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THE ATMOSPHERE OF VENUS

V. I. MOROZ

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CONTENTS

CONSIDERABLE progress was made in recent years in the investigations of Venus. The largest contributions were made by the results of direct measurements with the aid of the space probes used to sound the atmosphere of Venus in 1967 and 1969, namely Venera-4, Venera-5, and Venera-6. Interesting data were obtained as a result of the flight of the American space probe Mariner-5.* Earth-based investigations also contributed much that is new. Only on the basis of an analysis of the entire aggregate of the results obtained by different methods is it possible to understand the nature of this planet, and although the picture is still incomplete, it will be useful to make a certain summary. The present review covers literature published to the end of 1969 and partly in 1970.

1. CHEMICAL COMPOSITION

a) Chemical gas analyzers and spectroscopy—relative advantages and disadvantages. The landing modules (LM) of V4, V5, and V6 employed simple gas analyzers, based principally on the use of chemical absorbers[1,2]. Such an instrument consists of two identical volumes, one of which contains a substance absorbing the investigated gas, while the other does not absorb this gas. Both volumes are simultaneously filled with atmosphere samples, and the pressure is then measured in them. The difference is equal to the partial pressure of the investigated gas. The threshold sensitivity of the method is of the order of 1%. More subtle devices, such as mass spectroscopy and gas chromatography, could not be used during the first stage, owing to the peculiarities of the LM (small payloads, g-forces, low information contents).

Earth-based spectroscopy in its most developed form (Fourier spectrometry with high resolution, spectrographs at the Coude foci of large telescopes) makes it possible in individual cases to observe impurities with relative contents down to 10^{-8}, but it turned out to be almost powerless when it came to solving the problem of the main component of the atmosphere of Venus. Until recently, if we disregard the work of Connes et al.[3], it was assumed that CO_2 constituted not more than 10% of the atmosphere of Venus, and the remainder was assumed to be nitrogen (see, for example,[4,5]).

A spectroscopic determination of the absolute contents of the gases in the atmosphere of Venus is made difficult by the cloud layer. The cloud layer complicates the process of formation of the spectral lines, and we know too little concerning it to be able to account for this influence correctly. Usually one of two idealized models is used.

1) Simple reflection model. It is assumed that the cloud layer has a sharp boundary, optically equivalent to a solid surface, and that the absorption bands are formed in a purely gaseous atmosphere that lies above this boundary (Fig. 1a). The application of ordinary growth curves (see, for example,[4-6] to the measured intensities of the lines and bands use the amount u of the absorbed gas in a vertical column of unit cross section above the cloud layer (expressed in cm-atm, m-atm, or km-atm) and the total pressure p_c on the upper boundary of the clouds.

2) Model with scattering (Fig. 1b). It is assumed that the absorption bands are formed inside the cloud layer. Exact growth curves[7,8] exist for certain particular cases (isotropic scattering) and make it possible to determine the parameter

$$K=\frac{n_a}{\sigma} \tag{1.1}$$

which is the ratio of the concentration of the absorbing atoms to the volume scattering coefficient (cm^{-1}) and

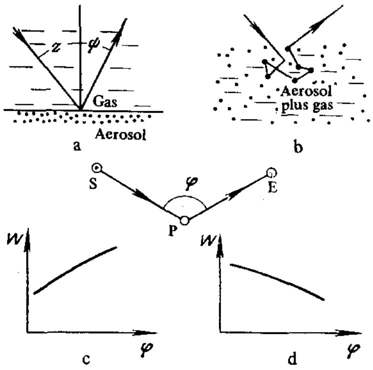

FIG. 1. Conditions for formation of spectral lines in the simple-reflection model (a) and in the model with scattering (b); c and d—dependence of the line intensity on the phase angle φ for both models.

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*For brevity we shall designate these space probes V4, V5, V6, and M5.

the average pressure p_c' inside the cloud layer. One of the obligatory additional assumptions is that the parameter K is constant over the entire thickness of the atmosphere. The result depends appreciably on the choice of the scattering indicatrix; if it is not spherical, one can use similarity relations[9] or approximate solutions[4,10] based on Rozenberg's theory[11].

The criterion for the choice between the two models is the dependence of the equivalent width (or of the total intensity) of the absorption band on the phase angle φ (the sun—planet—observer angle). Figures 1c and 1d demonstrate the form of this dependence for models 1 and 2. The observations[4,10,12,13] correspond to Fig. 1d, and therefore the model with scattering is assumed to be better founded.

The relative contents of different gases are obtained from spectroscopy of the atmosphere of Venus with good accuracy independently of the model. It suffices to choose the lines of two gases, close in intensity and in wavelength. At equal intensities we have

$$\frac{n_1}{n_2}=\frac{S_2}{S_1}, \tag{1.2}$$

where n_1 and n_2 are the concentrations and S_1 and S_2 the integral absorption coefficients of the compared gases. It is necessary to introduce also a small correction that depends on the line width.

b) $\underline{CO_2}$. Carbon dioxide was discovered in the atmosphere of Venus by Adams and Dunham (1932, see[14]). The gas analyzers on V4, V5, and V6 (see Table I) have shown that CO_2 is the main component of the atmosphere of this planet[1,2,15,16].

In the region 1—2.5 μ, the spectrum of Venus contains a large number of CO_2 bands (Fig. 2). From the weak bands one obtains in the simple-reflection model

$$u_{CO_2}\approx 1\ \text{km}\cdot\text{atm} \tag{1.3}$$

(Kuiper, see[14]) and the total pressure on the boundary of the cloud layer is

$$p_c\approx 0.15\ \text{atm}. \tag{1.4}$$

Table I. Chemical composition of the atmosphere of Venus
Molecules observed reliably

Molecule	Direct measurements of f		Spectroscopy				Note
	$V4^{1,2}$	$V5, V6^{15,16}$	f	u, cm-atm**	Band or line μ	Author	
CO_2	0.90 ± 0.10	0.95 ± 0.02	≈ 1 *)	10^5	0.78; 0.86	14	
CO	—	—	10^{-4}	13	2.35	3, 4, 19, 20	
HCl	—	—	$6\cdot10^{-7}$	$2\cdot10^{-2}$	1.74	3	
HF	—	—	$5\cdot10^{-9}$	$2\cdot10^{-4}$	1.29; 2.44	3	
H_2O			0—10^{-4}	0—0.012	0.82; 1.13; 1.38; 1.9	22–27	Level 0.15 atm
	$1\cdot10^{-3}$	9—$25\cdot10^{-3}$				15, 16	0.6
	0.5—$7\cdot10^{-3}$					1, 2	1.5
			$8\cdot10^{-3}$		$1.35\cdot10^4$	32–34	3—10

*In the simple-reflection model.
**Given in $g\text{-}cm^{-2}$ for H_2O.

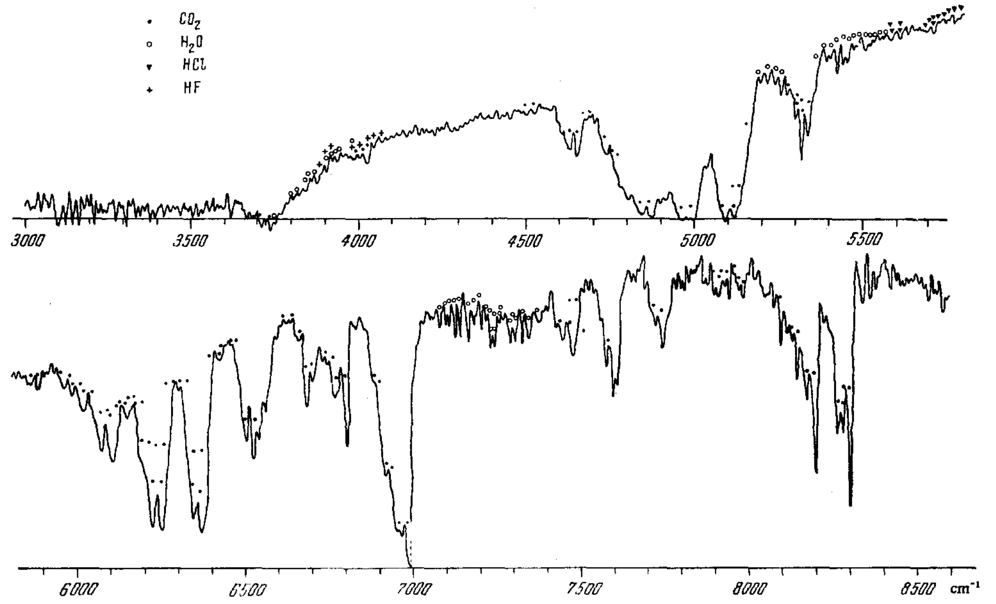

FIG. 2. Spectrum of Venus, obtained with the aid of a Fourier spectrometer on a high-flying airplane [21]. The Venus/moon intensity ratio is given, so as to exclude residual effects of the earth's atmosphere. The abscissas represent the wave numbers. All strong absorption bands belong to CO_2.

A correct estimate of p_c based on the growth curves was obtained only very recently[3], and gives a relative concentration

$$f_{CO_2}=\frac{p_{CO_2}}{p_c}\approx 1, \qquad (1.5)$$

which agress with the direct measurements. This agreement is hardly accidental; it is possible that the simple-reflection model is suitable for use in the spectroscopy of Venus, while a phase dependence of the form shown in Fig. 1d is not due to scattering at all but to the macrostructure of the cloud layer[17,18].

c) CO. The $\lambda 2.35\ \mu$ band of CO (first overtone) was observed by Sinton[19] and Moroz[20] with a resolution 50–100 Å. Connes et al.[3] obtained it by means of a Fourier spectrometer with a resolution 0.4 Å. According to[3,4,7,20]

$$f_{CO}\approx 10^{-4} \qquad (1.6)$$

with an uncertainty by a factor of two.

d) HCl and HS. Observation of such exotic compounds in the atmosphere of Venus turned out to be a sensation. It was made possible only by the use of the Connes Fourier spectrometer. According to[3]

$$f_{HCl}\approx 6\cdot 10^{-7}, \qquad f_{HF}\approx 5\cdot 10^{-9}. \qquad (1.7)$$

e) H_2O. The chemical gas analyzers on V4, V5, and V_6 gave a relative water-vapor concentration

$$f_{H_2O}\approx 0.01 \qquad (1.8)$$

at a level where $p\approx 1$ atm. There were many spectroscopic determinations[21-27], but the results were contradictory. Some authors indicate only the upper limit[21,22,27], which is sometimes very low (down to 1 μ of precipitated water), while others give large values near 100 μ[23-36]. However, even these relatively large quantities correspond to small relative concentrations

$$f_{H_2O}\approx 10^{-4}. \qquad (1.9)$$

Of course, these data pertain to the cloud layer, but the contradiction is not eliminated, since at the temperature of the cloud layer $T\approx 240-250°K$ (see Chap. 2) under saturation conditions, f_{H_2O} is larger by at least one order of magnitude than the estimate (1.9).

In the radio band, water vapor has an absorption line λ 1.35 cm. The radio emission spectrum of Venus, however (Fig. 3), has not been sufficiently well determined. Some authors[28,29] find this line, and others[30,31] do not. The upper limit for the 1.35-cm line and the continuous spectrum in the region $\lambda < 10$ cm[32-34] is

$$f_{H_2O}< 0.01. \qquad (1.10)$$

f) O_2. According to the gas-analyzer data from V5 and V6, the upper limit of the O_2 content is

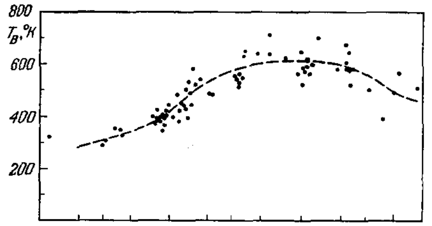

FIG. 3. Spectrum of radio emission of Venus. The accuracy of the individual points relative to T_B is on the average 10–20%.

$$f_{O_2}< 4\cdot 10^{-3}. \qquad (1.11)$$

Spectroscopy[35-37] gives an even lower value

$$f_{O_2}< 4\cdot 10^{-5}. \qquad (1.12)$$

Prokof'ev and co-workers[38,39] do not refute this upper limit, but assume that their spectrograms contain very weak Doppler satellites of the O_2 lines, belonging to the atmosphere of Venus.

g) N_2, inert gases. Direct measurements[1,2] yielded for them the upper limit

$$f_{N_2}+f_I\leqslant 0{,}1; \qquad (1.13)$$

and somewhat lower values were even obtained with V5 and V6. Kozyrev[40] reported emissions of N_2^+ in the spectrum of the ashen light of Venus (see Chap. 6), but the ambiguity of this identification was later demonstrated[41].

h) Other gases. In Table I we list information on the contents of gases definitely present in the atmosphere of Venus. For the remaining molecules there are upper limits (spectroscopic, with the exception of $N_2 + I$). They are given in Table II.

Lewis[48] calculated the upper limits for a number of molecules, starting from the assumption that the atmosphere is in chemical equilibrium with the lithosphere. The results are given in Table III. They are of illustrative character, for the calculations were based on the assumption of a surface pressure $p_s = 6$ atm and on the presence of large amounts of N_2 (66% N_2, 33% CO_2).

2. TEMPERATURE, PRESSURE, DENSITY

a) Direct measurements. On the landing modules of V4, V5, and V6, the temperature was measured with resistance thermomenters and the pressure with aneroid pickups[49,50]. An ionization density meter was used to measure the density on V4[49], and a tuning-fork meter was used on V5 and V6[15]. The results of V4 are reported in[49,50,51], and those of V5 and V6 in[52]. Direct experiment yielded T, p, and ρ as functions of the time during the course of the parachute

Table II. Chemical composition of the atmosphere of Venus. Spectroscopic upper limits

Molecule	$u<$, cm-atm*	$f<$	Spectral region, μ	Laboratory pressure, atm	Author	Reduced thickness in the earth's atmosphere, cm-atm
N_2 + inert gases	7	0,07	—	—	15, 16	$6{,}4\cdot 10^5$
O_2	4	$4\cdot 10^{-5}$	0,77	?	37	$1{,}6\cdot 10^5$
O_3	$5\cdot 10^{-4}$	$5\cdot 10^{-9}$	0,25	?	42	0,3
C_3O_2	0,05	$5\cdot 10^{-7}$	0,2–0,3	?	44	
CO	0,001	10^{-8}	2,44	0,93	44	
NH_3	0,001	10^{-8}	1,5–2,5	?	44	
H_2S	2	$2\cdot 10^{-4}$	1,58; 1,94	?	43	
SO_2	0,003	$3\cdot 10^{-8}$	0,3	0,04	45	
NO	0,1	10^{-6}	0,2–0,3	?	42	
NO_2	$6\cdot 10^{-2}$	$6\cdot 10^{-8}$	0,2–0,3	?	42	
N_2O	2	$2\cdot 10^{-5}$	2,1; 2,25	1	**)	0,8
HCHO	0,3	$3\cdot 10^{-6}$	0,3	?	46	
CH_4	0,1	10^{-6}	2,32	?	3	1,7
$(CH_2)_2$	3	$3\cdot 10^{-5}$	2,32	1	21	
$(CH_3)_2$	2	$2\cdot 10^{-5}$	2,35	0,09	21	
C_2H_2	0,1	10^{-6}	1,5–2,5	?	3	
HCN	0,1	10^{-6}	1,5	?	3	
CH_3F, CH_3Cl	0,1	10^{-6}	1–2,8	?	3	

*Simple reflection model, correction for air mass taken into account.

**Obtained by us by comparing the spectrum of Venus observed from an airplane (Kuiper et al. [21]) with the laboratory spectrum (Cruikshank [47]).

Table III. Chemical composition of the atmosphere of Venus. Theoretical upper limits[48]

Molecule	f/f_{CO_2}	Note*	Molecule	f/f_{CO_2}	Note*
H_2	$7 \cdot 10^{-6}$	A	C_2H_6	$3 \cdot 10^{-17}$	A
COS	$1{,}5 \cdot 10^{-6}$	B	SiF_4	$1 \cdot 10^{-19}$	B
H_2S	$3 \cdot 10^{-7}$	B	S_2O	$4 \cdot 10^{-20}$	A
NH_3	$3 \cdot 10^{-7}$	B	CH_3F	$3 \cdot 10^{-21}$	A
CH_4	$4 \cdot 10^{-8}$	A	SO	$5 \cdot 10^{-22}$	A
CS_2	$4 \cdot 10^{-11}$	A	S	$6 \cdot 10^{-26}$	A
S_2	$3 \cdot 10^{-13}$	A	Cl_2	$6 \cdot 10^{-27}$	A
HCN	$4 \cdot 10^{-14}$	A	SO_3	$8 \cdot 10^{-32}$	A
C_2H_6	$4 \cdot 10^{-15}$	A	N_2O	$1 \cdot 10^{-32}$	A
NO	$1 \cdot 10^{-16}$	A	S_8	$3 \cdot 10^{-35}$	A
SO_2	$1 \cdot 10^{-16}$	A	O_2	$5 \cdot 10^{-35}$	A

*A—calculated from equilibrium with other gases; B—calculated from equilibrium with minerals.

descent of the landing module. The absolute altitude reconciliation on V5 and V6 was with the aid of measurements with a radar altimeter. The relative heights were determined by integrating the equation of hydrostatic equilibrium

$$dp = -\rho g \, dz. \tag{2.1}$$

This gives for the instants of time t_1 and t_2 and altitude difference

$$z(t_1) - z(t_2) = -\frac{1}{g} \int_{p_1}^{p_2} \frac{dp}{\rho}. \tag{2.2}$$

The same difference can be obtained independently from the equation of motion

$$Mg = A\rho v^2/2, \tag{2.3}$$

whence

$$z(t_1) - z(t_2) = \left(\frac{2Mg}{A}\right)^{\frac{1}{2}} \int_{t_1}^{t_2} \frac{dt}{\rho^{1/2}}, \tag{2.4}$$

where M is the mass, v = dz/dt the velocity, and A the aerodynamic cross section of the LM.

One can determine the velocity of the vertical motions by differentiating and subtracting the right-hand sides of (2.2) and (2.4). Measurements gave an upper limit on the order of several $m\text{-}sec^{-1}$. Since T, p, and ρ are connected by the equation of state

$$p = (\rho / m_H \bar{\mu}) kT, \tag{2.5}$$

we can determine the average molecular weight $\bar{\mu}$, or, if it is known from other measurements (chemical composition), we can use any two parameters to monitor the third. At sufficient measurement accuracy it is possible, in principle, to determine g(R) and the distance R to the center of the planet from Eqs. (2.2)–(2.5)[53].

Figures 4 and 5 show the functions p(z) and T(z), obtained from measurements on V4, V5, and V6. Figure 6 shows the adiabats for pure CO_2 and a mixture 90% CO_2 + 10% N_2. The dependence of c_p on p and on T was taken from tables[56,57]. The adiabats approximate the observations well at T < 350°K, the latter being in better agreement with the experimental curves than the former. The adiabatic gradient in the region T < 350°K denotes that at large depths the atmosphere is in convective equilibrium (at least to a level T = 600°K). The measurements pertain to the night side of Venus.

Altimeter signals from V5 and V6 show that both stations ceased to operate at an altitude 20 ± 7 km. Extrapolation along the adiabat yields a surface temperature

$$T_s = 750 \pm 40^\circ \mathrm{K} \tag{2.6}$$

and a pressure

$$p_s = 93 \pm 25 \text{ atm}. \tag{2.7}$$

The lower limit of the temperature is

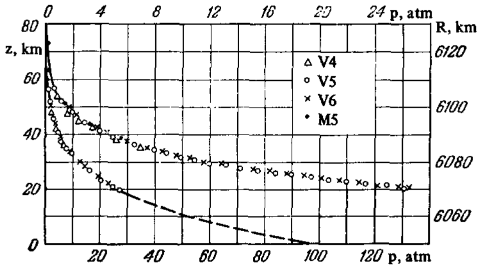

FIG. 4. Pressure as a function of the altitude as measured with V4, V5, V6 and M5. The upper and lower scales are for the upper and lower curves, respectively. Left ordinates—height, right ordinates—distance to the center. Extrapolation downward along the adiabat of 90% CO_2, 10% N_2. The measurement data were taken from [15].

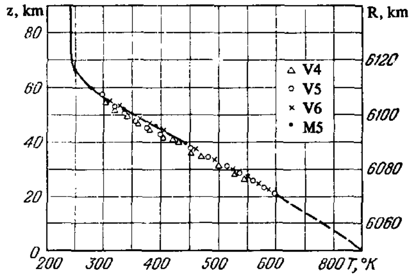

FIG. 5. Temperature as a function of the altitude as measured on V4, V5, V6, and M5 [15].

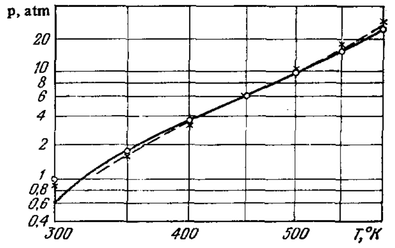

FIG. 6. The function P(T) as measured on V4, V5, V6. Solid line—V4 and V5, dashed—V6, crosses—adiabat for pure CO_2, circles—for the mixture 90% CO_2 + 10% N_2.

$$T_s = 600°\text{K} \tag{2.8}$$

and the upper limit of the pressure (from isothermal extrapolation at $T_S = 600°\text{K}$) is

$$p_s = 125 \pm 40 \text{ atm}. \tag{2.9}$$

b) Refraction measurements. An electromagnetic wave passing through an atmosphere (see Fig. 7) with refractive index m, experiences as a result of the refraction a phase shift

$$\Phi(R) = \frac{1}{\lambda} \int\limits_{-\infty}^{+\infty} (m-1)\, dx \tag{2.10}$$

(the integral is taken here along the line of sight) and is deflected and attenuated. The attenuation is due to the fact that the divergence of the beam increases. At small deflection angles, the attenuation in the isothermal atmosphere is equal to

$$f(R) = \frac{E_0}{E} = 1 + \frac{2L_2\,(L_1 - L_2)}{L_1 H}\, n_0 c_0 e^{-\frac{z}{H}} \left(\frac{\pi R}{2H}\right)^{\frac{1}{2}}, \tag{2.11}$$

where E_0 is the illumination in the absence of refraction, E the illumination attenuated by the refraction, L_1 the distance from the source to the observer, L_2 the distance from the planet to the observer, n_0 the numerical concentration of the molecules in the base of the atmosphere (or at some reference level), $c_0 = (m-1)/n_0$, m is the refractive index corresponding to the numerical concentration n_0, R is the shortest distance from the center of the planet to the line of sight, z is the height over the reference level, and H is the altitude scale.

By measuring $\Phi(R)$ or $f(R)$, we can obtain $n(R)$ and H. In practice this method was used twice: during the occultation of Regulus by Venus in 1960, and during the flight of M5 behind the disk of Venus in 1967. During the occultation of Regulus the function $f(R)$ was obtained in the region z = 120 km (R = 6170 km). A reduction of the measurements[58,60] yielded

$$n \approx 6 \cdot 10^{13} \text{ cm}^{-3} \quad (z = 120 \text{ km}),\ H = 6 \pm 2 \text{ km} \tag{2.12}$$

whence

$$190 < T < 380°\text{K}, \tag{2.13}$$

if CO_2 is not dissociated and $\bar{\mu} = 42.4$ (90% CO_2 + 10% Ne).

The concentration estimate (2.12) corresponds to the level at which $f(R) = 2$. In the case of refraction of radio waves from a spaceship, the factor $(L_1 - L_2)/L_1 \approx 10^4$, and a similar attenuation is obtained at concentrations 5×10^{17} cm^{-3}. This makes it possible to sound deeper layers of the atmosphere, and on M5 the $n(R)$ dependence was obtained down to the level of critical refraction ($P \approx 4.5$ atm), where the radius of curvature of the beam is equal to R. The phase shift $\Phi(R)$ was measured during the "radio occultation" of M5. Figures 4 and 5 show the results of the reduction of these measurements, in the form of the functions $p(R)$ and $T(R)$[60,61]. They are in splendid agreement with the direct measurements, if one assumes the radius of the solid surface

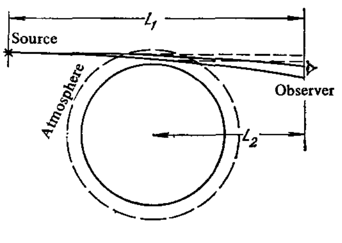

FIG. 7. Diagram of refraction experiment.

$$R_s = 6050 \text{ km} \tag{2.14}$$

which was obtained from earth-based radar measurements[62-65].

The upper isothermal section of the Mariner curve, corresponding to T = 240°K, causes doubts. In fact, the temperature of the isothermal stratosphere is equal to

$$T_{sf} = \frac{T_e}{\sqrt[n]{2}}, \tag{2.15}$$

where T_e is the equilibrium temperature of the planet and n = 4, if the absorption coefficient for the thermal radiation of the planet does not depend on the wavelength (the gray approximation). The equilibrium effective temperature is calculated from the condition

$$\sigma T_e^4 = \frac{1}{4}(1 - A_i)\frac{E_\odot}{r^2}, \tag{2.16}$$

where $E_\odot$ is the solar constant, r the distance to the sun in astronomical units, and A_i the integral albedo. According to Irvine[66] we have for Venus

$$A_i = 0{,}77 \pm 0{,}07, \tag{2.17}$$

whence

$$T_e = 238 \pm 16°\text{K} \tag{2.18}$$

and

$$T_{sf} = 192 \pm 12°\text{K}. \tag{2.19}$$

If the heat balance in the real stratosphere of Venus is determined by radiative transport in the λ 15 μ CO_2 band (see Ch. 6), then the stratosphere is not isothermal: it is warmer in the lower part and colder in the upper than in the gray approximation. Thus, an isothermal stratosphere with a temperature 240°K cannot be explained from the point of view of heat balance. If the stratosphere is isothermal, then it should be colder (approximately 200°K), and if it has in the lower part a temperature of 240°K, then there will be no isothermy.

c) Infrared temperatures. 1) Region of thermal radiation ($\lambda > 3\ \mu$). The estimate (2.16) represents the calculated temperature of the outgoing radiation. Let us compare it with the measured value. In the region $\lambda > 3\ \mu$, measurements were made with filters (see[4]), and spectral measurements were also made in the 8–14 μ window[69-72]. In the 8–14 μ region, the average brightness temperature of the disk[67,78] is

$$\bar{T}_B = 220 \pm 10°\text{K}, \tag{2.20}$$

i.e., it is close to T_e (see (2.18)). At the center of the disk we have

$$T_B = 230 \pm 10°\text{K}. \tag{2.21}$$

The darkening towards the limb is in accord with the law

$$T_B \propto \sqrt[8]{\cos\theta}, \tag{2.22}$$

where θ is the angle between the normal and the direction to the observer, down to $\cos\theta = 0.5$[68,69,71], and at $\cos\theta < 0.5$ it becomes still smoother[80]. The nighttime temperature hardly differs from the daytime temperature, and in accordance with certain observations it even exceeds it by several degrees[79].

The color temperature in the 8–14 μ region coincides with T_B. Individual measurements exist in the 3–4 μ[81] and 18–22 μ windows (Low, private communication). They give close values of T_B. Consequently, the absorption coefficient differs little in a very wide range of wavelengths, something that can occur only if the absorption is due to solid or liquid particles. In other words, the infrared brightness temperature is the temperature of the cloud layer (with a small correction necessitated by the emissivity). The angular dependence (2.22) is determined mainly by the angular dependence of the emissivity[82–84]. The contribution of the temperature gradient is small if it plays any role at all.

2) Region of reflected radiation. It is possible here to determine the rotational and vibrational temperatures of the CO_2 bands. The mean value as determined from the weak λ 8689 Å band[85] is

$$T_r = 291 \pm 10°\mathrm{K} \qquad (2.23)$$

in the model with scattering and

$$T_r = 317 \pm 10°\mathrm{K} \qquad (2.24)$$

in the simple-reflection model. Sometimes, lower values are obtained, close to 240°K[3,86,87] and sometimes higher, up to 450°K (only from the very weak bands near 0.8 μ[88]). The scatter is apparently due to time variations of the effective depth of formation of the bands.

d) Spectroscopic and polarization determinations of the picture. Spectroscopic pressures can be determined from the growth curve and from the line width. The growth curve gives an effective pressure

$$p_e \approx 0.1 - 0.2\ \text{atm}. \qquad (2.25)$$

Regardless of the model, the line widths turn out to be as a rule much narrower than the apparatus contour (<0.1 cm⁻¹), but in some cases[88] large widths are observed, corresponding to pressures of several atmospheres. A correlation was observed here with the rotational temperatures, and it is possible to use such data to construct the p(T) diagram; it agrees with the direct measurements, although the scatter is large (Fig. 8).

Plots of polarization of Venus against the phase in ultraviolet rays have a maximum near $\varphi = 90°$, which is characteristic of Rayleigh scattering[89–92]. The magnitude of this maximum gives the estimate

$$p_c \approx 0{,}02 - 0{,}05\ \text{atm}. \qquad (2.26)$$

in the simple-reflection model.

e) Radio emission of Venus. Radar. A review of the results of investigations of the radio emission of Venus and of radar experiments up to 1965–1966 can be found in[3,93]. Table IV gives results, known to the author, of measurements of the brightness temperature of the Venus in the radio band, averaged over the disk, that were obtained subsequently and not included in the cited reviews. Complete data on the observed spectrum of the radio emission of Venus are shown in Fig. 3.

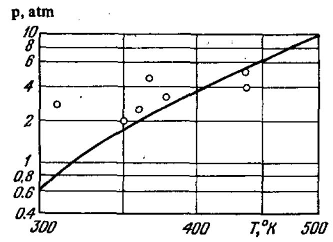

FIG. 8. The function p(T) as obtained from direct measurements (continuous line) and from spectroscopic measurements (circles).

It is seen from Fig. 3 and Table IV that the brightness temperature at centimeter and decimeter wavelengths is much higher than given by optical methods. It had already become clear in the early sixties that this fact can be attributed only to a high temperature of the surface and of the lower layers of the atmosphere[3,93].

The most important progress in the theoretical interpretation of the spectrum of the radio emission of Venus was attained recently as a result of laboratory investigations of the absorption of radio waves in CO_2, N_2, and H_2O at high temperatures and pressures (Ho, Kaufman, and Thaddeus[106]). The absorption is connected with rotational and translational transitions induced by pressure, and the absorption coefficient of the radio waves in a mixture of CO_2, N_2, and H_2O is

$$\alpha = p^2\nu^2 \left(\frac{273}{T}\right)^5 (15.7 f_{CO_2}^2 + 3.90 f_{CO_2} f_{N_2} + 0.085 f_{N_2}^2 + 1330 f_{H_2O}) \cdot 10^{-8}\ \mathrm{cm}^{-1}, \qquad (2.27)$$

where p is the pressure (atm), ν is the wave number (cm⁻¹), and f_{CO_2}, f_{N_2}, and f_{H_2O} are the relative contents. The brightness temperature at a certain point of the planetary disk is

$$T_B(\mu, \gamma, \lambda) = \varepsilon(\mu, \gamma, \lambda)\, T_s e^{-\frac{\tau_0(\lambda)}{\lambda}} + \int_0^{\tau_0} \frac{T(\tau)}{\mu} e^{-\frac{T(\tau)}{\mu}} d\tau + C, \qquad (2.28)$$

where $\tau(\lambda) = \int_0^x \alpha(\lambda, x)\,dx$—is the optical thickness reckoned downward, μ the cosine of the angle between the normal and the direction to the observer, τ_0 the total optical thickness of the atmosphere, T_S is the surface temperature, $\epsilon(\mu, \gamma, \lambda)$ is the emission coefficient of the surface, and γ is the angle between the direction of the polarization of the receiving system and the line joining the center of the disk with the radiating element. The first term represents the surface radiation attenuated by the atmosphere, the second represents the radiation of the atmosphere, and C is the increment due to reflection of the radiation of the atmosphere from the surface. The $\epsilon(\mu, \gamma)$ dependence is calculated from the Fresnel formula. The

Table IV. Brightness temperature of Venus in the radio band (results not included in[3,93])

λ, cm	$\overline{T}_B$, °K	Reference	λ, cm	$\overline{T}_B$, °K	Reference
0,86	423	94	4,52	656±35	99
1,18	400	95	6,0	706±45	100
1,28	418	95	6,0	635±30	101, 102
1,35	436	95	7,5	617±18	97
1,43	451	96	21,2	591±30	103
1,58	477	95			
2,07	485+60	96	21,3	+72 646 −100	104
3,12	553±21	97			
3,75	646+72 −100	98	49,1	580±100	105

brightness temperature averaged over the disk is

$$\bar{T}_B = \int_0^1 \int_0^{2\pi} T(\mu, \gamma)\, \mu\, d\mu\, d\gamma. \qquad (2.29)$$

At large τ_0, only the second term of (2.28) is effective, and in the approximation of a semiinfinite atmosphere we can assume

$$\bar{T}_B = \frac{3}{2} \int_0^{\tau_0} T(\tau)\, e^{-\frac{3}{2}\tau}\, d\tau. \qquad (2.30)$$

The maximum contribution to the integral of (2.30) is made by the layer at which $\tau \approx \tau_{\text{eff}} = 2/3$. The smaller the wavelength, the smaller the corresponding τ_{eff}. This is the reason why the brightness temperature decreases with wavelength in the 5–0.1 cm range (see Figs. 3 and 11). At this level of p and T we can assume

$$\alpha_{\text{eff}} = \frac{\tau_{\text{eff}}}{H} = \frac{2}{3}\, \frac{m_H \bar{\mu} g}{kT} \approx \frac{3 \cdot 10^{-4}}{T}. \qquad (2.31)$$

From (2.27) and (2.31) it follows that

$$p \left(\frac{273}{T}\right)^2 = \frac{a}{\nu}, \qquad (2.32)$$

where the constant $a = a(f_{CO_2}, f_{N_2}\ f_{H_2O})$ is a function of the composition only. Table V lists the pressures corresponding to $\tau = 2/3$, as functions of the wavelength, calculated from formula (2.32). They determine the effective altitude level of radiation of given wavelength. Putting $T = T_B$ in (2.32), we can regard this table as the function p(T) obtained from the radio-emission spectrum, and verify the validity of the employed theory by comparing it with the direct measurements, with a particular check of the premises employed in the theory concerning the chemical composition. Comparison with Fig. 6 shows that the agreement is satisfactory, especially if one takes the third column, which differs from the two others by the addition of 1% H_2O. The agreement is violated at $\lambda > 5$ cm, where the approximation of a semiinfinite atmosphere is not sufficiently exact. In the region $\lambda \geq 10$ cm, the main source of the radiation is already the surface.

Usually the problem is solved in reverse order: one specifies the model of the atmosphere (in accordance with direct measurements), the temperature of the surface (from the radio emission in the decimeter band) and its dielectric constant (obtained by radar), and the T_B spectrum is calculated in accordance with the rigorous formula (2.29) for some particular composition[32,33,107,108]. The results are as a rule the same: to reconcile the calculated spectrum with the observed one it is necessary to have in the atmosphere 0.5–1% H_2O vapor. It is clear from (2.27) that this small admixture of H_2O noticeably increases the absorption coefficient.

The authors of[32,33,108] use for the H_2O absorption coefficient expressions different from (2.27), but this does not influence the final result very strongly.

The maximum of T_B in the observed radio-emission spectrum occurs in the region 5–15 cm. At longer wavelengths, T_B decreases. In the theoretical spectrum, no maximum is obtained at all at dT/dz = const, if we disregard reflection from the surface (the term C in (2.28)). Allowance for C produces a certain "small hump" in the required region, but its magnitude is insufficient to reconcile the result with the observations. Agreement can be obtained by assuming either a decrease of ϵ at decimeter wavelengths[33] or a decrease of the gradient dT/dz in the surface layer[107-110]. According to Gale et al.[107], it suffices to assume the existence near the surface of an isothermal layer (T = 670 ± 20°K) with thickness

$$z_{is} = 7 \pm 2 \text{ km} \qquad (2.33)$$

A close result was obtained by Vetukhnovskaya et al.[108] and by Rzhiga[109].

Independent information concerning the absorption of radio waves in the atmosphere of Venus can be obtained by radar. Figure 9 shows the dependence of the effective reflection cross section σ_e of the planet on the wavelength[110,111]. This dependence yields the

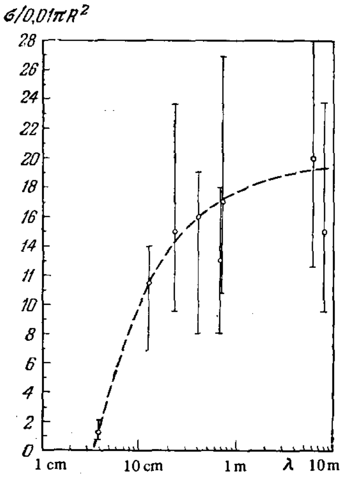

FIG. 9. Radar cross section of Venus (in units of $0.01\pi R^2$) as a function of the wavelength.

Table V. Pressure at the altitude level $\tau = 2/3$ as a function of the wavelength (effective altitude level of the emission yield in the approximation of a semiinfinite atmosphere)

ν, cm^{-1}	λ, cm	T_B, °K	Effective pressure, atm		
			pure CO_2	90% CO_2 + 10% N_2	90% CO_2 + 9% N_2 + 1% H_2O
100	0.01	(200) *)	(0.015)	(0.016)	(0.011)
10	0.1	280	0.09	0.1	0.07
5	0.2	300	0.6	0.7	0.5
2	0.5	350	2.3	2.5	1.8
1	1	400	5.7	6.3	4.5
0.5	2	550	22	24	17
0.2	5	600	64	70	45

*There are no observations. A stratosphere temperature close to the estimate (2.19) is assumed.

total optical thickness of the atmosphere as a function of the wavelength (see, for example,[112]):

$$\tau_0(\lambda) = 2\ln\frac{\sigma_e(\lambda)}{\sigma_e(\infty)} = (1.01 \pm 0.04)\left(\frac{3.8}{\lambda}\right)^2. \qquad (2.34)$$

From observations at 10 cm with an interferometer, carried out in two polarizations[113] it is possible to estimate τ_0 (10 cm) = 0.17 ± 0.03[108] (see also[112]), in splendid agreement with (2.34).

A joint analysis of the spectrum of the radio emission and of the radar cross sections makes it possible to obtain sufficiently reliable conclusions concerning the temperature and pressure at the surface (Table VI). Model (c) of Table VI, which implies a 100° gap between the temperatures at the surface and at the lower limit of the atmosphere (inversion: the atmosphere is hotter than the surface), is extremely unlikely.

Table IV and Fig. 3 show values of T_B at inferior conjunction, i.e., *for the nighttime side of the planet*. What is the difference in T_B between the nighttime and daytime sides, i.e., is there a dependence of T_B on the phase? At 10 cm, where the main contribution is made by the surface, there is no difference within 2–3%[93]. At centimeter and millimeter wavelengths, the data are contradictory, and will not be discussed here in detail. The observations at 8 mm can serve as an example. Even recently, all experimenters obtained higher values of T_B on the daytime than on the nighttime side, and this was confirmed with the Pulkovo radio telescope by measurements of the brightness distribution over the disk[114]. Yet the latest observations of A. D. Kuz'min et al. with the Serpukhov 22-m antenna of the Physics Institute of the Academy of Sciences have shown the absence of a phase variation, within 1%, at this wavelength.

Another problem is whether there is a noticeable temperature difference between the pole and the equator. Several series of measurements were performed with an interferometer having a variable base. The results are likewise contradictory. Clark and Kuz'min[113] found, from observations at 10 cm, that the pole is 150°K colder than the equator, but later observations at 11 cm (Sinclair et al.[114a]) gave an upper limit of 15° for the difference between the pole and the equator. At 3 cm, Berge and Greison[97] obtained an upper limit of 30°.

We shall show in Ch. 5 that the thermal inertia and the general circulation of the atmosphere of Venus should smooth out the temperature differences to a few degrees in the lower layers of the troposphere, so that in the $\lambda \gtrsim 3$ cm. range there should be no noticeable phase variation or latitude dependence of T_B. However, at the millimeter wavelengths radiated by the higher layers of the troposphere, the possibility of inhomogeneities of this kind in the brightness temperature cannot be excluded by a priori considerations.

It is clear from (2.28) that even if the thermodynamic temperature is constant at all points of the planet (at a given z), the brightness temperature should vary from the center to the limb of the disk. If the surface radiates, T_B drops off towards the limb owing to the decrease of the emissivity toward the limb; if the atmosphere radiates, a similar effect is produced by the vertical temperature gradient. The variation of T_B from the center towards the limb was investigated at 8 mm[114] and 1.35 and 1.9 cm (with Mariner-2, the first American space ship to fly by venus) and at 3 cm[97], 10 cm[113], and 21 cm[104].

f) Estimates of T_S based on chemical equilibrium. From the spectroscopic upper limits of H_2S and COS, Lewis[48] obtained T_S > 560°K assuming that the surface is isothermal. If the maximum surface temperature is T_S = 700°K, then, according to Lewis's estimate, T_S > 540°K at the coldest point.

g) Working model of the atmosphere up to an altitude of 120 km. Models of the atmosphere of Venus, published soon after the flights of V4 and M5 (see, for example,[3]), gave correct estimates of the temperatures of the surface and of the lower atmosphere (based

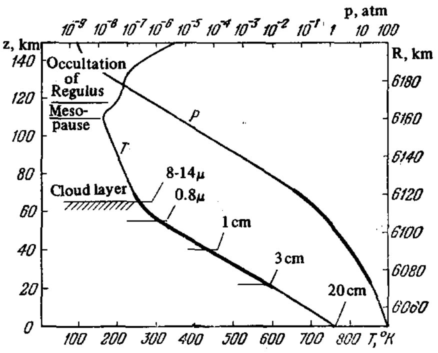

FIG. 10. Structure of the atmosphere of Venus up to an altitude of 120 km. The effective radiation altitude levels for different wavelengths are shown.

Table VI. Surface temperature T_S and pressure at the surfact p_S from radio-astronomy and radar data

Model of atmosphere	Author	T_s, °K	p_s, atm
a) Adiabatic down to the surface	Gale et al. [107] Vetukhnovskaya et al. [108]	720±20 700±50	— 65±20
b) Adiabatic with isothermal section at the surface	Gale et al. [107] Vetukhnovksaya et al. [108], Rzhiga [109]	670±20 650 650	100±20 —
c) Adiabatic with narrow inversion layer at the surface	Rzhiga [109]	650	—

on radio-astronomical data), but the uncertainty in the pressure was very high. A surface pressure close to 20 atm was assumed probable, but values 5 times larger or 5 times smaller could not be excluded. Successful investigations carried out with the aid of interplanetary automatic stations have made it possible to construct a much more reliable model of the atmosphere in the region of altitudes at which the measurements were performed[52]. In Table VII, where the model of the atmosphere of Venus from 0 to 120 km is given, the region reliably investigated by cosmic means is printed in boldface, and in Fig. 6 it is represented by a heavy line. This model agrees with all the results of the earth-based observations, both radio-astronomical and optical.

Since the vertical temperature profile in the atmosphere is determined by its heat balance, and the latter is determined to a considerable degree by the chemical composition, a refinement of the composition obtained in space experiments has made possible more reliable upward and downward extrapolations of the model. In the region below 20 km, both the radio-astronomical observations and the theory of heat balance (see Chap. 3) point either to an adiabatic temperature profile or to one close to adiabatic. We used an adiabat.

Above 70 km, we used the vertical profile obtained by McElroy from detailed calculations of the heat balance in the atmosphere above the clouds[175]. The profile calculated for 90% CO_2 and 10% N_2 was assumed. It differs very little from the profile calculated for pure CO_2. The models presented here and in[52] differ somewhat both in the character of the approximation of the experimental data and in the assumptions used during the extrapolation. In[52], in the altitude range $z < 45$ km, the gradient is somewhat higher than adiabatic. This excess is not realistic from the theoretical point of view, and at the same time does not go beyond the limits of measurement errors. Therefore below 45 km we approximated the measurement results with an adiabat (for the aforementioned composition). The model of[52] differs seriously in the mesopause region, which is warmer than assumed by McElroy. It can be shown that this leads to an incorrect value of the height of the maximum of the electron concentration in the ionosphere. The only experiment yielding direct information on the temperature of the mesopause of Venus is the occultation of Regulus, but as already indicated, its results have low reliability.

3. CLOUD LAYER

The cloud layer of Venus was investigated only by optical methods (photometry, polarimetry, and spectroscopy). They make it possible to estimate the average dimension of the particles, the refractive index, the scattering indicatrix, the albedo of single scattering, the average volume coefficient of scattering, and the total optical thickness, and give definite albeit not unambiguous material for estimating the chemical composition.

The radius of the outer limit of the cloud layer (the apparent radius of the planet), according to Vaucouleur[115] is $R_C = 6120 \pm 8$ km, so that its height is

$$z_c = 68 \pm 8 \text{ km} \tag{3.1}$$

In the visible region of the spectrum, the contrasts on the disk are very small. They are much larger near 3500 Å—dark spots are observed here with lifetimes on the order of several days[116-119]. Ultraviolet photographs give a larger radius of Venus[117]:

$$\begin{gathered} R_c(\text{UV}) = 6145 \text{ km or} \\ z_c(\text{UV}) \approx 95 \text{ km} \end{gathered} \tag{3.2}$$

The clouds reflecting in the near-ultraviolet region lie higher and are optically thin to visible light.

We do not know how deep the clouds extend downward. If they are produced by condensation of some single substance, the geometric thickness of the layer hardly exceeds 10 km. If several substances condense, or if the clouds (at least in the lower layers) consist of dust, they can extend all the way down to the surface.

a) Average dimension of particles and refractive index. These parameters were estimated from the phase variation of the brightness[122,123] (the dependence of the stellar magnitude on the phase angle φ) and the phase variation of the polarization. The second method

Table VII. Working model of the atmosphere of Venus up to an altitude of 120 km

z, km	T, °K	p, atm	ρ, g-cm^3	n, cm^{-3}	Note
0	750	93,0	$6,3\cdot 10^{-2}$	$8,8\cdot 10^{20}$	Surface
5	713	69,0	5,0	7,0	
10	675	50,3	3,8	5,4	
15	636	35,0	2,8	4,0	
20	**596**	**25,2**	**2,18**	**3,07**	End of operation of V5 and V6
25	**556**	**17,3**	**1,58**	**2,23**	
30	**515**	**11,5**	**1,15**	**1,62**	
35	**472**	**7,40**	**$8,10\cdot 10^{-3}$**	**1,14**	
40	**428**	**4,50**	**5,50**	**$7,74\cdot 10^{15}$**	
45	**382**	**2,60**	**3,58**	**5,05**	
50	**340**	**1,43**	**2,16**	**3,04**	
55	**304**	**0,71**	**1,25**	**1,70**	Start of operation of V4, V5, and V6
60	**274**	**0,34**	**$6,6\cdot 10^{-4}$**	**$9,0\cdot 10^{18}$**	
65	**250**	**0,14**	**2,7**	**3,6**	Upper limit of the cloud layer
70	240	$5,5\cdot 10^{-2}$	1,2	1,6	
80	220	$7,0\cdot 10^{-3}$	$1,7\cdot 10^{-5}$	$2,3\cdot 10^{17}$	
90	200	$7,4\cdot 10^{-4}$	$1,9\cdot 10^{-6}$	$2,6\cdot 10^{16}$	
100	180	$6,2\cdot 10^{-5}$	$1,8\cdot 10^{-7}$	$2,5\cdot 10^{15}$	
110	160	$4,9\cdot 10^{-6}$	$1,6\cdot 10^{-8}$	$2,2\cdot 10^{14}$	Mesopause
120	205	$3,9\cdot 10^{-7}$	$1,0\cdot 10^{-9}$	$1,4\cdot 10^{13}$	Occultation of Regulus

is more reliable. Sobolev[91] obtained from the polarization curves of Gehrels and Samuelsen[92], assuming m = 1.5, an average radius

$$\bar{r} \approx 1.2 \text{ atm} \tag{3.3}$$

and the indicatrix shown in Fig. 11. The coefficient of the second term of the expansion of the indicatrix in Legendre polynomials (it characterizes the degree of its elongation) is

$$x_1 = 2.1. \tag{3.4}$$

The model of a uniform semiinfinite atmosphere was used, and an indicatrix of the type $1 + x_1 \cos \gamma$ (see[124,125]) was assumed for scattering of higher orders, starting with the second. The homogeneity requirement means that the albedo of single scattering and the indicatrix do not vary with depth.

Figure 12 shows equal-polarization curves in coordinates φ and $1/\lambda$, obtained by Coffeen and Gehrels[89] from observations at 10 wavelengths. Coffeen[90] calculated the polarization curves by the Mie theory in coordinates φ and $1/\lambda$ for a wide range of values of the complex refractive index

$$m = m_0 + m_1 i. \tag{3.5}$$

Higher-order scattering leads principally to a decrease of the absolute values of the degree of polarization, but influences the position of the inversion points (at which the polarization reverses sign) only little. Comparing the calculated and observed inversion curves, Coffeen found

$$\bar{r} = 1.25 \pm 0.25 \text{ atm} \tag{3.6}$$

the real part of the refractive index

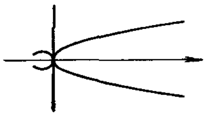

FIG. 11. Average scattering indicatrix of the particles of the cloud layer, calculated from polarization observations [126].

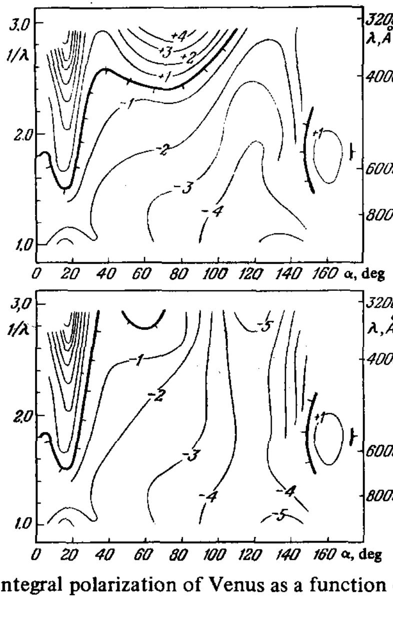

FIG. 12. Integral polarization of Venus as a function of φ and $1/\lambda$ [90].

$$1.43 < m_0 < 1.55 \tag{3.7}$$

and the upper limit of the complex part

$$m_1 < 10^{-3}. \tag{3.8}$$

Notice should be taken of the excellent agreement between the results of Sobolev and Coffeen, who used two entirely different approaches to the problem.

b) Albedo of single scattering. This quantity can be obtained directly from the spherical albedo A_B, if the atmosphere is assumed to be homogeneous and semiinfinite. According to Sobolev[126]

$$A_B = 1 - 4\sqrt{\frac{1-a}{3-x_1}}. \tag{3.8'}$$

Table VIII lists the values of A_B as given by Irvine[66] and the values of $1 - a$ calculated at $x_1 = 2.1$. We have neglected the dependence of x_1 on λ.

The fourth column gives the volume coefficient of absorption of the material of the particles

$$\varkappa \approx \frac{1}{2r}\frac{1-a}{a}. \tag{3.9}$$

Strong absorption takes place in the blue and in the ultraviolet (see also[127]). The accuracy of A_B is ±7%, and therefore K/σ and κ are determined only accurate to one half of an order of magnitude.

Figure 13 shows a plot of $A_B(\lambda)$, extrapolated towards longer and shorter wavelengths in accord with the spectroscopic data, without allowance for the difference in the phase curves, which were not determined outside the range 0.3–1.0 μ.

Table VIII. Spherical albedo of Venus[66] and albedo of single scattering

λ, Å	A_B	$1 - a \approx \frac{K}{\sigma}$	$\varkappa$, cm^{-1}
3 147	0.45	0.0165	66
3 590	0.52	0.0129	52
3 926	0.56	0.0109	42
4 155	0.66	0.0063	25
4 573	0.77	0.0030	12
5 012	0.79	0.0025	10
6 264	0.94	0.0002	0.8
7 227	0.93	0.0003	1.2
8 595	0.89	0.0007	2.8
10 630	0.89	0.0007	2.8

K and σ–volume coefficients of absorption and scattering.

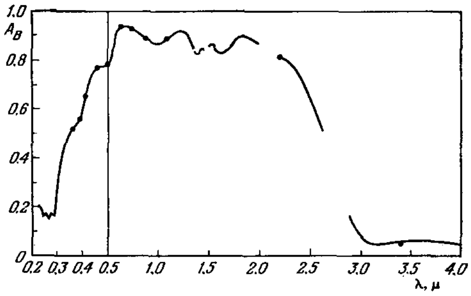

FIG. 13. Spherical monochromatic albedo of Venus in the range 0.2–0.4μ [120]. Ordinates–partial pressure of H_2O, calculated without allowance for condensation (complete mixing, three values of f_{H_2O}), the temperature (extrapolated in accordance with the McElroy model, see Chap. 6), and the saturated-vapor pressure.

c) Absorption or finite thickness? In terrestrial clouds, in the visible and in the near infrared (to 1.3 μ), the value of $1 - a$ is very small, about 10^{-7}[128]. If $1 - a = 0$, then the albedo of a semiinfinite ($\tau_0 = \infty$) scattering medium is $A_B = 1$. The difference between A_B and 1 can be attributed to two causes: 1) true absorption, 2) finite optical thickness τ_0. Actually both causes can be in operation, but we shall consider only these extreme cases. If the first possibility is realized, we obtain from the curve the values of $1 - a$ and κ, as was done above. If the second possibility is realized, we can estimate τ_0. We can use for this purpose the results of Romanova[129], who calculated the albedo of optically thick (but finite) layers for a scattering indicatrix close to that of Venus. The result is

$$\tau_0 = 65^{+50}_{-20}. \tag{3.10}$$

d) Average volume coefficient of scattering. This quantity can be estimated by two methods. 1) Since the thickness of the cloud layer Δz can hardly be smaller than 10 km and does not exceed the height of its upper limit $z_C \approx 65$ km, we have

$$\sigma = \frac{\tau_0}{\Delta z} \approx 3 \cdot 10^{-5}\,\text{cm}^{-1} \tag{3.11}$$

accurate to within one half an order of magnitude. 2) From spectral observations of the CO_2 band we obtain in the model with scattering

$$K = \frac{n}{\sigma} = 2 \cdot 10^4\,\text{cm} \cdot \text{atm} \cdot \text{cm}^{-1} \tag{3.12}$$

at the level where $p \approx 0.2$ atm[7]. Assuming $f_{CO_2} = 0.9$, we have

$$\sigma = \frac{0{,}9p}{kTKL} = 2.5 \cdot 10^{-5}\,\text{cm}^{-1}, \tag{3.13}$$

which practically coincides with (3.11). Here $L = 2.65 \times 10^{19}$ cm^{-3} is the Lohschmidt number and k is Boltzmann's constant. In terrestrial clouds, σ is larger by one order of magnitude.

The numerical particle concentration is

$$N \approx \frac{\sigma}{2\pi r^2} \approx 300\,\text{cm}^{-3}, \tag{3.14}$$

and the mass density is

$$\rho = \frac{4}{3}\pi r^3 \rho_0 N \approx 2 \cdot 10^{-9}\,\text{g-cm}^{-3}; \tag{3.15}$$

Here $\rho_0 \simeq 1$ g-cm^{-3} is the density of the material. The total content of the scattering material in a vertical column with 1 cm^2 cross section is

$$B = \rho\Delta z = \frac{2}{3}\tau_0 \rho_0 r \approx 5 \cdot 10^{-3}\,\text{g-cm}^{-2}. \tag{3.16}$$

e) Chemical nature of particles. It is necessary above all to ascertain whether the cloud layer consists of dust or of condensates. There are three objections to dust: 1) the high transparency of the particles in the 0.6–1 μ region is too high for dust; 2) the high altitude of the upper limit of the clouds—the pressure here is smaller by three orders of magnitude than at the surface; 3) the constancy of the infrared brightness temperature (and consequently of the altitude of the upper limit of the clouds if they are made of dust); in the case of condensation clouds, the altitude can vary, and the temperature is maintained automatically constant.

Thus, by inductive reasoning, we conclude that the cloud layer of Venus is made up of condensates, at any rate in the upper part. The assumed low wind velocities (see Chap. 5) suggest that the amount of dust in the atmosphere of Venus is small. What gases can condense in the cloud layer? The temperature of its upper boundary (with correction for the emissivity 0.7–0.9) is 230–250°K. This is too much for the condensation of CO_2.

The next obvious candidate is H_2O. In order for H_2O to condense in a cloud layer at altitudes 60–70 km, it is necessary to have

$$f_{H_2O} > 5 \cdot 10^{-3} \tag{3.17}$$

(see Fig. 14). Thus, if one is to believe the gas analyzers of V4, V5 and V6 ($f_{H_2O} \approx 1 \times 10^{-2}$), then condensation takes place, and if the spectroscopy is to be believed ($f_{H_2O} \approx 1 \times 10^{-4}$), then there is no condensation. An estimate of the critical value of f_{H_2O} depends on the choice of T(z), but it can hardly go below 5×10^{-4} (see[130,131]). In the model of T(z) chosen by us (see Chaps. 6 and 8) at $f_{H_2O} = 10^{-3}$, condensation takes place only in the mesopause. At $f_{H_2O} = 1 \times 10^{-2}$ we have

$$B = 0.6\ \text{g-cm}^2 \tag{3.18}$$

and at $f_{H_2O} = 5 \times 10^{-3}$

$$B = 0.15\ \text{g-cm}^{-2}, \tag{3.19}$$

but the altitude of the cloud layer turns out to be too high. If H_2O does condense, it can be only in the ice phase, and there can be no liquid-drop water in the atmosphere of Venus.

The ice-cloud hypothesis is enticing, but encounters many difficulties: 1) spectroscopy yields too low a value of f_{H_2O} at which there is no saturation and condensation (after all, spectroscopy measures the content of water vapor precisely in the clouds); 2) the absorption coefficients (Table VIII) in the region $\lambda < 0.6$ μ are too large for ice; 3) the absorption bands of ice at 1.5 and 2 μ are absent; 4) the refractive index of ice, m = 1.31, lies outside the limits given by the estimate (3.7). We shall discuss argument (3) in greater detail. Bottema et al.[132,133] obtained the spectrum of Venus from a balloon with extremely low resolution and announced that they had found, as it were, an ice band at

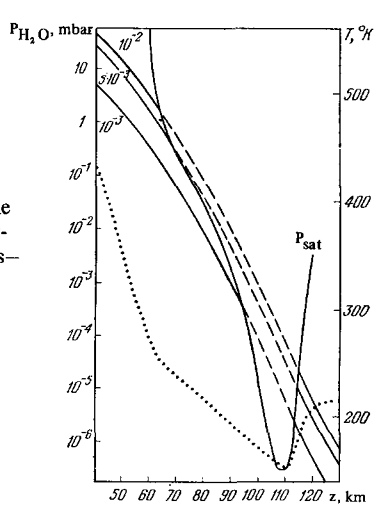

FIG. 14. Conditions for the condensation of H_2O in the atmosphere of Venus. Solid lines—partial pressure H_2O, dashed—temperature.

$2\ \mu$. This was patently in error, since the strongest triad of CO_2 bands is located here (see Fig. 2) and it is impossible to find any other bands here. Pollack and Sagan[134] proved by approximate calculations that the bands at 1.5 and $2\ \mu$ should be weak in the case of particles of micron dimensions, and Plummer attempted to demonstrate this experimentally[135]. Plummer's experiments are not convincing, since, judging from the published curves, they pertain to layers with small albedo and τ_0; as to calculations, there exist exact computer calculations of Hansen and Cheyney[136]. They show that layers of ice particles with micron dimensions, having a large albedo (and large τ_0) give a sufficiently strong depression (of depth up to 20%) at $1.5\ \mu$, which is not observed in the spectrum of Venus.

Many other possibilities were proposed: polymerized C_3O_2[137], hydrocarbons[138], mercury chloride and magnesium chloride[139], and ammonium chloride[140]. None are suitable, since the reflection spectra of these substances are in no way similar to the spectrum of Venus. For example, hydrocarbon clouds should give a strong band at $2.4\ \mu$[141], which is missing from the spectrum of Venus.

The only substance with a suitable spectrum (and geochemically hopeful) was pointed out by Kuiper[44]. This is partially hydrated iron chloride $FeCl_2 \cdot 2H_2O$. The equilibrium concentration of H_2O for this substance is $f_{H_2O} \approx 10^{-6}$, which is a great difficulty.

Even more uncertain is the question of the composition of the ultraviolet clouds. Kuiper proposed NH_4Cl[44]. If it is recognized that they are close in altitude (90–100 km) to the mesopause, then it seems probable that it is precisely they which consist of ice particles, since condensation should occur here even at $f_{H_2O} \approx 10^{-4}$ (see Fig. 14).

4. THERMAL REGIME

As we have seen above (Chap. 2), the surface temperature of Venus lies in the range $650 < T_S < 750°K$. What physical mechanism produces this high temperature? At different times, four hypotheses were advanced:

1) The greenhouse effect (first proposed by Sagan, 1960[142])

2) Fractional heating (the Opik aeolospheric hypothesis, 1961[143]),

3) The Goody and Robinson circulation mechanism (1966[144]),

4) Heating by internal heat (Kuz'min, 1964[145], Hansen and Matsushima, 1967[146]).

Let us consider them in sequence.

a) The greenhouse hypothesis. According to this hypothesis, the high surface temperature is due to the fact that the atmosphere transmits the solar radiation well and the thermal radiation of the planet poorly. In other words, the volume absorption coefficient k_ν is small in the region $\lambda < 2\ \mu$, where solar radiation predominates, and is large in the region $\lambda > 2\ \mu$, where thermal radiation of the surface and of the atmosphere of the planet predominates. The properties of the greenhouse model are best illustrated using the example of "gray two-stream approximation." It presupposes that the absorption coefficients for the solar and planetary radiations k_1 and k_2 are independent of the wavelength, although they are not equal to each other. If $k_1 = 0$, we can regard the surface as a source of heat, emitting a flux

$$F = \frac{1}{4}(1 - A_i)\frac{E_\odot}{r^2}, \tag{4.1}$$

where A_i is the integral albedo, $E_\odot$ is the solar constant, and r is the distance from Venus to the sun. This flux remains unchanged, passing through the atmosphere to the outside, and the temperature under conditions of radiant equilibrium is a function of the optical thickness only:

$$T^4 = \frac{E}{2\sigma}\left(\frac{3}{2}\tau + 1\right). \tag{4.2}$$

Formulas (4.2) is a rigorous solution of the transport equation for a semiinfinite atmosphere in the Eddington approximation[147]. For an optically thick but finite atmosphere, it yields an approximate solution[148]. The optical thickness τ is reckoned downwards. If the total optical thickness for the planetary radiation is $\tau = \tau_S$, then the surface temperature is

$$T_s = \left[\frac{F}{2\sigma}\left(\frac{3}{2}\tau_s + 1\right)\right]^{\frac{1}{4}}. \tag{4.3}$$

The flux F can be expressed in terms of the effective temperature of the planet:

$$\left(\frac{T_s}{T_e}\right)^4 = \frac{1}{2}\left(\frac{3}{2}\tau_s + 1\right). \tag{4.4}$$

Assuming $660 < T_S < 750°$ and $T_e = 228°K$, we have

$$87 < \tau_s < 155. \tag{4.5}$$

In formula (4.3) at $\tau_S \gg 1$, the quantities F and τ_S enter in the form of a product, and if part of the radiation is absorbed in the cloud layer, the flux at the surface is $F_S < F$, and the optical thickness τ_S should be proportionally increased[149].

If

$$n = \frac{k_1}{k_2} < 1, \tag{4.6}$$

then we have for a semiinfinite atmosphere (see[150])

$$\frac{T(\tau, n, \mu)}{T_e} = \left[ng\left(\tau, \frac{n}{\mu}\right)\right]^{\frac{1}{4}}, \tag{4.7}$$

where μ is the cosine of the angle of incidence of the solar rays, $g(\tau, n)$ is the normalized Neumann solution of the inhomogeneous Hopf equation

$$g(\tau, y) = \frac{3}{4}\int_0^\infty f(\tau)e^{-yt}y\,dt\left[f(t) - \int_0^\tau f(t)e^{y(t-\tau)}y\,dt\right], \tag{4.8}$$

and $f(\tau)$ is the normalized solution of the homogeneous Milne equation. Figure 15 shows the solutions of Eq. (4.6 at $\mu = 1$. Just as Eq. (4.2), Eq. (4.6) is of course suitable at large τ_S for the estimate of the surface temperature. From the curves of Fig. 15 it is possible to estimate the values of τ_S and n for which one can obtain an isothermal ($dT/d\tau = 0$) section of the T(z) curve near the surface. At $T_S/T_e \approx 3–3.5$, this will be the interval

$$\begin{gathered} 500 < \tau_s < 1000, \\ \frac{k_1}{k_2} = n \approx 5 \cdot 10^{-3}. \end{gathered} \tag{4.9}$$

The solutions (4.2) and (4.7) were obtained for the

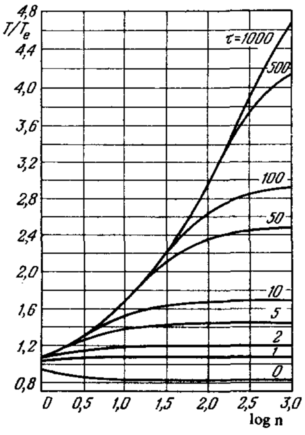

FIG. 15. Greenhouse effect in a gray planetary atmosphere at $k_1 \neq 0$. The temperature as a function of log n at normal incidence of the sun's rays ($n = k_1/k_2$ is the ratio of the absorption coefficients for the solar and planetary radiation).

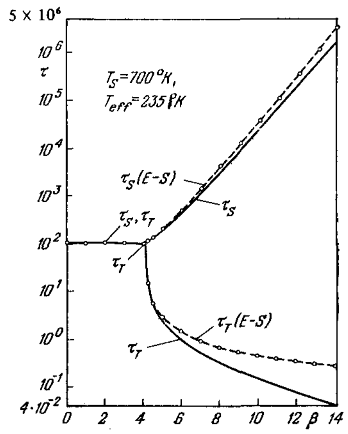

FIG. 16. Total optical thickness τ_S and optical depth of the tropopause τ_T as functions of the stability parameter β. The solid lines were obtained by the method of discrete ordinates, and the dashed lines in the Eddington approximation. For Venus, $\beta = 6$.

state of radiant equilibrium. The atmosphere, however, is unstable in the state of radiant equilibrium (convection arises) if

$$\left(\frac{dT}{dz}\right)_{rad} < \left(\frac{dT}{dz}\right)_{ad} = -\frac{g}{c_p}, \tag{4.10}$$

where $(dT/dz)_{rad}$ is the radiative gradient $(dT/dz)_{ad}$ is the adiabatic gradient, g is the acceleration due to gravity, and c_p is the specific heat at constant pressure. In the Eddington approximation (see[151])

$$\left(\frac{dT}{dz}\right)_{rad} = -\frac{3}{16}\frac{(\alpha+1)g\tau}{R\left(\frac{1}{2}+\frac{3\tau}{4}\right)}, \tag{4.11}$$

where R is the gas constant and α is the exponent in the dependence of the absorption coefficient on the pressure

$$k(p) = k_0\left(\frac{p}{p_0}\right)^{\alpha}. \tag{4.12}$$

The atmosphere is stable against convection if

$$\beta = \frac{c_p}{R}(\alpha+1) < 4 \tag{4.13}$$

in a semiinfinite atmosphere or

$$\beta < 4 + \frac{8}{3\tau_{s,\,rad}} \tag{4.14}$$

in an atmosphere of finite thickness, where $\tau_{s,rad}$ is the optical thickness obtained assuming radiant equilibrium.

If $\beta > 4 + (8/3\tau_{s,rad})$, then convection arises in the region

$$\tau > \tau_T = \frac{8}{3}\frac{1}{\beta-4}; \tag{4.15}$$

here τ_T is the level of the tropopause, above which radiant equilibrium exists and below which there is convective equilibrium. Since $|dT/dz|_{ad} < |dT/dz|_{rad}$, convection leads to the increase of τ_s necessary to ensure the observed ratio T_s/T_e. Watson[151] investigated the influence of convection on τ_s for a gray atmosphere, using the more exact solution of the transport equation (the method of discrete ordinates). The results are presented in Fig. 16.

For Venus, β depends on p and T, and consequently on τ. Assuming a certain mean value $c_p/R = 5.4$ at $\alpha = 0$, we get

$$\beta = 5{,}4, \qquad \tau_s \approx 250; \tag{4.16}$$

At $\alpha = 1$, which is apparently closer to reality, we have

$$\beta = 6.4, \qquad \tau_s \approx 700. \tag{4.17}$$

To determine whether the greenhouse model can explain the high temperature of the surface of Venus, it is necessary to answer two questions: 1) Is the transmission of the atmosphere sufficiently low for planetary radiation? 2) Is it sufficiently high for solar radiation?

1) Long-wave transmission (opacity to planetary radiation). In the atmosphere of Venus there are two types of significant sources of opacity: gas and aerosols. Let us consider gas first. In the atmosphere of Venus, the main gaseous absorbers are CO_2 and H_2O. The absorption is carried out by rotation-vibrational bands (and also by the pure rotational band of H_2O). If k_ν depends on the frequency, then the problem of radiant transport is made much more complicated, especially if k_ν varies rapidly, as is the case in rotational-vibrational bands. In the case of Venus, it is important that k_ν depends on the pressure and on the temperature. The dependence of k_ν on the pressure and on the temperature has been investigated experimentally in insufficiently wide ranges and with extensive averaging of k_ν over the spectrum (several times 10 cm^{-1}). The absorption is described here by formulas of the type

$$\tau = cu^m p^l, \tag{4.18}$$

where u is the amount of absorbing matter, p the pressure, and m and l constants not equal to unity. $m \neq 1$ denotes that the optical thicknesses are not additive, i.e., if there are two adjacent layers u_1 and u_2, then $\tau(u_1 + u_2) \neq \tau(u_1) + \tau(u_2)$. Under such conditions the transport equation is not applicable.

Some authors[152-155], assuming m = 1, considered radiant transport of heat in the atmosphere of Venus with allowance for the frequency dependence of k_ν. However, their results are based on averaging k_ν over large intervals. With such crude formulation of the problem, it suffices to use the gray approximation, calculating the average absorption coefficient, in one manner or another, as was done by Ginzburg and Feĭgel'son[149]. The heat is carried out by the radiation principally in several transparency windows, the positions of which are shown in Fig. 17. The principal role

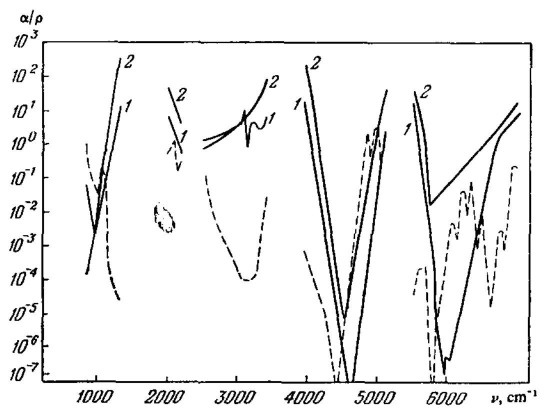

FIG. 17. Mass absorption coefficients (α/ρ) of CO_2 and H_2O in the transparency windows of the atmosphere of Venus (p = 1 atm; averaging over 50 cm^{-1}) [152,153]. 1–α_{H_2O}/ρ at T = 300°K; 2–T = 600°K; dashed lines–CO_2 at 300°K.

is played by three windows, for which the average absorption coefficients are given in Table IX.

The average mass coefficients of absorption ρ_a were calculated by the procedure of[149]. The optical thickness of the atmosphere for planetary radiation is

$$\bar{\tau} = \frac{1}{g}\int_0^{p_s} [k'_{CO_2} f'_{CO_2} \psi_{CO_2}(p) + k'_{H_2O} f'_{H_2O} \psi_{H_2O}(p)]\, dp, \qquad (4.19)$$

Here f′ is the relative content by mass and $\psi(p)$ is the dependence of the absorption coefficient on the pressure. Table X gives the average optical thicknesses calculated under two assumptions with regard to the H_2O content (0 and 0.01) and three assumptions concerning the character of the function $\psi(p)$. It is assumed that p_s = 100 atm.

Table X shows that if the dependence of the absorption coefficient on p is taken into account, then $\tau > 700$ even in the absence of H_2O. Comparing this with the estimate (4.17), we see that the optical thicknesses are sufficient for heating the surface of Venus to a temperature of about 700°K. This result, however, is not very reliable, since it is based on experimental data obtained at CO_2 amounts and pressures smaller by one order of magnitude than in the atmosphere of Venus. Sagan[156] estimates the optical thickness of the atmosphere of Venus for planetary radiation at several hundreds.

We proceed to the other possible cause of infrared opacity, the aerosols. In Chap. 3 we found the total optical thickness of the cloud layer in the region of 1 μ, $\tau_0 \approx 65$, at an average particle radius $\bar{r} \approx 1.25$ μ. For such particles in the region of 10 μ the attenuation cross section is smaller by approximately a factor of 5 than for visible light[157]. If these particles are absorbing (a ≈ 0.2–0.3), then at 10 μ the optical thickness for absorption is

$$\tau_s(10\,\text{atm}) \approx \frac{1}{5}\tau_0(1\ \text{atm}) = 13. \qquad (4.20)$$

Such properties would be possessed, for example, by a cloud layer made up of ice particles. It is clear from the foregoing that the contribution of a cloud layer with similar parameters to the greenhouse effect cannot be appreciable. However, the estimate of τ_0 obtained in Chap. 3 is, strictly speaking, only an upper bound. In the presence of true absorption in the region $\lambda < 1$ μ, the optical depth may turn out to be much larger both here and farther into the infrared region. Samuelson[157,158] considered a semiinfinite atmosphere with optical parameters the same as in the cloud layer, and found that in its deep layers the temperature can reach 500–700°K. The aerosol absorption depends little on the wavelength and is well described by the model of a gray atmosphere in the two-stream approximation. We have already verified above that it is possible to obtain the required values of T_S at $\tau_S \approx 1000$ and $k_1/k_2 \approx 5 \times 10^{-3}$ (the estimate (4.9)), which is very close to Samuelson's results[157]. The aerosol mass in a column of unit cross section is

$$B = \frac{4}{3}\pi r^3 \rho_0 \frac{\tau}{2\pi r^2} \approx \frac{2}{3}\tau r \rho_0 \approx 0{,}2\ \text{g-cm}^{-2}. \qquad (4.21)$$

2) Short-wave transmission (transparency to solar radiation). We have verified above that the first condition necessary for the greenhouse effect (sufficient infrared opacity) can be satisfied. Let us consider a second question: what fraction of the solar radiation absorbed by the planet (1 – A_i = 0.23) can reach the surface? Approximately half the solar radiation lies in the 0.6–1.3 μ band, where the albedo reaches a maximum (see Table VII and Fig. 15). If there is no true absorption in this band, then half of the solar radiation (the unreflected part) passes through the cloud layer. The Rayleigh scattering in the atmosphere below the cloud cannot attenuate it appreciably: at p = 100 atm and near λ 1.1 μ, the Rayleigh optical thickness is

$$\tau \approx 0.5. \qquad (4.22)$$

In general, the presence of a Rayleigh-like atmosphere under the clouds influences only the estimate of the optical thickness of the cloud layer (it decreases it), but not the total transmission of the atmosphere, which is equal to 1 – A_B in the absence of true absorption. A deviation of the true absorption of solar radiation

Table IX. Average mass coefficients of absorption in the transparency windows of the atmosphere of Venus [149]

Interval, μ		2.1–2.4	3.3–4.2	8–10	Over three intervals
$k' = \frac{k}{\rho_a}$, cm^{2}g^{-1} *)	CO_2	$4.7 \cdot 10^{-4}$	$1.9 \cdot 10^{-4}$	$4.5 \cdot 10^{-4}$	$1.27 \cdot 10^{-3}$
	H_2O	0.11	0.15	0.10	0.44

ρ_a–density of absorbing substance.

Table X. Average optical thickness of the atmosphere of Venus for planetary thermal radiation (absorption by CO_2 and H_2O

Composition		τ		
CO_2	H_2O	$\psi(p) = 1$	$\psi(p) = \sqrt{\frac{p}{1\ \text{atm}}}$	$\psi(p) = \frac{p}{1\ \text{atm}}$
0.9	None	$1.2 \cdot 10^2$	$7.2 \cdot 10^2$	$7.8 \cdot 10^3$
0.9	0.01	$5.6 \cdot 10^2$	$3.4 \cdot 10^3$	$3.6 \cdot 10^4$

from zero in some section of the spectrum is in itself of no great importance for the greenhouse effect. As we have seen above (Fig. 15), it suffices to maintain a definite small ratio k_1/k_2.

However, if dust raised from the surface is present in such an atmosphere, then it is very doubtful that it consists of transparent particles. It is precisely the possible presence of dust in the lower dense layers of the atmosphere which is the most frequent argument raised against the greenhouse model.

b) Aeolospheric model. This present argument was discussed in very great detail by Opik[143], and caused him to advance, in opposition to the greenhouse model, the so-called aeolospheric model. Opik proposed that the solar radiation is absorbed in the region of the upper boundary of the cloud layer; the circulatory motions that result from this absorption encompass the entire atmosphere, and part of the absorbed energy is transferred to the surface due to viscosity. The atmosphere is opaque to both the solar and planetary radiation. It is clear, however, that there is no need here for viscosity at all. If the circulation mixes the atmosphere, then the vertical motions should be adiabatic, since the energy does not go outside, owing to the infrared opacity. If the mixing extends to the surface, then the adiabatic gradient is established in the entire thickness of the atmosphere, starting with the cloud layer.

c) Circulation hypothesis. Developing further similar considerations with the aid of the mathematical formalism of dynamic meteorology, Goody and Robinson have shown[144,159] that such a model is perfectly viable. The influx of solar heat is sufficiently large, and the circulation in the relatively high layers of the atmosphere is accompanied by a slow ascending flow (in the equatorial zone), which encompasses the entire atmosphere. Its velocity is large enough to enable it to prevail over the diffusion of heat and momentum, ensuring the adibaticity of the motion, and as a result a high temperature can be produced at large depths by reversible processes without the penetration of solar energy. In the first variant of their work, they considered the circulation between the subsolar and antisolar points[144], but they later went over to the equator—pole circulation, since the thermal inertia of the atmosphere is too large and there are practically no diurnal effects (see Chap. 5).

d) Hypothesis of heating by internal heat. Kuz'min[145] and subsequently in general form Fesenkov[160] advanced the hypothesis that the surface of Venus can be heated by an internal-heat flux much larger than that of the earth. The flux proposed by Kuz'min (10^4 erg-cm^{-2} sec^{-1}) would produce a temperature of 6000°K at a depth of 2 km below the surface ([4], p. 252). Obviously, such fluxes need not be considered.

The heat flux on the earth's surface is 60 erg-cm^{-2} sec^{-1}. Let us assume that a like flux obtains on Venus. This is smaller by a factor 10^3 than the flux obtained by the planet from the sun. We have seen above that in the greenhouse model it is necessary to have $\tau_s \approx 200-1000$ in order to ensure the observed surface temperatures. If the heat flux on the surface is smaller by 10^3, this means that to produce the observed temperature $T_s \approx 700-750°K$ we need

$$\tau_s \approx 2 \cdot 10^5 - 1 \cdot 10^6. \tag{4.23}$$

Such an optical thickness can be produced by gas absorption only with dust. Its content should be

$$B = 40 - 200\,\text{g-cm}^{-2}. \tag{4.24}$$

This is a very large amount of dust, but nevertheless this possibility has been discussed[146].

The choice between the considered hypotheses is a matter for the future, but the second and fourth can be discarded with a tremendous probability. Only the greenhouse and circulation hypotheses can be regarded more or less on a par.

5. GENERAL CIRCULATION

The equatorial regions of a rotating planet obtain, averaged over the year, more heat than the polar regions. A temperature difference ΔT_φ is produced between the equator and the pole, with ascending flow at the equator and descending at the pole. Both flows form a closed system and transfer energy from the equator to the pole, thereby smoothing out the difference.

Ten years ago it was assumed that Venus has synchronous rotation, i.e., the same side of the planet faces the sun constantly. In this case the circulation should transfer heat from the subsolar point to the antisolar point. It was established by radar methods that Venus rotates slowly, to be sure, but not synchronously. The sidereal period of the revolution is[161]

$$P = 243^d.1 \pm 0.2; \tag{5.1}$$

the direction of rotation is retrograde, and since the period of revolution is 224.7^d, the duration of the solar day $P_\odot$ is determined from the equation

$$\frac{1}{P_\odot} = \frac{1}{243.1} + \frac{1}{224.7}, \tag{5.2}$$

whence

$$P_\odot \approx 117^d. \tag{5.3}$$

During the Venusian year, the observer sees two sunrises and two sunsets. At first glance it appears probable that the night side of the planet should be much colder than the day side. The thermal inertia of the atmosphere, however, is so large that it does not have time to cool appreciably during the night. A column of unit cross sections has a heat content

$$Q = mc_p = \frac{pc_p}{g} \approx 5 \cdot 10^{11}\,\text{erg-cm}^{-2}; \tag{5.4}$$

where m is the mass of the column. The flux lost per unit time is

$$F=\frac{1}{4}(1-A_i)\frac{E_\odot}{r^2}=1.5\cdot10^5\ \text{erg-cm}^{-2}\text{sec}^{-1}.$$

The amplitude of the diurnal temperature oscillations is

$$\frac{\delta T}{T}=\frac{FP_\odot}{2Q}\approx1.5\cdot10^{-3}. \qquad (5.5)$$

Consequently, the diurnal temperature difference is approximately 1°.

a) Temperature difference between the equator and the pole. The difference ΔT_φ depends on the extent to which heat is transferred effectively from the equator to the pole. Golitsyn[162], using similarity and dimensionality methods, has shown that

$$\Delta T_\varphi\approx\left(\frac{\pi}{2}\right)^{\frac{1}{2}}\left(\frac{\lambda_1}{k\alpha}\right)^{\frac{1}{4}}\frac{F^{\frac{9}{16}}}{\sigma^{\frac{1}{16}}c_p^{\frac{3}{4}}}\left(\frac{R}{m}\right)^{\frac{1}{2}}, \qquad (5.6)$$

where $\lambda_1 = 1$ for slowly rotating planets, $k \approx 0.1$ is the coefficient of utilization of the atmospheric "heat engine," $\alpha = T_e/T_s$, σ is the constant of the Stefan-Boltzmann law, and R is the radius of the planet.

Substituting in (5.6) $F = 1.5 \times 10^5$ erg-cm^2 sec^{-1}, $c_p = 10^7$ erg-g^{-1} deg^{-1} and $R/m = 5.3 \times 10^3$, we obtain

$$\Delta T_\varphi\approx2^\circ\text{K}, \qquad (5.7)$$

i.e., the temperature difference ΔT_φ is small.

Apparently, the horizontal temperature gradients in the latitudinal and longitudinal directions on Venus are comparable, and the circulation motions have a complicated character. Besides the heat transfer from the equator to the pole, a noticeable role may be played also by transfer from the subsolar to the antisolar point.

b) Average velocity of atmospheric motions. Following Golitsyn[162], we calculate the average velocity of the atmospheric motions

$$v=\left(\frac{\pi}{2}\right)^{\frac{1}{2}}\left(\frac{k}{\lambda_1}\right)^{\frac{1}{4}}\frac{\sigma^{\frac{1}{16}}F^{\frac{7}{16}}}{c_p^{\frac{1}{4}}}\left(\frac{R}{m}\right)^{\frac{1}{2}}\approx10^2\,\text{cm-sec}^{-1}, \qquad (5.8)$$

i.e., the wind velocities on Venus should be very low.

c) Time scale of atmospheric motions. This quantity indicates, roughly speaking, how rapidly the weather changes. It is equal to

$$\tau=\frac{R}{v}\approx6\cdot10^6\ \text{sec} \qquad (5.9)$$

i.e., it is close to the duration of the solar day on Venus. Formulas (5.7), (5.8), and (5.9) have been derived under the assumption that the solar radiation is absorbed by the surface. If the solar radiation does not reach the surface, then k decreases by two or three orders of magnitude. The velocities are then decreased by a factor 3–5, and ΔT_φ and τ are accordingly increased.

Goody and Robinson[144] and Stone[163] considered detailed models of the meridional circulation. According to their estimates, $\Delta T_\varphi \approx 10^\circ$K, which is close to (5.8); the characteristic motion velocities are much higher, $v \approx 10$ m-sec^{-1}, but they pertain to sufficiently high layers of the atmosphere.

Data on the winds in the lower atmosphere of Venus are practically nonexistent. There is, however, curious information concerning the motion of ultraviolet clouds.

d) Motions of ultraviolet clouds. Ultraviolet clouds (see Chap. 3) are quite stable, and the period of rotation can be determined from their displacement from day to day. It turns out to be[164,165]

$$P(\text{UV})\approx4^d; \qquad (5.10)$$

the direction of revolution is retrograde. This means that at the altitude at which the ultraviolet clouds are localized (≈90 km) there occur horizontal motions that are approximately constant in direction and in magnitude, with a velocity

$$v=\frac{2\pi R}{P(\text{УФ})}\approx100\,\text{m-sec}^{-1}. \qquad (5.11)$$

The nature of this wind is not clear. A qualitative explanation was offered by Schubert and Whitehead[166], who performed the following model experiment. They moved a Bunsen burner on a circular path under a cylindrical reservoir filled with mercury. The cylinder started to rotate in the opposite direction with a velocity higher than that of the burner. The reason, according to[166], is that the thermal diffusion prevails over viscous diffusion. A similar situation can take place also in the atmosphere of Venus at the altitudes of the ultraviolet clouds.

6. UPPER ATMOSPHERE

Experimental data on the upper atmosphere of Venus are limited. They can be divided into two groups:

1) Measurements of the electron density on M5[167].

2) Measurements of the intensity of the resonant emission of the atmosphere in the lines L_α λ 1215 Å and OI λ 1304 Å [168-170].

a) Ionosphere. Figure 18 shows the vertical distribution of the electron density in the atmosphere of Venus, obtained as a result of measurement of refraction with the aid of the spaceprobe M5. The measurements were performed by a two-frequency method[171] at 49.8 and 423.3 MHz. The plasma refractive index

$$m=\sqrt{1-\frac{e^2n_e}{2\pi m_ef^2}} \qquad (6.1)$$

depends strongly on the frequency f. Beats between the lower frequency and the 2/17 subharmonic of the higher

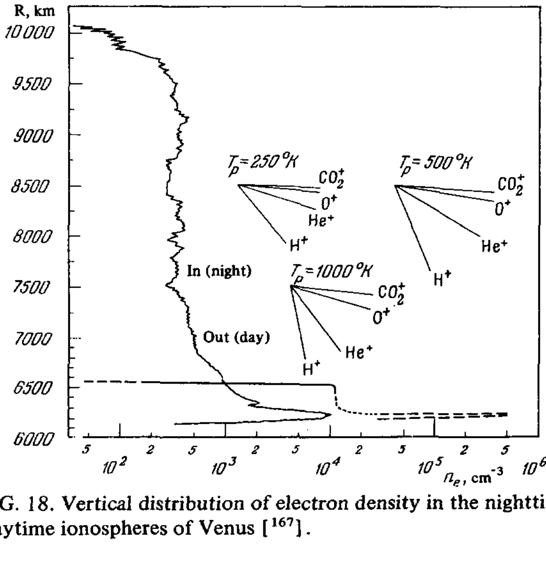

FIG. 18. Vertical distribution of electron density in the nighttime and daytime ionospheres of Venus [167].

frequency were registered on the spaceprobe. The beats gave the phase difference introduced by the ionosphere on passage of the lower frequency.

Figure 18 demonstrates the large difference between the structures of the daytime and nighttime ionospheres. The daytime ionosphere at 150 km altitude forms a narrow layer with a maximum $n_e \approx 5 \times 10^5$ cm^{-3}. At 550 km, n_e decreases rapidly to zero. The maximum of the nighttime ionosphere is located at the same altitude, ~150 km, but here $n_e \approx 10^4$ cm^{-3}. This is followed by a smooth decrease of n_e to the level $n_e \approx$ cm ≈ 500 cm^{-3}, and above that n_e remains practically constant up to an altitude of approximately 3500 km. The difference is due to two factors: 1) the absence, during the night, of the ionizing solar radiation (as a result of which n_e is lower at the nighttime maximum), 2) the absence of planetary magnetic field. According to measurements on V4 and M5[172-174], the dipole field of Venus itself is not more than 10^{-4} of the terrestrial field. The solar wind presses the ionosphere, which is not protected by a magnetic field, to the surface of the planet, thus explaining the sharp boundary of the daytime ionosphere at 550 km. This is the level of the plasmopause—the boundary between the solar plasma and the ionosphere.

More exact profiles in the regions of the day time and night time maxima can be found in[175-177].

The straight-line bunches of Fig. 18 show the altitude distributions of the ion concentration, calculated for different temperatures. In the nighttime ionosphere at altitudes 150—250 km, the main ion is CO_2^+ at a temperature

$$250 < T_i < 500° \mathrm{K}. \tag{6.2}$$

In the 350—2000 km region, the main ion is He^+ and

$$620 < T_i < 970° \mathrm{K}, \tag{6.3}$$

and at higher altitudes the main ion is H^+ (and probably D^+, see below) and

$$625 < T_i < 1100° \mathrm{K}. \tag{6.4}$$

The absence of the O^+ ion at altitudes below 250 km means that the CO_2 does not dissociate in this region. Detailed calculations of McElroy[175] and Steward[177] have shown that even a 10% dissociation of CO_2 cannot be reconciled with the observed n_e profile. Addition of 10% N_2 likewise already causes noticeable deviations from the observed profile, and this gives for the N_2 content an upper limit independent of the direct measurements.

b) <u>The problem of CO_2 dissociation.</u> The CO_2 dissociation is also small in the atmosphere of Mars[178]. It was generally assumed even recently[4,179,180] that the photochemical equilibrium of CO_2 is determined by the reactions

$$CO_2 + h\nu \rightarrow CO + O(^1D), \quad \lambda < 1700\,\text{Å}, \tag{6.5}$$

$$O(^1D) \rightarrow O(^3P) + h\nu, \quad \lambda = 6300\,\text{Å}, \tag{6.6}$$

$$CO + O + M \rightarrow CO_2 + M, \tag{6.7}$$

$$O + O + M \rightarrow O_2 + M, \tag{6.8}$$

$$O_2 + h\nu \rightarrow O + O, \quad \lambda < 1800\,\text{Å}. \tag{6.9}$$

The rate of reaction (6.7) is quite low, and CO_2 should dissociate above the level $n \approx 10^{14}$ cm^{-3} ($z \approx 120$ km) if there are no faster reverse reactions.

Warneek[181] and also Young and Ung[182] have established in laboratory experiments that the rate of CO_2 dissociation by ultraviolet radiation is much lower than it should be in the reactions (6.5)—(6.8). They advanced the hypothesis that it is necessary to add to them the reactions

$$CO_2 + O(^1D) \rightarrow CO_3^*, \tag{6.10}$$

$$CO_3^* + CO \rightarrow 2CO_2. \tag{6.11}$$

Donahue[183] noted, however, that the rate coefficient of the reaction (6.10) and the lifetime of the complex CO_3^* that are necessary for such an interpretation turn out to be improbably large. Another possible reverse reaction is

$$CO + OH \rightarrow CO_2 + H. \tag{6.12}$$

The CO_2 dissociation rate is approximately 2×10^{11} cm^{-2} sec^{-1}. This means that the rate of formation of H should be the same. Yet the concentrations of H measured in the atmosphere of Venus (from the emission in Lα) lead to formation rates 2×10^6 cm^{-2} sec^{-1}. Donahue[183] proposed a third possibility, the reaction

$$2CO + O_2 \rightarrow 2CO_2, \tag{6.13}$$

which is very slow in dry gases, but is catalyzed by very small H_2 impurities ($\approx 10^{-6}$).

The low content of atomic oxygen in the upper atmosphere of Venus is evidenced by measurements of the intensity of the resonant emission of the triplet OI 1304 Å on V4[169]. They have shown that an altitude of 300 km (at which the spaceprobe entered the shadow) the concentration $n_0 < 2 \times 10^3$ cm^{-3}. This is smaller by almost six orders of magnitude than in the case of complete dissociation of the CO_2.

On the other hand, a rocket experiment[170] yielded the spectrum of Venus with low resolution, in which a bright detail (≈ 5 kilorayleigh) is observed in the region of 1300 Å. If it belongs to OI, then partial dissociation of CO_2 takes place all the same. The region of the OI emission apparently is located below 300 km.

c) Temperature of the upper atmosphere. The calculated profile of the electron concentration is very sensitive to the quantity ϵ, the coefficient of conversion of the ionizing solar radiation into heat. Hogan and Steward[184,185] found for Venus and Mars

$$\varepsilon \approx 0.3, \tag{6.14}$$

from which they obtained the temperature in the exosphere

$$T_{ex} = 650 \pm 70° \mathrm{K} \tag{6.15}$$

for the solar activity level corresponding to the time of flight of M5. This temperature is established in the region above 200 km. Precisely the same temperature is obtained from the profile of the neutral-hydrogen concentration[186].

McElroy[175,176], and also Stewart[177] calculated the total profile of the temperature in the atmosphere of Venus under different assumptions concerning the composition. A check on these calculations is the altitude of the ionospheric maximum. Good agreement with the data of Mariner-5 is obtained only for pure CO_2 or for CO_2 with a small admixture of N_2 (on the order of 10%).

The critical level ($n \approx 10^8$ cm^{-3}) is reached at the altitude

$$z_c = 230\ \text{km} \tag{6.16}$$

The temperature of the mesopause (z_m = 110 km) is determined from the condition that the entire energy coming from the higher layers is radiated here in the λ 15 μ band of CO_2. It is equal to

$$T_m = 160^\circ\,\text{K}. \tag{6.17}$$

Between the tropopause and the mesopause (in the stratosphere) there is an approximately constant temperature gradient

$$\frac{dT}{dz} = 2^\circ\ \text{K/km} \tag{6.18}$$

i.e., unlike the gray approximation (see Chaps. 3 and 4), the stratosphere is not isothermal. The thermal regime in the stratosphere of Venus was considered also by Bartko and Hanel[187]. They assume that appreciable diurnal variations of the temperature are possible in the stratosphere.

One more important level in the atmosphere is the turbopause, which separates the regions of total mixing (homosphere) and diffusion separation (heterosphere). In the homosphere, the altitude scale for all gases (with the exception of the rapidly dissipating hydrogen) is the same and corresponds to an average molecular weight $\bar{\mu}$, and in the heterosphere the altitude scale for each gas is determined by its own molecular weight. If the position of the turbopause corresponds on Venus to the same total numerical concentration as on Earth, then its altitude is

$$z_{\text{turb}} = 140\ \text{km} \tag{6.19}$$

d) Hydrogen corona. Deuterium. Figure 19 shows the intensities of the emission of the upper atmosphere of Venus, measured with an ultraviolet photometer mounted on board M5[186]. Similar results were obtained with an analogous instrument on V4[189]. Measurements with a CaF_2 filter cover the region of the L_α line. The intensity with this filter is much higher than with the two others (which do not encompass L_α), and what was apparently observed here was the emission of the solar L_α line, scattered by hydrogen in the outer layers of the atmosphere of Venus. This very same outer part of the atmosphere can be called the hydrogen corona of Venus. An analogous hydrogen corona ("geocorona") is possessed also by the earth. Its presence is explained by the fact that the concentration of H in the heterosphere, owing to the small atomic weight, varies with altitude much more slowly than those of O, N_2, and all the more CO_2.

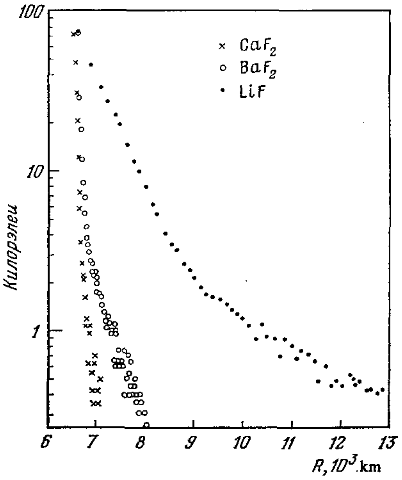

FIG. 19. Intensity of ultraviolet emission of the upper atmosphere of Venus as measured on M5 [186]. The short-wave limits of the filters are: CaF_2–1250Å, LiF–1050Å, BaF_2–1350Å.

From the curve for CaF_2 of Fig. 19 it is clearly seen that above 3000 km (R > 9000 km) the decrease of the intensity with increasing altitude proceeds at a slower rate than in the lower layers. The altitude scale H = kT/mg doubles for hydrogen at this altitude.

Such a change can be due to two causes: either a change of T by a factor of two, or a change of m. The change of m means that below 3000 km the emission comes either from deuterium or from H_2 molecules:

$$\begin{gathered} H_2 + h\nu \rightarrow H(2p),\ \lambda < 850\ \text{Å}, \\ H(2p) \rightarrow H(1s) + h\nu(L_\alpha). \end{gathered} \tag{6.20}$$

The hypothesis that T changes encounters great difficulties. It means that below 3000 km we have $T \approx 325^\circ$K, which does not agree with the ionospheric data (see the estimate (6.3)). In addition, at this altitude the atmosphere has a very small optical thickness for ultraviolet radiation of the sun, and cannot be heated by the latter. Stewart[177] suggested that the source of the hot atoms of the hydrogen is the solar wind, but it was shown that this source is insufficient[188].

The molecular-hydrogen hypothesis requires very large H_2 concentrations ($n \approx 10^9$ at 1000 km altitude), owing to the smallness of the flux of solar radiation in the region $\lambda < 850$ Å. Such an extended atmosphere of molecular hydrogen should be ionized, and its temperature would be not less than 2000°K[189], so that this hypothesis likewise does not hold.

The only reasonable possibility is a ratio $n_D/d_H > 1$ at altitudes $z < 3000$ km. A large n_D/n_H ratio can be obtained as a result of a difference in dissipation rates. Donahue[183,189] proposed

$$\left(\frac{n_D}{n_H}\right)_0 = \left(\frac{n_D}{n_H}\right)_c \frac{F_D}{F_c}, \tag{6.21}$$

where $(n_D/n_H)_0$ is the n_D/n_H ratio below the turbopause, $(n_D/n_H)_c$ is the same in the exosphere, and $n_{Hc}F_H$ and $n_{Dc}F_D$ are the fluxes of the dissipating H and D atoms at the critical level. T = 700°K, he obtained $(n_D/n_H)_0 \approx 5 \times 10^{-4}$. On earth, this ratio is equal to 1.4×10^{-4}. Donahue did not take into account, however, the fact that the fluxes $n_{Hc}F_H$ and $n_{Dc}F_D$ are limited by diffusion. Allowance for diffusion (McElroy and Hunten[188]) yields

$$\frac{n_D}{n_H} \approx 0.1 \tag{6.22}$$

in the lower atmosphere of Venus. Thus, the relative content of deuterium on Venus should be much higher than on earth. From earth-based spectroscopic observations (using the HCl lines), the upper limit is also found to be about 0.1.

The most detailed analysis of the L_α observations on M5 is contained in the paper of Wallace[190]. He has shown that the exosphere of Venus has a noticeable asymmetry—on the nighttime side the deuterium concentration is smaller by one order of magnitude than on the daytime side, and that of hydrogen is smaller by a factor of 2.

The rate of dissipation of hydrogen is limited in practice only by the rate of dissociation of the H_2O, which occurs near the mesopause. In spite of the fact that the exosphere temperature is half as high, the conditions for dissipation are approximately the same as on earth, because the ionized hydrogen atoms escape freely from the atmosphere of Venus. Their effective atomic weight (together with the electrons) is 0.5, and Venus does not have the magnetic field that retains the protons in the earth's atmosphere.

e) Helium. Judging from the altitude scale of the nighttime ionosphere, helium should be the main component at altitudes 300—2000 km[176]: This leads to a relative concentration $f_{He} \approx 2 \times 10^{-4}$ in the homosphere. The absolute content of He in the atmosphere of Venus is larger by four orders of magnitude than on earth. This is explained by the fact that the helium does not dissipate from the atmosphere of Venus, owing to the low temperature at the critical level and to the high ionization potential[191].

f) Emission of the upper atmosphere. From visual observations[192] it is known that the dark side of Venus sometimes emits a reddish-brown ashen light. This is most probably an aurora or skyglow. Kozyrev[40] and Newkirk[193] obtained spectrograms of the ashen light with emission bands, but these observations are not reproducible, and apparently the spectrum of the ashen light varies in time no less strongly than its brightness. Kozyrev found a large number of emissions with intensity of approximately 20 rayleigh which are well identified with CO_2 bands excited by proton beams[41], and there are no grounds for doubting their reality, but a spectrogram of this type was obtained only once. Newkirk[193], using a coronograph for the observations, found traces of emission near 4415 Å, likewise with approximate density 20 rayleigh; during the next inferior conjunction, this emission was not found[194]. Goody and McCord[195] attempted to measure the brightness of the glow of the night side of Venus with the aid of a photoelectric photometer with a set of interference filters, but obtained only an upper limit (approximately 100 rayleigh in the 200 Å band).

Observations of the skyglow, twilight, and auroras are among the most powerful tools for study of the upper atmosphere of the earth. A task for the nearest future is to apply this technique to investigations of other planets. McElroy and Strobel[176] gave a list of expected skyglow emissions of Venus. Among them are the well known forbidden lines OI λ5577 and 6300 Å. The intensities of the emissions expected in the skyglow spectrum are low, within 100 rayleigh, but one can expect much more, up to 100 kilorayleigh, in twilight and in auroras. The most promising here are observations from orbiting spacecraft, but even earth-based techniques are not hopeless. An extraeclipsing coronagraph with narrow filters would make it possible to observe emissions with intensity 50—100 rayleigh.

7. ORIGIN AND EVOLUTION

Venus and the earth are close in mass and radius, and their distances to the sun differ by only 30%, but their atmospheres have turned out to be quite dissimilar. The reasons should be sought in the conditions for the formation and evolution of the atmospheres.

According to the general conviction (see, for example[196-199]), the planets were the result of accretion of solid particles in a cold protoplanetary cloud surrounding the sun during the epoch of its formation. The earth retained the primordial atmosphere, similar in composition to the protoplanetary cloud (hydrogen, helium, inert gases, especially neon) only for a short time. The primordial atmosphere was almost completely lost (with the possible exception of the heavy inert gases) and replaced by a secondary atmosphere consisting of products of volcanic activity.

Table XI lists the comparative characteristics of the atmospheres of Venus and of the earth. For the earth, besides the composition of the present-day atmosphere, there is given the summary content of volatile compounds in the atmosphere, hydrosphere, and sedimentary rocks after Rubey[200], i.e., the total amount of the volatile compounds released as a result of volcanic activities during the entire geological history of the earth.

It is clear from Table XI that the difference between the total amounts of CO_2 and probably also N_2 on Venus and on the earth is small. The difference in the amounts of O_2 is explained by the fact that this gas is continuously supplied on earth by the biosphere. The biosphere has also transformed a greater part of the CO_2 on earth into sedimentary rocks. There remains only one question, but a very difficult one—why is there such a large difference in the amounts of H_2O, what did Venus do with its oceans? After all, H_2O is the main component of volcanic gases. Cameron[201] and Fesenkov[160] advanced the idea that since there is little H_2O in the atmosphere of Venus, it means that this atmosphere is not of volcanic origin, but is a remnant of the protoplanetary nebula, i.e., the astmosphere of Venus is not secondary, as on earth, but primordial. This is quite doubtful. The primordial atmospheres were retained by the giant planets, but they have an entirely different composition, and their principal components are hydrogen and helium. Were the atmosphere of Venus a remnant of a primordial nebula, it would contain much neon.

Holland[202] advanced the hypothesis that during the

Table XI. Compositions of the atmospheres of the earth and Venus

Gas	Earth		Venus, atm
	Contemporary atmosphere, atm	Atmosphere plus hydrosphere plus sedimentary rocks, atm	
H_2O	0.001	300	0.01—1
CO_2	0.0003	20	100
N_2	0.8	1	< 10
O_2	0.2	0.2	< 0.00005

stage of accretion of Venus, there were no ice particles (owing to the higher temperature), with the result that the solid matter of the planet contains much less H_2O than on earth. Gold stated in refutation that an appreciable H_2O fraction should be contained in hydrated silicates[203].

Rasool[125] believes that the lithosphere of Venus released as much H_2O as that of earth, but that almost the entire mass of H_2O was dissociated, the hydrogen was thrown into interplanetary space, and the free oxygen reacted with CH_4 and CO, transforming them into CO_2. In order for this mechanism to operate, it is necessary to have a much higher temperature in the mesopause than indicated in Chap. 7 (the McElroy model, 160°K). It is not excluded that the mesopause of Venus is warmer than in the McElroy model, or else was warmer earlier. The latter means that during later epochs, when the mesopause had "cooled down," the release of H_2O was greatly slowed down (the terrestrial rate of H_2O production is about 100 atm in 10^9 years). If the large value of the ratio n_D/n_H in the atmosphere of Venus is confirmed, this will be a decisive argument favoring the dissipative mechanism of removal of H_2O, since the dissipation of D proceeds much more slowly.

It is not excluded, finally, that Venus went through a stage of at least partial surface melting and that its atmosphere was formed in equilibrium with the molten material of the upper mantle. Then the bulk of the H_2O should have remained dissolved in the magma[59], while CO_2 and N_2, the solubility of which is much lower, should have gone into the atmosphere. It must be assumed here that soon after the solidification of the crust, the rate of release of volcanic gases was slower by several orders of magnitude than on earth.

8. PROSPECTS FOR FURTHER RESEARCH

Investigations of Venus with the aid of interplanetary automatic stations, carried out in 1967–1969, gave most valuable physical information on the atmosphere of this planet. We know now, in general outline, the composition of the atmosphere of Venus, We must learn much more, however, in order to understand how the atmosphere has acquired its features. What are the next questions on the agenda?

1) What is the mechanism of heating of the atmosphere—the greenhouse effect alone or "greenhouse" plus mechanical mixing? What is the role played in the heat balance by the aerosol component?

2) What is the chemical composition of the cloud-layer particles, and what are the vertical and horizontal structures of this layer?

3) How did the atmosphere evolve, why is there so much CO_2 and so little H_2O?

To answer these questions, we need new and more detailed investigations with the aid of soft-landed, flyby and orbiting spaceprobes, and also terrestrial—astronomy methods. No orbiting spaceprobes have as yet been sent to Venus (or to any other planet), but there is hope that this will occur in the not too remote future. NASA is preparing two such probes to be sent to Mars in 1971. Orbiting ships would also be quite effective, in the author's opinion, for the investigation of Venus. They do not make it possible to measure directly the parameters of the atmosphere by direct methods (for this purpose, of course, soft-landed vehicles are irreplacable), but on the other hand they can investigate by indirect means (optical and radiophysical) practically the entire planet with a spatial resolution far exceeding the capabilities of earth-based astronomy.

Using soft-landed vehicles, it is urgently necessary to continue measurements of the main parameters of atmosphere down to the surface. From a determination of the content of the main components of the atmosphere it would be interesting to go on to the minor components, although technical difficulties are unavoidable. It is necessary to develop accessible methods for a direct analysis of the chemical composition of the particles in the cloud layer. In principle, relatively simple optical pickups can be used to attempt to measure the distribution of the intensities of the direct solar radiation and the scattered and reradiated planetary radiation, as functions of the altitude. This information is important for the investigation of the heat balance and the stratification of the clouds (these questions are most likely related).

Using orbiting stations (and also flyby probes), but at a much shorter available observation time), it is possible to investigate the atmosphere below the clouds with the aid of radio telescopes. This method can yield, in principle, the distribution of the temperature over the surface of the planet (the latitudinal and diurnal effects), and also at different levels of the atmosphere. Photometry and spectrscopy during a flyby would give information on the macrostructure and microstructure of the cloud layer in its upper part, and possibly also its chemical composition. To investigate the minor components of the atmosphere and processes in its upper layers it would be extremely interesting to obtain the emission spectra of the night sky in the visible, near-infrared, and ultraviolet regions. It is necessary to verify by ultraviolet spectroscopy whether the excess of deuterium in the upper atmosphere, indicated by the results of M5, is real.

The foregoing list of possible experiments does not represent any official program and is not exhaustive. We have simply listed certain possibilities that are more or less realizable in the foreseeable future.

Many problems can be solved by methods of earth-based investigations. High-resolution spectroscopy of Venus as a means of investigating the chemical composition has almost exhausted its capabilities. The most informative region of the spectrum, $\lambda 1-2.5\ \mu$, has already been obtained with a resolution close to 10^5. The possibility of new identifications is limited here not by the shortage of spectral information on the planet, but by the absence of laboratory spectra of many molecules with the same resolving power.

At the present day, the list of problems faced by earth based observations is approximately as follows:

1) The widths of the CO_2 lines, to determine more accurately the pressure in the cloud layer (a resolution of approximately 10^6 is needed), 2) a more accurate dependence of the albedo on the wavelength in the region $\lambda > 1\ \mu$ (to identify the material of the

cloud layer), 3) the variation of the CO_2 bands over the disk and the dependence of their intensity on the phase, 4) the spectrum of the ashen light (at a good instrumental level—with an extraeclipsing coronograph, or a differential photometer), 5) the spectrum of the radio emission in the millimeter ($\lambda < 8$ mm) and submillimeter bands (this yields the structure of the stratosphere).

There is no doubt that spaceprobes are the most effective means of investigating planets, but they call for very large expenditures. Apparently, earth-based and subcosmic methods (i.e., balloons and airplanes) will remain during the next decade, as before, an important source of information on Venus and Mars, let alone other planets. In the study of Venus and Mars we have passed, in the main, the stage of the "first onslaught." We are faced with a period of detailed investigations, which call for long efforts on a very wide front.

SUPPLEMENT

THE RESULTS FROM VENERA-7

The main task of the automatic station "Venera-7" (see "Pravda" of 27 January 1971) was to land on the surface of the planet and to measure directly the pressure and temperature down to the surface. An analysis of the telemetry information and data on the descent velocity (obtained from the Doppler shift of the radiosignal frequency) has shown that the station reached the surface of Venus, and that the transmitter continued to operate 23 minutes after landing. The entire measurement program could not be performed, owing to breakdown of the on-board commutator, but the most valuable part of the information, the change of the temperature during the descent of the station to the very surface, was obtained. In conjunction with the data on the rate of descent as a function of the time, this has made it possible to obtain the altitude dependence of the temperature. The temperature of the atmosphere at the surface is $T = 748 \pm 20^\circ K$,

The temperature gradient remains, within the limits of measurement errors, adiabatic down to the surface. Thus, the correctness of the extrapolation of the measurements of V4, V5, and V6 downward along the adiabat, assumed in constructing (Chap. 2) the working model of the atmosphere of Venus, has been confirmed. We recall that this extrapolation yields $T = 750^\circ$ and a surface pressure $p = 92$ atm. The hypothesis that the temperature gradient possibly decreases in the layers next to the surface, mentioned in Chaps. 2 and 4, is refuted.

"Venera-7" was the first space vehicle to transmit scientific information from the surface of another planet. Its flight initiated direct experiments on the surface of Venus.

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