Theory of the Earth

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Chapter 4. The Lower Mantle and Core

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Abstract:

The lower mantle starts just below the major mantle discontinuity near 650 km. The depth of this discontinuity varies, perhaps by as much as 100 km and is variously referred to as the "650-km discontinuity" or "670-km discontinuity." In recent Earth models there is a region of high velocity gradient for another 50 to 100 km below the discontinuity. This is probably due to phase changes, but it could represent a chemical gradient. The "lower mantle proper" therefore does not start until a depth of about 750 or 800 km. Below this depth the lower mantle is relatively homogeneous until about 300 km above the core-mantle boundary. If there is a chemical difference between the upper and lower mantle then, in a convecting dynamic mantle, the boundary will not be at a fixed depth. This clarification is needed because of the controversy about whether slabs penetrate into the lower mantle or whether they just push down the discontinuity.
core.

4

The Lower Mantle and Core

I must be getting somewhere near the centre of the earth. Let me see: that would be four thousand miles down. I think—

—ALICE

The lower mantle starts just below the major mantle discontinuity near 650 km. The depth of this discontinuity varies, perhaps by as much as 100 km and is variously referred to as the "650-km discontinuity" or "670-km discontinuity." In recent Earth models there is a region of high velocity gradient for another 50 to 100 km below the discontinuity. This is probably due to phase changes, but it could represent a chemical gradient. The "lower mantle proper" therefore does not start until a depth of about 750 or 800 km. Below this depth the lower mantle is relatively homogeneous until about 300 km above the core-mantle boundary. If there is a chemical difference between the upper and lower mantle then, in a convecting dynamic mantle, the boundary will not be at a fixed depth. This clarification is needed because of the controversy about whether slabs penetrate into the lower mantle or whether they just push down the discontinuity.

COMPOSITION OF THE LOWER MANTLE

Several methods can be used to estimate the composition of the lower mantle from seismic data; perhaps the most direct is to compare shock-wave densities at high pressure of various silicates and oxides with seismically determined densities. The shock-wave Hugoniot data must be corrected to adiabats. There is a trade-off between temperature and composition, so this exercise is nonunique. Materials of quite different compositions, say $(Mg,Fe)SiO_3$ (perovskite) and $(Mg,Fe)O$, can have identical densities, and mixtures involving different proportions of MgO, FeO and SiO_2 can satisfy the density constraints. In addition, the density in the Earth is not as well determined as such parameters as the compressional and shear velocities. Nevertheless, many authors have used density alone to argue for specific compositional models for the lower mantle or to argue that the mantle is chemically homogeneous. The density of the lower mantle and the density jump at 650 km are very weak constraints on the chemistry of the lower mantle or the change in chemistry between the upper and lower mantle. Arguments based on viscosity and mean atomic weight are even weaker. The mineralogy of the lower mantle is even harder to determine since oxide mixtures, such as $MgO + SiO_2$ (stishovite), have densities, at high pressure, similar to compounds such as perovskite having the same stoichiometry.

The bulk modulus K_s can be determined by differentiating shock-wave data, $(\rho\partial P/\partial\rho)_s$, but this, of course, is subject to uncertainties. Nevertheless, using both p and K_s in comparisons with seismic data reduces the ambiguities. A more direct comparison uses the seismic parameter Φ, which can be determined from both seismology and shock-wave data:

$$\Phi = \left(V_p^2 - \frac{4}{3}V_s^2\right) = (\partial P/\partial\rho)_s$$

It has been shown that a chondritic composition for the lower mantle gives satisfactory agreement between shock-wave and seismic data (Anderson, 1977). Pyrolite, with 46 percent SiO_2, can not simultaneously satisfy both Φ and ρ. Watt and Ahrens (1982) also concluded that the SiO_2 content of the lower mantle is closer to chondritic than pyrolitic.

Another approach is to extrapolate seismic data to zero pressure with the assumption that the lower mantle is homogeneous and adiabatic (Anderson and Jordan, 1970; Anderson and others, 1971; Butler and Anderson, 1978). A variety of equations of state are available (discussed in Chapter 5) that can be used to fit ρ, G, K_s, V_p and V_s in the lower mantle, and the zero-pressure parameters can be compared with values inferred or measured for various candidate minerals and compositions. Although a large extrapolation is required, there is a large range of compressions available over the extent of the lower mantle, and the parameters of the equations of state can be estimated more accurately than they can over the available range of compressions in most static experiments. The temperature corrections to be applied to the extrapolated lower-mantle values are, of course, uncertain. Butler and Anderson (1978) concluded that pure "perovskite," $MgSiO_3$, was consistent with the seismic data. A range of $(Mg,Fe)SiO_3$ compositions is also permitted because of the uncertainty in the moduli of "perovskite" and lower mantle temperature.

The next approach is to use measured or inferred values of physical properties of various candidate lower-pressure phases (such as perovskite or magnesiowiistite) and to extrapolate to lower-mantle conditions. This method suffers from an extensive reliance on systematics involving analog compounds. Gaffney and Anderson (1973) and Burdick and Anderson (1975) concluded that the lower mantle was richer in SiO_2 than the upper mantle or olivine-rich assemblages. Bass and Anderson (1984) found that pyrolite and $(Mg,Fe)SiO_3$ (perovskite) gave similar results at the top of the lower mantle. The relatively homogeneous part of the lower mantle, however, does not set in until about 800 km depth.

In all of these approaches there is a trade-off between temperature and composition. If the lower mantle falls on or above the 1400° adiabat, then chondritic or pyroxenitic compositions are preferred. If temperatures are below the 1200°C adiabat, then more olivine ("perovskite" plus (MgFe)O) can be accommodated. A variety of evidence suggests that the higher temperatures are more appropriate.

Velocity-density systematics can also be applied to the problem (Anderson, 1970a). There are systematic variations between density, velocity and mean atomic weight $\overline{M}$. The lower mantle has higher Φ and ρ than inferred for the high-pressure forms of olivine and peridotite. This has been used to argue for iron enrichment in the lower mantle. It was later recognized that these systematics could not be applied through a phase change that involves an increase in coordination. An increase in coordination involves a large increase in density but only a small increase, or even a decrease, in seismic velocity (Anderson, 1970b). This weakens the arguments for FeO enrichment in the lower mantle but strengthens the arguments for SiO_2 enrichment.

Attempts to compute velocity throughout the mantle, assuming chemical homogeneity but allowing for phase changes, have not satisfied the seismic data, at least for an olivine-rich composition (Lees and others, 1983). A difference in composition between the upper and lower mantle is implied. The sharpness of the 650-km discontinuity implies either a univariant phase change, for which there is no laboratory evidence, or a compositional boundary. The absence of earthquakes below 690 km is indirect evidence, although inconclusive, for a chemical boundary that prevents penetrative convection.

The seismic velocities in the upper 150 km of the lower mantle exhibit a high gradient. This is probably due to the continuous transformation of garnet solid solution (garnet plus majorite) to "perovskite" and y-spinel to "perovskite" plus (Mg,Fe)O. Reactions involving the ilmenite structure may also be involved.

The mantle between about 800 and 2600 km depth appears to be relatively homogeneous, although a slight increase with depth of FeO may be permitted (Gaffney and Anderson, 1973; Anderson, 1977).

Region D", just above the core-mantle boundary, has a different gradient than the overlying mantle and may contain one or more discontinuities. It is also laterally inhomogeneous, causing scatter in the travel times and amplitudes of seismic waves that traverse it. It may be a region of high thermal gradient and small-scale convection, but its properties cannot be entirely explained by thermal boundary theory. Phase or compositional changes or both are probably involved. There is also some evidence that Q in this region is lower than in the overlying mantle (Anderson and Given, 1982).

D" is a logical site for a chemically distinct layer. Light material from the core can be plated to the base of the mantle, and if denser than the mantle, there it will remain. Chemically dense blobs from the mantle would also settle on the core-mantle boundary. As the Earth was accreting, the denser silicates, as well as iron, would probably sink through the mantle. A basaltic layer at the surface would transform to eclogite at high pressure and could sink to the protocore-mantle boundary, unimpeded by the spinel–post-spinel or garnet-perovskite phase changes until the Earth was Mars-size or larger. This assumes that the perovskite phase change in eclogite occurs at a higher pressure than in Al_2O_3-poor material and the high-pressure form of eclogite is less dense than Al_2O_3-poor assemblages. Subduction today probably cannot provide material to the lower mantle. D" may therefore be the site of ancient subducted lithosphere.

In the inhomogeneous accretion model the deep interior of the Earth would be initially rich in Fe and CaO-Al_2O_3-rich silicates. D" may therefore be more calcium- and aluminum-rich than the bulk of the mantle. At D pressures this may be denser than "normal" mantle (Ruff and Anderson, 1980).

The seismic parameter Φ_o for the lower mantle ranges from about 61 to 63 km^2/s^2, depending on the temperature assumed. For comparison $MgSiO_3$ (perovskite), Al_2O_3 and SiO_2 (stishovite) are 63, 63.2 and 73.7 km^2/s^2, respectively.

(Mg,Fe)O ranges from 40.7 to 47.4 km^2/s^2 for reasonable ranges in iron content. Increasing FeO decreases Φ unless Fe is in the low-spin state (see discussion below). Therefore, it appears that MgO and SiO_2 in approximately equal molar proportions are implied for the lower mantle.

There is a slight drop in Poisson's ratio across the 650-km discontinuity. Temperature and pressure both increase Poisson's ratio in a homogeneous material, so this drop is an indication of a change in chemistry or mineralogy. Increasing the packing efficiency of atoms in a lattice and increasing coordination both serve to decrease the Poisson's ratio (Anderson and Julian, 1969). Spinels and garnets, the major minerals of the transition region, have zero-pressure Poisson's ratios of about 0.24 to 0.27. Stishovite and most perovskites have values in the range 0.22–0.23. The difference is about that observed across the discontinuity. MgO has a very low value, 0.18, and is estimated to be about 0.25 at 670 km. The observed value at the top of the lower mantle is about 0.27.

Two measures of homogeneity are dK/dP and the Bullen parameter (B.P.). These are tabulated in the Appendix for the Preliminary Reference Earth Model (PREM) of Dziewonski and Anderson. In homogeneous self-compressed regions we expect dK/dP to be a smoothly decreasing function of depth and B.P. to be close to unity. These conditions are satisfied approximately between 770 km and 2500 km depth. Velocity gradients are very low in the lower 200 km of the mantle. The region at the top of the lower mantle has high gradients, possibly due to the garnet-perovskite or garnet-ilmenite transitions.

The partitioning of trace elements into a $(Mg,Fe)SiO_3$-perovskite should be evident in upper-mantle chemistry if a deep (greater than 700 km) magma ocean existed or if material from the deep mantle is brought into the upper mantle. The trace-element patterns of the refractory elements can, however, be explained by partitioning between melts and the common upper-mantle minerals. This suggests a chemically zoned planet, formed by a low-pressure zone refining process, and a chemically isolated lower mantle.

CaO and Al_2O_3

According to arguments based on cosmic abundance, the major components of the lower mantle are MgO and SiO_2. CaO and Al_2O_3 are likely to be the next most abundant components, but their concentrations are expected to be low, particularly if the material in the lower mantle has experienced low-pressure melting and removal of the basaltic components. There may be regions, however, such as D", that are enriched in refractories such as CaO and Al_2O_3. CaO and Al_2O_3 have densities and elastic properties similar to those inferred for the lower mantle, and therefore appreciable amounts may be accommodated without affecting the seismic properties. Thus they are essentially invisible and arbitrary amounts can be accommodated. Ca-rich perovskites, however, may have lower Φ than $(MgFe)SiO_3$-perovskite (Ruff and Anderson, 1980) and this may contribute to the low seismic gradients observed in D". The low density of $CaSiO_3$-perovskite, compared to $(Mg,Fe)SiO_3$-perovskite, may prevent Ca-rich material such as eclogite from sinking into the lower mantle. The CaO content of planets constructed from new solar abundances of the refractories (see Chapter 1) is higher than chondritic and the CaO/Al_2O_3 ratio is higher.

$CaSiO_3$ transforms to a perovskite structure with a density about the same or slightly greater than $MgSiO_3$-perovskite (Ringwood, 1975, 1982). At lower pressures $CaSiO_3$ combines with $MgSiO_3$ to form diopside and Ca-garnets. CaO also combines with Al_2O_3 to form compounds at low pressure, but Ringwood (1975) argued that Al_2O_3 will not be accommodated in $CaSiO_3$-perovskite related compounds. CaSiO, $\cdot\, xAl_2O_3$ (garnet) will therefore disproportionate to $CaSiO_3$(perovskite) + xAl_2O_3 at high pressure. This transformation occurs at a much higher pressure than the $CaSiO_3$-perovskite transformation. In spite of the above comments, Liu (1977) apparently synthesized a phase $Ca_2Al_2SiO_7$ related to perovskite with a density of 4.43 g/cm^3. Weng and others (1983), on the basis of high-pressure measurements on the system $MgSiO_3 \cdot$ CaSiO, . Al_2O_3, concluded that CaSiO, forms a separate phase. Shock-wave measurements (Svendsen, 1987) on CaSiO, and $CaMgSi_2O_6$ give high-pressure phases with densities consistent with either mixed oxides or perovskite that have zero-pressure densities of 4.0 to 4.13 g/cm^3. At high pressure the measured densities are considerably less than the lower mantle. There was no evidence for a superdense phase. In fact, CaO-rich material approaches the density of the lower mantle only for high iron contents (about 18 mole percent $FeSiO_3$) (Svendsen, 1987). The disproportionation products of CaSiO, . xAl_2O_3 will be even less dense because of the low density of Al_2O_3. The low inferred zero-pressure density for $CaSiO_3$ from high-pressure shock-wave experiments suggests that CaSiO, has not completely transformed even at pressures as high as 1.8 megabars.

On balance, it appears that $CaSiO_3$-perovskite exists as a separate phase in the lower mantle. At lower pressure CaSiO, forms garnet solid solutions with $(Mg,Fe)SiO_3$. xAl_2O_3 and therefore disproportionation reactions are involved in the formation of $CaSiO_3$-perovskite. Because of the broad stability interval of garnet, the transformation pressure is much higher than for pure $CaSiO_3$. $CaSiO_3$-perovskite, because of its low density relative to $(Mg,Fe)SiO_3$-perovskite and apparently high transformation pressure, does not contribute much to the negative buoyancy of eclogite in subducted slabs. Stishovite has a ρ_0 of 4.29 g/cm^3 and will serve to substantially increase the density of subducted quartz-bearing eclogite, but not SiO_2-poor basalt/eclogite.

Ringwood assumed that the grossularite portion of garnet disproportionates to $CaSiO_3$ (perovskite) plus Al_2O_3 at depths above 670 km. However, this assemblage is less dense than the lower mantle or $(Mg,Fe)SiO_3$(perovskite). This suggests that the basaltic and pyroxenitic portions of subducted lithosphere, and the eclogite cumulates formed in early Earth history are trapped in the upper mantle. Depleted peridotite is also trapped in the upper mantle because of its low density. Subducted slabs will tend to depress a chemical interface at 650 km, and convection will also deform this boundary. In fact, the "650-km" discontinuity may vary in depth by more than 80 km. Depths reported in seismological studies range from 640 to 720 km. The sharpness of the discontinuity is consistent with a chemical discontinuity.

Garnet has an extensive stability field in a silicate of eclogite composition; the transformation from garnet to perovslute is probably not complete until about 750 km. This is much deeper than the transformation in olivine and Al_2O_3-poor pyroxene. Grossularite plus $CaSiO_3$ forms a garnet solid solution that is probably stable to 900 km (Liu, 1979) and is considerably less dense than $(Mg,Fe)SiO_3$ (perovskite), which is stable at shallower depths. The bulk density of eclogite at 650 km is therefore less than the density of the lower mantle. An eclogite layer is gravitationally stable at midmantle depths. Transformations in CaO and Al_2O_3-rich silicates probably contribute to the high velocity gradient found between the 400- and 650-km discontinuities.

It is often assumed that overridden oceanic lithosphere disappears out of the bottom of the Wadati-Benioff zone. Some aspects of continental geology, however, invoke the presence of subducted lithosphere 1000–3000 km into the continental interior (Dickinson and Snyder, 1978). The thermal lifetime of overridden oceanic lithosphere is very long. It heats up with time but cools the adjacent mantle, so that if the slab remains in the upper mantle it should show up as a high-velocity anomaly. If subducted material is trapped in the upper mantle, the western Atlantic and the Brazilian and Canadian shields will be underlain by oceanic lithosphere that represents Jurassic Pacific Ocean. In fact, these parts of the world are in geoid lows and have high upper-mantle velocities. The fate of subducted oceanic lithosphere is intimately related to the problems of whole-mantle versus layered mantle convection and chemical inhomogeneity of the mantle. There is, as yet, no convincing evidence that slabs sink into the lower mantle.

Low-Spin Fe^{2+}

Two alternate electronic configurations, high-spin and low-spin, are possible for Fe^{2+}. The high-spin (H.S.) state is usually stable in silicates and oxides at normal pressures. The ionic radius of the low-spin (L.S.) state is much smaller than the high-spin state, and a spin-pairing transition is induced by increased pressure. A large increase in density accompanies this phase transformation. For example, the volume change accompanying a phase change in Fe_2O_3 at 500 kbar, attributed to the high-spin–low-spin transition, is 11–15 percent.

Gaffney and Anderson (1973) proposed that spin-pairing is likely in the mantle at depths below 1700 km and perhaps at higher levels as well. The small ionic radius of Fe^{2+}(L.S.) probably means that Fe^{2+} will not readily substitute for Mg^{2+} under lower-mantle conditions. Additional Fe^{2+} (L.S.)O-bearing phases will form with high densities and bulk modulus. Assuming that Fe^{2+} spin-pairing occurs below 670 km, Gaffney and Anderson (1973) showed that the lower mantle could be enriched in FeO and SiO_2 relative to the upper mantle. The magnesium-rich phases of the lower mantle may be relatively iron free:

$$MgFeSiO_4 \rightarrow MgSiO_3 \text{ (perovskite)} + FeO(L.S.)$$

which would facilitate the entry of FeO into molten iron and removal to the core.

The possible presence of low-spin Fe^{2+} in the lower mantle complicates the interpretation of seismic data in terms of chemistry and mineralogy. The lower mantle may be chondritic or "solar" in major elements or it may be residual refractory material remaining after extraction of the basaltic elements, calcium, aluminum and sodium. In the latter case it would be expected to be depleted in the radioactive elements, uranium, thorium and potassium. At very high pressure FeO may become metallic and, therefore, readily enter the core.

REGION D"

The lowermost 200 km of the mantle, region D", has long been known to be a region of generally low seismic gradient and increased scatter in travel times and amplitudes. Lay and Helmberger (1983) found a shear-velocity jump of 2.8 percent in this region that may vary in depth by up to 40 km. They concluded that a large shear-velocity discontinuity exists about 280 km above the core, in a region of otherwise low velocity gradient. The basic feature of a 2.75 $\pm$ 0.25 percent velocity discontinuity is present for each of several distinct paths. There appears to be a lateral variation in the velocity increase and sharpness of the structure, but the basic character of the discontinuity seems to be well established. Wright and Lyons (1981) found a rapid increase in compressional wave velocity of 2.5 to 3.0 percent about 200 km above the core-mantle boundary.

D" may represent a chemically distinct region of the mantle. If so it may vary laterally, and the discontinuity in D" would vary considerably in radius, the hot regions being elevated with respect to the cold regions. A chemically distinct layer at the base of the mantle that is only marginally denser than the overlying mantle would be able to rise into the lower mantle when it is hot and sink back when it cools off. The mantle-core boundary, being a chemical interface, is a region of high thermal gradient, at least in the colder parts of the lower mantle.

I argued earlier that neither the peridotitic nor the eclogitic portions of subducted oceanic lithosphere can sink into the lower mantle. However, while the Earth was accreting, conditions would have been more favorable for deep subduction of eclogite. D" may therefore be the repository for ancient subducted lithosphere. Likewise, light material from the core may have underplated the mantle. In either case D" would be more refractory (Ca-, Al-, Ti-rich) than the average mantle.

Because the core is a good conductor and has low viscosity, it is nearly isothermal. Lateral temperature variations can be maintained in the mantle, but they converge at the base of D". This means that temperature gradients are variable in D . In some places, in hotter mantle, the gradient may even be negative in D". Regions of negative shear velocity gradient in D" are probably regions of high temperature gradient and high heat loss from the core.

THE CORE

The core is approximately half the radius of the Earth and is about twice as dense as the mantle. It represents 32 percent of the mass of the Earth. A large dense core can be inferred from the mean density and moment of inertia of the Earth, and this calculation was performed by Emil Wiechert in 1891. The existence of stony meteorites and iron meteorites had earlier led to the suggestion that the Earth may have an iron core surrounded by a silicate mantle. The first seismic evidence for the existence of a core was presented in 1906 by Oldham, although it was some time before it was realized that the core does not transmit shear waves and is therefore probably a fluid. It was recognized that the velocity of compressional waves dropped considerably at the core-mantle boundary. Beno Gutenberg made the first accurate determination of the depth of the core, 2900 km, in 1912, and this is remarkably close to current values. The mantle-core boundary is sometimes referred to as the Gutenberg discontinuity and sometimes as the CMB.

Although the idea that the westward drift of the magnetic field might be due to a liquid core goes back 300 years, the fluidity of the core was not established until 1926 when Jeffreys pointed out that tidal yielding required a smaller rigidity for the Earth as a whole than indicated by seismic waves for the mantle. It was soon agreed by most that the transition from mantle to core involves both a change in composition and a change in state. Subsequent work has shown that the boundary is extremely sharp. There is some evidence for variability in depth, in addition to hydrostatic ellipticity. Variations in lower-mantle density and convection in the lower mantle can cause at least several kilometers of relief on the core-mantle boundary. The outer core has extremely high Q and transmits P-waves with very low attenuation. Evidence that the outer core is mainly an iron-rich fluid also comes from the magnetohydrodynamic requirement that the core be a good electrical conductor.

Although the outer core behaves as a fluid, it does not necessarily follow that temperatures are above the *liquidus*. It would behave as a fluid even if it contained 30 percent or more of suspended particles. All we know for sure is that at least part of the outer core is above the solidus or eutectic temperature and that the outer core, on average, has a very low rigidity and low viscosity. Because of the effect of pressure on the liquidus temperature, a homogeneous core can only be adiabatic if it is above the liquidus throughout. An initially homogeneous core with an adiabatic temperature profile that lies between the solidus and liquidus will contain suspended particles that will tend to rise or sink, depending on their density. The resulting core will be on the liquidus throughout and will have a radial gradient in iron content. The core will be stably stratified if the iron content increases with depth.

Inge Lehmann (1936) used seismic data from the "core shadow" to infer the presence of a higher velocity inner core. Although no waves have yet been identified that have traversed the inner core unambiguously as shear waves, indirect evidence indicates that the inner core is solid (Birch, 1952). Julian and others (1972) reported evidence for PKJKP, a compressional wave in the mantle and outer core that traverses the inner core as a shear wave, but this has yet to be confirmed. Early free-oscillation models (Jordan and Anderson, 1973) gave very low shear velocities for the inner core, 2 to 3 km/s, and some models (Backus and Gilbert, 1970) had entirely fluid cores. More recent models give shear velocities in the inner core ranging from 3.46 to 3.7 km/s (Anderson and Hart, 1976; Dziewonski and Anderson, 1981).

Gutenberg (1957) suggested that the boundary of the inner core is frequency dependent and, therefore, that the inner core might be a highly viscous fluid rather than a crystalline solid. The boundary of the inner core is also extremely sharp (Engdahl and others, 1970). The Q of the inner core is relatively low, and appears to increase with depth.

The high Poisson's ratio of the inner core, 0.44, has been used to argue that it is not a crystalline solid, or that it is near the melting point or partially molten or that it involves an electronic phase change. However, Poisson's ratio increases with both temperature and pressure and is expected to be high at inner core pressures, particularly if it is metallic (Anderson, 1977; Brown and McQueen, 1982). Some metals have Poisson's ratios of 0.43 to 0.46 even under laboratory conditions.

Table 4-1 presents numerous properties of the core.

Composition of the Core

Butler and Anderson (1978) fit a variety of equations of state to the seismic data for the outer core. Third-order finite strain theory was shown to be inadequate, and the best fits

TABLE 4-1
Properties of Core

Symbol	Property	Outer Core	Inner Core	Uncertainty
R	Radius (km)	3480	1221	
P	Pressure (Mbar)	1.36	3.29–3.64	
ρ	Density (g/cm^3)	9.90–12.17	12.76–13.09	
ρ_o		6.6–6.73	7.6	±0.2
K_s	Bulk modulus (Mbar)	6.4–13.0	13.4–14.3	
K_o		1.2–1.4		
G	Shear modulus (Mbar)	<0.02	1.57–1.76	±0.2
K'_o		4.3–4.8	1.76	±0.2
V_p	Compressional velocity (km/s)	8.06–10.36	11.03–11.26	
V_s	Shear velocity (km/s)	–0	3.50–3.67	
V,	Bulk velocity (km/s)	8.06–10.36	10.26–10.44	
V_{ϕ_o}		4.3–4.6		50.35
γ	Griineisen ratio	1.7	1.6	20 pct.
c_p	Specific heat (erg/g.K)	5×10^6		10 pct.
α	Expansivity (K^{-1})	10^{-5}		30 pct.
k	Thermal conductivity (erg/cm.K.s)	4×10^6		×2
σ	Electrical resistivity ($\mu\Omega$cm)	100–160		×2
v	Shear viscosity (cm^2/s)	8×10^{-3}		$\times 10^2$
T_m	Melting temperature (K)	2600–5000	6150–7000	
R,	Magnetic Reynolds number	200–600		$\times 10^2$
	Decay time (years)	15,000		
	Ohmic dissipation (W)			
	Poloidal	10^8		
	Toroidal	10^{10}–10^{12}	10^{11}	
	Heat loss (W)	10^{12}–10^{13}		
	Rotation rate (rad S^{-1})	7.29×10^{-5}		
	Westward drift	0.2°/yr		
	Dipole in core (Wb m^{-2})	3.8×10^4		
H_r	Poloidal field (gauss)	6		
H,	Toroidal field (gauss)	50–2400	$<10^6$	
μ	Permeativity	1		
	Heat of fusion (erg/g)	4×10^9		
	Ekman number	10^{-15}		
	Reynolds number	3×10^8		
	Rossby number	4×10^{-7}		
	Magnetic Rossby number	2×10^{-9}		

Verhoogen (1973), Ruff and Anderson (1980), Stevenson (1981), Jacobs (1975), Dziewonski and Anderson (1981), Melchior (1986), Gubbins (1977).

were obtained for fourth-order finite strain, Bardeen's equation of state and an equation of state involving an exponential repulsive potential. Their best fits for the region between 2200 and 3200 km radius gave the following values for zero-pressure quantities:

$$\rho_o = 6.60\text{–}6.71 \text{ g/cm}^3$$

$$K_o = 1.22\text{–}1.40 \text{ Mbar}$$

$$V_{\rho_o} = 4.30\text{–}4.57 \text{ km/s}$$

$$\Phi_o = 18.5\text{–}20.9 \text{ km}^2/\text{s}^2$$

$$K'_o = 4.5\text{–}4.8$$

These are uncorrected for temperature and therefore represent high-temperature values. Butler and Anderson concluded that a pure iron-nickel core has too high a density and too low a bulk sound velocity to be compatible with the seismic data. A lighter alloying element that increases the bulk sound speed seems to be required. The pressure derivative at K_o at P = 0 is K'_o and this appears to have normal values.

If the ratios of nonvolatile elements in the Earth are similar to those in the Sun and chondritic meteorites, then an iron-rich core is required. Some early workers proposed that silicates may undergo metallic phase changes and that material of high density, high electrical conductivity and

TABLE 4-2
Properties of Iron

Property	Units	Value
ρ_0	g/cm^3	7.02 (liq. at 1810 K)
		8.35 (ε)
α	K^{-1}	11.9×10^{-5} (liq.)
K_0	Mbar	1.40
		0.85 (liq.)
		1.95 (ε)
V_{ϕ_0}	km/s	3.80
γ	—	2.2–2.4
Electrical resistivity	$\mu\Omega$cm	140
Thermal conductivity	erg/cm K s	3.22×10^6
Shear viscosity	poises	3×10^{-3} (liq. at MP)

Ahrens (1979), Jeanloz and Knittle (1986), Stevenson (1981).

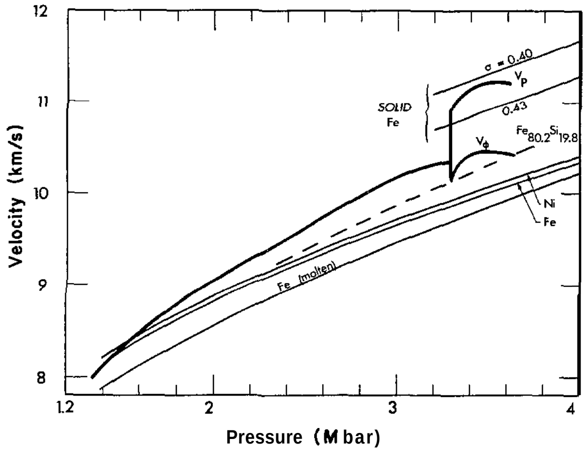

FIGURE 4-2
Compressional velocities (V_Φ) in the outer core and compressional (V_p) and bulk sound speeds (V_Φ) in the inner core (heavy lines) compared to estimates for iron and nickel. Values are shown for two Poisson's ratios σ in the inner core (after Anderson, 1977).

low melting point might be' formed from silicates at high pressure. However, material of sufficiently high density has not been observed in any shock-wave or static-compression experiment on silicates or oxides, and the iron hypothesis is the most reasonable one. Properties of pure iron are listed in Table 4-2.

Figures 4-1 and 4-2 show that the properties of the core closely parallel the properties of iron but that a light alloying element is required that also serves to increase the compressional wave velocity. This alloying element should also serve to decrease the melting point, since the melting point of pure iron is probably higher than temperatures in the outer core. Elements such as nickel and cobalt are likely to be in the core, but if they occur in cosmic ratios with iron

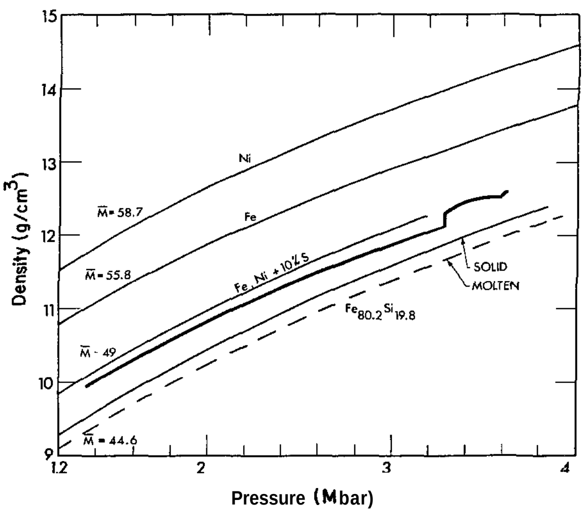

FIGURE 4-1
Estimated densities of iron, nickel and some iron-rich alloys, compared with core densities (heavy line). The estimated reduction in density due to melting is shown (dashed line) for one of the alloys (after Anderson, 1977).

they will not affect the seismic properties and melting temperature very much. Candidate elements should dissolve in iron in order to affect the melting point and to avoid separating out of the core. Material held in suspension could reduce the velocity, but unless the core is turbulent, or the particles are very small, such material would rapidly settle out because of the presumed low viscosity of the core. This mechanism cannot be ruled out completely, because new suspended material may be constantly replenished by convection across the liquidus or by erosion of the lower mantle and inner core.

Candidate materials, based on cosmic abundances alone, are hydrogen, helium, carbon, nitrogen, silicon, magnesium, oxygen and sulfur. The volatiles hydrogen, helium and possibly carbon, nitrogen and sulfur, which form volatile compounds under appropriate conditions, are depleted in the Earth relative even to the amount in the infalling planetesimals because of devolatilization during the accretional process. Silicon and magnesium are likely to partition strongly into the silicate phase, in preference to iron, at core pressure just as they do at low pressure. Some carbon, nitrogen, silicon and sulfur may enter the core since they form iron alloys. Sulfur and oxygen (perhaps as FeO or some other oxide) appear to be the strongest candidates for large concentrations in the core.

Sulfur depresses the melting point substantially (– 1000°C) at low pressure. Shock-wave results indicate that 6 to 12 percent of sulfur can explain the density in the core (Anderson, 1977; Ahrens, 1979). This range has been confirmed by more recent data (Brown and McQueen, 1982). The density of a-iron (7.87 g/cm^3) is much greater than the sulfides of iron; compare, for instance, FeS (troilite), 4.83 g/cm^3; FeS (sphalerite structure), 3.60 g/cm^3; FeS (wiirtzite structure), 3.54 g/cm^3; FeS_2 (pyrite), 5.02 g/cm^3; and FeS_2 (marcasite), 4.89 g/cm^3. The seismic velocities of molten iron-sulfur alloys are unknown; velocities in solid sulfides are greater than in the corresponding metals, but it is not clear if this carries over to the molten state. Pyrite, for example, has a V_p of about 8 km/s, compared to 6 km/s for pure iron. Pyrrhotite has a bulk sound velocity $(\partial P/\partial\rho)_s$, or c, about 20 percent greater than ε-iron at high pressure (Brown and others, 1984). Zero-pressure ultrasonic data on pyrite (FeS_2) give a c_o of 5.23 km/s, which is much higher than the shock-wave speed of 3.45 km/s for pure iron at zero pressure. The bulk sound speeds in such sulfides as CdS and ZnS are 40 to 45 percent greater than in the metal. The approximate zero-pressure bulk sound speed of an FeS-Fe core is 3.9 km/s. For Ringwood's FeSi core the corresponding value is about 4.2 km/s. Butler and Anderson (1978) and Anderson and others (1971) estimated that c_o in the outer core is 4.35 to 5.2 km/s. Anderson and others (1971) estimated 3.1 to 3.7 km/s for shocked iron-nickel alloys with a possible further decrease of 7 to 15 percent to allow for melting. Values estimated for pyrrhotite are 4.4–4.9 km/s (Brown and others, 1984).

Thus, sulfur appears to have the appropriate characteristic to be the light alloying element in the core. Sulfur, however, is a volatile element and will tend to be lost upon accretion. Other volatiles that are unlikely to be sequestered in the core are also depleted in the crust-mantle system relative to carbonaceous chondrites, and it is difficult to argue that sulfur is immune to this depletion process. The mantle is not particularly depleted in chalcophiles relative to other volatiles.

The depletion of sulfur in the crust-mantle system, relative to carbonaceous chondrites, is quite remarkable, roughly 10^{-3}. This is about an order of magnitude more depletion than other volatiles such as thallium, lead, bismuth and indium. The depletion is comparable to that of siderophile refractories such as rhenium, osmium and iridium. It is not clear at this point whether it is primarily the volatile nature of sulfur that prevented it from being accreted by the Earth, or its siderophile nature that allowed it to be removed efficiently to the core as FeS. Some sulfur could have been incorporated into the early Earth as the refractory CaS. In order to explain the density of the core, about 20 percent to 50 percent of the cosmic complement of sulfur must have been retained by the Earth, and this seems excessive considering the depletion of other volatiles.

An Earth composed of cosmic or chondritic abundances gives the proper mantle/core mass ratio if the core composition is about Fe_2O or 50 mole percent FeO. This also gives about the right density for the core. It would be of interest to know if the hypothetical intermetallic compound Fe_2O is stable at high pressure.

Goto and others (1982) estimated values of ρ_o and c_o for a high-pressure phase of Fe_2O_3 of 6.22 g/cm^3 and 6.7 km/s, respectively. This can be compared with the zero-pressure values of 6.6 + 0.15 g/cm^3 and 4.35 + 0.35 km/s estimated by Butler and Anderson for the outer core. Earlier estimates for ρ_o gave 6.4 to 7.2 g/cm^3. Brown and McQueen (1982) obtained $\rho_o = 8.28$ g/cm^3 and $c_o = 4.64$ km/s for ε-iron. The high-pressure form of FeO has an estimated density of 6.7–8.4 g/cm^3 (McCammon and others, 1983). At the core-mantle boundary p and c are approximately 9.9 g/cm^3 and 8.1 km/s. Values estimated for liquid iron at comparable pressures are 10.8 g/cm^3 and 7.5 km/s (Brown and McQueen, 1982). Thus, it appears that iron alloyed with oxygen will have lower density and higher velocity than pure iron.

Reactions such as

$$FeO \rightarrow \frac{1}{3}(Fe_2O_3 + \varepsilon\text{-}Fe)$$

$$FeO \rightarrow \frac{1}{4}(Fe_3O_4 + \varepsilon\text{-}Fe)$$

may be energetically favorable at high pressure and could permit the FeO component of mantle silicates to disproportionate and remove ε-iron to the core. The low-spin transition in Fe^{2+} would favor the creation of separate iron-rich phases (Gaffney and Anderson, 1973), which might then be involved directly in the above reactions. Ahrens (1979) concluded that the density of the core permitted 7–8 percent oxygen, slightly less than the allowable range for sulfur.

If iron-bearing silicates can disproportionate to separate iron-rich phases at high pressure, then it may be possible to form a core without invoking reduction of iron oxides at the surface or having free iron drain through the upper mantle. The presence of siderophiles in the upper mantle and water at the surface both argue against free iron near the surface, at least in the terminal stages of accretion. If fully oxidized material, such as carbonaceous chondrites, accreted to form the Earth, then there must be a mechanism for reducing the high fayalite content of meteoritic olivine to values appropriate for the mantle and, at the same time, preventing the complete stripping of siderophiles from the upper mantle.

One apparent problem with the oxygen-rich core hypothesis is the very limited solubility of oxygen in molten iron at low temperatures and pressures. However, at high temperature and pressure molten iron can dissolve a considerable amount of oxygen. At 2400°C, for example, molten iron can contain 40 mole percent FeO (Ohtani and Ringwood, 1984). At 2800°C molten iron in equilibrium with $(Mg_{0.8}Fe_{0.2})O$ is predicted to contain about 40 mole percent of FeO. Solubility of FeO in molten iron also increases sharply with pressure. The Fe-FeO phase diagram should resemble a simple eutectic system above about 20 GPa (McCammon and others, 1983). The solubility of FeO in molten iron in equilibrium with $(Mg_{0.8}Fe_{0.2})O$ at 2500°C increases from 14 mole percent at P = 0 to 25 mole percent at 20 GPa. Since the core is presumably in equilibrium with the silicates and oxides at the base of the mantle, it is likely that the core contains considerable oxygen. It appears that the core can dissolve enough FeO to explain its low density and to considerably lower its melting point.

Ringwood (1966) rejected hydrogen, helium, carbon, oxygen and nitrogen as important elements in the core because they form interstitial solid solutions with iron and would therefore not decrease the density. The applicability of this argument to molten iron at core pressures and temperatures is obscure, but it led Ringwood to favor silicon as the light element in the core. Ringwood also argued strongly against sulfur. By putting some silicon in the core and vaporizing more silicon in the terminal stages of accretion, he managed to generate an olivine-rich mantle from cosmic abundances. If the core of the Earth is formed *in situ* by reduction, the reaction products, H_2 and CO, plus silicon would form a massive atmosphere totaling more than half the mass of the core, and an efficient dissipation mechanism must be postulated.

Balchan and Cowan (1966) determined the density of shocked iron-silicon alloys at conditions comparable to those in the core and concluded that their results were consistent with a core containing 14 to 20 percent silicon in iron by weight. The zero-pressure, room-temperature densities of these compositions are 7.02 and 7.25 g/cm^3. The zero-pressure bulk sound speed, c_o, of the iron-silicon alloys lies between 4.1 $\pm$ 0.4 (4 percent silicon) and 5.4 $\pm$ 0.1 km/s (20 percent silicon). These values bracket estimates for the core.

There are other possible meteorite-based models for the core. Mixing of 40 percent carbonaceous I, 46 percent ordinary, and 14 percent iron meteorites, for example, yields the proper core-mantle ratio. The mean atomic weight and the zero-pressure density of the resulting core are 50.5 and 6.34 g/cm^3, respectively; the sulfur content is 14 percent by weight. The density will be reduced by up to 5 percent upon melting. The mantle, for this mix, contains 18.4 percent by weight of FeO. Another approach is to reduce some of the FeO and SiO_2 of a carbon-, sulfur- and H_2O-free type I carbonaceous chondrite in order to obtain a mantle composition similar to pyrolite and to obtain the proper silicate/metal or mantle/core ratio. The resulting core has 11 percent silicon by weight, a mean atomic weight of 50.4, and a zero-pressure density of 6.24 g/cm^3; the last two values are very close to those estimated above for the iron-sulfur core. There is no particular reason, however, for postulating a mantle that is deficient in silicon, as in the pyrolite model.

Thus it appears that silicon, oxygen and sulfur all serve to decrease the density and increase the velocity of iron. These estimates are very crude and do not completely take into account phase changes or melting. The point is that the various alloys all have similar physical properties. More shock-wave and static-compression data on mixtures may be able to resolve the possibilities, particularly if accurate values for c_o can be determined. The "chondritic coincidence," the fact that the core is in contact with the mantle, the depletion of the Earth in volatiles and the high solubility of FeO in molten iron at high temperature and pressure all favor oxygen as the major light element in the core. A possible implication is that the lower mantle, in particular region D", may be deficient in FeO. If FeO has been preferentially stripped out of the lowermost mantle, then the parts so affected would be rich in MgO and pyroxene relative to primitive mantle; for example,

$$(Mg_{0.5}Fe_{0.5})_2\ SiO_4 \rightarrow FeO + MgSiO_3$$

$$\text{Iron-rich olivine} \rightarrow \text{Wüstite} + \text{Enstatite}$$

The motivation for placing silicon in the core is that the upper mantle is deficient in silicon relative to cosmic abundances. However, there are magmatic processes for concentrating olivine in the shallow mantle and seismic evidence in favor of a chondritic Mg/Si ratio for the mantle as a whole. The melting point of Fe + S + O, at high-pressure, has not yet been determined. This is likely to be much lower than Fe + O or Fe + S and the core may have a much lower temperature than generally assumed.

The Inner Core

The inner core has a radius of 1222 km and a density about 13 g/cm^3. It represents about 1.7 percent of the mass of the Earth. The density and velocity jumps at the inner core–outer core boundary are large enough, and the boundary is sharp enough, so that the inner core boundary is a good reflector of short-period seismic energy.

There is a jump in V_p at the boundary, but the bulk sound speed $\sqrt{K/\rho}$ is nearly continuous. The increase in V_p may therefore be almost entirely due to the presence of a rigidity term, that is, $V_p = \sqrt{(K + 4/3G)/\rho}$, with no change in composition (Figure 4-2).

Because of the small size of the core, it is difficult to determine an accurate value for density. The main constraint on composition is therefore the compressional velocity. Within the uncertainties the inner core may be simply a frozen version of the outer core, Fe_2O or FeNiO, pure iron or an iron-nickel alloy. If the inner core froze out of the outer core, then the light alloying element may have been excluded from the inner core during the freezing or sedimentation process. An inner core growing over time could therefore cause convection in the outer core and may be an important energy source for maintaining the dynamo.

The possibility that the outer core is below the liquidus, with iron in suspension, presents an interesting dynamic problem. The iron particles will tend to settle out unless held in suspension by turbulent convection. If the composition of the core is such that it is always on the iron-rich side of the eutectic composition, the iron will settle to the inner core–outer core boundary and increase the size of the solid inner core. Otherwise it will melt at a certain depth in the core. The end result may be an outer core that is chemically inhomogeneous and on the liquidus throughout. The effect of pressure on the liquidus and the eutectic composition may, however, be such that solid iron particles can form in the upper part of the core and melt as they sink. In such a situation the core may oscillate from a nearly chemically homogeneous adiabatic state to a nearly chemically stratified unstable state. Such complex behavior is well known in other nonlinear systems. The apparently erratic behavior of the Earth's magnetic field may be an example of chaos in the core, oscillations controlled by nonlinear chemistry and dynamics.

Since the outer core is a good thermal conductor and is convecting, the lateral temperature gradients are expected to be quite small. The mantle, however, with which the outer core is in contact, is a poor conductor and is convecting much less rapidly. Seismic data for the lowermost mantle indicate large lateral changes in velocity and, possibly, a chemically distinct layer of variable thickness. Heat can only flow across the core-mantle boundary by conduction. A thermal boundary layer, a layer of high temperature gradient, is therefore established at the base of the colder parts of the mantle. That in turn can cause small-scale convection in this layer if the thermal gradient and viscosity combine to give an adequately high Rayleigh number. It is even possible for material to break out of the thermal boundary layer, even if it is also a chemical boundary, and ascend into the lower mantle above D''. The lateral temperature gradient near the base of the mantle also affects convection in the core. This may result in an asymmetric growth of the inner core. Hot upwellings in the outer core will deform and possibly erode or dissolve the inner core. Iron precipitation in cold downwellings could serve to increase inner-core growth rates in these areas. These considerations suggest that the inner-core boundary might not be a simple surface in rotational equilibrium.

The orientation of the Earth's spin axis is controlled by the mass distribution in the mantle. The most favorable orientation of the mantle places the warmest regions around the equator and the coldest regions at the poles. Insofar as temperatures in the mantle control the temperatures in the core, the polar regions of the core will also be the coldest regions. Precipitation of solid iron is therefore most likely in the axial cylinder containing the inner core.

There are two processes that could create a solid inner core: (1) Core material was never completely molten and the solid material coalesced into the solid inner core, and (2) the inner core solidified due to gradual cooling, increase of pressure as the Earth grew, and the increase of melting temperature with pressure. It is possible that both of these processes have occurred; that is, there was an initial inner core due to inhomogeneous accretion, incomplete melting or pressure freezing and, over geologic time, there has been some addition of solid precipitate. The details are obviously dependent on the early thermal history, the abundance of aluminum-26 and the redistribution of potential energy. The second process is controlled by the thermal gradient and the melting gradient. The inner core is presently 5 percent of the mass of the core, and it could either have grown or eroded with time, depending on the balance between heating and cooling. Whether or not the core is thermally stable depends on the distribution of heat sources and the state of the mantle. If all the uranium and thorium is removed with the refractories to the lowermost mantle, then the only energy sources in the core are cooling, a growing inner core and further gravitational separation in the outer core.

In the inhomogeneous accretion model the early condensates, calcium-aluminum-rich silicates, heavy refractory metals, and iron accreted to form the protocore (Ruff and Anderson, 1980). The early thermal history is likely to be dominated by aluminum-26, which could have produced enough heat to raise the core temperatures by 1000 K and melt it even if the Earth accreted 35 Ma after the Allende meteorite, the prototype refractory body. Melting of the protocore results in unmixing and the emplacement of refractory material (including uranium, thorium and possibly ^{26}Al) into the lowermost mantle. Calculations of the physical properties of the refractory material and normal mantle suggest that the refractories would be gravitationally stable in the lowermost mantle but would have a seismic velocity difference of a few percent.

Depending upon the available heat energy, the iron core could have been either completely or partially molten at the time of unmixing. Therefore, the present solid inner core could be remnant solid iron (or iron-nickel) from the segregation event, or it may have grown through geologic time from the precipitation of the solid phase from the fluid core.

Ruff and Anderson (1980) proposed that aluminum-26 dominated the early thermal history and that long-lived radioactive heat sources are distributed irregularly in the lowermost mantle and drive the fluid motions in the core that are responsible for the geodynamo. The anomalous lower-mantle velocity gradient suggests chemical inhomogeneity and/or a high thermal gradient. The seismic evidence for lateral variation at the base of the mantle is evidence for either variable temperature or varying composition. A new driving mechanism, differential cooling from above, was proposed to sustain the dynamo.

The lowest seismic velocity regions of the lowermost mantle are preferentially located in the equatorial regions. If these are due to high temperature, then downwellings in the outer core will be preferentially located in high latitudes where the lowermost mantle appears to be coldest. Lateral variations in D'' temperature, temperature gradient and radioactivity probably control the pattern of convection in the core, even if the dynamo is not driven from above.

Eventually, one would hope to see similarities in lower-mantle tomographic maps and maps of the magnetic field. Temperature differences in D" and at the top of the core may also generate contributions to the magnetic field by the thermoelectric effect.

MANTLE-CORE EQUILIBRATION

Upper mantle rocks are extremely depleted in the siderophile elements such as cobalt, nickel, osmium, iridium and platinum, and it can be assumed that these elements have mostly entered the core. This implies that material in the core had at one time been in contact with material currently in the mantle, or at least the upper mantle. Alternatively, the siderophiles could have experienced preaccretional separation, with the iron, from the silicate material that formed the mantle. In spite of their low concentrations, these elements are orders of magnitude more abundant than expected if they had been partitioned into core material under low-pressure equilibrium conditions. The presence of iron in the mantle would serve to strip the siderophile elements out of the silicates. The magnitude of the partitioning depends on the oxidation state of the mantle. The "overabundance" of siderophiles in the the upper mantle is based primarily on observed partitioning between iron and silicates in meteorites. The conclusion that has been drawn is that the entire upper mantle could never have equilibrated with metallic iron, which subsequently settled into the core. Various scenarios have been invented to explain the siderophile abundances in the mantle; these include rapid settling of large iron blobs so that equilibration is not possible or a late veneer of chondritic material that brings in siderophiles after the core is formed. The trouble with the latter explanation is that the siderophiles do not occur in the mantle in chondritic ratios, although they are not fractionated as strongly as one would expect if they had been exposed to molten iron. Some groups of siderophiles do have chondritic ratios.

Brett (1971) took another look at this problem. He argued that the iron-rich liquid involves the system Fe-S-O and looked at the partitioning of several metals (Co, Cu, Ni, Ga and Au) between this liquid and olivine and basaltic melts (Table 4-3). The calculated abundances for the silicate phase were remarkably close to upper-mantle abundances, and thus it appears that protocore material could have been in equilibrium with the upper mantle. Further, a protocore containing sulfur and oxygen seems likely.

Since the upper-mantle siderophile abundances fit a local equilibration model, the implication is that the upper mantle has not been mixed with the rest of the mantle since core formation. The partition coefficients depend on temperature, pressure and oxidation state, and it is unlikely that they are constant throughout the mantle. This is relevant to the question of whole-mantle versus layered-mantle convection and the chemical isolation of the lower mantle from the upper mantle. Since core formation was an early process, the implication is that subsequent convection did not homogenize the mantle. When a larger number of siderophile elements is considered, the original problem reemerges.

The highly siderophile elements (Os, Re, Ir, Ru, Pt, Rh, Au, Pd) have high metal-silicate partition coefficients and therefore strongly partition into any metal that is in contact with a silicate. These elements are depleted in the crust-mantle system by almost three orders of magnitude compared to cosmic abundances but occur in roughly chondritic proportions. If the mantle had been in equilibrium with an iron-rich melt, which was then completely removed to form the core, they would be even more depleted and would not occur in chondritic ratios. Either part of *the* melt remained in the mantle or part of the mantle, the part we sample, was not involved in core formation and has never been in contact with the core. Many of the moderately siderophile elements (including Co, Ni, W, Mo and Cu) also occur in nearly chondritic ratios, but they are depleted by about an order of magnitude less than the highly siderophile elements. They are depleted in the crust-mantle system to about the extent that iron is depleted. These elements have a large range of metal-silicate partition coefficients, and their relatively constant depletion factors suggest, again, that the upper mantle has not been exposed to the core or that some core-forming material has been trapped in the upper mantle.

It is not clear why the siderophiles should divide so clearly into two groups with chondritic ratios occurring among the elements within, but not between, groups. The least depleted siderophiles are of intermediate volatility, and very refractory elements occur in both groups.

TABLE 4-3
Partitioning Between Sulfide Melt and Silicates

M	$M_{sulfide}/M_{silicate}$ (1)	$M_{Fe}/M_{silicate}$ (2)
Ni	150–560	1700
Cu	50–330	330–50
Co	7–80	200
Ga	4	—
Ge	—	1000
Re	2×10^3	$\sim 10^5$
Au	10^4	$\sim 10^5$
W	100	—
Ir	400	—
Mo	10^5	—
P	200	—
Ag	250	—
Pb	16	—

(1) Brett (1984), Jones and Drake (1985).
(2) Ringwood (1979).

THE MAGNETIC FIELD

The magnetic fields of planets and stars are generally attributed to the dynamo action of a convecting, conducting core. The study of the interaction of a moving electrically conducting fluid and a magnetic field is called *magnetohydrodynamics.* Magnetic fields entrained in a conducting fluid are stretched and folded by the fluid motion, gaining energy in the process, and thus acting as a *dynamo,* a device that converts mechanical energy into the energy of an electric current and a magnetic field. A moving conductive fluid can amplify a magnetic field. The dynamo mechanism does not explain how the magnetic field originated, only how it is amplified and maintained in spite of the losses caused by the dissipation of the associated current. Fluid motions of the conducting liquid in the presence of the magnetic field induce currents that themselves generate the field. The fluid motions may be due to a variety of causes including precession, thermal convection and chemical convection.

The magnetic field can be visualized as lines of force, the closed loops along which a compass needle aligns itself. The strength of the field in any given volume can be represented by the density of lines in the volume. One may regard the field lines as being "frozen" into the conducting fluid or attached to the particles of which the fluid is composed. The field moves with the fluid, and the stretching of the field lines corresponds to a gain in strength of the field. The energy of the motion of the particles is converted into the energy of a magnetic field, and induced electromotive forces drive the current associated with the field.

The first requirements for a magnetohydrodynamic dynamo are the presence of a magnetic field and an electrically conducting fluid capable of supporting the currents associated with the field. The second requirement is a pattern of fluid motion that amplifies the magnetic field. The naturally occurring combination of *nonuniform rotation* and *cyclonic convection* seems to be particularly effective since these occur in planets, stars and galaxies, all of which can exhibit magnetic fields. A rotating body containing a convecting fluid exhibits differential rotation and cyclonic convection.

The dipolar magnetic field of the Earth is associated with circular electric currents of about 2×10^9 amperes flowing from east to west in the molten iron core. Local anomalies in the field, having dimensions of several thousand kilometers and amplitudes of about 10 percent of the main field, change slowly with time, drifting westward at about 20 km per year. This surface drift rate corresponds to a fluid velocity at the surface of the core of about a meter per hour. The nondipole field is generally attributed to a dozen or so cells in the core. The most obvious explanation for the slow rotation of the core is the action of the Coriolis force on the rising and sinking fluid in the convective cells. The conservation of angular momentum requires that the angular velocity of the rising fluid decrease as it moves further from the spin axis. Therefore, the surface of the core rotates faster at high latitudes than at low latitudes and the inner part of the core rotates faster than the surface.

The primary magnetic field in the core is an east-west field, at right angles to the mean component of the field at the surface. It is called the *azimuthal* or *toroidal jield;* the part lying in the planes through the axis is called a *meridional field.* The azimuthal field is created by the stretching of the north-south lines of force of the dipole field as they are carried around in the rotating fluid of the core. The part of a field lying near the axis is carried around further than the parts lying away from the axis; this nonuniform rotation stretches the north-south lines in an east-west direction. As the field lines are carried around, the azimuthal field gets stronger. The amplification continues until it is balanced by the tension of the magnetic lines of force or the resistive decay of the associated electric current. The azimuthal field in the core may be hundreds of times stronger than the dipole field observed at the surface, perhaps 100 gauss or more. The dipole field observed at the surface of the Earth is therefore a secondary effect of the azimuthal field, which is shielded from view by the insulating mantle.

In the 1930s T. *G.* Cowling proved that fluid motions cannot generate a perfect dipole field or any field with rotational symmetry about an axis. However, cyclonic convection can generate a dipole field. Cyclonic motion raises and rotates the lines of force of the azimuthal field, deforming them into helixes. Intermittent cyclonic convection generates a net dipole field. The essential ingredient for the generation of a field is that the motion of the fluid be helical with the field rotating about its direction of motion as it streams along.

A constraint on the terrestrial dynamo is that it must amplify the dipole field at a rate high enough to balance the decay of the field by ohmic dissipation. The magnetic field in a current-carrying body decays in a characteristic time that is proportional to the conductivity times the cross-sectional area. The strength of the magnetic field is determined by the number of times the field lines can be wrapped around the Earth in their lifetime.

Another property of the Earth's magnetic field is its ability to reverse its polarity abruptly in 1000 years or less, at apparently random intervals of about 10^5 to 10^7 years. Reversals of the magnetic field might be caused by sudden increases in the velocity of convection in the core. This in turn might be triggered by convection in the mantle, through instabilities in the thermal boundary layer at the base of the mantle or changes in core-mantle coupling caused by convection-induced irregularities in the shape of the boundary. There is some evidence that magnetic field variations are correlated with plate tectonic and magmatic events. Reversals might also be the result of the intrinsic nonlinear behavior of the core: nonperiodic chaotic behavior.

Due to the mathematical difficulties in treating the complete dynamical system, dynamo models derived for the terrestrial and astrophysical magnetic fields are generally kinematic models. The kinematic approach neglects the equations of fluid motion and heat transfer and considers just the hydromagnetic equation,

$$\frac{\partial \mathbf{B}}{\partial t} = \overline{\nabla} \times (\mathbf{V} \times \mathbf{B}) + \eta^{-2} \mathbf{B}$$

where B is the magnetic field, V is the fluid velocity, and η is the magnetic diffusivity. A particular velocity field is prescribed along with an initial magnetic field, and a regenerative solution to the above equation is then sought. This approach has yielded several successful models (see Levy, 1976 for a review). The successful velocity fields found vary from large-scale nearly axisymmetric motions to small-scale turbulence with a particular statistical nature. Cyclonic fluid motions with a radial component of velocity have appeared in several dynamo models (Parker, 1983, Levy, 1972) and in the limit of small length scale can be likened to the turbulent dynamo model. The cyclonic model also has the capability of producing a self-reversing dynamo (Levy 1976), an important observed feature of the Earth's magnetic field. Although other velocity fields can produce a dynamo, the cyclonic model is particularly pertinent to models driven by differential heating from above.

Kinematic models can thus describe velocity fields necessary for a dynamo, but they do not indicate the source of fluid motions. These are usually assumed to result from thermal or chemical convection within the core. Efforts toward a dynamic treatment including thermal convection have produced a few results, notably the "convective rolls" dynamo of Busse (1964) and the Rossby wave dynamo of Gilman (1969) as extended by Braginsky and Roberts (1973).

One strong constraint on the geodynamo is that adequate energy be supplied to maintain the magnetic field. Due to ohmic losses, energy must be supplied to the magnetic field through the velocity field. Since the magnetic field has existed at nearly the same intensity for at least 2.7 billion years (McElhinny and others, 1968) and the decay time for the fundamental mode of the magnetic field has been estimated at about 10,000 years (Cox, 1972), there has seemingly been a near-constant energy supply over geologic time.

Gubbins reviewed the energy requirements of the magnetic field and provided lower and upper bounds on the energy supply. The upper bound is of order 10^{20} erg/s, which is the observed surface heat flux, and the lower bound is 2×10^{17} erg/s by consideration of conduction and electric currents. This requires an energy source acting over geologic time of considerable size. The precessional dynamo has been eliminated on the basis of energy constraints (Rochester and others, 1975). Latent heat released from the supposed growth of the inner core is marginal as an energy supply, but this mechanism produces motions restricted to near the inner core (Verhoogen, 1973) unless the precipitation mechanism discussed in previous sections is operative. The secular variations of the magnetic field require substantial fluid motions in the outermost core (Elsasser, 1946).

The only other potential energy sources are radiogenic heating and, possibly, gravitational mechanical stirring. It was in the context of searching for an energy source that potassium-40 was suggested to be in the outer core (Lewis, 1971). This suggestion is rather arbitrary and is not consistent with any known differentiation process. The observational evidence argues against significant potassium in the metallic phase at low pressures. This issue is still controversial, but aside from whether or not potassium would partition into the metallic phase is the problem of the amount required. Murthy and Hall (1972) required three-fourths of the potassium within the Earth to be segregated into the metallic core. To partition that amount of potassium into the core is inconsistent with any accretion and evolution model for the Earth.

The idea of a mechanically stirred core has been suggested. The basic idea is that the inner core has grown continuously over the age of the Earth by precipitating Fe and Ni, excluding the lighter element from the inner core. This process releases a lighter fraction near the inner core boundary, which then causes fluid motions. If the inner core has grown with time and if there is a compositional difference between the inner and outer core, this process may well occur. However, it is not clear that it would be important for the magnetic field, particularly if the core is stratified. The quantitative calculation of the potential energy release (Loper and Roberts, 1977) assumed an adiabatic temperature gradient throughout the core over geologic time, and this assumption conflicts with many recent results, including those of Gubbins. Any stability within the core, even if only in the outermost part, seriously affects the gravitational energy available for fluid motions. An alternative to this model for inner core growth is continual freezing out of metallic iron from a sub-liquidus outer core, which then sinks to the inner core (Figure 4-3). This only works if core compositions are on the Fe side of the eutectic composition or if an intermetallic compound such as Fe_2O or FeNiO is stable at high pressure near the liquidus temperature.

At present there is no consensus on the energy source, or on details of the fluid motions. A considerable advance would be made if the topographies of the outer- and inner-core boundaries could be mapped and if lateral seismic velocity variations in D" and the outer core could be mapped. Seismic tomography is relevant to these questions.

Of the terrestrial planets, Earth and probably Mercury possess substantial intrinsic magnetic fields generated by core dynamos, while Venus and Mars apparently lack such fields. Thermal history calculations suggest that sulfur must be present in the core of Mercury if it is to be molten and capable of sustaining a dynamo.

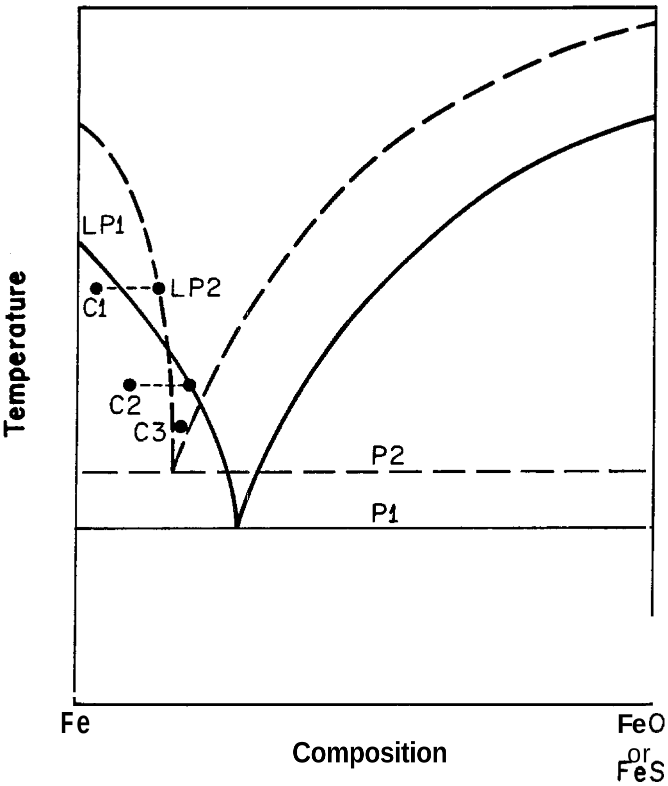

FIGURE 4-3
Possible eutectic phase relations for the core at two pressures. Three possible core compositions are shown (C1, C2 and C3). P1 and P2 are solidus curves for low and high pressures; LP1 and LP2 are the corresponding liquidus curves. For C1 the core is closer to the liquidus at low pressure; the reverse is true for C2. For C3 the core is above the liquidus at high pressure. Depending on the composition and the effect of pressure on the phase relations, one can have the solid content of the core increase or decrease with depth. If the solid particles become large enough they will settle out, giving a compositionally stratified core that may be gravitationally stable or unstable. An adiabatic temperature gradient may alternate with chemical homogeneity.

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