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Internal Constitution of Mars[1]

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Models for the internal structure of Mars that are consistent with its mass, radius, and moment of inertia have been constructed. Mars cannot be homogeneous but must have a core, the size of which depends on its density and, therefore, on its composition. A meteorite model for Mars implies an Fe-S-Ni core (12% by mass of the planet) and an Fe- or FeO-rich mantle with a zero-pressure density of approximately 3.54 g/cm^3. Mars has an iron content of 25 wt %, which is significantly less than the iron content of the earth, Mercury, or Venus but is close to the total iron content of ordinary and carbonaceous chondrites. A satisfactory model for Mars can be obtained by exposing ordinary chondrites to relatively modest temperatures. Core formation will start when temperatures exceed the eutectic temperature in the system Fe-FeS (~990°C) but will not go to completion unless temperatures exceed the liquidus throughout most of the planet. No high-temperature reduction stage is required. The size and density of the core and the density of the mantle indicate that approximately 63% of the potential core-forming material (Fe-S-Ni) has entered the core. Therefore, Mars, in contrast to the earth, is an incompletely differentiated planet, and its core is substantially richer in sulfur than the earth's core. The thermal energy associated with core formation in Mars is negligible. The absence of a magnetic field can be explained by lack of lunar precessional torques and by the small size and high resistivity of the Martian core.

Most calculations of the internal structure of Mars have assumed that the core of the planet, if there is any, has the density of iron or the density of the earth's core suitably corrected for the effects of pressure. On the basis of this assumption and the observed moment of inertia, it has been concluded that Mars is a nearly homogeneous body with, at most, an iron core of 1–6% of the planet's mass [*MacDonald*, 1962; *Binder*, 1969]. *Urey* [1957] and *Bullen* [1966] concluded that it is unlikely that Mars has a core. Thermal-history calculations [*Anderson and Phinney*, 1967; *Hanks and Anderson*, 1969] have been used to support the arguments against the development of an extensive molten-iron core in Mars, as has the lack of a detectable magnetic field. However, solar and meteoritic abundances suggest that iron is not the only candidate for the material in a planetary core. Cosmic abundances, for example, and melting relationships indicate that sulfur should be an abundant element in the core, particularly in small relatively cold planets.

The two measurable parameters that are pertinent to the internal structure of Mars are the mean density and the moment of inertia. These data can be used to determine only two parameters of the interior. The behavior of solids, such as silicates and metals, under pressure is

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fairly well known from high-pressure and shock-wave studies and from the known structure of the interior of the earth. For example, we know the equations of state and the locations of phase changes for most of the materials that might be expected to be important in the interiors of the terrestrial planets. With this information, we can completely define the structure of a two-zone planet (for example, one containing a mantle and a core) in terms of the zero-pressure densities of the mantle and the core and the radius of the core. For Mars two of these parameters can be found as a function of the third. This procedure is to be preferred to preassigning one of the parameters, since such preassignment would be equivalent to assuming that the composition of one of the regions of the planet is known. It is particularly dangerous to assign the density of the core, since the relative proportions of iron, nickel, sulfur, and silicon, elements that may be in the core in appreciable abundances, depend critically on conditions during accretion of the planet, present internal temperatures, degree of differentiation, and, of course, the composition and oxidization state of the original material. If silicon is the light alloying element in the earth's core [*Ringwood*, 1966], the density of the core will decrease as it grows; silicates are reduced only in the later high-temperature stages of accretion. Mars, being a smaller body than the earth, would have less silicon in its core, and the core would have a larger zero-pressure density. On the other hand, if sulfur is the light alloying element in the core [*Anderson et al.*, 1971], the density of the core will increase as it grows and becomes richer in iron because of the nature of the Fe-FeS phase diagram [*Anderson*, 1971]. Mars would thus be expected to have a core less dense than the earth. This effect would be compounded by the greater efficiency of Mars in retaining sulfur, by the smaller accretional energies involved in its growth, and by the inferred depletion of iron in Mars relative to the earth. In any case, a Martian core is unlikely to be of pure iron or of the same composition as the earth's core.

Accordingly, we have constructed a suite of planets that satisfy the observable properties for Mars and that are independent of compositional assumptions.

CALCULATIONS

For present purposes, we take an extremely simple form for the equation of state:

$$\rho = \rho_0 + a[1 - (r/R)]$$

where ρ_0 is the zero-pressure density of a given region of the planet, R is the radius of Mars, ρ is the density at radius r, and a is a constant taken as 0.565 g/cm^3. This equation gives results consistent with those of *Kovach and Anderson* [1965] and *Binder* [1969], who used a different approach. The mantle of Mars is presumably composed mainly of silicates, which can be expected to undergo one or two major phase changes, each involving a 10% increase in density. To a good approximation, these phase changes will occur at ⅓ and ⅔ of the radius of Mars. The deeper phase change will not occur if the radius of the core exceeds ⅓ of the radius of the planet. With these parameters we can solve for the radius and density of the core, given the density of the mantle and the observable mass, radius, and moment of inertia for Mars. The results are given in Table 1.

The curve in Figure 1 gives these results in terms of the density of the core and the radius of the core. The curve is the locus of possible Mars models. Clearly, the data can accommodate a small dense core or a large light core. The upper limit to the density of the core is probably close to the density of iron. This density value provides a lower limit to the radius of the core of 0.36 of the radius of Mars, or about 8% of the mass of the planet. To determine a lower limit to the density, one must consider possible major components of the core. Of the potential core-forming materials, Fe, S, and Ni are by far the most abundant elements, both in me-

TABLE 1. Parameters for Mars Models

R_c/R	ρ_m	ρ_c	M_c/M
0.20	3.557	20.1	*
0.25	3.554	12.6	*
0.30	3.550	9.22	0.06
0.33	3.548	7.93	0.07
0.40	3.541	6.30	0.10
0.50	3.522	5.27	0.17
0.60	3.492	4.80	0.26

*Not computed.

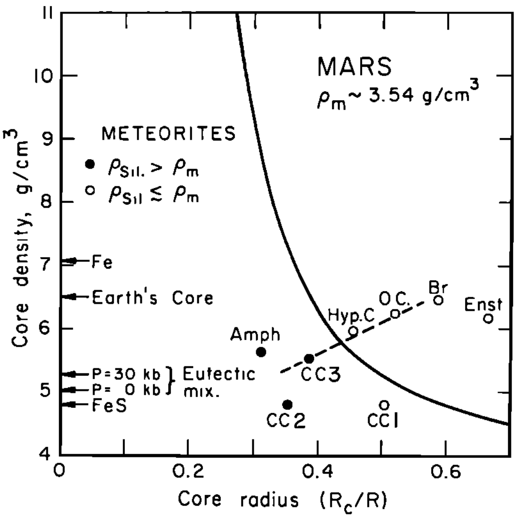

Fig. 1. Radius of the core versus the density of the core for planetary models that have the mass, radius, and moment of inertia of Mars. The density of the mantle varies along the solid curve (Table 1). The densities of the Fe-FeS eutectic compositions are determined from data of *Brett and Bell* [1969]. The points are for various meteorite classes with all the FeS and free Fe and Ni differentiated in a core. The dashed line shows how core density is related to core size in the Fe-FeS system. The density and amount of melt increase as the temperature is raised above the eutectic. The level of the line is adjusted to ordinary chondritic abundances. The slope is calculated from the phase diagram.

teorites and in the sun. In ordinary high-iron chondrites, the free iron content averages 17.2% by weight. The FeS content is approximately 5.4% (3.4% Fe, 2.0% S), and the Ni content is 1.6%. A planet assembled from such material, if completely differentiated, would yield a core of 24% of the mass of the planet, with Fe:S:Ni in the approximate proportions of 21:2:2 by weight. Low-iron chondrites would yield a core of 15% of the mass of the planet, with proportions of 12:1:2.

Carbonaceous chondrites have little or no free iron but contain 7–25% by weight FeS and about 1.5% Ni. The average core size for a planet made of carbonaceous chondrites would be 15% by mass, Fe:S:Ni being in the proportions 9:5:1.5. An absolute minimum core density can probably be taken as 4.8 g/cm^3. This value corresponds to a pure FeS core with a fractional core radius of 0.6 and a fractional mass of 26%. On these grounds, the mass of the Martian core can be considered to lie between 8 and 24% of the mass of the planet.

This range can be narrowed considerably by further consideration of the compositions of meteorites. The points in Figure 1 represent most of the major categories of stony meteorites. The size and density of the 'core' have been computed from the amounts of iron, sulfur, and nickel in the meteorite. No single class of meteorites, fully differentiated into core and mantle, would satisfy the data for Mars, although carbonaceous chondrites and hypersthene (low iron) chondrites come close. The open circles indicate silicate (mantle) densities less than the inferred density for the mantle of Mars; the closed circles indicate silicate densities that are too high. The meteorites above the curve can be migrated downward and to the left by placing some of the iron of the core in the mantle and thereby increasing the density of the mantle and decreasing the density and radius of the core. Physically, the result would correspond to a meteorite model that has been only partially differentiated, i.e., has undergone an incomplete separation of mantle and core. If chondrites are an appropriate guide to the composition of Mars, possible core sizes would be further restricted to 12–15% by mass. Alternatively, the closed circles could be migrated to the locus of possible Mars models by reducing some of the FeO in the silicate phase and allowing the iron to enter the core. This procedure is much more drastic and requires high temperatures and, probably, the presence of carbon to effect the reduction. It is interesting that the meteorites in question do contain substantial amounts of carbon. The resulting cores would be about the same size as was previously inferred.

A third possibility would be to assemble Mars from a mixture of meteorites that fall above and below the curve. The meteorites below the curve, however, are relatively rare, although the earth may not be collecting a representative sample.

The first alternative seems particularly attractive because of its simplicity and because it follows naturally from the phase relations in the Fe-FeS system for the compositions found in ordinary and enstatite chondrites. That is to say, the iron content of these meteorites is on the iron-rich side of the eutectic composition, and partial melting would produce a sulfur-rich melt and leave iron behind in the mantle.

Many other alternatives are possible. The mantle need not be homogeneous; all parts of the mantle need not contribute equally to the core. Some parts of the mantle may be more effectively stripped of their iron than other parts. The composition of the melt depends on temperature and pressure. Unfortunately, the data restrict us to a discuussion of two parameter models. In addition, the phase diagram for the Fe-FeS system is not known at high pressures. High-pressure phases will certainly intervene, but it is unlikely that the qualitative arguments given above will be invalidated.

DISCUSSION

Phase relations in the system Fe-FeS have been worked out by *Hansen and Anderko* [1958] and by *Brett and Bell* [1969]. The eutectic temperature is 990°C and is relatively insensitive to pressure, at least to 30 kb. The eutectic composition shifts from about 69 wt % Fe at zero pressure to 74 wt % Fe at 30 kb. The eutectic phase diagram for the Fe-FeS system is shown schematically in Figure 2. Several points should be noted. First, if sulfur is the light alloying element in the earth's core, estimates of its density and temperature would place it well above the liquidus temperature in this system. This result would suggest that the separation of core and mantle is probably well advanced or complete in the earth. Second, for compositions to the left of the eutectic composition the melt and presumably the core become progressively richer in iron and denser as temperature increases and melting proceeds. The contrary is true for FeS-rich systems, such as carbonaceous chondrites. For ordinary and enstatite chondrites partial melting accompanied by complete separation of melt from the solid would result in an iron-rich mantle and a sulfur-rich core. An estimate of the conditions in the Martian core are shown in the hatched region of Figure 2.

Figure 3 shows temperature profiles in Mars as a function of time from thermal-history calculations [*Hanks and Anderson,* 1969]. Also shown are estimates of the iron melting curve

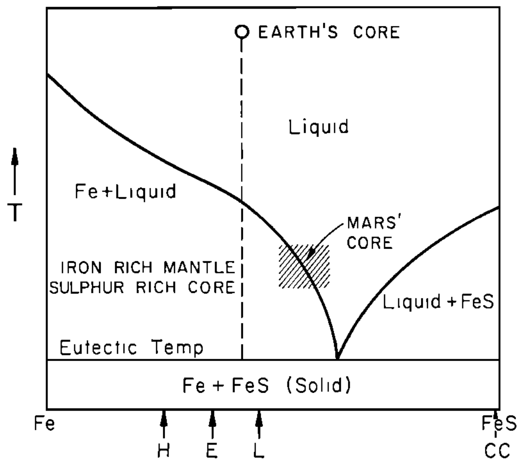

Fig. 2. Schematic diagram for the system Fe-FeS, adapted from *Brett and Bell* [1969] and *Hansen and Anderko* [1958]. The compositions of high-iron (H), low-iron (L), enstatite (E), and carbonaceous chondrites (CC) are shown along the bottom. These compositions are calculated from the free Fe and FeS contents of these meteorites. The shape of the diagram and the temperature scale depend on pressure. The melting temperatures of the end members (Fe and FeS) increase with temperature. The eutectic temperature is fairly independent of temperature, but the eutectic composition migrates to the left (i.e., becomes richer in iron) as pressure increases.

and the eutectic and liquidus curves. For the model much of the interior of Mars is between the eutectic and liquidus temperatures. This condition suggests partial rather than total melting and incomplete separation of potential core-forming material. Preliminary calculations on the thermal effect of core formation indicates that this source of energy is negligible for Mars.

We have argued above, qualitatively, that partial melting of certain classes of meteorites will move them toward the locus of Mars models. The phase diagram for the iron-sulfur system allows us to compute, at least at low pressures, the relation between the relative amount of melt and its composition or density. The dashed line in Figure 1 shows how these two quantities are related for a certain starting composition close to that of ordinary chondrites. At the intersection of the curves the mass of the core is 12% of the mass of the planet and the density of the core is 5.85 g/cm^3. An average ordinary chondrite contains 11.7% Fe, 1.3% Ni, and 5.9% FeS, or 19% potential core-

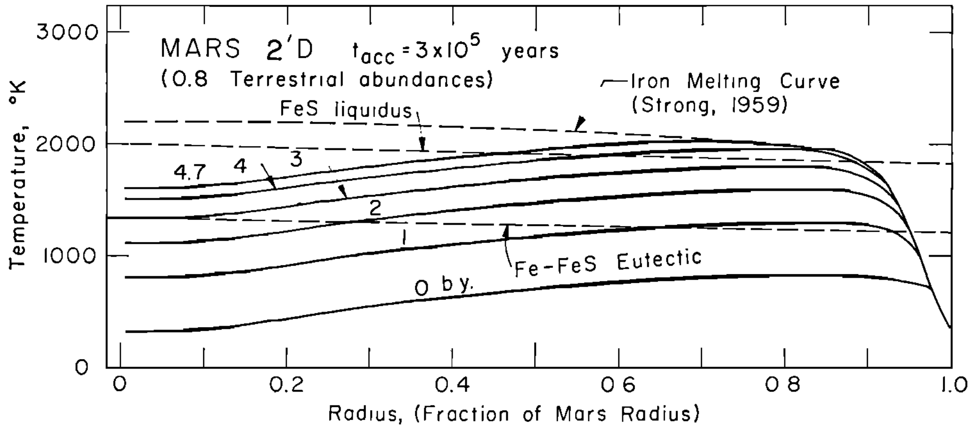

Fig. 3. Temperatures in Mars as a function of radius and time (adapted from *Hanks and Anderson* [1969]). Also shown are estimates of the melting temperature of pure iron and the eutectic and liquidus temperatures in the Fe-FeS system. The thermal effects of convection, latent heat, and core formation have been ignored.

forming material. From the phase diagram and the assumption that the Ni/Fe ratio stays constant, we determine that the core consists of 5.7% Fe, 0.6% Ni, and 5.7% S, relative to the mass of the planet. Thus, the total core is 12% of the mass of the planet. Most of the original sulfur has entered the core, but considerable amounts of Fe and Ni remain solid and in the mantle. The density of the silicate phase plus the residual iron and nickel yields a mantle density of 3.54 g/cm^3, which agrees well with the required value (Table 1). Thus, it appears that a satisfactory model can be obtained for Mars by melting 63% of the potential core-forming material in an ordinary chondrite. This situation would hold if internal temperatures in Mars averaged about 1300°–1600°C. Bronzite or high-iron chondrites provide mantles that are slightly too dense (3.62 g/cm^3) when migrated back to the Mars locus by the above procedures.

An alternative meteorite model for Mars that would accommodate higher temperatures and complete separation of core and mantle can be constructed. For example, if type III carbonaceous chondrites are mixed with ordinary chondrites to give the correct density, the core will be 12% by weight and will have a density of 5.78 g/cm^3. These values are close to the values required to satisfy the mass and moment of inertia of Mars. The high density of the mantle in this example is a result of the high FeO content of type III carbonaceous chondrites. Type II carbonaceous chondrites and amphoterites (Soko-Banjites) also have high FeO contents.

One interesting result of this model is that the total iron content of Mars is almost independent of the size of the core or assumptions on how the iron is distributed. For example, the total inferred iron content for Mars varies from 25 to 28% for models in which the mass of the core ranges from 10 to 26%. This range of core sizes just about covers the range of core densities from pure iron to pure FeS. This range of values for Mars can be compared with the iron content of the earth (about 35%) and verifies previous conclusions that Mars is less rich in iron than the earth.

Table 2 gives the composition of the mantle and core of Mars for the two meteorite models just discussed. Model 1 is the meteorite mix model, with 75% type III carbonaceous chondrites and 25% ordinary chondrites. Complete segregation of the core from the mantle is assumed. Model 2 is the partially differentiated ordinary chondrite model, with 63.5% of the potential core-forming material melted and settled to the core. Both models contain 25% iron, and both have a core of about 12% of the mass of the planet.

The Orgueil type I carbonaceous chondrite has been considered by some authors to be the best available sample of primordial material.

TABLE 2. Mars Models

Element	Model 1*	Model 2†
	Mantle	
SiO_2	34.9	38.3
MgO	23.9	23.9
FeO	21.2	12.0
Al_2O_3	2.7	2.7
CaO	2.2	1.9
Na_2O	0.7	0.9
K_2O	0.06	0.1
Cr_2O_3	0.5	0.4
MnO	0.2	0.3
TiO_2	0.1	0.1
P_2O_5	0.3	0.2
NiO	0.2	0.0
H_2O	0.8	0.3
Fe		6.0
Ni		0.7
FeS		0.2
Total	87.8	88.0
	Core	
Fe	4.6	5.7
Ni	1.1	0.6
FeS	6.1	5.7
Total	11.8	12.0

*Mix; 75% type 3 carbonaceous chondrites, 25% ordinary chondrites.

†Partially differentiated ordinary chondrite; ρ_c = 5.85, ρ_m = 3.54.

This meteorite is extremely rich in such low-temperature condensates as H_2O and FeS, as well as carbon and 'organic matter.' If we ignore the water and carbon compounds, a fully differentiated planet of this composition would have a core of 15% by mass, composed mainly of FeS (13.6% FeS, 1.4 Ni), and a mantle with a density of about 3.5 g/cm^3. However, this meteorite also contains about 19% H_2O, most of which must have escaped if Mars is to be made up primarily of this material. Otherwise, the mantle would not be dense enough.

Figure 4 gives several possible Mars models in terms of density as a function of fractional radius. These models are consistent with the mean density and moment of inertia. The range of possible solutions can be considerably reduced, as mentioned before, if chondritic compositions are used as an additional constraint.

MAGNETIC FIELD

The presence of a substantial core rich in molten iron, implied by previous sections, and the observed rapid rotation of Mars provide two of the conditions that are thought necessary for the generation of a magnetic field. One

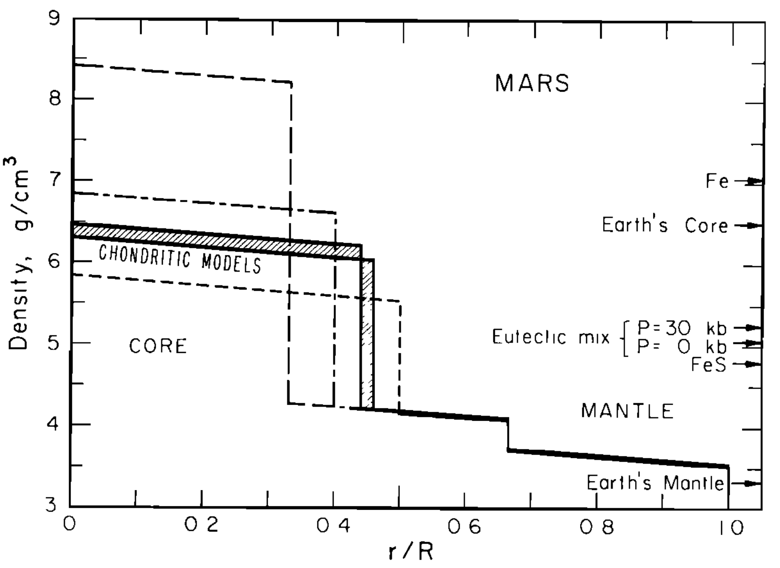

Fig. 4. Density versus radius for some representative Mars models that have the proper mean density and moment of inertia. Models with chondritic compositions fall in the hatched region. Note that all models have approximately the same mantle density.

might, therefore, expect Mars to have a magnetic field. However, a differential precessional torque acting on the core and mantle is probably the ultimate driving mechanism for the geomagnetic dynamo [*Malkus,* 1968]. The moon and the sun are much less effective in generating such a torque for Mars because of the distance of Mars from the sun and Mars' lack of a substantial moon.

The small size of the Martian core and, if sulfur is abundant, its high resistivity would lower the magnetic Reynolds number. These effects alone could eliminate the possibility of regenerative dynamo action. Other factors that must be considered are the core's viscosity, which is highly dependent on temperature and pressure, and the shielding effect of an iron-rich mantle. All things considered, it seems unlikely that Mars ever had a substantial magnetic field of internal origin.

CONCLUSIONS

The present astronomical data for Mars require that it have a dense core.

The size of the core and its density can be traded off. By using the density of pure iron and the density of pure troilite (FeS) as reasonable upper and lower bounds for the density of the core, its radius can be considered to lie between 0.36 and 0.60 of the radius of the planet.

The zero-pressure density of the mantle is confined to the range 3.54–3.49 g/cm^3. This range implies an FeO content of 21–24 wt % unless some free iron has been retained by the mantle.

If chondrites are an appropriate guide to the major element composition of Mars, the core of Mars is smaller and less dense than the core of the earth, and the mantle of Mars is denser than that of the earth.

Mars contains 25–28% iron, independent of assumptions about the over-all composition or distribution of the iron. The earth is clearly enriched in iron when compared with Mars or most classes of chondritic meteorites.

Meteorite models for Mars that contain 25 wt % iron and 12 wt % core have been constructed.

If sulfur is retained in Mars, the presence of a core is not inconsistent with the low temperatures inferred from thermal-history calculations.

The absence of a magnetic field can be explained by the lack of significant lunar torques. The presence of large amounts of sulfur in the core and its small size, as compared with the earth's core, also serve to suppress the importance of dynamo action in the Martian core.

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REFERENCES

Anderson, D. L., Sulfur in the core: Implications for the earth and Mars, *Comments Earth Sci. Geophys., 1,* 133–137, 1971.

Anderson, D. L., and R. A. Phinney, Early thermal history of the terrestrial planets, in *Mantles of the Earth and Terrestrial Planets,* edited by S. K. Runcorn, pp. 113–126, Interscience, New York, 1967.

Anderson, D. L., C. Sammis, and T. Jordan, Composition and evolution of the mantle and core, *Science, 171,* 1103–1112, 1971.

Binder, A. B., The internal structure of Mars, *J. Geophys. Res., 74,* 3110–3118, 1969.

Brett, R., and P. Bell, Melting relations in the Fe-rich portion of the system Fe-FeS at 30 kb pressure, *Earth Planet. Sci. Lett., 6,* 479–482, 1969.

Bullen, K. E., On the constitution of Mars III, *Mon. Notic. Roy. Astron. Soc., 133,* 229–238, 1966.

Hanks, T. C., and D. L. Anderson, The early thermal history of the earth, *Phys. Earth Planet. Interiors, 2,* 19–29, 1969.

Hansen, M., and K. Anderko, *Constitution of the Binary Alloys,* 2nd ed., pp. 1–720, McGraw-Hill, New York, 1958.

Kovach, R. L., and D. L. Anderson, The interiors of the terrestrial planets, *J. Geophys. Res., 70,* 2873–2882, 1965.

MacDonald, G. J. F., On the internal constitution of the inner planets, *J. Geophys. Res., 67,* 2945–2974, 1962.

Malkus, W. V. R., Precession of the earth as the cause of geomagnetism, *Science, 160,* 259, 1968.

Ringwood, A. E., Chemical evolution of the terrestrial planets, *Geochim. Cosmochim. Acta, 30,* 41–104, 1966.

Strong, H. M., The experimental fusion curve of iron to 96,000 atmospheres, *J. Geophys. Res., 64,* 653–660, 1959.

Urey, H. C., *Progress in Physics and Chemistry of the Earth,* vol. 2, p. 46, Pergamon, New York, 1957.

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