MODELS OF THE INTERNAL STRUCTURE OF VENUS

V. N. ZHARKOV

Institute of the Physics of the Earth, USSR Academy of Sciences, Moscow, U.S.S.R.

(Received 18 July, 1983)

Abstract. A survey is made of the physics of the interiors of Venus. The introduction explains the main concepts used in the construction of models of Venus and the history of the question; observational data are gathered and analyzed. The method of constructing the models of the planet is explained and earth-like models of Venus and parametrically simple PVM models are discussed. Within the compass of a physical model of Venus, the thermodynamics of the mantle and core is constructed and questions are discussed concerning the heat conduction, temperature distribution in the lithosphere and the thermal flux from the interior of Venus, the electrical conduction and mechanical quality, and large-scale steady stresses in the mantle of Venus. A rheological model of the crust and mantle is constructed. In conclusion, the question as to the distribution of radioactivity and convection in the interior of the planet is discussed.

1. Introduction. Observational Data

Venus belongs to the planets of the Earth group, which also includes Mercury, Earth, Mars, and the Moon. All the planets of the earth-like group, including Venus and Earth, are relatively small. As a result, in the process of their formation they were unable to retain the hydrogen–helium component which is the most common in space. Moreover, all these planets have a deficit of water, methane, and ammonia – low-boiling compounds which are rather common in space. The main components of the planets of the earth-like group are silicates, iron, and compounds between iron and sulfur.

According to modern theories, the planets, satellites, asteroids, and comets were formed as a result of the evolution of a protoplanetary cloud, which at an early stage of its existence was a gas/dust cloud. In constructing the model of a planet, it is important to have an idea as to the chemical composition of the protoplanetary cloud in the region of formation of the planet. Theoretical calculations of the condensation of the protoplanetary cloud contain indications as to the possibility of chemical fractionation of iron, sulfur, and radioactive sources as a function of the temperature conditions of condensation at various distances from the Sun (Grossman and Larimer, 1974; Lewis, 1972). One of the main problems in constructing models of the earth-like planets is their comparative analysis and the production of quantitative estimates of the concentration of dust component in the original gas/dust cloud, as well as confirming the conclusions of the space-chemical schemes of condensation of this cloud.

By its mechanical parameters – mass M, mean radius R, and mean density ρ_0 – Venus is reminiscent of a twin planet of Earth. Thus it is perfectly natural that the first modern model of Venus, by Harold Jeffreys (1937), was based on the first modern model of Earth (Bullen, 1936). At that time, which now appears to us infinitely remote, the value

The Moon and the Planets **29** (1983) 139–175.   0165–0807/83/0292–0139$05.55.

of the mass $M = 4.91 \times 10^{27}$ g was exaggerated by 1% and the mean radius $R = 6150$ km was known with a good precision. In the double-layer model of the planet, the density ρ in the mantle varied from 3.29 to 5.44 g cm^{-3}, and that in the core from 9.6 to 11.1 g cm^{-3}; at the boundary between the mantle and the core, the pressure P was 1.24 Mbar, while at the center of Venus it was 2.4 Mbar; the mass of the core was 1.06×10^{27} g, and its radius was 2910 km. Comparing these figures with the parameters of the latest models of Venus (Table II), we may conclude that the mechanical model of the planet in 1937 [$\rho(l)$, $p(l)$, $g(l)$, where g is the acceleration of gravity and l is the depth) was very fortunate.

After the work of Jeffreys, in the classical monograph of Harold Urey (1952), a considerable enlargement was made of the conceptual basis, being used everywhere since then to construct models of planets, including those of the Earth group. Harold Urey emphasized the importance of cosmo-chemical data and cosmogonic concepts. It became customary to compare the composition of Venus not only with that of the Earth, but also with that of the meterorites. Interest grew toward the problem of the distribution of the concentration of iron and the other main elements in the planets of the Earth group. A survey of the works carried out in the 50's and 60's is given in the books of Levein (1970) and Bullen (1975). The basic conclusion of these works is that the interior structure of Venus is similar to that of Earth.

In very recent times, an increased interest has again been noted for the investigation of the internal structure of Venus (cf. Ringwood and Anderson, 1977; Kalinin and Sergeyeva, 1979; Zharkov and Zasurskiy, 1982). In the works of Ringwood and Anderson (1977) and Kalinin and Sergeyeva (1979) it was shown that an earth-like model of Venus should have an average density which is about 2% larger than the observed average density. In the work of Kozlovskaya (1982), carried out by our suggestion, a large number of mechanical models of Venus are considered in order to identify discrepancies in its composition as compared with the average chemical composition of Earth. The conclusions of this work and of a work devoted to the construction of a physical model of Venus (Zharkov and Zasurskiy, 1982; Zharkov *et al.*, 1981) will be described in greater detail below. In the work of Anderson (1980) a new interpretation was proposed for the average density of Venus, which is reduced from that of the Earth. Previously this fact had been interpreted as an indication of a difference in the bulk content of iron (Kovach and Anderson, 1965), sulfur (Lewis, 1972), or in the degree of oxidation of the mantle (Ringwood and Anderson, 1977). However, if we assume that Venus has a very thick outer basalt shell, and the corresponding basalt fraction of Earth by subduction into the mantle is present there in the Eclogite phase (Anderson, 1979), then the depressed average density of Venus is more likely produced by tectonic, rather than cosmo-chemical factors.

Since the time of the first publications of Jeffreys and Urey, geophysics has greatly changed its aspect (Zharkov and Trubitsyn, 1978, 1980; Schubert, 1979; Phillips and Ivins, 1979; Stacey, 1977; Zharkov, 1983). Now, we are interested not only in a mechanical model of the planet – i.e., the distribution of $\rho(l)$, $p(l)$ and $g(l)$ in its interior, but also to not a lesser extent in a physical model which gives the distribution of many physical parameters such as the heat capacity, coefficient of thermal expansion, adiabatic temperatures, coefficients of heat conduction and effective viscosity, etc. In connection with the development of works on the hydrodynamics of the interiors of the planets, the problem of determining the distribution of the temperature in these interiors has undergone significant changes (Toksöz *et al.*, 1978; Schubert, 1979; Zharkov, 1983). All these matters will be discussed below.

The accumulation of data on Venus and its analysis reveals not only a similarity between Venus and Earth, but also an important difference. There gradually emerges the fact that each planet is a unique individual.

In fact, both planets have different atmospheres, different histories of rotation and, consequently, different histories of tidal evolution. The Earth has a substantial magnetic field. The question as to the presence of a magnetic field on Venus is debatable, tending to indicate that the planet lacks such a field (Russell, 1980; Russell *et al.*, 1980). The magnetic moment of Venus is estimated at a maximum of about 5×10^{-5} of the magnetic moment of Earth (Russell *et al.*, 1980). The question as to whether the core of Venus is solid or liquid (not even mentioning the presence of a solid internal core) is important but not trivial. It is not yet possible to provide unambiguous answers to all these questions. But it is obvious that these problems can only be solved by constructing an actual model of the planet and the history of its evolution.

The tectonic type of development is also different for Venus and Earth (Masursky *et al.*, 1980). This is revealed in their structure and thickness of outer layers, ultimately resulting in the fact that Venus is aseismic (Zharkov *et al.*, 1981).

To a first approximation, the outer rigid layer of Earth (the lithosphere) can be divided into the oceanic and continental parts with thicknesses of ~ 80 km and ~ 200 km, respectively. But a more important difference between these type of lithosphere consists in their structure and thermal regime. The oceanic lithosphere, forming roughly 0.7 of the surface layer of Earth, contains a basalt crust with an overall thickness of ~ 6 km, while the average thickness of the continental crust is 35 km. The oceanic lithosphere represents a thermal boundary layer, created in the rift zones of the mid-oceanic ridges and engulfed in other areas, known as zones of subduction. Thus, the plate-tectonic regime of the Earth leads, on the one hand, to an effective cooling of the planet by the continual creation of a thermal boundary layer, the oceanic lithosphere; and on the other hand to a continual exchange of crust material between the crust and the mantle.

The tectonic regime of Venus and the other planets of the Earth group differs from the plate-tectonic. As a consequence of this fact, the Moon and, apparently, Mercury, Venus, and Mars should have significantly thicker crust layers than the Earth. The thickness of the lunar crust is ~ 60–100 km. The thickness of the crust in the other planets of the Earth group is probably within the same limits (this is mere supposition at present, which it is highly interesting to confirm).

A displacement of the center of the geometrical figure of a planet with respect to its center of masses may be interpreted as an indication of considerable regional variations

TABLE I

Observational data and parameters of the planetary figures of the earth-like group

	Venus	Earth	Mars	Mercury
Mass M, 10^{27} g	4.869	5.974	0.6422	0.3302
Equatorial radius R_e, km	6051.53* 6051.54*	6376	3400	
Mean radius R, km	6051.5	6371	3390	2439
Mean density, ρ_0, g cm^{-2}	5.25	5.514	3.94	5.44
$I^* = I/MR^2 \approx C/MR_e^2$	0.334*	0.33076	0.365 (0.375)**	–
Period of rotation τ, days	243.16	1.00	1.027	58.646
$q = 4\pi^2 R_e^3/GM\tau^2$	6.1×10^{-8}	3.47×10^{-3}	4.6×10^{-3}	1×10^{-6}
J_2, 10^{-6}	4.0 ∓ 1.5^a 5.97 ∓ 3.2^b	1082.64	1959	80 ± 60
J_2/q	65 ± 25 98 ± 52	0.31	0.43	80 ± 60
J_2^0, 10^{-6}	–	1072	1830	–
$J_2 = (J_2 - J_2^0)$, 10^{-6}	4.0 ± 1.5^a 5.97 ± 3.2^b	10	130	80 ∓ 60
α^{-1}		298.26	191.19	
e^{-1}		298.18	151.75	
C_{22}, 10^{-6}	$-(0.33 \pm 0.81)$	1.565	-54.9	–
$-S_{22}$, 10^{-6}	1.74 ± 0.74	0.894	-31.3	–
f, 10^{-6}	7.1	7.2	253	–
$g_0 = GM/R^2$, cm s^{-2}	887	982	373	370
$M = B_0 . R^3$, 10^{22} Gs cm^3	$< (0.43 \pm 0.2)$	7800	3.5	5
B_0, γ	< 2	30950	64	350
Temp. on surface, K(°C)	733(460)	277(3)	255(– 18)	435(182)

[a] From data of Akin *et al.* (1978).
[b] From data of Ananda *et al.* (1980).
* Values calculated theoretically, cf. text.
** Value for equilibrium model of Mars.

in the thickness of the crust. In Venus the distance between both centers is ~ (440 ± 120) m, which is much less than the correspond differences for Earth, the Moon, and Mars, which are of the order of a kilometer (Masursky *et al.*, 1980). Consequently, the variation of thickness of the Venusian crust is less than that of the other planets of the Earth type. This same fact may be regarded as one of the indications that the outer layers of Venus are closer to spherical symmetry than those of Earth.

The observational data for Venus is collected in Table I, where for comparison analogous information is given for Earth, Mars, and Mercury. Table I shows the masses M, the mean and equatorial radii R and R_e, the mean densities ρ_0, the non-dimensional moment of inertia $I^* = I/MR^2$, the period of rotation τ, the first coefficients for the expansion of the gravitational potential in spherical functions J_2 (Akim *et al.*, 1978; Ananda *et al.*, 1980), C_{22} and S_{22} (non-normalized coefficients) (Ananda *et al.*, 1980). At present, for Venus, an expansion of the gravitational field up to the sixth order inclusive ($m \leqslant n \leqslant 6$) has been found (Ananda *et al.*, 1980), although the accuracy of the determination is not large for any of the coefficients except J_2. Table I includes the value of the small parameter of the theory of figures† q, the dynamic flattening (the flattening of the outer equipotential surface of the gravitational potential of the planet α), and the geometrical flattening e, given by

$$q = \frac{\omega^2 R_e}{GM} = \frac{4\pi^2 R_e^3}{GM\tau^2}, \quad \alpha = \tfrac{3}{2}J_2 + \tfrac{1}{2}q, \quad e = \frac{R_e - R_p}{R_e}; \tag{1}$$

where ω is the angular speed of rotation, G is the gravitational constant, and R_p is the polar radius. For planets close to hydrostatic equilibrium (e.g., Earth or Mars), the mean radius R to a first approximation is expressed by R_e, in the formula $R = (1 - \alpha/3)R_e$. As is known (Zharkov and Trubitsyn, 1978; Zharkov, 1983), for an equilibrium planet the values of q and J_2 are of the same order of smallness. Referring to Table I, we discover that, for Venus, J_2 is larger than q by (65 ± 25) or (98 ± 52) times, while for Mercury it is (80 ± 60) times. Consequently, we may assert that Venus and Mercury are the most non-equilibrated planetary bodies in the solar system. This fact is evidently not random, since the rotation of both planets in the past was greatly retarded by tidal friction. If we assume that, for an effectively equilibrated Venus, the ratio of $J_2/q \sim 0.3$ (i.e., the same as the Earth, cf. Table I), then it is possible to determine $\alpha_0 = \frac{3}{2}J_2 + \frac{1}{2}q_0 \sim 3.17J_2 \sim (12.7 \text{ or } 18.9) \times 10^{-6}$. The corresponding equatorial radii R_e for Venus are shown in Table I. Thus we see that, for Venus, R_e should practically coincide with R.

The usual method of determining the moment of inertia of a planet from given J_2 and q is based on the formula of Radau–Darwin (Zharkov and Trubitsyn, 1978), asserting that

$$I^* = \frac{I}{MR^2} = \frac{2}{3}\left\{1 - \frac{2}{5}\left[5\left(1 - \frac{3}{2}\frac{J_2}{\alpha}\right) - 1\right]^{1/2}\right\}, \tag{2}$$

based on the assumption that the planet is close to a hydrostatic equilibrium. The scale of disequilibrium of Venus and Mercury prevents us from finding out their moments of inertia in this manner. The constants of precessions $H = MR_e^2 J_2/C = (C - A)/C$, where C and $A \approx B$ are the polar and equatorial moment of inertia, are also unknown for Venus and Mercury, and it is not clear whether H can be determined for both planets in the foreseeable future. Hence, it is not apparently possible to find the moment of inertia of Venus and Mercury from such observations in the near future.

The young Venus and Mercury, at an early epoch – when their rotation was not yet retarded by tidal friction – rotated much more quickly with a period of ~ 10 hr (Zharkov and Trubitsyn, 1978). Thus, the small parameter of the theory of figures of these planets, inversely proportional to the square of the period of rotation ($q \sim \tau^{-2}$), was much larger (roughly 4 orders of magnitude) for the young planets, than the present values. The observed value of J_2 for Venus and Mercury is approximately 70 times greater than q,

† q is equal to the ratio between the centrifugal acceleration at the equator $\omega^2 R_e$ and the gravitational acceleration GM/R_e^2. The larger q, the more strongly the centrifugal forces distend the planet in the direction of extension of the equatorial plane and the planet is thus contracted along the polar axis.

which may be regarded as certain relict values of this quantity, pertaining to the early and much larger values of q, when the rotation of the planets was not yet retarded by tidal friction to the present extent. And, since the shells of both planets had been able to cool considerably and become excessively rigid (or excessively viscous), the planetary figure 'froze', as it were, at a certain remote epoch and therefore does not conform to the present angular speed of rotation of the planet. If we solve the formula of Radau–Darwin with respect to the period of rotation from

$$\tau_{J_2} = \left\{ \frac{\pi}{\rho_0 G J_2} \left[\frac{5}{6.25(1-1.5I^*)^2 + 1} - 1 \right] \right\}^{1/2}, \tag{3}$$

where ρ_0 is the mean density, we are then able to estimate τ_{J2} for the epoch when the corresponding equilibrium figure of the planet was 'fixed', as well as the value of J_2, which has been retained to the present day. Assuming for the moment of inertia of Venus $I^* = 0.334$, a value obtained from model calculations (cf. item 2, Table II; also entered in Table I), we can find a certain paleoperiod of rotation of Venus, τ_{J2} (Venus) $\approx 16.9^{+0.3}_{-0.4}$ days or $13.9^{+6.5}_{-2.7}$ days, where J_2 has been taken from (Akim *et al.*, 1978) and (Ananda *et al.*, 1980). The obtained result suggests that Venus rotated more rapidly in the past. The period of rotation of the young Venus was probably even less and is equal to ~ 10 hr, although the disequilibrium of the planet, corresponding to so rapid a rotation, was apparently long ago lost from the 'memory' of Venus due the 'ductility' of its mantle and core.

Since the interior of the planets of the Earth group deviates from the state of hydrostatic equilibrium, the difference between their major momenta of inertia with respect to the axes in the equatorial plane, is not equal to zero. This difference can be calculated from the formula (Zharkov and Trubitsyn, 1978)

$$f = \frac{B-A}{MR^2} = 4\sqrt{C_{22}^2 + S_{22}^2}. \tag{4}$$

The results are shown in Table I. We see that the value f for Venus is small and close to the value of f for Earth. This indicates that the density distribution in both planets is close to the axisymmetrical, with good accuracy. Moreover, this fact strengthens the idea that the large non-equilibrium J_2 for the planet represents a relict value which corresponds to the more rapid rotation of Venus in a certain earlier epoch, as we have mentioned above.

Table I shows the values of the hydrostatic portion of the quadrupole moment J_2^0 for Earth and Mars (Zharkov, 1983) and the non-equilibrium value of the quadrupole moment $\Delta J_2 = J_2 - J_2^0$. For Venus and Mercury, $J_2 \gg J_2^0$ and $\Delta J_2 \sim J_2$. The value of the quantity ΔJ_2 enables an estimate of large-scale static tangential stresses in the interiors of the planets of the Earth group (Zharkov and Zasurskiy, 1981).

TABLE II

Parameters of Venus models for different compositions of the mantle and core

Venus crust: $M_c = 0.0183M$, $\Delta l_c = 70$ km, $\rho_c = 2.89$ g cm^{-3}							Earth PEM-C $\Delta l_c = 35$ km
	Core – MCE			Core – Fe			
$\Delta\rho$%	0	−4	−8	0	−4	−8	0
ρ_0, g cm^{-3}	3.26	3.13	3.00	3.26	3.13	3.00	3.30
M_1, %	22.3	23.3	24.3	22.3	23.3	24.3	18.4
M_2, %	47.8	41.7	35.7	49.8	43.9	38.1	49.0
l_1, km	481	548	618	581	548	618	420
l_2, km	756	825	900	756	825	900	670
M_{core}, %	29.9	35.0	40.0	27.9	32.8	37.6	32.6
r_{core}, km	3210	3388	3548	3076	3239	3394	3486
ρ, g cm^{-3}	{5.39	5.17	4.96	5.45	5.24	5.03	5.55
	{9.59	9.45	9.32	10.42	10.30	10.17	9.91
P_c, kbar	1158	1066	978	1231	1144	1058	1354
ρ, g cm^{-3}	11.7	11.8	11.9	12.3	12.4	12.5	13
p, kbar	2878	2987	3087	3006	3131	3246	3632
$I^* = I/MR^2$	0.334	0.328	0.321	0.333	0.326	0.319	0.330 89
Σ Fe(⊕ − ♀)	2.2	1.7		0.5–0.4			

Note: ρ_0 = Density of mantle under normal conditions, M_1 and M_2 = masses of upper and lower mantle, l_1 and l_2 = depth of first and second phase transitions in mantle; MCE = material of the core of the Earth; the last row gives the difference between the total iron content in the models of Earth and Venus with mantles of pyrolite composition.

2. Models of Venus. Earth-like Models

The creep limit of rocks is ~ 10^3 bar, or even less under conditions of high temperatures and 'geological' time intervals. Therefore the planetary figures are close to the hydrostatic equilibrium, in the sense that the non-equilibrium stresses in the majority of the planetary interiors is much less than the hydrostatic pressure $p(r)$, where r is the distance to the center of the planet. As a result, in analyzing the models, it is possible to use the equation of hydrostatic equilibrium

$$\frac{\mathrm{d}p}{\mathrm{d}r} = -\frac{GM(r)}{r^2}\rho(r) = -g(r)\rho(r), \tag{5}$$

and the equation for $M(r)$, the mass of a sphere of radius r

$$\frac{\mathrm{d}M}{\mathrm{d}r} = 4\pi r^2\rho(r). \tag{6}$$

The boundary conditions assert that

$$\begin{aligned} &\text{on the surface} \quad r = R, \quad M(R) = M, \\ &\text{in the center} \quad r = 0, \quad M(0) = 0; \end{aligned} \tag{7}$$

in which M and R are the known mass and mean radius of the planet. Unfortunately, the moment of inertia cannot be found from the observations of Venus: and, therefore, the analysis of its model does not make use of the equation for the moment of inertia, as is done for Earth and Mars. In order to close the set of Equations (5) and (6), we must join an equation of state of the form

$$\rho = \rho(p). \tag{8}$$

We must point out the difference in the determination of $\rho(r)$ for Earth and for the other planets. In the case of Earth, from seismology we know the quantity

$$\Phi = K_S/\rho = V_P^2 - \tfrac{4}{3}V_S^2$$

as a function of the radius (K_S is the adiabatic modulus of compression, V_P and V_S are the velocities of the longitudinal and transverse seismic waves). By use of the equation of Adams–Williamson, it is now possible to calculate an actual model of the Earth and thus determine the equation of state of the material of Earth $p(\rho)$, using only geophysical data (Bullen, 1975; Zharkov and Trubitsyn, 1978). In the case of the planets, the quantity Φ is unknown and therefore it is necessary to know the equation of state $p(\rho)$, which produces the law by which the material of the planet is compressed beneath the weight of the overlying strata. The analysis of the models of Venus makes use of the equation of state for the material of Earth, as well as the equation of state of Fe, FeS, MgO, FeO, SiO_2, Al_2O_3, etc., determined from dynamic and static experimental data (Zharkov *et al.*, 1975; Zharkov and Trubitsyn, 1978).

Let us explain the specific procedure of constructing a model of Venus in accordance with the works of (Kozlovskaya, 1982; Zharkov *et al.*, 1981; Zharkov, 1983.† Few data is available for Venus. Under such conditions – considering that Venus is a twin planet of Earth in regard to the mass and radius – it is reasonable for the construction of a model of the planet to use the equation $p(\rho)$ for Earth as the initial equation of state (Dziewonski *et al.*, 1975; Zharkov and Trubitsyn, 1978). For Earth, at present, there also exist more detailed models (Dziewonski and Anderson, 1980). However, in constructing a model of Venus it is not important which of the similar models of Earth is taken as the foundation. The convenience of such a choice, furthermore, automatically takes into account the influence of temperature on the equation of state, as the temperature distribution in both planets is evidently similar for depths greater than ~ 200 km. Specifically, the dependence $p(\rho)$ was chosen as the equation of state for the PEM-C model (Dziewonski *et al.*, 1975). In constructing models for the silicate mantle of Venus, both 'reduced' and 'weighted' equations of state were used in relation to $p(\rho)$ of PEM-C. These curves are shown in Figure 1 and are described by the simple formulae

$$\rho(p) = \rho_1(p) + \Delta\rho, \qquad \Delta\rho = \text{const.},$$

while $\rho_1(p)$ is the equation of state as per the PEM-C model. Thus, all the curves of

† In the analysis of the models of the planet, its mean radius is assumed to be $R \approx 6050$ km.

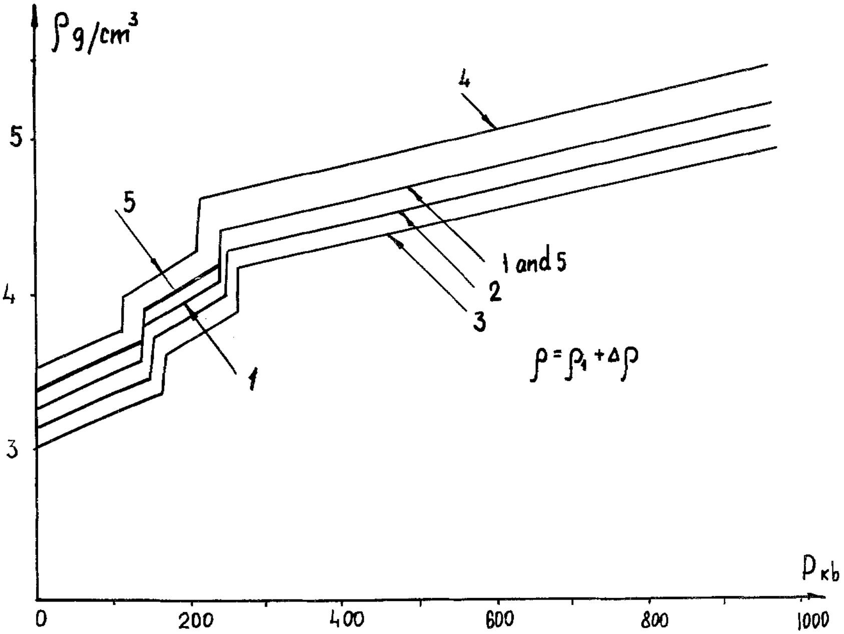

Fig. 1. Set of equations of state of $\rho(p)$ for analysis of silicate mantles of models of Venus. Curves (1–4) for models with minimum crust, $\Delta l_K = 38$ km, $M_K \approx 1\%M$, $\bar{\rho}_K = 2.8\,\mathrm{g\,cm^{-3}}$. (1) [$\Delta\rho = 0$, $\rho(p) = \rho_1(p)$], (2) [$\Delta\rho = -4\%\rho_1$], (3) [$\Delta\rho = -8\%\rho_1$], (4) [$\Delta\rho = 8\%\rho_1$], (5) ($\Delta\rho = 0$) for models with maximum crust, $\Delta l_K = 127$ km, $M_K \approx 3.3\%M$, 15% of the material of the upper mantle is melted. $\rho(p)$ obtained by interpolation for intermediate values of Δl.

Figure 1 are produced by a parallel shift along the ρ axis by the amount $\Delta\rho$. Specifically, $\Delta\rho$ was varied within limits of $\pm$ 8% of ρ_1, while the actual value was assigned to the models with $\Delta\rho < 0$, which corresponds to the cosmo-chemical data, in accordance with which the content of iron in the silicates of the mantle should diminish systematically in the transition from Mars to Mercury (Lewis, 1972). The models of Venus, constructed with equations of state having $\Delta\rho < 0$ are interpreted as possessing a deficit of iron in the silicates of the mantle. A reduction of $\Delta\rho$ by 1% corresponds to a reduction of 1.4% in the content of iron in the silicates of the mantle. In accordance with the data of Venera 10, the density of the surface rocks on Venus is $2.8 \pm 0.1\,\mathrm{g\,cm^{-3}}$, which corresponds to the basaltic rocks (Surkov *et al.*, 1977), cf. also (Masursky *et al.*, 1980). We shall cite below models of the planet with *a priori* given thickness of the crust $\Delta l_c = 70$ km. The reasons for assuming so thick a crust have been discussed in the introduction. The mass of the crust is $M_c \approx 1.8\%M$ and corresponds to a melting of 8.2% of the material of the upper mantle (the boundary of the upper mantle is located at the depth of the second phase transition). The selection of the equations of state for the Cytherean core is more complicated. Here, also, $\rho(p)$ for the PEM-C model was assumed at the start [$\rho_{\mathrm{MCE}}(p)$, where MCE is the material of the core of the Earth]. Moreover, according to the cosmo-chemical data, the cores of the planets of the Earth group may contain FeS, nor is it

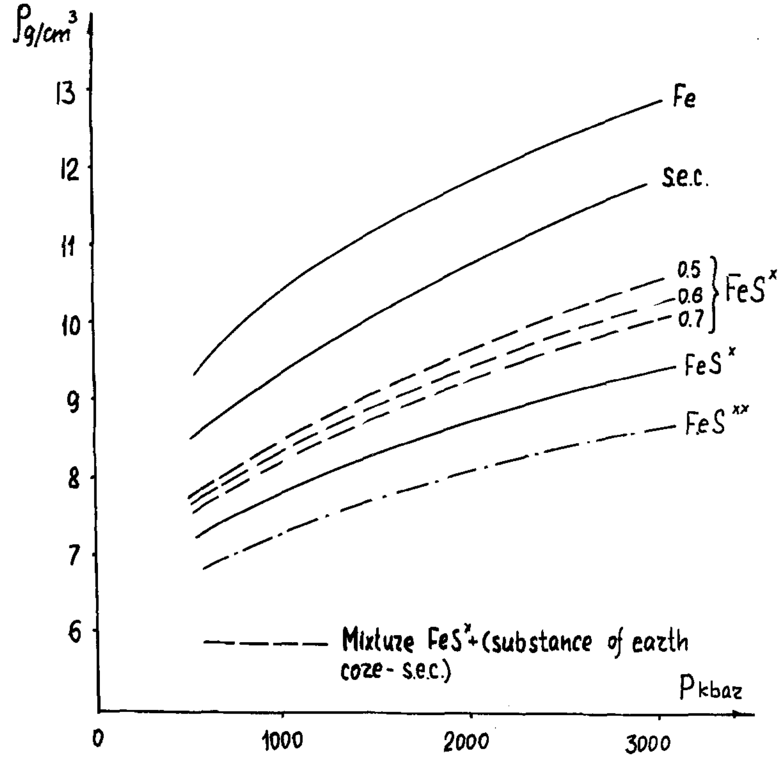

Fig. 2. Set of equations of state for analysis of the core of Venus models: molten iron Fe, MCE – material of the core of Earth, FeS* and FeS** – with an allowance for the thermal expansion and value of the coefficient of thermal expansion, $\alpha = 2 \times 10^{-5}$ and 4×10^{-5} deg^{-1}, respectively; mixture of MCE and FeS*, in which the portion of FeS* comprises 0.5, 0.6, and 0.7.

excluded. that the cores of Venus and Mercury may consist of pure iron. Therefore the equations of state, shown in Figure 2, include the alloys Fe – FeS, FeS, MCE and Fe. For Fe, Figure 2 shows a curve for the adiabatic temperature distribution in the core with temperature at the mantle/core boundary equalling 3500 K. The equations of state were taken from Zharkov *et al.* (1975) and Zharkov and Trubitsyn (1978). The equation of state, taken for the high-pressure phase of iron, is practically identical with the equation for ϵ-Fe, obtained in (Brown and McQueen, 1981). The equation of state for FeS differs somewhat from that obtained in Ahrens (1979), but this is not important, due to the small content of FeS in the core of Earth being much less in the core of Venus. According to modern cosmo-chemical theories, the content of FeS in the interiors of the planets of the Earth group should diminish in the transition from Mars to Mercury. Since the admixture of FeS to Fe in the core of the Earth is not large (of the order of 10%), if it is even present at all, it is logical to use the equations of state for the MCE and for Fe as the two extreme cases when constructing the models of Venus. Table II shows the main parameters of the models of Venus with a core of MCE and a core of molten iron,

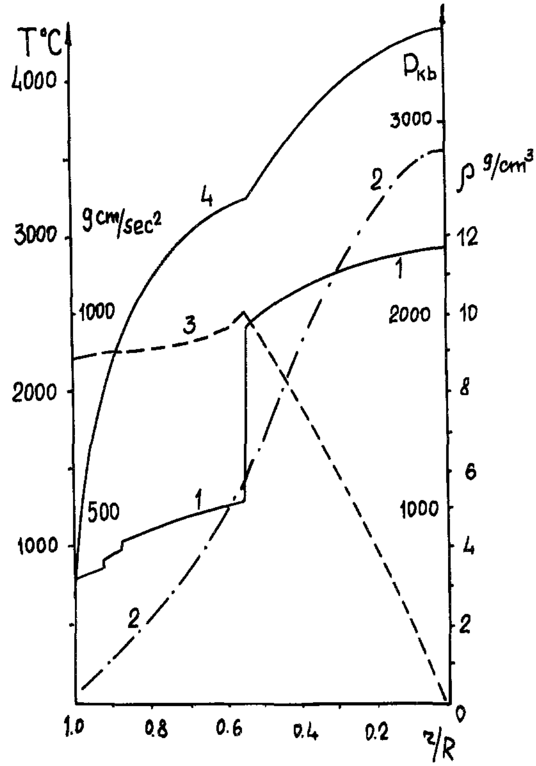

Fig. 3. An earth-like model of Venus. Curves 1, 2, 3, and 4 are distribution of the density ρ, pressure p, gravity g, and temperature T, respectively, along the radius. In the core the temperatures follow the adiabatic curve.

and for comparison data for the PEM-C model of the Earth are also shown. Figure 3 shows the distribution of density, pressure, gravity, and the temperature in an earth-like model of Venus (numerical parameters given in Table II, $\Delta\rho = 0$, core of MCE). The temperature distribution was obtained from *a priori* considerations. Assuming a thickness of 200 km for the Cytherean lithosphere, we shall take a temperature of ~ 1200 °C for this depth. At the boundary between the mantle and the core the temperature was assumed to be ~ 3500 K (~ 3230 °C). The temperatures in the core were regarded as adiabatic and were computed from Equation (27). As a result, the temperature in the center of Venus was found to be about 4670 K (4400 °C). The trial temperature distribution in Figure 3 in the mantle is close to the distribution obtained in (Toksöz *et al.*, 1978), while in the core it is appreciably higher and at the center of Venus the discrepancy attains ~ 1200 °C. It is natural that the question as to the temperature distribution in the interior of Venus is even more vague than the interior of Earth (Zharkov, 1983). We shall discuss the temperature distribution of Venus in greater detail in the section devoted to convection in the mantle of the planet.

In order to classify the silicate mantle of Venus into mineralogical zones, we use the

TABLE III
Mineralogical zones in the mantle of Venus

	Depth, km	Principal mineral phases		
Upper mantel	70		Pyroxene + garnet	
(Zone B)	olivine zone 480	Olivine (α-phase)	+ Al_2O_3 → garnet	
Transitional zone	Spinel zone	β-phase → spinel		
(Zone C)		(γ-phase)	Garnet	
	760 zone of ilmenite	Ilmenite → perovskite +	Ilmenite	Ilmenite
	and perovskite	ferropericlase	perovskite	perovskite
	1000	(Mg, Fe)O		
Lower mantle	Perovskite zone	Perovskite +	Perovskite	Perovskite
(Zone D)		(Mg, Fe)O		
	2840			

phase diagrams of the systems Mg_2SiO_4–Fe_2SiO_4, $MgSiO_3$–$FeSiO_3$ and $MgSiO_3$–Al_2O_3. These data have been collected in books (Zharkov and Trubitsyn, 1978, 1980) and were basically gathered by Akimoto *et al.* (1975) and Liu (1977). The mineralogical zones in the mantle of Venus are given in Table III. In comparing the structures of Venus and the Earth, we can only conclude that both planets are alike except for the thicker crust of Venus and the deeper location in it of the boundaries of the first and second phase transitions in the mantles. Referring to Table II, we see that the content of iron in the earth-like models of Venus is 2% less than that in the Earth. In the introduction we pointed out the tectonic interpretation of this fact in the work of Anderson (1980). Only in models of Venus with a core of molten iron is the overall content of Fe the same as in the Earth. Despite the insufficient data on Venus, we may still form the provisional conclusion that the general tendency of monotonic decrease of the general content of iron in the transition from Mercury to Mars seems to have been disturbed. This is somewhat surprising in the context of modern theories as to the condensation of the protoplanetary cloud and the subsequent formation of the planets. If this finding is confirmed, it will be necessary to treat it as a new boundary condition in the problem of the origin and evolution of the planets of the Earth group.

2.1. THE PARAMETRICALLY SIMPLE MODEL OF VENUS – PVM†

For Venus, as for Earth, it is useful to construct a parametrically simple model (PVM), in which the density distribution $\rho(x)$ and the velocities of the longitudinal $V_p(x)$ and transverse $V_s(x)$ volume waves are given by piecewise-continuous analytical functions of the non-dimensional radius $x = r/R$. The continuous segments of the distributions are described by polynomials in x, not larger than the third degree. Knowing the distribution of density $\rho(r)$ (Fig. 3) in an earth-like model of Venus and describing the continuous segments of this distribution by polynomials in the radius, it is easy to construct a seismic

† PVM – Parametric Venus Models.

TABLE IV

Parametrically simple model of Venus

l, km	ρ, g cm^{-3}	V_p, km s^{-1}	V_s, km s^{-1}
70–481	$7.374 - 4.146X$	$27.17 - 19.74X$	$14.4 - 10.4X$
481–756	$10.101 - 6.871X$	$19.32 - 10.59X$	$13.54 - 9.21X$
756–2840	$6.77 - 2.467X - 0.266X^2$	$14.84 - 0.074X - 5.011X^2$	$6.83 + 2.65X - 3.95X^2$
2840–6050	$11.742 - 0.17X - 5.402X^2 - 3.642X^3$	$9.98 - 0.66X - 9.125X^2 - 0.74X^3$	

$\rho_{crust} = 2.8\,\mathrm{g\,cm^{-3}}$; $x = r/R$; $R = 6050$ km

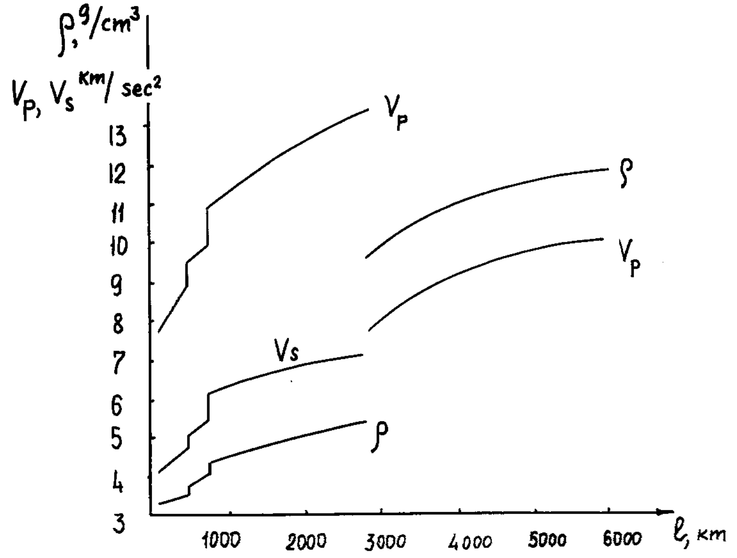

Fig. 4. A parametrically-simple Venus model (PVM). The distributions of V_P, V_S, and ρ are shown.

model of the planet. For this, we should use the functions $V_p(\rho)$ and $V_s(\rho)$ of the PEM-C models. A PVM model, designed in this fashion, is characterized by Table IV. This is shown graphically in Figure 4 (Zharkov and Zasurskiy, 1982).

3. A Physical Model of Venus

In this section we shall explain the thermodynamics of the mantle and core and estimate the coefficients of thermal and electric conductivity, the mechanical quality Q_μ, and the large-scale static stresses in the mantle. A rheological model of the crust and mantle is considered.

3.1. The Thermodynamics of the Mantle

Let us consider the thermodynamical parameters of the mantle on the basis of the Debye model (Zharkov and Trubitsyn, 1978, 1980; Zhakov and Kalinin, 1971; Zharkov and Zasurskiy, 1980a, b). Let us define the average speed of sound V in terms of the velocity of the seismic waves V_p and V_s by

$$\bar{v} = 3^{1/3}[1 + 2(V_P/V_S)^3]^{-1/3} V_P; \tag{9}$$

then, assuming an average atomic weight of $A \approx 22$ for the mantle of Venus, we shall find the Debye temperature by means of the standard formula (cf Zharkov and Zasurskiy, 1980a, b)

$$\theta = 0.924 \times 10^{-3} \bar{v} \rho^{1/3}. \tag{10}$$

In this equation, θ is found in K with $\bar{v}$ in cm s^{-1} and ρ in g cm^{-3}. The second function needed to calculate the thermodynamic coefficients is γ, the Grüneisen parameter

$$\gamma = \frac{\mathrm{d}\ln\theta}{\mathrm{d}\ln\rho}. \tag{11}$$

In the classical limiting case of $T > \theta$, the specific entropy S is related to θ by the simple formula

$$S = \frac{R}{A}[4 - 3\ln\theta/T], \quad \frac{\theta}{T} < 1, \tag{12}$$

where R is the universal gas constant. From this, we obtain the adiabatic equation ($S = \text{const.}$)

$$T_{ad} = T_{ad0}(\theta/\theta_0), \tag{13}$$

where the index 0 designates the value with respect to the initial counting level, usually the depth $l_0 = 100$ or 200 km, where T_{ad} is supposed to be ≈ 1500 K. For the Debye model, Equation (13) is equivalent to the standard thermodynamic formula for an adiabatic gradient

$$\left(\frac{\mathrm{d}T}{\mathrm{d}l}\right)_{ag} = \frac{g\alpha T}{C_p}, \tag{14}$$

where T is the absolute temperature, l is the depth, g is the acceleration of gravity, α is the volume coefficient of thermal expansion, and C_p is the specific heat capacity at constant pressure.

The fusion temperature of the planetary interiors is usually determined by means of the formula of Lindemann (cf. Zharkov and Trubitsyn, 1978)

$$T_m(\rho) = T_{m0}(\theta/\theta_0)^2(\rho_0/\rho)^{2/3}, \tag{15}$$

where the index 0 refers to the starting level of the measurement. The distributions of $\theta(l)$, $T_{ad}(l)$, and $T_m(l)$, thus calculated, along with the trial temperature $T(l)$, are shown in Figure 5. The inset of Figure 5 shows a graphical image of the Grüneisen parameter $\gamma(l)$, found from Equations (11) and (10), and the PVM model (Table IV). Since the function $\gamma(l)$ is determined by differentiation of the function $\theta(l)$ [Equation (10)], it may contain errors which are difficult to spot. Therefore, along with function $\gamma(l)$, calculated in the above manner, it is useful to employ a function which is determined from classical averaged models of Earth, which have a continuous velocity distribution without jumps in the zone of phase transitions. The function $\bar{\gamma}(l)$, calculated for classical models for the mantle of the Earth, decreases uniformly from a value of ~ 1.9 at a depth of 100 km to values of ~ 1.0 km at the boundary between the mantle and core of Earth. The averaged (smoothed-out) Grüneisen parameter for the mantle of Venus $\bar{\gamma}(l)$ is likewise shown in the inset of Figure 5 for comparison. The PVM model (Table IV) allows the calculation of the adiabatic modulus of compression K_S from

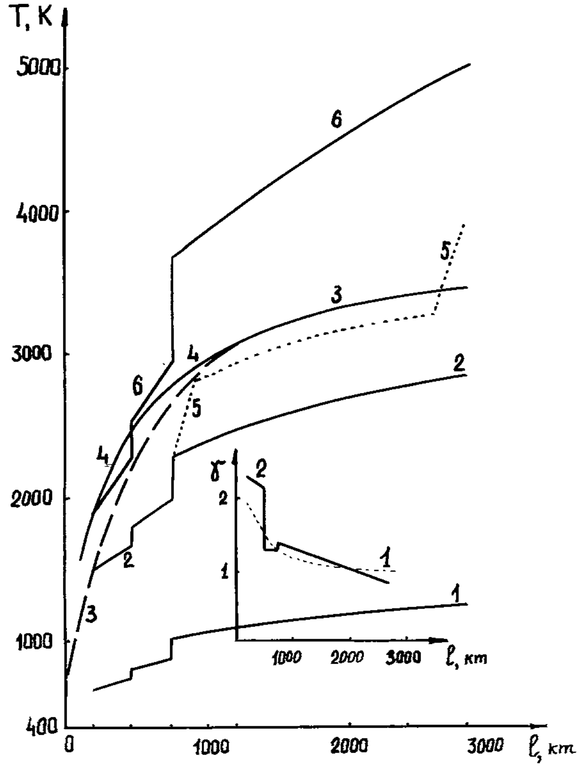

Fig. 5. The temperature in the mantle of Venus. (1) Debye temperature, (2) adiabatic temperatures, (3) trail smoothed-out temperatures, (4) temperatures obtained in analysis of the thermal history of the planet (Maeva, 1969), (5) temperatures with thermal boundary layers at the borders of the lower mantle, (6) melting temperature of the mantle. The inset shows the Gruneisen parameters as a function of the depth. (1) $\bar{\gamma}$ as per the smoothed-out classical models, (2) γ as per PVM models.

$$K_S = \rho\Phi = \rho(V_P^2 - \tfrac{4}{3}V_s^2), \tag{16}$$

where $\Phi(l)$ is the seismic parameter of the mantle of Venus. In the classical limiting case $(T > \theta)$, the Debye model gives, for the thermal pressure (Zharkov and Kalinin, 1971)

$$p_T = \frac{3RT\gamma\rho}{A}; \tag{17}$$

and, correspondingly, for the thermodynamic derivative

$$\left(\frac{\partial p}{\partial T}\right)_\rho = \frac{3R\gamma\rho}{A}. \tag{18}$$

From given K_S, the thermodynamic formula can be used to calculate K_T from

$$K_T = K_S - \frac{T}{\rho C_v}\left(\frac{\partial p}{\partial T}\right)^2; \quad C_v = \frac{3R}{A} \tag{19}$$

or, by using Equation (18), we can find

$$K_T = K_S - \frac{3R\rho\gamma^2 T}{A}. \tag{20}$$

The coefficient of thermal expansion α can be calculated by means of the formula

$$\alpha = K_T\left(\frac{\partial p}{\partial T}\right)_\rho \approx \frac{3R\gamma\rho}{A}K_T, \tag{21}$$

while the heat capacities C_p and C_v are related by

$$C_p \approx C_v(1+\gamma\alpha T). \tag{22}$$

The ratios between the coefficients of compressibility β_T and β_S, the heat capacities, and the moduli of compression and related by the thermodynamic formula

$$\frac{C_p}{C_v} = \frac{\beta_T}{\beta_S} = \frac{K_S}{K_T}. \tag{23}$$

Values of the thermodynamic parameters of the mantle of Venus, calculated by formulas (9)–(23), are summarized in Table V; Table V shows the jumps of the Debye temperature at the first ($l_1 = 481$ km) and second ($l_2 = 756$ km) phase transitions, $(\Delta\theta)_1 = 55$ K and $(\Delta\theta)_2 = 126$ K, and the corresponding jumps in the adiabatic temperatures $(\Delta T_{ad})_1 = 124$ K and $(\Delta T_{ad})_2 = 285$ K. Equation (12) allows the estimation of the entropy jumps upon phase transitions in the mantle of Venus

$$\Delta S = S_2 - S_1 = \frac{3R}{A}\ln\frac{\theta_1}{\theta_2}. \tag{24}$$

Using equation (24) we obtain the slope of the curve of the phase equilibrium

$$\lambda = \frac{\mathrm{d}p}{\mathrm{d}T} = \frac{\Delta S}{V_1 - V_2} = -\rho_1\rho_2\frac{\Delta S}{(\rho_2-\rho_1)} = -\frac{\rho_1\rho_2}{\Delta\rho}\frac{q}{T}. \tag{25}$$

This was found to equal ~ 51 bar K^{-1} and ~ 90 bar K^{-1} at the first and second phase transition in the mantle, respectively. According to Equation (25), the thermal effect at the first phase transition ($T_1 \sim 2160$ K) is equal to $(-q)_1 \approx 5.9 \times 10^8$ erg g^{-1}, which corresponds to a heating by $(\Delta T)_1 \sim (-q)_1/C_p \sim 49$ K ($C_p \sim 1.2 \times 10^7$ erg g^{-1}). The corresponding values for the second phase transition ($T_2 \sim 2660$ K) are $(-q)_2 \approx 13.4 \times 10^8$ erg g^{-1}, $(\Delta T)_2 \sim (-q)_2/C_p \sim 112$ K.

The PVM model, as well as the PEM models for Earth, is a schematized and simplified model of the planet. In particular, the jumps in the parameters at the boundaries of the first and second phase transitions in the mantle of Venus should in fact be obliterated, the same as in the mantle of Earth at $(\Delta l)_1 \sim 50$ km and $(\Delta l)_2 \sim 100$ km (Zharkov and Zasurskiy, 1980a, b).

TABLE V

Values of the thermodynamic parameters of the mantle of Venus

l km	ρ g cm^{-3}	p Mbar	θ K	T_{ad} K	T_m K	γ	Φ km^2 s^{-2}	K_S Mbar	μ Mbar	$\bar{\gamma}$	$F(\rho)$	β_T, 10^{-6} bar^{-1}	α, 10^{-5} deg^{-1}
200	3.365	0	663	1500	1862	2.28	40.1	1.35	0.64	1.9		0.81	6.07
304	3.453	0.0895	697	1576	2018	2.25	43.7	1.57	0.71	1.75	1.36	0.77	5.52
480	3.55	0.1416	747	1690	2281	2.19	49.6	1.76	0.82	1.53	2.19	0.65	4.13
480	3.763	0.1416	802	1814	2524	1.3	57.1	2.15	0.96	1.53	2.63	0.52	3.52
626	3.924	0.1891	846	1914	2735	1.3	59.1	2.32	1.09	1.37	3.85	0.48	3.01
756	4.077	0.2391	890	2013	2949	1.3	60.8	2.48	1.22	1.28	5.00	0.4	2.48
756	4.377	0.2395	1016	2298	3668	1.39	69.4	3.04	1.64	1.28	7.26	0.36	2.34
1212	4.62	0.4266	1087	2459	4050	1.26	79.2	3.66	1.9	1.08	12.27	0.29	1.7
1817	4.918	0.6816	1164	2633	4456	1.09	91.9	4.52	2.22	1.03	15.90	0.23	1.4
2843	5.388	1.158	1274	2882	5017	0.78	111.7	6.02	2.73	1.01	21.89	0.17	1.11

$C_V = 1.187 \times 10^7$ (erg/g · deg)

3.2. The Thermodynamics of the Core

Let us discuss the thermodynamics of the core of Venus on the basis of the Debye model. The Grüneisen parameter of the core of Earth has been studied in Zharkov and Zasurskiy (1980b), cf. also Zharkov (1962) or Jacobs (1975). There it was shown that $\gamma(\rho)$ in the core diminishes smoothly from 1.6–1.5 at the boundary with the mantle of the Earth down to ~ 1.3 at the center of the Earth. Roughly the identical range of variation of the function $\gamma(\rho)$ can also be assumed for Venus. In consideration of the weak dependence of the function $\gamma(\rho)$ on the density, the same as in Zharkov and Zasurskiy (1980a), we shall assume for Venus a constant mean value $\bar{\gamma} = \bar{\gamma}(\rho) = 1.45$. Then the Debye temperature of the interior of Venus can be computed by the equation

$$\theta(\rho) = 1070(\rho/\rho_0)^{\bar{\gamma}} = 1070(\rho/\rho_0)^{1.45}\,\text{K}, \tag{26}$$

where $\rho_0 = 9.59\,\text{g cm}^{-3}$ is the density of the core at the boundary with the mantle, ρ, in g cm^{-3} and θ is in K. In the calculation of $\theta_0 = 1070\,\text{K}$, the Poisson coefficient was assumed to be $\sigma = 0.3$, while the mean molecular weight of the core material was $A = 56$. In accordance with Equations (13) and (26), the adiabatic curve of the core of Venus is of the form

$$T_{ad}(\rho) = T_{ad0}(\theta/\theta_0) = T_{ad0}(\rho/\rho_0)^{1.45}. \tag{27}$$

We shall define the equation of the melting curve of the core of Venus by the formula of Lindemann [Equations (15) and (26)], of the form

$$T_m(\rho) = T_{m0}(\theta/\theta_0)^2(\rho_0/\rho)^{2/3} = T_{m0}(\rho/\rho_0)^{2.24}. \tag{28}$$

In Equations (27) and (28), T_{ad0} and T_{m0} are the temperatures at the beginning of the adiabatic curve and the melting curve, respectively. Using Equations (27) and (28), we find it easy to compute the adiabatic gradient and the gradient of the melting curve form

$$\frac{\mathrm{d}T_{ad}}{\mathrm{d}l} = 1.45T_{ad0}\,\frac{g}{v_P^2}\left(\frac{\rho}{\rho_0}\right)^{1.45}, \quad \frac{\mathrm{d}T_m}{\mathrm{d}l} = 2.24T_{m0}\,\frac{g}{v_P^2}\left(\frac{\rho}{\rho_0}\right)^{2.24}, \tag{29}$$

where g is the gravity, v_P is the velocity of longitudinal waves in the molten core of Venus. In the Earth, the adiabatic curve of the convective external core intersects the melting curve at the boundary with the solid internal core. The question as to whether Venus possesses a solid internal core remains open at present. However, since both planets are very similar, it is reasonable to assume that Venus also has a small solid internal core, the boundary of which we shall arbitrarily place at a depth of ~ 5000 km. Then the adiabatic curve of the outer molten core of Venus should intersect the melting curve at $l = 5000$ km. This circumstance was used to construct Figure 6. Figure 6 shows trial adiabatic curves for the core of Venus with initial temperatures T_{ad0} at the boundary with the mantle equaling 3500, 4000, and 4500 K. The corresponding temperatures at the boundary between the outer and inner core will be 4550, 5200, and 5850 K. Also shown by broken lines are the corresponding melting curves of the core, computed from

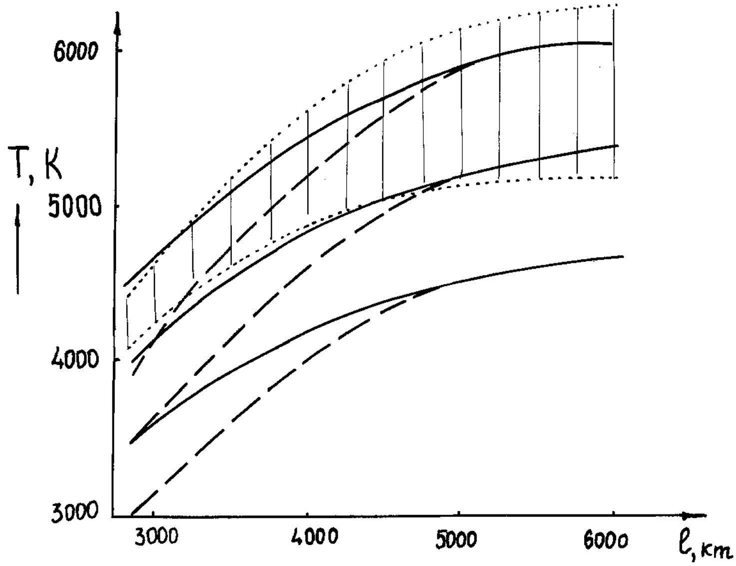

Fig. 6. Trial adiabatic temperatures and melting curves of the core. Solid lines of the adiabatic curve with values of T_{ad0} at the core/mantle boundary equal 3500 K, 4000 K and 4500 K. Dashes – melting curves with values of T_{m0} at the core/mantle boundary equaling 3000 K, 3450 K and 3900 K. Also shown is the experimental zone for the iron melting curve after Brown and McQueen (1981).

Equation (28). The values of the melting temperature at the boundary with the mantle were found to be 3000, 3450, and 3900 K. Figure 6 also shows an experimental zone for the melting temperatures of iron according to the data of Brown and McQueen (1981). We can see that, for the molten core of Venus with a composition close to pure iron, only the adiabatic curve with initial temperature at the boundary between the core and mantle equalling $T_{ad0} \sim 4500$ K is realistic. From this, the problem arises – how to coordinate such high temperatures with a convective mantle of the planet. This question is discussed in greater detail in Section 4.

A knowledge of the Grüneisen parameter γ and the Debye temperature θ allows the calculation of all the thermodynamic coefficients of the core by the formulas given in the preceding paragraph, $A = 56$. The results of the calculations are summarized in Table VI.

Let us now calculate the heat flux, emitted by the core of Venus by the mechanism of molecular thermal conduction (since the core by hypothesis is in the convective state, this will provide the lower limit for the heat flux from the core). For this, let us multiply $\mathrm{d}T_{ad}/\mathrm{d}l$ [Equation (29)] at the boundary with the mantle by the coefficient of thermal conduction of the core:

$$\begin{aligned} LT\sigma_c &= \kappa_c, \quad (L = 0.245 \text{ erg} \cdot \text{s}^{-1} \cdot \text{Ohm} \cdot \text{K}, \\ \sigma_c &= 3 \times 10^3 \text{ Ohm}^{-1} \cdot \text{cm}^{-1}, \; T = 3500 \text{ K}, \\ \kappa_c &= 0.26 \times 10^7 \text{ erg} \cdot \text{cm}^{-1} \cdot \text{s}^{-1} \cdot \text{K}^{-1}), \end{aligned}$$

TABLE VI
Values of the thermodynamic parameters of the core of Venus

l, km	ρ, g cm^{-3}	p, Mbar	θ, K	T_{ad} K	K_S, Mbar	β_T, 10^{-7} bar^{-1}	α, 10^{-5} deg^{-1}
2843	9.591	1.158	1071	3480	5.62	1.89	1.21
3027	9.858	1.323	1114	3620	6.21	1.71	1.11
3330	10.25	1.586	1178	3830	7.17	1.47	0.97
4237	11.10	2.268	1324	4300	9.69	1.08	0.74
5447	11.67	2.805	1424	4027	11.56	0.9	0.62
6050	11.74	2.874	1436	4666	11.69	0.88	0.6

where σ_c is the coefficient of electrical conduction of the core, L is the Lorentz constant (Zharkov and Tribitsyn, 1978), and the area of the surface of the core is $4\pi R_c^2$, where $R_c = 3210$ km. As a result, we find $Q_c = 2.86 \times 10^{19}$ erg s^{-1} $= 0.9 \times 10^{27}$ erg y^{-1}.

3.3. Thermal Conductivity. Temperature Distribution in the Lithosphere and the Heat Flux from the Interior of Venus

Estimates of the coefficient of thermal conduction of the metal core of Venus have been given in a preceding paragraph. A more complicated problem is the estimation of the thermal conduction of the mantle of the planet. Below we shall discuss the molecular coefficient of thermal conduction, which determines the heat transfer in the lithosphere and the thermal boundary layers of the planetary interior (cf. Section 4).

When calculationg the temperatures in the lithosphere of Earth, often for the coefficient of thermal conduction κ in the upper mantle we use formulas which smooth out the experimental data in the temperature range of 500–1400 K (Schatz and Simmons, 1972). These formulae presuppose that the properties of the mantle are approximated by dunite with an olivine composition $(Mg_{0.9}Fe_{0.1})_2SiO_4$.

The effective thermal conduction of the mantle is represented in the form of the sum of the lattice κ_L and radiant κ_R parts, $\kappa = \kappa_L + \kappa_R$, as

$$\kappa_L = 10^3/(74 + 0.5\,T), \tag{30}$$

$$\kappa_R = \begin{cases} 2.3 \times 10^{-3}(T - 500) & \text{when } T > 500\,\text{K}, \\ 0 & \text{when } T < 500\,\text{K}, \end{cases} \tag{31}$$

where κ is in units of W m^{-1} K^{-1} (1 W m^{-1} K^{-1} $= 10^5$ erg cm^{-1} s^{-1} K^{-1} $= 2.39 \times 10^{-3}$ cal cm^{-1} s^{-1} K^{-1}), and the temperature T is in degrees Kelvin. An extensive summary of data concerning κ has been gathered in Chermak (1981), where the experimental data is smoothed out by means of the formula

$$\kappa = \kappa_0/(1 + C\hat{T}), \quad \text{where } \hat{T} = T - 273. \tag{32}$$

Expressions (30), (31) and (32) yield practically coincidental results when $\kappa_0 = 2.5$ W m^{-1} K^{-1} and $C = -0.00025$ °C^{-1}. Chermak considers that, in the temperature range of

0–700 °C, for a crust of the Earth consisting of an upper "granitic" layer and a lower "basaltic" layer an excellent aproximation is obtained by setting $\kappa_0 = 3\,\mathrm{W\,m^{-1}\,K^{-1}}$ and $C = 8 \times 10^{-4}\,{}^\circ\mathrm{C}^{-1}$ in Equation (32). In the temperature interval 200–600 °C the thermal conduction of basalts is approximately constant at $\sim 2\,\mathrm{W\,m^{-1}\,K^{-1}}$. This was used to obtain averaged values of κ_0 ($\mathrm{W\,m^{-1}\,K^{-1}}$) and C ($10^{-40}\,\mathrm{C}^{-1}$) for granitoid, gabbroid, and ultrabasic rocks, respectively, equaling (κ_0 and C) 2.93 ± 0.51 and 11.90 ± 0.51; 2.10 ± 0.47 and 2.54 ± 0.50; 4.51 ± 0.70 and 14.8 ± 2.56. For conditions corresponding to depths of $l >$ 100–200 km the coefficient of thermal conduction has not been studied experimentally. In this connection, it is usually supposed to be constant and equal to $8 \times 10^{-3}\,\mathrm{cal\,cm^{-1}\,s^{-1}\,K^{-1}} = 3.33\,\mathrm{W\,m^{-1}\,K^{-1}}$, which is close to the value of $3.5\,\mathrm{W\,m^{-1}\,K^{-1}}$, produced by the formula of Schatz and Simmons [Equations (30) and (31)] for $T = 1473$ K. However we have no basis to suppose that the coefficient of thermal conduction is constant at $l >$ 100–200 km. In the absence of experimental data, it is reasonable to use the theoretical indications for this purpose. The theoretical expression for κ_L (cf. Zharkov, 1958; Zharkov and Trubitsyn, 1978) is of the form

$$\kappa_L = \kappa_{L,l_0}\left(\frac{T_{l_0}}{T_l}\right)F_l(\rho), \quad F_l(\rho) = \left(\frac{\rho_{l_0}}{\rho_l}\right)^{1/3}\left(\frac{\gamma_{l_0}}{\gamma_l}\right)^2\left(\frac{\theta_l}{\theta_{l_0}}\right)^3, \tag{33}$$

where l_0 is the reference depth, l is the actual depth; ρ, θ and γ are the density, the Debye temperature, and the Grüneisen parameter, respectively; T is the absolute temperature. According to the data of Schatz and Simmons, when $\hat{T} \sim 1200$ °C the components of κ_L and κ_R provide roughly the same contribution to κ. The question as to the values of κ_R in the depths of the mantle has not been decided and there is reason to suppose that κ_R does not appreciably increase there. Therefore, we shall assume that $\kappa_R = 4 \times 10^{-3}$ cal $\mathrm{cm^{-1}\,s^{-1}\,K}$ when $l \geqslant 200$ km and $\kappa_{L,200} = 4 \times 10^{-3}\,\mathrm{cal\,cm^{-1}\,s^{-1}\,K^{-1}}$. It is then reasonable to calculate the further variation with depth by means of the relation

$$\kappa(l) = 4 \times 10^{-3}\left[1 + \left(\frac{T_{200}}{T_l}\right)F_l(\rho)\right]\mathrm{cal\,cm^{-1}\,s^{-1}\,K^{-1}}, \tag{34}$$

where $F_l(p)$ is defined by formula (33) and given in Table V. At the base of the mantle, $T_{2340} \lesssim 3T_{200}$, $F_{2840} \sim 22$, and we obtain the estimate $\kappa(l = 2840) \sim 4\kappa(l = 200)$. Consequently, in the mantle of Venus, $\kappa(l)$ may increase by several times.

The temperature distribution in the outer 200 km layer of Venus (considering this the lithosphere of the planet) can be found by means of solving the steady-state thermal conduction equation for the flat layer

$$\frac{\mathrm{d}}{\mathrm{d}l}\left(\kappa\frac{\mathrm{d}T}{\mathrm{d}l}\right) + A(l) = 0, \tag{35}$$

where l is the depth, A is the evolution of heat in $\mathrm{W\,cm^{-3}}$. The distribution of radioactive impurities in the crust is assumed to be a homogeneous or exponential function of the depth

$$A(l) = A_0 e^{-l/D} \tag{36}$$

with a scale of $D \sim 8$–10 km. Let us consider a two-layered model of $\Delta l_1 = 70$ km and $\Delta l_2 = 130$ km. We shall consider the upper layer to be basalt, $\kappa_1 = 2 \times 10^{-2}\,\mathrm{W\,cm^{-1}\,K^{-1}}$; the lower consists of ultrabasic rocks, $\kappa_2 = 3.3 \times 10^{-2}\,\mathrm{W\,cm^{-1}\,K^{-1}}$; the temperature value at the surface is $T_0 = 733$ K. We shall consider the temperature value at the lower boundary of the lithosphere to be a free parameter, varying from 1473 to 1873 K, $T_{200} \sim$ 1473–1873 K.

According to modern geochemical theories (O'Nions *et al.*, 1979; Wasserburg and De Paolo, 1979; Zharkov, 1983), the upper and lower mantle of the Earth do not exchange material or radioactive impurities between themselves (the boundary of the upper and lower mantle is situated at the depth of the second phase transition, for Venus at ~ 760 km). We shall also presuppose a similar theory in respect of Venus. When the thick Cytherean crust melted out of the upper mantle of the planet, all the radioactive impurities have been virtually concentrated in the crust; and, therefore, we may presume that the concentration of sources of heat in the lower layer of the lithosphere is 0, $A_2 = 0$. Let us assume that the rate of generation of heat in the non-differentiated upper mantle of Venus is the same as for the non-differentiated upper mantle of Earth, $\sim 4.8 \times 10^{-6}\,\mu\mathrm{W\,kg^{-1}}$ (O'Nions *et al.*, 1979). Supposing that almost all the sources of heat have been transported from the upper mantle into the basalt crust, we obtain the value of A_0 in Equation (36), where we assumed that $D = 10$ km and $A_0 = 1.13 \times 10^{-6}\,\mu\mathrm{W\,cm^{-3}}$. The other limiting case corresponds to a uniform distribution of radioactivity in the crust. Then, $A_1(l) = \bar{A} = 1.62 \times 10^{-7}\,\mu\mathrm{W\,cm^{-3}}$. The solutions of the problem for $A_1(l)$ [Equation (36)] and $A_1(l) = \bar{A} = \text{const.}$ provide similar results. For simplicity, we shall give the formulas for a uniform distribution of radioactivity in the crust, $A_1(l) = \bar{A} = 1.62 \times 10^{-7}\,\mu\mathrm{W\,cm^{-3}}$. By l_1 and l_2 we shall designate the thickness of the crust and lithosphere, respectively; $T(l_1) = T_1$, $T(l_2) = T_2$, $T(0) = T_0$. Then, the heat flux from the interior of Venus q_0 is equal to

$$q_0 = \frac{\kappa_1 \dfrac{(T_2 - T_0)}{l_1} + \dfrac{\bar{A} l_1}{2}\left[1 + 2\dfrac{\kappa_1}{\kappa_2}\dfrac{(l_2 - l_1)}{l_1}\right]}{1 + \dfrac{\kappa_1}{\kappa_2}\dfrac{(l_2 - l_1)}{l_1}} \tag{37}$$

This is determined by the temperature value T_2 at the lower boundary of the lithosphere and by the thickness of the crust l_1. Under the assumed values of the parmeters $T_0 = 733$ K, $T_2 = 1473$–1873 K†, $l_1 = 70$ km, $l_2 = 200$ km, $\kappa_1 = 2 \times 10^{-2}\,\mathrm{W\,cm^{-1}\,K^{-1}}$, $\kappa_2 = 3.3 \times 10^{-2}\,\mathrm{W\,cm^{-1}\,K^{-1}}$, $\bar{A} = 1.62 \times 10^{-7}\,\mu\mathrm{W\,cm^{-3}}$, we obtain $q_0 = (18.5$–$23.8)$ $\mathrm{erg\,s^{-1}\,cm^{-2}}$. This is much less than that found in the work of (Toksöz *et al.*, 1978) $q_0 \sim 77\,\mathrm{erg\,s^{-1}\,cm^{-2}}$ (for Earth, $q_0 \sim 61.5\,\mathrm{erg\,s^{-1}\,cm^{-2}}$). If we reduce the thickness of

† The lower boundary of the lithosphere can be arbitrarily defined on the isothermic surface $T_2 = 1473$ K (1200 °C). The choice of $T_2 = 1873$ K (1600 °C) physically denotes that the lithosphere includes the upper thermal boundary layer of convective cells in the upper mantle.

the lithosphere by two times, $l_2 = 100\,\text{km}$, while retaining the values of the other parameters, then the heat flux from Venus is somewhat increased, $q_0 \sim (23.5\text{–}32.5)\,\text{erg s}^{-1}\,\text{cm}^{-2}$. Let us now consider what happens when the thickness of the crust is reduced by two times, $l_1 = 35\,\text{km}$ (corresponding to $\bar{A} = 3.24 \times 10^{-7}\,\mu\text{W cm}^{-3}$); when $l_2 = 200\,\text{km}$, we obtain $q_0 = (20.8\text{–}26.7)\,\text{erg s}^{-1}\,\text{cm}^{-2}$; when $l_2 = 100\,\text{km}$, the flux increases somewhat, $q_0 = (28.6\text{–}39.3)\,\text{erg s}^{-1}\,\text{cm}^{-2}$. As a general deduction from the above we may regard the heat flux from the interior of Venus as included in the interval

$$q_0 \sim (20\text{–}40)\,\text{erg s}^{-1}\,\text{cm}^{-2}, \tag{38}$$

i.e., 2–3 times less than that from the interior of the Earth. This is due to the high temperature of the surface of Venus and the low thermal conduction of its thick basalt crust.

In this particular 2-layer model of the lithosphere, the temperature in the crust is given by the expression

$$T(l) = T_0 + \frac{q_0}{\kappa_1} l - \frac{\bar{A}}{2\kappa_1} l^2, \quad l \leqslant l_1, \tag{39}$$

while in the lower lithosphere it is a linear function of the depth l.

3.4. THE ELECTRICAL CONDUCTION

Estimates of the coefficient of electrical conductivity of the metallic core of Venus are given at the end of the paragraph devoted to the thermodynamics of the core.

The question as to the electrical conductivity of the mantle of Venus can only be treated in a qualitative sense, while the estimates themselves only at best indicate the order of magnitude. With these reservations, the same as in the case of the mantle of Earth, an estimation of the electrical conductivity of the Venusian asthenosphere ($100\text{–}200 < l < 760\,\text{km}$) can be assumed (Zharkov and Trubitsyn, 1980) in the form

$$\sigma_a \sim 10^{-4}\text{–}10^{-3}\,\text{Ohm}^{-1}\,\text{cm}^{-1}, \tag{40}$$

while for the lower mantle,

$$\sigma(l \sim 10^3\,\text{km}) \sim 10^{-1}\,\text{Ohm}^{-1}\,\text{cm}^{-1}$$

and

$$\sigma(l \sim 2840\,\text{km})/\sigma(l \sim 1000\,\text{km}) \sim 10^2. \tag{41}$$

3.5. THE MECHANICAL QUALITY FACTOR Q_μ

Only hypotheses can be formed as to the mechanical quality factor Q_μ, as for the electrical conductivity of the mantle. Due to the high temperatures, the quality factor of the Cytherean crust should not be more than

$$Q_\mu \sim 300\text{–}100 \quad \text{when } l \lesssim 70\,\text{km}.$$

The quality factor of the asthenosphere should be even less, i.e.,

$$Q_\mu \sim 80\text{–}200 \quad \text{when } 70 \lesssim l \lesssim 760\,\text{km}.$$

In the lower mantle, the quality factor should be even higher, i.e.,

$$Q_\mu \sim 200\text{–}500, \qquad 800 < l < 2800;$$

although, exactly as in the case of Earth, we may expect a zone of low Q_μ at the bottom of the mantle and, probably, at the boundary between the lower and upper mantle. If Venus has a molten core, its quality factor should be large, $Q \gg 10^3$.

For a solid interior core (if it exists!), the quality factor should be much less, $Q_\mu \sim (100\text{–}500)$, than that for a molten core.

3.6. LARGE-SCALE STEADY STRESSES IN THE MANTLE OF VENUS

Due to the disequilibrium of the planets of the Earth group, i.e., the fact that the planetary figure deviates from the equilibrium figure, large-scale steady tangential stresses are created in their interiors, which are maintained over the course of geologic intervals of time $(10^8\text{–}10^9)$ yr by rigid zones of the crust and mantle. The deviation of the planet from the hydrostatic-equilibrium state can be explained by data on the gravitational field of the planet (Zharkov and Trubitsyn, 1978). The non-hydrostatic nature of the planet is characterized by non-hydrostatic values of the coefficients in the expansion of the external gravitational potential. In the case of Earth and Mars, the maximum stresses are occasioned by the non-hydrostatic nature of the quadrupole moment $\Delta J_2 = (J_2 - J_2^0)$ (cf. Table I). The other coefficients for these planets are appreciably less than ΔJ_2. In the case of Venus, the quantity $\Delta J_2 \sim J_2$ is of the order of J_3 and appreciably larger than the other coefficients† (Ananda *et al.*, 1980). Considering this and the fact that this entire discussion is by its nature qualitative, we shall estimate the scale of the tangential stresses in the interior of Venus, regarding them as mainly determined by the quantity ΔJ_2.

We shall now explain the method which enables the estimation of the steady tangential stresses in the planets of the Earth group by virtue of $\Delta J_2 \neq 0$. Let us consider a homogeneous model of the planet with a surface represented by a sphere of radius equalling the mean planetary radius R. On the surface of our model we shall situate a thin layer with mean density equalling that of the planet, while the thickness of this layer and its distribution over the surface of the planet shall be chosen to produce the non-equilibrium portion of the quadrupole field of the planet, which is equal to

$$\Delta V = -\frac{GM}{r}\left(\frac{R}{r}\right)^2 \Delta J_2 P_2(\cos\theta), \tag{42}$$

where we have used the standard designations and θ is the polar angle, which equals the complement of the latitude. Then it is easy to determine the distribution of the amplitude of the sought layer $\epsilon_2(\theta)$ over the surface of the planet

$$\epsilon_2(\theta) = \epsilon_{20} P_2(\cos\theta), \qquad \epsilon_{20} = -\tfrac{5}{3} R \Delta J_2, \tag{43}$$

† The coefficient J_3 has only just been determined with a very large error and naturally its value may be considerably lowered by a future refinement.

where P_2 is the second Legendre polynomial. The presence of the weight layer at the surface of the planet leads to non-hydrostatic stresses in its interior. It is clear that the stresses are proportional to the linear amplitude of the layer $|\epsilon_{20}|$, its density ρ_0, and gravity g_0, i.e., the stress is proportional to the weight of the layer, adjusted to the unit of area. The maximum tangential stresses for a homogeneous planet model are obtained in the center. In due time, these have been found by Jeffreys to be equal to

$$(\tau_2)_{\max} = \tfrac{1}{2} g_0 \rho_0 R (J_2 - J_2^0) \quad \text{when } r = 0. \tag{44}$$

Using the data in Table I, we find it easy to calculate $(\tau_2)_{\max}$ for all the planets of the Earth group:

	Mercury	Venus	Earth	Mars
$(\tau_2)_{\max}$, bar	19.6	5.6[a] 8.4[b]	17.2	32.4

[a] Data of Akim *et al.*, (1978).
[b] Data of Ananda *et al.*, (1980).

The homogeneous, continuously elastic model of the planet simplifies the actual situation too much. The actual planets have molten cores or sufficiently-warmed solid cores which cannot withstand non-hydrostatic loads for a prolonged time of $(10^8\text{–}10^9)$ yr. Therefore, a two-layer model consisting of a molten core with a mean density of ρ^i and a radius r_c and of an elastic silicate shell with a mean density of ρ, capable of withstanding non-hydrostatic tangential stresses over the course of long intervals of time, is more realistic. The core may consist of either the actual core or the core and a portion of the lower mantle which, by virtue of the high temperatures, cannot withstand non-hydrostatic loads over the course of geological intervals and thus behaves as a liquid over large intervals of time. The presence of a molten core results in the fact that the stresses will be displaced from it into the elastic shell where, depending on the radius of the effective molten core, they may considerably increase. Thus, we must solve a problem in the theory of elasticity concerning the stresses in a two-layer planet model, resulting from a ponderable layer arranged on its surface and providing the non-hydrostatic component of the quadrupole gravitational moment. The problem mentioned above has been solved for a two-layer model of Venus with average parameters of the core and mantle (Zharkov and Zasurskiy, 1981). The results of a calculation of the three principal tangential stresses τ_1, τ_2 and τ_3 are graphically shown in Figs. 7–9 for three values of the polar angle $\theta = \pi/2$, 0, and $\pi/4$, i.e., for the equatorial ($\theta = \pi/2$), polar ($\theta = 0$), and bisectorial ($\theta = \pi/4$) planes of the planet for ΔJ_2, according to the data of Akim *et al.* (1978). (For ΔJ_2, according to the data of Ananda *et al.* (1980), the values of the stresses will be $1\frac{1}{2}$ times greater). The largest tangential stresses, $\tau_{\max}$, are achieved on the equatorial plane of the planet (Figure 7) at the boundary between the mantle and core ($r_c = 3210$ km, while

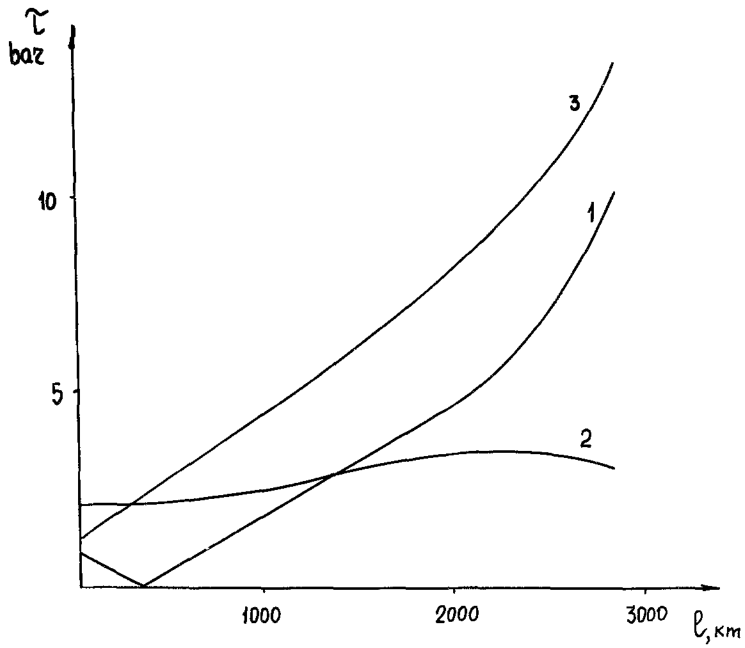

Fig. 7. Principal tangential stresses in the equatorial ($\theta = \pi/2$), polar ($\theta = 0$) (see Figure 8), and bisectorial ($\theta = \pi/4$) (see Figure 9) planes of the planet. For ΔJ_2, according to the data of (Akim *et al.*, 1978). (1) τ_1, (2) τ_2, (3) τ_3, when $\theta = 0$, $\tau_1 = \tau_2$, $\tau_3 = 0$.

$$\tau_{\max} \approx 1.45 \rho g_0 R \Delta J_2 \approx 13.5\,\text{bar}, \tag{45}$$

with an error of ± 40%, due to the uncertainty of J_2 (cf. Table I). Consequently, the presence of a molten core entails an increase in the stresses in the lower mantle of Venus by roughly 2.4 times over the stresses given above for the homogeneous model. As can be seen from Figures 7–9, the stresses in the silicate shell of Venus drop off rapidly from values of ~ 10 bar near the core to values of ~ 1.5 bar at a depth of ~ 100 km. The most important conclusion to be considered from the calculations should be the low level of stresses in the interior of Venus. This is yet another indication of the hot interior of the planet. The mean level of stresses in the lower mantle of Venus apparently lies in the interval of (3–10) bar. In the upper mantle of the planet ($l \lesssim 760$ km), excluding its lithosphere ($l \lesssim 200$–100 km), the viscosity of the planet is substantially lower (cf. next paragraph) than its viscosity in the lower mantle ($l > 760$ km), as a result of which the stresses should be displaced from the upper mantle into the lithosphere and lower mantle. The level of stresses in the asthenosphere of Venus (100–$200 \lesssim l \lesssim 760$ km) should be about 1 bar or less.

As is well known, in the Earth the thickness of the seismically-active surface layer is roughly 15 km. The geothermic gradient at the surface of Earth is ~ (20–30)° km^{-1}. Consequently, at the lower boundary of the seismically-active layer of Earth, the temperature is (300–450) °C. The mean temperature of the surface of Venus is 460 °C (cf. Table I). On

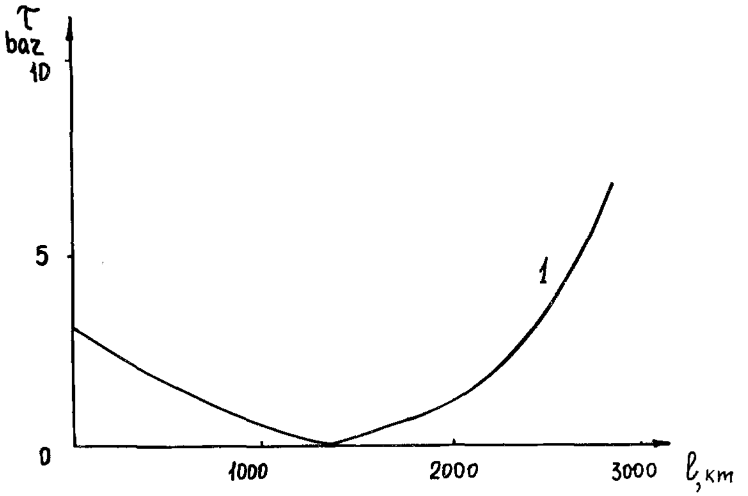

Fig. 8. See caption to Figure 7.

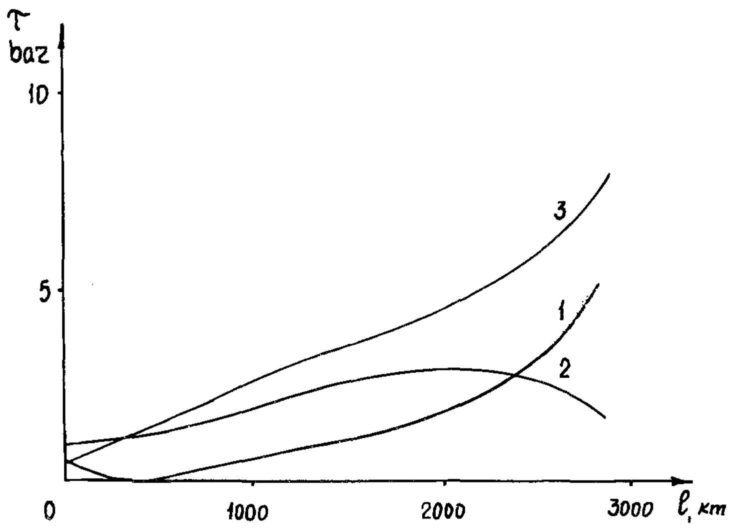

Fig. 9. See caption to Figure 7.

this basis, we may presume that Venus does not possess an outer seismically-active layer. In connection with this and the low level of tangential stresses in the interior of Venus, we may conclude that the planet is aseismic.

Due to the currents or convection in the mantle of Venus, the gradients of the hydrodynamic velocities are not equal to zero, which leads to dynamic viscous tangential stresses. We shall not consider these, due to the incomplete statement of the problem.

3.7. A RHEOLOGICAL MODEL OF THE CRUST AND MANTLE

A rheological model is constructed in the following manner (Murrell, 1976; Zharkov and Trubitsyn, 1980; Zharkov, 1982). The irreversible deformation due to the creep ϵ is represented in the form of three terms as

$$\epsilon = \alpha \lg t + \beta t^m + \gamma t, \tag{46}$$

where α, β, and γ are functions of the tangential stress τ, the temperature T, and the pressure p, and t is the time. The exponent m lies within the interval (1/3–1/2). The first two terms in Equation (46) describe the non-steady creep. Under small deformations $\epsilon \lesssim 0.01$ and temperatures of $T \lesssim 0.02T_m$, the first term is dominant, while for the conditions in the interior of Venus it is unimportant and will not be considered below; the third term describes the stage of steady creep, which appears under sufficiently high temperatures and deformations of $\epsilon > 0.1$. The non-steady creep is related to the deformational strengthening which is due to the presence of dislocations in crystals. The dependence of β and γ on τ and T has been established experimentally to be of the form

$$\beta = \beta_0 \left(\frac{\tau}{\mu}\right)^p e^{-H_1/kT} \simeq 10^{-4}\text{–}10^{-5} \quad \text{at } \tau \lesssim 1 \text{ kbar}, \tag{47}$$

$$\gamma = \gamma_0 \left(\frac{\tau}{\mu}\right)^n e^{-H_2/kT}, \tag{48}$$

where $n \simeq 2$–3 when $100 \lesssim \tau \lesssim 1$ kbar and $n > 3$ under the large stresses for oxides and silicates. In Equations (47) and (48), μ is the shear modulus, k is the Boltzmann constant, H_1 and H_2 are the enthalpies of activation, while the constant $p \simeq 1$ when $\tau < 1$ kbar and takes on larger values for larger stresses.

We shall first consider the non-steady creep and then move on to the steady creep. An important parameter of the medium is the time of non-steady creep t_{ss}, which is determined from the condition that the rate of non-steady creep $\mathring{\epsilon}_t$ has been reduced to the value of the rate of steady creep $\mathring{\epsilon}_{ss}$. Further, it is convenient to exploit the law than the total deformation $\epsilon_t = \beta t_{ss}^m \simeq 0.1$ under non-steady creep does not depend on the time of deformation, the temperature, or the stress. This leads to formulae which relate t_{ss} and the effective viscosity η_{ss}:

$$\eta_{ss} = (\tfrac{1}{3})^{(1+n)/2n} \left(\frac{\tau}{\mathring{\epsilon}_{ss}}\right) = (\tfrac{1}{3})^{(m+p)/2p} \left(\frac{\tau t_{ss}}{m\epsilon_t}\right), \tag{49}$$

or

$$t_{ss} \approx 2.5 \times 10^{-15} (\eta_{ss}/\tau), \tag{50}$$

where t_{ss} is in years when η_{ss} is expressed in poises and τ is in bars. The rheological model of the lithosphere is characterized by the time t_{ss} and the effective viscosity η_{ss}. The value of the parameter t_{ss} is determined by the fact that, if the duration of the tectonic process is less than t_{ss}, then there only takes place a non-steady creep and the integrated deformation is less than 0.1. The dependence of the creep on the pressure is included in the activation enthalpy H_1 and H_2.

Three approaches are used in this problem (Zharkov, 1983). One of the methods starts from the assumption that the activation enthalpy for self-diffusion H is proportional to the melting temperature – i.e., $H(p) \sim \xi T_m(p)$ – while the quantity $\xi \sim 20$–40 is very large. In geophysics, the melting temperature of the mantle and that of the core are estimated by means of the Lindemann formula (15) and we get

$$H(p) = H_{100}(\rho_{100}/\rho(l))^{2/3}[\theta(l)/\theta_{100}]^2, \tag{51}$$

where $l = 100$ km has been taken as the reference depth.

In the second method, the standard thermodynamic formula is used, of the form

$$H(p) = E^* + V^* p, \tag{52}$$

where E^* is the energy of activation (when $p = 0$) and V^* is the activation volume, both assumed to be constant. In fact, E^* and V^* are unknown functions of the density, and therefore, strictly speaking, Equation (52) can be used only for small pressures, $p \ll K_T$, where K_T is the isothermal modulus of compression.

In the third method, the equations

$$H(p) = E^*(\rho/\rho_0)^L, \qquad L = \frac{\partial \ln H}{\partial \ln \rho} \equiv \frac{V^* K_T}{E^*} \tag{53}$$

are used. The coefficient L is similar in meaning to the Gruneisen parameter – it is a weak diminishing function of the density $L = L(\rho)$ and can only be considered constant to a first approximation (Zharkov and Kalinin, 1971).

At low pressures, formula (53) is equivalent to Equation (52), while at high pressures it properly accounts for the effective dependence of L^* and V^* on the density.

Moreover, being an explicit dependence on the density, it enables a qualitative and proper estimation of the change in the activation enthalpy due to the density jump under the phase transitions in the mantle of Venus.

The question as to the effective viscosity, which characterizes the high-temperature steady creep, is complicated. The effective viscosity depends not only on the above-enumerated parameters, but also the partial oxygen pressure in the silicates, which is unknown. Therefore the treatment of this question has been simplified.

In silicate mantles, the fundamental role belongs to the two mechanisms of creep – diffusional and dislocational – and correspondingly two coefficients of viscosity η_1 and η_2 are introduced, from which a general expression is constructed for the effective viscosity of the mantle:

$$\eta_{\text{eff}} = \frac{\eta_1 \cdot \eta_2}{\eta_1 + \eta_2}. \tag{54}$$

Under stresses of $\tau > 1$ bar, $\eta_2 < \eta_1$ prevails (Zharkov, 1983). Employing the experimental data for peridotite (cf. Schubert, 1979; Murrell, 1976; Zharkov, 1983), and assuming a peridotite mantle, we can write for $\eta \equiv \eta_{\text{eff}} = \eta_2$ the simplified formula

$$\eta = 1.1 \times 10^3 \left(\frac{T}{1460\,\text{K}}\right)\left(\frac{1\,\text{bar}}{\tau}\right)^2 \left(\frac{\mu}{6.4 \times 10^5\,\text{bar}}\right)^3 e^{\frac{68960}{T} \cdot \left(\frac{\rho}{3.365}\right)^{2.5}} \text{ poise},$$

$$T > 1373\,\text{K}, \tag{55}$$

where T is the absolute temperature, τ, the value of the tangential stresses in bars; and μ is the modulus of shear in bars. The reference level was chosen to be the depth $l = 100$ km, where $p = 32$ kbar and the enthalpy at this depth $H^* = 126\,\text{kcal mole}^{-1} + pV^* = 136.2\,\text{kcal mole}^{-1}$, where, in conformity with Ross *et al.* (1978) and Schubert (1979), it was assumed that $V^* = 13.4\,\text{cm}^3\,\text{mole}^{-1}$, whereas L from Equation (53) is equal to 2.5. Equation (55) allows us to evaluate the distribution of the effective viscosity in the mantle of Venus, after assigning the temperature distribution and selecting a specific value of τ. Performing numerical experiments for the temperatures shown in Figure 5 (curves 3 and 5), and considering that in zones of high viscosity in the lower mantle stresses of 10–20 bar may be concentrated, while in the upper mantle the stresses are small (of the order of several bars). Taking into account the reduction in L in Equation (53) (if we compare H from Equation (53) with H resulting from Equation (51) then over the extent of the mantle $L = L(\rho)$ should diminish from 2.5 to 2.1), a rough estimate can be made for the viscosity of the lower mantle, amounting to

$$\eta \sim (10^{23}\text{–}10^{25})\text{ poise}, \quad \begin{matrix} \tau \sim (10\text{–}20)\,\text{bar}, \\ l > 750\,\text{km}, \end{matrix} \tag{56}$$

and, correspondingly, the viscosity of the upper mantle

$$\eta \sim (10^{20}\text{–}10^{21})\text{ poise}, \quad \tau \sim 1\text{ bar}, \quad (100\text{–}200)\,\text{km} < l < 750\,\text{km}. \tag{57}$$

In what follows we shall discuss the matter of the boundary between the upper mantle and the lithosphere, and the thickness of the Cytherean lithosphere. Using Equations (56), (57) and (50), we obtain estimates for the time of non-steady creep in the lower and upper mantle, given by

$$t_{ss} \sim \begin{cases} 2.5 \times (10^7\text{–}10^9)\,\text{yr} & l > 750\,\text{km}, \\ 2.5 \times (10^5\text{–}10^6)\,\text{yr} & (100\text{–}200)\,\text{km} < l < 750\,\text{km}. \end{cases} \tag{58}$$

A rheological model of the lithosphere can also be constructed after Murrell (1976) or Zharkov (1983). For 70 km of basalt crust, the rheological parameters for dolerite are assumed, while the mantle is modeled with a peridotite composition. Then, after assigning the level of stresses in the lithosphere and the temperature distribution, it is easy to calculate the time of non-steady creep t_{ss} and the effective viscosity, as functions of the depth.

4. Convection

A survey on convection in the mantles of the planets of the Earth group has been written by Schubert (1979). Schubert believes that convection transpires across the entire mantle in the planets of the Earth group. The present author maintains a different theory, in accordance with which convection takes place independently in the upper and lower mantles of the Earth (and Venus) and both convective systems interact on their boundary (Zharkov, 1983). Similar views are shared by MacKenzie, Richter, and Turcott.

We shall suppose that in Venus – as in the Earth – gravitational differentiation has long been accomplished, and does not contribute to the formation of the heat flux of the planet. In accordance with O'Nions *et al.* (1979) and Wasserburg and De Paolo (1979), an exchange of material between the upper and lower mantle of Earth can be disregarded, but the convection takes place independently in the upper and lower mantles (Zharkov, 1983). We shall suppose the same for Venus. The main sources of heat are the radioactive impurities in the crust and mantle, the original heat, and the heat flux from the core to the mantle, which has been estimated in the section on 'thermodynamics of the core', $Q_c = 2.86 \times 10^{19}\,\mathrm{erg\,s^{-1}} = 0.9 \times 10^{27}\,\mathrm{erg\,yr^{-1}}$.

We shall assume that the mean rate of radiogenic evolution of heat in the non-differentiated silicate mantle of Venus is the same as in the non-differentiated mantle of the Earth, $4.8 \times 10^{-6}\,\mu\mathrm{W\,kg^{-1}}$ (O'Nions *et al.*, 1979) (in fact, Venus may be depleted in potassium with respect to the Earth). The mass of the mantle of Venus, 3.4×10^{27} g, is roughly divided between the upper mantle + crust and the lower mantle, as 1.08×10^{27} and 2.33×10^{27} g, respectively. Since the lower and upper mantle do not exchange material, the radiogenic evolution of heat in the lower mantle coincides with the original value of $4.8 \times 10^{-6}\,\mu\mathrm{W\,kg^{-1}}$, while the upper mantle is depleted in radioactivity, due to the concentration of the latter in the crust (cf. Section 3.3). We shall ignore the radioactivity of the upper mantle of Venus and consider that the convection there is caused by the heat flux supplied from the lower mantle at the bottom. The radiogenic evolution of heat inside Venus as a whole, and individually in the upper mantle + crust and in the lower mantle is shown in Figure 10.

Let us consider the convection in the upper mantle of Venus, produced by a heat flux supplied from below F, the size of which is generally unknown. According to the estimations in Section 3.3, it is difficult to conduct a large heat flux across the lithosphere. In the estimates according to Equation (38), for the lower limit we shall assume $F_1 \sim 20\,\mathrm{erg\,s^{-1}\,cm^{-2}}$, and for the upper limit, $F_h \sim 40\,\mathrm{erg\,s^{-1}\,cm^{-2}}$. The complete heat flux will be obtained by adding to F the evolution of heat in the crust, equalling $\sim 10\,\mathrm{erg\,s^{-1}\,cm^{-2}}$, according to Figure 10. We shall model the upper mantle by a flat layer with a thickness $d_1 = 650$ km and constant physical parameters, i.e., with

$$\begin{aligned} &g \sim 10^3\,\mathrm{cm\,s^{-2}}, \quad \alpha \approx 3 \times 10^{-5}\,\mathrm{K^{-1}}, \quad C_p \approx 1.2 \times 10^7\,\mathrm{erg\,g^{-1}\,K^{-1}}, \\ &\rho_1 \approx 3.7\,\mathrm{g\,cm^{-3}}, \quad \chi = 10^{-2}\,\mathrm{cm^2\,s^{-1}}, \quad \nu \approx 3 \times 10^{20}\,\mathrm{cm^2\,s^{-1}}, \\ &F \approx (20\text{–}40)\,\mathrm{erg\,s^{-1}\,cm^{-2}}, \end{aligned} \tag{59}$$

where the standard designations χ and ν have been used – the coefficients of the thermal diffusivity and of the kinematic viscosity, respectively. Then, the flux Rayleigh number is given by

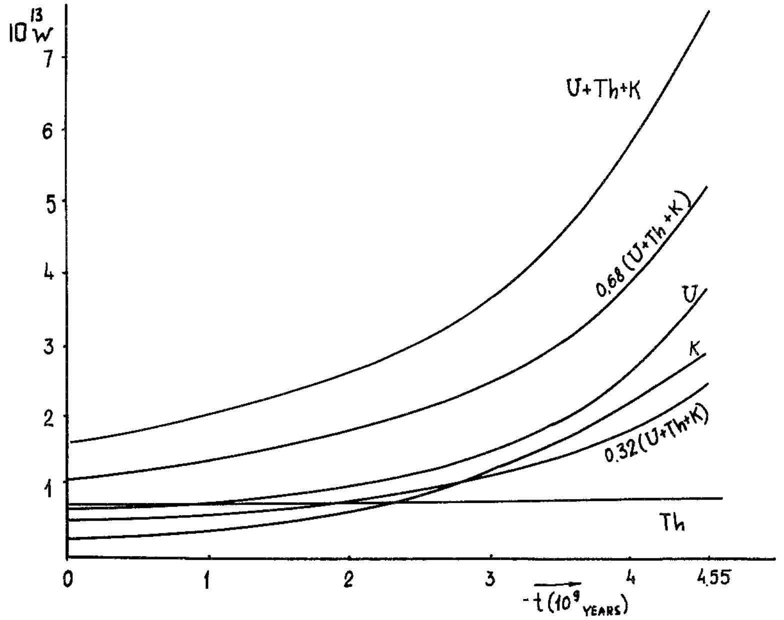

Fig. 10. Total radiogenic evolution of heat in the entirety of Venus. Shown separately are the evolution of heat from the decay of K, Th, U, and the sum of U + Th + K, the evolution of heat in the lower mantle 0.68 (U + Th + K) and in the upper mantle + crust 0.32 (U + Th + K).

$$R_F = \frac{\alpha g F d_1^4}{\rho C_p \chi^2 \nu} \tag{60}$$

which is equal to

$$R_F(1\text{–}2) \times 8 \times 10^6. \tag{61}$$

The critical value is $R_{Fc} \sim 10^3$, and thus estimate (61) indicates that the convection in the upper mantle of Venus is strongly developed and, at the boundaries of this mantle there should be formed thermal boundary layers with a thickness of δ_1, in each of which the temperature drop $\Delta T_1/2$ takes place, where ΔT_1 is the super-adiabatic temperature difference in the upper mantle. The Prandtl number, $Pr = \nu/\chi \sim 3 \times 10^{22}$, is practically equal to infinity and therefore all the convection parameters, averaged over the horizontal, can be expressed by the number R_F (Zharkov, 1983) as

$$Nu = \frac{d_1}{2\delta_1}, \quad \delta_1 \approx \frac{d_1}{2a_1^{3/4}} R_F^{-1/4}, \quad \Delta T_1 \approx a_1^{-3/4} \left(\frac{F}{\rho C_p \chi}\right)^{3/4} \left(\frac{\nu\chi}{\alpha g}\right)^{1/4}, \tag{62}$$

$$\bar{u}_1 \sim a_2 R_F^{1/2} \chi / d_1, \quad a_1 \sim 0.2, \quad a_2 \sim 0.1,$$

where $\bar{u}_1$ is the characteristic speed – the speed of the boundary layer or the speed of the hot ascending or cold descending fluxes. The Nusselt number Nu characterizes the effectiveness of the convective export of heat. Substituting in Equation (62) the values of the parameters (59) and (60), we find that

$$Nu \sim 16, \quad \delta_1 \sim (20.5\text{–}17.2)\,\text{km}, \quad \Delta T_1 \sim (184\text{–}309)\,\text{K},$$
$$\bar{u}_1 \sim (1.4\text{–}2)\,\text{cm yr}^{-1}. \tag{63}$$

The convective cells in the upper mantle of Venus are approximately isometric, and their horizontal dimension is of the order of the depth of the layer d_1. The characteristic time for the establishment of a steady convection is $\tau_{K1} \sim d_1^2/\pi^2\chi \sim 1.3 \times 10^9$ yr, which is not generally large; however, it is substantially larger than the time of export of heat from the upper mantle, which is $\tau_1 \sim d_1/\bar{u}_1 \sim 4 \times 10^7$ yr. The time of thermal relaxation of the lithosphere is also calculated as τ_{K1}: $\tau_1 \sim l_2^2/\pi^2\chi \sim 3 \times 10^7$ yr, where we have set $l_2 = 100$ km. Consequently, the time of thermal relaxation of the crust and upper mantle of Venus $\tau_{K,BM} \lesssim 10^8$ yr is small, as the intensity of radiogenic evolution of heat is virtually unchanged in such an interval of time.

With a flux of $F \sim (20\text{–}40)$ erg s^{-1} cm^{-2}, supplied from the lower mantle, weak thermal boundary layers are formed on the boundaries of the upper mantle, of a thickness of about 20 km and temperature drops of $\Delta T_{1/2} \sim (100\text{–}150)$ K. Finally, the convection in the upper mantle increases its effective thermal conduction by 16 times, and $Nu \sim 16$.

The independence of convection in the lower mantle from the convection in the upper mantle was apparently occasioned during its formation by a jump in the viscosity of two orders of magnitude at the depth of the second phase transition ($l \sim 750$ km). As a result of the separation of the crust from the upper mantle, this boundary is also a weak chemical boundary. For the lower mantle, we shall assume the values of the parameters

$$\begin{aligned} &g \sim 10^3\,\text{cm s}^{-2}, \quad \alpha \sim 1.5 \times 10^{-5}\,\text{K}^{-1}, \quad \rho \sim 4.9\,\text{g cm}^{-3}, \\ &q_2 \sim 2.35 \times 10^{-7}\,\text{erg cm}^{-3}\,\text{s}^{-1}, \quad d_2 \sim 2.08 \times 10^8\,\text{cm} \\ &C_p \sim 1.2 \times 10^7\,\text{erg g}^{-1}\,\text{K}^{-1}, \quad \nu \sim (1\text{–}10) \times 10^{23}\,\text{cm}^2\,\text{s}^{-1}, \\ &\chi \sim 3 \times 10^{-2}\,\text{cm}^2\,\text{s}^{-1}, \end{aligned} \tag{64}$$

where q_2 is the heat production in the lower mantle.

The heat flux from the core into the mantle is $Q_c = 2.86 \times 10^{13}$ erg s^{-1} cm^{-2} or $F_{c\text{-}m} = 22$ erg s^{-1} cm^{-2}. The production of heat in the lower mantle due to radioactivity, $Q_{HM} \sim q_2 V_{HM} \sim 11.3 \times 10^{19}$ erg s^{-1}, is four times as large as the heat conducted in the mantle from the core. If we further consider that the thermal inertia of the lower mantle is large, $\tau_{HM} \sim 10^9$ y, then – in accordance with Figure 10 – the effective production of heat here may be even larger. Thus the main contribution of heat to the lower mantle comes from internal heating.

The heat flux from the core into the mantle creates a thermal boundary layer at the bottom of the mantle (Zharkov, 1983). Let us assume that this layer is located at the limt of convective stability, when the critical value of the flux Rayleigh number is attained $R_{Fc} \sim 10^3$.

This permits us to estimate the thickness of the layer δ_3 and the temperature increment in it ΔT_3: namely,

$$\delta_3 \sim R_{Fc}^{1/4}\left(\frac{\rho C_p \chi^2 \nu}{\alpha g}\right)^{1/4} F_{c-m}^{-1/4} \sim 630\,\text{km},$$
$$\Delta T_3 \sim \frac{F_{c-m}\delta_3}{\rho C_p \chi} \sim 790\,\text{K}, \tag{65}$$

These estimates should obviously be regarded as upper limits of δ_3 and ΔT_3. Reducing these by a factor 2, we shall consider that $\delta_3 \sim (300\text{–}630)$ km and $\Delta T_3 \sim (400\text{–}790)$ K.

In the case of convection with internal heating at large Rayleigh numbers, a thin boundary layer forms near the upper boundary of the convective cell, which is converted into a narrow descending flux, adjacent to the right vertical boundary of the cell (in the case of a cell with clockwise motion). In this case, the horizontal speed of the boundary layer u_x, the vertical speed of the descending flux u_z, $u_z \sim u_x$, the approximately equal widths of the layers δ_2, and the Nusselt number Nu are expressed in terms of the non-dimensional number R_q (Zharkov, 1983)

$$R_q = \frac{\alpha g q_2 d_2^5}{\rho C_p \chi^2 \nu} \sim 2.6 \times 10^5,$$
$$u_x \approx 0.8 R_q^{2/5}\chi/d_2, \quad u_z \simeq 1.6 R_q^{2/5}\chi/d_2, \quad \delta_2 = 1.5 R_q^{-1/5} d_2, \tag{66}$$
$$Nu \approx 0.8 R_q^{1/5}.$$

Consequently,

$$u_x \sim 0.53\,\text{cm y}^{-1}, \quad u_z \sim 1\,\text{cm y}^{-1}, \quad \delta_2 \sim 260\,\text{km}, \quad Nu \sim 9.6. \tag{67}$$

Using the Nusselt number Nu, we shall determine the temperature increment ΔT_2 in the boundary layer at the upper boundary as

$$\Delta T_2 \sim q_2 d_2^2/\rho C_p \chi Nu \sim 600\,\text{K}. \tag{68}$$

In the convective cells of the lower mantle, there are two characteristic speeds. One is $u \sim u_x \sim u_z$, the speed of the boundary layer and descending flux; the second is U, the scale of the speed in the remainder of the cell.

The vertical component of the latter speed U_z can be determined by means of the law of conservation of mass $U_z d_2 \sim u_z \delta_2$; the quantity of matter flowing into the boundary layer is equal to the drain in the descending flux – i.e. to

$$U_z \sim u_z \frac{\delta_2}{d_2} \sim 1 \times \frac{260}{2080} \sim 0.125\,\text{cm y}^{-1}. \tag{69}$$

The characteristic time is $\tau_{HM} \sim d_2/U_z \sim 1.7 \times 10^9$ y, the time of transport of material from the bottom of the lower mantle to its boundary layer at depths of 800–1000 km.

This is the time of thermal inertia of the lower mantle. The physical meaning of τ_{HM} consists in the fact that the heat exported to the boundary layer during the modern epoch was formed at the bottom of the mantle τ_{HM} years ago, when the rate of radiogenic evolution of heat was significantly larger. According to Figure 10, the evolution of heat in the lower mantle of about (1.5–1.7) × 10^9 yr ago was ~ 1.6 × 10^{20} erg s^{-1}. Adding to this the flux from the core into the mantle, we find 1.9 × 10^{20} erg s^{-1}; dividing this by the area of the surface of Venus, we find $F \sim 40$ erg s^{-1} cm^{-2}.

Let us again consider the question of the thickness of the lithosphere of Venus, choosing a temperature at its base which is capable of sustaining a flux of $F \sim 40$ erg s^{-1} cm^{-2}.

Let the thickness of the lithosphere coincide with the thickness of the crust $l = l_1 =$ 70 km. Then, using Equation (37), it is easy to find that the temperature at the base of the crust is 2060 °C! The absurdity of this result indicates that a planet with a relief and a level of stresses in its interior of ~ 10 bar cannot possess a temperature of ~ 2060 °C at a depth of 70 km. The solution of this dilemma is the fact that only a portion of the heat evolved in the lower mantle is transmitted to the upper mantle, while another portion is used to heat up the lower mantle itself. Let half the heat of the lower mantle be transferred to the upper, the other half being used for its heating. When $F = 20$ erg s^{-1} cm^{-2}, we obtain a temperature of 1360 °C at the base of the crust, which is rather more reasonable. The other portion of the flux evolves 1.16 × 10^{37} erg in 4 × 10^9 yr. Dividing this by the heat capacity of the lower mantle, 2.8 × 10^{34} erg K^{-1}, we determine that it should be heated to 420 K, which is fully acceptable.

Let us briefly formulate the conclusions from the above estimates. The heat flux from the interior of the planet can hardly exceed the value of about 30 erg s^{-1} cm^{-2}. In the convective, adiabatic mantle, the super-adiabatic temperature increment is formed in the thermal boundary layers in the upper mantle $\Delta T_1 \lesssim 200$ K from Equation (63); in the lower mantle $\Delta T_2 \lesssim 600$ K from Equation (68) and $\Delta T_3 \lesssim$ (400–800) K from Equation (65). The adiabatic temperatures at the mantle/core boundary are ~ 2900 K (cf. Figure 5). Consequently, it is reasonable to consider that the actual temperatures at the mantle/core boundary lie within limits of ~ (3500–4500) K. The thermal boundary layers of the mantle are zones of reheating and, consequently, should be zones of reduced quality Q_μ.

References

Ahrens, T. J.: 1979, *J. Geophys. Res.* **84**, 985.

Akim, E. L., Vlasova, Z. P., and Chuico, I. V.: 1978, *Dokl. Adad. Nauk SSSR* **240** (3), 556.

Akimoto, S., Matsui, Y., and Syono, Y.: 1976, 'High-Pressure Crystal-Chemistry of Orthosilicates and the Formation of the Mantle Transition Zone', in R. G. J. Strens (ed.), *The Physics and Chemistry of Minerals and Rocks*, John Wiley, London, pp. 327–363.

Ananda, M. P., Sjogren, W. L., Phillips, R. J., Wimberly, R. N., and Bills, B. G.: 1980, *J. Geophys. Res.* **85**, 8303.

Anderson, D. L.: 1979, *Geophys. Res. Lett.* **6**, 433.

Anderson, D. L.: 1980, *Geophys. Res. Lett.* **7**, 101.

Brown, J. M. and McQueen, R. G.: 1981, 'The Equation of State for Iron and the Earth's Core', Preprint.

Bullen, K. E.: 1936, *Monthly Notices Roy. Astron. Soc., Geophys. Suppl.* **3**, 395.

Bullen, K. E.: 1975, *The Earth's Density*, Chapman and Hall, London.

Cermak, V.: 1981, 'The Geothermal Model of Lithosphere and the Regional Distribution of the Lithosphere Thickness for the Territory of U.S.S.R.'. Preprint. Submitted to *Irv. Acad. Nauk SSSR-Fiz. Zem.*

Cook, A. H.: 1980, *Interiors of the Planets*, Cambridge Univ. Press, Cambridge.
Dziewonski, A. M., Hales, A. L., and Lapwood, E. R.: 1975, *Phys. Earth Planet. Inter.* **10**, 12.
Dziewonski, A. m. and Anderson, D. L.: 1980, 'Preliminary Reference Earth Model', Preprint. Submitted to *Phys. Earth Planet. Inter.*
Grossman, L. and Larimer, J. W.: 1974, *Rev. Geophys. Space Phys.* **12**, 71.
Jacobs, J. A.: 1975, *The Earth's Core*, Academic Press, London.
Jeffreys, H.: 1937, *Monthly Notices Roy. Astron. Soc., Geophys. Suppl.* **4**, 62.
Kalinin, V. A. and Sergeyeva, N. A.: 1979, *Izv. Academy of Sciences – Fizika Zemli* **10**, 3.
Kovach, R. and Anderson, D. L.: 1965, *J. Geophys. Res.* **70**, 2873.
Kozlovskaya, S. V.: 1982, *Astronomical Vestnik.* **16**, N 1.
Levin, B. J.: 1970, 'Internal Constitution of Terrestrial Planets', in A. Dollfus (ed.), *Surfaces and Interiors of Planets and Satellites*, Academic Press, London.
Lewis, J. S.: 1972, *Earth Planet. Sci. Lett.* **15**, 286.
Liu, L.: 1977, 'The System Enstatite-Pyrope at high Pressures and Temperatures and the Mineralogy of the Earth's Mantle', Preprint.
Maeva, S. V.: 1969, *Astrophys. Lett.* **4**, 11.
Masursky, H., Eliason, E., Ford, P. G., McGill, G. E., Pettengill, G. H., Schaber, G. G., and Schubert, G.: 1980, *J. Geophys. Res.* **85**, No AB, 8232.
Murrell, S. A. F.: 1976, *Tectonophyscs* **36**, 5.
Phillips, R. J. and Ivins, E. R.: 1979, *Phys. Earth Planet. Inter.* **19**, 107.
Ringwood, A. E. and Anderson, D. L.: 1977, *Icarus* **30**, 243.
Ross, J. V., Lallemant, H. G., and Carter, N. L.: 1978, *EOS Trans. AGU* **59**, 374.
Russell, C. T.: 1980, *Rev. Geophys. Space Phys.* **18**, 77.
Russell, C. T., Elphic, R. C., and Slavin, J. A.: 1980, *J. Geophys. Res.* **85**, 8319.
Schatz, J. F. and Simmons, G.: 1972, *J. Geophys. Res.* **77**, 6966.
Schubert, G.: 1979, *Ann. Rev. Earth Planet. Sci.* **7**, 289.
Stacey, F. D.: 1977, *Physics of the Earth*, p. 414, John Wiley and Sons, New York.
Surkov, Yu. A., Kirnozov, F. F., Khristianov, V. K., Glazov, B. N., Ivanov, V. F., and Korchuganov, B. N.: 1977, *COSPAR Space Res.* **17**, 651.
Toksöz, M. N., Hsui, A. T., and Johnston, D. H.: 1978, *The Moon and the Planets* **18**, 281.
Urey, H. C.: 1952, *The Planets, Their Origin and Development*, Yale Univ. Press, New Haven.
Zharkov, V. N.: 1958, *Izv. Akad. Nauk SSSR, Geophys. Ser.*, No. **11**, 1342.
Zharkov, V. N.: 1962, *Trudy Inst. Fiziki Zemli AN, No.* **20** (187), 3.
Zharkov, V. N.: 1983, *Internal Structure of the Earth and Planets*, Gordon and Breach, New York.
Zharkov, V. N. and Kalinin, V. A.: 1971, 'Equations of State for Solid at high Pressures and Temperatures', Consultants Bureau, New York.
Zharkov, V. N. and Trubitsyn, V. P.: 1978, in W. B. Hubbard (ed.), *Physics of Planetary Interiors*, p. 388. Pachart, Tucson.
Zharkov, V. N. and Trubitsyn, V. P.: 1980, *Physics of Planetary Interiors*, 448 pp, Nauka, Moscow.
Zharkov, V. N. and Zasurskiy, I. Ya.: 1980a, *DAN-SSSR* **253**, 840.
Zharkov, V. N. and Zasurskiy, I. Ya.: 1980b, *Izv. Acad. Nauk SSSR-Fiz. Zem. No.* **5**, 80.
Zharkov, V. N. and Zasurskiy, I. Ya.: 1981, *Astronomical Vestnik.* **15**, 11.
Zharkov, V. N. and Zasurskiy, I. Ya.: 1982, *Astronomical Vestnik.* **16**, N 1.
Zharkov, V. N., Kozlovskaya, S. V., and Zasurskiy, I. Ya.: 1981, *Adv. Space Res.* **1**, 117.