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EFFECTS OF SELECTIVE FUSION ON THE THERMAL HISTORY OF THE MOON, MARS, AND VENUS *

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A comparative study on the thermal history of the Moon, Mars, and Venus was made by numerical solutions of the heat equation including and excluding selective fusion of silicates. Selective fusion was approximated by melting in a multicomponent system and redistribution of radioactive elements. Effects of selective fusion on the thermal models are (1) lowering (by several hundred degrees centigrade) and stabilizing the internal temperature distribution, and (2) increasing the surface heat-flow.

1. INTRODUCTION

A classic approach to the thermal history of a planet is to formulate it as an initial-boundary-value problem. First, a temperature distribution $T(\mathbf{x},t_0)$, $\mathbf{0} \leqslant \mathbf{x} \leqslant \mathbf{R}$ at some initial time t_0 is assumed, and a boundary condition $T(\mathbf{R},t)$ at the surface of the planet for all its history $t \geqslant t_0$ is specified. We then apply the principle of conservation of energy with an assumed heat transport mechanism. Traditionally the heat-conduction equation is used and is written in cartesian tensor notation as:

$$(1) \quad \rho c \frac{\partial T}{\partial t} = \frac{\partial}{\partial x_j}\left(K \frac{\partial T}{\partial x_j}\right) + A\,, \qquad (1)$$

where ρ = density, c = specific heat, T = temperature, t = time, x_j = jth component of position vector $\mathbf{x}$, K = thermal conductivity including radiative heat transfer, A = heat production per unit volume per unit time.

After specifying the values of ρ, c, K, and A, eq. (1) can be solved with the assumed initial and boundary conditions for the temperature distribution of the planet as a function of time: $T(\mathbf{x},t)$, $\mathbf{0} \leqslant \mathbf{x} \leqslant \mathbf{R}$ and $t \geqslant t_0$. The surface heat-flow of the planet can also be readily computed as a function of time by:

$$(2) \quad Q(t) = K \cdot \nabla T(\mathbf{x},t)\Big|_{\mathbf{x}=\mathbf{R}}\,. \qquad (2)$$

The validity of any thermal model can be tested because (1) the present surface heat-flow can be measured directly, and (2) the present temperature distribution can be deduced from geophysical observations. Thermal models which disagree with the observations must be rejected. However, that a thermal model which agrees with the observations can be found does not necessarily imply that the actual thermal history of the planet must have developed exactly as the thermal model. The thermal-history problem does not have a unique solution because a number of different assumptions may lead to the same results.

The simple heat-conduction model based on eq. (1), has been used by MacDonald [6, 7] to investigate the thermal history of the Moon and the terrestrial planets in great detail. Recently Lee [4, 5] has constructed more complicated thermal models than eq. (1). His results show that selective fusion of silicates can play a dominant role in the thermal history of the earth's mantle. The present letter is an extension of Lee's work to the Moon, Mars, and Venus. Its purpose is to show the differences in the development of internal

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temperature distribution and surface heat-flow between thermal models including and excluding selective fusion of silicates. The scope of the present letter, however, does not permit detailed discussion on the choice of parameters that enter into the thermal-history calculations and its effects on the results. Detailed investigations on the thermal history of the Moon, Mars, and Venus in the light of recent advances are now in preparation and will be published later.

2. THERMAL MODELS WITH SELECTIVE FUSION

Lee [4] has developed a thermal-history calculation for the earth which incorporates selective fusion and migration of radioactive elements. Extension to the Moon, Mars, and Venus follows simply by using parameters that are appropriate for these bodies.

All thermal models studied are spherically symmetric. Selective fusion of silicates is approximated by (1) melting in a multicomponent system, and (2) redistribution of radioactive elements by upward migration of magmas in a viscous medium. Latent heats for melting and freezing of a multicomponent system are allowed for in the computation of temperatures across the transition boundaries. Moving heat sources and heat transfer due to penetrative convection are taken into account in the finite-difference equation, which is equivalent to the following heat equation:

$$\rho \frac{\partial (cT')}{\partial t} = \frac{1}{r^2}\frac{\partial}{\partial r}\left(Kr^2 \frac{\partial T'}{\partial r}\right) + A - f\rho v \frac{\partial (cT')}{\partial r}\,, \quad (3)$$

where T' = adjusted temperature for latent heats if needed, r = radial distance, f = ratio of the mass of magma to the total mass, v = velocity of magma rising in a viscous medium, and ρ, c, t, K, and A have been defined in eq. (1). The heat equation is then solved by the Crank-Nicolson implicit method as described in ref. [4].

Thermal models for the Moon, Mars, and the mantle of Venus are assumed to consist of two components: 5% low-melting 'basalt' and 95% high-melting 'dunite'. The melting curves of 'basalt' and 'dunite'

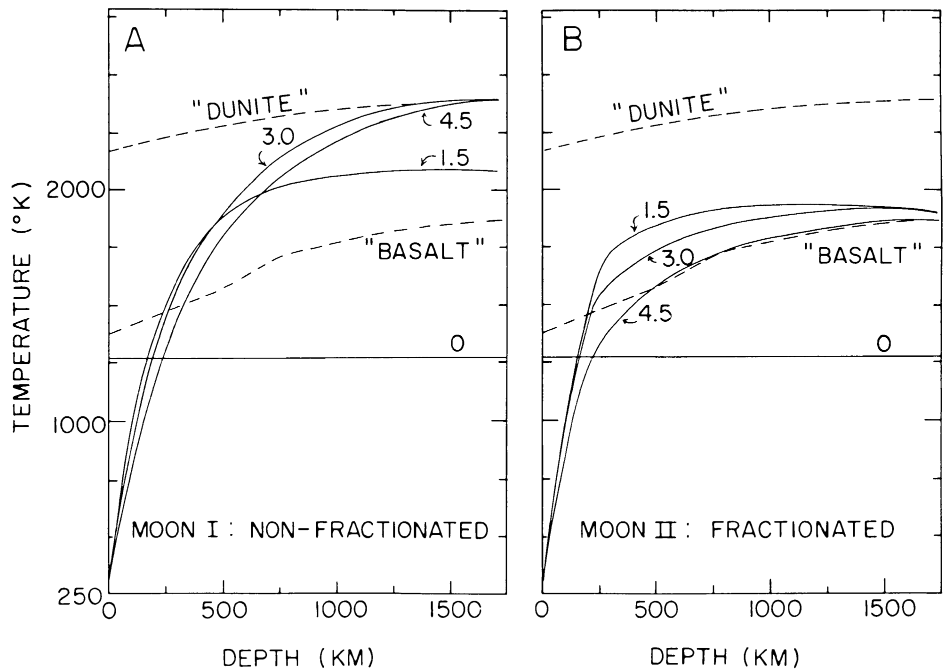

Fig. 1. Temperature developments in Moon I and Moon II at 0, 1.5, 3.0, and 4.5 billion years after the Moon was formed. Assumed melting curves for 'dunite' and 'basalt' are dashed. Data used for computation are identical for these two models except that fractionation of radioactive elements is allowed in Moon II but not in Moon I.

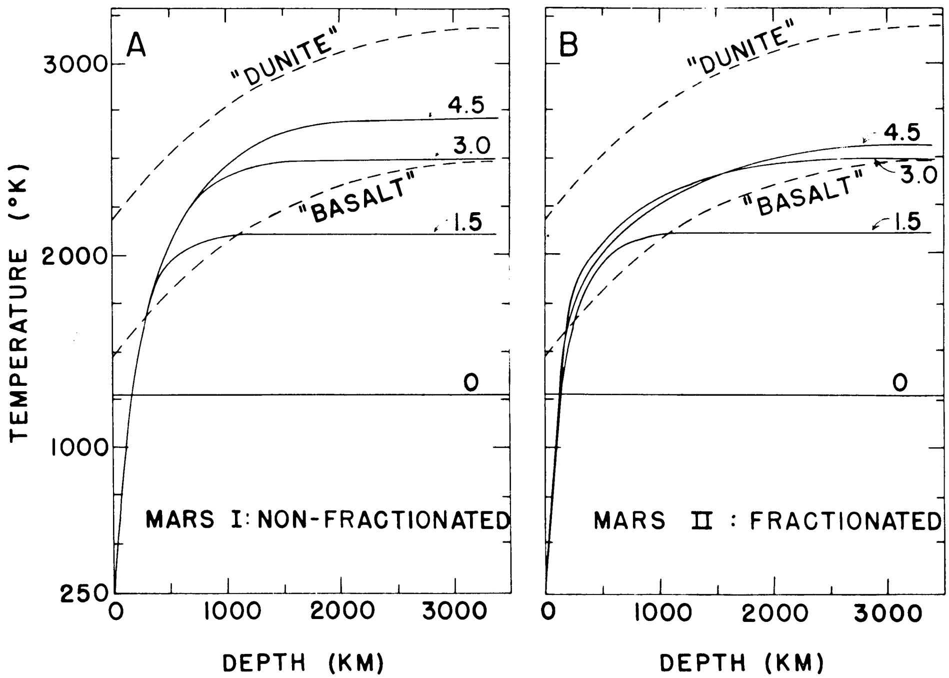

Fig. 2. Temperature developments in Mars I and Mars II at 0, 1.5, 3.0, and 4.5 billion years after Mars was formed. Assumed melting curves for 'dunite' and 'basalt' are dashed. Data used for computation are identical for these two models except that fractionation of radioactive elements is allowed in Mars II but not in Mars I.

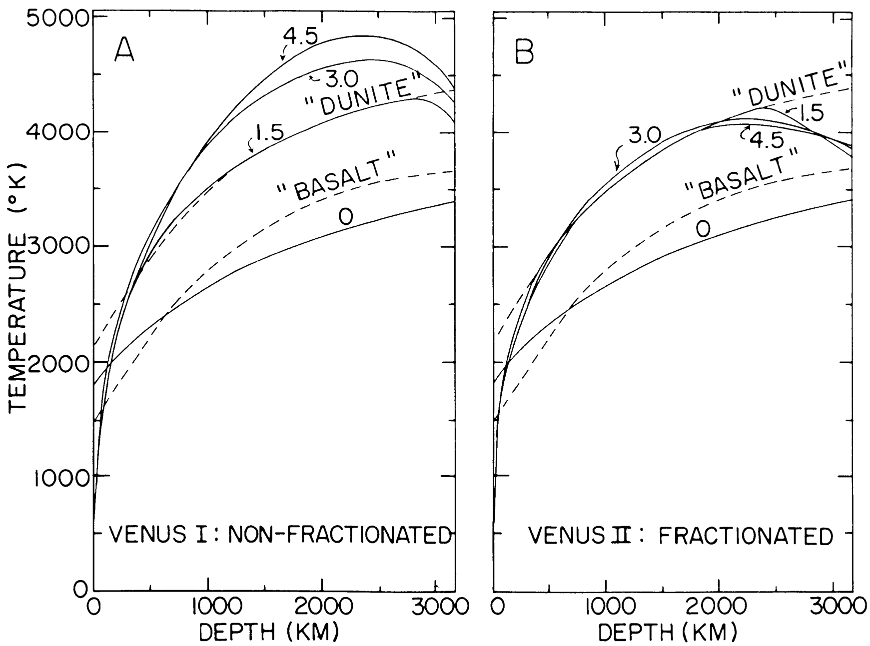

Fig. 3. Temperature developments in Venus I and Venus II at 0, 1.5, 3.0, and 4.5 billion years after Venus was formed. Assumed melting curves for 'dunite' and 'basalt' are dashed. Data used for computation are identical for these two models except that fractionation of radioactive elements is allowed in Venus II but not in Venus I.

Table 1
Common parameters in thermal-history calculations.

Age	4.5×10^9 years
Specific heat	1.2 J/g ^oK
Latent heat of fusion	400 J/g
Radiogenic heat generation (J/g yr)	^{238}U = 2.97
	^{235}U = 18.0
	^{232}Th = 0.82
	^{40}K = 0.94
Radioactive decay constant ($\times 10^{-10}$ yr^{-1})	^{238}U = 1.54
	^{235}U = 9.71
	^{232}Th = 0.499
	^{40}K = 5.5
Lattice conductivity	0.025 J/g sec ^oK
Index of refraction	1.7
Low-temperature opacity	10 cm^{-1}
Electrical conductivity at infinite temperature	10 ohm^{-1} cm^{-1}
Activation energy	3 eV

Table 2
Particular parameters in thermal-history calculations.

	Moon	Mars	Venus	
Radius (km)	1738	3400	6100	
Radius of core (km)	0	0	2900	
Density (g/cm^3)	3.34	4	table VIII of Jeffreys [3]	
Concentration of radioactive elements				
U ($\times 10^{-8}$ g/g)	1.17	1.17	1.94	in Venus's
Th ($\times 10^{-8}$ g/g)	3.65	3.65	6.06	mantle
K ($\times 10^{-4}$ g/g)	5.65	5.65	9.44	only
Surface temperature (^oK)	273	230	600	
Initial temperature (^oK)	1273	1273	fusion curve of 'iron'	

are taken from those measured for forsterite and basalt in the Moon models. They are extrapolated similar to those for the earth in the Mars and Venus models. These melting curves with depth are dashed in figs. 1, 2, and 3. Other data which enter into the thermal-history calculations are summarized in tables 1 and 2.

Table 1 shows parameters common to all thermal models. The ages for the Moon, Mars, and Venus are taken to be 4.5 billion years, the same as that for the earth. Typical values for specific heat, latent heat of fusion, radiogenic heat generations, radioactive decay constants, and thermal conductivity are adopted as shown in table 1. Radiative heat transfer is computed according to Clark [2].

Table 2 shows parameters which are appropriate for the thermal models of the Moon, Mars, and Venus. The radius and density are well established for the Moon, but are less certain for Mars and Venus because of the difficulty in measuring their solid discs. Models for the Moon and Mars do not have cores as suggested by their relatively low density. The core-mantle boundary for the Venus model and its density distribution are taken from Jeffreys [3]. The surface temperature for the Moon, Mars, and Venus are reasonably known, and typical values are adopted in their models as shown in table 2. For the Moon and Mars models, the radioactivity has been assumed to be that of Type I carbonaceous chondrites on a sample basis as determined by Morgan and Lovering [8]. For the Venus model, all radioactive elements are assumed to be in the mantle. Initial temperatures are uncertain, but we have assumed a uniform 1000^{o}C for the Moon and Mars models, and the melting curve of 'iron' for the Venus model.

Two thermal models each have been constructed for the Moon, Mars, and Venus. They are denoted as Models I (non-fractionated) and II (fractionated). Model I is the traditional model *without* selective fusion (but taking into account latent heats), whereas Model II is with selective fusion. Otherwise, Models I and II are identical.

3. RESULTS OF THERMAL-HISTORY CALCULATIONS

The purpose of the present calculations is to show the differences in the development of internal temperature distribution and surface heat-flow between thermal models including and excluding selective fusion of silicates. The Moon and the terrestrial planets are different in size and mass. The mean densities of the Moon, Mars, and Venus are 3.3, 4, and 5.1 g/cm^3, respectively, in comparison with 5.5 g/cm^3 for the earth. To a first approximation, we may infer that the Moon and Mars are like the earth's mantle, and that Venus has a core and is earth-like in structure. Thus the mean density of a planet suggests its internal structure upon which thermal models are constructed. The size of a planet greatly influences its thermal history because it affects the initial temperature distribution, melting temperatures, and thermal lag.

3.1. *The Moon*

The small size and low density of the Moon are well suited for modeling it after the earth's upper mantle. Assuming constant density and hydrostatic equilibrium, the maximum pressure reached within the Moon is about 46 kilobars. This low maximum pressure suggests that major phase transitions of silicates (such as the olivine-spinel transformation) are unlikely to occur within the Moon. The melting temperatures of silicates under the lunar pressure range have been measured. Consequently, interpretation of temperature with respect to melting or partial melting (which may play a significant role in redistribution of radioactive elements and thermal convection) is more definite for the Moon model than for planetary models.

In Moon I, no fractionation of radioactive elements is allowed. The temperature development is shown in fig. 1 (A). The deep interior of the Moon reaches the melting point of 'dunite' at 3.0 billion years, and is nearly molten at 4.5 billion years after the Moon was formed. The temperature development is very different if we allow for fractionation of radioactive elements, as shown in Moon II in fig. 1 (B). The temperature of the Moon in this model after 4.5 billion years (i.e., at present) reaches the melting temperature of 'basalt' at a depth of about 500 kilometers. The temperatures at 1.5 and 3.0 billion years, however, are higher than that at present. This suggests that magmas at depths of about 200 kilometers were present early in the Moon's history. This may explain Baldwin's suggestion [1] that the maria were formed by the flooding of pre-existing craters, and that in many craters a large time interval had elapsed between the formation of the crater and its filling by lava from below.

Surface heat-flow for both Moon models as a function of time is shown in fig. 4 (A). The heat-production flux of the Moon models is included for comparison. It is computed from the rate of radiogenic heat production divided by the Moon's surface area. Be-

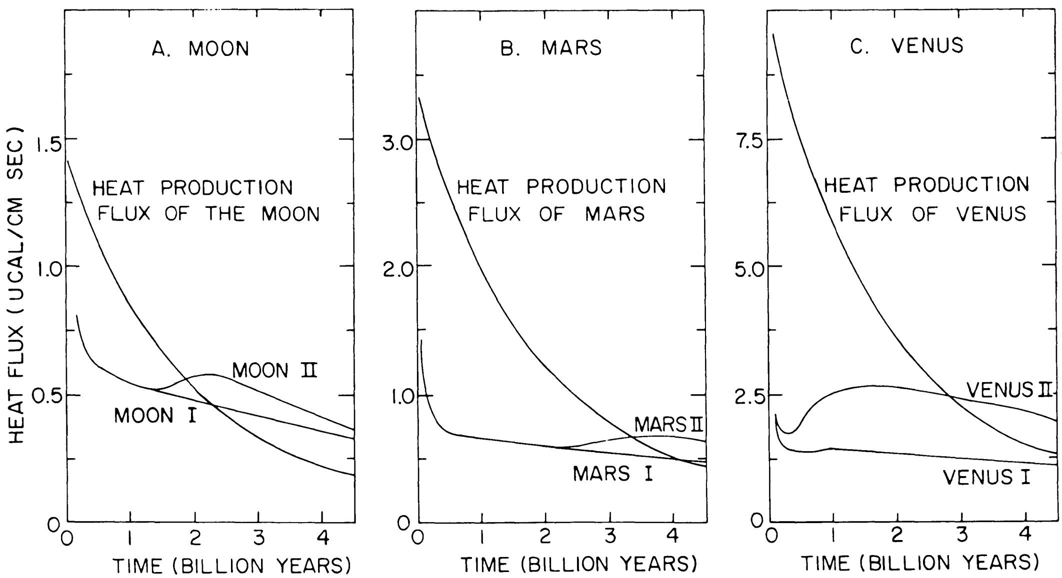

Fig. 4. Surface heat-flow for Moon I and Moon II, Mars I and Mars II, and Venus I and Venus II. Heat-production flux is computed from the rate of radiogenic heat production of a body divided by its surface area: A. the Moon, B. Mars, and C. Venus.

cause the Moon has a small ratio of mass to surface area, the heat-production flux is about 0.2 μcal/cm^{2}sec at present. Both Moon I and Moon II give a present surface heat-flow of about 0.3 μcal/cm^{2}sec, which is in agreement with Baldwin's estimate of 0.25 μcal/cm^{2}sec [1]. Fig. 4(A) also shows that the surface heat loss of the Moon models began to exceed the heat production at about 2 billion years after the Moon was formed. In other words, the Moon models suggest cooling of the Moon for the last 2.5 billion years, provided that its age is 4.5 billion years. This happens because the Moon is a small body so that its thermal lag is shorter than its age.

No definite Moon models can be made at present because constraints on its present temperature distribution and surface heat-flow are uncertain. Nevertheless, it is clear that the present temperature distribution of the Moon for a *fractionated* model is lower by some 600^{o}C than for a *non-fractionated* model. If we assume the initial temperature to be 0^{o}C instead of 1000^{o}C, then the subsequent temperature distributions do not reach the fusion curve of 'basalt' and thus no fractionation of radioactive elements takes place.

3.2. *Mars*

Assuming constant density and hydrostatic equilibrium, the maximum pressure reached within Mars is about 250 kilobars, which is equivalent to a depth of about 700 kilometers within the earth. Since a number of major phase transitions of silicates can occur within the pressure range of Mars, its internal structure may be complex like the earth. Discussion of melting or partial melting within Mars is further complicated by the necessity of extrapolation of melting curves.

The development of internal temperature for Mars I (non-fractionated) and Mars II (fractionated) is shown in fig. 2. Since fractionation is not very effective until after 3 billion years since the initial time, the present temperature distribution between these two Mars models does not differ much. The situation will be greatly different, however, if either the initial temperature is higher than the assumed 1000^{o}C or radioactivity is greater than the type I carbonaceous chondrites.

The development of surface heat-flow for both Mars models is shown in fig. 4 (B). The results suggest that cooling of these Mars models occurs only during the last billion years or less. The present surface heat-flow for Mars I is about 0.5 μcal/cm^{2}sec, and for Mars II about 0.65 μcal/cm^{2}sec. Since no surface heat-flow measurement (or even crude estimate) has been made for Mars, it is not feasible to test the models.

3.3. *Venus*

Although Venus is similar to the earth, there are two important differences:

(1) Venus appears to have a higher surface temperature of about 600^{o}K, and

(2) Venus has a lower melting point than the earth at corresponding depths for the same composition.

The first difference is due to Venus's thick atmosphere and its orbit being closer to the sun. The second difference is caused by Venus's having a lower internal pressure than the earth. This in turn is due to Venus's smaller size and lower density.

The radius of the Venus models is taken as 6100 km. Following Jeffreys [3], the Venus models have a core of radius 2900 km. Since radioactive elements do not seem to associate with the metal phase, they are all assumed to be in Venus's mantle.

The development of temperature distribution for Venus I (non-fractionated) and Venus II (fractionated) is shown in fig. 3. Venus I shows that the melting curve of 'dunite' has been exceeded at a depth of about 300 kilometers since 1.5 billion years after Venus was formed. On the other hand, Venus II shows that the present temperature reaches the melting point of 'dunite' only at depths between 1000 to 2000 kilometers. Extensive fractionation occurs in Venus II and this is consistent with the fact that Venus has a thick atmosphere. If seismic-velocity data were available for Venus, it would be easy to choose between these two Venus models.

The development of surface heat-flow for both Venus I and II is shown in fig. 4 (C). Venus I suggests that it is still heating up, whereas Venus II suggests that cooling has occurred since 1.5 billion years ago. If measurements of surface heat-flow were available for Venus, it would also be easy to choose between these two Venus models for they yield a present surface heat-flow which differs by a factor of two.

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