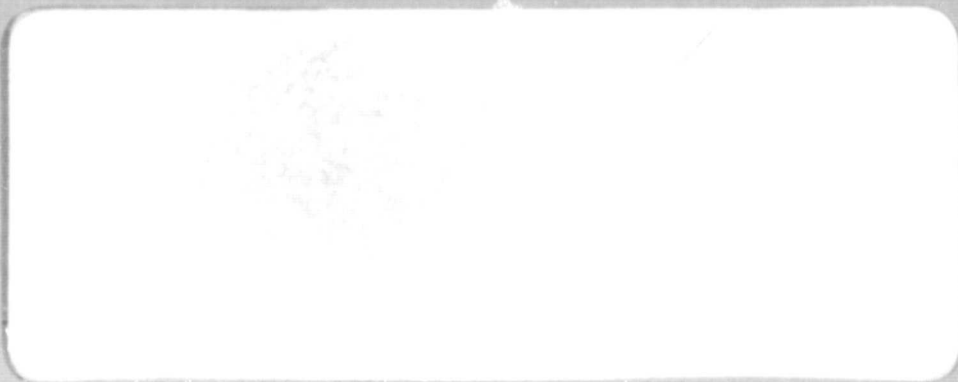# N O T I C E

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CALIFORNIA INSTITUTE OF TECHNOLOGY

PASADENA, CALIFORNIA

FINAL TECHNICAL REPORT

1 June 1967 - 30 June 1980

THEORETICAL INVESTIGATION OF LUNAR

AND PLANETARY INTERIORS

NASA RESEARCH GRANT NO.   NGL 05-002-069

FINAL TECHNICAL REPORT

Grant Number: NASA Research Grant No. NGL 05-002-069

Name of Contractor: California Institute of Technology
                                 Pasadena, California   91125

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NASA Technical Monitor: Joseph Boyce
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                                 Office of Space Science
                                 Code SL-4
                                 NASA
                                 Washington, D. C.   20545

Title: Theoretical Investigation of Lunar
                                    and Planetary Interiors

Period of Grant: 1 June 1967 - 30 June 1980

Period of Report: 1 June 1967 - 30 June 1980

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CONVECTION, COMPOSITION AND THE

THERMAL STATE OF THE LOWER MANTLE

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CONVECTION, COMPOSITION AND THE
THERMAL STATE OF THE LOWER MANTLE

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$\mathcal{J}4/77$

Abstract

We present a thermal model for the lower mantle which is constructed
from petrologically derived estimates of the temperature in the transition
zone and from an adiabat based on the thermal properties of MgO and SiO_2
measured at high pressures. Superadiabatic contributions to the geotherm
through the lower mantle, including those from possible phase transformations
but excluding those from thermal boundary layers, are negligibly small.
A thermal boundary layer is required at the base of the mantle in order
to satisfy our estimate of the lowest possible temperature in the core
(about 2800K); its thickness of about 100 km is well constrained. Even
so, our model suggests that we either require a significantly larger heat
flux from the core than has been considered reasonable (> 50 mWm^{-2}) or
we require the presence of a further thermal boundary layer within the
lower mantle in order to arrive at more likely core temperatures (about
3200-3500K). This latter alternative appears the more plausible, and
such a boundary layer requires a barrier to convection. Guided by the
present seismological evidence, we suggest that a thermal boundary layer
is associated with a chemical discontinuity either at the top of the
lower mantle or near its base (D" region). If seismological data require the
D" layer to be significantly thicker than about 100 km, this is support
for a chemically distinct D" region with multiple thermal boundary layers.
Otherwise, the results of this study suggest that the upper and lower
mantle may be chemically distinct, hence precluding whole-mantle convection.

Introduction

The conposition and dynamic state of the earth's lower mantle have
evoked considerable attention recently, in particular with reference to
arguments being presented for or aginst mantle-wide convection [Davies
1977; Jordan, 1977; O'Connell, 1977; Richter and McKenzie, 1978].
Similarly, it has been argued that seismological data, combined with
measured and extrapolated physical properties of candidate minerals, do
or do not require a difference in bulk chemical composition between
the upper and lower mantle [e.g. Birch, 1961; Press, 1968; Anderson, 1968,
1977; Anderson, et al. 1972; Mao, 1974; Davies, 1974; Ringwood, 1975;
Burdick and Anderson, 1975; Watt, et al. 1975; Liu, 1979]. These two
questions are, of course, coupled in that whole-mantle convection is not
compatible with a compositional change between the upper and lower mantle.

We propose a different approach for addressing this problem by
examining the thermal state of the lower mantle, and the ensuing impli-
cations for convection and chemical composition throughout the mantle.
We find that the temperature gradient through the lower mantle is well
enough constrained to indicate the existance of a thermal boundary layer at
the base of the mantle. Furthermore, our preferred solution requires a
second boundary layer elsewhere within the lower mantle, most likely
either close to its base or associated with the transition zone. This
strongly suggests the existence of a chemical transition which is a
barrier to convection either at the top of the lower mantle or in a complex
region at its base. A thermal structure with only one thermal boundary
layer within the lower mantle (presumably at the core-mantle boundary)
is only acceptable if the outer core is at a lower temperature than has
previously been estimated or if the heat flux from the core is significantly
larger than has been considered reasonable.

Adiabatic Gradients

We examine the following equation for the increase in temperature from the top of the lower mantle ($T(700)$) to the top of the outer core ($T(2900)$):

$$T(2900) - T(700) = [\Delta T_S + \bar{\beta}\Delta z] + \Sigma \; \Delta T(P.T.) + \Sigma \; \Delta T(B.L.). \qquad [1]$$

The bracketed term contains the nearly adiabatic temperature rise through the lower mantle, appropriate to a convecting and homogeneous system ($\bar{\beta}$ is the average superadiabatic gradient and Δz is the thickness of the lower mantle). The remaining terms allow for temperature increases associated with regions of phase transformations ($\Delta T(P.T.)$) and thermal boundary layers ($\Delta T(B.L.)$), respectively [cf; Verhoogen, 1965; McKenzie, et al. 1974]. We constrain first the left hand side of [1] and then evaluate each term on the right hand side in succession.

We presume throughout this discussion that the lower mantle is convecting, as is strongly indicated by the fact that it is very nearly adiabatic throughout [e.g. Jordan and Anderson, 1974; Dziewonski, et al. 1975; see also Stacey, 1977a]. Since an adiabatic gradient is not a steady solution to the conduction equation, it can exist only under very special circumstances and the most likely interpretation is that the planetary interior is adiabatic due to sufficiently vigorous convection.

The temperature at the top of the outer core is poorly constrained, however we estimate that a minimum plausible value is about 2800 K. We have estimated both liquidi and solidi for the Fe-S system under core conditions based on extrapolations of Usselman's [1975 a,b] high-pressure data for the eutectic and Liu's [1975] analysis of the melting curve of iron, corrected upward slightly to conform with the nonzero shear modulus observed in iron at 180 GPa under shock [Al'tshuler, et al. 1971]. Taking Ahrens' [1979]

bounds on the maximum sulfur content allowed in the core and assuming that the core is at a temperature at least as high as the mean between eutectic and liquidus, we arrive at a lowest estimate of about 2800K (under these conditions the outer core would already be a slurry consisting of at least 5-10% crystals according to our extrapolations). Our bound of 2800K is substantially less than most estimates of the temperature in the core [cf. Jacobs, 1975], and is less than or comparable to previously suggested lower bounds [e.g., Birch, 1972; Usselman, 1975b]. Furthermore, we have only considered sulfur as the light element in the outer core: sulfur is likely to decrease the melting temperature of iron significantly more than other plausible candidates [e.g., Brett, 1976], in particular oxygen which is currently considered a strong candidate [Ringwood, 1978]. In this case a minimum estimate for T(2900) would be above 3000-3200K, taking the zero pressure phase equilibria [Darken and Gurry, 1953] and assuming that the effect of pressure in the Fe-O system can be scaled to that in the Fe-S system [see also Ringwood, 1978]. Thus a more plausible value for T(2900) might lie between about 3200 and 3500K, in accord with recent analyses [e.g., Stacey, 1972, 1977b; Verhoogen, 1973; Usselman, 1975].

The temperature at the base of the transition zone can best be constrained petrologically, yielding estimates of about 1900-2000K for T(700) [Ahrens, 1973; Akimoto, et al. 1976]. For example, taking the transition from olivine to β-spinel as beginning at about 390 km depth [Burdick and Helmberger, 1978] provides a temperature "fixed-point" of about 1700K [e.g., Akimoto, et al. 1976]. This implies a temperature of about 2000K at the base of the transition zone, for plausible values of thermal expansion, specific heat and Clapeyron slope [cf. Verhoogen, 1965; Mao, et al. 1969; Schubert, et al. 1975; Navrotsky, et al. 1979]. Because of the sensitivity of the estimated temperatures on the detailed phase equilibria as well as

the uncertainties in our estimates of the thermo-
chemical properties in the transition zone, our values for T(700) are
somewhat uncertain, perhaps by as much as 100-200K. However a recent,
in situ determination of the olivine-spinel phase boundary [Akimoto, et al.
1977] suggests that our values of temperature may, if anything, be high.
Furthermore, independent evaluations of the temperature at the top of
the lower mantle are very compatible with our result [e.g., Banks, 1969;
Graham and Dobrzykowski, 1976; Watt and O'Connell, 1978]. Thus, we
estimate the difference in temperature between the top of the lower
mantle and the outer core to be no smaller than 800K, and more likely of
the order of 1300K.

If we assume, initially, that the lower mantle is homogeneous (in
phase and composition) the bracketed term in [1] should yield the rise
in temperature from top to bottom of the lower mantle. Following Jeffreys
[1930], dynamic stability requires a Rayleigh number less than about 10^3,
and this provides an estimate of the superadiabatic temperature increase through
the lower mantle of $\overline{\beta}\Delta z \sim$ 1K. Here, the constraints currently available
for the viscosity [Peltier and Andrews, 1976; see Table 1 for representative
parameter estimates] have been used and this estimate is only accurate to
about one order of magnitude. In any case, the superadiabatic gradient
is negligibly small.

The adiabatic temperature gradient is given by the Grüneisen parameter,
$\gamma = \left(\frac{\partial \ln T}{\partial \ln \rho}\right)_S$, as a function of density (ρ). We have determined γ for the
lower mantle from experimental data on MgO and SiO_2 throughout the pressure
range of interest. Shock-wave data on porous (or amorphous) and nonporous
samples yield values of γ directly as a function of pressure or density

[see Carter, et al. 1971, for example]; for MgO and SiO$_2$ these values

are quite similar as a function of pressure [data base: Wackerle, 1962;

Al'tshuler, et al. 1965; Hart and Skidmore, 1965; Anderson, 1967;

Jones, et al. 1968; Ahrens and Rosenberg, 1968; Carter, et al. 1971;

Trunin, et al. 1970, 1971; Podurets, et al. 1976). Figure 1 shows the

values of γ derived from these data for mixtures of the oxides corresponding

to olivine and pyroxene stoichiometry. The differences between these two

cases are not resolvable in this study. Taking MgO and SiO$_2$ at high

pressures to be representative of the earth's deep interior, this supplies

the ratio of temperatures at the top and bottom of the lower mantle along

an adiabat by applying seismologically determined density profiles

[e.g., Dziewonski, et al. 1975]. Consequently ΔT_s, the adiabatic

temperature increase across the lower mantle, corresponds to about 460K

(with an average $\overline{\gamma} = \frac{\Delta \ln T}{\Delta \ln \rho} \sim 0.88$), and the bracketed term in [1] amounts

to no more than about 500K. This is approximately 300K less than even

our limiting estimate of the left hand side of [1].

That our estimate of the Grüneisen parameter is similar to those

resulting from independent analyses is shown in Figure 1. The different

estimates do not strongly affect ΔT_S in that increasing $\overline{\gamma}$ from 0.88 to 1.0

raises the ratio of temperatures at the top and bottom of the lower mantle

by only 3%. However, our analysis has the advantage of being constrained

by experimental data measured at the appropriate pressures.

Superadiabatic Gradients

In the discussion above we have estimated the temperature at a depth

of 700 Km, a lower bound on the temperature at the core-mantle boundary

and the Grüneisen parameter, which specifies the adiabatic gradient between

these two points. Our results are illustrated in Figure 2 and it is apparent

that a single adiabat does not join the two temperature estimates. Therefore one or more regions of superadiabatic gradients are likely to exist between 700 km depth and the core-mantle boundary. The required superadiabatic temperature increase is so large as to further validate our initial assumption that convection is inevitable. We now explore how such superadiabatic gradients can. arise in a convecting mantle.

If regions of phase transformation exist within the convecting lower mantle, they could contribute superadiabatic temperature increases, depending on the enthalpies of reaction involved. The occurence of phase transformations has indeed been inferred from several seismological studies [e.g. Kanamori, 1967; Johnson, 1969; Vinnik and Nikolayev, 1970; Wright and Cleary, 1972; Butler and Anderson, 1978] and, although the evidence is still open to question [e.g., Wiggins, et al. 1973], as many as five seismic discontinuities have been identified [Johnson, 1969]. Whether these transitions are univariant or polyvariant, the additional change in temperature associated with each transformation is very nearly [Verhoogen, 1965]

$$\Delta T(P.T.) \sim \frac{T}{C} \left(\frac{dP}{dT}\right) \Delta V,$$

where T is the temperature, C an appropriate (averaged) specific heat, $\frac{dP}{dT}$ the Clapeyron slope and ΔV the volume change associated with the transformation. The inferred lower mantle discontinuities are known to be small (velocity jumps less than about 1%) and we take $\Delta V \sim 2\%$ as a generous upper limit. For typical values of the Clapeyron slope (2-3 MPa K^{-1}) we find $\Delta T(P.T.)$ no larger than 20-30K at each transition, and hence the total contribution from phase transformations to the temperature change

throughout the lower mantle amounts to no more than about 100-150K, as shown in Figure 2. Of course, if smaller (or negative) Clapeyron slopes are considered for any of the inferred transitions (as might be appropriate for transformations involving disproportionation) the total $\Delta T(P.T.)$ is smaller than our calculated values (or even negative). Clearly, the effect of phase transformations is insufficient to reconcile our two temperature estimates and we are therefore forced to consider further sources for superadiabatic temperature increases within the lower mantle.

An obvious place for non-adiabatic gradients would be immediately above the core-mantle boundary if a significant fraction of the earth's total heat flow is supplied to the mantle from the core. Heat flowing out of the core will result in a thermal boundary layer at the base of the mantle and we need an appropriate convection model to estimate the characteristic thickness and temperature change across this boundary layer. One complication is that the required model will have to take into account the effects of compressibility and spherical geometry when applied to such a deep layer but to date these two properties have only been studied separately, and in the case of spherical geometry over a very limited parameter range. Nevertheless the existing results are sufficient to evaluate the way in which the boundary layers from incompressible, plane layer convection models will be modified.

We will first consider the boundary layers that are found in plane layer calculations which assume negligible compressibility, uniform viscosity and a surface heat flux of 5.85 X 10^{-2} Wm^{-2}, which is close to the average measured terrestrial heat flow. The material properties of the convecting material are given in Table 1. The resulting horizontally averaged temperature for three models of heating are given in Figure 3.

When all of the heat is supplied from below, the averaged temperature structure consists of two equal boundary layers joined by an isothermal interior region. The characteristic boundary layer thickness δ is about 100 km and the associated temperature change $\Delta T(B.L.)$ is approximately 300K.

A simple boundary layer analysis [see McKenzie, et al. 1974] can be used to determine the dependence of the boundary layer properties on the heat flux and material properties of the layer:

$$\delta \propto \left(\frac{\rho C_p \kappa^2 \nu}{g\alpha} \right)^{1/4} F^{-1/4}$$ [2]

and

$$\Delta T(B.L.) \propto \left(\frac{\nu}{g\alpha\kappa^2 \rho^3 C_p^3} \right)^{1/4} F^{3/4} \; ,$$ [3]

where the symbols and their typical value for mantle materials are given in Table 1. Equations [2] and [3] are consistent with the idea that the thermal boundary transmits the flux F conductively:

$$\rho \; C_p \; \kappa\frac{dT}{dz} \sim \rho \; C_p \; \kappa\frac{\Delta T}{\delta} = F$$

An important point regarding the boundary layer properties is that uncertainties in the flux across a given boundary layer affect the estimated temperature change much more than the boundary layer thickness. This effect can be seen in Figure 3, which includes a case in which half the heating is from below and half from within (by internal heat sources). The thickness of the lower boundary layer is not much different than in the case fully heated from below, but the temperature change is substantially

reduced. The fully internally heated case (i.e., no flux from the core) is also shown in Figure 3 to illustrate the absence of any lower boundary layer under such conditions.

The next question is how does spherical geometry affect the results given above? A useful calculation relevant to this question is by Young [1974] who considered finite amplitude convection between isothermal shells. Figure 4 compares the horizontally averaged thermal structure of convection in a spherical shell to that of convection between plane isothermal boundaries. The spherical case has a relatively low Rayleigh number of about five times critical and a Prandtl number of five, which means that inertial terms in the momentum equation are not negligibly small. The plane layer case has a somewhat larger Rayleigh number of about 15 times critical and the Prandtl number is assumed to be infinitely large. The two calculations thus differ in more than geometry alone but the major effect on the thermal boundary layer structure does appear to be simply related to the geometric differences. The plane layer case has symmetric boundary layers. In the spherical case both the inner and the outer boundaries transmit the same total amount of heat, but since they differ in surface area by a factor of about three, the heat flux per unit area is three times greater across the lower boundary. As one would expect from relations [2] and [3] this difference in heat flux has little effect on the boundary layer thickness but results in an almost three times greater temperature drop across the lower boundary layer. As a first approximation, the effect of spherical geometry can be taken into account by simply using the local heat flux for estimating boundary layer properties.

The effect of compressibility is illustrated in Figure 5. The incompressible temperature profile shown for reference is the same as in Figure 3 but now using a specific depth extent of 700 Km. The thermal

structure is taken from a recent study of convection in a compressible material by Jarvis and McKenzie [1979]. The importance of compressibility in such a calculation is measured by the ratio of the layer depth d to the thermal scale height H_T ($H_T = C_p/g\alpha$). This scale height is the vertical distance a parcel of material must be displaced for its temperature to change by a factor e due to adiabatic compression. The case used to illustrate the effect of compressibility on the average thermal structure assumes $d/H_T = 0.5$, which results in a reasonable large effect. Comparing the two calculations one can see that the interior temperature is no longer constant in the compressible case, and in fact, the interior temperature gradient is found to be almost exactly adiabatic as we have taken it to be in our earlier discussion. The boundary layers are not very much affected by the compressibility because they are small compared to H_T. When considering the mantle, we argue that compressibility will not greatly affect the thermal boundary layers because $\delta/H_T \ll 1 (\delta \sim 100$ Km, $H_T \sim 5000$ Km) but the interior temperature gradient joining the boundary layers must be considered to be adiabatic.

Guided by the discussion on the effects of sphericity, we will assume that 1/4 of the total terrestrial heat flow arises originally in the core. The heat flow per unit area out of the core is then almost exactly equal to the surface heat flow and we expect, based on the arguments given above, that the boundary layer at the base of the mantle will be about 100 km thick and with a temperature change of approximately 300K. A smaller flux from the core would result in a smaller boundary layer, both in thickness and in superadiabatic temperature gain (equations [2] and [3]). Similarly, because we have ignored the temperature-dependence of viscosity within the thermal boundary layer, our values of δ and ΔT may be somewhat large. However, taking our best estimate of the

temperature at the top of the mantle, combined with our adiabatic gradient
(± contributions from phase transformations), we can satisfy estimates
of the minimum core temperature once we include a thermal boundary layer at
the bottom of the mantle.

Independently, there is considerable evidence from seismology [see
Cleary, 1974, for example] that the lowermost mantle [the D'' region,
extending to about 100-200 km above the core-mantle boundary: Bullen,
1949] is distinct from the overlying mantle and does not behave as an
adiabatic and homogeneous region. The seismological anomalies corresponding
to this region may therefore reflect the properties of a thermal boundary
layer. We note that our estimated average temperature gradient through this
layer ($\sim$ 3K km^{-1}) is probably smaller by a factor of 2 to 4 than the critical
gradient required to make the gradients in density or seismic parameter
vanish; however dispersion may significantly affect the seismologically
observed velocities. Allowing the values quoted in Table 1 to vary within
reasonable limits does not significantly change our estimate of the thermal
boundary layer thickness, therefore if the seismological data require an
anomalous layer thicker than about 100 km at the base of the mantle, this
should be taken as evidence that something more than a single thermal
boundary layer is present.

Chemical Composition

We have presented our best estimates of the temperature field through
the lower mantle and have found that a solution can be derived which is
in agreement with our minimum bounds on the core temperature. We are
required, however, to insert a significant thermal boundary layer at the
core-mantle interface (Figure 6a). This conclusion is not affected by
the presence (or lack thereof) of phase transformations in the lower
mantle because we found that such transitions have a minor if not negligible

effect on the overall geothermal gradient. We note that the thermal boundary layer near the earth's surface involves a much larger ΔT than that at the base of the mantle due largely to the temperature dependence of the effective viscosity.

Although this model is acceptable, it barely satisfies our bounds and we find it a rather extreme solution. If the lower mantle contains only one boundary layer, we estimate a temperature at the top of the core which is very low: about 2800 - 3000K. Such temperatures can not be considered unreasonable, however they are significantly lower than recent estimates [e.g., Verhoogen, 1973; Stacey, 1977b; Bukowinski, 1977] and especially previous results ranging between 4000 and 5000K [cf. Jacobs, 1975].

One possible factor might come from our having underestimated the heat flux from the core. Since $\Delta T(B.L.)$ is proportional to $F^{3/4}$ (equation [3]), a larger flux might allow a large enough temperature jump across the boundary layer to yield more plausible core temperatures. If the total flux from the core is comparable to the total heat flux through the earth's surface then our mantle temperatures are essentially compatible with our best estimates for the core. Such a large flow of heat from the core has, however, been generally considered unreasonable [e.g., Stacey, 1972; Verhoogen, 1973]: our value of 1/4 of the total output at the surface represents the maximum for steady state. This is within a factor of two of the estimated minimum [Stacey, 1972, 1977a] and is probably a fairly good estimate.

Thus, we find that if the lower mantle contains only one boundary layer at its base, the core must be at a significantly lower temperature than has previously been considered (essentially consistent with Tolland's [1974] conclusions). Alternatively, if a temperature of about 2900-3000K is considered to be too low for the outer core, the lower mantle must contain

other boundary layers in addition to that at the core-mantle boundary. We conclude, therefore, that a second boundary layer within the lower mantle is suggested by our results up to this point.

From our analysis we can not constrain at which depth such a boundary layer might be found, although we can rule out the possibility that there could be many such boundary layers within the lower mantle since the resultant temperatures in the core would most likely be well above the melting point of iron at the inner core - outer core boundary.

We consider two depths, as shown in Figures 6b and c, as the most plausible locations for such a boundary layer based on seismological evidence: either at the transition zone or near the base of the mantle. The lowermost mantle, for example, is known to be anomalous and it has been inferred to contain significant inhomogeneities [e.g., Phinney and Alexander, 1969; Davies and Sheppard, 1972; Kanasewich, et al. 1973; Haddon and Cleary, 1974; Doornbos, 1976; van den berg, et al. 1978]. Thus, if layer D" is chemically distinct from the overlying mantle it could contain a boundary layer both at its top and bottom, and the temperature could increase by about 900K within the lowermost 200-300 km of the mantle, easily satisfying currently accepted temperatures for the core (Figure 6b). Such a model has, in principal, already been proposed [e.g., Sacks and Beach, 1974; Snoke and Sacks, 1976] and it has been shown to be compatible with some of the seismological data. However, numerous difficulties remain [e.g., Husebye, et al. 1976; Miller, et al. 1977] and this model must be considered speculative at present. An attempt to explain the observations from the lowermost mantle, such as the velocity anomalies and apparent scattering, in terms of a simple (single) boundary layer could provide an important seismological test on the constitution of the lower mantle.

The remaining possibility is to place a thermal boundary layer at the top of the lower mantle (associated with the transition zone) as well as at its base (Figure 6c). Again, this possibility can not be critically assessed for lack of data, however it has important implications in that it precludes whole mantle convection. Such a boundary layer would most likely be associated with a change in chemical composition from upper to lower mantle [cf. Richter and Johnson, 1974] and is therefore consistent with suggestions to this effect. Indeed, if the independent arguments in support of a change in composition across the transition zone are correct [e.g., Anderson, 1968, 1977; Press, 1968; Anderson and Jordan, 1970; Anderson, et al. 1972; Burdick and Anderson, 1975; Liu, 1979], then a boundary layer would be expected at the top of the transition zone and temperatures through the lower mantle would be some 500-700 K higher than we have shown in Figure 2. With regards to experimental, high pressure phase equilibria, the situation at present is that no adequate explanation has apparently been found for the large velocity and density jump at the base of the transition zone [e.g., Hart, et al. 1977; Burdick and Helmberger, 1978] in terms of polymorphic phase transformations alone [Anderson, 1976; Liu, 1979], and therefore a change in bulk chemical composition is strongly implicated for the transition zone. Similarly, recent isotopic and trace element data can be interpreted to suggest the existence of a separate, relatively undepleted geochemical reservoir underlying a more depleted upper mantle reservoir [DePaolo and Wasserburg, 1976; O'Nions, et al. 1978, 1979].

Conclusions

The central point of this study is that there is good evidence for regions of non-adiabatic gradients within the lower mantle and these are

most likely confined to thermal boundary layers. We have evaluated
the adiabatic thermal gradient through the lower mantle by using high
pressure experimental data on the Grüneisen parameter and have found that
this gradient is compatible with a lower bound on the temperature in the
core only after allowing for a thermal boundary layer at the core-mantle
boundary. Even with a thermal boundary layer at the core-mantle boundary
we require a very low core temperature, and therefore we discussed the
possibility of further non-adiabatic regions within the lower mantle.
Our analysis relies upon having properly estimated the temperature at
the top of the/mantle and other constraints on the nature of the transition
zone from high-pressure experimental petrology would be very useful in this
regard. However our geotherm can be arrived at by several lines of reasoning
and it is therefore considered to be adequately constrained.

We can not resolve the effects of possible phase transformations
within the lower mantle since they appear to have only minor consequences
for its thermal state, within the scope of our study. The thermal boundary
layer properties depend on the thermal flux assumed from the core, and our
conclusions regarding the need for more than one thermal boundary layer within
the lower mantle, may not be valid if a significantly larger flux is allowed
from the core than has previously been considered acceptable. Alternatively,
we find that either the temperature in the core must be lower than has been
considered plausible, or there must be a barrier to convection within the
lower mantle which results in additional thermal boundary layers. We find
this latter possibility more plausible, however we can not constrain the
depth at which such a barrier may be found. Guided by seismological evidence,
we suggest the existence of a transition in chemical composition either at
the top of the lower mantle or toward its base (D" region) as being the

most likely cause of extra thermal boundary layers. This, then, provides us with a large enough increase in temperature across the lower mantle to allow plausible core temperatures (about 3500K). If a complex zone with multiple thermal boundary layers and separate convection within it is required to explain the observed (anamalous) properties of the D" region then we estimate an increase in temperature of about 1000K within the lowermost 200-300 km of the mantle. Alternatively, the transition zone is a likely candidate for a region across which the bulk chemical composition changes; this could readily explain a thermal boundary layer located there.

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Table 1

Lower Mantle Properties

Density:	$\rho \sim 4$ to 5.5 Mg m^{-3}	(1)
Acceleration of Gravity:	$g \sim 10$ ms^{-2}	(1)
Specific Heat at Constant Pressure:	$C_p \sim 1.2 \times 10^3$ J Kg^{-1}K^{-1}	(2)
Kinematic Viscosity:	$\nu \sim 1$ to 10×10^{17} m^2 s^{-1}	(3)
Thermal Diffusivity:	$\kappa \sim 1$ to 2×10^{-6} m^2 s^{-1}	(4)
Coefficient of Thermal Expansion:	$\alpha \sim 1$ to 2×10^{-5} K^{-1}	(5)
Heat Flux From Core:	$F = 5.85 \times 10^{-2}$ W m^{-2}	(6)

Sources

(1) Measured

(2) Dulong-Petit value

(3) Peltier and Andrews [1976]

(4) Schatz and Simmons [1972]; Kieffer [1976]

(5) This study: $\alpha = \dfrac{\gamma C_p \rho}{K_s}$; also, Stacey [1977b]

(6) Assumed equivalent to the flux through the earth's surface:

see text.

Figure Captions

Figure 1   Grüneisen parameter through the lower mantle based on experimental high pressure data for MgO and SiO_2 (this study) compared with the theoretical formulations of Stacey [1977b] and Brennan and Stacey [1979].

O and P refer to mixtures of the oxides corresponding to olivine and pyroxene stoichiometry, respectively.

Figure 2   The temperature profile for a homogeneous and adiabatic lower mantle (solid line) is compared with profiles including the effects of possible phase transformations (dashed line) and a thermal boundary layer at the core-mantle boundary (dotted). Fixed reference points in the transition zone are shown, along with our best estimate and lowest plausible values of temperature in the core.

Figure 3   Horizontally averaged temperature as a function of depth from numerical simulations of convection in an incompressible fluid layer. [McKenzie, Roberts and Weiss, 1974]. The flux at the upper surface is 5.85×10^{-2} Wm^{-2} and relevant material properties are given in Table 1.

   a) All heating from below

   b) All heating internal

   c) Half heating from below, half internally heated

The boundary layer thickness δ is measured from the boundary to the first point where $dT/dz = 0$.

Figure 4   Horizontally averaged temperature from numerical calculations

of convection between isothermal boundaries.

    a) Plane layer: $Ra = 10^4$, $Pr = \infty$

    b) Spherical shell: $Ra = 4 \times 10^3$, $Pr = 5$ and ratio of

       inner radius to outer radius of 0.6 [axially symmetric

       mode, taken from Young, 1974]

$Ra = \dfrac{g\alpha\Delta T\, d^3}{\kappa\nu}$, $Pr = \nu/\kappa$ and $\Delta T = T_2 - T_1$

The vertical axis is depth in fractions of d, the total

layer depth.

Figure 5   Horizontally averaged temperature for both incompressible

and compressible convection.

    a) Same as in Figure 3a but with depth of 700 Km

    b) Compressible case [Jarvis and McKenzie, 1978;

       case B 10]

Assumes depth = 700 Km, $d/H_T = 0.5$ and flux of 5×10^{-2} Wm^{-2}

from below.

Figure 6   Temperature profiles through the earth's mantle: three

acceptable models.   Thermal boundary layers (B.L.) are indicated

in the upper and lower mantle, while dynamic barriers associated

with possible chemical transitions in the lower mantle are shown

as dashed lines.   The preferred core temperature and its plausible

lower bound are also shown.   Note the fundamental asymmetry of

the profiles caused by the large decrease in temperature through

the lithosphere: an effect which has no counterpart at depth

within the earth and which is due to temperature-dependent

rheology.

Figure 6 (continued)

(a) Adiabat through lower mantle with one thermal boundary layer at the core-mantle boundary and with the core at a relatively low temperature. Taken from Figure 2.

(b) Adiabat with two thermal boundary layers near the base of the mantle corresponding to a seismologically complex and chemically distinct D'' region.

(c) Adiabat with a thermal boundary layer at the top and the bottom of the lower mantle. In this case, the lower and upper mantle are separate chemical and dynamic systems.
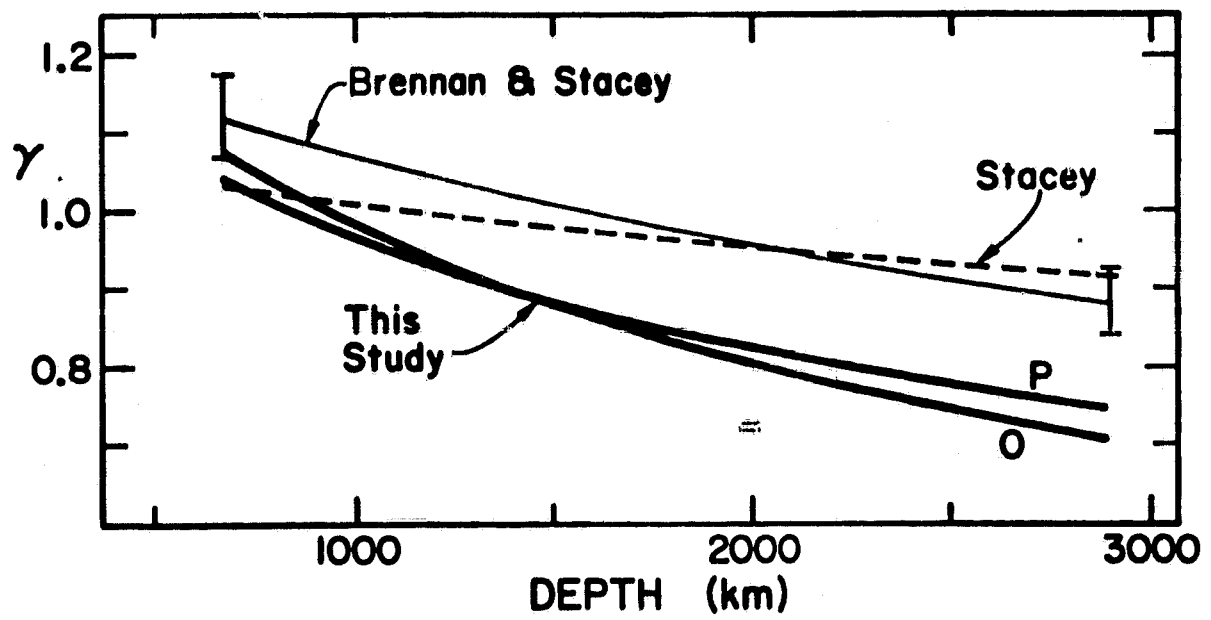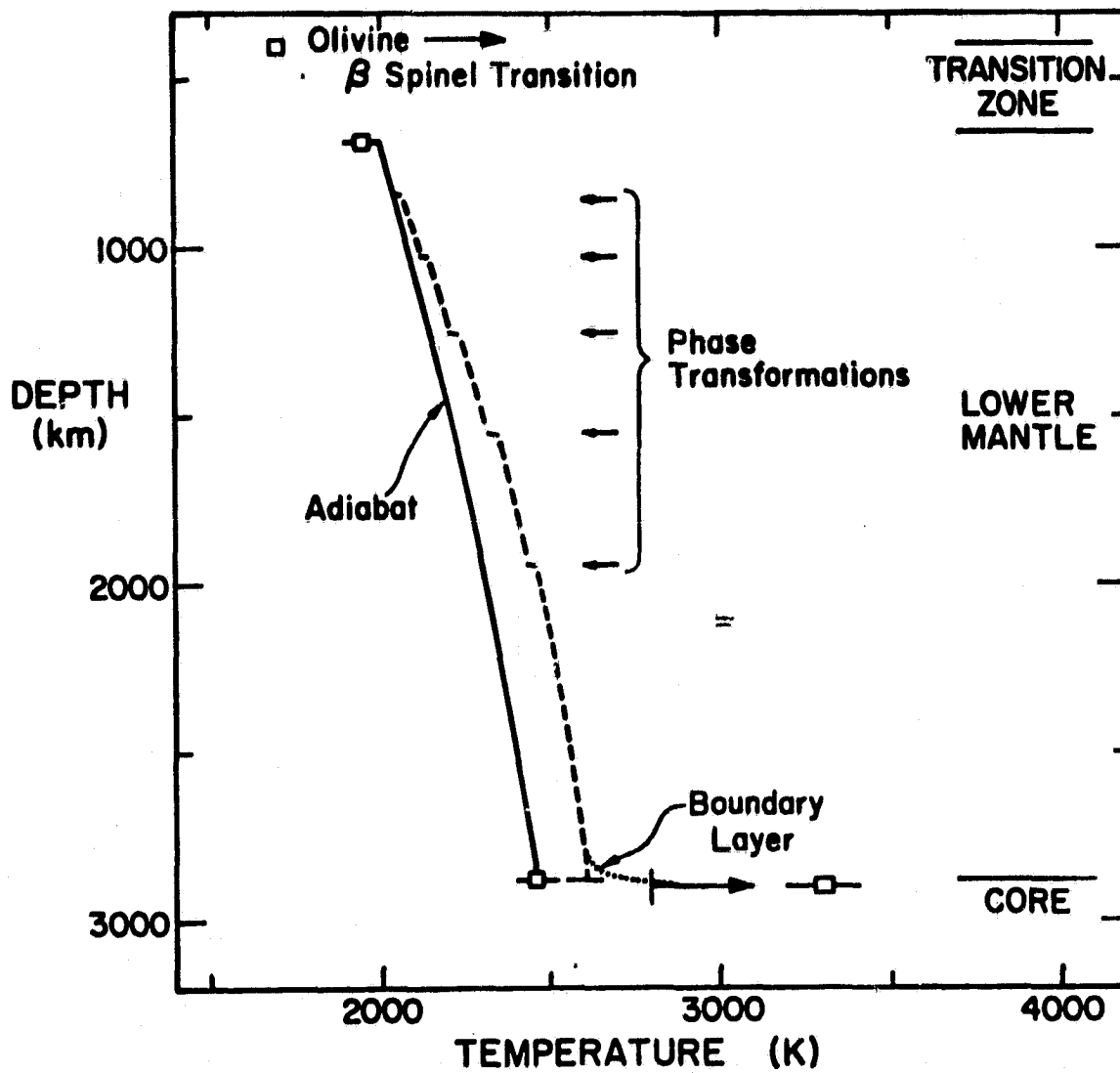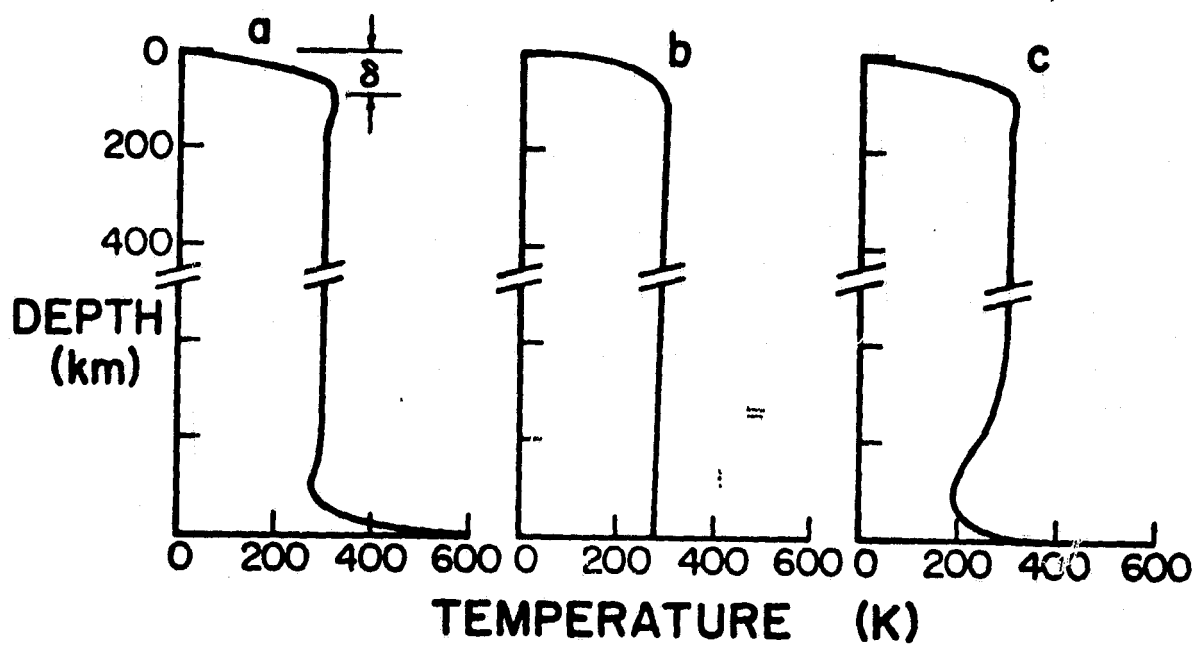
Figure 1

Figure 2

Figure 3

DEPTH

TEMPERATURE

Figure 4.

Figure 5

Figure showing temperature versus depth profiles for three mantle models (a, b, c). The horizontal axis is TEMPERATURE (K), ranging from 0 to beyond 3000. The vertical axis is DEPTH (km), ranging from 0 to 3000.

Labels include: LITHOSPHERE, TRANSITION ZONE, UPPER MANTLE, LOWER MANTLE, CORE, (D"), B.L., Olivine → β Spinel, Adiabat, Chemical & Dynamic Boundaries.
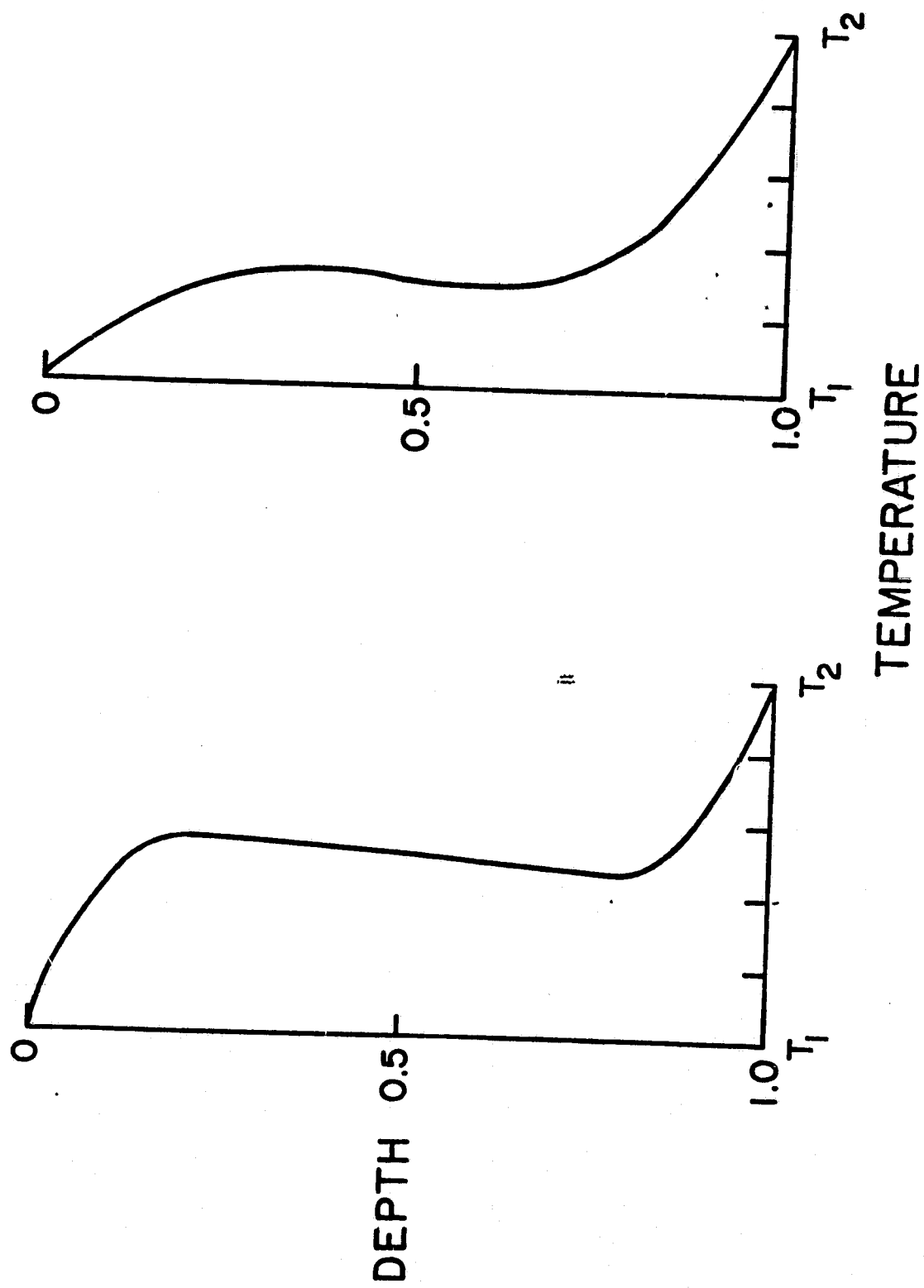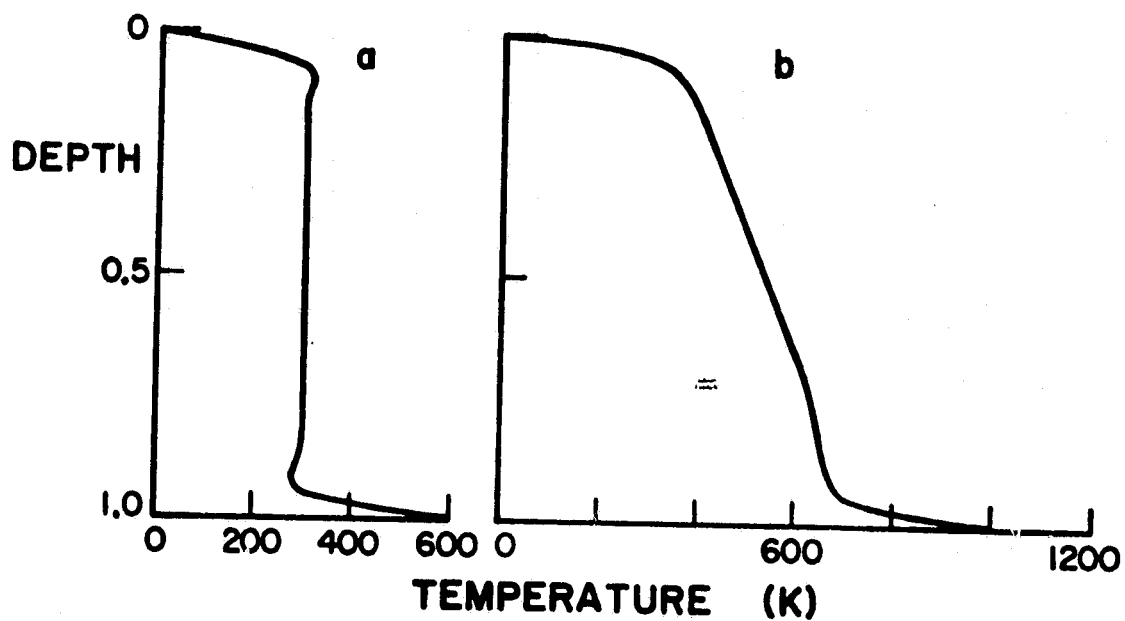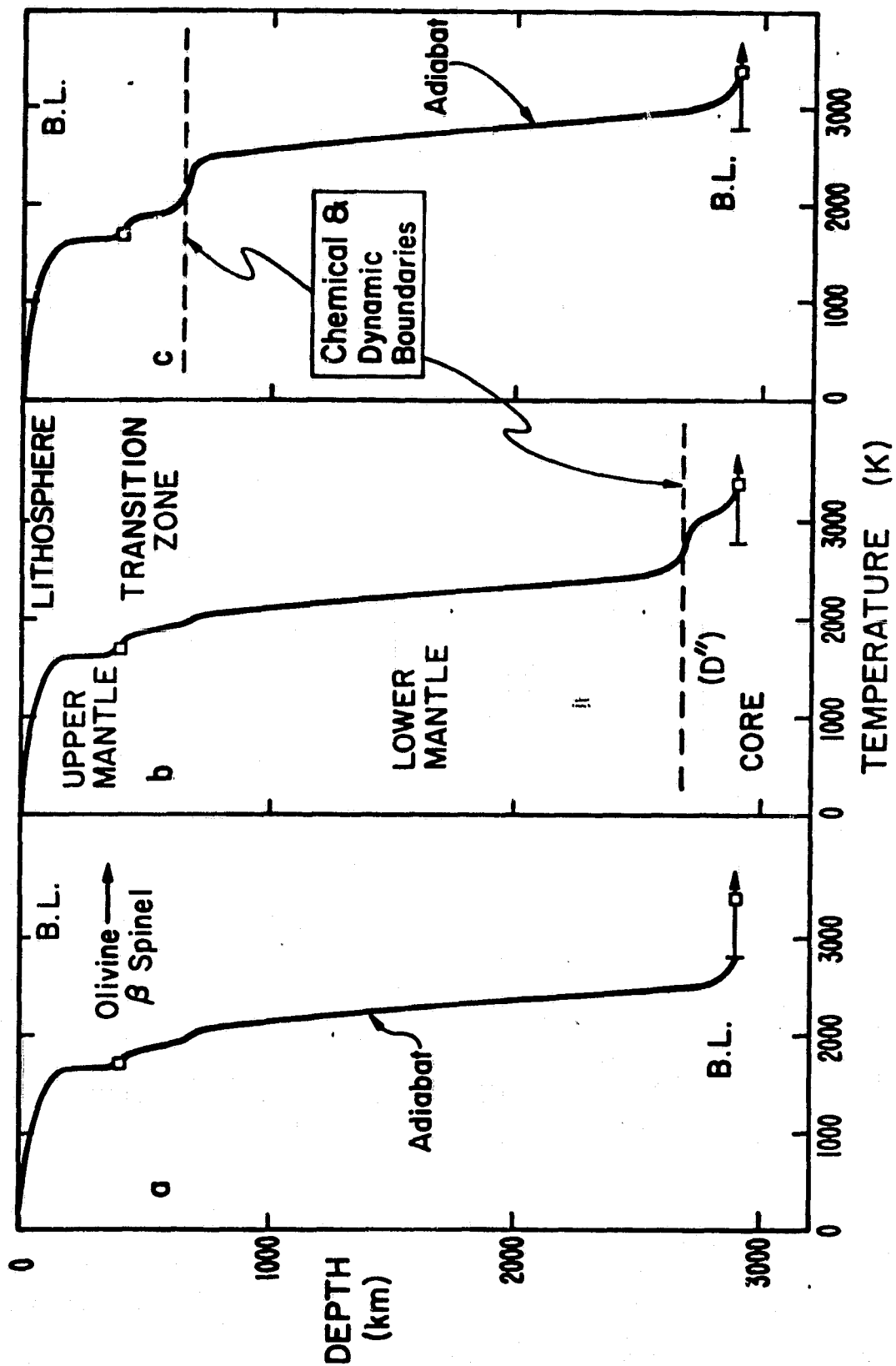
PLATE TECTONICS ON VENUS

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PLATE TECTONICS ON VENUS

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Abstract

The high surface temperature of Venus implies a permanently buoyant lithosphere and a thick basaltic crust. Terrestrial style tectonics with deep subduction and crustal recycling is not possible. Overthickened basaltic crust partially melts instead of converting to eclogite. Because mantle magmas do not have convenient acesss to the surface the ^{40}Ar abundance in the atmosphere should be low. Venus provides an analog to Archean tectonics on the Earth.

Introduction

The surface temperature of Venus is about 450 K warmer than the surface of the Earth. This affects the buoyancy, thermal expansion, thermal conductivity and, hence, the thermal evolution and ultimate fate of the lithosphere. The pressure in the upper mantle of Venus is at least 12% less than at equivalent depths in the Earth's upper mantle. Thus, the depth of melting, the locations of upper mantle phase changes and the viscosity of the upper mantle will be different for the two planets. The buoyancy and thermal properties of the lithosphere control the style of plate tectonics and the associated time and length scales.

Most discussions of the comparative tectonics between Earth and Venus address only the differences in viscosity.

The oceanic lithosphere of the Earth cools as it ages and it eventually becomes denser than the underlying mantle. This instability develops after about 40 m.y, (Oxburgh and Paramentier, 1977). Once cooling reaches a depth of 50 km the garnet pyroxenite or eclogite transformations may make a substantial contribution to the negative buoyancy of the lithosphere. Continents and oceanic plateaus resist subduction because of their thick low-density crusts. The fate of the terrestrial lithosphere, therefore, depends on both chemistry and temperature. The purpose of this paper is to investigate the implications for Venus tectonics of its high surface temperature. We will conclude that the surface thermal boundary layer on Venus is permanently buoyant and that the reversible part of mantle convection occurs below about 100 km.

The terms lithosphere and thermal boundary layer will be used for the cold outer layer of a planet. Neither term is strictly correct since the former has a strength connotation and the density of the latter is usually assummed to depend on temperature alone. The thickness of the mechanical or rheological lithosphere depends on temperature and, possibly, composition and is proportional to the thickness of the thermal lithosphere that is discussed here.

Lithospheric cooling

The average temperature of the oceanic lithosphere decreases about 660°C as it ages. The thickness of the conductively cooled thermal

boundary layer increases at a rate controlled by the thermal conductivity and the difference between the surface and interior temperatures. Since the crustal and harzburgite portions of the boundary layer are both less dense than the underlying mantle a gravitational instability can only occur after the conductive cooling penetrates for a sufficient distance, ~30 km, into the denser upper mantle portion of the boundary layer. In some petrological models the lower lithosphere is pyroxenitic. Upon cooling, plagioclase and spinel pyroxenite convert to garnet pyroxenite or eclogite with a substantial increase in density. These phase changes also require cooling to depths in excess of 30 kilometers.

The equilibrium thickness,δ, of a conductive boundary layer is

$$\delta = K\Delta T/\dot{Q} \qquad (1)$$

where K is the thermal conductivity, $\Delta T/\delta$ is the average thermal gradient and $\dot{Q}$ is the mantle heat flow. For K of 7×10^{-3} cal/cm sec°C, ΔT of 1300°C and Q of 0.6 μcal/cm^{2}sec, values appropriate for the terrestrial oceanic lithosphere, δ is about 150 km. For a lithosphere composed of 6 km of basalt, 24 km of harzburgite and a lower lithosphere composed of undepleted mantle, the equilibrium boundary layer for $\Delta T =$ 1300°C is 1% denser than underlying mantle and therefore gravitationally unstable. Phase changes in the lower lithosphere may contribute further to the negative buoyancy.

For a given mantle temperature and heat flow the equilibrium thickness of the boundary layer on Venus is reduced by two effects; K

is about 30% lower at the higher temperature (Schatz and Simmons, 1972) and ΔT is reduced by 460°C. This reduces δ on Venus to 42 km and its density, assumming similarity with oceanic lithosphere, is 2% _lighter_ than the underlying mantle. This is a result of the smaller amount of cooling, the higher proportion of basalt and harzburgite in the boundary layer and the lack of significant cooling where it is most required, i.e. the lower lithosphere. Furthermore, phase transformations in the lower lithosphere cannot contribute at high temperature and low pressure. Basalt has about one-half the thermal conductivity of ultramafic rocks. This reduces the thickness of the boundary layer to about 27 km if it is mainly basaltic.

The buoyancy of the Venus lithosphere, even for a relatively thin crust, is greater than the buoyancy of young oceanic lithosphere on Earth. Since a thicker crust for Venus is probable, it is certain that the surface thermal boundary layer for Venus is permanently buoyant and stable against subduction.

The temperature gradient in the conductive layer can be written

$$dT/dz = (1/K)(\dot{Q}_s - Az) \qquad (2)$$

where $\dot{Q}_s$ is the surface heat flow and A is the heat production rate in the layer. With basaltic conductivities and radioactivities ($A \sim 10^{-14}$ cal/cm^3 sec) the temperature and thermal gradient at 20 km are 950°C and 23°C/km respectively and the solidus of dry basalt will be exceeded at depths shallower than 40 km. Eclogite is not stable in the mantle of Venus at depths shallower than about 100 km (Anderson, 1980).

Since crust cannot subduct and the garnet-rich assemblages are well below the boundary layer, all partial melt products of mantle differentation that make their way into the outer 100 km or so will remain there. The crust is therefore likely to be thicker than either the oceanic crust or the average crustal thickness on Earth. It may be thicker than either the mechanical lithosphere or the thermal boundary layer. Crust which is overthickened by compression, buckling, thrusting, or collision will melt at its base. The resulting plutonism will further increase the density contrast between crust and mantle and increase the radioactivity and incompatible trace element content of the parts of the crust so affected.

The characteristic time for thickening of the boundary layer is

$$\tau = (\rho C_p / 4 \ K) \delta^2 \tag{3}$$

which for $\rho = 3.3$ g/cm^3, $C_p = 0.25$ cal/g°C and $\delta = 150$ km is ~200 m.y. for the Earth. For Venus the characteristic time is 20 m.y. or less.

Therefore, Venus has a thick, low-density and permanently buoyant crust-lithosphere that very quickly reaches thermal equilibrium. Upper mantle temperatures are high, resulting in low viscosities.

The basalt-eclogite transformation will occur below about 100 km. The driving mechanism for mantle convection may therefore be partly chemical. Basaltic melt rising in an upcurrent will freeze as eclogite if it remains below about 100 km, and provide the negative buoyancy required for overturn. The 670 km discontinuity which halts eclogite subduction on Earth (Anderson, 1979a,c) will occur at about 800 km in

Venus giving a convecting layer about 700 km thick. This should generate topographic and gravity anomalies of about 1400 km wavelength.

Equation (1) suggests that plate tectonics in the Archean on Earth may have been similar to present day tectonics on Venus. The small δ in early Earth history was primarily due to the high $\dot{Q}$ from radioactive decay, which was at least three times larger 4×10^9 years ago. The thermal boundary layer thickens as $\dot{Q}$ decreases until eventually it becomes unstable and overturns, destroying the early geological record and setting the stage for the present style of tectonics (Anderson, 1979a,b, 1980). This overturn event would allow much of the accumulated ^{40}Ar in the upper mantle to escape and the subsequent steady-state ridge/trench style of tectonics allows continuous outgassing of the terrestrial mantle.

Early Evolution of Earth and Venus

The initial evolution of Earth and Venus were likely quite similar. The high accretional energies would result in partial melting of the mantle and upward transport of a picritic melt. Crystallization of this upper mantle melt layer, or magma ocean, yields a thin plagioclase-rich crust and a deep eclogite cumulate layer. An olivine-orthopyroxene cumulate layer may form at intermediate depths. This kind of scenerio is well documented for the Moon except that pressures are too low for extensive eclogite fractionation. Eclogite fractionation removes Al_2O_3 from the melt and reduces the thickness of an early anorthositic crust. Eclogite cumulates sink no deeper than 670 km on Earth because of the intervention of the ilmenite and perovskite phase changes in a

peridotitic mantle (Anderson, 1979c). An eclogite cumulate layer will be deeper and smaller in Venus because of the effect of temperature and pressure on the phase boundaries and the limited stability field of eclogite in the upper mantle. Nevertheless, the early tectonics, geochemical differentation and outgassing of the two planets were likely to have been similar.

If the outer layer becomes unstable as the planet cools, the situation changes. Subducted slabs affect both the flow and thermal regime of the upper mantle. The sinking of the cold boundary layer deep into the upper mantle drags cold isotherms to depth and allows hot replacement material to rise at ridges. This leads to relatively large lateral and vertical temperature differences in the upper mantle and rapid convection in which the surface boundary layer participates.

The mantle of the Earth contains approximately 15-20% of a basaltic component. If the mantle were well differentiated this would give a crustal layer 350-450 km in thickness. Only the top 50 km would be buoyant, because of the eclogite transformation, but this is still more than 4 times the present average crustal thickness. The lunar and Martian crusts are much closer to being the appropriate thickness for a well differentiated planet in spite of the fact that accretional heating must have been less for these bodies. The difference, of course, is due to the continuous removal of basaltic crust from the surface implying a long-term storage reservoir at depth. The size of this reservoir has been estimated to represent about 20% of the mantle (Anderson, 1980).

Samples thought to be representative of the upper mantle are depleted in Al_2O_3, CaO and SiO_2 compared to cosmochemical estimates of

mantle composition. They are also depleted in trace elements which are retained by the eclogite minerals, garnet and clonopyroxene. The degree of depletion implies an eclogite layer about 450 km thick which is probably deeper than 200 km.

The ^{40}Ar abundance for Venus is about an order of magnitude less than for the Earth (Pollack and Black, 1979). This suggests that late outgassing has been less efficient for Venus than for the Earth in spite of the higher surface and upper mantle temperatures. This is easily understood with the present model. On Venus, mantle melts and their volatiles mainly have access to the surface through relatively transient rifts. Because of the conservation of near surface material these rifts can only form when the crust compresses or thickens somewhere else on the planet. The amount of early outgassing on the two planets may have been similiar but Venus did not experience the early lithospheric overturn event or the outgassing associated with continuous crustal renewal at ridge axes.

Conclusions

The high surface temperature of Venus has several important tectonic implications. The most significant is the small amount of cooling that the lithosphere experiences before it reaches thermal equilibrium. This increases its buoyancy and long-term stability at the surface when compared with the terrestrial oceanic lithosphere. Since crust cannot be recycled into the mantle the thickness of the basaltic crust on Venus is much greater than on Earth. The combination of a thick crust and high temperatures decreases the thermal conductivity of

the lithosphere. This means that for a given mantle heat flow the temperature gradient in the lithosphere is greater than would be the case for a cold lithosphere with a thin crust. This leads to high upper mantle temperatures, low viscosities and the possibility of partial melting at relatively shallow depths.

Acknowledgements

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HOTSPOTS, BASALTS AND THE EVOLUTION OF THE MANTLE

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Hotspots, Basalts and the Evolution of the Mantle

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Revised: July 23, 1980

Abstract

Trace element concentration patterns of continental and ocean island basalts and of mid-ocean ridge basalts are complementary. The relative sizes of the source regions for these fundamentally different basalt types can be estimated from the trace element enrichment/depletion patterns. Their combined volume occupies most of the mantle above the 670 km discontinuity. The separate source regions are the result of early mantle differentiation and crystal fractionation from the resulting melt. The MORB source evolved from an eclogite cumulate which lost its late stage enriched fluids at various times to the shallower mantle and continental crust. The MORB source is primarily garnet and clinopyroxene, while the continental and ocean island basalt source is a garnet peridotite that has experienced secondary enrichment. These relationships are consistent with the evolution of a terrestrial magma ocean.

Introduction

Hotspots, or plumes, have not yet been fitted satisfactorily into either the tectonic or geochemical framework for the evolution of the mantle. 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 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 we designate MORBS, are still uncertain. It appears to be very large, homogeneous, global in extent, and to be the result of a previous differentiation or fractionation process that depleted it, relative to other source regions, of most of the incompatible elements. 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% 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 continental interiors, rift zones, oceanic islands, anomalous ridge segments and island arcs, mantle xenoliths from kimberlites and alkali basalts, and kimberlites themselves, all indicate the presence of a mantle reservoir with trace element concentrations that are also complementary to trace element concentrations in MORB. This reservoir apparently is also global and can provide magmas to a variety of tectonic environments including all types of plate boundaries as well as plate interiors. These magmas, of which alkali basalts are one example, are not nearly as voluminous as MORB but they indicate the presence of an enriched region or layer which may be substantial compared to the continental crust. We shall designate this as the hotspot or plume reservoir, or simply PLUME (5), and assume that it is the source of alkali basalts, nephelinites, melilitites, basanites, kimberlites and continental tholeiites which it provides by varying degrees of partial melting (6,7). One of the diagnostic signatures of these basalts is the rare-Earth element (REE) pattern which implies a source enriched by about 10 in LREE and 3 in the HREE, relative to chondrites (6,7,8).

PLUME basalts also have high ratios of Rb/Sr, Rb/K, Ba/K, Nd/Sm, Ba/Nd, Ba/Sr, Nb/Zr, $^{87}Sr/^{86}Sr$, and $^{144}Nd/^{143}Nd$, and high concentrations

of H_2O, CO_2, Ti, K, Rb, Sr, Ba, La, Nd, Th, and U relative to MORBS. These characteristics are shared by continental and ocean island basalts and are also evident, albeit diluted, in back-arc basins, island arcs, and transitional or anomalous ridge tholeiites. Other characteristics that may be common but which have been studied in only a few hotspot locations are high abundances of Cl, F, Br and the primordial rare gases (9). The process that led to the enrichment of PLUME will have depleted other, presumably deeper, regions of the mantle. The absolute and relative sizes of these enriched and depleted reservoirs bear directly on the problem of the evolution of the mantle and, in particular, on the question of whether there is a large primitive reservoir in the mantle.

Basalt source regions

The continental crust is extremely enriched in the incompatible trace and minor elements. Continental tholeiites, basanites, nephelenites, alkali basalts, kimberlites and basalts from oceanic islands are also enriched. The complement to the depleted mid-ocean ridge source region may, therefore, be much more voluminous than just the continental crust. The trace element and isotopic affinities of these rocks, regardless of their tectonic setting, suggests that they are all derived from a similar source region. This reservoir, PLUME, has previously been referred to as continental mantle, the enriched source region or primitive mantle. Its location is uncertain but it appears to be a global layer. Proposals for its location include the uppermost (6,10) and lowermost (11) mantle. In trace element ratios

such as Ba/Nd, Nd/Sm, Ba/Sr, Ba/K and Rb/K there is a progressive
increase from ocean island and island arc basalts, basalts from
anomalous ridge segments, continental tholeiites, alkali basalts to
kimberlites. Mid-ocean ridge basalts have much lower ratios and
tholeiites from back-arc basins are generally slightly higher than MORB.
Trace element concentrations vary systematically through the
compositional spectrum olivine melilitite, olivine nephelinite,
basanite, alkali basalt, and olivine tholeiite and can be explained by
varying degrees of partial melting ranging from 4 to 25% from a common
source region which is enriched in the strongly incompatible elements
(Ba, Sr, Th, U, LREE) by 10x chondritic and in the moderately
incompatible elements (Ti, Zr, Hf, Y, HREE) by a factor of about 3
(6-8). Mid-ocean ridge basalts are derived from a very different source
which is depleted in the large-ion lithophile (LIL) elements. The
complementary nature of the two source regions is possibly due to the
migration from the MORB source region of a melt or fluid with
incompatible element concentrations similar to those of kimberlite.
This fluid depletes the MORB source region and enriches the
complementary mantle reservoir. The upper mantle low-velocity zone
(LVZ) is a likely repository of these volatile and trace element
enriched fluids and, therefore, a possible source region for the
enriched magmas which are characteristic of the plume or hotspot source.
Kimberlites come from depths as great as 220 km, consistent with an
origin near the bottom of the LVZ. The xenoliths in kimberlites and
alkali basalts represent mantle fragments from various shallower depths
and also generally exhibit enrichment of the type inferred for the plume

source region. There is now a wealth of evidence for uppermantle metasomatism (12, 30) and enrichment events at various times. The source of the enriching fluid has not been much discusssssed.

The complementary nature of continental and oceanic tholeiites is illustrated in the center part of Figure 1. Alkalii basalts, nephelinites, melilitites and basanites have similar but more enriched patterns than continental flood basalts, consistent with their derivation from the same source region by smaller degees of partial melting. Continental and abyssal tholeiites exhibit reflection symmetry about a line corresponding to a 6 to 7 fold enrichment over average mantle concentrations.

Continental and ocean island basalts are relatively depleted in those elements that are retained by garnet and clinopyroxene. The reverse is the case for abyssal tholeiites. Since garnet and cpx are reduced or eliminated during the large amounts of partial melting usually inferred for the formation of tholeiites (6), the inverse ga-cpx pattern of plume basalts must be related to a prior history involving eclogite fractionation or invasion of PLUME by a fluid that was in equilibrium with a ga-cpx assemblage.

The origin of island arc basalts is still controversial but they have PLUME affinities in such trace element ratios as Rb/K, Rb/Sr, Ba/K, Ba/Sr and Ba/Nd. These ratios are closer to ocean island basalts, alkali basalts and continental flood basalts than they are to mid-ocean

ridge basalts. Back-arc basin basalts are intermediate to MORB and hotspot magmas, suggesting a mixture from a deeper MORB source and an overlying enriched source region. Ocean island tholeiites also appear to be mixtures both in trace element and isotopic ratios (2). The fact that both MORB and basalts with continental affinities are available at ridge enrivonments, island arcs, oceanic islands and well developed continental rifts suggests a compositionally stratified mantle. The order of appearance of these basalt types is consistent with the shallower layer being enriched in the incompatible trace elements. In a stratified mantle the deeper layer should be denser and, therefore, more garnet rich. The, ratio of concentrations in MORB to those in continental tholeiites indicates that the MORB source region, although depleted in most of the incompatible trace elements, has selectively retained those that are most comfortable in the garnet and clinopyroxene lattices. This suggests that MORBS is an eclogite or garnet pyroxenite cumulate.

The composition and volumes of the two source regions

The concentrations of some key incompatible trace elements in various magma types, normalized to average mantle concentrations, are shown in Figure 1. Relative to the average mantle, continental tholeiites, alkali basalts, kimberlites and the continental crust are all most enriched in K, Rb, Ba, La and U. These are the elements that are most discriminated against by the major mantle minerals, olivine, orthopyroxene, clinopyroxene and garnet, i.e., they have the lowest

mineral/melt partition coefficients. The other elements are also strongly rejected by olivine and orthopyroxene, having partition coefficients of less than 0.03 for these minerals (6). On the other hand, garnet and clinopyroxene have partition coefficients greater than about 0.1 for Y, Nd, Sm, Sr, and Yb. These are the least enriched elements in the magmas with continental and ocean island affinities and the most enriched in MORB. It appears that garnet and clinopyroxene are abundant minerals in the MORB source region, and that the fluid which enriched PLUME was previously in equilibrium with a ga-cpx rich reservoir.

The high concentration of Y and Yb in MORB relative to other elements and relative to concentrations in basalts from PLUME, the enriched source region, is particularly significant. These elements have mineral/melt partition coefficients greater than unity for garnet. Mass balance calculations suggest that most of the terrestrial inventory of Y and Yb may be in MORBS. This can only be accomplished if garnet is a dominant phase and if most or all of the mantle has experienced differentiation. Partial melting of primitive mantle concentrates Y and Yb into the melt; crystallization of this melt at moderate presssure would concentrate these elements into an eclogite cumulate.

The relative volumes of the two source regions can be inferred from Figure 1 if it is assumed that they, plus the continental crust, give an undifferentiated terrestrial pattern for the very incompatible elements. Using 0.56% for the mass of continental crust relative to the mantle (3)

we compute that the mass of MORB must exceed the tholeiitic fraction of
PLUME by a factor of 12 to 15. Continental tholeiites represent about
20% melting of a peridotitic source region (6) and MORB must be mixed
with at least 15% olivine to recover the composition of its parent magma
(13,14). The residual crystals in both cases are assumed to be depleted
because of their very low partition coefficients (6,8). The MORB source
region is, therefore, at least 2.8 to 3.5 times the mass of PLUME and
the enrichment of the combined source regions, relative to primitive
mantle, is about 3.3 to 3.8. Note that MORB is depleted only in a
relative sense. Compared to average mantle abundances it is enriched.
This suggests that MORBS plus PLUME are complementary to the remainder
of the mantle from which they have presumably been removed by partial
melting. The enrichment factor implies whole mantle differentiation and
a residual mantle about 2-1/2 times the size of the combined reservoirs.

By assuming that the very incompatible elements have been entirely
fractionated into the two source regions it is possible to estimate
their total mass. For example, if La is depleted in the lower mantle
the combined source regions represent 26 to 30% of the mantle. This
corresponds to a thickness of 560 to 640 km in the upper mantle, or a
region extending upwards from the 670 km discontinuity to a depth of 27
to 110 km. If we take the transition region, 220-670 km, to be the
depleted source region (20), then PLUME is 130-160 km thick, about the
average thickness of the low-velocity zone (LVZ).

A similar calculation for the other strongly incompatible elements

(K, Rb, Ba, U) gives a range of 21-38% for MORBS plus PLUME. These are upper bounds since it has been assumed that the whole mantle has been processed and depleted. This, however, is a good approximation if olivine and orthopyroxene are the main residual phases. It is significant that the above estimates of the relative and absolute sizes of the principal mantle reservoirs correspond to the main subdivisions of the upper mantle, the LVZ and the transition region. The 670 km depth corresponds to a major seismic discontinuity and the maximum depth of earthquakes.

Other elements can be used to estimate the sizes of the individual reservoirs. The partition coefficients of Yb and Y are such that they strongly prefer the garnet structure (15). Both are enriched in MORB relative to PLUME, suggesting a higher portion of garnet in the former. If the entire mantle complement of Y and Yb reside in MORBS, then this would represent 14-17% of the mantle or an upper mantle equivalent thickness of 300 to 364 km. Sc is partitioned into garnet and clinopyroxene by a factor of 8 relative to olivine and orthopyroxene. Using this ratio of enrichment for the MORB source relative to the rest of the mantle it would constitute 26% of the mantle.

K, Rb, Ba, La and U have the highest relative concentrations in PLUME. These are the elements with the lowest partition coefficients for the major mantle minerals and those that are most likely to be concentrated into PLUME by melts or metasomatic fluids. The PLUME abundances of these elements can account for the entire mantle inventory

if PLUME is 9 to 19% of the mantle.

Using the above estimates of the relative sizes of the two source regions and Ganapathy and Anders (37) values for terrestrial abundances, we can now estimate the total abundances of the trace elements in the various reservoirs. The continental crust contains more than 26% of the mantle plus crust inventory of K, Rb, and Ba and less than 8% of Sr, Y, Zr, Nb, Sm, Yb, and Hf. The "depleted" MORB source contains more than 50% of the Earth's inventory of Y, Zr, Nd, Sm, Yb, Hf and Th and more than 20% of Nb, La and U. The PLUME source region, although generating highly enriched magmas contains only about 10 to 20% of the Earth's K, Rb, Sr, Nd, Sm, Yb, Th, and U. It is the primary repository for Ba and La and has small total abundances of Y, Zr, Nb, and Hf (less than 7%). The crust and the two upper mantle reservoirs account for about 75% of the heat production from K, Th, and U. Therefore, only 25% of the Earth's heat flow comes from the lower mantle. About 40% is due to MORBS. The continental crust and PLUME each contribute about 18%.

This distribution of heat sources affects the style of convection in the two reservoirs. Assumming PLUME to be the shallower reservoir it is primarily heated from below and will therefore be characterized by narrow ascending plumes. MORBS is primarily heated from within and will be characterized by broader ascending regions and narrow descending jets or "slabs".

The mineralogy of the two source regions

Assuming that the two source regions are related, an apparent partition coefficient can be determined by forming the ratio of the concentrations in MORB to those in continental tholeiites (CFB). This is shown in Figure 2 along with mineral/melt partition coefficients for garnet, clinopyroxene, orthopyroxene and olivine. Tholeiites are used in this comparison since they presumably are products of large degrees of partial melting. They will, therefore, have the trace element pattern of their source and a relatively uniform enrichment. The ratio of the concentrations in the two reservoirs can be explained if the MORB source is composed mainly of garnet and clinopyroxene and PLUME has been enriched with fluids that were in equilibrium with this layer and therefore depleted in such elements as Zr, Nb, Y, Yb, and Hf.

MORB is enriched with respect to continental tholeiites in Y and Zr and only slightly depleted in Hf and Yb. Both MORB and continental tholeiites represent rather large degrees of partial melting of their respective source region. A garnet peridotite would be expected to lose most of its garnet and clinopyroxene under these circumstances. The above results indicate that primary garnet must be a minor mineral in PLUME but a major component of the MORB source. MORB is depleted in the most incompatible elements, such as K, Rb, Ba, and U but not as much as would be expected if it were composed entirely of residual crystals. This suggests that MORBS evolved from a melt, presumably the result of an early differentation event, and consists of a mixture of melt with

excess garnet which has settled into it during crystallization, i.e. it is an orthocumulate. The upper part of the mantle would consist of lighter, cumulate and residual, crystals and late stage fluids which impart an inverse garnet signature to this region.

This sequence of events, based on trace elements, is precisely what O'Hara et al. (14) proposed on the basis of petrological and major element considerations. They concluded that the parent magma for ocean island tholeiites had experienced a previous history of eclogite extraction. The remaining magma evolved to ocean island tholeiite by olivine fractionation. On the basis of major element (20) and trace element chemistry, I suggest that the eclogite cumulates constitute the source region for mid-ocean ridge basalts. The parent magma from which both MORBS and PLUME evolved by crystal fractionation would be picritic, the result of extensive partial melting of a primitive garnet peridotite mantle. Similar considerations have led to the concept of a magma ocean on the moon (41). Because of the higher pressures in the Earth's mantle a deep eclogite cumulate is the analog of the floating plagioclase cumulate that forms the lunar highland crust.

In order to see if this holds up quantitatively for the trace elements we investigate the following model. The primitive mantle is split into two reservoirs by 15% partial melting and melt extraction to the surface. The melt fraction reservoir is enriched by about a factor of 6 in the incompatible elements but, because of the low partition coefficients, has nearly primitive ratios of Rb/Sr and Sm/Nd.

Crystallization of the melt yields an eclogite cumulate layer which is modelled as a 50:50 mix of garnet and melt, proportions appropriate for an orthocumulate. Part of the final 5% melt fraction, a melt that is in equilibrium with eclogite, is removed to deplete the cumulate layer and enrich the overlying mantle. The mixing ratios were adjusted so that the former can yield MORB by 15% olivine fractionation and the latter can yield CFB by 20% partial melting (38).

Results of this model are shown as solid points in Figure 2. They are consistent with MORB being the result of nearly complete melting of a source region which represents high pressure cumulates from the melt fraction of the primary differentiation, i.e. eclogite or garnet pyroxenite.

Prior to providing depleted MORB this region lost its late stage fluids or, alternatively, crystallized completely and then lost its early melt fraction on a subsequent reheating cycle. PLUME was enriched by upward migration of this fluid. Thus, two stages of differentation and fractionation are required in order to concentrate the LIL in the two upper mantle reservoirs. Isotopically, the MORB source evolves as nearly primitive mantle until it transfers its incompatible elements to PLUME. It subsequently evolves as a depleted reservoir. Basalts from PLUME will have time-integrated depleted, enriched or "primitive" isotopic ratios, depending on when the enrichment occurred.

Locations of the two source regions

There is variety of evidence that suggests that the plume source region is shallow. Anomalously low seismic velocities can be traced to a depth of about 250 km under Yellowstone but deeper velocities appear to be normal (17). The variable and small amounts of partial melting required to generate melilitites, nephelenites, basanites and alkali basalts (6) from a common source region are consistent with diapirs rising from various shallow depths. Tholeiites are the result of larger degrees of melting, consistent with adiabatic ascent from deeper levels. The xenoliths entrained in kimberlites and alkali basalts are volatile and trace element rich and these are samples from the upper 200 km of the mantle (12). Intuitively, we expect that volatiles will migrate upward and be trapped by the cold upper mantle. Volatile and LIL enriched magmas occur not only under continents but also at island arcs where the downgoing slab perturbs the upper mantle. Plume type basalts occur in continental rifts. These rifts evolve to oceanic ridges, with isolated oceanic island hotspots, when the rifting has led to the formation of an ocean basin.

There are also some suggestive geometric constraints. The volume of the low-velocity zone (LVZ) is adequate to provide the plume basalts but not the voluminous MORB (10). The preferred mode of convection in a region with a high temperature gradient and a rapidly varying viscosity takes the form of hexagonal cells with upwelling centers (18). This seems to be consistent with hotspot patterns (19) and the small areal

extent of hotspots. In more homogeneous regions of the mantle, where the temperature gradient is smaller and the viscosity more uniform, linear rolls are a possible mode of heat transport. Linear ridge systems and the uniformity of oceanic tholeiites therefore suggest a deeper source, one that is below the large vertical and lateral variations that occur in the upper 200 km. The sink of oceanic lithosphere, judging from the depth distribution of earthquakes, appears to be between 200 km and 670 km. The distinctive isotopic and trace element signature of MORBS can be maintained if it is also between these depths. Upper mantle temperatures are closest to the melting point between about 150 and 250 km depth and this is therefore the depth range where it is most likely for diapirs to originate.

It is difficult to estimate the depth at which partial melting first occurs. Seismic data from several hotspots give low velocities down to at least 150 km (17,21). Eruption temperatures of 1300°C and large degrees of partial melting require initiation of melting below 150-200 km (22). There is some evidence for crystal fractionation in melts as deep as 280 km (23). Mantle fragments brought up by alkali basalts and kimberlites do not, in general, have the trace element pattern required for the source region of MORB. Therefore, PLUME magmas may originate from shallow depths but MORBS must be deeper than 200 km. In a gravitationally stratified mantle one would expect the deeper layer to be denser and therefore rich in garnet. The inability of young buoyant lithosphere to subduct below about 220 km suggests a density increase at this depth and this may be the boundary between the two

source regions (20). Taking all evidence into account it appears that PLUME may be coincident with the LVZ and diapirs rising from the transition region supply the magmas that evolve to MORB.

A terrestrial magma ocean

The complementary LIL element patterns of the major terrestrial magma types is reminiscent of lunar data which has led to the widely accepted concept of crystal fractionation and cumulate formation in a magma ocean or vast lava lakes (41). The various source regions are attributed to cumulate and residual fluid layers that resulted from crystallization of a ~300 km thick magma ocean that, in turn, was derived from very early melting of at least half of the moon. If a body as small as the moon experienced such extensive differentiation and fractionation, then the Earth should have as well. The amount of partial melting required, ~15%, to explain the enrichment of the upper mantle reservoirs on Earth, is, in fact, relatively modest. Although the energy of accretion of the Earth is much greater than that of the moon, the greater size of the Earth results in high Rayleigh number convection and rapid increase of the melting curve with depth. This plus the latent heat buffer and the high melting temperature of olivine may prevent more extensive melting on a global scale. A 15% melt implies a magma ocean ~400 km deep. The pressures in the Earth are greater than on the moon. Early eclogite fractionation at depth therefore preempts the extensive plagioclase fractionation that resulted in the early lunar anorthositic crust.

Crystallization of a magma ocean will proceed from the base because of the relative slopes of the adiabat and the liquidus. The near liquidus phases at depths greater than 60 km are garnet and clinopyroxene (14, 16) and they will form an eclogite or garnet pyroxenite cumulate layer. Although there is a seismic discontinuity in the mantle near 400 km, eclogite is denser than residual garnet peridotite to depths of 670 km (20). These cumulates will therefore sink to this depth, displacing residual mantle upwards. The shallower part of the mantle will therefore consist of olivine and pyroxene cumulates and residual fluid from the magma ocean and displaced residual lower mantle. The top of the eclogite layer could be as shallow as 220 km (20) or as deep as 370 km based on considerations of this paper. The mantle discontinuity near 400 km, which is usually attributed to the olivine-spinel phase change could therefore be either a chemical discontinuity or could represent the completion of the garnet pyroxenite (eclogite) to garnetite (garnet solid-solution) phase change (42).

Isotopic ratios of the two source regions

The study of Sr and Nd isotopes places important time constraints on the evolution of mantle reservoirs (24-26). The continental crust and the depleted reservoir have mean ages of 1.5 b.y. (3). Kimberlites and continental flood basalts, although enriched in LIL, have been attributed to a primitive reservoir (2, 3, 24, 27). This interpretation is not required by the data as we show later. Some mantle samples have been derived from ancient enriched reservoirs (28). Other samples come

from reservoirs which apparently have been enriched only recently (28,29).

Consider a primitive mantle that partially melts and separates into two reservoirs, as before, by upward removal of the melt. The lower mantle consists of residual crystals and is therefore depleted in the incompatible elements. The melt fractionates into a deep eclogite cumulate layer and a shallow olivine-orthopyroxene cumulate layer. A 5% melt fraction is transferred from the deeper to the shallower layer at various times (40). The Nd and Sr isotopic ratios for the two reservoirs are shown in Figure 3. The central horizontal scale gives the ages of the depletion and enrichment events. The theoretical $\epsilon_{Nd}-\epsilon_{Sr}$ correlation lines agree with the data and indicate that the mid-ocean ridge basalt source region was depleted at times between 1.5 and 2.5×10^9 years ago. A redistribution of LIL makes it possible to satisfy the mantle isotopic data even if the primary differentiation occurred early in the history of the Earth. Thus, the trace element, petrological and thermal constraints on the evolution of the mantle and the various reservoirs are not contradicted by the isotopic data. The type of model investigated here is similar to previous ideas of mantle metasomatism and trace element redistribution (30).

Implications for mantle evolution

Isotopic studies indicate that the two major mantle reservoirs have been isolated for more than 1 or 2 b.y. (2,3). The evidence presented

here from trace elements, and previously from major element and seismic considerations (20), suggest that the mantle is chemically stratified and that the various regions of the mantle are complementary products of terrestrial differentiation. The present emphasis has been on magmas from these reservoirs but similar conclusions result from the study of solid fragments from the mantle (14, 20). O'Hara et al. (14) pointed out that the restricted number of phases found in bimineralic eclogites from kimberlites indicate that they are either crystal accumulates or crystalline residue developed in contact with a liquid. This is exactly the situation inferred from the trace element patterns for the source region of abyssal tholeiites. Eclogite and garnet peridotite xenoliths from kimberlites may represent samples from MORBS and PLUME, respectively. They give satisfactory average mantle compositions for the major oxides when combined in the portions indicated by the trace elements (20).

Although whole mantle convection, in the conventional sense, is precluded in a chemically stratified mantle, transfer of material into and out of the various reservoirs is possible because of the large volume change associated with partial melting and phase changes, e.g., basalt-eclogite. The oceanic part of the plate tectonic cycle may be summarized as follows. Partial melting in the eclogite layer allows diapirs to rise to the base of the oceanic lithosphere. Nearly complete melting occurs during adiabatic ascent. This is possible because of the proximity, in temperature, of the liquidus and the solidus in eclogite. Peridotite diapirs can only partially melt because of their high

liquidi. The resulting melt fractionates in near surface magma chambers to form tholeiitic melts which are light enough to rise to the surface. The pyroxenite residue forms the lower oceanic lithosphere. As the lithosphere cools and thickens the lower part transforms to garnet pyroxenite or eclogite which is denser than the underlying mantle. The oceanic lithosphere becomes gravitationally unstable and it returns to its source region which lies between about 220 or deeper and 670 km depth. Partial melting in the shallow enriched peridotite layer generates continental and ocean island basalts and a harzburgite residue, both of which are lighter than their source region. These PLUME products remain in the crust and upper mantle. Some volatiles are returned to PLUME by subducted sediments and hydrothermally altered oceanic crust. The high temperature gradient in the thermal boundary layer at the PLUME-MORBS interface brings temperatures there close to the melting point (20). When mature oceanic lithosphere passes over a hotspot it is quite likely that the lower pyroxenite part will contribute xenoliths to the alkalic magmas and, by remelting, contribute to the trace element and isotopic signatures of ocean island basalts. The evolution of the mantle is shown schematically in Figure 4.

It has generally been assumed that basalts of all kinds represent partial melts of peridotites. Mantle compositions based on this premise have much lower abundances of SiO_2, Al_2O_3, and CaO than models based on cosmochemical considerations. The evidence used to construct the petrological models is obtained from the upper 200 km of the mantle and the models are therefore strictly only valid for the uppermost mantle.

A thick eclogite layer serves to increase the abundances of the above components and can reconcile the cosmochemical and petrological interpretations. The possibility of an eclogite layer in the upper mantle was also discussed by Press (43).

The idea that eclogite may be the source for oceanic basalts is an old one (31) but has not been in favor in recent years. The main objection is that limited partial melting of eclogite does not generate a tholeiite. Extensive or complete melting is required and this has been thought to be unlikely.

However, eclogite has an extremely small melting interval (about 60°C) compared to peridotite (16, 32). Since diapirs cannot rise out of an eclogite layer into a less dense peridotite layer until they are already extensively molten, it requires only a small additional temperature rise, relative to the liquidus, to complete the melting. This can be accomplished in a rising diapir. It is still unknown why melting initiates in the first place. One possibility is insulation by the thick continental lithosphere.

Hotspots and ridges in the Atlantic and Indian oceans were beneath continental lithosphere prior to 200 m.y. ago. A large number of hotspots are presently under the relatively stationary continent of Africa. This suggests that continental insulation, which prevents mantle heat from being efficiently removed to the surface, may be the cause of the partial melting that ultimately results in ridge and

hotspot volcanism. The thick, 150 km, continental lithosphere (20) prevents diapiric uprise and extensive melting may be possible until the continent rifts and rapid spreading and heat removal can take place. It is not so obvious that this explanation holds for the ridges and hotspots in the Pacific. The thick crust under the oceanic plateaus in the central Pacific or a thick oceanic lithosphere, due to a period of slow spreading, could also serve to insulate the mantle, raise temperatures and cause extensive in-situ melting prior to diapiric ascent. Parts of the Pacific rim continents such as Alaska, Mexico, Central America and S.E Asia may also have been located in the the central Pacific prior to 200 m.y. ago. Once initiated, the rise of upper mantle diapirs should be rapid (33), possibly rapid enough to avoid crystal fractionation en route.

Conclusions

Garnet and possibly clinopyroxene control the complementary trace element patterns of MORB and plume basalts. I suggest that the source region for MORB was formed from an eclogite cumulate layer resulting from crystallization of the picritic melt fraction of the original differentiation of the Earth. The plume source is enriched in those elements that would be concentrated in a melt that was removed from a garnet rich region. This fluid, the result of an early partial melt or a late stage fluid from crystallizing eclogite in the MORB source, infiltrated the shallow mantle plume source at various times. This gives plume basalts an inverse garnet trace element signature relative

to MORB. The MORB and plume source regions together make up about 25-30% of the mantle. This requires that most, if not all, of the mantle has been processsed to obtain the observed enrichment. There is no need to invoke a large primitive reservoir in the mantle or deep, lower mantle plume sources. On the contrary, the plume source region appears to be shallow and may be coincident with the LVZ. The upper mantle transition region, 220-670 km, represents about 21% of the mantle and the LVZ is about one-third as large. This is about the ratio of the sizes of the two upper mantle reservoirs which is required to satisfy the trace element data. The MORB source is probably an eclogite cumulate while the plume source is probably garnet peridotite. Crystallization of a magma ocean would give this kind of upper mantle stratigraphy. Nd and Sr isotopic data indicates that the depletion of that part of the MORB source region being sampled today occurred over the interval 1.5 to 2.5 x 10^9 years ago. The fluid involved in the depletion/enrichment events, and in the formation of the continental crust, appears to be a partial melt or late stage residual fluid from an eclogite cumulate layer. The lower mantle is residual peridotite and, although it may be, convecting, it no longer communicates with the shallower reservoirs.

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   Latin and German antecedents as discussed in D. L. Anderson,
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38. The two source regions (MORBS and PLUME) are constructed as

follows. For primitive mantle (PM) we use the

Ganapathy-Anders (37) values and extract 15% melt (ME).

The LIL concentrations in ME and residual mantle (RM) are

calculated using partition coefficients appropriate for a

peridotitic residue (3, 6). The eclogite cumulate (EC)

is a 50:50 mix of garnet and equilibrium intercumulus

fluid (EIF). To obtain the depleted LIL pattern for

MORBS half of the final 5% melt (LSM) is removed.

Partition coefficients for an eclogite residue are

used at this stage. LSM is the enriching or metasomatic

fluid for PLUME which is composed of 10% LSM and 90%

RM. Thus, PM evolves to ME and RM by partial melting;

ME evolves to EC and a silica-undersaturated fluid by

eclogite fractionation. EC minus LSM yields MORBS.

RM plus LSM yields PLUME.

39. ϵ_{Nd} and ϵ_{Sr} are the $^{143}Nd/^{144}Nd$ and $^{87}Sr/^{86}Sr$

ratios expressed as the fractional deviation in parts of 10^4

from those in a primitive, undifferentiated reference reservoir (2).

Thus, $\epsilon_{Nd} = 0$ implies a primoridal mantle with the following

qualifications. Small degrees of partial melting fractionate Rb/Sr

and Sm/Nd in both the melt and the residual crystals and they

subsequently evolve as enriched and depleted reservoirs,

respectively. For large degrees of partial melting, > 15%, the

melt has near primitive ratios of Rb/Sr and Sm/Nd. Even large

amounts of olivine and orthopyroxene fractionation from such a melt

do not much affect these ratios. Garnet and clinopyroxene

orthocumulates, i.e., nearly equal proportions of crystals and melt,

also give nearly unfractionated ratios. These results follow from

the small mineral/melt partition coefficients and the complementary

concentrations in garnet and its intercumulus fluid.

40. The following model gives the observed $\epsilon_{Nd}-\epsilon_{Sr}$

correlation. As before, 15% melt is extracted from the primitive

reservoir. The proto-MORBS cumulate is 60% melt, 40% garnet. This

is slightly different than the ratios used in the previous

calculation since we are now requiring a fit for only Rb, Sr, Sm and

Nd. For present purposes this difference is not significant. The

depletion of this reservoir is modelled by extracting 5% melt at

various times. This melt extract is used to form the continental

crust and to enrich, or metasomatise, PLUME. The enriched upper

mantle reservoir is modelled as 4% melt extract from the eclogite

cumulate, 21% of the original melt and 75% of the depleted residue of the original differentiation. The enrichment event occurs at various times.

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44. This research was supported by NASA Grant No. NGL05-002-069. I would like to thank E. Stolper, A. E. Ringwood, P. Wyllie, A. Boettcher, G. J. Wasserburg, C. Allegre, S. Jacobson, L. Silver, H. Taylor, and D. Green for helpful conversations. R. Oxburgh, J. G. Schilling, P. Wyllie and G. Ernst reviewed earlier versions of the manuscript. Art Boettcher kindly provided a most useful preprint co-authored with J. O'Neil, and made some important suggestions. I thank an anonymous reviewer for his editorial comments. Contribution Number 3432, Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena, California 91125.

Figure Captions

Figure 1. Normalized trace element concentrations in the continental
crust (dots), continental basalts and mid-ocean ridge basalts
(2, 6, 8, 34, 35). All concentrations are normalized to
terrestrial values of Ganapathy and Anders (37), recalculated
to mantle equivalents. The lower curves for continental
tholeiites and MORB are normalized to mantle concentrations
after removal of the continental crust, as are the curves for
alkali basalts and kimberlites. The complementary nature of
MORB and the other magma types is evident. The continental
crust is not the only "enriched" reservoir.

Figure 2. Crystal/liquid partition coefficients for garnet,
clinopyroxene, orthopyroxene and olivine (6). Heavy line is
ratio of concentrations in primitive mid-ocean ridge basalts
(36) and continental tholeiites (2, 6). This indicates that
MORB results from melting of a garnet and clinopyroxene rich
source region and that the continental tholeiite (CFB) source
region has experienced eclogite extraction. The points are
for a model (38) in which the proto-MORB source region is an
eclogite orthocumulate from the initial melt fraction of
primitive mantle differentiation. The continental and ocean
island basalt, or plume, source region is depleted residue
enriched by a late stage melt from the proto-MORB cumulate
layer.

Figure 3. Present ϵ_{nd} - ϵ_{sr} values (39) for an eclogite cumulate (solid line) which has been depleted at various times by removal of a melt fraction (40). This melt fraction enriches the complimentary reservoir (dashed line). Data from references 2, 24, 25, 29, and 30.

Figure 4. A model for the evolution of the mantle. Primitive mantle (1) is partially molten either during accretion or by subsequent whole mantle convection which brings the entire mantle across the solidus at shallow depths. LIL elements are concentrated in the melt. The deep magma ocean (2) fractionates into a thin plagioclase-rich surface layer and deeper olivine-rich and garnet-rich cumulate layers (3). Late stage melts in the eclogite cumulate are removed (4) to form the continental crust (c.c.), enrich the peridotite layer and deplete MORBS, the source region of oceanic crust (o.c.) and lower oceanic lithosphere. Partial melting of PLUME (5) generates continental flood basalts (CFB), ocean island basalts (OIB) and other enriched magmas, leaving a depleted residue (harzburgite) that stays in the upper mantle.
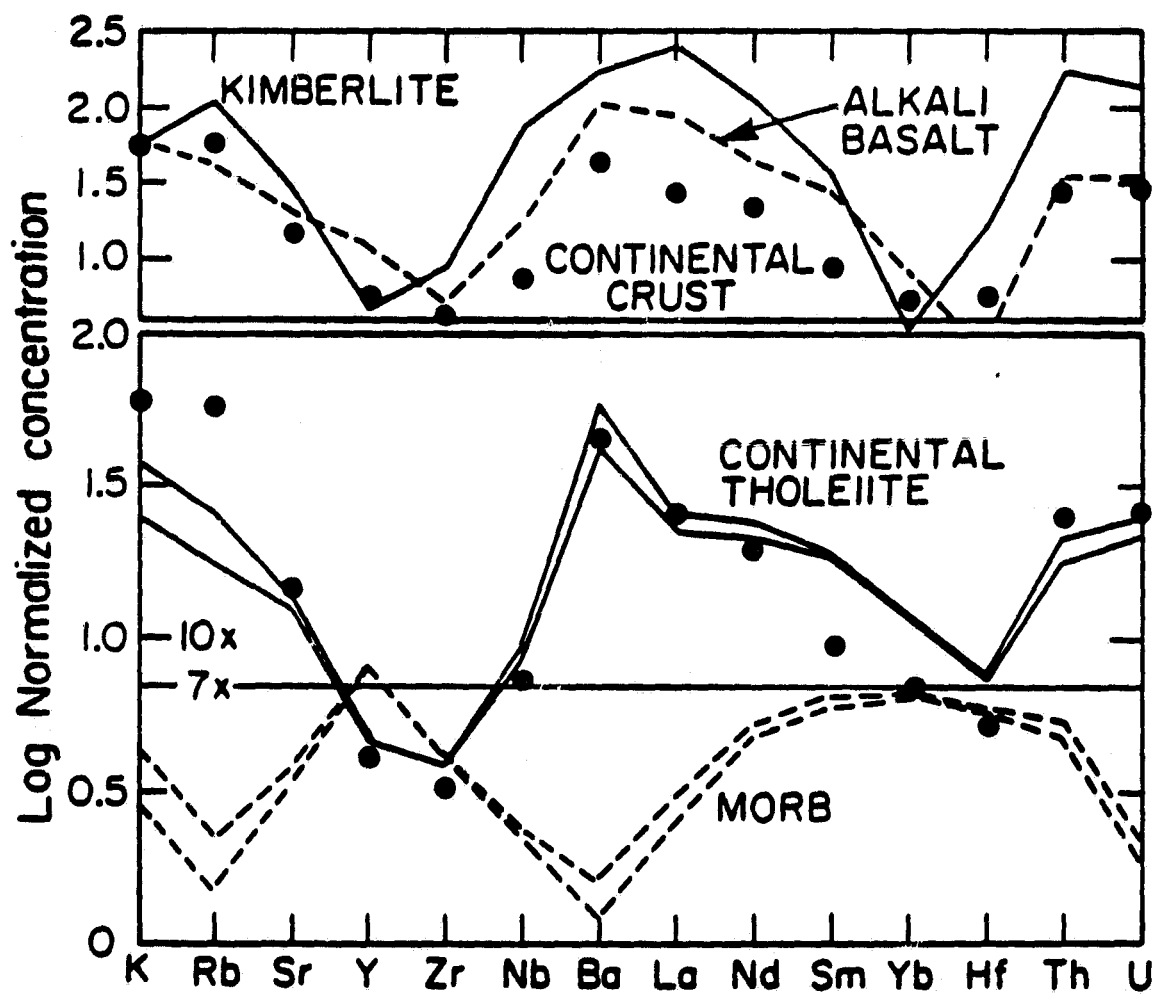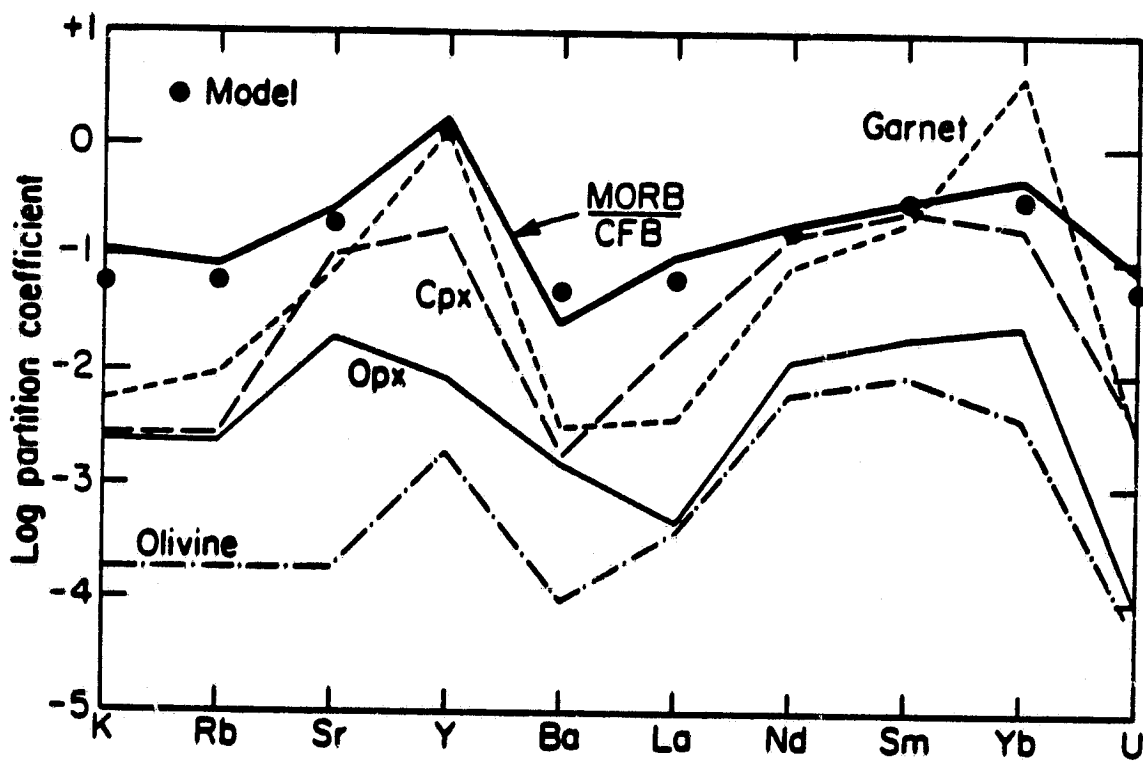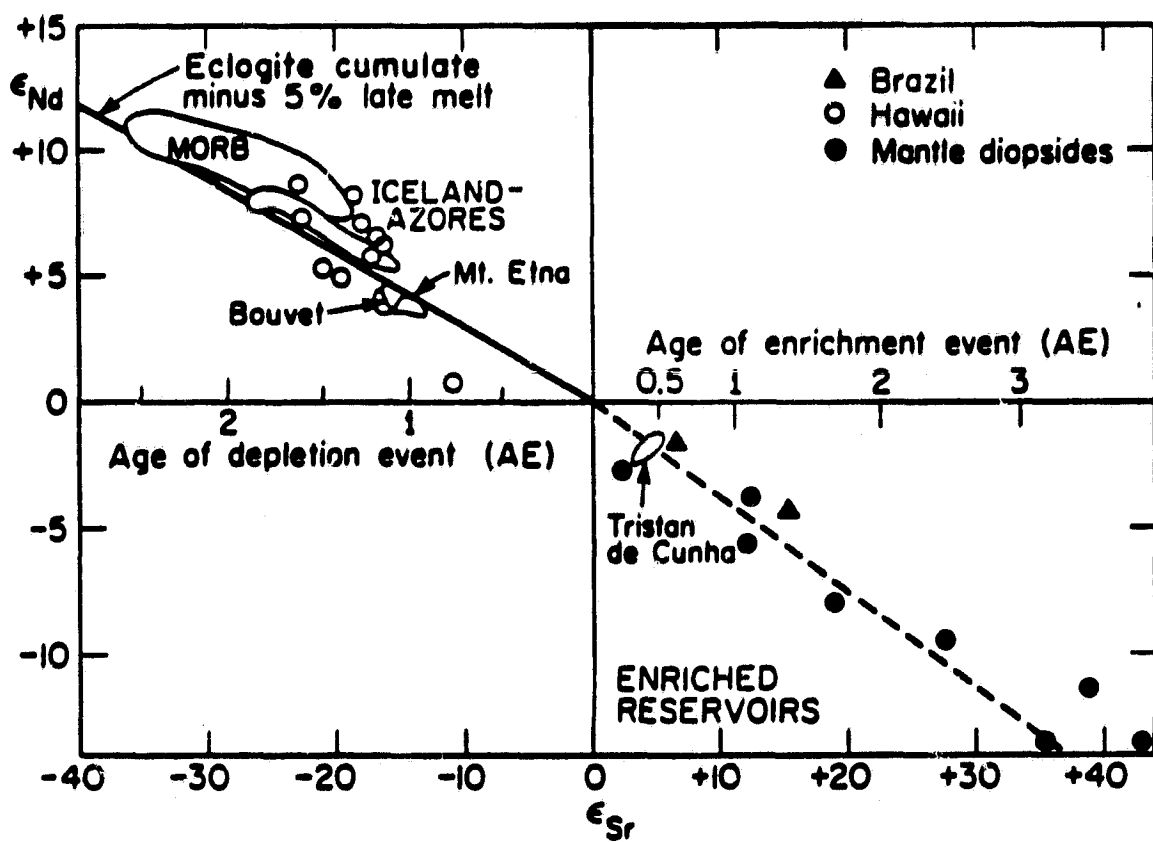
Fig. 1

Fig. 2

Fig. 3

CFB
OIB

MORB

| ① | ② | ③ | ④ | ⑤ |

PRIMITIVE MANTLE

MELT FRACTION

RESIDUE

PRIMARY DIFFERENTIATION

COOLING AND GRAVITATIONAL STRATIFICATION

BASALT

PERIDOTITE

ECLOGITE

C.C.

O.C.

LIL MORBS

DEPLETION-ENRICHMENT EVENTS

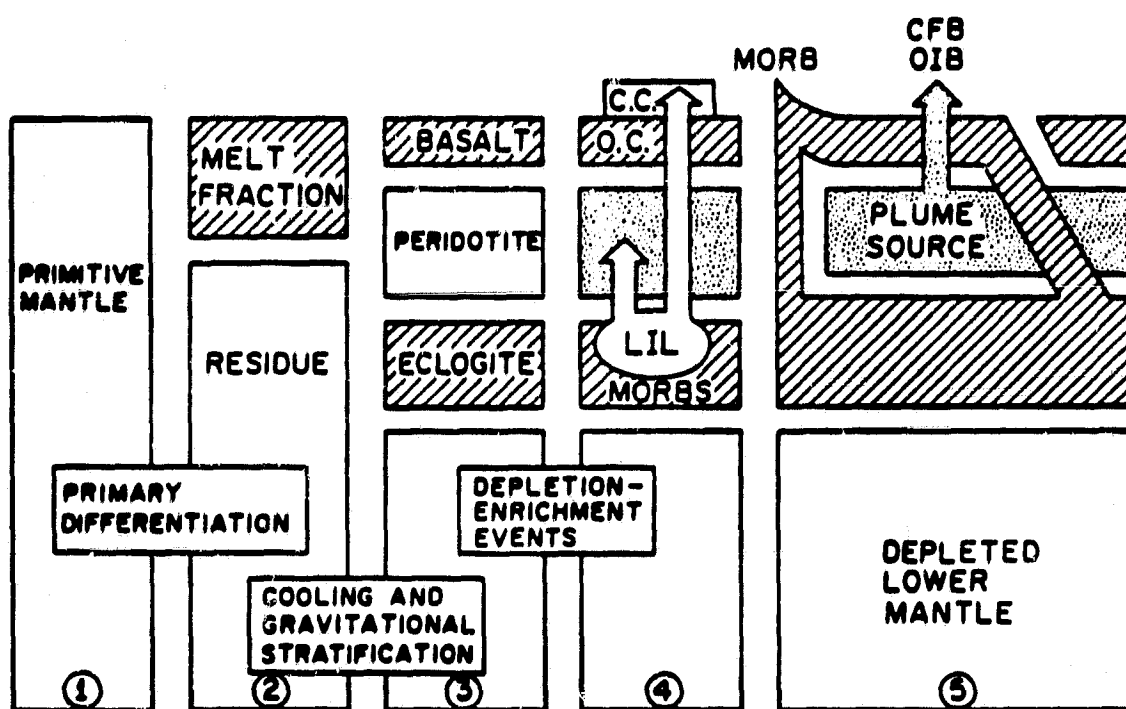PLUME SOURCE

DEPLETED LOWER MANTLE

Fig. 4

BASALTIC VOLCANISM CONTRIBUTION

Don L. Anderson

Basaltic Volcanism Contribution

by

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$\angle - 2$

IV. Evidence from Geophysics

The earth is the best studied planet and it is reasonable to test
cosmological models against the structure and composition of the earth
as deduced from seismology or even to proceed in the opposite direction.

The earth appears to be a differentiated body with the crust, mantle,
core and inner core being the main subdivisions. Presently forming oceanic
crust can be understood as a differentiated, or partial melt fraction,
of the upper mantle and petrological models of the upper mantle agree
well with the seismological data (1, 2). There is also no great difficulty
in understanding most other crustal rocks in terms of a similar origin
involving either primitive melts or melting of reworked crustal material.
The older massive anorthosite complexes may, however, involve processes
or materials that were unique to the earliest history of the earth.

The seismic data for the upper mantle (above 400 km) is well explained
by a mixture of olivine, pyroxene and garnet in the proportions suggested
from petrological studies. The seismic discontinuity at 400 km can be
explained in terms of the olivine-β-spinel phase change in a predominately
olivine upper mantle, and there is no need to invoke any changes in com-
position down to at least 600 km. The general increase in seismic velocity
and density is well explained by self-compression and by fairly well
understood phase changes in the olivine-pyroxene-garnet system. The olivine
successively transforms to the β-spinel and then to the spinel structure,
involving a total density increase of about 10%.

The pyroxene and garnet components interact to form a complex garnet

solid solution so that at 650 km the mantle is composed of about equal
parts of spinel and garnet.

The 650 km discontinuity in the mantle represents a large increase
in both density and seismic velocity. It cannot be attributed to any
known phase change in the olivine component (12) and must therefore involve
the pyroxene-garnet component if it is to be interpreted in terms of a
phase change alone. There is evidence, however, that the lower mantle
does not have the same composition as the upper mantle and that the 650
km discontinuity represents a chemical, as well as a phase, boundary (1-7).
The absence of earthquakes below 650 km can be interpreted in terms of a
density barrier, due to a change in chemistry, although other interpretations
are possible. The seismic data for the lower mantle is consistent with
SiO_2 and FeO enrichment, compared to the upper mantle, and with a composition
close to ordinary chondrites (1).

The lowermost mantle has unusual seismic properties and appears to
differ in composition from the bulk of the lower mantle. It can be ex-
plained in terms of Al_2O_3 and CaO enrichment (7) which implies enrichment
in U and Th, providing a mechanism for driving the core dynamo by differential
heating from above.

The outer core is molten and has the properties of iron with about 10%
of a lighter element or compound which is probably S or O. The inner core
is solid and, within the uncertainties of present data, can have the same
composition as the outer core or be pure Fe or Fe-Ni.

The relative amount of iron in the earth is significantly greater
than chondritic abundances. It is also greater than inferred for the Moon,
Mars and possibly Venus and less than Mercury. In terms of the condensation
hypothesis iron is a refractory. The earth is then presumably also enriched

in other refractories such as Ca-Al silicates and oxides and U and Th.
The lowermost mantle may still be enriched in these refractories.

The bulk of the lower mantle appears to be closer to pyroxene than to
olivine stoichiometry and closer to chondritic than upper mantle composition,
at least in terms of the Mg, Fe, and Si ratios (1). This is an important
boundary condition which is not generally considered in cosmochemical or
petrological modeling. The bulk compositions of the terrestrial planets
probably have higher Fe/Mg and Si/Mg ratios than the present terrestrial
upper mantle which is probably a differentiate of the lower mantle. Further-
more, from thermal and energy considerations this differentiation is likely
to be more efficient on the earth. The source region for basalts on
smaller bodies is likely also to be enriched in Fe and Si relative to the
source region of terrestrial basalts. In addition, the source regions of
the refractory planets, Mercury and the Moon, are probably enriched in Ca
and Al relative to both the earth and chondrites.

The origin of the earth's core is difficult to understand in the frame-
work of the homogeneous accretion hypothesis. However, the central concen-
tration of U and Th and possibly [26]Al makes it easy to melt the core and
maintain the dynamo and the associated refractories can explain the lower-
most mantle (8, 9, 13).

Planetary Modelling

Although the earth can be used as a guide in constructing models of the terrestrial planets the above differences should be kept in mind, i.e., 1) the earth is enriched in iron and possibly other refractories compared to other inner solar system objects with the exception of Mercury and 2) the upper mantle of the earth is not representative of the average composition of the mantle. The phase changes in a purely pyroxene planet are quite different from those in a mainly olivine planet.

As a first cut, however, the density depth profiles, and hence the mean density and moment of inertia of the terrestrial planets can be constructed from the terrestrial density-pressure curve. Thermal models can be used to correct for differences in temperature. This affects not only the density versus depth curves but also the locations of the temperature sensitive phase boundaries. This can have a large effect on the inferred mantle composition.

This approach was used by Anderson (10) on Mars and it was concluded that the Martian mantle was enriched in FeO compared to the terrestrial upper mantle and that a generally satisfactory model could be constructed with a chrondritic starting material. The moment of inertia implies that Mars has a small core. The data can be satisfied by an incompletely differentiated carbonaceous chrondrite (minus most of the water) or a mixture of chrondrites.

When the iron content of a planetary mantle is increased the location of the equilibrium phase boundary between olivine and the post-olivine phases shifts to lower pressures or shallower in the planet. The effects

of composition and temperature on the location of phase boundaries are
more important than refinements in the equations of state. In small
planets, such as the Moon and Mars,temperature is as important as pressure
in controlling the density but temperature, of course, is much more uncertain
than pressure.

Pressure vs. depth is almost an invarient in planetary models that
have the same radius, mass and moment of inertia. It can be computed
from the assumption of hydrostaticity. Even if the planet is grossly out
of hydrostatic equilibrium the effective pressure, which controls the
density, can be estimated quite accurately. The actual density at a given
depth is another matter. There is a trade-off between temperature and
composition. When we have only two parameters, such as mean density and
moment of inertia, we can only determine two parameters and these can be
for example, two of the following three: density of the mantle, density
of the core and radius of the core. If information about the crust is
available, from gravity or seismology, this can be stripped off.

For small planets, where pressure is not great, a very simple equation
of state can be used, such as the Murnighan or Birch-Murnighan or, as
discussed above, the "terrestrial equation-of-state." The zero-pressure
density in these equations is a variable, to be determined by the mass
and moment of inertia, or a parameter set by assumptions about composition.

A simple equation of state and one which is adequate for small planets
is the Murnaghan equation,

$$P = \frac{K_o}{n} [(\rho/\rho_o)^n - 1]$$

where K_0 is the zero-pressure bulk modulus, ρ is the density, ρ_0 is the zero pressure density and n is the pressure derivative of the bulk modulus, $n = (\partial K/\partial P)_0$, which is generally from 3 to 4. To a first approximation the effect of temperature can be included simply by writing,

$$P = (k(T)/n) \; [(\rho/\rho_0(T)^n - 1]$$

A higher order equation is the Birch-Murnighan equation which is a member of a class of equations which can be written in general form,

$$\rho = 3K_0 \; (m-n)^{-1}[(\rho/\rho_0)^{1/3(m+3)} - (\rho/\rho_0)^{1/3(n+3)}]$$

where m is generally 1 or 2 and n is 2 or 4 with n < m. Again, to first order the effect of temperature can be included in K_0 and ρ_0. Equivalent expressions can be written for the seismic velocities (11).

The pressure is,

$$P(z) = \int \rho(z) \; g(z) \; dz$$

where,

$$g = \frac{GM(r)}{r^2}$$

and z and r are depth and radius and g is the gravity.

Much more sophisticated equations have been applied to the earth but they depend on a detailed knowledge of the seismic velocities which are unknown on the other planets.

For the moon the effect of pressure is small and a direct comparison of seismic velocities can be made with only a small correction for pressure. Uncertainties in the temperature and temperature coefficients of the physical parameters are the main sources of uncertainty in constructing and inter-

preting models of Mars, Moon, Mercury and the asteroid parent bodies.

Thermal History Modelling

In order to calculate the thermal evolution of a planet the following must be known:

1) The temperature and velocity of the infalling material,

2) The rate of infall and the energy redistribution upon impact and embedment,

3) The ratio of retainment and loss of impacting material as a function of size, volatility, velocity and time,

4) The thermal and elastic properties of the material as a function of temperature and pressure,

5) The surface boundary condition and opacity of the environment as a function of time,

6) The radioactive abundances, half-lives and absolute time of initiation of accretion. Short-lived radiogens may be important if accretion is early and rapid,

7) The mode of cooling, i.e. conduction, radiation, convection,

and 8) The latent heats of all phase changes involved in impact and compression processes.

Clearly, approximations and assumptions must be made. The end member assumptions are a) that all kinetic energy is retained in the body and used up as specific heat and latent heat and b) the body grows in radiative equilibrium with the surroundings. The efficiency with which a body retains energy brought in depends on how deeply the particles are embedded, how much material is vaporized or ejected and how rapidly they are covered up. Thus, crater mechanics and ballistics are involved and it is important to know the size and impact velocity of the accreting material. Thus, a

large amount of information is required before the calculation can be started and some of these depend on cosmology, i.e. where are the particles coming from and what has been their previous history?

The question of homogeneous versus inhomogeneous accretion is important at the earliest stages. When vaporization and escape are important the planet retains only the refractories and "homogeneous" accretion actually leads to an inhomogeneous planet. Whether a planet starts off as refractory depends not only on the temperature in the nebula and the composition of the planetesimals but on where in the solar system the particles are originating. Advocates of pure homogeneous accretion will have to demonstrate that incoming planetesimals either escape vaporization or, if not, that the volatiles can be retained by the planet. These conditions can be expected to vary with time.

The present internal temperatures of a planet are under a little bit better control since the planet strives to evolve to an equilibrium configuration. A hot planet will simply approach equilibrium faster than a cold planet. The present internal temperatures simply represent a balance between heat production, conduction and convection, the latter two being a function of the initial temperatures. The boundary conditions are the surface temperature and if the gravity field is known, an estimate of the thickness of the cold outer boundary layer. The age and depth of origin of magmas is a potential constraint. High thermal gradients can be expected in the outer layers and a near adiabatic gradient at deeper layers. The temperature of the bulk of the interior is therefore set by the thickness and conductivity of the outer shell.

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