6 November 1981, Volume 214, Number 4521 SCIENCE

Models of the Earth's Core

D. J. Stevenson

In *Voyage to the Center of the Earth*, Jules Verne's poetic elaboration of geology published in 1864, the professor's nephew Axel has a dream in which "the granite rocks softened; solid matter turned to liquid under the action of intense heat; water covered the surface of the globe, boiling and volatilizing; steam enveloped the earth." Verne's fantastic yet scientifically prescient writings reflected the high level of interest in the interior and history of the earth among 19th-century scientists (*1*). In the subsequent century, fact has replaced much (but not all) of the fantasy. Nobody has yet made a field trip to the earth's core and there is almost certainly no material reaching the earth's surface from the core. However, the combination of seismology, high-pressure experiment and theory, geomagnetism, fluid dynamics, and current views of terrestrial planetary evolution lead to strong constraints on core models.

Summary. Combined inferences from seismology, high-pressure experiment and theory, geomagnetism, fluid dynamics, and current views of terrestrial planetary evolution lead to models of the earth's core with the following properties. Core formation was contemporaneous with earth accretion; the core is not in chemical equilibrium with the mantle; the outer core is a fluid iron alloy containing significant quantities of lighter elements and is probably almost adiabatic and compositionally uniform; the more iron-rich inner solid core is a consequence of partial freezing of the outer core, and the energy release from this process sustains the earth's magnetic field; and the thermodynamic properties of the core are well constrained by the application of liquid-state theory to seismic and laboratory data.

The synthesis presented here is devoted to the defense of the following propositions.

Core formation was contemporaneous with earth accretion. Partial burial of the gravitational energy of infall as the earth grew from planetesimals would have rapidly ensured internal temperatures sufficient for the formation and downward migration of a liquid iron alloy. Large-scale asymmetries may have played an important role in core emplacement, and the core formation process necessarily has implications for the present core.

The core cannot be in chemical equilibrium with the mantle. This follows very generally from the inevitable pressure dependence of chemical equilibria and from a consideration of boundary-layer dynamics.

The outer, liquid core is predominantly iron but cannot be almost purely iron. A pure iron core would have long since frozen (if it could form at all). There must be at least one alloying constituent which substantially depresses the melting point. Sulfur and possibly oxygen are likely minor constituents responsible for this.

The inner core–outer core boundary represents a thermodynamic equilibrium between a liquid alloy and a predominantly iron solid. This is the most plausible explanation for the existence of a small, solid inner core. Furthermore, growth of the inner core over geologic time provides the most probable energy source for sustaining the geomagnetic field.

Thermodynamic and transport properties of the outer core can be estimated from liquid-state theories. Seismic data can be "inverted" to infer properties of the mean interatomic potential. This can then be used to predict other useful quantities.

The outer core is probably almost adiabatic and almost uniform in composition. Seismic data, hydromagnetics of the core, and thermal evolution studies are consistent with this hypothesis.

None of these propositions is revolutionary, but neither are they universally accepted by geophysicists. Birch's admonition (*2*), concerning the dangers of excessive confidence when discussing the earth's interior, still merits attention. My intent is to present a coherent picture which explains most of the data with the fewest ad hoc assumptions. No attempt is made to discuss exhaustively all the alternatives.

Core Formation

Core formation cannot be divorced from the more general and contentious issues of solar system and planetary formation. In most recent scenarios, the forming solar system is modeled as an accretion disk: a very oblate nebula of gas and dust containing a central concentration of matter that eventually becomes the sun (*3, 4*). In Cameron's models (*3*), the nebula undergoes numerous gravitational instabilities in the gas phase, leading to the formation of a large number of giant, gaseous protoplanets. This model encounters difficulties in explaining the terrestrial planets (*5*) and is not pursued here. The preferred models are those developed by Safronov (*6*) and others (*7*) in which condensation and

The author is an associate professor of planetary science in the Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena 91125.

 0036-8075/81/1106-0611$01.00/0

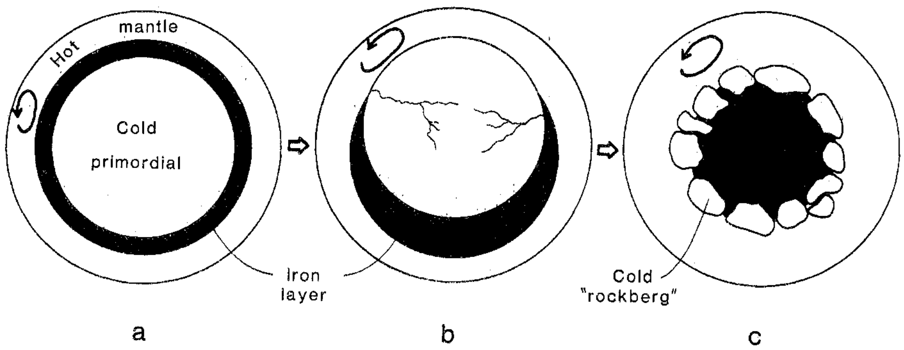

Fig. 1. Hypothetical scenario for the early stages of core formation. At this stage, the protoearth may be much smaller than the final earth. (a) A cold, undifferentiated primordial core is overlain by the denser iron that has accumulated from the partially molten mantle above. (b) A spontaneous asymmetry (*23*) must develop. Large, nonhydrostatic stresses on the resulting primordial core lead to deformation and even fracturing. (c) The debris from this is distributed in the form of "rockbergs" around the newly formed core. Thermal equilibration of these rockbergs with the surroundings is by thermal diffusion and takes longer than earth accretion.

settling of solid grains to the nebula midplane occur, followed by gravitational instability of the resulting dust layer. Planetesimals approximately 1 kilometer in radius form and undergo collisional evolution, leading to the formation of larger bodies. In a gas-free environment, the accretion time of the earth is perhaps 10^7 to 10^8 years (*7*). Accretion in a dense, gaseous environment has also been proposed but is difficult to reconcile with rare-gas systematics, especially ^{36}Ar (*8*). In the absence of an efficient physical or chemical mechanism for large-scale separation of iron from silicates in the solar nebula (*9*), the earth is most likely to have accreted in an approximately homogeneous fashion, with incoming planetesimals containing both iron and silicates in roughly solar (chondritic) abundances (*10*). These planetesimals might possess differentiated structures (*11*), in which case the iron is already present as a core with a radius of a few tens of kilometers or even more. Alternatively, collisional evolution may lead to an intimate mixture of iron and silicates. In any event, the metallic iron will be substantially disseminated after impact (*12*). A strongly heterogeneous accretion, in which a terrestrial iron core forms directly, is difficult to reconcile with the present-day fluid, convecting outer core of the earth (*13*).

For an accretion time of 10^7 to 10^8 years, surface temperatures of the protoearth may be low (*14*). However, only a small fraction of the gravitational energy of infalling planetesimals needs to be buried during impact to ensure high internal temperatures. Let $T(r')$ be the temperature at a radius $r' < r$, the outer radius of the growing planet. If no internal heat transport occurs, then (*15*)

$$T(r') = \frac{h}{C_p}\left[\frac{GM(r')}{r'} + \frac{1}{2}v_\infty^2\right] + T_a \quad (1)$$

where h is the portion of impact energy retained as buried heat, C_p is the specific heat, G is the gravitational constant, $M(r')$ is the mass within radius r', v_∞ is the velocity at infinity of an incoming planetesimal, and T_a is the ambient temperature. This neglects previous heating events within planetesimals and assumes that earth accretion occurs after the decay of short-lived radiogenic nuclides (such as ^{26}Al) but so rapidly that the important long-lived sources (^{40}K, ^{235}U, ^{238}U, and ^{232}Th) have not yet provided significant heating.

The value of h is not known, but is not likely to be very close to either 0 or 1 (*15*). If it is assumed that accretional heating of the outer regions of the moon is needed to explain the anorthositic highlands, then $h \sim 0.5$. Even for $h \sim 0.1$, regions at a depth of ~ 10 km in the protoearth could have approached the melting point when the body was only the size of Mars. Melting of disseminated iron would occur before melting of the major silicate fraction, especially if any alloying constituent (for instance, sulfur) is present to depress the melting point. Liquid iron would have readily penetrated downward through the silicate matrix (*16*) but since Eq. 1 shows that deeper levels were probably colder, the iron could not have immediately migrated to the center of the planet. Rather, it would have formed a spheroidal layer, the bottom of which would necessarily have been at its freezing point. Below this layer would have existed a zone consisting of an undifferentiated, cold primordial mixture of materials, and the mantle above the iron-rich layer would have been iron-depleted and well stirred by thermal convection and Rayleigh-Taylor instabilities initiated by incoming iron-bearing planetesimals.

Elsasser (*17*) proposed that core formation could eventuate from a Rayleigh-Taylor instability in which large "blobs" of iron (perhaps hundreds of kilometers in radius) grow and migrate downward through the less dense, deeper regions. Although Rayleigh-Taylor instabilities undoubtedly exist, their characteristic length scale is likely to be much smaller than Elsasser proposed because of the very strong temperature dependence of the silicate viscosity (*18*). Consider an inviscid, fluid layer of density $\rho + \Delta\rho$ overlying an infinite layer of density ρ and shear kinematic viscosity $\nu = \nu_0 \exp(-z/L)$, where z is the vertical coordinate and $z = 0$ at the interface. (The simple analytic form for the viscosity is chosen for computational convenience.) The situation of interest is positive L (viscosity increasing rapidly downward). The fastest growing modes of the resulting Rayleigh-Taylor instability are found (*19*) to have a characteristic length scale $\sim L$ and growth rate $\sim 0.25(\Delta\rho g L/\nu_0)$, where g is the gravitational acceleration. Typically, L is a few kilometers and the blobs sink slowly because of the need to warm the region beneath them. The time scale for forming a core is found to be several hundred million years, despite the exothermic nature of the process. The long time scale is directly attributable to the small thermal diffusivity and the absence of any other process for transporting heat downward (*20*).

Another model, proposed by Vityazev and Mayeva (*21*), involves the downward migration of a spherical iron layer by melting of the iron immediately below the interface. Silicate grains are thereby released from the primordial core and can float up through the iron layer to merge with the mantle above. The problem with this mode of core formation is again the slow thermal diffusion: heat must be conducted into the interface from above to supply the latent heat of fusion Q_L. The migration velocity of the iron layer is found to be at most $\sim K\Delta\rho g/Q_L$ (*22*) and the corresponding time of core formation $\sim 10^9$ years.

Although the Elsasser blob model and the Vityazev and Mayeva sinking layer model can work, given enough time, there exist processes for iron migration that are more rapid and depend on the non-Newtonian rheological properties of the material. Figure 1 shows the spontaneous asymmetry that can occur on a time scale of hours for a rigid, low-density core overlain by an inviscid, high-density layer. This process has also been invoked to explain the nearside-farside asymmetry of the moon (*23*). In the earth (unlike the moon), the nonhydrostatic stresses that develop because of this spontaneous symmetry breaking are capable of deforming or even fracturing the primordial core (*24*). The iron can then migrate downward. Migration along iron-filled cracks (analogous to "magma fracturing") may also play a role (*25*).

In a sense, the details of core emplacement are not important since Eq. 1 suggests that they probably refer to a time when the protoearth was much smaller than its final size. It is more interesting to consider the process by which most of the core mass is accumulated. This is later in the accretionary process, when the entire interior of the earth is hot because of the combined effects of accretion and the gravitational energy of core growth (*26*). The suggested physical picture at this stage is indicated in Fig. 2.

In this almost steady-state process, incoming iron-bearing planetesimals disseminate on impact. Lateral heterogeneities occur and will be especially important for the largest planetesimals, but these can only aid downward migration of the iron. Even if the distribution of iron-bearing impact debris is rather uniform, foundering or Rayleigh-Taylor instabilities ("subduction") of the outer boundary layer, ~ 1 to 10 km in thickness (*27*), must occur. As this material migrates into hotter, subsurface regions, the iron alloy is mobilized and can percolate to form "diapirs." The typical size of such iron blobs in this scenario is ~ 1 km, possibly more. Their Stokes velocity downward depends on the viscosity of the surrounding mantle material. In a steady-state scenario, this is determined by the requirement that the excess heat of core formation is transported out by thermal convection. The required heat flow is in the range 1000 to 3000 ergs per square centimeter per second (*26*), and thermal boundary-layer theory predicts a convective heat flow of 2500 $(10^{16}/\nu)^{1/3}$ erg/cm^2-sec (*28*). The viscosity in the deep mantle is then in the range 10^{16} to 10^{17} cm^2/sec, for which partial but not necessarily complete melting of the silicates is required (*29*). Even for $\nu = 10^{17}$ cm^2/sec, the time for a 1-km blob to reach the core is less than 10^6 years, small compared to the accretion time, thereby justifying the steady-state picture.

Although the initiation of the core-forming process may be difficult because of the presence of cold, central material, most of the subsequent migration of core-forming material would have been efficient and would have easily kept pace with the accretion process. Rapid core formation is not inconsistent with lead isotope data (*30*) and with the evidence of a geomagnetic field in rocks as old as 3.5 billion years (*31*).

Core-Mantle Disequilibrium

The extent to which the core and mantle are now in chemical equilibrium is a contentious issue (*32*) involving both thermodynamic and fluid dynamic considerations. One conclusion can be reached independent of the details of core formation: almost all the chemical equilibration with mantle material during core formation would have been at pressures significantly different from the present core-mantle boundary pressure of 1.36 megabars. Since the partitioning of species between phases is likely to be strongly pressure-dependent (*33*), it follows that the newly formed core (especially the inner regions) is unavoidably out of equilibrium with the mantle above.

For the particular model of Fig. 2, an even stronger statement can be made: the only chemical equilibrium between solid silicates and liquid iron alloy or between two immiscible liquids took place at low pressures. This can be proved as follows: consider an iron blob of radius R migrating through the silicate mantle at velocity ν. The distance over which diffusive equilibrium between the iron and silicates can occur is limited to a diffusive boundary layer about $(Dt)^{1/2}$ in thickness, where D is the solute diffusivity in the silicate phase and $t \sim R/\nu$ is the time that the two phases are in contact. The total volume of mantle material that achieves diffusive equilibrium with the iron blob is then $\sim \pi R(DR/\nu)^{1/2}d$, where d is the depth of the mantle. Since the total volume of iron that must traverse the mantle is about one-quarter of the mantle volume, I conclude that the fraction f of the mantle that underwent chemical equilibrium after the formation of the large blobs is of order

$$f \sim 10^{-2}\left(\frac{D}{10^{-7}\ \text{cm}^2/\text{sec}}\right)^{1/2} \times \left(\frac{\nu}{10^{16}\ \text{cm}^2/\text{sec}}\right)^{1/2}\left(\frac{1\ \text{km}}{R}\right)^{5/2} \quad (2)$$

For likely parameter choices (*34*), $f << 1$. Chemical equilibrium can occur at low pressures, however, in regions where the iron is percolating through a silicate matrix.

After core formation, chemical equilibrium by diffusion across the core-mantle boundary is strongly inhibited. Experiments on double-diffusive systems (*35*), in which both heat and solute are transported across an interface between two convective layers, indicate that the ratio of solute redistribution to thermal buoyancy redistribution is proportional to $(D/K)^{1/2} << 1$, where $K \sim 10^{-2}$ cm^2/sec is the thermal diffusivity.

It should be stressed, however, that these conclusions are contingent on diffusion-limited processes. If partial melting in the deep mantle provides a melt that is miscible in an iron alloy, then the greater degree of high-pressure chemical equilibrium is conceivable (*36*). This may be important for understanding the composition of the core.

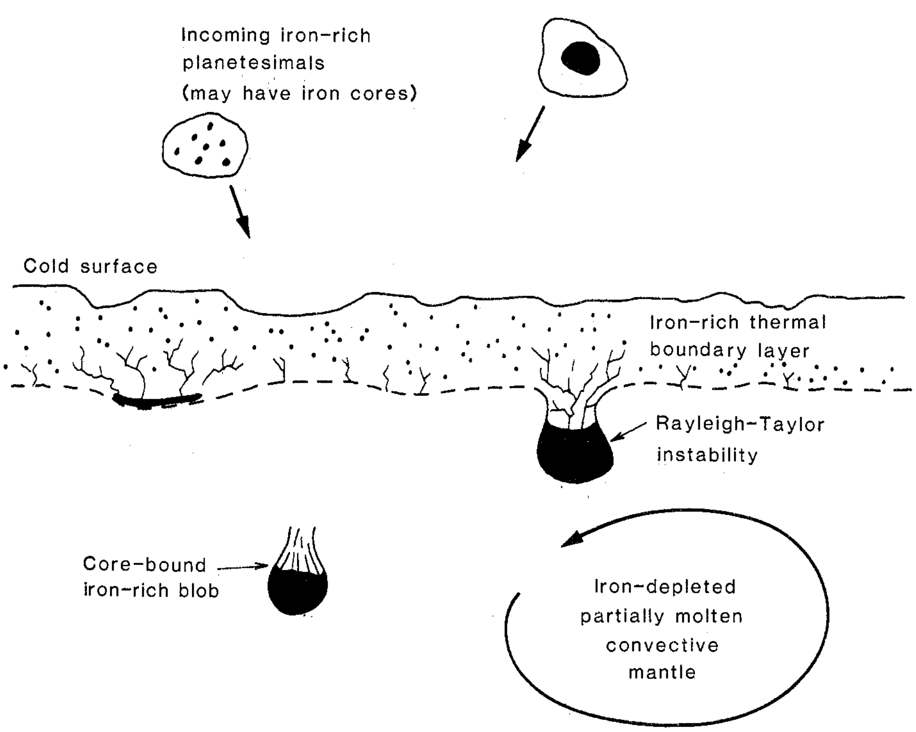

Fig. 2. Iron migration at a later (steady-state) phase of core formation. Below the thermal boundary, iron percolates into diapirs, which migrate down rapidly, ensuring incomplete chemical equilibration with the mantle environment.

Core Composition

Seismic data (both body waves and the earth's free oscillations), the astronomically determined moment of inertia, and the excellent assumption of hydrostatic equilibrium provide strong constraints on the pressure, density, and state of matter within the earth (*37*). Within an uncertainty of only a few kilometers, the radius of the core is 3480 km (where the density changes from 5.57 to 9.90 grams per cubic centimeter at a pressure of 1.36 Mbar). The region between about 1220 and 3480 km does not sustain shear waves and is conventionally described as a liquid (*38*). The inner core of radius 1220 km is generally believed to be solid, and there may be a discontinuity of 0.3 to 1.1 g/cm^3 at the inner core–outer core boundary (*39*). The density and pressure at the center of the earth are about 13 g/cm^3 and 3.64 Mbar, respectively.

The strongest argument for a predominantly iron core relies on a comparison of these data with laboratory shock wave data (*40*). Since there are numerous elements (for example, vanadium) or mixtures with a pressure-density relationship approximately consistent with that of the earth's core, one must always appeal to the cosmochemical argument as well: iron is the only sufficiently abundant element of appropriately high density to explain the core's properties (*41*). The density of pure iron at core pressures in a shock experiment is 8 to 11 percent more than the core density at the same pressure, as illustrated in Fig. 3. (The uncertainty represents the range of recent seismic models and the range of shock wave results; these have similar uncertainties.) The comparison is complicated by the differences in temperature and possible difference in thermodynamic state.

Neither shock temperatures nor core temperatures are known to better than about ± 1000 K, although they are fortuitously similar, both being around 4000 K at 2 Mbar. Shock temperatures in iron are estimated from porous Hugoniot data (*42*) and actual temperatures in the earth's core can be estimated from appropriately chosen mantle and core adiabats (*43*). This is illustrated in Fig. 4, together with the melting curve of pure iron estimated from the recent work of Brown and McQueen (*44*).

Evidently, the best estimates of the earth's core temperature lie below the melting curve of pure iron. This is not inconsistent with a fluid outer core, since the outer core density is too low for pure iron and an admixture of lighter elements can depress the melting point (just as salt depresses the freezing point of water). However, this raises another difficulty with the comparison of shock wave data and seismic data: the shock wave data may refer to the solid state (*45*). Combining this uncertainty with the uncertainty in temperatures, and assuming a coefficient of thermal expansion of 1×10^{-5} K^{-1} (Table 1), I conclude that the core density is less than that of pure, liquid (actually metastable) iron at the same temperature and pressure by 5 to 12 percent. Brown and McQueen reach an essentially identical conclusion.

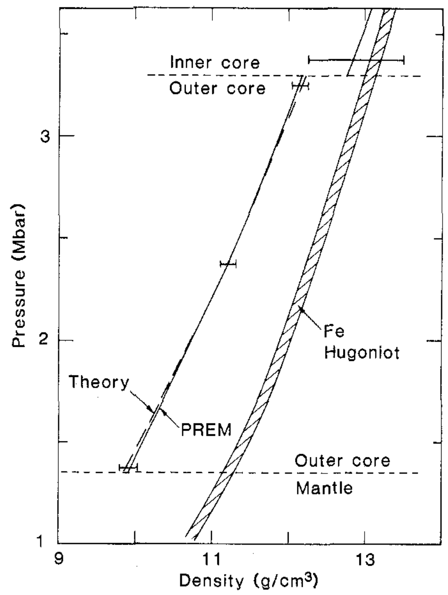

Fig. 3. Comparison of shock wave data with seismic data and theory for pressures and densities in the earth's core. Shaded region represents the range of densities for each pressure obtained in shock wave experiments on pure iron (*40*). Line labeled PREM represents a particular recent seismic model (*37*) for the pressure-density relationship in the core. Error bars represent the variation between various tabulated models. Notice the large uncertainty in the inner core (*39*). The dashed line is a theoretical liquid state model discussed in the text.

What other element or elements mixed with iron could explain this density difference? The presence of nickel (~ 4 percent by mass) is likely but does not substantially change the density (*46*). It is likely that any element with a substantially lower atomic number than iron ($Z = 26$) can reduce the density at high pressure (*47*). Possible candidates of sufficient cosmic abundance to be potentially important are H, He, C, N, Si, Mg(O), O, and S. To be acceptable, any candidate must pass two additional tests: (i) it should be capable of high-pressure alloying with iron, including the ability to depress the melting point by ≥ 1000 K, and (ii) it should partition in sufficient amounts into the low-pressure core-forming liquid iron. As discussed previously, high-pressure partitioning is likely to be inefficient even if it is strongly preferred thermodynamically.

Helium clearly fails the first test (*48*), while oxygen and MgO are doubtful. Ringwood (*49*) has argued for the solubility of oxygen in the core, partly on the basis that FeO becomes metallic at pressures less than those present in the core. This metallization is not clearly indicated by existing shock wave data (*50*), but neither is it definitely excluded (*51*). Metallization of FeO may not be a necessary condition for alloying (although it is surely a sufficient condition); the main requirement is that the specific volume of oxygen must be small. Oxygen has previously been proposed for the core (*52*) but with weaker arguments. The solubility of MgO, suggested by Alder (*53*), is plausible but poorly constrained by existing data. Hydrogen is likely to alloy with iron at high pressure (*54*). The remaining candidates (C, N, Si, and S) form alloys at low pressure and are likely to alloy at high pressure also.

Only sulfur clearly passes the second test for these candidates. Hydrogen, helium, oxygen, and MgO are too insoluble. Carbon and nitrogen are only incorporated in small amounts in iron in any plausible condensation from the solar nebula (*55*). Other more volatile forms of carbon and nitrogen may be outgassed during impact. Silicon is not likely to partition into the iron phase in preference to oxides, except possibly at high pressure (*56*). Ringwood's model (*57*) for an Fe-Si core is not kinetically prohibited, since disequilibrium between core and mantle is very likely, but the very specific accretion and core formation scenario envisaged by Ringwood is not likely to have occurred. (Nevertheless, there is surely some Si in the core, and perhaps not a negligible amount for density considerations.) Sulfur remains as an element for which a low melting point alloy can occur [the eutectic of Fe-S is at about 1000°C for $0 \leq P \leq 55$ kbar (*58*)] and the partitioning is favorable at low pressure. Shock wave results for $Fe_{0.9}S$ and FeS_2 (*59*) suggest that 8 to 10 percent sulfur by weight is sufficient to explain the core density. This is at least a factor of 3 less than the cosmic abundance. Ringwood's arguments against sulfur (*49*) are based in part on the earlier estimate of 15 percent sulfur, a value that may be too large for any model in which the earth accreted only a small fraction of volatile-rich palentesimals.

The probable availability of sulfur, the

Table 1. Thermodynamic properties of the outer core.

Parameter	Liquid iron* at atmospheric pressure and temperature $T_{M,Fe}$	Outer core (theory) P = 1.5 Mbar, $T \simeq 3000$ K	Outer core (theory) P = 3 Mbar, $T \simeq 4000$ K	Theoretical uncertainty
Gruneisen γ	2.44	1.7†	1.6†	± 20 percent
Heat capacity, C_v (k per atom)	4.0	3.8†	3.9†	± 10 percent
Coefficient of thermal expansion, α (K^{-1})	1.22×10^{-4}	1.0×10^{-5}	1.0×10^{-5}	± 30 percent
Electrical resistivity (μohm-cm)	140	160	160	×2
Thermal conductivity (erg/cm-K-sec)	3.22×10^6	4×10^6	4×10^6	×2
Kinematic shear viscosity (cm^2/sec)	3×10^{-3}	8×10^{-3}	8×10^{-3}	×10

*Data from (63) or (93). †Includes electronic corrections, as calculated in (79).

strongly depressed eutectic of Fe-S, and the modest amount required to explain the core density make sulfur a strong candidate for the primary light element in the core. Oxygen may also be important, provided partial melting of the deep, primordial mantle occurs (as discussed above). It should be stressed, however, that there is no reason to believe that the core is a particularly "clean" system. The combined effect on the density of small amounts of Si, C, N, MgO, and H may be comparable to or greater than the effect of S or O.

The Inner Core

Figure 4 indicates that the melting point of pure iron exceeds the actual temperature in the deep core by 1000 to 2000 K. If we suppose that the minor light constituents in the core partition primarily into the liquid phase, so that the inner core is almost pure Fe (or Fe-Ni), then thermodynamic equilibrium between solid inner core and liquid outer core is defined by (60)

$$\mu_0^l(P,T) + kT \sum \ln (1 - C_i) = \mu_0^s (P,T) \quad (3)$$

where μ_0^l and μ_0^s are the chemical potentials of pure liquid and solid iron, respectively, at temperature T and pressure P, k is Boltzmann's constant, C_i is the number fraction of species i in the liquid phase and ideality of mixing is assumed. To lowest order, the melting point is then depressed by an amount $\Delta T = -(kT/\Delta S)\Sigma_i \ln (1 - C_i)$, where ΔS is the entropy of melting. For $\Delta S = k \ln 2$ (61) and $\Sigma_i C_i = 0.15$ (appropriate to 8 percent S by weight, for example) $\Delta T \sim 1100$ K, about the required amount. (It is somewhat smaller than Fig. 4 suggests, but compatible with the large error bars indicated.) It is therefore entirely reasonable to suppose that the inner-outer core boundary represents thermodynamic equilibrium of the liquid outer core with pure, solid iron.

This calculation also assumes that the outer core alloy is more iron-rich than the eutectic composition. If the coexisting solid phase at the eutectic is FeS and if the sulfur-rich fluid also obeys ideal mixing, then the eutectic composition is the simultaneous solution of Eq. 2 with

$$\mu_{FeS}^l (P,T) + kT \ln [4x (1 - x)] = \mu_{FeS}^s (P,T) \quad (4)$$

where $C_i = x$ for sulfur and all other $C_i = 0$ in Eq. 2. If both entropies of melting are k ln 2, then the eutectic composition x_e satisfies

$$(1 - x_e)^{1 - \beta} (4x_e)^{-\beta} = 2^{1-\beta} \quad (5)$$

where $\beta \equiv T_{M,Fe}/T_{M,FeS}$, the ratio of the melting points for pure Fe and pure FeS, respectively. For $\beta = 1.1$, appropriate to the experimentally determined melting points at 60 kbar (58), $x = 0.275$, compared with the experimentally determined value of 0.36. The difference is a consequence of nonideality. Any reasonable extrapolation of the FeS melting curve (62) predicts that $\beta \sim 1$ at the inner core–outer core boundary, in which case the eutectic sulfur fraction substantially exceeds the sulfur content of the outer core. (An analogous calculation is possible for an iron-oxygen alloy.) Even at $\beta = 0.7$ (corresponding to $T_{M,FeS} \sim 8000$ K at 3.2 Mbar), the eutectic sulfur fraction exceeds 8 percent by weight. Any vestige of nonideality only strengthens the argument for an outer core that is more iron-rich than the eutectic composition. Theoretical calculations of the kind discussed below may help constrain the extent of nonideality.

This model for the inner core has important implications for core energet-

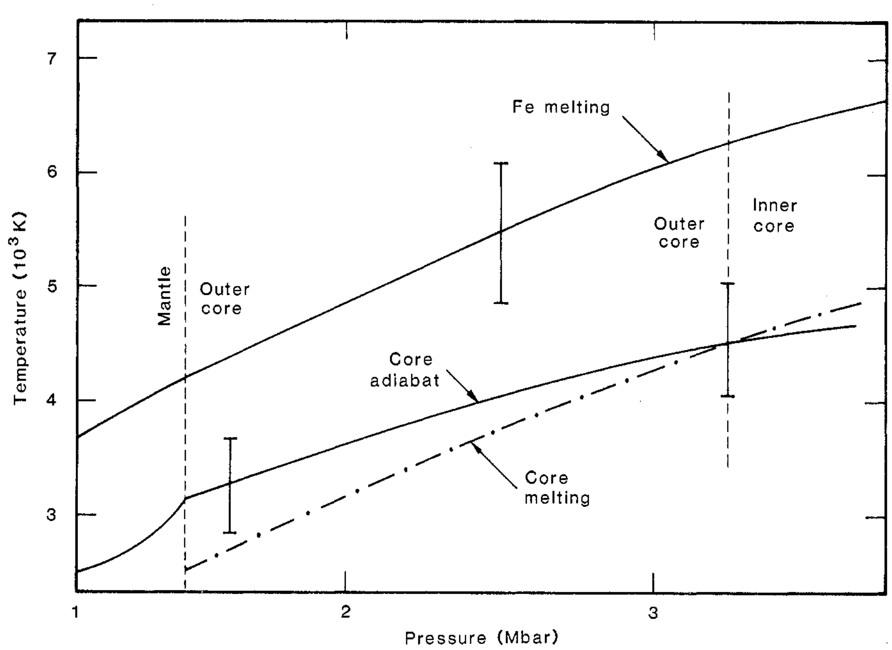

Fig. 4. Temperatures in the core. The pure iron melting curve and its associated uncertainty are from (44). The estimated core adiabat (43) and its associated uncertainty is almost certainly at a substantially lower temperature. The curve labeled core melting represents the thermodynamic equilibrium between pure, solid iron and the outer core alloy and is chosen to cross the core adiabat at the inner core-outer core boundary. Calculation of this curve is discussed in the text.

ics (63). As the core cools over geologic time, the inner core grows and the sulfur content of the outer core increases. A net release of gravitational and latent heat occurs. The gravitational energy release associated with the upward redistribution of the light elements is available for generating the earth's magnetic field. The history of the geomagnetic field may thus be ultimately tied to the history of the earth's inner core (64, 65).

Thermodynamics and Transport Properties of the Outer Core

Any discussion of the outer core must involve the theory of liquids. It is tempting to suppose that the properties of crystalline materials provide an adequate measure of liquid properties at high pressure because of the small difference in volume between phases, but a number of thermodynamic variables, including the thermodynamic Gruneisen γ (66), may be sensitive to the state of the material. I shall discuss here the application of the modern theory of liquids (67, 68) to the outer core.

There are three principal ingredients in the application of this theory. (For simplicity, a one-component fluid is described here, but all aspects are generalizable to multicomponent systems.) First, one needs a pair potential $\varphi(r)$, which is a measure of the pairwise interaction energy between atoms. In a metal, the identification of φ is complicated because of itinerant electrons. Iron is a transition metal and the determination of φ from first principles is intractable (69) but, as explained below, there are reasons to believe that φ can be estimated from existing equation-of-state data. X-ray diffraction experiments (70) also provide constraints. The second ingredient in a liquids theory is the pair distribution function $g(r)$, defined as the relative probability of finding two atoms separated by a distance r. ["Relative" means that $g(r) \equiv 1$ for an uncorrelated system, such as an ideal gas.] As a first approximation, it is common to use the hard-sphere $g(r)$, which is the pair correlation function for an assemblage of rigid spheres (for instance, billiard balls) and is characterized by a single adjustable parameter $\eta \equiv (\pi/6)n\sigma^3$, called the packing fraction, where n is the number density of spheres, each of diameter σ. The hard-sphere system has been extensively studied and its properties are well known (67). The third ingredient in a liquids theory is the statistical mechanical apparatus that relates the microworld of g,φ to the macroworld of measurable thermodynamic variables.

One way to proceed is by applying liquid-state perturbation theory (67), for which the Helmholtz free energy F is given by

$$F = F_{\text{el}}(\rho,T) + F_{\text{hs}}(\eta) + \frac{n}{2}\int \varphi(r)\, g_{\text{hs}}(r)\, d^3r \qquad (6)$$

where el refers to the itinerant electron contribution and hs to the reference hard-sphere system. The choice of the reference hard spheres is governed by the Gibbs-Bogoliubov inequality, expressed as a variational principle

$$\frac{\partial F}{\partial \eta} = 0 \text{ for each } n,T \qquad (7)$$

Assuming that F_{el} is small (71), it is possible to invert the seismic data for a pair potential φ. Figure 3 shows the theoretical equation of state for an inverse power law potential $\varphi \propto r^{-m}$, and $m = 9.5 \pm 0.5$, which provides the best fit to the seismic data. (More complicated potentials have been tried, but with only marginally improved fits.) If the outer core is assumed to be a single-component fluid, then this inversion is constrained by the requirement that the inner core–outer core boundary is at the freezing point and the temperature thereby obtained is 6500 ± 500 K at 3.2 Mbar (72). This is clearly too high for the actual temperature in the core, although it is compatible with the likely melting point of pure Fe (see Fig. 4). This provides further evidence for the necessity of alloying constituents.

Once the best-fitting potential is obtained, all other thermodynamic properties can be estimated. Some trial calculations have been made for binary systems (73), but only the melting point is greatly affected; other thermodynamic variables such as the Gruneisen γ (66) are primarily determined by the best-fitting equivalent single-component φ. It is also possible to use this φ and the corresponding variationally chosen $g(r)$ (or its Fourier space equivalent, the structure factor) to estimate transport properties. All these results are summarized in Table 1. The value of γ is significantly larger than most previous solid-state estimates, but this is not very surprising since liquids (including liquid iron at low pressure) frequently have significantly larger values of γ than the coexisting solid. The atomic contribution to C_v is close to the expected solid value, unlike my previous claim (68), which was based on a cruder (nonvariational) model. The estimate for α follows from the thermodynamic identity $\gamma \equiv \alpha K_T/\rho C_v$, where K_T is the isothermal incompressibility and C_v is the heat capacity. The electrical resistivity is estimated from Ziman's theory, as applied by Jain and Evans (74), but using my variationally determined structure factor. The thermal conductivity is obtained by application of the Wiedemann-Franz relation (75). The kinematic shear viscosity is obtained from the nondimensionalized form of molecular dynamics simulations for inverse power law potentials (76), but is not much different from simpler estimates (77). The shear viscosity is potentially much more uncertain than the other parameters, and the bulk viscosity is even more uncertain (78).

There is another way to compare liquids theory with seismic data. It can be shown (68) that for a fluid outer core of uniform composition

$$\frac{dK}{dP} \sim 5 - \frac{5.6P}{K} \qquad (8)$$

(insensitive to whether the thermal conditions are isothermal, adiabatic, or intermediate). This equation is consistent with the seismic data, all recent tabulations of which give $3 \lesssim dK/dP \lesssim 4.5$ everywhere in the outer core, except possibly in the lowermost 600 km (where the seismic resolution is very poor). Another interesting but less rigorous result can be obtained from liquids theory if it is assumed that the shape of $g(r)$ is invariant along the melting curve. It can then be shown (68) that

$$\frac{d \ln T_M}{d \ln \rho} = \frac{\gamma_a C_{v,a} - k}{C_{v,a} - 3/2\, k} \qquad (9)$$

where T_M is the melting temperature, γ_a is the atomic contribution to the thermodynamic Gruneisen parameter, and $C_{v,a}$ is the atomic contribution to the heat capacity per atom. If $C_{v,a} = 3k$ (the classical high-temperature harmonic oscillator value) then Eq. 9 becomes essentially identical to Lindemann's law (62). This law is the best present procedure for extrapolating melting laws beyond existing data. Assuming the validity of the law, it can then be shown (68, 79) that the melting curve for any single-component system is steeper than the adiabat provided $\gamma > 2/3$. Mathematically

$$\gamma \equiv \left(\frac{\partial \ln T}{\partial \ln \rho}\right)_s < \frac{d \ln T_M}{d \ln \rho} <=> \gamma > 2/3 \qquad (10)$$

Furthermore, $\gamma > 2/3$ is expected for any strongly repulsive potential $[(-d \ln \varphi/d \ln r) \gtrsim 2]$, including any potential compatible with the equation of state of the core. This result essentially disproves the "core paradox" of Higgins and Kennedy (*80*), in which they proposed that the melting curve was actually less steep than the adiabat. Since the inequalities in Eq. 10 are satisfied, there is no problem with the interpretation of the inner core given above. In other words, the core must freeze at the center first rather than at the outer edge.

Dynamic State of the Core

The earth possesses a large magnetic field and the only known process capable of sustaining this field is a hydromagnetic dynamo (*81*). This implies a fluid, metallic region in nonuniform motion and imposes constraints on the thermal and compositional states. The requirement for substantial vertical motions (*82*) strongly suggests that at least a large radial zone of the core is close to adiabatic and compositionally uniform. Partial justification for this assertion is obtained by considering the consequences of its attempted refutation. Suppose the region of dynamo generation were stably stratified (by being either subadiabatic or "bottom heavy" because of compositional stratification). Large-scale vertical motions are then strongly inhibited and no efficient mechanism then exists for generating a poloidal field from a toroidal field. Oscillatory motions may exist, but they are highly inefficient for dynamo generation (*83*). If we define the buoyancy frequency $N \equiv [(-g/\rho)\ \delta(d\rho/dr)]^{1/2}$, where $\delta(d\rho/dr)$ is that part of the density gradient which provides stability (excluding the contribution from self-compression), then a dynamo is difficult to sustain if $N\tau >> 1$, where $\tau \gtrsim 10^3$ years is the characteristic time scale of large-scale core motions or ohmic dissipation. The requirement that $N\tau$ not be large effectively limits the static stability to very small values (for instance, an average deviation from an adiabatic temperature structure of less than about 1 part in a million is required to give $N\tau \lesssim 1$; local and boundary-layer deviations could be much larger).

If it is supposed that the dynamo region is unstably stratified, then a related argument can be made: large-scale motions are uninhibited by viscosity and can grow at an initial rate $\approx \sigma \equiv [(g/\rho)\ \delta(d\rho/dr)]^{1/2}$. This state of instability can be sustained provided $\sigma\tau$ is not too large. (If $\sigma\tau >> 1$, then rapid redistribution of the unstable density distribution would ensue, with the system relaxing toward a state of nearly neutral stability.) As before, average deviations exceeding about 1 part in a million are not likely. This result can also be reached by an application of the mixing-length theory of convection, even with allowance for the complications of magnetic field and rotation (*84*). It should be stressed, however, that these conclusions can be applied only to the region of dynamo generation, not necessarily to the entire outer core. More complex situations can be envisaged, for example, layered convection in a system with opposing gradients of thermal and compositional buoyancy (*85*). This would arise in a core formation model where fluid added later is intrinsically lighter than the earlier core-forming fluid. (Mantle layering could arise for similar reasons.) However, there is no compelling evidence from seismic data or from analyses of fields at the core-mantle boundary for regions of stability in the core (*86*). Stable regions would not be expected to develop in any scenario where core formation led to a vigorously stirred initial state, because growth of the inner core provides both compositional and thermal destabilization (*87*).

The recent realization that the earth is almost certainly cooling down (*88*) has essentially removed any difficulty with generating the geomagnetic field. The presence of intrinsic heat sources such as ^{40}K is neither essential nor likely (*89*) and the inner core has emerged as an essential part of the generation process, at least for the earth (*90*).

The Future

The history of science teaches us that nothing is as simple as it seems from afar. The earth's core is not likely to be an exception. The synthesis presented here should be regarded as a progress report on an area of research that still lacks some crucial data and insight. Some of these deficiencies may be remedied in the next decade.

Four areas can be identified in which future progress is both essential and likely; geo- and cosmochronology, seismological determinations of core structure, fluid dynamics of the core and mantle, and condensed matter physics (especially diamond cell anvil experiments).

Isotope geochemistry and cosmochemistry provide essential data for constraining speculations on the way in which planets are put together, differentiate, and evolve. Extinct radioactivities constrain the time scale of solar system formation (*91*), lead isotope studies constrain the rapidity of core formation (*30*), and studies of the samarium-neodymium system provide tantalizing glimpses of the way in which the earth's mantle may be layered (*92*). None of the inferences is straightforward, but the abundance of data and high precision of the measurement techniques provide optimism for future revelations.

Seismology is another data-rich area in which important information on the core awaits extraction, provided mantle and near-surface structure can be adequately characterized. Free oscillation data are likely to play an especially important role, particularly for constraining the nature of the inner core.

Core dynamics suffers from the current unavailability of geophysically plausible (as distinguished from mathematically elegant) models for generation of the earth's magnetic field. The most plausible energy source is identified and the efficacy of the dynamo process is not in doubt, but the diversity of hydromagnetic processes in a rotating fluid is intimidating. More optimistically, it is likely that we can learn much about the relationships between inner and outer core and between outer core and mantle by experimental and theoretical work on compositionally layered, convecting systems. Work is ongoing in these areas [for instance, see Fearn *et al.* (*65*)].

Last, and perhaps most important, tremendous improvements in our understanding of the behavior of materials at high pressure are anticipated. Static and shock compression experiments provide the capability for a variety of property measurements at a range of temperatures and pressures which include conditions in the earth's core. Phase diagrams can be delineated and the partitioning of constituents determined (either directly in a diamond cell or indirectly by application of thermochemical arguments). Transport properties such as electrical conductivity are also readily measurable. The diamond cell anvil technique, in particular, is still largely untapped and is expected to provide a wealth of thermodynamic data for high pressures in the next decade. Interpretations of the inner core and outer core and their relationship should be greatly improved.

References and Notes

1. S. G. Brush, *Ann. Sci.* **36**, 225 (1979).
2. F. Birch, *J. Geophys. Res.* **57**, 227 (1952); on p. 234: "Unwary readers should take warning that ordinary language undergoes modification to a high pressure form when applied to the interior of the earth, e.g.,

High pressure form	Ordinary meaning
Certain	Dubious
Undoubtedly	Perhaps
Positive proof	Vague suggestion
Unanswerable argument	Trivial objection
Pure iron	Uncertain mixture of all the elements"

3. A. G. W. Cameron, *Moon Planets* **18**, 5 (1978).
4. D. N. C. Lin and J. Papaloizou, *Mon. Not. R. Astron. Soc.* **191**, 37 (1980).
5. A. G. W. Cameron, in *Protostars and Planets*, T. Gehrels, Ed. (Univ. of Arizona Press, Tucson, 1978), p. 453.
6. V. S. Safronov, *Evolution of the Protoplanetary Cloud and Formation of the Earth and Planets* (Nauka, Moskow; Israel Program for Scientific Translations, Jerusalem, 1972).
7. G. W. Wetherill, *Annu. Rev. Astron. Astrophys.* **18**, 77 (1980).
8. ——, *Icarus* **46**, 70 (1981).
9. The temperatures for condensation of iron and of the major silicate phases from the solar nebula are so similar [to within a few tens of degrees; see *(10)*] that efficient separation is almost inconceivable at that stage. Aerodynamic drag may separate small solid bodies according to density and perhaps explain the iron-rich planet Mercury [S. Weidenschilling, *Mon. Not. R. Astron. Soc.* **180**, 57 (1977)], but this is not likely to be relevant for the earth.
10. L. Grossman and J. Larimer, *Rev. Geophys. Space Phys.* **12**, 71 (1974).
11. J. A. Wood, in *Asteroids*, T. Gehrels, Ed. (Univ. of Arizona Press, Tucson, 1979), p. 860.
12. The strength and gravitational binding energy of a frozen iron core within a planetesimal ~ 100 km in radius is small compared with the available kinetic energy for impact velocities exceeding several kilometers per second. The motion of such a body is negligibly retarded by an atmosphere.
13. D. L. Anderson and T. C. Hanks [*Nature (London)* **237**, 387 (1972)] attempt to overcome this difficulty by heating the solid core with an inner core of Ca-Al-U refractories containing radiogenic heat sources. However, such an inner core can spontaneously migrate upward through a solid yet viscous outer core because of its lower density *(23)* without melting the core. Only disseminated ^{26}Al could melt the core since the heat of long-lived radiogenic heat sources can be transported out by subsolidus convection *(88)*. It is unlikely that the earth formed so quickly that ^{26}Al was important *(7)*. There is also no way of adding and mixing a light element subsequently to explain the present fluidity of the outer core. The arguments against extreme heterogeneous accretion thus appear very strong.
14. See, for example, B. M. Jakosky and T. J. Ahrens, *Proc. 10th Lunar Sci. Conf.* (1979), p. 2727. The presence of a blanketing atmosphere may modify this conclusion.
15. W. M. Kaula, *J. Geophys. Res.* **84**, 999 (1979); *Geol. Assoc. Can. Spec. Pap. 20* (1980), p. 25.
16. Application of Darcy's law [see, for example, D. Walker, E. M. Stolper, J. F. Hays, *J. Geophys. Res.* **83**, 6005 (1978)] suggests a melt migration velocity of ~ 0.02 cm/sec, much faster than the accretion process. This calculation assumes ~ 20 percent porosity (generated by the melting of disseminated free iron) and a silicate grain size of 0.1 cm.
17. W. M. Elsasser, in *Earth Science and Meteoritics*, J. Geiss and E. Goldberg, Eds. (North-Holland, Amsterdam, 1963), p. 1.
18. The effective viscous response of any solid medium is a thermally activated process involving an activation energy much larger than the thermal energy per atom. Typical silicate rheology laws [J. W. Tullis, *Rev. Geophys. Space Phys.* **17**, 1137 (1979)] predict $-35 \lesssim d \ln \nu / d \ln T \lesssim -25$, where ν is the kinematic shear viscosity and T is the absolute temperature.
19. Details are unpublished but are a straightforward extension of existing analyses of Rayleigh-Taylor instabilities [see, for instance, H. Ramberg, *Phys. Earth Planet. Inter.* **1**, 427 (1968)] or postglacial rebound for an earth mantle of nonuniform viscosity [C. Brennan, *J. Geophys. Res.* **79**, 3993 (1974)].
20. The length scale L is approximately the thermal diffusion precursor length scale K/v divided by $|d \ln \nu / d \ln T|$, where v is the blob velocity and K is the thermal diffusivity, equal to 10^{-2} cm^2/sec for silicate rock. (The characteristic distance that heat can diffuse in τ years is $\sim 5 \times 10^{-3} \tau^{1/2}$ km. Even for $\tau \sim 10^9$ years, this distance is much less than the radius of the earth. This is one of the reasons why convection is inevitable almost everywhere in all planets.)
21. A. V. Vityazev and S. V. Mayeva, *Izv. Acad. Sci. USSR Phys. Solid Earth* **12**, 79 (1976).
22. D. J. Stevenson, unpublished results. Here K is the liquid iron thermal diffusivity, $\sim 10^{-1}$ cm^2/sec.
23. ——, *Nature (London)* **287**, 520 (1980).
24. Stress levels of $\sim \Delta\rho g h$ develop in such an asymmetry, where $\Delta\rho$ is the density difference between iron and silicates and h is the depth of the iron layer. For $\Delta\rho \sim 4$ g/cm^3 and $g \sim 10^3$ cm/sec^2, this stress exceeds typical strengths of competent materials ($\sim 10^9$ dyne/cm^2) for $h \gtrsim 2$ km. (The material will be stronger than typical undifferentiated meteorites because of the high-pressure hydrostatic environment.)
25. Application of magma fracturing concepts [see, for example, O. H. Muller and M. R. Muller, *Proc. 11th Lunar Planetary Sci. Conf.* (1980), p. 1979] suggests the rapid propagation of long, narrow cracks. However, the iron will freeze because the predicted crack width (~ 10 cm) implies rapid ($\sim$ hours) thermal equilibration with surrounding rock. This time may be too short for efficient downward migration. N. H. Sleep (private communication) has suggested that migration along conduits may nevertheless be very important.
26. G. F. Davies, *J. Geophys. Res.* **85**, 7108 (1980). The energy of core formation is equivalent to a temperature rise of ~ 1500 K for the whole earth or $\gtrsim 5000$ K for the core alone [F. M. Flasar and F. Birch, *ibid.* **78**, 6101 (1973); G. H. Shaw, *Phys. Earth Planet. Inter.* **16**, 361 (1978)].
27. The same calculation as in *(20)* but with $\tau \sim$ accretion time.
28. For a review of thermal boundary-layer theory applied to mantle convection, see D. L. Turcotte and E. R. Oxburgh, *Annu. Rev. Fluid Mech.* **4**, 33 (1972).
29. The transition between glasses or dense suspensions ($\upsilon \lesssim 10^{14}$ cm^2/sec) and bodies with a rigid matrix and permeating melt is poorly understood. Single-component silicates typically have (extrapolated) melting point viscosities of 10^{15} to 10^{16} cm^2/sec *(18)*. See also A. Arzi, *Tectonophysics* **44**, 173 (1978).
30. A. J. Gancarz and G. J. Wasserburg, *Geochim. Cosmochim. Acta* **41**, 1283 (1977); R. T. Pidgeon, *U.S. Geol. Surv. Open-File Rep. 78-701* (1978), p. 334.
31. M. W. McElhinny and W. E. Senanayake, *J. Geophys. Res.* **85**, 3523 (1980).
32. R. Brett, *Rev. Geophys. Space Phys.* **14**, 375 (1976); A. E. Ringwood, *Origin of the Earth and Moon* (Springer-Verlag, New York, 1978), chaps. 3 and 8.
33. Internal pressures within the earth are comparable to the bulk moduli of major earth-forming materials, implying a strong effect of pressure on thermodynamics within the earth.
34. Diffusion in a silicate melt is very strongly temperature-dependent. A value of $D \lesssim 10^{-6}$ cm^2/sec is likely [A. W. Hofmann, in *Physics of Magmatic Processes*, R. B. Hargraves, Ed. (Princeton Univ. Press, Princeton, N.J., 1980), chap. 9]. If the silicates are partially or primarily solid then D is even smaller.
35. P. F. Linden and T. G. L. Shirtcliffe, *J. Fluid Mech.* **87**, 417 (1978).
36. However, the deep mantle is generally the least likely place for partial melting (except, perhaps, in a boundary layer) because the adiabat that characterizes a convective state is less steep than the melting point gradient. Partial melting is thus more likely to be a consequence of pressure release melting near the earth's surface than deep-seated.
37. A. M. Dziewonski and D. L. Anderson, *Phys. Earth Planet. Inter.* **25**, 297 (1981).
38. The seismologist's definition of a liquid does not coincide with the theoretical physicist's definition of a liquid. (For example, liquids can propagate shear waves at sufficiently high frequencies.) The distinction is probably unimportant, but see *(78)*.
39. T. G. Masters, *Geophys. J. R. Astron. Soc.* **57**, 507 (1979).
40. T. J. Ahrens, *Science* **207**, 1035 (1980).
41. Recent isentropic compression of quartz (SiO_2) with a superstrong magnetic field [A. I. Pavlovskii *et al.*, *JETP Lett.* **27**, 264 (1978)] yielded a high-pressure (~ 1.5 Mbar) density consistent with that of the earth's core. This enigmatic result is inconsistent with all existing experimental and theoretical systematics for the high-pressure behavior of materials. If correct, it could provide support for the generally discounted theory that the earth's core is a high-pressure phase transition of mantle material [W. H. Ramsey, *Mon. Not. R. Astron. Soc. Geophys. Suppl.* **5**, 409 (1949)]. There are numerous difficulties with this hypothesis, so that one is tempted to question the experimental results. Independent confirmation or refutation is needed.
42. R. Jeanloz, *J. Geophys. Res.* **84**, 6059 (1979).
43. An independent calculation very similar to that of R. Jeanloz and F. M. Richter [*J. Geophys. Res.* **84**, 5497 (1979)]. Thermodynamic parameters for the outer core are taken from Table 1.
44. J. M. Brown and R. G. McQueen, *Geophys. Res. Lett.* **7**, 533 (1980). They inferred melting in their shock wave experiments from the vanishing of the shear modulus (a decrease of the acoustic velocity at 2.5 ± 0.1 Mbar to a value approximately equal to the hydrodynamic sound speed). The implications of these results are discussed in J. M. Brown and R. G. McQueen, *Adv. Earth Planet. Phys.*, in press.
45. Static high-pressure data on ϵ-iron [H.-K. Mao and P. Bell, *J. Geophys. Res.* **84**, 4533 (1979)] are not available at a sufficiently high pressure or with a sufficient pressure calibration to determine this.
46. R. G. McQueen and S. P. Marsh, *J. Geophys. Res.* **71**, 1751 (1966).
47. Low-pressure solids do not always undergo density reduction when light atoms are added because these atoms may occupy interstitial sites. However, fluids and solids (but especially fluids, which are typically more ideal) are likely to undergo density reduction when any light elements are added at megabar pressures. At such pressures, overlap of orbitals is unavoidable and the atomic radii as conventionally defined cease to have meaning. (Carbon is probably a metal, for example.) The density of the medium is then likely to be primarily a function of mean atomic weight. Shell structure effects are undoubtedly still important, even at pressures ~ 100 Mbar [E. N. Avronin, B. K. Vodolaga, L. P. Volkov, A. S. Vladimirov, V. A. Simonenko, B. T. Chernovolyuk, *JETP Lett.* **31**, 685 (1980)] but are not likely to overwhelm this trend.
48. Application of the calculated pseudopotential for He [D. J. Stevenson, *J. Phys. F* **9**, 791 (1979)] to the earth's core leads to a predicted solubility of $\sim 10^{-3}$ mole fraction.
49. A. E. Ringwood, *Geochem. J.* **11**, 111 (1977).
50. R. Jeanloz and T. J. Ahrens, *Geophys. J. R. Astron. Soc.* **62**, 505 (1980).
51. I. Jackson and A. E. Ringwood, *ibid.* **64**, 767 (1981).
52. V. A. Dubrovskiy and V. Pan'kov, *Izv. Acad. Sci. USSR Phys. Solid Earth* **7**, 48 (1972); K. E. Bullen, *Nature (London)* **243**, 68 (1973).
53. B. J. Alder, *J. Geophys. Res.* **71**, 4973 (1966).
54. D. J. Stevenson, *Nature (London)* **268**, 130 (1977).
55. J. S. Lewis, S. Barshay, B. Noyes, *Icarus* **37**, 190 (1979).
56. R. Brett, *Geochim. Cosmochim. Acta* **35**, 203 (1971).
57. A. E. Ringwood, *ibid.* **30**, 41 (1966).
58. T. M. Usselman, *Am. J. Sci.* **275**, 291 (1975).
59. T. J. Ahrens, *J. Geophys. Res.* **84**, 985 (1979).
60. L. D. Landau and E. M. Lifshitz, *Statistical Physics* (Addison-Wesley, Reading, Mass., 1969), chap. 9.
61. S. M. Stishov, *Sov. Phys. Usp.* **17**, 625 (1975).
62. For example, the thermodynamic version of Lindemann's law [F. D. Stacey, *Geophys. Surv.* **3**, 175 (1979)] with a free-volume Gruneisen γ *(59)*.
63. J. A. Jacobs, *Nature (London)* **172**, 297 (1953); D. E. Loper, *Geophys. J. R. Astron. Soc.* **54**, 389 (1978); D. Gubbins and T. G. Masters, *Adv. Geophys.* **21**, 1 (1979).
64. D. Gubbins, T. G. Masters, J. A. Jacobs, *Geophys. J. R. Astron. Soc.* **59**, 57 (1979); D. J. Stevenson, G. Schubert, P. Cassen, R. T. Reynolds, *Lunar Planet. Sci. Abstr.* **11**, 1091 (1979).
65. D. R. Fearn, D. E. Loper, P. H. Roberts, *Nature (London)* **292**, 232 (1981).
66. The thermodynamic Gruneisen γ is defined as $(\partial \ln T / \partial \ln \rho)_s$, the isentropic (adiabatic) logarithmic change of temperature with density. This is an important dimensionless parameter for calculating temperatures and estimating melting curves *(62)*.
67. J. A. Barker and D. Henderson, *Rev. Mod. Phys.* **48**, 587 (1967).
68. D. J. Stevenson, *Phys. Earth Planet. Inter.* **22**, 42 (1980).
69. Approximations exist, however [W. A. Harrison, *Electronic Structure and the Properties of Solids* (Freeman, San Francisco, 1980), chap. 20]. Band structure calculations [M. S. T. Bukowinski and L. Knopoff, *Geophys. Res. Lett.* **3**, 45 (1976)] are not applicable to the liquid state and do not provide a φ.
70. Y. Waseda and K. Suzuki, *Phys. Status Solidi B* **57**, 351 (1973); Yu. N. Vorob'ev, G. S. Yur'ev, E. A. Klimenkov, B. A. Baum, Yu. A. Bazin, *Izv. Vyssh. Uchebn. Zaved. Fiz.* **1**, 146 (1977).
71. This is a reasonable assumption for some thermodynamic properties [including the compressibility; see *(68)*] but possibly not for some others (for instance, specific heat).
72. Derived from existing data on computer simulations of inverse power law fluids [J. P. Hansen and I. R. McDonald, *Theory of Simple Liquids* (Academic Press, New York, 1976)].
73. Theoretical formulation as described in D. J. Stevenson, *Phys. Rev. B* **12**, 3999 (1975).

74. A. Jain and R. Evans, *Nature (London) Phys. Sci.* **235**, 165 (1972).
75. This states that the electrical conductivity (erg/cm-K-sec) times the electrical resistivity (μohm-cm) is about 2.45 × 10^5 T, where T (K) is the temperature [N. W. Ashcroft and N. D. Mermin, *Solid State Physics* (Holt, Rhinehart & Winston, New York, 1976), p. 20].
76. W. T. Ashurst and W. G. Hoover, *Phys. Rev. A* **11**, 658 (1975).
77. R. F. Gans, *J. Geophys. Res.* **77**, 360 (1972).
78. D. L. Anderson, *Nature (London)* **285**, 204 (1980).
79. J. C. Jamieson, H. H. Demarest, D. Schiferl, *J. Geophys. Res.* **83**, 5929 (1978).
80. G. Higgins and G. C. Kennedy, *ibid.* **76**, 1870 (1971).
81. H. K. Moffatt, *Magnetic Field Generation in Electrically Conducting Fluids* (Cambridge Univ. Press, Cambridge, 1978).
82. F. H. Busse, *J. Geophys. Res.* **80**, 278 (1976).
83. D. Gubbins, *Geophys. Res. Lett.* **2**, 409 (1975).
84. D. J. Stevenson, *Geophys. Astrophys. Fluid Dyn.* **12**, 139 (1979).
85. J. S. Turner, *Buoyancy Effects in Fluids* (Cambridge Univ. Press, Cambridge, 1973), chap. 8.
86. The analysis of K. Whaler [*Nature (London)* **287**, 528 (1980)] is consistent with no convection at the top of the core but is also consistent with geostrophic balance in a convecting, rotating fluid. The analysis of E. R. Benton [*Phys. Earth Planet. Inter.* **20**, 111 (1979)] imposes constraints on upwelling motions with large horizontal length scales.
87. D. R. Fearn and D. E. Loper, *Nature (London)* **289**, 393 (1981).
88. H. N. Sharpe and W. R. Peltier, *Geophys. Res. Lett.* **5**, 737 (1978); D. J. Stevenson and J. S. Turner, in *The Earth, Its Origin, Evolution and Structure*, M. W. McElhinny, Ed. (Academic Press, New York, 1979), p. 227.
89. Partitioning of potassium into the core has some theoretical support [M. S. T. Bukowinski, *Geophys. Res. Lett.* **3**, 491 (1976)], but the observed mantle depletion is shared by several other elements that would not partition into the core fluid and is more probably explained by volatility (*49*). See also M. Somerville and T. J. Ahrens, *J. Geophys. Res.* **85**, 7016 (1980); T. Kaiser, D. Piepgras, G. J. Wasserburg, *Earth Planet. Sci. Lett.* **52**, 239 (1981).
90. The absence of an inner core may explain why Venus has no substantial magnetic field [D. J. Stevenson, *Eos* **60**, 305 (1979)].
91. G. J. Wasserburg, D. A. Papanastassiou, T. Lee, in *Early Solar System Processes and the Present Solar System* (Societa Italiana di Fisica, Bologna, Italy, 1980), p. 144.
92. D. J. DePaolo, *Eos* **62**, 137 (1981).
93. *Handbook of Chemistry and Physics* (CRC Press, West Palm Beach, Fla., ed. 56, 1975).
94. I acknowledge useful comments and corrections from J. M. Brown, S. Cooperman, G. Davies, W. M. Kaula, D. E. Loper, and N. H. Sleep. This work was supported in part by the President's Fund, California Institute of Technology and by NASA grant NGL05-007-002 at the University of California, Los Angeles. Contribution number 3622 of the Division of Geological and Planetary Sciences, California Institute of Technology.

Space Shuttle: A New Era in Terrestrial Remote Sensing

James V. Taranik and Mark Settle

Summary. The space shuttle will carry its first scientific cargo into orbit on its second test flight. The seven experiments to be conducted during this flight—investigations related to continental geology, atmospheric chemistry, meteorology, marine biology, and plant physiology—will demonstrate the potential usefulness of the shuttle for research in the earth and life sciences.

The space shuttle successfully completed its first orbital test flight in April 1981 and ushered in a new era of manned space flight. This new space transportation system possesses the capability to routinely carry men, equipment, and experiments into low earth orbit. The shuttle will be used to deploy and retrieve earth-orbiting satellites, and it will also be used as an orbital laboratory in which highly specialized experiments can be conducted in the weightless and vacuum conditions of space.

Remote sensing studies of the earth, an obvious application of the shuttle's capabilities, have recently been overshadowed by popular interest in industrial and military applications of the system. The shuttle will provide an ideal orbital platform for collecting experimental remote sensing data as well as for testing advanced sensor technology as it becomes available (*1*). In recognition of the potential utility of the shuttle for earth-related research, NASA has placed a high priority on demonstrating the shuttle's capability as a platform for terrestrial remote sensing observations. In 1976 the second orbital test flight of the shuttle was designated as an earth-viewing mission, and an announcement of opportunity was issued to solicit proposals for experiments. Six investigations were selected from a total of 32 proposals. A seventh experiment was added to the mission more recently. These experiments will involve several different types of scientific investigations related to the study of continental geology, atmospheric chemistry, meteorology, marine biology, and plant physiology. The experiments are collectively referred to as the OSTA-1 payload because most are managed by NASA's Office of Space and Terrestrial Applications.

The OSTA-1 payload represents the first scientific cargo that the shuttle will carry into orbit (Fig. 1). Five of the experiments selected for the mission are mounted on an engineering model of the Spacelab pallet manufactured by the European Space Agency (Fig. 2). The pallet, located in the shuttle's cargo bay, weighs 1218 kilograms, and the experiments mounted on it weigh 1016 kilograms and require 1452 watts of electrical power. Pallet experiments can be operated directly by the astronauts or by ground controllers at the Johnson Space Center in Houston, Texas. The other two experiments are located in the crew compartment.

The second shuttle test flight is scheduled for launch in early November from the Kennedy Space Center in Florida. The primary objective of the test flight program is to evaluate the performance and flight characteristics of the shuttle itself. Launch procedures and reentry maneuvers have been changed significantly from the first flight in order to obtain additional engineering flight data. The shuttle will be placed in a 137- to 142-kilometer circular orbit with an inclination of 38° during the period of actual data collection. The orbiter will fly in an inverted position with its cargo bay doors open, facing the earth's surface, for a total of 88 hours during the 5-day (120-hour) mission. All of the scientific data collected during the mission will be removed from the orbiter within 72 hours of its landing at Roger's Dry Lake Bed (NASA's Dryden Flight Research Facility) in California.

James Taranik is the program scientist for the experiments to be conducted during the second flight of the space shuttle and chief of the National Aeronautics and Space Administration's Non-Renewable Resources Program. Mark Settle is the program scientist for the Non-Renewable Resources Branch. Both authors are with the Resource Observation Division, Office of Space and Terrestrial Applications, NASA Headquarters, Washington, D.C. 20546.

0036-8075/81/1106-0619$01.00/0