Internal Structure and Properties of Mars

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Theoretical physical models of the Martian interior are presented in the light of new and revised data and constraints. These models include thermal evolution, densities, and seismic wave velocities. The interior of Mars appears to be Earth-like in many respects. Although thermal models indicate that Mars has passed its peak of evolution it may still have an asthenosphere and may be moderately active tectonically. Mars has an Fe-FeS core with a radius of 1500-2000 km. The mantle is enriched in FeO with an olivine composition of about Fo72. Theoretically determined seismic wave velocities are relatively well constrained in the mantle with upper-mantle $P_s$ velocities ranging from 7.64 to 7.80 km/sec. However, there are wide variations in $V_p$ in the core dependent on composition. The shadow zone due to the core is larger than the Earth's.

In this paper we present a set of physical models of the Martian interior based on available data. Some of the results presented are extensions of work published in Johnston et al. (1974; hereafter referred to as Paper 1). However, improved theoretical techniques, revised constraints, and additional data that have become available since 1974 warrant another study of the structure and properties of the interior of Mars.

Interest in the structure of Mars has a long and involved history. Early studies of the thermal evolution include Urey (1951, 1952), MacDonald (1962), Kopal (1962), Lee (1968), Anderson and Phinney (1967), Hanks and Anderson (1969), and Reynolds and Summers (1969). More recent papers (Binder, 1969; Binder and Davis, 1973; Anderson, 1972; Ringwood and Clark, 1971; Toksoz and Johnston, 1974, 1976; Johnston et al., 1974) have calculated density and compositional models including a metallic core. The debate over whether or not Mars has a core still continues, with Strangway et al. (1976) postulating a planet with an undifferentiated deep interior.

In this paper, thermal models are briefly presented followed by density and seismic velocity models. The final section summarizes the main features of the Martian structure.

DATA ON THE MARTIAN INTERIOR

Theoretical calculations of planetary interior models require constraints based on observed data to limit the problem to realistic models. Although Mars is not as strongly constrained as the Earth or Moon, data primarily from Mariner 9 have enabled us to revise and improve earlier thermal and density models. Many of these data are concerned with the interpretation of Mariner 9 photographs which yield evidence of surface evolution history, volcanism, and differentiation of the planet (Masursky, 1973; Soderblom et al., 1974; Arvidson, 1974; Chapman, 1974; Jones, 1974; Malin, 1976; Murray et al., 1972; Mutch and Head, 1975). These types of
data are particularly useful in constraining thermal history models. A possible evolution history proposed by Masursky (1973) and its relation to thermal models is discussed in Paper I. It appears that Mars, like the Moon (probably the Earth, Venus, and Mercury also), suffered through a period of intense bombardment about 4 b.y. ago (Tera and Wasserburg, 1976) shortly after origin. This implies that a crust had formed prior to this time, requiring relatively high initial temperatures. The later stages of Martian evolution show episodes of tectonic and volcanic activity. Tharsis volcanism probably occurred about a billion years ago (Soderblom et al., 1974) and is a major constraint in the thermal evolution and temperature models of Mars.

Perhaps the most important data concerning the present structure of Mars are the mass, $M$, and moment-of-inertia factor, $C/Ma^2$. These roughly determine the density variation within the planet and indicate the presence of a high-density core. While the mass of Mars is well determined, the moment-of-inertia factor is not. The interpretation of the second zonal harmonic of the gravitational field ($J_2$) in terms of the hydrostatic and nonhydrostatic contributions depends critically on assumptions of crustal structure (Binder and Davis, 1975; Weir, 1975; Reasenberg, 1977). Binder and Davis (1973), assuming an isostatically compensated equatorial bulge of 8 km, estimate $C/Ma^2$ ranging from about 0.374 to 0.370.

Analysis of the gravity field and topography of Mars (Phillips and Saunders, 1975) indicates that the surface may be divided into two distinct groups: older crust isostatically compensated at a relatively shallow depth; and the Tharsis plateau, Chryse, and Amazonis lowlands, which are only partially compensated. The Tharsis plateau is the most prominent feature associated with an uncompensated gravity anomaly. These data suggest a description of Mars as a spheroid nearly in isostatic equilibrium plus an extra mass represented by the Tharsis region. Reasenberg (1977) uses this model to calculate the nonhydrostatic contribution of Tharsis to $J_2$. Thus, assuming that the rigidity supporting Tharsis could be relaxed, Reasenberg finds that the body, now in hydrostatic equilibrium, would have a moment-of-inertia factor of 0.3654 ± 0.001. This model is further justified in that the optical and dynamic flattening calculated are in much better agreement than those for other models. We therefore adopt for this paper the value of $C/Ma^2$ obtained by Reasenberg. This value is significantly lower than the 0.377 used in Paper I. It alters the density models presented there in such a way as to emphasize an increased density contrast between the shallow layers and deep interior of Mars, possibly implying a larger core radius. Other physical parameters for Mars which are used in the thermal and density calculations are presented in Table I. A detailed discussion of these parameters may be found in Paper I.

### Table I

**Parameters Used in Mars Models**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Radius (km)</td>
<td>3389</td>
</tr>
<tr>
<td>Mean density (g/cm$^3$)</td>
<td>3.90</td>
</tr>
<tr>
<td>$C/Ma^2$</td>
<td>0.3654</td>
</tr>
<tr>
<td>Heat of fusion (mantle) (J/g)</td>
<td>400</td>
</tr>
<tr>
<td>$C_p$ (mantle) (J/g°C)</td>
<td>1.2</td>
</tr>
<tr>
<td>Surface temperature (°C)</td>
<td>-20</td>
</tr>
<tr>
<td>$U$ concentration (ppb)</td>
<td>15</td>
</tr>
<tr>
<td>Th/U ratio</td>
<td>4</td>
</tr>
<tr>
<td>K/U ratio</td>
<td>50,000</td>
</tr>
</tbody>
</table>

**THERMAL MODELS**

The implied evolution history and presence of a core discussed in the preceding section place strong constraints on the thermal history of Mars. The presence of a weak magnetic field (Dolginov et al., 1972) also seems to imply that present-day internal temperatures should produce a
molten or partially molten core. Condensation models of the solar nebula (Grossman and Larimer, 1974) suggest that the Martian core is enriched in FeS, and thus may contain small amounts of the heat-producing radioactive isotope $^{40}\text{K}$ (Lewis, 1971; Goettel, 1972).

Theoretical conduction thermal models consistent with the known constraints concerning the evolution of Mars and core formation are presented in Paper 1. Hsui et al. (1976) have calculated thermal models for Mars including the effect of solid-state convection by simultaneously solving the equation of motion and the time-dependent energy equation in spherical geometry. Since the process of core formation is not well understood, its effect upon convective structure is difficult to determine. Thus, prior to core formation, the calculations are carried out in the manner described in Paper 1, with similar initial conditions and physical parameters (Table I). After the separation of the core is completed, the core temperatures are assumed to follow an adiabatic gradient determined by the temperature at the core–mantle boundary. As in Paper 1, the presence of $^{40}\text{K}$ in the core is allowed.

The evolutionary history of Mars from Hsui et al. is shown in Fig. 1. It is assumed that core formation takes place before 1.5 b.y. after origin. The planet continues to heat up slightly after core formation but then starts to cool. Large tensile features of fractured plains (Carr, 1974) may have been associated with the heating-up period. A partially molten region in the upper mantle exists up to 1 b.y. ago. The thick lithosphere, however, resists lateral motions explaining the absence of plate tectonics. The short dashed line shows the top of local partial melt associated with the upwelling of convective currents. If water is present near the Martian surface, the solidus will be suppressed and partially molten regions may exist at a depth of 250 km at the present time. This could produce some present-day tectonic activity. There is a region of solid-state convection at the lower part of the mantle. The core is completely molten.

The present-day temperature profiles from both the conduction and convection...
models are interpreted to be bounds on the temperature distribution within Mars. These are used in the determination of density models and are shown in Fig. 2. Also shown are the dry-mantle solidus (extrapolated peridotite; Green and Ringwood, 1969) and an extrapolated Fe-FeS eutectic melting curve calculated by Usselman (1975a,b). In that convection most likely occurs within planetary interiors, we used the convective temperature profile in the determination of density and seismic wave velocities.

**TABLE II**  
**Parameters Used in Density Models**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper mantle (Fo$_{50}$)</td>
<td></td>
</tr>
<tr>
<td>$K_a$ (mbar)</td>
<td>1.255</td>
</tr>
<tr>
<td>$dK/dT$ (kbar/°K)</td>
<td>0.174</td>
</tr>
<tr>
<td>$\alpha(T = 20^\circ C)$ (°C$^{-1}$)</td>
<td>$2.4 \times 10^{-5}$</td>
</tr>
<tr>
<td>Lower mantle (spinel Fo$_{50}$)</td>
<td></td>
</tr>
<tr>
<td>$K_a$ (mbar)</td>
<td>1.863</td>
</tr>
<tr>
<td>Core</td>
<td></td>
</tr>
<tr>
<td>$K_a$ (mbar)</td>
<td>1.2</td>
</tr>
<tr>
<td>$\alpha(T = 20^\circ C)$ (°C$^{-1}$)</td>
<td>$7.3 \times 10^{-5}$</td>
</tr>
</tbody>
</table>

**DENSITY MODELS**

Theoretical density models for Mars can be constructed using the mass and moment-of-inertia factor as constraints if a knowledge of the mantle and core compositions is assumed. These models are calculated for a warm, compressible, layered planet following the technique described in Paper 1 and in Solomon and Toksöz (1973). Since the core composition is not well determined, we present a range of possible density profiles.

The density models are derived by numerically integrating the equation of hydrostatic equilibrium where the density, $\rho$, is related to the pressure, $P$, by an equation of state, assumed to be the isothermal Birch-Murnaghan equation (Birch, 1938, 1952; Murnaghan, 1937):

$$P = \frac{3}{2} K_0 [\rho_0^{-7/2} - (\rho/\rho_0)^{7/2}]$$  (1)

$K_0$ and $\rho_0$ are the isothermal bulk modulus and density at $P = 0$ and are computed at the temperature, $T$, at depths given by the thermal models shown in Fig. 2. The thermal expansion coefficient may be temperature dependent and is based on extrap-
olated data from experimental measurements for the appropriate material. Values for these parameters are listed in Table II.

Each model assumes a 50-km crust with a density of 3.0 g/cm³ (Phillips et al., 1973) which amounts to about 3% of the total mass of Mars. The bulk modulus and thermal expansion used are typical of terrestrial crustal rocks (Clark, 1966, p. 94).

As a starting model for the mantle, we assume a composition of Fo₇₅. Values of $K₀$ and $dK/dT$ taken are given by Chung (1971). The use of Fo₇₅ as a mantle model was based on preliminary density calculations in order to roughly estimate the STP mantle density, and thus its equivalent olivine composition.

A phase change from olivine to spinel phase may occur where the temperature profile crosses the $P$–$T$ transition (Anderson, 1967a), as shown in Fig. 3. For the conduction thermal model, this occurs at a depth of about 1500 km and for the convection model at a depth of 1200 km. The standard temperature and pressure spinel-phase density is constrained to be 10% higher than the STP olivine-phase density. The elastic parameters for this layer are again taken from Chung (1971) and are listed in Table II. Other possible mantle phase changes would have smaller effects and are not considered in this paper.

The core composition (and therefore density) is considered to be a free parameter to be determined by the calculation. Core compositions considered range from Fe₀.₈S to FeS. An FeS composition has been proposed by Lewis (1972) for the Martian core, and Fe₀.₈S is similar to compositions proposed for the Earth (Usselman, 1975b; King and Ahrens, 1973).

Usselman (1975a) has shown from experimental determinations of the Fe–FeS eutectic temperature that a solid–solid phase transformation occurs in FeS at about 52 kbar. This is consistent with the high-pressure phase (hpp) evident from shock compression studies of pyrrhotite reported by King and Ahrens (1973). Assuming no further transitions, this is the phase present in the Martian core. Recent experimental work suggests a value of $K₀$ for the hpp of 1.2 mbar and a zero-pressure density increase of 15% over the low-

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**Fig. 3. Present-day temperature profiles from the conduction thermal model (solid line) and the convection model (dashed line) superimposed on the stability fields of olivine and spinel phase.**
pressure phase (Ahrens, 1976; Pichulo et al., 1976).

Thermal expansion in the core is temperature dependent and is taken to be an extrapolation of values reported for γ iron (Basinski et al., 1955), as discussed by Siegfried and Solomon (1974). It is assumed that the melting of the core material has little effect at elevated pressures on the physical parameters used in the density calculations.

Density models are calculated for core compositions of FeS, FeS$_2$, Fe$_7$S$_8$, and FeS for both conductive and convective temperature profiles and fixed mean density of the planet. In Fig. 4, models of varying core compositions are plotted as a function of $C/Na^2$, STP mantle density, and core depth. From these one may select a density model that best fits the moment-of-inertia factor. Valid density models satisfying both $C/Na^2$ and the mean density are obtainable in every case with core radii varying from 1500 to 2050 km and STP mantle densities ranging from 3.47 to 3.58 g cm$^{-3}$. The Martian mantle is more dense than the Earth's upper mantle, implying an enrichment in FeO as predicted by condensation models of the solar nebula (Lewis, 1972). However, such models that predict an FeS core in Mars along with extreme enrichment in FeO of the mantle appear to be inconsistent with the data. FeO (FeO + MgO) values obtained from the density models range from 0.19 for an FeS core to 0.27 for an Fe$_7$S$_8$ core. Lewis (1972), on the other hand, predicts a value of 0.5. It is almost inescapable that some free iron is present in the Martian core.

Further variations in density and core radius may be obtained from Fig. 4 for different assumed values of $C/Na^2$. 

Fig. 4. Standard temperature and pressure upper-mantle density shown as a function of moment of inertia and depth to the core for varying core compositions (lower right corner) and for both temperature models. The horizontal dashed line is the moment-of-inertia factor used in this paper.
Density profiles for valid models satisfying $C/Ma^2 = 0.3654$ are shown in Figs. 5 and 6 for the conduction and convection temperatures, respectively. Comparison of these figures shows the range of density variations due to extreme models of temperature and composition considered. We prefer models based on convective profiles (Fig. 6) since they represent more realistic thermal processes. Several features stand out. First, the change in core density is reflected primarily in core radius, with relatively small changes in the STP mantle density. This is important because it allows us to set rather specific bounds on the mantle composition and seismic velocities that will be discussed later. Second, close examination of these figures reveals that in the upper mantle, thermal expansion is dominant over compression resulting in a slight density inversion of about 2% at a depth of about 250–300 km. This is the
The same region where a small partial pressure of water could induce local partial melting, as discussed earlier.

**VELOCITY AND TRAVEL TIME MODELS**

Seismic wave velocity as a function of depth in the Martian mantle may be estimated using seismic equations of state empirically determined for terrestrial rocks. Most of these have velocities dependent on the mean atomic weight, $\bar{m}$, and the density, $\rho$.

The most common seismic equation of state is Birch's law (Birch, 1966), which is of the form

$$ V = \alpha(\bar{m}) + b \rho. \quad (2) $$

Simmons (1964) proposed a generalized form of the equation for $P$ waves that will be used in this paper:

$$ V_P = -0.98 + 0.7(21 - \bar{m}) $$

$$ + 4.00[\text{CaO} - 2.76\rho] \quad (3) $$

reflecting the fact that Ca-rich rocks have a higher $V$ than other rocks with similar $\bar{m}$.

Wave velocities in the core are determined from the calculated bulk modulus based on composition and density models. Since the core is molten (the shear modulus, $\mu = 0$),

$$ V_{core} = (K \rho)^{1/2}. \quad (4) $$

**TABLE III**

**Chemical Analyses for Mantle Models (Weight Percent)**

<table>
<thead>
<tr>
<th>Oxide</th>
<th>Pyrolite</th>
<th>Mars</th>
<th>Fe$_3$ core</th>
<th>FeS core</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO$_2$</td>
<td>45.16</td>
<td>39.03</td>
<td>42.32</td>
<td></td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>3.54</td>
<td>3.06</td>
<td>3.32</td>
<td></td>
</tr>
<tr>
<td>FeO</td>
<td>8.04</td>
<td>20.53</td>
<td>13.82</td>
<td></td>
</tr>
<tr>
<td>MgO</td>
<td>37.47</td>
<td>32.38</td>
<td>55.11</td>
<td></td>
</tr>
<tr>
<td>CaO</td>
<td>3.08</td>
<td>2.66</td>
<td>2.89</td>
<td></td>
</tr>
<tr>
<td>K$_2$O</td>
<td>0.13</td>
<td>0.11</td>
<td>0.12</td>
<td></td>
</tr>
<tr>
<td>Fe$_3$O$_4$</td>
<td>0.46</td>
<td>0.40</td>
<td>0.43</td>
<td></td>
</tr>
</tbody>
</table>

Mean atomic weight 22.48 22.12

The mean atomic weight of the mantle may be found from a consideration of its oxide composition as discussed in Paper 1. We assume that the mantle may be modeled as a terrestrial pyrolite (Ringwood, 1966) with an STP density of 3.35 g cm$^{-3}$ with enough FeO added as wustite ($\rho = 5.745$ g cm$^{-3}$) to obtain a density equivalent to the STP upper-mantle density obtained from the theoretical density models. Given the resulting composition, the mean atomic weight may then be determined for use in (3).

We will compute seismic velocity models using the convective temperature profile and two compositions based on Fe$_3$S and FeS cores. These represent two extremes in core radius and density. Table III lists the upper-mantle oxide compositions and mean atomic weights for those models. The lower mantle is assumed to have an $\bar{m}$ equal to that of the upper mantle. Furthermore, we propose a two-layered crust with an upper layer of 20 km and $V_P = 6$ km/sec and a 30-km lower layer with $V_P = 7$ km/sec, although such an assumption is not necessary for the models.

Shear wave velocities, $V_s$, may be found by adopting a reasonable value for Poisson's ratio, $\sigma = f(V_P, V_s)$, or by using Anderson's (1967b) equation of state using the seismic parameter $\phi = (21\rho \bar{m})^2$. Given $V_P$ and $\phi$, $V_s$ may then be determined. However, the use of this equation of state does not necessarily give independent results for the shear modulus (Anderson, 1976). Thus $V_s$ is not strongly constrained.

The range for compressional and shear wave velocities calculated in the manner described above, bounded by core compositions of Fe$_3$S and FeS, are shown in Fig. 7. Some important characteristics of velocity models are readily apparent. First, the variation of possible velocities in the Martian mantle is relatively small because of the insensitivity of mantle density to varying core density. While mantle densities are lower for the FeS core compared to
Fig. 7. Range of seismic wave velocities as a function of depth for both P and S waves (shaded areas), calculated using the densities shown in Fig. 6. The range is bounded by models with core compositions of FeS (dashed line) and Fe3S2 (solid line).

For the Fe3S2 core, the mean atomic weight is also lower and thus the velocities are higher. Second, a large velocity contrast exists at the core-mantle boundary and this produces a seismic shadow zone which is more prominent than that which exists in the Earth. Upper-mantle (\(P_s\)) velocities range from 7.64 to 7.80 km/sec. A low-velocity zone associated with the density inversion occurs at about 250 km. Velocity increases at the base of the mantle are due to the olivine-spinel phase change. In reality this boundary may be smeared out and less sharp than shown on the model.

Poisson's ratio generally varies from about 0.30 at the base of the crust to 0.35 near the core-mantle boundary. The high value reflects the enrichment of FeO in the Martian mantle.

To demonstrate the effects of these models on seismic wave propagation, travel times were calculated for the FeS core composition. \(P\)-wave ray paths for a surface source (Marsquake or meteorite impact) shown in Fig. 8. Travel time versus distance for P waves is shown in the same figure. In the mantle, seismic ray propagation and the travel-time curve are well behaved, except for a reverse branch due to the velocity increase at the bottom. There is no shadow zone from the upper-mantle low-velocity zone due to the compensating effect of curvature, i.e., \(r/V\) is monotonically decreasing with depth. However, relatively lower amplitudes might be expected from 15 to 45°. The most striking feature of these models, of course, is the shadow zone due to the core. For an FeS core the range is from 90 to 152°, while for an Fe3S2 core this zone extends from 115 to 170°. The shadow zone for Mars is larger than the
Earth’s due to a greater velocity contrast between the mantle and core.

**SUMMARY AND CONCLUSIONS**

We have presented a suite of physically realistic models of the interior of Mars that satisfy the available constraints. What has emerged is a picture of the Martian interior that is similar to the Earth’s in many respects. Some inferences that can be made on the basis of these models are:

1. Thermal history models indicate that Mars has passed its peak of evolution. However, it has an asthenosphere (about 250 km deep) and, most likely, a convecting lower mantle. The planet may still be moderately active tectonically.

2. Mars has a core with a radius of about 1500–2000 km with a composition on the Fe-rich side of the Fe FeS zero-pressure eutectic composition. More sulfur-rich cores do not satisfy the constraint of the moment of inertia while simultaneously providing an FeO-rich mantle. The presence of a molten core is consistent with an internally generated magnetic field, however, the low observed field strength may be due to either stratification within the core or less conductive material due to a higher sulfur content relative to the Earth’s core. Alternate models of magnetic field generation infer distinctly different models (Strangway et al., 1976).

3. The mantle of Mars is enriched in FeO relative to that of the Earth and corresponds to an olivine composition of about Fo75. The mantle is essentially homogeneous with the olivine-to-spinel phase transition occurring at depths ranging from 1200 to 1500 km dependent primarily on the present-day temperature profile. The upper mantle has a density inversion due to the dominance of thermal expansion over compression. This temperature increase could produce an asthenosphere at a depth of about 250 km. Local partial melting could occur in this zone if small amounts of water exist in the upper-mantle system.

4. Theoretical seismic wave velocity models based on the density models exhibit narrow bounds on the mantle velocities but wide variations in the core dependent on composition. Typical upper-mantle $P_s$ velocities range from 7.64 to 7.80 km/sec. A mantle low-velocity zone does not produce a shadow zone. The shadow zone for the core, however, extends from about 100 to 160°.

*Note added in proof.* Inorganic analyses of surface samples from Viking sites 1 and 2 have been reported by Clark et al. (1976) and Baird et al. (1976). The samples are characterized by high iron, moderate magnesium, calcium, and sulfur, and low aluminum abundances. A comparison of these values with the mantle models proposed in this paper (Table III) shows reasonable agreement, especially for the enriched Fe. A simple ascending differentiation of the mantle material will result in a substantial depletion of MgO (Maderazzo, 1977: Carmichael et al., 1974), giving a surface olivine composition of nearly Fo75.

Although sulfur is not included in our mantle model, the whole planet is enriched in sulfur relative to Earth abundances when the core is included. Density considerations preclude an 6 $10^3$, S03 abundance, as observed in the surface samples for the mantle.

The relatively high concentration of sulfur may be due to a process proposed by Huguenin (1976) in which sulfates and other salts formed by photochemical weathering precipitate from a brine and are enriched in a thin surface layer.

**ACKNOWLEDGMENTS**

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