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The Thermal Conditions of Venus

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ABSTRACT

This paper examines models of Venus' thermal evolution. The models include the core which is capable of solidifying when the core's temperature drops below the liquidus curve, the mantle which is proposed as divided into two, independent of the convecting layers (upper and lower mantle), and the cold crust which maintains a temperature on the surface of the convective mantle close to 1200°C. The models are based on the approximation of parametrized convection, modified here to account for new investigations of convection in a medium with complex rheology.

Venus' thermal evolution, examined from the point when gravitational differentiation of the planet was completed (4.6 billion years ago), is divided into three periods: (1) adaptation of the upper mantle to the thermal regime of the lower mantle: approximately 0.5 billion years; (2) entry of the entire mantle into the asymptotic regime approximately three to four billion years; (3) asymptotic regime. The parameters of a convective planet in an asymptotic regime are not dependent on the initial conditions (the planet "forgets" its initial state) and are found analytically. The thermal flux in the current epoch is ~ 50 ergs $cm^2 s^{-1}$. We consider the connection between the thermal regime of Venus' core and its lack of magnetic field. After comparing the current thermal state of thermal models of Venus and Earth, together with the latest research on melting of the Fe-FeS system and the phase diagram of iron, we propose that Venus' lack of its own magnetic field is related to the fact that Venus' core does not solidify in the contemporary epoch. The particular situation of the iron triple point (γ – ϵ – melt) strengthens this conclusion. We discuss the thermal regime of the

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Venusian crust. We demonstrate that convection in the lower portion of the crust plays a minor role in regions with a particular crust composition, but that effusive or intrusive heat transport by melt, formed from melting of the crust's lower horizons, is the dominant mechanism for heat transport to the surface.

MODIFIED APPROXIMATION OF PARAMETRIZED CONVECTION

Models of Venus' thermal evolution, calculated in approximation of parameterized convection (APC), were examined in the works of Schubert (1979); Turcotte *et al.* (1979); Stevenson *et al.* (1983); Solomatov *et al.* (1986); and Solomatov *et al.* (1987). In parameterizing, dependencies were used that were obtained from studying convection in a liquid with constant viscosity. They were inferred to be true in cases of more complex rheology. New numerical investigations of convection in media with rheology that is more appropriate for the mantle (Christensen 1984a, b, 1985a, b) and theoretical research (Solomatov and Zharkov 1989) necessitated the construction of a modified APC (MAPC).

Let the law of viscosity be expressed as (Zharkov 1983):

$$\eta = b/\tau^{m-1} \exp \left[A_o/T \left(\rho/\rho_o\right)^L\right] \tag{1}$$

where τ denotes the second invariant of the tensor of tangential stresses, ρ is density; b and L are considered here to be the constants within the upper and lower mantles; m $\approx$ 3; A_o denotes the enthalpy of activation for self diffusion (in K); and where $\rho = \rho_o$ is the reference value of density selected at the surface of each layer.

To describe convection in such a medium, following Christensen (1985 a), we will use two Raleigh figures:

$$Ra_o = \frac{\alpha g \rho \Delta T d^3}{\chi \eta_o}, \tag{2}$$

$$Ra_T = \frac{\alpha g \rho \Delta T d^3}{\chi \eta_T}, \tag{3}$$

where α denotes the thermal expansion coefficient; g is the acceleration of gravity; ΔT is the mean superadiabatic temperature difference in the layer; d denotes layer thickness, χ is the coefficient of thermal diffusivity; and η_o and η_T are defined by the formulae:

$$\eta_o = \frac{b}{\tau_o^{m-1}} \exp \frac{A_o}{T_o}; \tau_o = \eta_o \frac{x}{d^2} \tag{4}$$

$$\eta_T = \frac{b}{\tau_o^{m-1}} \exp \frac{A_o}{\overline{T}} \left(\frac{\overline{\rho}}{\rho_o}\right)^L. \tag{5}$$

The dependence of the Nusselt number (determined in terms of the thickness of the thermal boundary layer δ) on Ra_o and Ra_T, where m = 3, is parameterized by the formula

$$Nu = \frac{d}{2\delta} = aRa_o^{\beta_o} Ra_T^{\beta_T}. \qquad (6)$$

In the case of free boundaries (lower mantle):

$$a = 0.29, \quad \beta_o = 0.37, \quad \beta_T = 0.16. \qquad (7)$$

In the case of fixed boundaries (upper mantle):

$$a = 0.13, \quad \beta_o = 0.35, \quad \beta_T = 0.15. \qquad (8)$$

The theoretical expression has the following appearance (Solomatov and Zharkov 1989):

$$Nu = \pi - \frac{3m+4}{2(m+2)} Ra_0^{\frac{m-1}{m+2}} Ra_T^{\frac{1}{m+2}} = 0.23 Ra_0^{0.4} Ra_T^{0.2}. \qquad (9)$$

It is obtained from the balance of the capacities of viscous dissipation and buoyancy forces, and is in good agreement with (7) and (8).

The heat flow at the upper or lower boundary of the layer is equal to

$$F_i = \frac{æ\Delta T_i}{\delta}, \qquad (10)$$

where ΔT_i denotes the temperature difference across the thermal boundary layer. Velocities at the boundary (u), mean tangential stresses in the layer (τ) and mean viscosity ($\overline{\eta}$) are estimated by the formulae (m=3):

$$u = \frac{\pi}{4}\frac{xd_2}{\delta}, \qquad (11)$$

$$\overline{\tau} = \left(\frac{\pi u b}{d} \exp\frac{A}{T^o}\left(\frac{\overline{\rho}}{\rho_o}\right)^L\right)^{1/3}, \qquad (12)$$

$$\overline{\eta} = \frac{b_2}{\overline{\tau}} \exp\frac{A_o}{\overline{T}}\left(\frac{\overline{\rho}}{\rho_o}\right)^L. \qquad (13)$$

The mean temperature of the layer, $\overline{T}$, and the temperature of the top of the lower thermal boundary layer, T_L, can be calculated from the adiabatic relationship through the temperature in the base of the upper thermal boundary layer, T_u:

$$\overline{T} = nT_u; \; T_L = n_L T_u, \qquad (14)$$

where n and n_L are constants.

With large Ra_T/Ra_o, the very viscous upper thermal boundary layer becomes reduced in mobility, taking an increasingly less effective part in convection, and the previous formulae are not applicable. Solomatov and Zharkov (1989) estimated that the transition to a new convective regime occurs when

$$Ra_T \gtrsim Ra_{Ttr} \approx 2Ra_o^2;\ m = 3. \tag{15}$$

DESCRIPTION OF THE MODEL

There is no unequivocal answer to the question of whether convection in the Earth's mantle (or Venus') is single layered or double-layered. Previous works explored the single-layered models of convection. Possible differences in the single- and double-layered model of Venus' thermal evolution were discussed by Solomatov *et al.* (1986) and Solomatov *et al.* (1987). We propose here that convection is double-layered, and the boundary division coincides with the boundary of the second phase transition at a depth of approximately 756 kilometers (Zharkov 1983).

The thermal model of Venus (Figure 1a) contains a cold crust, whose role is to maintain the temperature in its base at approximately 1200°C (melting temperature of basalts), the convective upper mantle, the convective lower mantle, and the core. An averaged, spherically symmetric distribution, T (r,t) is completely determined by the temperatures indicated in Figure 1b.

The thermal balance equations for the upper mantle, the lower mantle, and the core are written as:

$$\frac{4}{3}\pi(R_L^3 - R_{12}^3)\rho_1 C_{p1}\frac{d\overline{T}}{dt^1} = 4\pi R_{12}^2 F_{12} - 4\pi R_L^2 F_L, \tag{16}$$

$$\frac{4}{3}\pi(R_{12}^3 - R_c^3)\rho_2 C_{p2}\frac{d\overline{T}}{dt^2} = \frac{4}{3}\pi(R_{12}^3 - R_c^3)\rho_2 Q_2 - 4\pi R_{12}^2 F_{21} + 4\pi R_c^2 F_c, \tag{17}$$

$$-\frac{4}{3}\pi R_c^3 C_{pc}\rho_c\frac{d\overline{T}}{dt^c} + Q_c\frac{dm}{dt} = 4\pi R_c^2 F_c. \tag{18}$$

Indices "1," "2," and "C" relate, respectively, to the upper mantle, lower mantle and the core. $\overline{T}$ denotes the mean layer temperature, F_L is the heat flow from the mantle under the lithosphere. The heat flow at the surface of the planet is obtained by adding to F_L, thermal flow generated by the radiogenic production of heat in the crust (~11 erg cm^{-2}s^{-1}). The radius of the lithosphere boundary (R_L) differs little from the radius of the planet (R_o) so that $R_L \approx R_o$. It is supposed that almost all of the radioactive elements of the upper mantle migrated into the crust when the

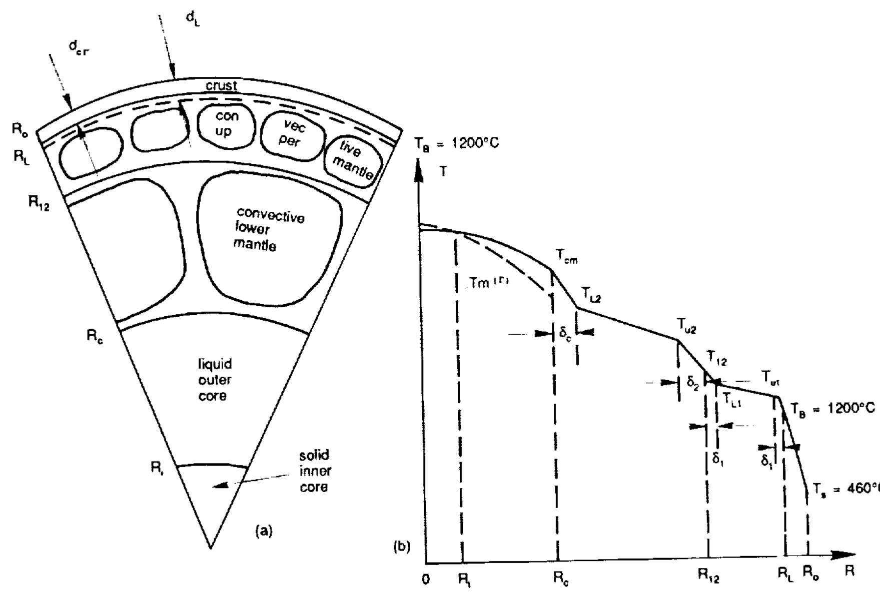

FIGURE 1 (a) Diagram of Venus's internal structure. Radii are indicated for: the planet—R_o; the base of the lithosphere—R_L; the boundary between the upper and lower mantles—R_{12}; the core—R_c; and the solid internal core —R_i. The dashed line illustrates the boundary of the lithosphere, which is an isothermic surface of T_B = $1200°C$ = const; d_{cr} and d_L denote the thickness of the crust and the lithosphere. (b) Schematic, spherically symmetric temperature distribution in the cores of Venus. Reference temperatures are indicated for: the surface—T_s; lithosphere base—T_B; base of the upper thermal boundary layer of the upper mantle—T_{u1}; peak of the lower boundary layer of the upper mantle—T_{L1}; boundary between the upper and lower mantles—T_{12}; base of the upper boundary layer of the lower mantle—T_{u2}; peak of the lower boundary layer of the lower mantle—T_{L2}; and the boundary between the core and the mantle—T_{cm}. Thicknesses of the thermal boundary layers are given: δ_1 for the boundaries of the upper mantle; δ_2 and δ_c for the boundaries of the lower mantle. The dashed line indicates the core melting curve, which intersects the adiabatic temperature curve at the boundary between the outer, liquid and inner, solid core.

crust was melted. F_c denotes the heat flow from the core. The heat flow at the boundary between the upper and lower mantles with a radius of R_{12} is continuous: $F_{12} = F_{21}$.

Heat production in the lower mantle (Q_2) is defined by the sum

$$Q_2 = \sum_{i=1}^{4} Q_{oi} \exp\left[\lambda_i(t_o - t)\right] \tag{19}$$

where Q_{oi} and λ_i denote the current heat production of the radioactive isotopes K, U, and Th for one kilogram of undifferentiated silicate reservoir of the mantle and their decay constants. The concentrations of K, U, and Th are selected in accordance with O'Nions *et al.* (1979): U = 20 mg/t, K/U = 10^4, Th/U = 4.

The term, Q_c dm/dt, in (18) describes heat release occurring when the core solidifies after the core adiabat drops below the core liquidus curve. It is supposed that the core consists of the mixture, Fe-FeS, and as solidification begins from the center of the planet, sulfur remains in the liquid layer, reducing the solidification temperature. The value Q_c is composed of the heat of the phase transition and gravitational energy.

According to the estimates of Loper (1978); Stevenson *et al.* (1983), and Solomatov and Zharkov (1989), $Q_c = (1\text{-}2) \cdot 10^{10}$erg g^{-1}.

Mean temperature of the adiabatic core:

$$\overline{T}_c \approx n_c T_{CM} \tag{20}$$

where $n_c \approx 1.2$, the mean core density is $\overline{\rho} = 10.5$ g cm^{-3}, $C_{PC} = 4.7 \cdot 10^6$ erg g^{-1} K^{-1} (Zharkov and Trubitsyn 1980; Zharkov 1983).

The formulae for the melting $T_m(\rho)$ and adiabatic $T_{ad}(\rho)$ curves are written as follows:

$$T_m(\rho) = T_o(1 - \propto x)\left(\frac{\rho}{\rho cm}\right)^{2.24}, \tag{21}$$

$$T_{ad}(\rho) = T_{cm}\left(\frac{\rho}{\rho_{cm}}\right)^{1.45}, \tag{22}$$

$$\frac{\rho}{\rho_{cm}} - 1.224 - 0.009405\frac{r}{R_c} - 0.1586\left(\frac{r}{R_c}\right)^2 - 0.05672\left(\frac{r}{R_c}\right)^3. \tag{23}$$

Here T_o denotes the temperature at which pure iron melts at the boundary core of the radius R_c, where $\rho = \rho_{CM} = 9.59$ g cm^{-3}; $\alpha \approx 2$; x is the mass portion of sulfur in the liquid core, depending upon the radius of the solid core R_i, and the overall amount of sulfur in the core, x_o:

$$x(R_i) = \frac{x_o R_c^3}{R_c^3 - R_i^3} \tag{24}$$

The intersection of (21) and (22) defines the radius of the solidified core R_i (Figure 1b).

Convection in the upper mantle is parameterized by the MAPC with the parameters (7) (for fixed boundaries) in the lower mantle; and by the same formulae with the parameters (8) (for free boundaries). However, the difference between (7) and (8) is not very substantial. The parameters for the upper mantle are: $b_1 = 4.3 \cdot 10^{15}$dyne^{3}cm^{-6}s, $A_{o1} = 6.9 \cdot 10^4$K,

$\alpha_1 = 3 \cdot 10^{-5}K^{-1}$, $\bar{\rho}_1 = 3.7$ g cm^{-3}, $\chi_1 = 10^{-2}$cm^{-2}s^{-1}, $æ_1 = 4.5 \cdot 10^5$erg cm^{-1}s$\cdot$ 10^7erg g

and for the lower mantle:

$b_2 = 1.1 \cdot 10^{17}$dyne^{3}cm^{-6}s, $A_{o2} = 1.3 \cdot 10^5$K, $\alpha_2 = 1.5 \cdot 10^{-5}K^{-1}$, $\bar{\rho}_2 = 4.9$ g cm^{-3}, $\chi_2 = 3 \cdot 10^{-2}$cm^{-2}s^{-1}, $æ_2 = 1.8 \cdot 10^6$erg cm^{-1}s^{-1}K^{-1}, g = 900cm s^{-2}, $C_{p2} = 1.2 \cdot 10^7$erg g^{-1}, $n_2 = 1.13$, $n_{L2} = 1.26$.

NUMERICAL RESULTS AND ASYMPOTOTIC SOLUTION OF THE MAPC EQUATIONS

The evolution of the planet began approximately 4.6 billion years ago. The initial state of the planet was a variable parameter. The initial value of T_{u2} is the most significant, since T_{u1}, due to the low thermal inertia of the upper mantle, rapidly adapts to the thermal regime of the lower mantle ($t < 0.5$ billion years). The core has little effect on the evolution of the planet in general and the majority of models do not take into account its influence. T_{u2} was selected as equal to 2500, 3000, 3500 K. The upper value is limited by the melting temperature of the mantle, since a melted mantle is rapidly freed from excess heat.

The upper mantle adapts to the thermal regime of the lower mantle for the first approximately 0.5 billion years (Figure 2a, b). Then, after approximately three to four billion years the entire mantle enters an asymptotic regime which is not dependent upon the initial conditions. The evolution picture in general is similar to the one described by Solomatov *et al.* (1986); and Solomatov *et al.* (1987). The planet is close to an asymptotic state in the present epoch. Contemporary parameters of the models are:

T_{u1} = (1700-1720) K,	u_1 = (2.1-2.4)cm yr^{-1}
T_{12} = (2500-2530) K,	u_2 = (0.8-1.0)cm yr^{-1}
T_{u2} = (2840-2870) K,	$\bar{\tau}_1$ = (5-6) bars,
F_L = (35-40)erg cm^{-2} s^{-1},	$\bar{\tau}_2$ = (110-120) bars,
σ_1 = (28-30)km,	$\bar{\eta}_1$ = (1-2)10^{21} poise,
σ_2 = (130-140)km,	$\bar{\eta}_2$ = (3-10)10^{22} poise.

According to the criterion (15) MAPC are applicable throughout the entire evolution, just as with the quasistationary criterion (Solomatov *et al.* 1987).

The asymptotic expression for F_L in the first approximation is formulated as (Solomatov *et al.* 1987):

$$F_L = F_Q\left(1 + \frac{t_{in}}{t_r}\right), \tag{25}$$

where F_Q denotes the thermal flow created by the radioactivity of the lower mantle, and t_r is the characteristic time of decay:

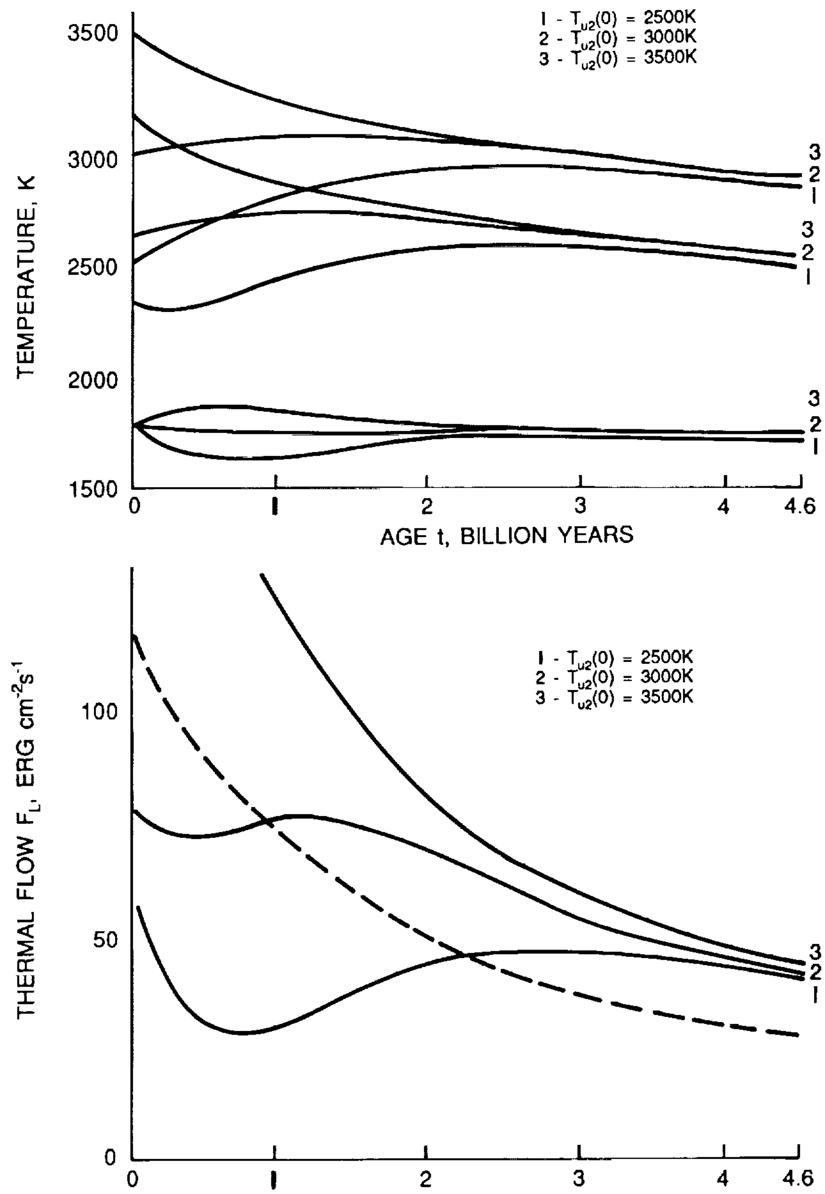

FIGURE 2 (a) Evolution of the base temperatures, T_{u1}, T_{12} and T_{u2}, with differing initial conditions. The core is not taken into account and is considered to be $T_{cm} = T_{L2}$ (Figure 1). (b) Evolution of the thermal flow under the lithosphere, F_L, with differing initial conditions. The thermal flow to the surface is obtained by adding $\sim$ 11 erg $cm^{-1}s^{-1}$ to F_L, generated by the radioactive elements of the crust. The dashed line indicates the thermal flow generated by radioactive elements of the lower mantle.

$$t_r = -\left(\frac{dlnQ_2}{dt}\right)^{-1}, \tag{26}$$

equal $\sim 5.67 \cdot 10^9$ years at the current time.

The time scale of thermal inertia of the mantle is equal to

$$t_{in} = t_{in1} + t_{in2}(1+\delta), \tag{27}$$

where t_{in1} and t_{in2} are the time scales for the inertia of each of the layers individually:

$$t_{in1} = \frac{M_1 n_1 C_{p1} \Delta T_1}{M_2 Q_2}\left(1+\beta_1+\frac{\beta_1 A_1 (T_{u1}-T_B)}{T_{u1}^2}\right)^{-1}, \tag{28}$$

$$t_{in2} = \frac{n_2 C_{p2} \Delta T_2}{Q_2}\left(1+\beta_2+\frac{\beta_2 A_2 (T_{u2}-T_{12})}{T_{u2}^2}\right)^{-1}, \tag{29}$$

$$\xi = \frac{\Delta T_1}{\Delta T_2}\frac{1+n_{L1}\bar{s}}{\bar{s}}\frac{1+\beta_o+\beta_2-1/3(\beta_o-2\beta_2)A_{o2}(T_{u2}-T_{12})T_{u2}^{-2}}{1+\beta_1+\beta_1 A_1(T_{u1}-T_B)T_{u1}^{-2}}, \tag{30}$$

M_1 and M_2 denote the layer masses, $\bar{s} = R_{12}^2 / R_L^2$. The values of (27) through (30) are calculated in a zero approximation: $F_L = F_Q$, and $A = A_o/n\ (\rho/\rho_o)^L$.

The time scale for thermal inertia, as compared with models based on the conventional APC (Solomatov *et al.* 1986, 1987), has increased from $\sim 2.5 \cdot 10^9$ years to $\sim 3.5 \cdot 10^9$ years. The mantle and core temperatures obtained are somewhat lower (by 300 K near the core). In the models with the core, $T_{cm} \approx 3720$K and $F_c \approx 15$ erg cm^{-2}s^{-1} in the contemporary epoch. Since the adiabatic value is $F_c \approx 30$ erg cm^{-2}s^{-1}, there is no convection if solidification is absent, and a magnetic field cannot be generated (Stevenson *et al.* 1983).

MAGNETISM AND THE THERMAL REGIME OF THE CORES OF EARTH AND VENUS

We will estimate the temperatures in the Earth using the MAPC. MAPC is not applicable to the Earth's upper mantle, since convection in the Earth involves the surface layer, and rheology is, in general, more complex. However, MAPC is applicable to the Earth's lower mantle. Let us assume a value for the temperature at the boundary between Earth's upper and lower mantles of

$$T_{12} = (2300-2500)K, \tag{31}$$

(Zharkov 1983), and the thermal flow at this boundary is

$$F_{21} \approx \frac{R_o^2}{R_{12}^2}\left(F - F_{cr} + C_p M_1 \frac{d\overline{T}_1}{dt}\right) \approx 80 erg cm^{-2} s^{-1},$$

where $F - F_{cr} \approx 70$ erg $cm^{-2}s^{-1}$ is the medium thermal flow from the Earth, after deducting heat release in the crust (according to the estimates of Sclater *et al.* 1980), $d\overline{T}_1/dt \approx$ -100K/billion years. (Basaltic Volcanism Study Project 1981).

We will assume the following values for the rheologic parameters:

$$b = 6.6 \cdot 10^{16} dyne^3 cm^{-6} s, A = 1.5 \cdot 10^5 K, and A_o = 1.3 \cdot 10^5 K.$$

The remaining parameters are listed in Zharkov (1983).

As a result, we have:

T_{u2} = (2800-2900) K,
T_{L2} = (3600-3800) K,
T_{cm} = (3800-4000) K,
δ_2 = 100 km.

The value of T_{cm} - T_{L2} depends upon F_c, which we will assume to be equal to the adiabatic value of $F_{ad} \approx 30$ erg $cm^{-2}s^{-1}$.

Therefore, the temperature at the boundary of the Earth's core, T_{CME} is greater than for Venus (T_{CMV}) by a value of

$$T_{CME} - T_{CMV} = (100 - 300)K. \quad (33)$$

In order for Venus' core not to solidify, it is necessary that the adiabat of the Venusian core not drop below the solidification curve during cooling. Otherwise, solidification of the core will cause the core to mix by chemical or thermal convection, and it will trigger the generation of a magnetic field (Stevenson *et al.* 1983). We will estimate the difference between T_{CME} and the temperature, T_{crV} (which is critical for the beginning of solidification), at the boundary of Venus' core, with a single pure iron melting curve, T_m (P) and a single equation of the state for iron ρ(P).

T_{CME} is found from the intersection of the adiabat of the Earth's core (22) and the liquidus curve (23), with a sulfur content in the core of x_E at the boundary of the Earth's inner core. T_{crV} is obtained from the intersection of the adiabat of Venus' core (22) and the liquidus curve (23) with a sulfur content of x_V. We then obtain (Solomatov and Zharkov 1989):

$$T_{CME} - T_{crV} = (+300) \div (-300), K, \quad (34)$$

where $x_E - x_V = 0 \div 0.07$. Therefore, if $x_V \gtrsim x_E - (0 \div 0.02) = 0.07 \div 0.12$; (with $x_E = 0.09 \div 0.12$; Aherns 1979), the core of Venus is not solidifying at the present time, and a magnetic field is not being generated.

Complete solidification of the core would have led to the absence of a liquid layer in the core and would have made it impossible for the magnetic field to be generated. However, for this, the temperature near the boundary of Venus' core should have dropped below the eutectic value, which, according to the estimates of Anderson *et al.* (1987) is ~ 3000 K and according to Usselman's estimate (1975) is ~ 2000 K. Such low temperatures of the core seem to be of little probability.

Pressure in the iron triple point, $\gamma - \epsilon -$ 1, (liquid) approaches the pressure in the center of Venus. This is significant for interpretation the absence of the planet's own magnetic field. According to the estimates of Anderson (1986), $P_{tp} \approx 2.8$ Mbar (Figure 3), but is also impossible to rule out the larger values of P_{tp}. In the center of Venus, $P_{cv} \approx 2.9$ Mbar; at the boundary of the solid inner Earth's core, $P_{IE} = 3.3$.Mbar; and in the Earth's center, $P_{CE} = 3.6$ Mbar (Zharkov 1983). If $P_{cv} \lesssim P_{tp}$, the conclusion that there is no solidification of Venus' core is further supported, since, in this case, the core's adiabat (critical for the beginning of solidification) drops several hundred degrees lower. This stems from the fact that reduction in the temperature of solidification of the mixture, $\gamma-$ Fe $-$ FeS is greater than for the mixture $\epsilon-$ Fe $-$ FeS by $\Delta S_\epsilon / \Delta S_\gamma$ times, where $\Delta S_\epsilon / \Delta S_\gamma$ denotes the ratio of enthropy jumps (during melting) equal to ~ 2, according to Anderson's estimates (1986).

The latest experimental data on the melting of iron allow us to estimate the melting temperature in the cores of Earth and Venus. Figure 3 shows melting curves obtained by various researchers, and the $\gamma - \epsilon$ boundary, computed by Anderson (1986). With x = 0.09-0.12 and $\alpha = 1 \div 2$ (formula 21), the full spread of temperatures at the boundary of the Earth's core, leading to the intersection of the liquidus and adiabat curves, is equal to

$$T_{CME} = 3500 \div 4700K, \tag{35}$$

- with data from Brown and McQueen (1986) and

$$T_{CME} = 4300 \div 5400K, \tag{36}$$

- with data from Williams *et al.* (1987).

The effect of pressure on viscosity of the lower mantle reduces the effective Nusselt number and increases the temperature of T_{CME} and T_{CMV} by ~ 300 K (Solomatov and Zharkov 1989). We obtain the estimate

$$T_{CME} = 3800 \div 4300K, \tag{37}$$

which is the best fit with (35).

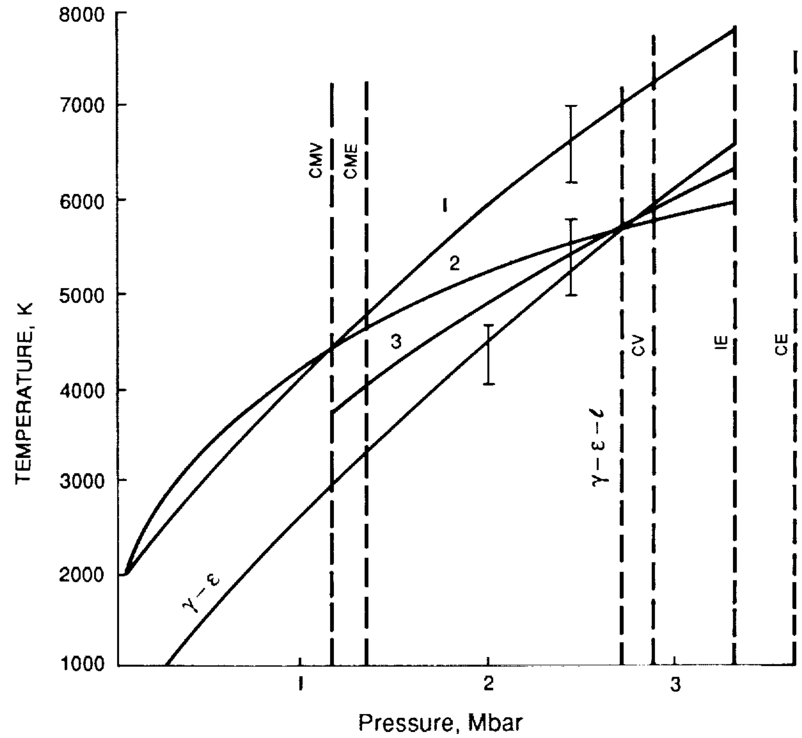

FIGURE 3 Pure iron melting curves according to various data and the boundary of the phase transition γ—Fe—ϵ—Fe. The figures indicate melting curves: 1 is from the study by Williams *et al.* (1987); 2 is from the study by Anderson (1986) and experimental data from Brown and McQueen (1986); and 3 is based on the Lindeman formula with the Grunehaisen parameter of G = 1.45 = const, using experimental data from Brown and McQueen (1986). The transition boundary, γ—ϵ, was constructed in the study by Anderson (1986) and together with the melting curve gives us the location of the triple point. The vertical segments illustrate errors in determining temperature for curves 1 and 2. The boundary labels are as follows: CM is the boundary between the core and the mantle; C denotes the center of the planet; I is the boundary of the solid inner core; and the final letter indicates Earth (E) or Venus (V). γ - ϵ - l denotes the location of the triple point.

THE THERMAL REGIME OF THE VENUSIAN CRUST

What are the mechanisms by which heat is removed from Venus's interior to the surface? Global plate tectonics, like Earth's, are absent on the planet (Zharkov 1983; Solomon and Head 1982), although in individual, smaller regions, it is possible that there are features of plate tectonics (Head and Crumpler 1987). A certain portion of heat may be removed by the mechanism of hot spots (Solomon and Head 1982; Morgan and Phillips

1983). Conductive transport of heat through the crust clearly plays a large role. However, with a large crust thickness ($\gtrsim$ 40 km), as indicated by data from a number of works (Zharkov 1983; Anderson 1980; Solomatov *et al.* 1987), the heat flow of $F \approx$ (40-50) erg $cm^{-2}s^{-1}$ triggers the melting of the lower crust layers and removal of approximately one half of heat flow by the melted matter. The latter flows to the surface as lava or congeals in the crust as intrusions.

We shall discuss the possibility of solid-state convection in the crust as an alternative mechanism of heat removal through the Venusian crust.

The crust is constituted of an upper, resilient layer with a thickness of d_e and a viscous one with a thickness of $d_v = d_{cr} - d_e$.

The crust thickness of d_{cr} is constrained in our model by the phase transition of gabbro-eclogite, since eclogite, with a density higher than that of the underlying mantle, will sink into the mantle (Anderson 1980; Sobolev and Babeiko 1988). The depth of this boundary is dependent upon the composition of basalts of Venus' crust. We assume that d_{cr} = 70 km (Yoder 1976; Zharkov 1983; Sobolev and Babeiko 1988). It is considered that radioactive elements are concentrated in the upper portion of the crust, and the primary heating occurs via the heat flow from the mantle of $F_L \approx 40$ erg $cm^{-2}s^{-1}$. The boundary (d_e) of the resilient crust is defined as the surface of the division between the region effectively participating in convection and the nonmobile upper layer. Crust rheology is described by law (1) with the parameters from Kirby and Kronenberg (1987). Four modeled rocks are considered: quartz diorite, anorthosite, diabase and albite. Parameter m in (1) is close to 3 for them, so that the formulae MAPC with parameters (8) for fixed boundaries is fully suitable for estimating. In addition, criterion (15) is used to define the boundary, d_e. At this boundary, which is regarded as the upper boundary of the convective portion of the crust, the temperature is equal to

$$T_o = 733 + 20d_e(km), K. \tag{38}$$

The following physical parameters are assumed (Zharkov *et al.* 1969):

$$\rho = 2.8g\ cm^{-3}, \alpha = 2 \cdot 10^{-5}K^{-1},\ æ\ = 2 \cdot 10^5 erg\ cm^{-1}s^{-1}K^{-1}.$$

Computations have demonstrated that the thickness of the resilient crust is 20-30 kilometers. The mean temperature of the convective layer of the crust has been calculated at $\sim$ 1600K, 1700K, 1900K and 2000K, respectively, for quartz diorite, anorthosite, diabase, and albite, and exceeds the melting temperature for basalts by hundreds of degrees. This means that convection does not protect the crust from melting, and heat is removed by the melted matter.

We can estimate that portion of the heat which is removed by convection. For this, let us assume the temperature in the base of the crust to be T_B = 1500K (approximately the melting temperature). We obtain the Nusselt number from the formulae of MAPC, and find that it is ~1.7; 1; 0.6; and 0.5, respectively, for quartz diorite, anorthosite, diabase, and albite. For η = const, and for complex rheology, as in (1), (Christensen 1984a and 1985a), convection begins at Nu ~ 1.5-2, as calculated according to the formulae of APC (MAPC). Thus, where $d_{cr} \approx$ 70km, perhaps only quartz diorite is convective, removing 25-30 erg $cm^{-2}s^{-1}$, while 10-20 erg $cm^{-2}s^{-1}$ is removed by the melt. The rate of convective currents is three to five millimeters per year. It can be demonstrated that

$$Nu \sim (d_{cr} - d_e)^{0.9} \tag{39}$$

With $d_{cr} \sim$ 120 km, convection also begins in anorthosite, while in quartz diorite it occurs virtually without melting.

These estimates show that convection in the crust may play some role in individual regions, depending on the thickness and composition of the crust. It is not excluded that in individual regions, convection in the crust makes its way to the surface. The bulk of the heat is, apparently, removed by melted matter. The rate of circulation of material from the crust is, with this kind of volcanism, 50-100 km^3yr^{-1}. This is three to five times greater than crust generation in the terrestrial spreading zones. Another process by which basalt material circulates is where new portions of melted basalt reach the crust from the upper mantle, and basalt returns back to the mantle in the eclogite phase. This process may trigger the accumulation of eclogite in the gravitationally stable region between the upper and lower mantles. It may lead to the chemical differentiation of the mantle. This process has been noted for the Earth by Ringwood and Irifune (1988).

Figure 4 illustrates various processes involved in heat and mass transport.

The geological structures observed on Venus may be related to these processes. Flat regions may be tied to effusive, basalt volcanism. Linear structures in the mountainous regions may be related to horizontal deformations which are an appearance of convections in the crust or mantle. Ring structures may stem from melted intrusion or hot plume lifted towards the surface from the bottom of the upper mantle.

Convection both in the mantle and in the crust is, apparently, nonstationary (as on Earth). This nonstationary nature stems from instabilities occurring in the convective system. The characteristic time scale for such fluctuations is t ~ d/u, where d ~ 10^8-10^9 cm is the characteristic dimension, and u ~ 1 cm per year is the characteristic velocity. Therefore, characteristic "lifespans" for various occurrences of instability are t ~ 10^8-10^9 years.

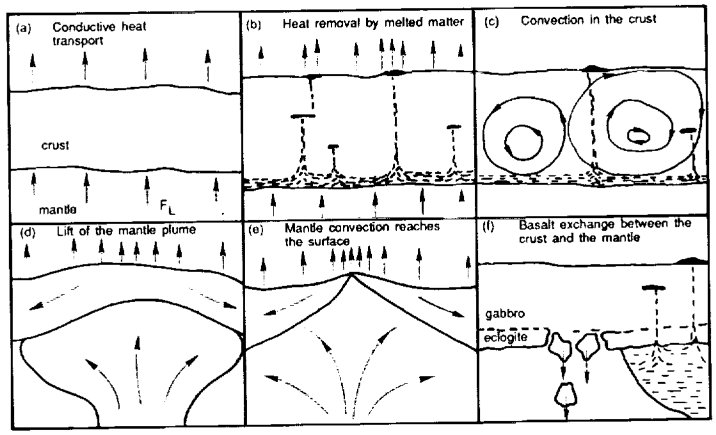

FIGURE 4 Process of heat transport through the Venusian crust: (a) transfer of heat carried from the mantle by the mechanism of conductive thermal conductivity; (b) heat removal by the melted matter which is formed as the basalt crust melts. The melted matter may flow to the surface or form intrusions; (c) convection in the crust which does not reach the surface or reaching the surface; (d) lifting of hot plume from the bottom of the upper mantle to the crust of Venus, triggering enhanced heat flow, a flow of crust material, and the sinking of the crust; (e) involvement of the crust in mantle convection with the formation of spreading zones; (f) basalt exchange between the crust and the mantle. Basalt is formed when the upper mantle partially melts and is returned back as eclogite.

Regional features of tectonic structures, thermal flows, volcanic activity and so on, can exist for this length of time.

CONCLUSION

1. Modification of the approximation of parameterized convection for the case of nonNewtonian mantle rheology led to no marked increase in the time scale of thermal inertia of the mantle from two to three or three to four billion years in comparison with the usual parameterization. The thermal flow at the surface of Venus of $\sim$ 50 erg cm^{-2}s^{-1}, is the product of radiogenic heat release from the mantle (50%), heat release in the crust (20%), and cooling of the planet (30%). These figures for Earth are approximately 40, 20, and 40% for the double-layered convection models. Temperatures in the upper portion of the mantle are approximately 1700 K, which is 50-100 K greater than on Earth. Given the existing uncertainties in the concerning parameters, T_{cm} $\sim$ 3700-4000 K at the core boundary and may, possibly, be greater. This temperature is 100-300 K less than for the Earth.

2. The magnetic field on Venus is absent. This is most likely due to the lack of core solidification and, respectively, the lack of energy needed

to maintain convection in the liquid core. For this, it is enough for the sulfur content in Venus' core to be even a little bit less than on the Earth (but not more than 20-30% less). This conclusion is stronger if the triple point in the phase diagram for iron, $\gamma - \epsilon - 1$, lies at pressures that are greater than in Venus' center, but approximately less than in Earth's center.

3. At the values obtained for heat flows from Venus' mantle, the crust melts, and the melted matter ($\lesssim$ 100 km^3 per year) removes about one half of the entire heat. The remainder is removed conductively. In individual regions, depending on the crust thickness and composition, a portion of the heat may be removed by convection, reducing crust temperature and the portion of heat removed by the melted matter. Flow velocities comprise several millimeters per year. Due to the insufficiently high temperature of the surface, convection in the crust separates from the surface as a highly viscous, nonmobile layer with a thickness of 20-30 kilometers. In individual regions, crust convection may emerge to the surface. Basalt circulation also occurs by another way: the basalt is melted out of the upper mantle and returned back in the form of eclogite masses, which sink into the lighter mantle rock. It is possible that this process triggers the accumulation of eclogite at the boundary between the upper and lower mantles, resulting in the chemical separation of the mantle. These processes may explain the formation of various geological structures on Venus.

4. The nonstationary nature of convection in Venus' mantle and crust determine regional features of tectonic, thermal and volcanic appearances on the surface of the planet which have a characteristic duration of $\sim 10^8$-10^9 years.

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