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SOME ASPECTS OF CORE FORMATION IN MERCURY

Sean C. Solomon
Department of Earth and Planetary Sciences
Massachusetts Institute of Technology
Cambridge, Massachusetts 02139

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ABSTRACT

Evidence for a large metallic core in Mercury is all indirect: the internal magnetic field may imply a convective dynamo; the surface geology is suggestive of large-scale differentiation; and thermal history calculations based on cosmochemical models for Mercury predict core formation. The presence of a core will not be confirmed by an accurate measure of J_2, probably two orders of magnitude larger than the hydrostatic value.

Core infall on Mercury would be accompanied by an increase in planetary radius of 13 km, an increase in the mean internal temperature of 700 K, and substantial melting of the mantle. If Mercury's core differentiated from an originally homogeneous planet, such an event must predate most of the present surface. Subsequent to core formation, Mercury's radius has decreased by about 2 km due primarily to cooling of the lithosphere, in agreement with the estimate by Strom and others of the amount of planetary contraction based on photogeologic measurements of length, dip and throw on the global system of lobate scarps.

A convective dynamo mechanism for Mercury's magnetic field is in apparent conflict with cosmochemical models that do not predict a substantial source of heat, most probably radiogenic, in Mercury's core. Without such a heat source, the core would solidify within about 1 b.y. after core infall, producing an unacceptably large contraction in Mercury's radius.

INTRODUCTION

Of fundamental importance to discussions of internal structure and tectonic evolution of a terrestrial planet is the question of a dense central core. The size of the high density core in the earth, in Mars and in the moon is a major distinguishing feature of their relative structures, and presumably of their formation histories. The redistribution of gravitational and thermal energy during core formation can have a profound effect on the surface of the planet.

Because of the detection of an apparently intrinsic magnetic field in the planet Mercury (Ness et al., 1974; 1975a,b) and in light of several deductions about Mercury's history made after study of Mariner 10 photographs of the planet's surface (Murray et al., 1974a, b, 1975; Trask and Guest, 1975; Strom et al., 1975), several questions dealing with the nature and history of any large metallic core in Mercury need to be asked. Does Mercury have a core? If so, how big is it? Is it fluid or solid? What was the time scale for core formation? What would be the geologic consequences of core formation, and are these consistent with surface geology? This paper offers some of the answers to these questions.

IS THERE A CORE?

If a metallic core has been completely differentiated from a silicate mantle in Mercury, then the size of such a core may be determined. Necessary for such a calculation are the mass and radius of the planet, both sufficiently well known (Smith et al., 1970; Ash et al., 1971; Howard et al., 1974), a nominal temperature profile, and assumed equations of state for both the mantle and core. Siegfried and Solomon (1974) estimated that a fully differentiated core in Mercury would have a radius equal to 75 percent of the planetary radius and a mass equal to 66 percent of Mercury's total mass. These figures are weakly sensitive to the assumed temperature distribution, to the equation of state parameters, and to the adopted values for the zero pressure densities of the mantle and core, the latter reflecting the amount of nickel and any lighter elements alloyed with iron. As a result, the iron abundance in Mercury, about 60 weight percent in the models of Siegfried and Solomon (1974), has a probable uncertainty of 6 to 10 weight percent (Reynolds and Summers, 1969; Kozlovskaya, 1969).

In the absence of seismological information about a planetary interior, the only unequivocal evidence for a planetary core is a dimensionless moment of inertia C/MR^2 substantially less than 0.40. The moment of inertia cannot be observed directly, however. For the earth and the moon,

C/MR^2 is obtained from a combination of second-degree coefficients in the spherical harmonic expansion of the gravitational potential and relative differences among principal moments obtained from an analysis of the response of the planet's motion to external torques. For Mars, C/MR^2 is generally obtained from J_2, the second-degree zonal coefficient in the expansion of the gravity field, using the hydrostatic theory of Clairaut and Radau (e.g., Binder, 1969). The non-hydrostatic shape of Mars makes the assumption of hydrostatic equilibrium questionable, and modifications of the assumption lead to reduced values for C/MR^2 and larger cores (Binder and Davis, 1973).

There is some promise of extracting a reliable value of J_2 for Mercury from the three Mariner 10 encounters (Howard et al., 1974; Esposito et al., 1975). Can the measurement of J_2 be used to estimate C/MR^2 using hydrostatic theory? Almost certainly it cannot.

Because Mercury rotates only very slowly, the nonhydrostatic contributions to the second degree coefficients in the planet's gravitational potential are much larger than the hydrostatic contribution. An illustration of this point is given in Table 1. For the earth, moon and Mars, J_2 is well known and the hydrostatic contribution J_{2H} can be calculated either from the known value of C/MR^2 and hydrostatic theory, as for the earth (Khan, 1969) and the moon (Jeffreys, 1970), or from a physical model for the nonhydrostatic contribution, as for Mars (Binder and Davis, 1973). The nonhydrostatic fraction, J_{2NH}, of J_2 is simply $J_2 - J_{2H}$. A rough estimate of J_{2NH} for Mercury may be obtained using Kaula's (1966) equal stress hypothesis, according to which nonhydrostatic contributions to spherical harmonic coefficients of the gravitational potential should scale as the inverse square of surface gravity g. By that hypothesis, J_{2NH} for Mercury should be between 10^{-5} and 10^{-4}, roughly two orders of magnitude greater than J_{2H} for hydrostatic models of Mercury, either with or without cores (Siegfried and Solomon, 1974). Preliminary estimates of J_2 for Mercury (Esposito et al., 1975; R.D. Reasenberg, personal communication, 1975) fall approximately in the range of J_{2NH} predicted in Table 1.

Thus even after J_2 for Mercury is determined to great accuracy, C/MR^2 and the constraint that quantity places on core size will not be known without making additional measurements of the response of Mercury's motion to solar torques, measurements requiring a precision that can be attained only by observations from the planet's surface (Peale, 1975). Proof that a core does or does not exist in Mercury, therefore, will not be forthcoming for some time.

There are several lines of evidence, all of them indirect and assumption dependent, to suggest nonetheless that a core has formed in Mercury:

(1) Mercury has an internal magnetic field (Ness et al., 1974, 1975a, b), and a convective dynamo in a fluid conducting core is the preferred mechanism for field generation (Ness et al., 1975b; Stevenson, 1974, 1975).

(2) Large areas on Mercury's surface are covered with smooth plains, which from stratigraphic relationships and similarities to lunar maria are thought to be largely volcanic in origin (Murray et al., 1974a, b; Trask and Guest, 1975; Strom et al., 1975). The inferred melting necessary to produce widespread igneous activity and the close similarity of the surfaces of Mercury and the moon suggest that Mercury has an iron-rich core and a silicate crust and mantle (Murray et al., 1974a, b, 1975).

(3) If Mercury has retained its full solar system complement of uranium and thorium, as suggested by the most specific of the current chemical models for solar nebula condensation and planetary accretion (Lewis, 1972, 1973; Grossman, 1972; Grossman and Larimer, 1974), then differentiation of a core is predicted for all likely initial temperature distributions and thermal conductivity values (Siegfried and Solomon, 1974).

For the remainder of this paper, a large iron-rich core in Mercury is regarded as probable. We examine next some possible histories of core formation and evolution.

TIME SCALE FOR CORE FORMATION AND THE STATE OF THE CORE

Segregation of core from mantle in Mercury could have altered drastically the planet's geology, as is discussed in a later section. The present state of Mercury's core may be intimately related to the planet's magnetic field. It is therefore important to understand the time scales for various chapters in the history of the core and to relate these times, where possible, to events recorded at the planet's surface. Three times will be of particular interest in this paper: the time when core formation began, the duration of core infall, and the time between core formation and core solidification.

To model the process of core formation in Mercury, several assumptions are necessary. For the calculations reported here, it is assumed that the planet began as a body homogeneous on a large scale. Such a premise is likely only if condensation in the primitive solar nebula was a much faster process than planetary accretion, and while a case can be made for such a hypothesis on both chemical (Lewis, 1972, 1973) and mechanical (Safronov, 1972; Weidenschilling, 1974) grounds, the issue is far from settled. If there were a pronounced chemical heterogeneity in Mercury after accretion, particularly a varying content of metallic iron with radius (Grossman and Larimer, 1974), many of the arguments of this paper would be qualitatively unchanged but the detailed calculations would require revision.

Core formation is assumed to follow Elsasser's (1963) scenario for core infall in the earth. To quantify Elsasser's model it is necessary to postulate a necessary criterion for segregation of metal from silicate. In our calculations (Siegfried and Solomon, 1974) this criterion is taken to be local melting of the metal phase. This is formally different from Elsasser's (1963) treatment in which the viscosity of the silicate material controls segregation. Such a distinction is not quantitatively significant as the melting curves for iron and silicates in Mercury are likely similar (Siegfried and Solomon, 1974) and close approach to silicate melting is necessary in Elsasser's scenario for rapid core differentiation. The Elsasser model for the earth has been modified by the postulate that core infall began when local temperatures exceeded the Fe-FeS eutectic (Murthy and Hall, 1970), a temperature significantly below the iron melting curve (Brett and Bell, 1969; Usselman, 1975). Such a possibility is not expected to be important for Mercury if current cosmochemical models for the terrestrial planets (Lewis, 1972; Grossman, 1972), which predict negligible sulfur and volatiles in Mercury, are approximately valid.

The thermal evolution of Mercury prior to, during and following core formation is modeled as in Siegfried and Solomon (1974). Only a few minor modifications are made in our earlier procedure. Adiabatic compression is included as an initial heat source. The surface temperature is fixed throughout the planetary history at 380 K (Morrison, 1970; Ulich _et al._, 1973; Briggs and Drake, 1973). The adopted ratios for Th/U and U/Fe are 3.7 and 6.4×10^7, respectively, the latter taken as a representative "cosmic" ratio from the carbonaceous and L-type chondrites (Mason, 1971). With such a U/Fe ratio, the mean present day uranium abundance in Mercury is 38 p.p.b., slightly lower than the figure used by Siegfried and Solomon (1974) in most of their models.

The time when core differentiation begins in Mercury is critically dependent on the nature of the early sources of heat in the planets, about which very little is known. In the absence of significant heat from accretional energy or from a solar or other extraplanetary source, core formation would not begin for at least 1 b.y. after planetary formation (Siegfried and Solomon, 1974). This is illustrated by the thermal model in Fig. 1, for which all parameters were chosen to favor early differentiation but for which no _ad hoc_ early heating processes are postulated. The initial temperature is taken to be 1400 K, the approximate condensation temperature for most of the planet (Lewis, 1972), plus an adiabatic compression term. The thermal conductivity of the metal-silicate mix is taken to be the lower Hashin-Shtrikman bound (see Siegfried and Solomon, 1974).

Core formation begins at 1.2 b.y. after planetary origin in the model of Fig. 1 and is complete by 1.8 b.y.. The conversion of gravitational energy to heat (Birch, 1965) accelerates core infall somewhat, but the event is by no means the catastrophe proposed for the earth (Elsasser, 1963; Birch, 1965). The total heat gained during differentiation is equivalent to a mean temperature rise of less than 700 K, as discussed in a later section. A duration of about 500 m.y. for core segregation is typical of thermal models with a nearly flat initial temperature profile. The figure is controlled by the rate of heat production, including gravitational energy, and is not sensitive to shifting upwards or downwards the temperature-depth curve used as the criterion for local metal-silicate separation.

Because the thermal conductivity of the iron core is high and because all radioactive heat sources are presumed to remain in the silicate phase during differentiation, the core cools rapidly in the model of Fig. 1 and is solid 1.5 b.y. after core infall is complete. Such a core solidification time is characteristic of the thermal models we have considered, and should more properly be treated as a lower bound since solid state convection in the silicate mantle, a process not included in these calculations, might hasten solidification greatly (Cassen, 1975).

The process of core formation leads to melting of most of the mantle of Mercury (Fig. 1), the effects of which would surely modify greatly the planetary surface. If the surface of Mercury is as old as has been inferred from photogeology (Murray et al., 1974a, b, 1975), at least as old as the period of heavy bombardment of the lunar surface 4.0 b.y. ago, then core formation would have to have been complete at least 1 b.y. earlier than in Fig. 1 and the core, for an otherwise similar thermal model, would have been solid for the last 2.5 b.y..

Until viable alternatives to the convective dynamo are proposed as mechanisms for generating Mercury's magnetic field, thermal models such as that in Fig. 1 which predict a solid core should be regarded with some disfavor. What additional sources of heat are available that would keep the core in the model of Fig. 1 molten or partly molten until the present time? Tidal dissipation is one possible source, since Mercury's rotation is clearly a product of tidal evolution (Pettengill and Dyce, 1965; Peale and Gold, 1965; Colombo, 1965; Goldreich and Peale, 1966). The tidal heating is dependent on the original rotation period of the planet, on the time scale for deceleration, and on the distribution of anelasticity with depth and time, none of which are known well enough to estimate whether tidal heating could have prevented core solidification, though this possibility is probably doubtful.

Radioactive heat sources in the core could maintain a molten state. Suppose that during core segregation, separation of U and Th into the silicate component did not operate at perfect efficiency. What fraction of U and Th would have to be trapped in the core to keep the metal molten at present for a thermal model otherwise similar to that of Fig. 1? If 10 percent of the total U and Th were uniformly distributed in the core, core solidification would be retarded by 1 b.y. but the core at present would be solid. Only if 20 percent or more of Mercury's U and Th were in the core would the core be at least partly melted at present. Such inefficient fractionation of U and Th during differentiation is geochemically impausible.

Heating due to K^{40} has been ignored in the calculations here and in Siegfried and Solomon (1974) because of the predictions of cosmochemical models (Lewis, 1972; Grossman, 1972) that Mercury should be composed dominantly of materials that condensed at temperatures higher than that at which potassium-bearing phases first appear as solids. Because of the suggestion (Lewis, 1971; Goettel, 1972) that K^{40} may be an important heat source in the earth's core, it is of interest to ask how much potassium in Mercury's core would be necessary to prevent core solidification. Toksöz and Johnston (1975), using models similar to those of Siegfried and Solomon (1974) but with earlier core formation, estimate that 156 p.p.b. K^{40} in the core is the minimum necessary for the iron to be at least partly molten at present. This corresponds to a total potassium abundance in the planet of about .1 percent by weight, or a bulk K/U ratio of about 20,000. Such a high K/U ratio would require substantial revision of cosmochemical models for the terrestrial planets.

A fluid iron core is a necessary but by no means a sufficient condition for a convective dynamo. Stevenson (1975) has examined the stronger condition that the thermal gradient in the core be at least as steep as an adiabat. The details of such a condition are not certain even for the earth (Higgins and Kennedy, 1971; Birch, 1972; Kennedy and Higgins, 1973), but for Mercury about half the heat sources in the planet must be retained in the core, according to Stevenson (1975), for convection to be permitted. If this radioactive heat in the core is provided by K^{40}, a roughly chondritic K/U ratio (Wasserburg *et al.*, 1964) is implied.

An alternative to radioactive heat sources in the core is to distribute the gravitational heating over more of the planet's history. One such model is shown in Fig. 2. The initial temperature is *ad hoc*, but is chosen so that the near surface regions are hot and the deep interior approaches the equilibrium black-body temperature at Mercury's distance from the sun. The nature of the energy source for heating the exterior of the planet is not considered here; several large impacts or the unipolar induction heating model of Sonett *et al.* (1968) are two possibilities.

The upper 350 km of Mercury in the model of Fig. 2 are initially melted and rapidly cool and differentiate. Metal-silicate differentiation proceeds downward slowly, because the heat released by gravitational infall is only a small fraction of the heat necessary to melt the deep interior. Core formation extends over a period of about 4 b.y., and the core is partly molten at present. In such a thermal model, the near-surface igneous activity would be expected to be confined to the early history of the planet and the partly fluid iron core would satisfy at least a minimum condition for dynamo generation of Mercury's magnetic field.

A final possibility to reconcile the postulates of early core formation and a currently fluid core is that in addition to iron and to nickel and other siderophiles, Mercury's core contains one or more additional elements such as a sulfur which substantially lower the solidus temperature of core material (Brett, 1975). Unless such elements have been incorporated at greater than trace amounts, however, the core would be molten only within a thin outer layer. Whether a dynamo can be generated in a thin fluid shell is a serious question. The order-of-magnitude calculations of Stevenson (1975) indicate that the magnetic Reynolds number would likely be subcritical for fluid metallic layers much thinner than 100 km. To maintain convection in such a layer would still require heat sources within the core.

GEOLOGICAL CONSEQUENCES OF CORE FORMATION

Because more is known about the surface of Mercury than about the interior, a situation unlikely to be changed by new information in the future, it is important to relate the preceding discussion to Mercury's geology. What changes does core formation produce on Mercury's surface? What constraints does the known or inferred surface geological history place on the thermal history and on core formation? Two major consequences of core differentiation in Mercury are examined in this section: the change in planetary volume and the change in internal temperatures.

Volume changes with time in Mercury are of especial interest because of the global system of lobate scarps identified in the Mariner 10 photographs (Murray et al., 1974b; Strom et al., 1975). From the morphology and dimensions of these scarps and from their transection relationships, the features have been interpreted as thrust faults indicative of planet-wide compressive stresses. Strom et al. (1975) estimate that the observed displacements on the faults represent a 1 to 2 km decrease in the radius of Mercury. The compressive stage in Mercury's history apparently began at least by the end of the period of heavy bombardment of the surface (Trask and Guest, 1975) and continued after the later (but perhaps overlapping) period of emplacement of the smooth plains, thought to be of volcanic origin (Strom et al., 1975). Both

Murray et al. (1974b) and Strom et al. (1975) speculate that the planetary compression may be the result of core shrinkage.

Two effects contribute to the volume change associated with core formation: the change in compression associated with the redistribution of mass and the thermal expansion due to the conversion of gravitational potential energy to heat. For purposes of discussion, the thermal model of Fig. 1 is used to illustrate the magnitude of the volume change associated with these two phenomena.

For a planet fully differentiated into mantle and core with a temperature distribution given by the present-day profile in Fig. 1, the core occupies 66.5 percent of the planetary mass for likely values of the zero-pressure densities of mantle and core material (Siegfried and Solomon, 1975). The change in planetary volume due only to the redistribution of mass upon core differentiation may be determined by finding the radius of the homogeneous planet with the same composition and same temperature profile and with a mass equal to Mercury's present mass. This exercise is illustrated in Fig 3, which is a mass-radius diagram for planets isochemical with the assumed present-day model for Mercury (Siegfried and Solomon, 1974) and with temperature distributions taken from Fig. 1. The equation of state of the metal-silicate mix follows the procedures of Siegfried and Solomon (1974). No account is taken of possible volume changes associated with chemical reactions, in particular with melting or with solid-solid phase changes.

The mass-radius curve labeled 4.6 b.y. gives the change in planetary radius $\Delta R/R$ due only to the rearrangement of mass: the radius increases by 0.36 percent, or 8 km, upon core infall. The expansion of the planet during differentiation is due to the greater compressibility of the silicate fraction and to the greater pressures in the outermost 700 km of the planet for the undifferentiated state (see Fig. 9 of Siegfried and Solomon, 1975). Birch (1965) and Flasar and Birch (1973) obtained a similar result for the earth.

An additional increase in planetary radius is caused by thermal expansion. The planet is hotter after core infall than prior to differentiation (see below) and there is a thermal expansion added to the expansion calculated above at fixed temperature distribution. This extra ΔR may be obtained from Fig. 3 by taking the difference in radius, at fixed mass, predicted by the mass-radius curves for temperature distributions just prior to (1.2 b.y.) and immediately following (1.8 b.y.) core segregation, giving $\Delta R = .0030\ R$. The total change in radius associated with core formation is $\Delta R = .0066\ R$ or over 13 km. This expansion would produce an increase in surface area of 1.3 percent, a figure that would be increased slightly if account had been taken of solid-solid and solid-liquid phase changes concurrent with differentation. The total expansion is not sensitive to the particular thermal model used to construct Fig. 3, i.e., to the precise timing of core infall.

The pronounced increase in Mercury's surface area would not go unnoticed at the planet's surface. Huge rift valleys, grabens and other tensional features would be expected of a lithosphere subjected to the stresses associated with such a volume change. The absence of such features on Mercury's surface today (Murray et al., 1974a, b) indicates either that (1) Mercury never expanded by the large amount calculated above or (2) the effects of the expansion have been erased by subsequent surface-modifying events. If case (1) holds, then the scenario of homogeneous accretion followed by core differentiation would have to be replaced by one of inhomogeneous accretion and more or less in situ core development (Grossman and Larimer, 1974; Murray et al., 1975). If case (2) holds, then core infall must have been substantially complete prior to the period of heavy bombardment of Mercury's surface, tentatively identified (Murray et al., 1975) with the similar period on the moon lasting until 3.9 to 4.0 b.y. ago (e.g., Tera et al., 1974). Choice between these two alternatives must await information beyond what is now available. We might note that in the model of Fig. 2, planetary expansion would be spread over 4 b.y., probably a basis for the model's rejection.

Subsequent to completion of core differentiation, Mercury cools. As a consequence there is a slow thermal contraction of the planet, with by far the greatest contribution to the contraction coming from the lithosphere. The amount of this contraction may be determined by comparing the mass-radius curves for 1.8 and 4.6 b.y. in Fig. 3: $\Delta R = -.0010$ R or 2 km. (A comparison of the mass-radius curves for differentiated planets with the appropriate temperature distributions gives the same result.) Thus thermal contraction of Mercury following core formation is consistent in time and in magnitude with the contraction necessary to have produced the global pattern of lobate scarps (Strom et al., 1975).

For the volume change following completion of core infall, neglect of volume changes associated with chemical reactions may be a critical omission. If the core has partly or wholly solidified, the additional volume decrease may have been considerable. The relevant reaction at core pressures in Mercury ($P > 70$ kbar, Siegfried and Solomon, 1974) is Fe liquid $\rightarrow$ γ-Fe (Bundy, 1965), for which no measurements of the specific volume change are available. Birch (1972) has estimated the volume change associated with melting γ-Fe to be 0.42 cm^3/mole at zero pressure and 0.36 cm^3/mole at 50 kbar. Using the latter figure and a molar volume of 7.02 cm^2 for γ-Fe (Birch, 1972) gives a volume change of slightly more than 5 percent upon core solidification. Mercury's core occupies 43 percent of its volume, so that complete solidification of the core would introduce a total decrease in the planetary volume of 2.2 percent, or a decrease in the radius of .73 percent (18 km).

If the specific volume change upon melting γ-Fe is even approximately correct, then solidification of Mercury's core can be excluded as a possibility for the last 3.9 to 4.0 b.y. and, if the time scale for solidification determined above is valid, for its entire history. If the core began by melting and gravitational infall of molten metal, then the core is still largely molten at present and one of the heat sources discussed earlier must be present in the core. On the other hand, if the core largely accreted in situ as solid metal, either it is still solid or it melted very early in Mercury's history and has stayed fluid. Use of volume change as a definitive constraint on core history, however, should await direct measurement of the thermodynamic properties of the γ-Fe melting reaction.

The change in gravitational energy during core formation is converted largely to heat. This heat is added in appropriate increments in the thermal history calculational scheme (Siegfried and Solomon, 1974), but it is useful also to estimate the mean temperature rise in the planet equivalent to the total additional thermal energy. The difference in the gravitational potential energy

$$\Omega = 4\pi \int_0^R g(r)\ \rho(r)\ r^3\ dr$$

(Birch, 1965) between an undifferentiated Mercury (1.2 b.y. in the thermal model of Fig. 1) and a differentiated planet immediately following core segregation (1.8 b.y. in Fig. 1) is 5.5×10^9 erg/g. Ignoring differences in the strain energy of the two states (Birch, 1965) and using a specific heat appropriate to the mass-weighted values for iron and appropriate silicates (see Siegfried and Solomon, 1974), this energy is equivalent to a rise in the mean temperature of Mercury of 680 K. For comparison, the rise in mean temperature after core infall is 2300 K in the earth (Flaser and Birch, 1973) and less than 300 K in Mars.

The distribution of this heat with radius is somewhat arbitrary, but it likely that partial melting at shallow depths in the planet was one consequence of core fractionation. Probably some igneous activity at Mercury's surface would result, but on thermal grounds the surface need not have suffered the massive melting and disruption that has often been suggested for the earth following core formation. The inference from photogeology that Mercury's near surface regions are moon-like (Murray et al., 1974a, b), however, would require efficient differentiation of the shallow portion of Mercury prior to events recorded in the present morphology of the surface.

DISCUSSION

There are three major constraints on the evolution and present state of a metallic core in Mercury: (1) Mercury's magnetic field, (2) Mercury's surface geology, and (3) cosmochemical models for the terrestrial planets. All three constraints imply that a core is present. Individually these constraints and the inferences made from them impose various and often conflicting conditions on the history of the planet and of the core.

If the core in Mercury differentiated from an originally homogeneous planet, then an expansion in planetary radius of 13 km and a rise in mean temperature of about 700 K would result from core infall. Because the surface of the planet would be substantially altered by such an event, core differentiation must predate the oldest geological units comprising a major fraction of Mercury's surface, probably either the intercrater plains or the heavily cratered terrain mapped by Trask and Guest (1975). Murray et al. (1975) postulate that the heavily cratered terrain records a period of intense bombardment of all of the terrestrial planets at about 4.0 b.y. ago. If core formation was completed in the first 500 m.y. of Mercury's history, an early source of heat is necessary. Accretion of small planetesimals is probably too slow to trap much gravitational energy as heat (Safronov, 1972; Weidenschilling, 1974), but late impacts by large bodies, solar effects and tidal dissipation are all possibilities. The age of Mercury's surface is controversial (Chapman, 1975), however, and a younger age would relax the requirement for such early heat sources.

If the cosmochemical models of Lewis (1972, 1973) or Grossman (1972) are taken literally, then Mercury's core is almost entirely iron-nickel with little or no light elements added and the radioactive heat in Mercury is provided dominantly by uranium and thorium with only negligible potassium. For such chemical models, solidification of the core would follow completion of core infall within 1 b.y. or less. Thus constraints (2) and (3) above are incompatible with the most common inference from constraint (1), that Mercury has a fluid convecting core. The only other currently viable mechanism for internal magnetic field generation is permanent magnetization (Ness et al., 1974, 1975a, b; Stevenson 1974, 1975). Most thermal models for Mercury probably preclude such an explanation, however. Only a thin outer shell, perhaps 50 to 80 km thick (Figs. 1 and 2), is currently below the Curie temperature in models in which core differentiation has proceeded to completion, and the magnetization required of such a shell would be very high (Ness et al., 1975b). If segregation of metal from silicates has been incomplete in the outer portions of the planet, however, then the higher Fe content would give a greater thermal conductivity, and thus a thicker shell at sub-Curie temperatures, as well as a potential for greater remanent magnetization. Such a possibility may still permit permanent magnetization as an explanation of the observed field.

Taking a contrary view, if the inference of a convective dynamo is taken literally, then a substantial source of heat in the core not predicted by current cosmochemical models must be postulated. Gravitational energy and tidal dissipation are two likely heat sources, but both were probably spent early in Mercury's history and thus would not have prevented subsequent core solidification. Radioactive heating is the most probable heat source, and cosmochemical models for Mercury incorporating such a necessity deserve considerable attention in the future.

Most likely Mercury has a large metallic core. Very probably the core is molten, both because of the magnetic field arguments and because shrinkage of the core upon solidification can probably be excluded for most of Mercury's history by its surface geology. An obvious implication of the discussion of this paper is that the simplest comprehensive models of chemistry, of magnetism, or of surface history of the terrestrial planets are unlikely to survive in complete detail the quantitative tests made possible by new data from these bodies.

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Table 1. Hydrostatic and nonhydrostatic contributions to J_2 for the terrestrial planets.

Planet	J_2 $\times 10^3$	g cm/sec^2	J_{2H} $\times 10^3$	J_{2NH} $\times 10^5$	J_{2NH} scaled to Mercury $\times 10^5$
Earth	1.08264[a]	980	1.07166[b]	1.098	7.7
Moon	0.2048[c]	162	0.00928[d]	19.55	3.9
Mars	1.966[e]	373	<1.94[f]	>3	>3
Mercury		370	.0003 to .0005[g]		

[a]Gaposchkin (1974)

[b]Khan (1969)

[c]Sjogren (1971)

[d]Jeffreys (1970)

[e]Sinclair (1972)

[f]Binder and Davis (1973)

[g]Siegfried and Solomon (1974)

FIGURE CAPTIONS

Fig. 1. A model for the thermal evolution of Mercury. The initial temperature is 1400 K (after Lewis, 1972) plus a contribution from adiabatic compression. Other input parameters are discussed in the text or follow Siegfried and Solomon (1974). The time since planetary origin, in billions of years, is shown adjacent to the corresponding temperature profile.

Fig. 2. An alternative thermal history model for Mercury. Because the (arbitrary) initial temperature distribution is strongly peaked toward the surface, planetary differentiation is spread over a substantially longer time than in Fig. 1. Other parameters are identical for the two models.

Fig. 3. Mass-radius diagrams for planets composed of a homogeneous mixture of 66.5 weight percent Fe-Ni ($\rho_0 = 7.97\ g/cm^3$) and 33.5 weight percent "silicate" ($\rho_0 = 3.2\ g/cm^3$), for temperature distributions at various times taken from Fig. 1. The radius of such planets for a mass equal to Mercury's mass may be used to infer the radius changes during various phases of the planet's history (see text).

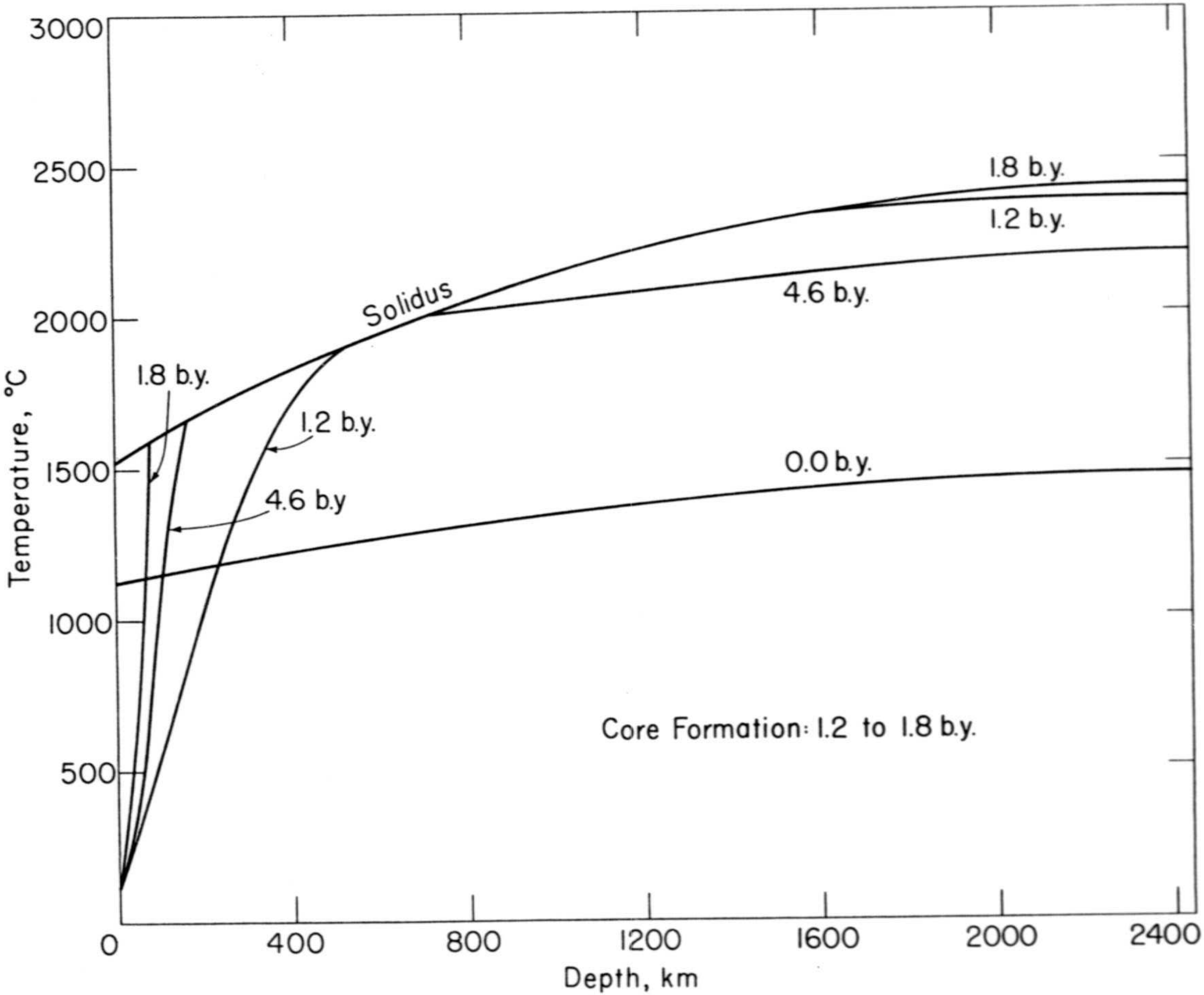
3000
2500
2000
1500
1000
500
Temperature, °C
0
400
800
1200
1600
2000
2400
Depth, km
Solidus
1.8 b.y.
1.2 b.y.
4.6 b.y
1.8 b.y.
1.2 b.y.
4.6 b.y.
0.0 b.y.
Core Formation: 1.2 to 1.8 b.y.

Fig. 1

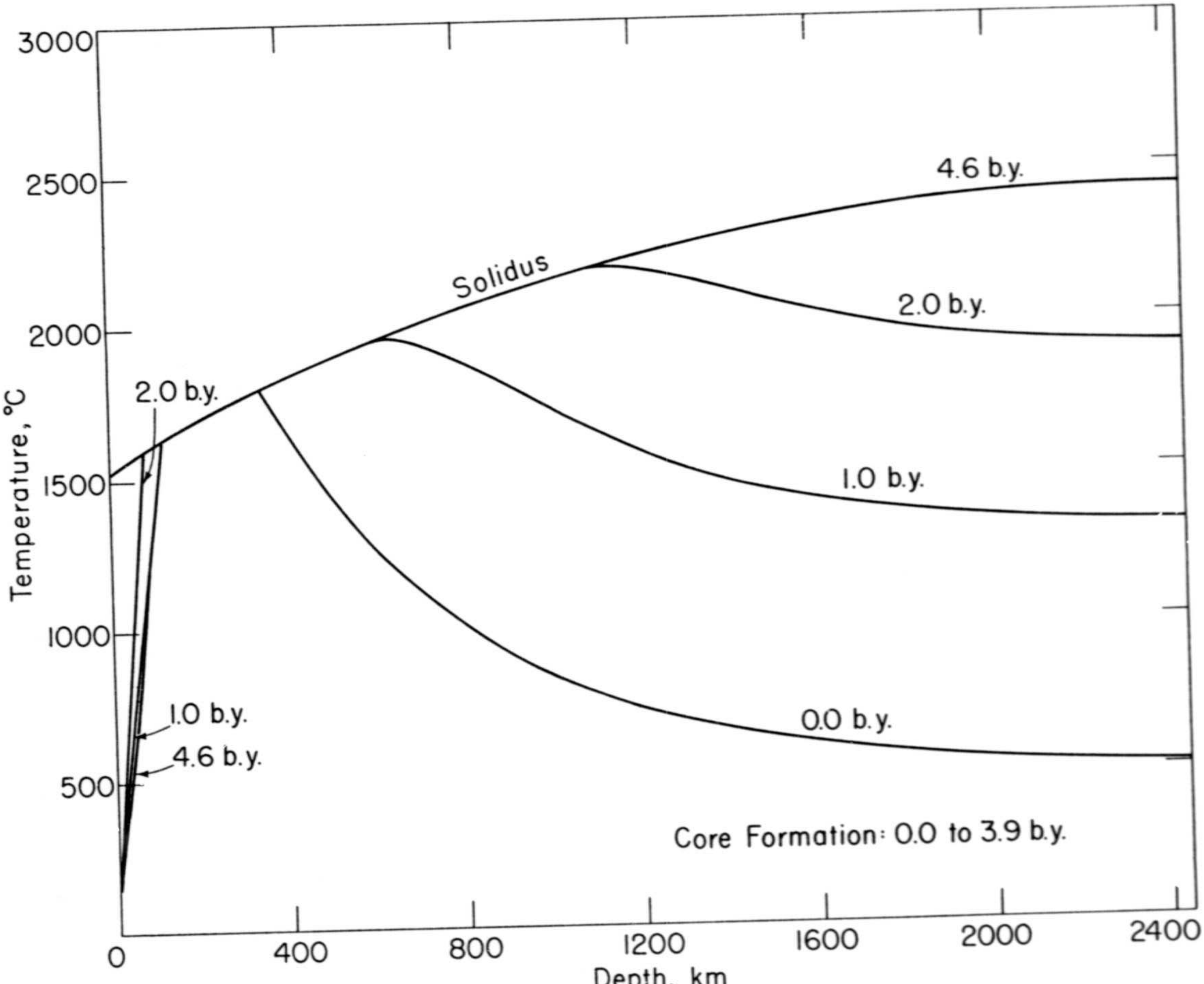
3000
2500
2000
1500
1000
500
Temperature, °C
0
400
800
1200
1600
2000
2400
Depth, km
Solidus
4.6 b.y.
2.0 b.y.
1.0 b.y.
0.0 b.y.
2.0 b.y.
1.0 b.y.
4.6 b.y.
Core Formation: 0.0 to 3.9 b.y.

Fig. 2

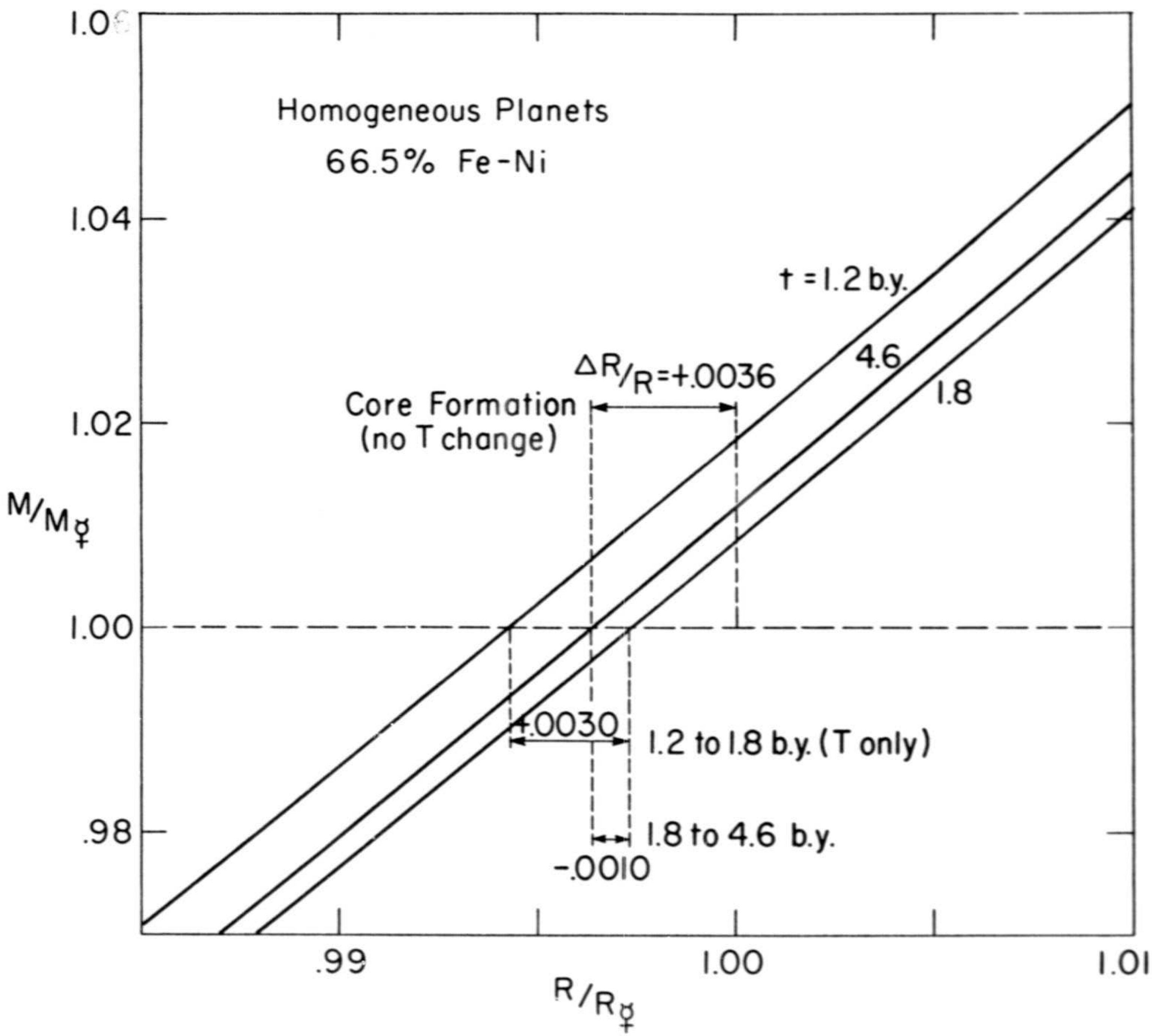
Homogeneous Planets
66.5% Fe-Ni
t = 1.2 b.y.
4.6
1.8
ΔR/R=+.0036
Core Formation
(no T change)
+.0030
1.2 to 1.8 b.y. (T only)
1.8 to 4.6 b.y.
-.0010
M/M☿
1.04
1.02
1.00
.98
.99
1.00
1.01
R/R☿

Fig. 3