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THE COMPOSITION OF THE TERRESTRIAL PLANETS

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ABSTRACT

The densities and figures of the Moon and the terrestrial planets are important inputs into the highly theoretical problem of the origin and evolution of the solar system and the highly practical problem of computing detailed trajectories of planetary probes. Feedback from trajectory data has important implication in the solution of both of these problems. The Mariner IV Space Probe supplied several pieces of information that reopen the question of the internal constitution of Mars.

The new mass and radius combine to give a better estimate of the density of Mars and, hence, its composition. Mars is a particularly critical object in discussions of the origin of the terrestrial planets since we also know its ellipticity and can therefore discuss the internal distribution of density. The absence of a magnetic field verified our suspicions concerning its lack of a metallic core and the tenuous atmosphere and nature of the surface supplied complimentary information regarding the lack of chemical differentiation in the interior of this planet.

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INTRODUCTION

Kovach and Anderson (1965)[1] showed that the Earth, Mars, and Venus could be identical in composition if the larger estimates of the densities of these latter two bodies were accepted. In order to satisfy both the density and moment of inertia for Mars they were forced to undifferentiate the planet, i.e. mix the mantle and core material together. Mars, then, represented a primitive, undifferentiated body with iron mixed uniformly in the mantle. Anderson and Phinney (1966)[2] showed that thermal history calculations were consistent with this view of Mars. A homogeneous body with the mass of Mars, terrestrial abundances of radioactive elements, and an initial adiabatic thermal gradient will not exceed the melting point of iron in 4.5×10^9 years while an identical body, the mass of the Earth will exceed the melting point of iron early in its history and will differentiate a metallic core; the accompanying release of gravitational energy will lead to differentiation of the silicates and volatives giving a crust and atmosphere. This basic conclusion is strengthened if the iron in Mars is oxidized rather than elemental.

The new determinations of the density of Mars and the preferred value for the density of Venus are both less than appropriate for a planet having the same composition as the Earth.

Birch (1961)[3] and Anderson (1966)[4] showed that the mean atomic weight, $\overline{M}$, is an appropriate measure of composition in discussions of equations of state of silicates and this also follows from theoretical considerations such as Thomas-Fermi-Dirac (MacDonald and Knopoff, 1958[5]). Accordingly, we will discuss the composition of the terrestrial planets in terms of this parameter.

Acknowledgment is made to Robert L. Kovach who assisted in some of the calculations which are reported in this paper.

DENSITY DISTRIBUTION OF THE EARTH

The P-V relations for the mantle and core are shown in Figure 1; for comparison are shown theoretical and experimental equations of state for iron. The central pressures of the Moon, Mars, and Venus are also indicated. The density versus pressure curve for the mantle is not monatomic because of the phase changes experienced by silicates at pressures of the order of hundreds of kilobars. A complete collapse of silicates to close packed cubic oxides involves a density increase of some 20%. By suitably combining the mantle and core equations of state it is possible to construct planets of arbitrary size and mean atomic weight. It is necessary, of course, to assign a mean atomic weight to the mantle and core. Ultrasonic and shock wave measurements on silicates and metals supply the raw data for this calibration.

Figure 2 gives density versus the parameter Φ for oxides, silicates and elements of various mean atomic weights. The parameter Φ is defined as

$$\Phi = \alpha^2 - \frac{4}{3}\beta^2 \quad (1)$$

$$= \frac{\partial P}{\partial \rho} = \frac{K}{\rho} \quad (2)$$

where α, β, P, ρ, and K are, respectively, compressional velocity, shear velocity, pressure, density and bulk modulus. It is available from ultrasonic, static compression, and shock wave data and is also known throughout the Earth. From data of this type $\overline{M}$ for the mantle is estimated as 22.4 ± 0.5. Birch (1961) estimated the mean atomic weight of the mantle to be about 22.5 ± 0.5. The mean atomic weight of the core lies between 45.7 and 53.2.

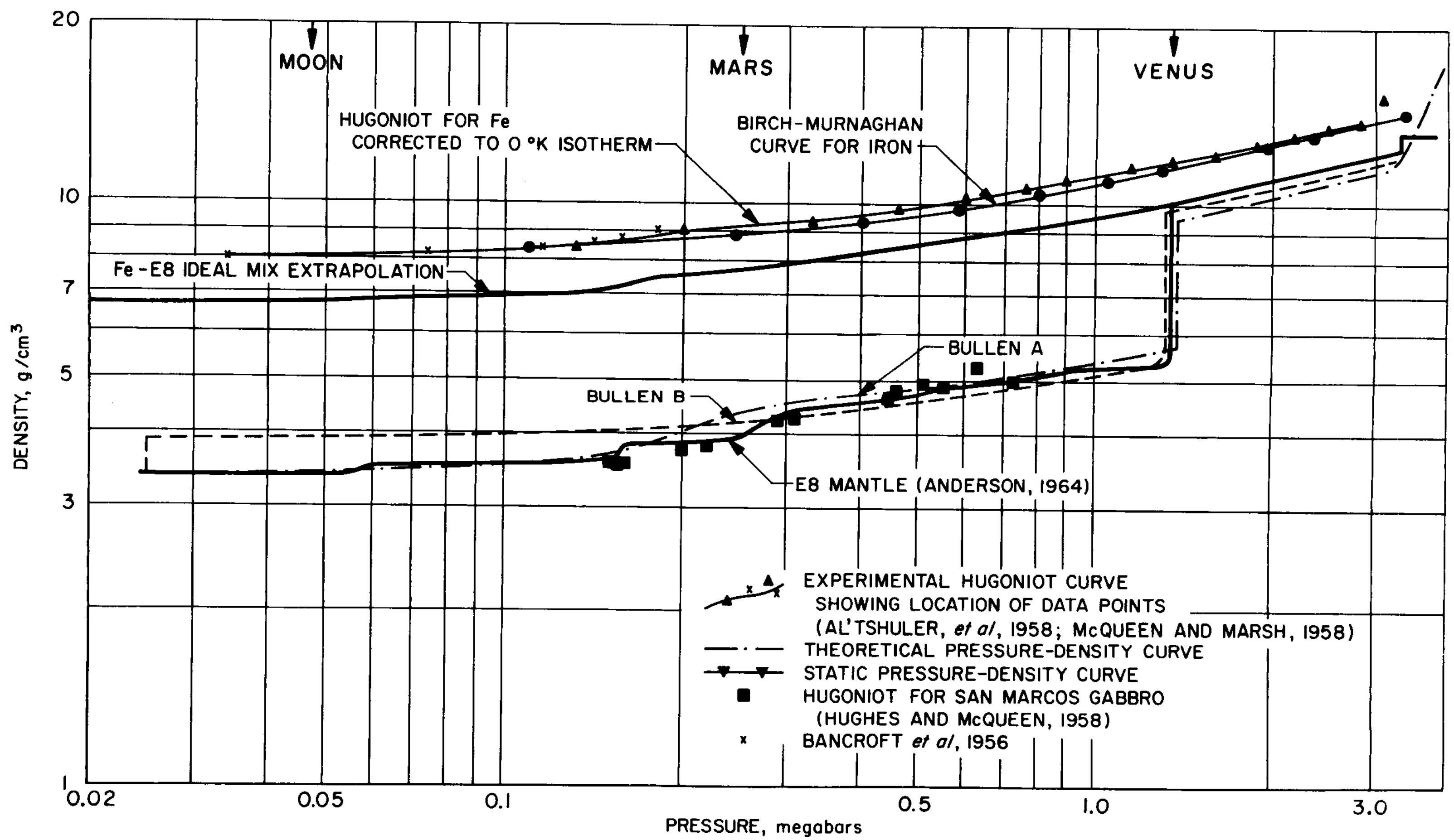

Figure 1. P-V Relationship for the Mantle and Core of the Moon, Mars, and Venus. (Kovach and Anderson, 1965)

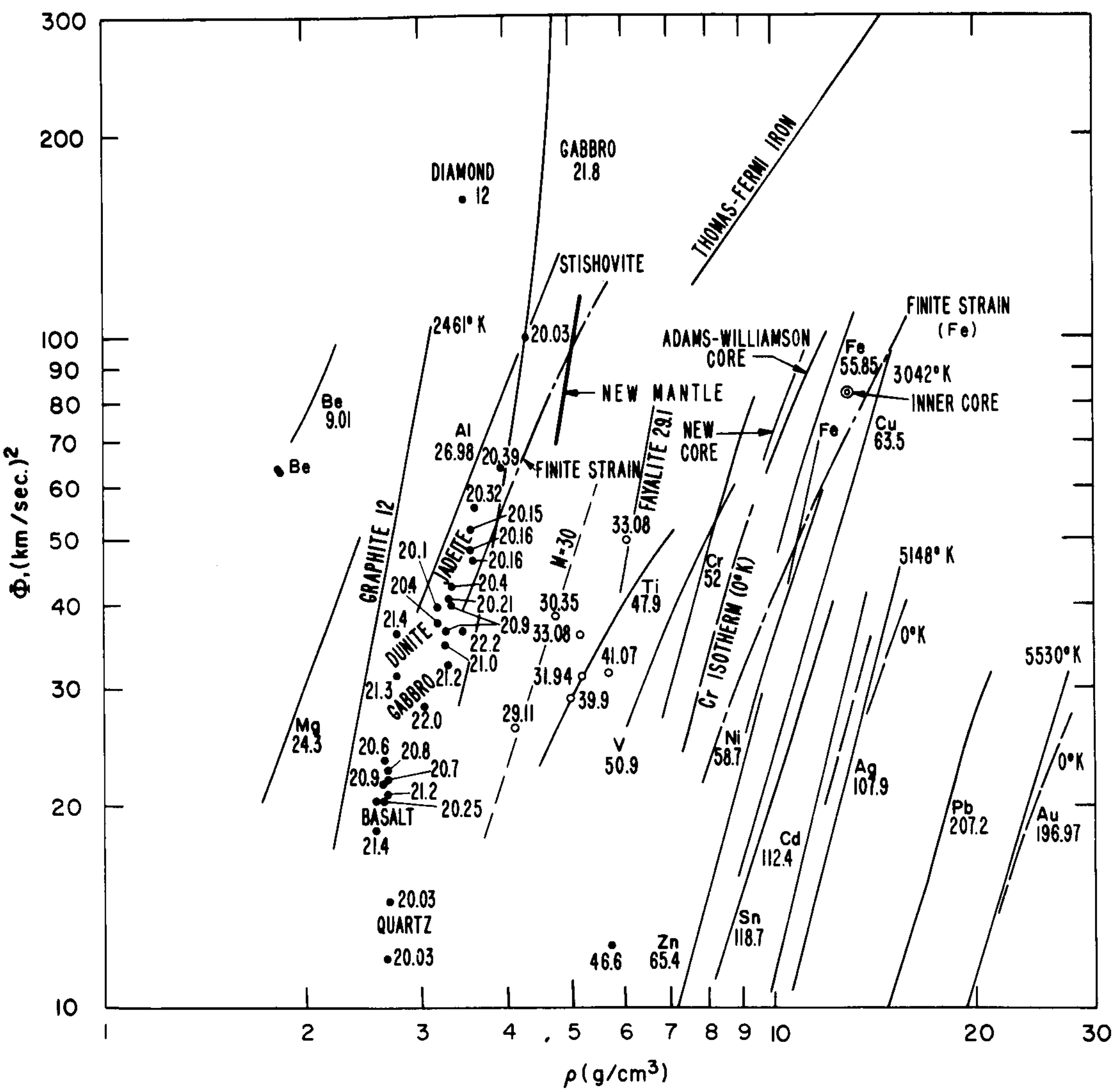

Figure 2. Density-Parameter Φ Relationship for Oxides, Silicates, and Elements of Various Mean Atomic Weights.

In this work we will take the mean atomic weight of the mantle and core to be 22.4 and 47.0 respectively. These numbers will undoubtedly require revision as more shock wave data becomes available and as our knowledge regarding the density in the Earth improves, but our conclusions regarding the difference in mean atomic weight of the terrestrial planets depends only on the density-mass relationships between these bodies.

APPLICATION TO THE PLANETS

Figure 1 in Anderson and Phinney (1966)[2], (hereafter referred to as Paper I) gives density versus mass (relative to Earth) for planets of various mean atomic weights constructed according to the scheme outlined in the previous section and the assumption of hydrostaticity. These planetary models are homogeneous, i.e. density varies with depth due to self-compression and solid-solid phase changes alone. There is no compositional stratification. The core of the Earth is about 32.5% of the total mass and this yields

$$\bar{M} = \left[(0.325/47.0) + (0.675/22.4)\right]^{-1} = 27.0 \qquad (3)$$

as the mean atomic weight of the whole Earth. The new measurements of the mass and radius of Mars (Kliore et al, 1966[6], Dollfus, 1966[7]) give a density between about 3.96 and 4.10. Simple interpolation gives $\bar{M} = 25.3 \pm 0.4$ for the mean atomic weight of Mars, i.e. 1 to 2 units less than the Earth. The error bars associated with the densities of the planets in Figure 1, Paper I, indicate the spread of measured radii as summarized in Kovach and Anderson (1965)[1]. The preferred density for Venus is about 5.1 g/cm^3 which gives a mean atomic weight of about 25.5. This estimate is for a homogeneous Venus. If it is differentiated, as implied by the thermal history calculations in Paper I, this estimate will be raised by about 0.1. The density of the Moon, 3.34 g/cm^3, gives $\bar{M} \sim 22$.

The variation in mean atomic weight can be attributed to:

1) differences in the iron content-- commonly expressed in terms of the Fe/Si ratio,

2) differences in degree of oxidation of the iron,

3) differences in the amount of retained volatiles.

The thermal history calculations in Paper I would be consistent with any of these interpretations except, perhaps, with the retention of volatiles in the necessary amount by the Moon. Chemical arguments might be expected to resolve the ambiguity except that, on the basis of such arguments Urey favors alternate 1) and Ringwood favors alternate 2).

Since the material in the Earth's core is less dense than pure iron, we will designate it as "core material" and discuss hypothesis 1) in terms of this quantity, rather than iron, not implying that this material is necessarily concentrated at the center of the planet. The weight fraction of core material in the Earth is about 0.325 or 32.5%. The mean atomic weight of Mars if attributed to a deficit of core material, implies a content of 21%. A similar calculation for Venus gives about 25% core material. In the case of Mars, the ellipticity requires that this material be essentially evenly distributed throughtout its mantle, although a small central core cannot be ruled out on the basis of astronomical data. The thermal history calculations suggest that there is no central molten region. The mean atomic weight of the Moon, combined with the thermal history calculations indicates that it is very similar in composition to the Earth's mantle.

The variation in the mean atomic weight among Mars, Venus and the Earth can also be attributed to differences in oxidation state. This is the view taken by Ringwood (1966).[8] If this view is correct Mars is the most oxidized and the Earth is the most reduced of the three planets with Venus being intermediate. The Moon does not fit into this scheme and it almost certainly is deficient in iron.

The mean atomic weight is only a very crude measure of the composition of a material. Most common iron free silicates and oxides have roughly the same mean atomic weight. See, for example, Table I.

TABLE I

Mean Atomic Weight of Selected Minerals and Rocks

Periclase	MgO	20.2
Forsterite	Mg_2SiO_4	20.1
Quartz	SiO_2	20.03
Corundum	Al_2O_3	20.4
Granite		20.9
Basalt		20.7
Dunite		20.9
Gabbro		21.8
Fayalite	Fe_2SiO_4	29.1
Hematite	Fe_2O_3	31.9
Magnetite	Fe_3O_4	33.08

Of the common elements that combine with oxygen to form the oxides and silicates, Mg, Si, Al and Fe, only iron has a significantly different atomic weight to be distinguished from the others when combined with oxygen and it is only iron bearing rocks that have mean atomic weights greater than 21. Mg, Si, and Fe are by far the most abundant metals in terrestrial and meteoritic material.

Table II summarizes the mean atomic weights of various rocks and meteorites. In these calculations we have considered only SiO_2, MgO, FeO, Al_2O_3, CaO, Fe, Ni, FeS, and H_2O. These usually account for more than 98% of the mass of the material.

TABLE II

MEAN ATOMIC WEIGHTS OF VARIOUS ROCKS AND METEORITES

Carbonaceous Chondrites	Mean Atomic Weight
Type I minus H_20	24.0
Type II minus H_20	23.4
Type III	23.7
"Ordinary" chondrites	24.4
Silicate phase, chondrites	21.8
Chondrites, "high iron"	25.1
Chondrites, "low iron"	23.9
Enstatite chondrites	25.6
Reduced Carbonaceous Chondrite I	21.9

This table represents the range of values found for materials which may be representative of planetary interiors. It indicates the effect of iron content, water content and degree of oxidation. All of the entries in this table represent material that has been differentiated, depleted or enriched to some extent in certain compounds and none may be representative of "primordial" material. The mean atomic weight and the ratio of elemental iron compounds (Fe0, FeS) increases with total iron content. Note that common crustal and upper mantle rocks are deficient in iron when compared with meteorites and the implied composition of the lower mantle.

CONCLUSION

Enstatite chondrites and high iron (H) chondrites have approximately the same mean atomic weight as Mars and Venus. The Earth has a higher mean atomic weight than any common group of stoney meteorites. The Moon and the mantle of the Earth have approximately the same mean atomic weight as the silicate phase of chondrites. The only conclusion that is warranted is that the range of compositions found in meteorites is adequate to explain the range in mean atomic weight of the terrestrial planets, i.e. Mars, Earth, Moon, and Venus.

It is possible to account for the differences in mean atomic weights by differences in total iron content as discussed previously or by differences in the Free iron/total iron ratio.

REFERENCES

1. Kovach, R. L. and Anderson, Don L.: The Interiors of the Terrestrial Planets. J. Geophys. Res., 70, 1965, pp. 2873-2882.
2. Anderson, Don L. and Phinney, R. A.: Early Thermal History of the Terrestrial Planets. "TYCHO" Paper, 1966, TG # 21.
3. Birch, F.: Composition of the Earth's Mantle. Geophys. J., 4, 1961, pp. 295-311.
4. Anderson, Don L.: Latest Information from Seismic Observations. H. Gaskell, Ed., The Earth's Mantle, Academic Press, New York-London, (in press).
5. MacDonald, G. J. F. and Knopoff, L.: On the Chemical Composition of the Outer Core. Geophys. J., 1, 1958, pp. 284-297.
6. Kliore, A., Cain, D. L., and Levy, G. S.: Radio Occultation Measurement of the Martian Atmosphere over Two Regions by the Mariner IV Space Probe. Seventh Intern. Space Science Symp. Comm. on Space Research, Vienna, Austria, 1966.
7. Dollfus, A.: Personal Communication, 1966.
8. Ringwood, A. E.: Chemical Evolution of the Terrestrial Planets. Geochim. et Cosmochim. Acta, 30, 1966, pp. 41-104.