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A Decade of Progress in Earth's Internal Properties and Processes

Orson L. Anderson

Summary. A major component of the Inter-Union Commission on Geodynamics Project, labeled "Internal Properties and Processes," included certain experimental and theoretical research in tectonophysics, seismology, geochemistry, petrology, volcanology, and planetology. This review focuses on a few research areas in which there have been surprises and reversals. In particular, attention is given to the attempts to quantify the thermal profile in the earth's interior and the material properties of the earth's interior.

One decade has passed since the Inter-Union Commission on Geodynamics (ICG) Project was launched, and 7 years ago the U.S. Program for the Geodynamics Project was published (*1*). A major component of this program, which has received national and international attention, was labeled "Internal Properties and Processes." It is appropriate to review the progress that has been made in this component of the ICG now that the Geodynamics Project has ended.

A comprehensive account of the progress in tectonics, seismology, and geochemistry appropriate to this component of the ICG is found in the U.S. National Report to the International Union of Geodesy and Geophysics, which is prepared every 4 years (*2*), and so I will not attempt a detailed review of that here. Instead, I will focus on a few research areas in which there have been surprises or in which there has been an overturning of the wisdom of the early 1970's.

The U.S. Program for the Geodynamics Project (*1*) listed several problems in "Internal Properties and Processes" that loomed in the early 1970's. The major headings were thermal structure, dynamical models, material properties in the deep interior, a theory of rheology, and the nature of instabilities in the deep interior. Of these problems, the thermal structure [in particular, the temperature (T) distribution of the earth] was selected as critical and worthy of accelerated and focused effort: "Temperature is probably the most important parameter concerning the state of motion of the earth's interior" (*1*, p. 32).

The Thermal Regime

In 1970, future research on the thermal state required, or so it seemed, reliable ways of estimating the T profile. In other words, quantifying the unobservable internal T profile appeared to be an essential and feasible preliminary step in the task of interpreting the thermal regime.

That view is now no longer commonly held by those researching the thermal regime. Many now hold to the view that T is of secondary importance and that T has receded to the background as a "determined rather than a determining characteristic of the thermal regime of terrestrial planets" (*3*, p. 2). How this view changed is discussed below.

By 1970 many geophysicists no longer adhered to the long-entrenched belief that the thermal structure of a planet is controlled by heat conduction. Heat conduction theory imposed two insurmountable difficulties. First, the time (τ) required for a planetary-sized object formed from some initial T distribution to reach equilibrium is $\tau = R^2/k$, where R is the radius and k is the thermal conductivity. For the earth, τ is $\sim 10^{18}$ seconds, or longer than the age of the earth. Most of the original heat could not have dissipated, and for contemporary geodynamic processes a planetary-wide heat conduction theory means no significant heat flow from the interior while the process is occurring. The second difficulty is that the exact boundary conditions at the origin of the cooling cycle must be specified, and thus the earth's thermal regime is thereby tied closely to cosmology as well as to the radiogenic heat production (H).

A convection theory circumvented these difficulties. A substantial amount of heat is moved by mass motion in a convecting stream, and τ can be made small enough to be in consonance with

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the period of geological events. The convection process tends to give a planet a "fading memory" of past events, and so the present thermal regime is not rigidly fixed to cosmology. The boundary conditions of a convection theory of the thermal regime can be tied to present plate motion instead of to the complete specification of planetary initial conditions at the time of formation of the earth.

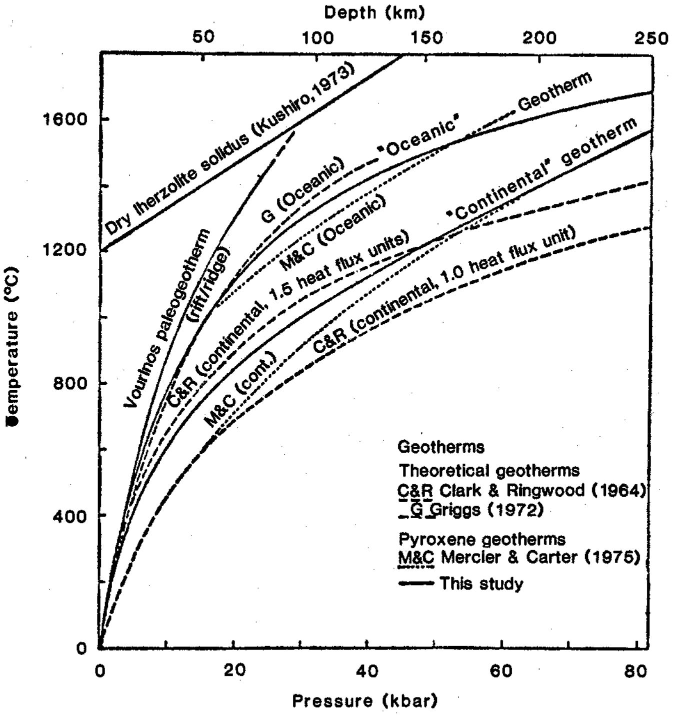

Fig. 1. Geotherms of the lithosphere computed and assembled by Mercier with emphasis on pyroxene geotherms (*6*).

But, although the earth's interior is thought to be in a state of convection, the earth's outer shell—the lithosphere—is still a thermally conducting medium. "Steady heat conduction is not satisfactory for planetary objects over a few hundred kilometers in radius . . . and above that size a [conductively controlled] rise in T makes a central region so fluid that it becomes unstable" (*3*, p. 17). The bottom of the lithosphere can be defined as the boundary between convective and conductive heat flow. The depth of the lithosphere then depends upon the size of the planet and the vigor of the convection.

Little was done in the past decade to define the thermal profile of the earth in its convective regions until 1977. Up to then, attention was focused on a description of the thermal regime in the lithosphere itself. Two different approaches were used. In the geophysical method, surface heat flow was used, and, by accounting for the radioactive rocks at the surface, researchers were able to find reliable methods of extrapolating T to about 100 kilometers. Typical of this approach is the work of Lachenbruch and Sass (*4*). The second method, that of petrology, came to be known as geothermometry and geobarometry. Typical of this work is that of Boyd (*5*). A recent summary of the geotherm as calculated from the compositions of coexisting pyroxenes is shown in Fig. 1 (*6*). The main result of these calculations is confirmation, largely supported by plate tectonics theory, of the idea that the thermal profile of the lithosphere depends very much on the tectonic regime at the surface.

The problem in computing T in the convective parts of the earth hinged on defining an elusive parameter, called the Grüneisen parameter (γ), as a function of depth. In an adiabatic thermal state, which is close to the conditions required by convection, the T gradient is related to the density gradient by the simple expression

$$dT/T = \gamma(d\rho/\rho) \qquad (1)$$

where ρ is the density.

The definition of γ as a function of volume (V) and T has occupied theoretical physicists for many years. A number of formulas have been derived. None of these formulas has been universally accepted, and some theorists have been pessimistic about any formula that is both sound theoretically and useful for computing γ. Two different formulas for γ have been used for computing T, both of which could be defined in terms of the parameters reported in seismic solutions of the earth. The two formulas used were presented by Irvine and Stacey (*7*) [see also (*8*)] and by Anderson (*9*). These formulas are equivalent to the empirical law

$$\rho\gamma^q = \text{const.} \qquad (2)$$

where q is near but not necessarily equal to unity, so that

$$d \ln T/d\rho = \gamma_0\rho_0{}^{1/q} \cdot \rho^{-(1/q\ +\ 1)}$$

where the zero subscript refers to values at the surface, and the T gradient in the convective parts of the earth is then defined in terms of three parameters: γ_0, ρ_0, and q. Anderson (*9*) showed that, because of the restricted range of compression, a lack of sensitivity in q exists when it is specified in terms of seismic results. He found that $0.8 < q < 2.2$ in the lower mantle, but he recommended $q = 1$. The formula derived by Irvine and Stacey, when evaluated on the basis of the seismic data of the earth's lower mantle, showed q to be in the lower range, perhaps slightly less than 0.8.

In order to find the T profile itself, Eq. 1 is integrated. But then an integrating constant is needed. The value of T at some depth within the convective zone is required. To obtain this integrating constant, one has to appeal to experimental results on a solid-state transition, one that can be identified as corresponding to a seismically determined transition. Two such transitions have been proposed, and, as a result, a number of T values for the earth's interior have recently been presented (*10, 11*). Stacey (*10*), using the experimental results on the compression of iron-sulfur systems to extrapolate to the P corresponding to the inner core–outer core transition, fixed T there (3900°C). Others (*11*) have used the experimental results of the phase transition of peridotite, found at pressures corresponding to the seismic transition zone (380 to 670 km), to fix T at 670 km (1960°C). These results are based on the experiments of Akaogi and Akimoto (*12*) and of Liu (*13*) on peridotite. Both these transitions are subject to criticism when used as the basis for the integrating constant because they depend upon an assumption about chemical composition in the mantle and in the core.

These recent results of attempts to define the T profile of the mantle are surprisingly close when one considers that the data and the theory of the various methods are quite different. The main result is that the d ln $T/d\rho$ is not high, and consequently the value of T at the mantle side of the core-mantle boundary is 3000 K or less. Thus, a large jump in T at the thermal boundary must be invoked if the T profile of the core is to resemble the properties of iron or iron diluted by a lighter element. This requires a thermal boundary with a large T jump at the core-mantle boundary (*11*) if the physics of the core and the mantle are to be simultaneously satisfied. A large change in T across the thermal boundary in turn requires a large heat flux passing from the core to the mantle. It is doubtful whether further work on γ can substantially improve the recent qualitative results on the thermal gradient in the convective parts of the earth. Substantial new progress on the T profile more likely depends on an improved understanding of the thermal boundary at the core-mantle interface and on an improved understanding of the heat-generating process in the core.

The basic issue, however, is whether the T profile must be quantified before the mantle convection can be quantified. Tozer (*14*) has stoutly maintained that such quantification of T is not necessary and that T is a derivative property of the conditions of flow; now many others share his view. Tozer's hypothesis is that flow in the convecting parts of the earth is self-regulated and that the strain rate is nearly constant and controlled by a critical value of the viscosity η, represented as η^*, which is near 10^{22} poises. This self-regulation arises because η is such a rapidly varying function of T. Tozer's argument is summarized as follows.

The heat-conducting process in the earth is so slow as compared to the convecting heat flow that, when convection stops, heat flow effectively stops and the interior T rises. "If the temperature inside the planet were initially 'high,' the associated viscosity would lead to a rapid convecting loss of heat, and a fall of T would slow the rate of cooling by increasing η. Alternately, if the T were initially 'low,' and convective motion inhibited by an enormous η, the interior would tend to warm up to quite high T" (*3*, p. 16). This self-regulation of convective flow inside a planet arises because of the enormous dependence of η upon T. The self-regulating value of η^* depends upon the material state: η^* would be more in an anhydrous rock than in a hydrous rock of the same composition.

There are several general properties of planets, which have been discussed by Tozer (*3*), that can be understood in terms of this theory. One is that the depth of the lithosphere is related to planet size. If η^* is common to terrestrial planets and the radiogenic heat production is similar, then the thickness of the lithosphere (d) depends on the pressure (P). That is, η^* is reached at the same pressure in each planet. For example, assume that d is 150 km in the earth, or about P = 4 gigapascals. At 4 GPa in the moon, d would be about 1000 km. This depth agrees quite well with the observations of deep moonquakes (*15*), which have been interpreted as shear fractures (*16*). This may explain why several identifiable plates occur on large terrestrial planets whereas smaller planets have only one plate. Another consequence is that the central T of a planet is substantially independent of H. The value of dT/dZ, where Z is depth, of the lithosphere is affected by H, but η^* is independent of H and therefore the central T profile is independent of H (*17*).

In 1975, McKenzie and Weiss (*18*) showed that the strong T dependence of η prevents large-scale changes in the mantle T over geologic time. Although this is substantially true, a more detailed examination showed that the average T of the mantle may decrease by 50° to 180° K per billion years (*19*) and that the heat content of the earth may be decreasing

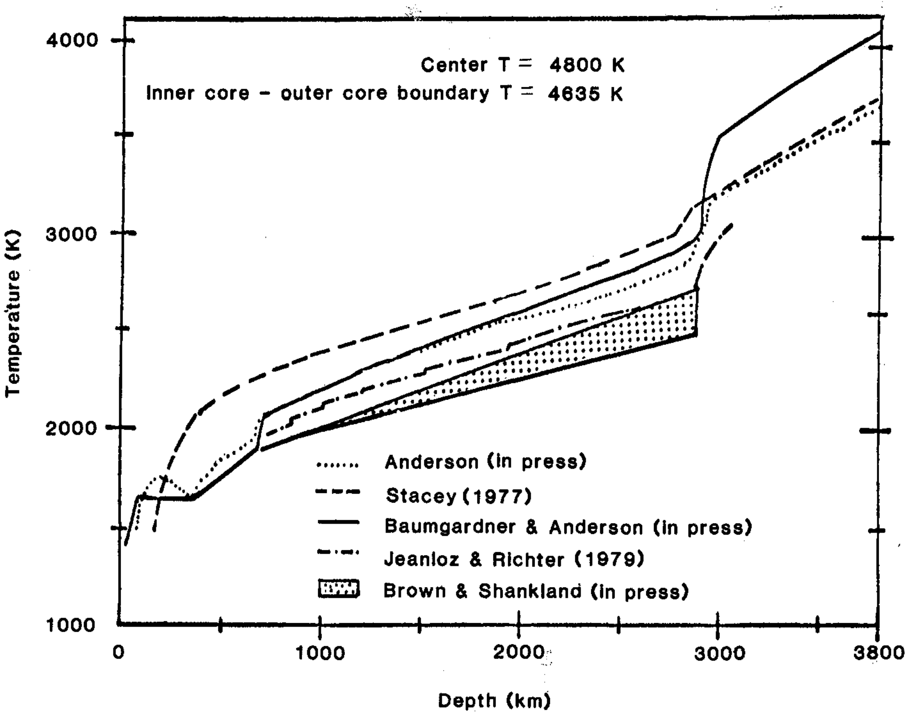

Fig. 2. Recently computed geotherms of the lower mantle. Sources are listed in (*10, 11, 17*).

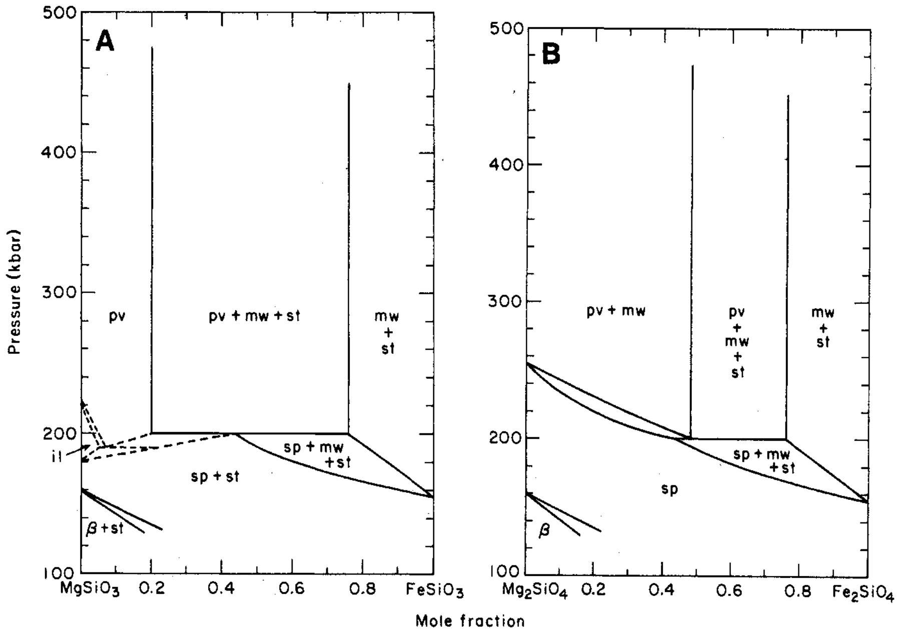

Fig. 3. (A) Pseudobinary diagram for pyroxene composition. Abbreviations: *pv*, perovskite; *mw*, magnetowüstite; *st*, stishovite; *sp*, spinel; β, β phase (of olivine composition); and *il*, ilmenite. Reproduced from (*25*). (B) Pseudobinary diagram for olivine composition; temperature = 1000°C. Reproduced from (*25*). [Courtesy Carnegie Institution of Washington, Washington, D.C.]

by 2 to 4 percent per billion years. This means that the convective speed of the mantle was only slightly larger in the remote past than now (*20*).

Tozer's hypothesis of constant η^* runs counter to several cherished views of the mantle. Four examples of such views follow. (i) The value of T in the mantle is close to the solidus, T_s; according to the η^* hypothesis, T can never exceed about 0.8 T_s. (ii) Convection is limited to the upper part of the mantle. (iii) The details of the physical parameters in a creep law (the relation between rate of strain and stress and temperature) count heavily in understanding the thermal engine of the earth. (iv) Details of the thermal origin and paleothermal history of the earth determine its present T distribution. According to the η^* hypothesis, the planet has a "fading memory" of its thermal origin. Figure 2 shows a plot of T versus Z in the mantle, comparing several recent theories.

Material Properties of the Deep Interior

High-pressure mineralogy. The U.S. Program for the Geodynamics Project (*1*, p. 155) defined the problem of this research area as follows: "What limits can be placed on the composition, mineralogy, and temperature of the earth's crust and mantle by comparing geological observations with laboratory measurements in relevent materials?" Measurements of the lattice constant versus P in minerals have rapidly advanced in the past decade. In answer to the above question, the experimentalists are now able to investigate a given hypothesis of the chemical composition of the upper mantle.

In 1973 it was suggested that a continuing experimental and theoretical program on the elastic constants of minerals, rocks, and aggregates be launched with minimum parameters of at least 30 kilobars (3 GPa) and 1000°C (*1*). Because of the explosive growth of research that far exceeded expectations early in the decade, materials are now measured up to 100 GPa and to over 2000°C (*21*). These spectacular advances are due largely to the use of the so-called diamond anvil pressure cells (*22*) coupled with laser technology and advanced x-ray technology borrowed from experimental physics. As a result, specific chemical models of the mantle can be tested to see if they conform to the ρ profiles required by seismic modeling. For example, Liu (*23*) has eliminated mixed oxides derived from olivine as a candidate material for the lower mantle; he has proposed instead peridotite in the perovskite structure (90 percent $(Mg_{0.9}Fe_{0.1})SiO_3 \cdot 10$ percent Al_2O_3). Mao *et al.* (*24*) also have discounted mixed oxides below 670 km and propose instead a perovskite structure with a small amount of iron plus an iron-rich magnetowüstite. Yagi *et al.* (*25*) presented the phase diagrams of pyroxene and olivine extended to 50 GPa (Fig. 3).

Research in this area is moving rapidly, and we may expect a narrowing of possibilities of mantle candidate chemistry. Currently, the important questions are the following: (i) What is the role of calcium compounds in mantle composition and structure? Analysis of shock-wave results on CaO indicate that the high-pressure B2 phase has a density such that calcium compounds are candidates for components of the lower mantle (*26*). (ii) Is the transition zone (380 to 670 km) a result of phase transitions, chemical interactions between minerals, or compositional changes (*23*)? (iii) Is there a significant effect of gravity separation upon mineral phases at the transition where phase separation takes place? And (iv) with regard to (ii) and (iii), is eclogite the major component in the lower part of the upper mantle (*27*)?

Equations of state. In addition to encouraging programs on elastic constants, the U.S. Program for the Geodynamics Project suggested that a better knowledge (theoretical and experimental) of γ is required (*1*). In 1973 a difficulty prevented the accumulation of equation-of-state (EOS) data on many minerals pertinent to geodynamic problems. High-quality crystals of the requisite size for experimentation were too often not available. The U.S. Program recommended consideration of a national facility for growing crystals. Two experimental developments circumvented this difficulty.

The application of Brillouin scattering to measure the velocities of sound in the diamond cell enabled researchers to measure very small crystals, thereby greatly increasing the catalog of information on the bulk modulus (K_0) (*28*). Even the single-crystal elastic constants of a tiny grain of stishovite have been measured in this way (*29*). The two most important constants for an EOS are K_0 and ρ_0. The other parameter is $K_0' = (\partial K_0/\partial P)_T$, which is sometimes measured in the diamond cell.

The application of the rectangular parallelepiped resonant method (RPR) has made it possible to measure the elastic constants of small crystals against T over a wide T range (*30*). Recently, the measurements on MgO by RPR have been extended so that elastic constants are now known from −200° to 1000°C (*31*). In principle, the T data acquisition problem is now resolved. A new bonus for EOS catalog data has been the resurgence of interest in the measurement of the thermal expansivity of minerals at high T by Suzuki (*32*).

There has been some progress in understanding γ in the last decade. Two theoretical formulas for γ are in current use (*7, 9*), but fundamental objections have been raised against both of them and these objections are not yet resolved. However, recent calculations of γ for the mantle and core from these two formulations agreed to within about 15 percent, so that computations of thermal gradients by Eq. 1 do not yield greatly different answers in terms of the two formulas for γ. Thermodynamic studies of the earth, based on current estimates of γ, seem justified in view of other much larger uncertainties.

On the experimental side, an important advance in the measurement of γ has been made by Kennedy and his coworkers (*33*). They measured γ directly by adiabatic decompression, using the simple formula

$$\gamma = T/K_T(\partial P/\partial T)_S \qquad (3)$$

where K_T is the isothermal bulk modulus at P and T, and S signifies constant entropy. This important experiment is very promising. The isotherms in the V–P plane for the two solids measured so far (NaCl and LiF) up to 3 GPa and 500°C (the limits of measurement) are parallel. It is easy to show that parallel isotherms occur when the thermal pressure (P_{TH}) is independent of V. This results in a significant simplification in the general EOS.

There are thermodynamic equivalents to the case of parallel isotherms, which allow preliminary tests of P_{TH} on V for other minerals. One of them is

$$(\partial K_T/\partial T)_V = 0, \text{ where } T > \theta \qquad (4)$$

(where θ is the Debye temperature). The other is the case for $q = 1$ in Eq. 2. High T elastic constant data measured by RPR and high T thermal expansivity data allow Eq. 4 to be tested. Anderson (*34*) carried out this test and concluded that, for many dense minerals important to geodynamic problems, Eq. 4 holds, and therefore $P_{TH} \neq f(V)$. This finding justified the assumption that the simplified EOS of the Hildebrand type may be appropriate for planetary interiors

$$P(V,T) = P_1(V) + P_2(T) \qquad (5)$$

where $P_1(V)$ is the isothermal EOS evaluated at $T = 0$; the well-known Birch-Murnaghan EOS is an example. The quantity $P_2(T)$ is P_{TH} and can be computed from the approximation $P_{TH} =$

$aT + b$ (34). Equation 5 was used to compute the T profiles of the planets with reasonably good results (17); the T profile so computed for the earth is presented in Fig. 2. The value of P_{TH} has been computed from Debye theory, and its value changes from about 12 GPa to about 15 GPa throughout the lower mantle (35). These values of P_{TH} were also found by direct inversion of the seismic data of the parametric earth model (PEM) (coupled with the use of the Birch-Murnaghan EOS) (36).

An important advance in the theory of the EOS has been the work of Bukowinski who, using fundamental quantum mechanics, has calculated energy (E) versus V for iron, potassium, and a few simple oxides. The value of P is found from E by a simple derivative. This new approach is very promising for the future of the EOS.

Rheology and fracture. In this important research area there has been a great deal of experimental progress because of its relevance to earthquakes. Excellent reviews are available (38), and so I will mention only a few points briefly.

Rheology has two aspects, mantle rheology and lithosphere rheology. The latter is characterized by lower T and very high η ($\eta \gg 10^{22}$ poises). Experimental measurements of strain rate ($\dot{\epsilon}$) in the laboratory indicate that $\dot{\epsilon} = c\sigma^n$, where σ is shear stress, n is ~ 3, and c is a quantity that varies exponentially with T, which is a measure of the activation energy. Tullis (39) has reviewed the experimental situation and the parameters for various cases of power law creep. The details of these varieties of power law creep, although very important in the upper parts of the earth, have not been applied to the deep mantle as much as was expected in the early days of the Geodynamics Project. Kolstadt *et al.* (40) have shown that one creep law—thermal-activated movement of dislocations—has Newtonian behavior at quite low strain rates and that is in agreement with the results of glacial rebound studies.

As an example of glacial rebound theory, Cathles (41) concluded that η is close to 10^{22} poises and a near constant throughout the mantle, except for a possible near surface low-viscosity channel where it may be less. Glacial rebound is often regarded as the most significant evidence that the earth's mantle can flow. Small variations in the gravity field, supported by satellite geodesy, indicate that the shear stresses supported by the convecting mantle are only a few tens of bars (42). Thus, after a decade, the arguments are still persuasive that the rheology of the mantle is essentially Newtonian (although slightly nonlinear) and that η is virtually constant throughout the lower mantle.

Dynamical Modeling

The U.S. Program for the Geodynamics Project (1, p. 43) recommended that "an intensive effort be made to stimulate coordinated work by applied mathematicians, numerical analysts, and geophysicists on the solution of realistic models of the circulation of matter in the earth's interior." Results in this research area have been gratifying, especially in the last half of the decade. Excellent reviews of the mantle convection have been published recently (43), and so my comments are limited to a few general remarks.

The lithosphere is the thin thermal boundary layer on the convecting cell in the earth's mantle. The plates on the earth's surface are part of the convecting cell and are thus very important boundary conditions for the dynamical model.

For most of the last decade, solutions to dynamic problems were restricted to limited parts of the globe or were nonspherical, two-dimensional models. Just recently, however, models have been developed that are spherical and global in extent and that correctly include the boundary conditions of the surface plates. The model of Hager and O'Connell (44) is an example of this general approach.

Great progress is certain to come as soon as these generalized models are perfected. Results from the general model of Hager and O'Connell lend support to the idea that convection cells extend through the entire mantle. Even so, the question of whether the upper convection cell extends only to the 670-km discontinuity is not settled, and there is some evidence that mantle phase changes may impede vertical flow through the phase boundary. The isotopic evidence indicates that there are layered sets of convection cells (45).

Instabilities

The low-velocity zone. The U.S. Program for the Geodynamics Project (1, p. 30) stated: "It is generally agreed there is a minimum in shear velocity that occurs near the lithosphere-asthenosphere boundary. There are many indications this minimum results from a zone of partial melting—what are the homogeneities . . . where is the bottom of the low velocity channel?" In the last decade there has been a pronounced shift in the consensus of what causes the low-velocity zone (LVZ). In the early 1970's the majority held that the minimum in shear velocity was due to a partial melting of the rock (46). A minority held that it was due in large part to an abrupt change in the thermal gradient near the top of the upper mantle (47).

The argument in favor of partial melting relied heavily on the observed low Q (high seismic wave absorption) in the LVZ, which was postulated as arising from acoustic absorption in the melted grains of the rock. But now the necessity for having a partial melt in the upper mantle is no longer demanded by the attenuation models of the LVZ. The current view is that the low Q probably arises from subsolidus solid-state absorption processes (48). Two views have recently been advanced for the LVZ. One is that the minimum in shear velocity (v_s) results from a sharp knee in the T profile, as the thermal regime changes from the conductive mode in the lithosphere to the convective mode in the asthenosphere (49). Another is that an acoustic absorption band associated with the subsolidus mechanisms of the low Q channel affects the v_s behavior (50). Further work in this field is presently inhibited by the lack of resolution in the seismic details of the LVZ.

Melting. An adequate theory of fusion seemed remote in 1973, although many researchers supported the Lindemann law of fusion (51), which was derived on the basis of a naïve lattice model. Two advances have given a good basis of support to the Lindemann law, which now appears to be based on assumptions far broader than given in the original derivation. Stacey and Irvine (52) derived a melting law by using a clear-cut thermodynamic basis, including the Clapeyron relation. Their result is very close to the Lindemann formula, but it has the important advantage of being based on thermodynamic equilibrium and not just on a theory of instability.

The Lindemann law, expressed in terms of γ, is

$$d \ln T_m / d \ln \rho = 2(\gamma - 1/3) \qquad (7)$$

where T_m is the temperature at melting. The Stacey and Irvine formula for solids is the same as Eq. 7 except that the 1/3 term is replaced by the term $\Delta V / V(1 + \gamma) = a$, where $\Delta V/V$ is the volume change upon melting (a few percent);

$$d \ln T_m / d\, ln\, \rho = 2(\gamma - a) \qquad (8)$$

In comparing Eq. 8 with Eq. 1, we see that the adiabatic gradient equals the melting gradient when $\gamma = 2a$. For $\gamma > 2a$, the T in a convective-flow region rises more slowly with depth than T_m.

Thus, the T of the deep mantle should be far below the T of the solidus if the current values of γ in the mantle are reasonably correct (*7, 9*).

Stevenson (*53*), using liquid theory, derived a formula for melting which also greatly resembles the Lindemann law:

$$d \ln T_m / d \ln \rho = (\gamma C_V - k)/(C_V - \frac{3}{2} k) \tag{9}$$

where k is the Boltzmann constant and C_V is the heat capacity at constant volume. Equation 9 approaches Eq. 7 as C_V approaches the classical value $3k$. It seems clear that the Lindemann law, with slight modifications, will become part of the future thermal theories of the earth's mantle and core.

The experimental details of the melting process seemed clearer in 1973 than they do now. It was known that CO_2 and H_2O were quite soluble in silicate liquids at pressures equivalent to those below 50 km (*54*). Consequently, it was assumed that at the beginning of melting the melting point was substantially depressed because of the presumed volatile content of the upper mantle (*55*), an idea that was convenient at the time that the LVZ was attributed to partial melting. The separate and combined effects of H_2O, CO_2, and other volatiles on melting and other phase transformations are now fairly well understood (*56*). The amount of the T depression depends upon the details of the chemical composition, including especially the volatiles. However, it is an obvious oversimplification to adopt a single geochemical model for the entire mantle, particularly because of the heterogeneous distribution of volatiles, alkalies, iron, and other components (*57*).

Solid-solid transition. The need for an adequate theory of solid-solid transitions was emphasized in the U.S. Program for the Geodynamics Project (*1*). But now only a little progress can be reported. In 1973 it was generally supposed that solid-solid transitions were associated with the disappearance of a shear-elastic constant at high P. Further work disclosed that the transitions, at least in simple diatomic ionic solids, take place when a particular shear constant is not zero but is small. The transition takes place when the shear elastic constant (c_{ij}) is a given fraction of the bulk modulus (K) ($c_{ij}/K = 0.15$ to 2.0) (*56*). Experimental confirmation of this criterion was reported, for example, for CaO (*58*).

On the experimental side the progress has been quite rewarding. Of special significance is the measurement by Akaogi and Akimoto (*12*) of the phase equilibrium in garnet lherzolite and in particular the stability field of olivine in its various phases. In this work they found that the pressure corresponding to the first jump in v_s at the transition zone, occurring at 380 km, has a corresponding T of 1400°C. This T is associated with the $\alpha + \gamma > \alpha + \beta$ transition (olivine to modified spinel structure). Thus for the peridotite model of the mantle a good T fix is 1400°C at the beginning of the transition zone.

Mao *et al.* (*24*) showed that iron partitions between perovskite and magnetowüstite in such a way that the iron content of pyroxene and olivine in the lower mantle has definite limits. Mao and his colleagues and Liu (*23*) have made a convincing case for the perovskite structure of $(Mg,Fe)SiO_3$ in the lower mantle. This is one of the most exciting results of the decade. The edge of research in this area now concerns the possibility of gravitational separation of perovskite and magnetowüstite in the lower mantle and the possibility of transitions or chemical reactions to phases denser than perovskites in the deep parts of the lower mantle.

Deep Structure

The contributions of seismological research to the understanding of the earth's deep interior have been reviewed elsewhere (*59*). Mention will be made here of a seismic solution that has been particularly useful to the theories of thermal structure described in this article. The PEM (*60*) has been frequently used for its density, velocity, and pressure distributions. They have been the primary data in many thermal models.

The PEM smooths out the seismic data so that thermodynamic manipulation is eased, but at the same time evidence for heterogeneity is obscured. Today there is mounting evidence that there may be several subtle transitions in the mantle, and further work along this line is under way (*61*). The resolution of possible lateral inhomogeneities is one of the most challenging and urgent problems facing seismologists.

The idea that the mean atomic weight ($\bar{m}$) is constant throughout the lower mantle has not been seriously challenged in the past decade. The evidence for constant $\bar{m}$ has been reviewed by Watt *et al.* (*62*). Heterogeneities in the lower mantle are therefore probably not due to variations in the iron concentration but rather to the relative changes in calcium-rich minerals and garnet structures, since $\bar{m}$ is insensitive to many calcium minerals and aluminum-rich minerals such as garnet mixed with accepted mantle minerals.

The Core

It is interesting that in the U.S. Program for the Geodynamics Project (*1*) there was little mention of the physics of the earth's core. It is clear from the progress made in understanding the thermal structure of the earth that the mantle cannot be understood apart from the core. The core must be producing a large thermal flux because of the suspected large thermal boundary at the core-mantle interface. A basic understanding of the chemistry of the core is needed if we are to understand this large flux. It was accepted a decade ago that the core is largely iron, modified by a lighter element; this concept has not changed. Much research has been devoted to understanding what this modifying element or elements can be. Recent candidates include sulfur, potassium, and oxygen (*63*).

Conclusions

In 1970 the main features of the earth's anatomy were known. Progress in the past decade has been devoted to elaborating upon the earth's anatomy but more importantly, has revealed glimpses of the planet's physiology. Enough has been learned to assure that dynamic problems of the planet will be a central focus in the coming decade.

The research results of the geodynamics of the mantle and core have implications for the current philosophical basis of geophysics, uniformitarianism. According to this philosophy, the present is the key to the past, and in principle the early moments of the earth's creation as a planet can be reconstructed if the present state of the earth is sufficiently well understood. If the thermal regime of the deep interior is controlled by a fixed viscosity, as maintained by Tozer (*3, 14*), then the physical properties of the planet's interior have only a "fading memory" of the initial conditions at the time of the formation of the planet. This conclusion is a setback for uniformitarianism because, in this case, the present however well understood is not reliable as a key to the distant past.

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Hotspots, Basalts, and the Evolution of the Mantle

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Hotspots, or plumes, have not yet been fitted satisfactorily into either the tectonic or the geochemical framework for the evolution of the mantle. A hotspot is a relatively stationary and long-lived thermal anomaly in the mantle and provides a variety of trace element enriched magmas to oceanic islands and continental interiors. From the point of view of mantle processes and the chemical evolution of the mantle, the alkali volcanism associated with hotspots is a very important process even if the abundance of such rocks is small as compared to the volume of abyssal tholeiites (*1*). The processes of creation and subduction of oceanic crust and lithosphere clearly account for most of the mass transport into and out of the upper mantle, and the study of mid-ocean ridge basalts (MORB) has placed important constraints on the part of the mantle that is providing these magmas. Trace element and isotopic considerations show, however, that the source region for MORB cannot be representative of the average composition of the mantle, nor can it have existed as a separate entity with its present characteristics for the full age of the earth.

The composition, location, and volume of this source region, which I designate as MORBS, are still uncertain. It appears to be very large, homogeneous, global in extent, and the result of an earlier differentiation or fractionation process that depleted it, relative to other source regions, of most of the incompatible elements, that is, those trace elements that are strongly concentrated into partial melts. On the other hand, it is clearly not lacking in a basaltic component. The continental crust is complementary to MORB in both trace elements and isotopic ratios (*2, 3*). Formation of the continents has presumably been at least partially responsible for the depletion of the so-called oceanic mantle. If the continental crust is the only enriched reservoir, then mass balance calculations indicate that only about 30 percent of the mantle has been processed, the rest remaining undifferentiated and primordial (*3, 4*).

A large primitive reservoir would be difficult to reconcile with other evidence regarding the early thermal evolution of planetary interiors. There is, however, evidence for an additional enriched reservoir in the mantle. Magmas from conti-

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