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Self-gravity, Self-consistency & Self-organization in Geodynamics & Geochemistry

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"...it is a privilege to see so much confusion."
Marianne Moore*, The steeple-jack*

It is widely believed that the results of seismology and geochemistry for mantle structure are discordant, with the former favoring whole-mantle convection and the later favoring layered convection. However, a different view arises from recognizing effects usually ignored in the construction of these models. Self-compression and expansion affect material properties that are important in all aspects of mantle geochemistry and dynamics, including the interpretation of tomographic images. Pressure compresses a solid and changes physical properties that depend on volume and does so in a highly non-linear way. Intrinsic, anelastic, composition and crystal structure effects also affect seismic velocities; temperature is not the only parameter. Shear velocity, or rigidity, is not a good proxy for density, temperature, composition or for other elastic constants such as bulk modulus. Scaling concepts play a central role in the analysis of complex systems and are particularly important in mantle dynamics, equations of state and wherever it is necessary to extend laboratory experiments to the parameter range of the Earth's mantle. Simple volume-scaling relations that permit extrapolation of laboratory experiments, in a thermodynamically self-consistent way, to deep mantle conditions include the quasiharmonic approximation but not the Boussinesq formalisms. Scaling relations can also be used when self-organized systems control boundary conditions and their own material properties, as in mantle convection. Whereas slabs, plates, and the upper thermal boundary layer of the mantle have characteristic thicknesses of hundreds of kilometers and lifetimes of the order of 100 million years, volume-scaling predicts values an order of magnitude higher for deep-mantle thermal boundary layers. This implies that deep mantle features are sluggish and ancient, consistent with previous investigations using depth and temperature parameterizations [Tackley, 1998]. Irreversible chemical stratification is consistent with these results, mainly because plausible temperature variations in the deep mantle cause density variations that are smaller than the likely density contrasts across chemical interfaces created by accretional differentiation and magmatic processes. Deep mantle features may be convectively isolated from upper mantle processes. The 650-km discontinuity is an isochemical phase change but major chemical boundaries may occur near 1000 and 2000 km depths. In contrast to standard geochemical models and recent

modifications, the deeper layers may not be accessible to surface volcanoes. Plate tectonics and surface geochemical cycles appear to be entirely restricted to the upper ~1000 km. There is no conflict between geophysical and geochemical data but a physical basis for standard geochemical mantle models, including the two-layer and whole mantle versions, and qualitative tomographic interpretations, has been lacking.

1. INTRODUCTION

There is a feeling of crisis and frustration amongst some workers in mantle geochemistry. Francis Albarede, in his Plenary Lecture of the European Union of Geosciences meeting in 2001, put it well;

"The paradigm of layered mantle convection was established nearly 20 years ago, mostly based on geochemical mass balance and heat budget arguments. It is now stumbling over the difficulty imposed by convection models to maintain a sharp interface in the mantle at mid-depth and by overwhelming tomographic evidence that at least some of the subducting lithospheric plates are currently reaching the core-mantle boundary. Discontinuities in the deep mantle… remain elusive…The present situation, however, remains frustrating because the reasons why the layered convection model was defended in the first place are still there and do not find a proper answer with the model of homogeneous mantle convection. First, the imbalance between heat flow and heat production (Urey ratio) requires that the deep mantle is rich in U, Th, and K. Second, the imbalance of some refractory lithophile elements between the composition of the Earth estimated with a homogeneous mantle and the composition of chondrites leaves a number of 'paradoxes' unresolved. Third, convective mixing should take place with a characteristic time of less than 1 Gy and should essentially wipe out mantle isotopic heterogeneities. In addition, frustrating evidence that the lower mantle hides a geochemical 'black box', with a non-primitive composition and hardly accessible to observation, is mounting."
[www.theconference.com/JConfAbs/6/Albarede.html]

Two years later Albarede (Albarede and Boyet, 2003) was still frustrated, and hints at the power of a false color image in changing minds that twenty years of developments in geophysics had not;

"For more than a decade, conflicting evidence between seismic tomography and isotope geochemistry of rare gases has thwarted the construction of a unifying convection model and blurred our vision of lower mantle chemistry and mineralogy. All body wave models vividly depict lithospheric plates penetrating the 660 km discontinuity..."

2. PRELUDE–ASSERTIONS & SEMANTICS

The purpose of this paper is to address the kinds of concerns raised above, and discussed more fully in Albared and van der Hilst (2002a,b), and to look beyond models motivated solely by isotopes, body waves and vivid images. These are simply small subsets of information that has long been available. Many unstated assumptions behind these models will be brought up front, reassessed and dropped. The idea that the mantle is subdivided into a small number of large, isotopically distinct, homogeneous and accessible reservoirs, separated by major seismic discontinuities, (e.g. DePaolo and Wasserburg, 1976; Allegre, 1982) has dominated thinking in mantle geochemistry and geodynamics for more than two decades, but it has never had a sound theoretical underpinning. The paradigm referred to above by Albarede has been called the standard model of mantle geochemistry (e.g.Hofmann, 1997 and many others). It is a two-box model with a semi-isolated leaky lower layer and a well-stirred and homogenized upper layer–above 650 km. This idea has generated a number of paradoxes and problems, including the current "crisis", and is responsible for a proliferation of increasingly complex, contrived and contradictory models for mantle structure, evolution and convection. The idea of an homogeneous, depleted (i.e. stripped of large-ion-lithophile–LILE–or crustal elements), well-stirred, upper mantle ('the convecting mantle') and an undepleted (retaining the original complement of all elements) gas-rich lower mantle, starting at the 650-km-deep olivine-spinel phase boundary, is a direct consequence of assumptions behind this model, not of data or calculation.

2.1 Conflicts and paradoxes

The notion that there is a conflict–"an enduring paradox"– between geophysics and geochemistry rather than simply strong evidence against the standard models of geochemistry is not new (Silver et al., 1988). There is, moreover, a large number of geochemical paradoxes–sometimes called enigmas or surprises–associated with what is nominally a model based on geochemistry. The view of conflict and paradox is widespread (e.g. van der Hilst et al., 1997; Coltice and Ricard, 1999; Becker et al., 1999; Helffrich and Wood, 2001; Ballentine et al., 2002). Few authors, however, itemize or evaluate the "conflicting" geophysical or geochemical evidence [see Appendix 1] but refer only to models which, in my view, are mainly a result of assumptions and *ad hoc* interpretations. My purpose here is to review the evidence and argue that falsification of a particular layered model dos not imply whole mantle convection.

van der Hilst et al. (1998) state their personal view thus; "*there is increasing consensus from seismological and geodynamical studies that slabs of subducted lithosphere sink deep into Earth's lower mantle and that present-day mantle convection is predominated by some form of whole mantle flow*". This was also the view of some in 1979 (e.g. Elsasser et al., 1979). According to Becker et al. (1999) "*…geochemists have argued for a layered mantle…whereas geophysicists…have supported a whole mantle view…[but] neither end-member scenerio is tenable*". These models will be referred to as the two-layer, or standard, model and the whole-mantle model; and together, as the standard models. Both are "whole mantle" in the sense that material from all depths in the mantle is assumed to be accessible to surface volcanoes. Both also assume that primordial undegassed regions still exist in the mantle. Some recent modifications to the standard models (e.g. van Thienen et al., 2005) retain the ideas that ocean island basalts must tap deep primitive reservoirs, that there must be deep radioactive-rich layers, and that slabs sink deeply into the lower mantle. Since, in the standard models, critical processes occur deep in the mantle, the effects of pressure are of prime importance. The standard models are built on geochemistry and seismic body-wave travel times, and do not address physical plausibility. Other geophysical techniques and physical considerations, however, have been brought to bear on the problem, with quite different results.

2.2 Geophysical data

Most geophysical data contradict the standard models [see Appendix 1]. For example, the geoid is inconsistent with a chemical discontinuity near 650 km (Hager et al., 1985). Mass balance calculations, dynamic topography [i.e. convectively induced topography, Wen and Anderson, 1997], and seismic data do not require or favor the kinds of models mentioned above [e.g. Anderson, 1989, 2002; Wen and Anderson, 1995; Trampert et al., 2004]. What can be ruled out are the original mass balance and primitive undegassed lower mantle models and models that ignore recycling (e.g. DePaolo and Wasserburg, 1976; Jacobsen and Wasserburg, 1979), radioactive-rich lower mantle models, whole-mantle convection schemes based on qualitative tomographic interpretations (e.g. van der Hilst et al., 1998), geoid models that do not satisfy long-wavelength topography, layered models that consider the 650-km phase boundary to be primarily a chemical or isotope boundary or the only plausible barrier to convection, and layered models with shear coupling across layers. It has been repeatedly suggested that if slabs appear to sink below 650 km they must sink to the core-mantle boundary. There are many other options. Mantle convection and stirring schemes based on heating from below and the Boussinesq approximation (see below) should probably not be viewed as definitive or realistic, for the mantle (Bunge et al., 2001; Tackley, 1998).

Different scenerios, mostly consistent with chemical stratification, have been proposed using petrological, major element and mineral physics considerations (e.g. Birch, 1952; Ringwood, 1966; Agee, 1990; Agee and Walker, 1988; Anderson, 1987b, 1989a,b; Mattern et al., 2005), and on the basis of quantitative analyses of tomography (e.g. Gu et al. 2001, Trampert et al., 2004; Anderson, 2002a; Ray and Anderson, 1994; Scrivner and Anderson, 1992; Wen and Anderson, 1995; Ishii and Tromp, 2004), sampling theory (Meibom and Anderson, 2003a,b), and dynamic topography and plate reconstructions (Wen and Anderson, 1997). These scenerios are based on inferences from data, not on a series of assumptions. They do not involve a "conflict between geophysics and geochemistry"; they are multidisciplinary, and evolved from mineral physics and petrology rather than isotope geochemistry. Geophysical data taken as a whole, including dynamic topography, quantitative tomographic interpretations, normal-mode data, plate reconstructions, and mineral physics, are consistent with a chemically and convectively inhomogeneous mantle but one that is different from the standard model that is currently under siege.

It is instructive that recent reviews and syntheses, and proposals for 'new mantle convection models' (e.g. Helffrich and Wood, 2001; Albarede and van der Hilst, 1999, 2002 a, b) do not refer to any of these petrological or mineral physics studies, or to the alternative and more quantitative tomographic interpretations referred to above. According to Helffrich and Wood (2001) "The seismic discontinuities at 410 and 660 km depth initially appeared to be likely boundaries for the compositional layering". This is not correct. The initial discovery of these discontinuities was immediately followed by calculations that showed that the depths and other features of these were consistent with isochemical phase changes, with the 410 km primarily due to the olivine-spinel phase change and the 650 km with a spinel-postspinel phase change with a negative Clapyron slope (Anderson,1967). Previously, the 400-1000 km depth interval was attributed to a spread-out phase change (e.g. Bullen, 1947; Birch, 1952; Ringwood, 1966). The phase change interpretation of the upper mantle seismic discontinuities was reviewed by Anderson (1969, 1970), Akimoto (1969) and Rinwood (1975). It was much later that isotope geochemists redefined the boundary of the lower mantle and attributed the 650 km discontinuity to a fundamental isotopic and undegassed boundary (e.g. DePaolo and Wasserburg, 1976); many geochemists have adopted this assumption, which is at the heart of the "crisis". For a review of the history of transition zone studies see http://www.mantleplumes.org/TransitionZone.html .

2.3 Why a crisis now?

Geophysical data and thermodynamic calculations have long been considered contradictory to the geochemical version of the layered-mantle model. Of note is the evidence for high-velocity material below 650 km (Jordan, 1977) and an analysis arguing that geoid data were inconsistent with a chemical discontinuity or a convective barrier near 650 km (Hager et al., 1985). Fundamental problems with the standard model–and alternatives to it, and the assumptions underlying it–had been developed by 1981 (e.g. Anderson, 1982a,b,c; Arnstrong, 1981; Zindler et al., 1984). Geochemical data, including isotopic data, have never required layered models, deep mantle or primitive reservoirs, deep recycling, or coincidence of isotope reservoir boundaries with seismic discontinuities (see, for example, Theory of the Earth, Anderson, 1989, hereafter TOE). It has been forgotten that these inferences are actually assumptions.

Geophysicists and petrologists have long known that the 650 km discontinuity could be explained well as an isochemical phase change (Anderson, 1967, 2002a; Akimoto, 1972). There has been no explanation of why a phase change should affect the isotopic composition of the mantle or why the process of crust extraction should have reached that depth at this point in time (some of the assumptions behind the standard model). Sixteen years ago Silver et al. (1988) were already referring to "Deep slabs, geochemical heterogeneity and the large-scale structure of mantle convection: Investigation of an enduring paradox" as a long-standing problem and Armstrong (1991) takes the story–which he calls a 'myth'–back even further. Isotopes, in fact, do not constrain the structure of the mantle and complete mass-balance calculations do not require a boundary at 650 km.

Why then do we suddenly have a crisis? What happened? Isotope geochemists who have long supported the standard model (e.g. Ballentine et al., 2002) most commonly refer to van der Hilst et al. (1997) and a widely reproduced dramatic color cross-section from this milestone paper. The visual impression of a tomographic image, and the plausibility or implausibility of layering and survival of isotopic heterogeneities in a convecting mantle are now at the heart of the perceived crisis. These are issues related to mineral properties, physics and scaling to large systems, not of isotope geochemistry or even seismology.

The view expressed by van der Hilst et al. (1998) –although widely held in the isotope geochemistry community–is not a consensus view of seismologists, primarily because body wave tomography is a powerful but imperfect tool. The results depend crucially on the ray geometry, which is constrained by the geometry of earthquakes and seismic stations, and the details of the mathematical techniques employed (Spakman and Nolet, 1988; Spakman et al., 1989; Shapiro and Ritzwoller, 2004). Moreover, the visual appearance of displayed results can vary depending on the color scheme and cross sections chosen. It is possible for the resulting images to contain artifacts that appear convincing (Spakman et al., 1989). Even further difficulties may result from the limitations of present algorithms, which do not correct completely for finite frequency, source and anisotropic effects.

Color cross sections are particularly ambiguous; although certainly vivid, they are not overwhelming, compelling or convincing evidence; they can be over-interpreted. There are issues of physical, geodynamic and petrological interpretations-whether 'blue' regions of the lower mantle–even if real–are unambiguous indicators of cold, dense material that started at the Earth's surface. This again is a mineral physics issue.

2.4 Ways out of the crisis, if there is one

Any inverse problem, including seismic tomography, must deal with limitations dictated by the distribution and quality of seismic data, and trade-offs between diverse structures within the Earth; it is likely that the Earth possesses a substantial component in the null space of any mix of seismic data (e.g. Shapiro and Ritzwoller, 2004). Methods are available to control the over-interpretation of data but these do not guarantee physical acceptability of the resulting model nor a model that resembles the real Earth. As Shapiro and Ritswoller emphasize, these limitations are fundamental. To produce realistic, physically plausible Earth models requires physical constraints to be applied (Shapiro and Ritzwoller, 2004).

Tomographic images are often interpreted in terms of an assumed velocity-density-temperature correlation, e.g. high shear velocity (blue) is attributed to cold dense slabs, and low shear velocity (red) are interpreted as hot rising blobs. There are many factors controlling shear velocity and some do not involve temperature or density. Cold, dense regions of the mantle can have low shear velocities (e.g. Presnall and Gudfinnsson, 2004; Trampert et al., 2004). Changes in composition can lower the shear velocity and increase the bulk modulus and/or density, as can verified by checking any extensive tabulation of elastic properties and densities of minerals. Ishii and Tromp (2004), for example, found negative correlations between velocity and density in the upper mantle.

A large range of seismological techniques and parameters are now available for the investigation of mantle structure; these include normal modes, surface waves, waveforms, cross-correlations, spectral density, matched filters, probabilistic methods, scattering, anisotropy, attenuation, moduli, and density. The geoid, dynamic topography and mantle response times provide further constraints. These paint a different–and richer–picture than is available from body wave travel times alone and the interpretation of the resulting color cross sections (e.g. Albarede and van der Hilst, 2002). For reviews of the situation regarding seismic modeling, and more current seismic views of the mantle–including uncertainties–see Lay (2005), Dziewonski (2004. 2005), Boschi and Dziewonski(1999); Vasco et al. (1994), Shapiro and Ritzwoller (2004); Julian (2005); Ritsema et al.(1999) and Ritsema (2005).

2.5 Assumptions and paradoxes

From its inception, the standard model had paradoxes; lead isotopes in general and the lead-paradoxes in particular, the helium-paradoxes and various heat flow and mass-balance paradoxes were unexplained. Paradoxes have, in fact, multiplied (e.g. Lenardic and Kaula, 1994; Hofmann, 1997; Anderson, 1998a,b; Ballentine et al., 1997; Bunge et al., 2001). Paradoxes are a result of paradigms and assumptions; sometimes we can make progress by dropping assumptions, even cherished ones. Sometimes new embellishments and complications to the standard model are made simply to overcome problems, or paradoxes, created by the original unphysical assumptions (e.g. Albarede and van der Hilst, 1999, 2002a,b; Ballentine et al., 2002; Kellogg et al., 1999).

Understanding how the mantle works requires a synthesis of petrology, mineral physics, seismology, geodesy, fluid dynamics, thermodynamics, heat flow, scaling relations and sampling theory; travel-times and isotopes are not enough. Mass balance and heat budget, constraints mentioned by Albarede, do not have spatial resolving power or depth perception. Tomography is an instantaneous snapshot of the Earth while heat flow integrates processes over some billion years or more, and throughout an unknown depth interval. Isotope geochemistry, tomography and convection simulations, in isolation, cannot constrain the dynamics of the mantle or the locations of 'reservoirs' (the other disciplines mentioned above have not been much involved in recent syntheses). However, quantitative tomographic and other geophysical data, and self-consistent convection calculations, guided by mineral physics and thermodynamics, can hopefully reduce the ambiguities.

In effect, the present contribution is the flip-side of a series of recent review and synthesis papers that present modifications to the standard models (e.g. Albarede and van der Hilst, 1999, 2002a,b; van der Hilst, 1997,1998), which represent one particular school of thought regarding mantle structure.

3. PHILOSOPHY AND GROUND RULES

In the following I discuss most of the issues raised in the Introduction (mass balance, heat budget, visual tomography, convection and mixing) but I concentrate on pressure, physics of materials, the initial state of the mantle, the role of complexity, and scaling and sampling theory–issues not much involved in the present debate but essential for its resolution. Debye theory, the Boussinesq approximation and the Rayleigh number must also be brought into the debate; their importance in scaling theory will be discussed below. These concepts are not explicitly recognized in the one- and two-dimensional mantle models formulated by body wave seismologists and isotope geochemists but the unstated assumptions behind these models are not always consistent with classical physics and thermodynamics .

These issues can be grouped into themes of self-gravitation (pressure), self-consistency (thermodynamics), and self-organization (thermo-chemical convection in a gravity field). The elements that have been missing in recent discussions of mantle evolution are accretion, petrology and mineral physics, and to some extent, statistics–particularly the central limit theorem. A major advance has recently been made by Trampert et al. (2004) with their probabilistic tomography and the integration of mineral physics and seismology.

Complexity is an emerging science at the edge of order and chaos. Thermodynamic systems, if unperturbed, evolve toward complete disorder; gravitating systems evolve to ordered layered structures. Far-from-equilibrium open systems become organized, but also become sensitive to small disturbances, changes in boundary conditions and modeling assumptions (e.g. Anderson, 2002b). For the Earth's mantle the Rayleigh number is one scaling parameter. It combines gravity, thermal and thermodynamic information to indicate where the mantle lies in the spectrum from static equilibrium–thermal and gravitational–through ordered to complex and chaotic. It cannot be estimated for the mantle without knowing the effects of pressure on the material properties and the style of mantle convection. In effect, it determines itself.

Complex systems have multiple states and sometimes convert from one to another with no apparent cause. This not only makes them difficult to model, but makes generalizations suspect. Is the Earth gravitational zoned into stable shells, or are some shells vigorously stirred and homogenized, as in the standard model? It is probable that mantle convection and plate tectonics are complex systems but it is not always clear which is the self-organizing agent, the mantle or the plates. If most heat sources and sinks are near the top of the system, if the lithosphere is stiff and if pressure is important, then plate tectonics is not just the surface expression of mantle convection; the top and bottom of the system are not equivalent. It is important to know, or investigate, the effects of pressure and the distribution of radioactive elements, in addition to standard concerns such as temperature and rheology. Pressure can make gravitational stratification irreversible, and the deeper layers inaccessible. This possibility alone resolves many of the geochemical paradoxes; there can be hidden or inaccessible regions or layers.

4. THEMES

4.1 Pressure and chemical stratification

Gravity is the restoring force in Rayleigh-Benard convection (but not in Benard-Marangoni surface driven convection). However, self-gravitation causes pressure, which compresses mantle minerals and changes their elastic, transport, dissipative and thermal properties. Pressure decreases interatomic distances in solids; this has strong non-linear effects on such properties as thermal expansion, conductivity, melting point and viscosity.

The complexity of realistic computer simulations of mantle evolution and dynamics is such that only a very small range of parameter space has been explored, and a thermodynamically self-consistent calculation of mantle convection has yet to be attempted (but see Ita and King, 1994, 1998). It would be useful to be able to restrict the number of models investigated to those that are thermodynamically consistent and plausible from a mineral physics point of view. Mineral physics and petrology-based models, although derived from simple physical rules, are structurally complex and therefore have not been tested by modelers, who have focused instead on simple starting models and steady-state conditions. If one side of Occam's razor is a simple model, the other side is simple rules.

The range of plausible lithologies in both the upper mantle and lower mantle is such that there may be only very small differences in the seismic velocities (e.g. Lee, 2003; Zhao and Anderson,

1994) particularly at high pressure. This is one reason for the controversy concerning lower-mantle composition, an element in the whole-mantle convection debate. The various candidate mantle rocks (pyrolite, chondrite, cosmic, perovskite) have similar elastic properties; also, chemistry trades off with temperature and mineralogy. Thus, compositional changes and interfaces are hard to detect compared to most phase changes. Some arguments for whole-mantle convection are based on the absence of obvious seismological evidence for layering and thermal boundary layers [TBLs], or the presence of high-velocity regions below 650 km [the arguments for and against layered models are given in Appendix 1].

It has recently become possible to resolve density as well as seismic velocities from seismic data and to separate the effects of temperature and composition (Ishii and Tromp, 2004, Trampert et al., 2004). In the deep mantle, low-shear wave velocity regions do not necessarily correspond to low density, low bulk modulus or high temperature. Variations in chemistry–possibly iron content–and mineralogy are as important as temperature (Trampert et al., 2004). The basic assumption in many tomographic interpretations, that low seismic velocity is always a proxy for high temperature and low density, is not valid. A counter-example is a cold dense eclogitic slab, which can be above its solidus at normal–or even cold–mantle temperatures, and hence have low seismic velocities. There is no correlation in properties between the upper mantle, mid-mantle, and lower mantle and no evidence for either deep slab penetration or continuous plume-like low-velocity upwellings (Ishii and Tromp, 2004, Trampert et al., 2004; Becker and Boschi, 2002); this is consistent with chemical stratification (Wen and Anderson, 1995,1997).

4.2 The initial state

Geodynamic and evolutionary models, and partial differential equations, require boundary conditions and initial conditions. The surface boundary condition is a continuously evolving system of oceanic and continental plates. The initial condition I prefer is based on the other edge of Occam's Razor. Although a homogeneous mantle with constant properties is the simplest imaginable assumption, no one has simply explained how the mantle may have arrived at such a state, except by slow, cold, homogeneous accretion. This is an unstated assumption in the standard models. The accretion of Earth was more likely to have been a violent high temperature process that involved repeated melting and vaporization and the probable end result was a hot, gravitationally differentiated body. That the Earth itself is efficiently differentiated there can be no doubt. Most crustal elements are in the crust, possibly all the ^{40}Ar–depending on the unknown potassium content–is in the atmosphere [TOE, Chapter 8; **http://resolver.caltech.edu/CaltechBook:1989.001]** and most of the siderophile elements are in the core. Given these circumstances, it is probable that the mantle is also zoned by chemistry and density. This raises more mineral physics questions e.g., What do chemical boundaries look like to seismology? What are the plausible ranges of densities of the silicate materials that form during accretion and mantle differentiation and are they adequate to maintain chemical stratification? When effects of pressure on mineral properties are taken into account, can the deep layers be accessible to surface volcanoes?

The assumed starting composition for the Earth is usually based on cosmic or meteoritic abundances. The refractory parts of carbonaceous, ordinary or enstatite chondrites are the usual choices (Ringwood, 1966; Javoy, 1995; Agee, 1990; TOE). These compositions predict that the lower mantle has more silicon than the olivine-rich buoyant shallow mantle and that only a small fraction of the mantle, or even the upper mantle, can be basaltic [e.g. TOE, Chapter 8]. The volatile components that are still in the Earth were most likely added to Earth as a late veneer after most of the mass had already been added and the planet had cooled to the point where it could retain volatiles. The other choices for starting compositions–considering the standard models–are 1. undegassed volatile-rich components with abundances of both refractory and volatile elements, including noble gases, the same as unfractionated carbonaceous chondrites (Kellogg and Wasserburg, 1990), and 2. whole-mantle compositions dictated by upper-mantle peridotites (the pyrolite model), the whole mantle convection model. The first option evolves to the standard model upon degassing and crustal extraction from the upper mantle.

A process of RAdial ZOne Refining (RAZOR) during accretion may remove incompatible and volatile elements and cause purified dense materials to sink (see TOE). Crystallizing magma oceans at the surface are part of this process. The formation of a deep reservoir by perovskite fractionation in a magma ocean, suggested by Agee and Walker (1988), is not necessary. The magma ocean may always have been shallower than the perovskite phase boundary but as the Earth accretes, the deeper layers will convert to high-pressure phases. There is no need for material in the upper mantle to have been in equilibrium with the dense phases that now exist at depth. Prior to the era of plate tectonics, the Earth was probably surfaced with thick crustal layers, which only later became dense enough to sink into the mantle. But because of the large stability field of garnet, there is a subduction barrier, currently near 600 km (Anderson, 1989b, 2002a). The great buoyancy of young and thick oceanic crust, particularly oceanic plateaus, and the low melting temperature of eclogite, and the subduction barrier to eclogite (and harzburgite) probably prevents formation of deep fertile and radioactive layers, even after the onset of plate tectonics.

The RAZOR process sets the initial stage for mantle evolution, including the distribution of radioactive elements. This step is often overlooked in geochemical and geodynamic models. The intial temperatures may have been forgotten but the stratification of major and radioactive elements may be permanent. An excellent summary of the initial conditions from a petrological point of view is given in Ringwood (1966).

4.3 *Petrological models*

Mantle models based on major element chemistry, mineral physics and petrology (e.g. Birch, 1952; Ringwood, 1979; Agee, 1990; Agee and Walker, 1988; Gasparik, 1997; Anderson, 1983,1987a) are more complex than the standard one- and two-layer models (e.g. Albarede and van der Hilst, 1999; Coltice and Ricard, 1999; Helffrick and Wood, 2001), hybrid models (e.g. Albarede and van der Hilst, 1999, 2002a,b), and convection models, which tend to ignore petrology, the effects of pressure, and the possibility of early and irreversible gravitational differentiation. A series of convection simulations with realistic physical and thermal parameters is needed to test these petrologic–and physics based–models; they are not based on the same kinds of assumptions that generated the paradoxes and crisis.

4.4 Sampling or stirring?

Assumption: the 'convecting' upper mantle is efficiently stirred and homogenized.
Corollary: non-MORB and enriched magmas come from the lower mantle.

The survival of layers, blobs, and reservoirs and homogenization by vigorous convective stirring, are issues in all current models of mantle structure and evolution. The existence and stability of narrow plumes are issues in models that assume strong heating from below and depth-independent properties. At high Rayleigh numbers, inhomogeneities in the mantle may be stretched, thinned, folded and stirred; this is the usual explanation for mid-ocean ridge basalts (MORB), which are noted for their homogeneity. Inefficient convective stirring leaves inhomogeneities in the mantle and this is sometimes used to explain 'anomalous' basalts in the single layer and mantle-blob schemes of homogeneous mantle convection (Coltice and Ricard, 1999; Becker et al., 1999), which are alternative to the standard model of isolated reservoirs. The usual assumption in these calculations is whole-mantle convection, uniform density, very high Rayleigh number (10^7-10^8), no pressure effects (the Boussinesq approximation–see below), steady-state, unidirectional stirring, Newtonian rheology, long stirring times and no plates or continents. Often the calculations are done in 2D Cartesian co-ordinates with uniform surfaces and no internal heating (e.g. van Keken et al., 2002, 2003). The experimenter has a great deal of control on the outcome.

For computational convenience mantle convection is often treated with the simplest scaling relation of all, the Boussinesq equations. These assume that density, or specific volume (V), is a function only of temperature (T) and that all other properties are independent of T, V and pressure (P).

In a convectively layered mantle with volume dependent properties, the effective Rayleigh number is low and chemical differentiation (gravitational stratification) rather than homogenization must be considered a possible outcome. The mantle may, left to itself, and allowed the necessary degrees of freedom–the essence of self-organization–behave in a way inconsistent with imposed boundary conditions, stirring history, and parameters. In the standard two-box model, stratification is due to removal of the crust from the upper layer and is unrelated to accretional or density stratification.

In spite of the assumptions underlying the standard model, isotope data do not constrain the shape, size or depth of mantle heterogeneities. The assumption that the whole upper mantle is efficiently stirred and homogenized underlies some of the geochemical paradoxes, and the rationale for recent hybrid models. Even if the mantle is convecting, the isotopic diversity of magmas may not be an issue of solid-state convection or large-scale stirring or layering (Zindler et al., 1984; Meibom and Anderson, 2003). Some stirring calculations give mixing times much greater than sampling times (Olsen et al., 1984), and much greater than the characteristic time of 1 Ga assumed by Albarede (2001). Another assumption in the standard model is that homogeneous basalts require a homogeneous source. The alternative to homogenization by convection in the solid state is magma blending during the sampling and eruption stage (TOE, p. 231), or the extraordinarily powerful central limit sampling theorem (Anderson, 2000a,b, 2001; Meibom and Anderson, 2003). It is not clear that the assumptions and conclusions in current models about homogenization, time constants, layering, spatial scales and steady-state are plausible or valid from a mineral physics perspective. In fact, there is a large range of results.

4.5 Distribution of radioactive elements

Assumption: the lower mantle is primordial, undegassed, or less degassed compared to the upper mantle; it is enriched or undepleted in U, Th and K.

Mass-balance and thermal constraints *are consistent with* the view that the radioactive elements are strongly concentrated into the crust and upper mantle (Clark and Turekian, 1979; Birch, 1952; Anderson, 1989, 2002a); [Note; the upper mantle–a seismological concept–is not equivalent to 'the MORB-source', 'the convecting mantle', 'the depleted upper mantle', DM or DUM of the standard models; the 'upper mantle' of the standard models–above 670 km–is not the same as the classical 'upper mantle' of seismology–above 1000 km (Bullen, 1947; Birch, 1952)]. There is a rapid decrease in the concentrations of the radioactive elements from the upper crust [U=2.8 ppm] through the mid-crust [1.6 ppm] and the lower crust [0.2 ppm] to lithospheric xenoliths [0.04–0.12 ppm] (Rudnick and Fountain, 1995; McDonough, 1994). Clark and Turekian [1979] suggested that this may be part of a general exponential decrease with depth in the mantle with a characteristic scale of 1000 km. They suggested that there may be essentially no radioactive elements in the deep mantle and there has been no subsequent evidence to refute this conjecture. This is a plausible but not unique interpretation, but it does contradict the common view that *the imbalance between heat flow and heat production…**requires** that the deep mantle is rich in U, Th, and K.* The "missing" U, Th and K may be in the upper mantle–in the EMORB components, in kimberlite and in other enriched components, some recycled (TOE, Chapter 8). Others have proposed deep radioactive-rich layers (Kellogg et al., 1999), assuming that the MORB-source fills up 'the upper mantle' as in the standard model. The conjecture that there are no enriched components in the upper mantle–that the whole upper mantle is uniform and depleted–is the source of some of the geochemical paradoxes. There is also the issue of whether one should use observed heatflow values or whether one should 'correct' them for a presumed missing hydrothermal component (Hofmeister and Criss, 2005); there may be no heatflow paradox or "missing" U, Th and K.

Some of the paradoxes in mantle dynamics can be traced to the assumption that the lower mantle has high concentrations of U, Th and K (e.g. Bunge et al., 2001). Convection models based on the Clark-Turekian or similar distributions (e.g. TOE) would undoubtedly behave differently than Boussinesq models with the reverse stratification and strong bottom heating. This type of model, and other types of realistic layered models, are harder to implement than the standard models and are beyond the capability of present programs. Modelers have been strongly influenced by the standard models and the vivid but perhaps misleading images that apparently show slabs plunging to the core-mantle boundary.

A plausible alternative layered model has almost all of the radioactivities, and other LILE, in the crust and upper mantle [e.g. TOE]. Secular cooling is also mainly in the upper mantle, so the existence of deep and large TBLs depends on the actual balance of heating and cooling of the upper layers.

4.6 Distribution of major elements

Assumption: the mantle is chemically homogeneous in the major elements, allowing whole mantle convection.

Seismological and mineralogical properties can be compared (1) by adiabatic decompression of seismologically determined mantle properties for comparison at zero pressure with laboratory data (e.g. Butler and Anderson, 1978; Anderson, 1989a; Stacey and Isaak, 2000), or (2) by extrapolation of measured mineral properties to lower mantle P-T conditions for comparison with seismologically determined velocity and density profiles (e.g. Sammis et al., 1970; Burdick and Anderson, 1975; Anderson and Bass, 1984; Duffy and Anderson, 1989; Zhao and Anderson, 1994). Conclusions using this approach are not robust; a variety of compositional models and temperatures are consistent with the data (Kiefer et al., 2002; Stacey and Isaak, 2001; Lee et al. 2004; Bunge et al., 2001; Karki et al., 2001; Organov 2001; de Silva et al., 2004), but this in itself explains some of the paradoxes such as the apparent absence–or invisibility–of compositional discontinuities.

Recent studies, imply or are **consistent with** an iron-rich high-temperature lower mantle (Lee et al., 2004; Stacey and Isaak, 2000; Mattern et al., 2004) or a lower layer that is chemically different from the upper mantle. Note that **consistent with** is not the same as **required** (Birch, 1952; Albarede, 2000). A chemically homogeneous mantle with no thermal boundary layers would have low potential temperatures in the deep mantle. Mattern et al. (2004) reviewed recent contributions in this field.

4.7 Self-consistency & self-organization

Thermal convection is a far-from-equilibrium self-organized thermodynamic system in a gravity field and is sensitive to initial conditions and changes in parameters (see, for example, Tackley, 2000; Anderson, 2002a,b). In the ideal convection calculation all parameters vary in a self-consistent way as a fluid parcel is advected into regions of different temperature and pressure. In the case of the mantle, the boundary conditions and material properties become part of the solution and artificially imposed conditions or preset parameters may not give results that are realistic or robust. The role of mineral physics in mantle geodynamics is to provide a way to test hypotheses and a way to assure physical consistency in tomographic and convection studies.

The parameters that control natural convecting systems at high temperature and pressure are interrelated. Thermodyamic variables are often indiscriminately assumed to be constant, or to vary independently, ignoring thermodynamic constraints that preclude such assumptions [Schubert, Turcotte, and Olsen, 2001].

Plate tectonics and mantle convection are different aspects of the same coupled system, yet mantle convection calculations do not exhibit plate tectonic behavior unless it is imposed by the modeler (Tackley, 2000). This may be because the top is not free to self-organize itself into plates (Anderson, 2001; Stein et al., 2004; Tackley, 2000) or that the system is driven–and organized–from the top.

4.7.1 What drives what, or is this a stupid question?

Assumption: mantle convection drives the plates; the plates ***are*** *mantle convection.*

If the plates are a self-organized-far-from-equilibrium [SOFFE] system and provide most of the buoyancy, dissipation and driving forces of geodynamics [Hager and O'Connell, 1979; Davies, 1988; Anderson, 2001,2002b], then mantle flow is organized by the plate-continent-slab system.

For a fluid with constant properties heated from below, driven to steady-state, the upper and lower boundary layers are symmetric and it makes no sense to talk about one part of the system driving another. At the onset of convection, however, it is clearly the newly unstable bottom boundary layer that drives the convection. The mantle is internally heated–at least the upper parts–and strongly cooled from the top, and plate motions are constantly evolving. If most of the driving forces (internal buoyancy) and dissipative forces (viscosity) are in the mantle then the mantle drives the system, even if plates and convection are one coupled system. The plates can be viewed simply as the upper boundary layer and the most visible part of the convection. If oceanic plates are the cold–and cooling–upper TBL [Davies, 1988], and if bending–and rubbing–of the plates is the main dissipative force (Conrad and Hager, 2001), and if there is no bottom heating–or if the buoyancy of the lower layer is small and slow to develop–then the dominant driving force is the negative buoyancy associated with cooling plates and subducted slabs. The importance of rising plumes depends on the conditions at the base of the system. Thus, it becomes important to know the time constants of the system and this involves the Rayleigh number, and the distribution of heating and cooling and dissipation.

Although material properties and time-scales can be estimated with scaling, details of the interactions in a non-linear self-regulating system are difficult to predict with simple scaling alone. The mantle may be dynamically a lot more interesting than the 'standard models' or current convection simulations. Scaling relations and the effects of extreme conditions raise the specter that the deep mantle may be in a completely different regime that the active layer at the top of the mantle. One cannot just assume that the upper mantle is vigorously convecting, or that the lower mantle is behaving as the upper mantle, or that sharp interfaces and TBL should be either obvious or precluded by convection calculations.

5. SCALING

It is often possible to progress in understanding complex systems that involve a variety of scales and conditions by identifying key variables that characterize the system on a particular scale, and postulating simple scaling relations between them. This may unify widely differing sets of data, a phenomenon called *universality*. When there is a single independent variable, these relations often take the form of power laws. Specific volume (rather than pressure, temperature or depth) is such a parameter. It may allow extrapolation of laboratory measurements of microscopic properties and behavior (mineral physics) to the pressure range and scale of the Earth's mantle. The various Gruneisen parameters of classical lattice-dynamic theory and Rayleigh-Nusselt number scalings are examples of such relations. The important parameters in mantle convection are density and expansivity; however, because seismology mainly measures seismic velocities, going from a tomographic model to a convective model requires scaling relations. Such relations, and the bridges they provide from mineral physics and petrology to seismology and geodynamics, are developed below. No such scaling relations exist between isotopic properties and seismology, or between the colors on a tomographic cross-section and the direction of mantle flow, such as are assumed in the standard models–two-layer or whole-mantle.

5.1 Scaling relations in seismology

The relative behaviors of density, shear velocity, and bulk sound speed in the mantle are now being determined and the results have implications for chemical and thermal structure (e.g. Masters et al., 2000; Trampert et al., 2001, 2004). Reflection and scattering coefficients also contain information about density contrasts. Prior to these developments it was common in seismology to attempt to infer density from the shear velocity and sometimes to assume or infer a relationship between shear velocity and compressional velocity, and shear velocity and temperature.

Although there are thermodynamic relationships between compressibility and volume there is little theoretical or empirical support for the idea that the shear velocity should be the control, or scaling, variable, particularly at high-pressure or high-temperature. Recent work has shown that the very large low-shear velocity features in the lower mantle do not have low density or low bulk modulus (Ishii and Tromp, 2004, Trampert et al., 2004). These authors were able resolve density as well as seismic velocities from seismic data and to separate the effects of temperature and composition.

5,2 Scaling relations in geodynamics

The scaling parameters in fluid-dynamics involve dimensionless numbers A measure of the vigor of convection and the distance from static equilibrium is the Rayleigh number, Ra. This is also the scaling parameter for the effect of size on the system. Estimates of Rayleigh numbers for the mantle often do not take into account the effect of layering and pressure on physical properties. In a spherical shell, convection occurs spontaneously when Ra exceeds about 2000 [Chandrasekhar, 1961]. Chaotic convection and efficient mixing is thought to require Ra of $>10^7$. Can realistic mantle models yield such values?

Mantle convection models with Ra of 10^7 or higher (Bunge et al., 2001; Parmentier et al., 1994), are assuming that the zero pressure values of thermal and transport properties maintain throughout the mantle. If one instead uses values estimated for the base of the mantle [Tackley, 1998] one derives a value of only about 4000. The Rayleigh number depends on the thickness of the convection layer cubed. Therefore, a chemically–or convectively–layered mantle will have a lower effective Rayleigh number than will a chemically homogeneous mantle with whole-mantle convection. If the lower 1000 km of the mantle is convectively isolated, Ra drops to 500. There is thus the possibility that the deep mantle convects sluggishly, episodically, or not at all. Clearly, in the case of the Earth's mantle, and in systems involving a pressure gradient in general, a single-system Rayleigh number is not an adequate description of convective style. In a self-organizing self-gravitating system it may not even be possible to assign the Rayleigh number in advance.

5.3 Volume as a scaling parameter

As far as physical properties are concerned, the main effects of pressure, temperature and phase changes are via volume changes. The thermal, elastic and rheological properties of solids depend on interatomic distances, or lattice volumetric strain, and are relatively indifferent as to what causes the strain (T, P, composition, crystal structure) [Anderson, 1989; Anderson, 1987; Birch, 1952, 1961]. This is the basis of the quasiharmonic approximation. In this approximation the effects of temperature are not ignored, but are assumed to affect the physical properties mainly through volume expansion. The intrinsic effects of temperature are treated as perturbations, as in the Debye theory of solids [Kieffer et al., 2002; Karki et al., 2001; Organov et al., 2001]. Intrinsic temperature effects are those that occur at constant volume. The quasiharmonic approximation is widely used in mineral physics but not in seismology or geodynamics where less physically sound relationships –they should not be called approximations since often the sign is wrong–are traditionally used.

A parameter that depends on P, T, phase (ϕ) and composition © can be expanded as;

$$M(P, T, \phi, ©) = M(V) + \varepsilon$$

where ε represents higher-order intrinsic effects at constant V. This is the basis of Birch's Law [Birch, 1961], the seismic equation of state [Anderson, 1987], laws of corresponding states, finite-strain equations, the Hildebrand equation of state and quasiharmonic approximations. Lattice dynamic parameters [Debye temperature and eigenfrequencies, ω], and thermodynamic and anharmonic parameters are interrelated via V. In some cases, the intrinsic effect of temperature is important, but the quasiharmonic approximation is a step away from the Boussinesq and related approximations that ignore the effects of volume change on most physical properties, or combine pressure and temperature effects in thermodynamically inconsistent ways.

5.3.1 Lattice theory

Anharmonicity causes atoms to take up new average positions of equilibrium, dependent on the amplitude of the vibrations and hence on temperature, but oscillations about the new positions of dynamic equilibrium remain nearly harmonic. At any given volume the harmonic approximation can be made so the characteristic temperature and frequency are not explicit functions of temperature. This is the essence of the quasiharmonic approximation (see, e.g., Anderson, 1987a, O. L. Anderson, 1995; Stacey, 1993).

In an harmonic solid there is no thermal expansion, the elastic constants are independent of temperature and pressure, and the free energy is independent of volume. In a real crystal, and in the quasiharmonic approximation, there is a close relationship between lattice thermal conductivity, thermal expansion and other properties that depend intrinsically on anharmonicity of the interatomic potential or the interatomic distances. In the Boussinesq approximation it is assumed that volume depends on temperature alone. The coefficient of thermal expansion and the lattice conductivity, which in the complete theory depend on volume, are assumed to be constants. In some interpretations of tomographic images it is assumed that temperature is the only variable. Although the images are in false color, red and blue are usually taken as proxies for hot and cold, or plumes and slabs.

The temperature effect on volume can be written in terms of the coefficient of thermal expansion, where constant pressure is implied;

$(d\, ln\, V / \alpha\, dT)$

All other derivatives, e.g.

$d \ln M / d \ln V$

are set to zero in the Boussinesq approximation.

5.4 Beyond Boussinesq

The effect of volume changes on thermodynamic properties are determined by dimensionless Gruneisen parameters (e.g. Stacey, 1992; Anderson, 1989a). Scaling parameters for volume-dependent properties [Anderson, 1987,1989a] can be written as power laws, $M \sim V^n$ or as logarithmic volume derivatives ($d\, ln\, M/d\, ln\, V$) about the reference state;

Lattice thermal conductivity	$- d\, ln\, K_L / d\, ln\, V \sim 4$
Bulk modulus	$- d\, ln\, K_T / d\, ln\, V \sim 4$
Thermal expansivity	$- d\, ln\, \alpha / d\, ln\, V \sim -3$
Viscosity	$- d\, ln\, \nu / d\, ln\, V \sim 40- 48$

Gruneisen constant $-d \ln \omega / d \ln V \sim 1$

These non-linear relations show how parameters might vary with temperature and pressure in a self-consistent way, rather than independently. Where T and P variations are replaced by *ln* V variations, these called Dimensionless Logarithmic Anharmonic (DLA) parameters [Anderson, 1987a]. The scaling relations are semi-empirical and the numerical values are estimated from experiment for the thermal properties and by experiment and geodynamic measurements for the viscosity.

In the upper mantle, T and phase changes (ϕ) mainly control V variations [e.g. TOE] while in the deep mantle it is P and ©. T is particularly important in the upper mantle because the coefficient of thermal expansion is large and increases with temperature and therefore with depth, in the upper part of the mantle. Other derivatives scale with the thermal expansion coefficient $\alpha = (d\, ln\, V/ dT)_{p.}$

5.4.1 Things that do not scale with volume

Shear velocity, rigidity, viscosity, radiative conductivity, and seismic attenuation have intrinsic temperature, compositional or structural dependencies in addition to volume dependent terms, if any. They are not simple functions of volume. Shear velocity has an anelastic term that depends on frequency and microstructure of the solid. Rigidity has a strong intrinsic temperature term (e.g. Anderson, 1987a). Rigidity is affected by iron substitution while the bulk modulus is insensitive to iron (TOE, Table 6.2). The DLA parameters for bulk modulus fall into a narrow range for most minerals while those for rigidity do not (TOE, Table 6.7). Nevertheless, the rigidity is an important parameter and may help resolve the ambiguity of mantle compositional and temperature models (e.g. Mattern et al., 2004).

Intrinsic, anelastic, composition and crystal structure effects affect the shear moduli. Shear velocity, or rigidity, is not a good proxy for density, temperature, composition or for other elastic constants such as bulk modulus. Even the 'compressional' velocity is affected by these considerations since it is more affected by rigidity changes than by bulk modulus, or compressibility, changes. The good correlation between shear velocity and compressional velocity changes in the upper mantle is not necessarily because the moduli are good quasiharmonic parameters but because anelastic effects are large, and partial melting strongly affects the rigidity. Chemical changes such as iron substitution or changes in mean atomic weight can have a large direct effect on density (and thus seismic velocity) and rigidity, without a corresponding effect on the bulk moduli. Partial melting can lower the rigidity with little effect on density; partially molten dense eclogitic sinkers can be 'red' even if they are colder than ambient mantle.

5.4.2 Viscosity

Viscosity is one of the most important–but most uncertain and most variable–parameters in mantle dynamics and evolution [see Appendix 2]. The buoyancy parameter $\alpha\delta T$, however, is more important in discussions of chemical stratification. The total variation of viscosity across the mantle may be 5 orders of magnitude Viscosity decreases strongly over the depth intervals 1000-1400 km and at 2000-2500 km (e.g. Forte and Mitrovica, 2001). The former interval also has anomalous thermal and FeO-gradients (Mattern et al., 2004) and seems to be a fundamental boundary, even barrier, in the mantle (e.g. Wen and Anderson, 1997).

In the outer parts of the Earth viscosity is strongly affected by water content, as well as temperature. The rapid drop in viscosity from the lithosphere to the asthenosphere is therefore not entirely a result of temperature . The low temperature rise across the bottom boundary layer of the mantle that results from internal heating and other factors (e.g. Lenardic and Kaula, 1994; Tackley, 1998) suggests a smaller viscosity drop across the bottom TBL. In general, the viscosity profiles support the idea of a tri-partite mantle, although there may be more than three chemically distinct regions of the mantle. Thin layers such as the crust, the olivine-rich buoyant perisphere and D" are not included.

6. APPLICATIONS

The specific volume at the base of the mantle is 64% of that at the top [Anderson, 1989; Dziewonski and Anderson, 1981]. Compression, composition and phase changes, and to some extent, temperature, are all involved. In classical lattice dynamics, and in the quasiharmonic and seismic-equation-of-state approximations, it does not matter, to first order, what causes the volume, or interatomic spacing to change, for such parameters such as bulk modulus, lattice conductivity, specific heat and expansivity. Activated processes e.g., diffusion and viscosity, and radiative conductivity, however, depend on temperature explicitly. The scaling theory reviewed above will be applied to a few situations relevant to the deep mantle; the Rayleigh number, the thickness and growth time of a deep thermal boundary layer, the 'detectability' of chemical interfaces, and the possibility of irreversible chemical layering.

6.1 Rayleigh numbers

The thermal boundary layer (TBL) thickness of a fluid cooled from above or heated from below grows as;

$$h \sim (\kappa\, t)^{1/2}$$

where κ is thermal diffusivity, $K_L/\rho c_p$ and t is time. The TBL becomes unstable, and detaches when the *local or sublayer Rayleigh* number

$$Ra_c = \alpha \, \mathbf{g} \, (\delta T) \, h^3/\kappa\nu$$

exceeds about 1000 [Howard, 1966; Elder, 1976] (g is acceleration due to gravity; ν is the kinematic viscosity = dynamic viscosity/ρ and δT is the temperature increase across the TBL). [the units are α=1/K, g=m/s^2, h=m, T=K, κ=m^2/s and ν=m^2/s giving m K m^3 s s/ K s^2 m^2 m^2 so that Ra_c is dimensionless]. Ra_c is related to the Nusselt number (Nu) and the system Rayleigh number by;

$$Ra_c = Ra/(Nu)^3$$

The combination $\alpha/\kappa\nu$ decreases with V, thereby lowering Ra_c at high P or low T. Viscosity alone may increase by a factor of 60 to 80 across the mantle due to compression plus phase changes alone (but see Mitrovica and Forte, 1997). Compressible flow, temperature-dependent viscosity and internal heating calculations give a δT across a deep TBL that is less than the surface δT [Tackley, 1998; Lenardic and Kaula, 1994], and much less if there are intervening boundary layers. This plus the upper-mantle water-weakening effect means that the viscosity at the base of the mantle is probably higher than in the asthenosphere, and the drop in viscosity across the TBL is less. Mantle flow driven by cooling of the top boundary layer and internal heating makes it difficult for cavity plumes to form in the bottom boundary layer (Nataf, 1991; Lenardic and Kaula, 1994).

The local Rayleigh number is based on the thickness of the thermal boundary layer. At the surface of the Earth the issue is complicated because of water (Hirth and Kohlstedt, 1996) and because the crust and refractory peridotite part of the outer shell are not formed by conductive cooling (they are intrinsically buoyant) and because the viscosity, conductivity and α at low P are strongly T-dependent. For parameters appropriate for the top of the mantle, treated as a constant viscosity fluid, the surface TBL becomes unstable at h of about 100 km [Elder, 1976]. The time-scale is about 10^8 years, approximately the lifetime of surface oceanic plates. The top boundary is very viscous, stiff and partly buoyant, and the instability (called subduction or delamination) is controlled, in part, by faulting and not viscous deformation (see Lenardic and Kaula, 1994); a viscous instability calculation is not entirely appropriate. For the bottom boundary layer the deformation is more likely to be purely viscous but there may also be intrinsic stabilizing density effects.

6.2 Tomography

The implication of the volume scaling plus the role of anelasticity is that temperature effects are much less important in controlling density and seismic velocities in the lower mantle than at the surface. At high pressure, temperature has little effect on density and other properties that depend on density. Seismic attenuation is low at mid-mantle depths and the anelastic contributions are expected to be low (TOE, Chapter 14). Elastic moduli and seismic velocities depend on density, or specific volume but because of low α at large compression [Birch, 1961; Chopelas and Boehler, 1992], temperature has little effect on seismic velocity in the lower mantle. Chemical and phase changes, anisotropy, and fluid phases should dominate seismic velocity variations.

The changes that influence measured and inferred properties near the surface of the Earth are anelastic, mineralogical, chemical and thermal (e.g. Goes and Govers, 2000; Lee, 2003; Perry et al. 2003) plus partial melting. The high near-surface temperature gradient creates a low-velocity, low-viscosity, high-attenuation and low-thermal-conductivity zone, usually equated with the asthenosphere, at depths of ~ 100 to 200 km. Anelastic dispersion and mineralogy dominate the variation of seismic velocity (e.g. Goes et al., 2000; Lee, 2003). At greater depths the temperature gradient and the effect of temperature on volume (V) and elastic properties becomes smaller but pressure continues to increase. At very high pressures, variations in density are mainly controlled by composition and mineralogy, which are the main controls on lattice volume when the coefficient of thermal expansion is low (e.g. Ishii and Tromp, 2004; Trampert et al. 2001). Composition, silicate and metallic melts and phase changes (including the low-spin transition in FeO and the post-perovskite phase change) become the important controls on V, buoyancy and seismic parameters. Activated and quantum effects have additional (intrinsic) temperature dependencies.

Purely thermal upwellings are expected to have low bulk modulus, low compressional velocity and low density. However, this does not seem to be the case (Ishii and Tromp, 2004, Trampert et al., 2004) for the large lower mantle features. There is a trade-off between temperature, composition and mineralogy, which these studies have attempted to resolve. The large-scale features have the appropriate dimensions to be thermal in nature but resemble more a chemically dense layer at high pressure, i.e., large-scale marginally stable domes with large relief on the boundaries. D'' would then be a very dense, probably iron-rich– layer, and the overlying 'layer', which might be called D', would be a less dense region trapped between D" and the rest of the lower mantle. Such density stratification may have been established during accretional melting of the Earth [Anderson, 1989a; Agee, 1990] by downward drainage of dense melts and residual refractory phases, and iron partitioning into phase that may include post-perovskite, low-spin iron-rich oxides and sufides and intermetallic compounds. The large low-shear-velocity features are more appropriately called 'domes', a geologically descriptive term, than 'megaplumes', which implies a thermal active upwelling with low-density and low bulk modulus.

6.3 Thermal boundary layers

An homogeneous fluid with constant properties, heated from below, will develop symmetric upper and lower thermal boundary layers. Downwellings and upwellings from the boundaries will have the same dimensions and time constants. However, the upper and lower thirds of the mantle have quite different spectral and spatial characteristics and correlations with plate reconstructions and this has led to the concept of a tri-partite mantle [Anderson, 2002a]. Whether these distinctively different tomographic regions differ in intrinsic chemistry and whether they can exchange material are matters of current debate. The prominent tomographic anomalies in the lower third of the mantle–the abyss– have very large dimensions [e.g. Gu et al., 2001], much larger than upper mantle slabs, consistent with the scaling (small ultra-low velocity zones at the base of the mantle [Garnero and Helmberger, 1995] are here interpreted as regions containing core or mantle fluids or reaction products).

When the whole lower TBL goes unstable we have a 'diapiric plume'. When a thin low-viscosity layer near the core feeds a plume head, we have a cavity plume; the physics is similar to a hot air balloon. Temperature dependence of viscosity is essential for the formation of cavity plumes with large bulbous plume heads and narrow plume tails. Temperature dependence and internal heating, however, reduce the temperature drop across the lower thermal boundary layer and the viscosity contrasts essential for this kind of plume (Nataf, 1991). It has been thought that pressure effects might permit the reestablishment of conditions necessary for cavity plume formation, but this does not seem to be the case (Tackley, 1998). Pressure broadens considerably the dimensions of diapir plumes and cavity plume heads. It is this 'pressure-broadening' that makes intuitive concepts about plumes (e.g DePaolo and Manga, 2003), based on unscaled laboratory simulations, implausible for the mantle (Anderson, 2004).

6.3.1 The lower thermal boundary layer

The presence of a lower TBL and the need to power the dynamo do not require that plumes exist or that they have the properties assigned to plumes in the current literature. The key questions are whether the dimensions and timescales of deep upwellings are of the order of hundreds of km and tens of millions of years, whether they rise to the surface, and whether they can give rise to the sorts of melting anomalies seen at the surface. The neglect of pressure and scaling effects is responsible for the widely held misconception that narrow rapidly upwelling cylinders are required by boundary layer and dynamo theory (e.g. DePaolo and Manga. 2003). This misconception is maintained by unscaled laboratory injection experiments and Boussinesq computer simulations (e.g. Cordery et al., 1997). The core can get rid of its heat by mechanisms other than ~200 km wide plumes extending to the surface.

The critical dimension of lower-mantle thermal instabilities, ignoring radiative transfer, is predicted from the above considerations to be about 10 times larger than at the surface, or about 1000 km. This is consistent with seismic tomography [Hager et al., 1985; Hager and Clayton, 1989; Tanimoto, 1990; Gu et al., 2001] and with compressible flow calculations with depth dependent properties [Tackley, 1998]. If there is an appreciable radiative component to the conductivity, or a chemical component to the density increase (i.e. chemical stratification), then the scale-lengths can be much greater.

The timescale of deep thermal instabilities scaled from the upper mantle value is $\sim 3 \times 10^9$ years. Radiative transfer and other effects [Chopelas and Boehler, 1992; Hofmeister, 1999] may increase thermal diffusivity, further increasing timescales. The surface TBL cools rapidly and becomes unstable quickly. The same theory, scaled for the density increase across the mantle, predicts large and long-lived features above the core. This, and the plausibility of chemical stratification, are the most dramatic effects of pressure and volume scaling. One further prediction is that the long wavelength geoid is very stable over time [e.g. TOE, Chapter 12].

The lower 1000 km of the mantle has unique spectral and spatial tomographic characteristics [Gu et al., 2001; Lay, 2005; Lay et al., 1998; Garnero, 2000; Trampert et al., 2004; Ishii and Tromp, 2004], viscosity (Forte and Mirovica, 2001) and thermal gradient (de Silva et al., 2004). This part of the mantle is probably stabilized against convective overturn by effects of pressure and composition [Tackley, 1998] and, possibly, by high thermal conductivity and by the low-spin transition in FeO [Gaffney and Anderson, 1973; Badro et. al., 2003]. The idea that the upper third and the bottom third of the mantle might be chemically distinct has been viewed with skepticism (e.g. van der Hilst, 2004); many still favor whole mantle convection, deep slab penetration and a homogeneous mantle because of visual impressions of some tomographic cross-sections and equation of state modeling of the lower mantle.

If the abyss represents about one-third of mantle thickness it will have an Ra reduction due to this effect alone of a factor of 30 relative to a reference state of whole mantle convection. A similar reduction is accomplished by viscosity increase alone. Together, these decrease the deep mantle Ra by about 10^3 compared to whole-mantle, constant-property, values. There may also be other chemical boundaries in the mantle [Anderson, 1979, 2002a] that would further reduce Ra and the δT across TBLs.

The predicted large-scale, longevity and sluggishness of lower mantle features are not entirely due to high viscosity. Low α at high P means that intrinsically dense layers may be permanently trapped; moderate jumps (~ 1%, depending on δT) in intrinsic density between layers in the mantle can stabilize chemical layering [Tackley, 1998; Anderson, 2002a]. Unreasonably high mantle temperatures do not occur in these trapped layers if most of the radioactivity is in the crust and upper mantle [Anderson, 1989; Anderson, 2002a]. Heat can also be conducted across the layer more efficiently at high P

The removal of material from deep layers by entrainment is controlled by ratios of density, viscosity, layer thicknesses, radioactivities and Rayleigh numbers, all of which are controlled, to some extent, by the convective process itself. This emphasizes the need for self-regulation rather than externally imposed boundary and material conditions. Acccessibility of deep layers, or zones, is an essential ingredient of the standard models and recent modifications to them.

6.3.2 Relation to plume heads

Paradoxes appear when physical properties are assumed to depend only on temperature, or when only one or two of the volume dependent terms are varied. If viscosity depends mainly on temperature, a thin, low-viscosity velocity boundary layer can form at the base of the thermal boundary layer and this layer is what feeds cavity plumes (Stacey and Loper, 1983). Nataf (1991) noted that basally heated convection in strongly temperature dependent media (the condition for the existence of cavity plumes) always leads to a much larger temperature drop across the cold upper boundary layer than across the hot lower boundary layer, and for these conditions, cavity plumes are unlikely (see also Lenardic and Kaula, 1994). Internal heating reduces the temperature and viscosity jump still further. Pressure dependence might restore the equivalence to the temperature drop across the upper and lower boundaries, but this does not seem to be the case (Tackley, 1998). A solution to this paradox was investigated by Lenardic and Kaula (1994). The solution itself is paradoxical; surface plates are made more deformable so they become unstable before they get too thick; they fall rapidly to the lower boundary where they create a subadiabatic temperature gradient above the superadiabatic gradient in the boundary layer, creating a large temperature and viscosity contrast. In other words, the condition for a cavity plume is created by destabilizing the upper layer, overcoming the temperature dependent viscosity, and thinning and cooling the lower boundary layer. Plume formation is triggered from above rather than from an instability in the boundary layer itself as in normal plume theory. In effect, plumes are ‘splashed out’ by impacting cold slabs. The net effect is that part of the effects of temperature dependence are negated and the systen resembles more the constant-property case. In the case of diapiric plumes, the effects of temperature are reversed by pressure, and only broad domes or diapirs form. Diapiric plumes, where the entire lower thermal boundary layer becomes unstable, have been considered unlikely on other grounds (e.g. Loper, 1985; Lenardic and Kaula, 1994).

The scales of lower mantle thermal diapiric instabilities are much larger than those often quoted for plume heads. This is because the volumes of plume heads are often assumed to be related to the sizes of large igneous provinces or the thickness of D”. Some published plume head dimensions are not the result of a convection calculation, or use the Boussinesq approximation, or are simply arbitrary in size [e.g. Cordery et al., 1997]. However, a plume head must obtain enough buoyancy to overcome lower mantle viscous resistance and must be large.

If the thickness of D’’ is controlled by compositional layering or phase changes (Lay et al., 1998) then one must look elsewhere for the scale of lower mantle thermal instabilities. da Silva et al. (2004) found that the thermal gradient in much of the lower mantle is superadiabatic. If this is interpreted as evidence for a thermal boundary layer then it is more than 1000 km thick, consistent with the scaling relations, and it must have taken a long time to form. Recall that heat is supplied to the CMB at about $1/10^{th}$ the rate at which heat is removed from the surface, so deep TBL instabilities are slow to form.

6.4 Radiative conductivity

If there is one parameter that could radically change current views of mantle convection it is radiative thermal conductivity. We do not yet know if it is important in the deep mantle. Radiative conductivity (K_R), a quantum effect, increases as T^3 and may increase considerably the conductivity of the deep mantle [Lubinova, 1958; Hofmeister, 1999]. This depends on grain size, purity, and iron partitioning.

K_R can significantly affect the thermal history of the mantle and the style of mantle convection [Dubuffet et al., 2002]. This is important since P and T now combine to increase non-convective heat transfer processes and to suppress the vigor of mantle convection and mixing. K_R at high T, and lattice conductivity, K_L, at large compression exert a stabilizing influence on the deep mantle [Tackley, 1998; Dubuffet, Yuen, and Rainey, 2002; Chopelas and Boehler, 1992]. These considerations may even make Ra subcritical.

6.5 STRATIFIED MANTLE?

According to Helffrich and Wood (2001) “there is no major change of chemical composition between the upper and lower mantles…studies are consistent with deep subduction and hence with whole-mantle convection… stratification seems, therefore, to be an increasingly difficult position to defend”. Can these assertions be defended?

The conflict, confrontation and crisis that permeate the isotopic literature may be a chimera. The whole-mantle-convection / deep-slab-penetration interpretation of tomographic cross sections is not unique or robust (Cizkova et al., 1999, Davaille, 1999; Dziewonski, 2004, 2005; Boschi and Dziewonski, 1999; Cizkova and Matyska, 2004; Hamilton, 2002) and it is inconsistent with geophysical evidence more broadly defined. Isotopic data do not require a layered mantle or a chemical boundary at 650 km. Petrology-based models tend to be gravitationally stratified, with buoyant olivine-rich layers at the top, dense perovskite- and iron-rich layers at depth, and intermediate eclogite or komatiitic regions (e.g. Agee, 1990; Anderson, 1983). In these models the mid-mantle, upper mantle and crust are complementary to the lower mantle and to each other, in composition. Abrupt seismic discontinuities are not automatically assumed to be reservoir or isotope boundaries. What distinguishes this class of model from the others is gravitational stratification by density and upward concentration of volatile and LILE, including U, Th and K. Another distinction is that the deeper layers are not accessible to surface volcanoes. Plate tectonics and geochemical cycles appear to be entirely restricted to the upper ~1000 km. These layered models do not have the paradoxes associated with the standard model or with qualitative tomographic interpretations, and they are consistent with the effects of volume changes discussed in this paper. The prediction of inaccessible regions is consistent with various mass imbalance calculations that are paradoxical in the standard models. It may be relevant that even the crust and the continental lithosphere have managed to stratify themselves by their intrinsic density (Rudnick and Fountain, 1995; Lee et al., 2003).

The seismic velocities of plausible materials in the upper mantle, the mesosphere [1000 to 2000 km depth] and the deep mantle are predicted to differ little from one another, even if the density contrasts are adequate to permanently stabilize the layering against convective overturn [Anderson, 2002a]. Since chemical discontinuities are almost invisible to seismology, compared to phase changes, and since even small chemical density contrasts can stratify the mantle, the possibility must be kept in mind that there may be multiple chemical layers in the mantle. The major seismic discontinuities in the mantle are due to mineralogical and phase changes, not chemical changes but this does not rule out chemical layering.

Radial gradients in seismic velocity were initially used to divide the mantle into upper mantle, transition zone and lower mantle (Bullen, 1947). The original, or classical, location of the boundary between the upper and lower mantles was at 1000 km, also known as the Repetti discontinuity [Bullen, 1947; Birch, 1952]. The transition zone (TZ), Bullen's Region C, extends from 400 to 1000 km. The 650-km discontinuity is thus in the transition region, as defined by Bullen and above the lower mantle proper. Spectral, spatial, scattering, surface wave, decorrelation, and amplitude characteristics of mantle heterogeneity have been used to further divide the mantle into major radial zones, some of which may be separated by chemical discontinuities [Bullen, 1947; Wen and Anderson, 1995, 1997; Gu et al., 2001; Hamilton, 2002]. These lithologic zones include a shallow depleted buoyant layer (continental lithosphere, perisphere), asthenosphere, upper mantle and TZ (in the original Bullen sense, Regions B and C), the mesosphere (1000 - 2000 km), the abyss (2000 km to the CMB), and D". None of these regions need be distinctive isotope reservoirs assumed by geochemists (e.g. TOE). If they are not, most of the geochemical paradoxes are eliminated. Albarede and van der Hilst (2002a.b) quoted the lack of evidence for global seismic interfaces anywhere between 650 km and D" as evidence against chemical stratification. But the evidence for chemical stratification is more subtle than they imply. There is, however, little evidence for structure in the mesosphere (e.g.Vidal et al.,2001; Gu et al.,2001) except at the variable depth boundary regions.

The tomographic structure of the mantle above the Repetti discontinuity–the upper mantle proper–correlates with present and past plate tectonics [Wen and Anderson, 1995; Becker and Boschi, 2002] and behaves as expected for a moderate-Rayleigh-number fluid cooled from above and driven by the plates and lithospheric architecture.

6.5.1 The Repetti discontinuity

A layered convection model with a chemical interface near 900 km at the base of Bullen's TZ explains both the geoid and the dynamic topography [Wen and Anderson, 1997] although other interpretations are possible (Richards and Hager, 1984; King and Masters, 1992). The evidence for stratification includes the mismatch between tomographic patterns and spectra between various depth regions (e.g. Gu et al. 2001; Trampert et al., 2004; Anderson, 2002a; Becker and Boschi, 2002; Ray and Anderson, 1994; Scrivner and Anderson, 1992; Wen and Anderson, 1995; Tanimoto, 1990) and evidence for slab flattening (Fukao et al., 1992, 2001). The degree 2 patterns of shallow- and deep-mantle seismic tomographic structures change across a depth of about 800 km (Tanimoto, 1990). Kawakatsu and Niu (1994) found a 920-km discontinuity near subduction zones and Shen et al. (2003) found a discontinuity near 1000 km depth beneath Iceland and Hawaii. Ritzwoller and Lavely (1995) showed that there is a significant decorrelation of radial mantle structures at a depth of about 1000 km. Wen and Anderson (1995) found good correlations between subducted slabs and seismic tomography in the 900-1100 depth range. The mantle does not become radially homogeneous and adiabatic until about 800 km depth (Dziewonski and Anderson, 1981) or deeper (Mattern et al., 2004). Mantle viscosity decreases strongly over the depth interval 1000-1400 km (Forte and Mitrovica, 2001). This interval also has anomalous thermal and FeO-gradients (Mattern et al., 2004). Anderson (1970) and Whitcomb and Anderson (1970) summarized the early data for reflectors near 900 km depth–and other depths–and Revenaugh and Jordan (1991) summarized later data. A variety of evidence therefore suggests that there might be an important geodynamic boundary, possibly a barrier to convection, and a thermal boundary at a depth of about 900-1000 km.

Chemical boundaries, in contrast to most phase-change boundaries, will not be flat, as assumed in some layered convection models, and will have little impedence contrast. The latter inference is based on plausible compositional differences between various lower-mantle assemblages. Complications between 650 and 1300 km depth (e.g. Mattern et al., 2004; Wen and Anderson, 1997) are perhaps related to slab trapping or thermal coupling (e.g. Cizkova and Matyska, 2004) and undulations in the Repetti discontinuity.

The lowermost 1000 km is equally rich in seismological detail and differs from the mesosphere in all respects (e.g. Gu et al. 2001; Trampert et al., 2004; Anderson, 2002a; Ray and Anderson,1994; Lay et al., 1998a; Garnero, 2000; Garnero and Helmberger, 1995).

7. SUMMARY

The petrological and mineral physics case for an inhomogeneous mantle (Birch, 1952; Anderson, 1983) and some sort of convective or chemical stratification is strong (e.g. Javoy, 1995; Agee, 1990; Agee and Walker, 1993; Duffy and Anderson, 1989; Anderson, 2002; Gasparik, 1997; Wen and Anderson, 1997; Meibom and Anderson, 2003; Lee et al., 2004). The seismological evidence is equally strong. A large amount of geophysical data and quantitative analysis by many workers from diverse disciplines is integrated into this synthesis. In contrast, the whole mantle convection view is based mainly on travel time tomography and visual impressions from color cross sections, and absence of obvious evidence for discontinuities and TBLs.

Quantitative interpretation of tomographic models suggest that compositional heterogeneity is particularly strong above 1000 km and below 2000 km depth (Gu et al., 2001), leading to the suggestion of a tri-partite mantle (Anderson, 2002). High seismic velocities are often attributed to low temperatures and sinking slabs but such intuitive scaling is probably not warranted. Quantitative constraints on mantle layering and composition combined with new mineral physics results reopen the contentious issues of iron and silicon enrichment–relative to pyrolite–in the deep mantle (e.g.TOE).

With regard to issues involved in the debate about whole mantle convection, locations of geochemical reservoirs and the style of mantle convection;

1. The velocity discontinuity at 650 km is primarily due to a isochemical phase change; the chemical change at this depth, if any, appears to be small (Anderson, 1967; Duffy and Anderson, 1989; Ito & Takahashi, 1989).

2. Chemical discontinuities in the mantle are hard to detect if they are not associated with changes in mineralogy. Chemical interfaces will have high-relief and be difficult to map by standard seismological techniques. An endothermic phase change acting in concert with chemical and viscosity changes near 1000 km may be very effective in stratifying the upper parts of the mantle.

3. Purely thermal and thermo-chemical instabilities in the deep mantle are predicted to be immense. This is a simple consequence of the effect of pressure on volume and the effect of volume on thermal and transport properties. Furthermore, scaling relations, and estimates of heating rates, suggest that lower-mantle thermal features are sluggish, slow to form, and long-lived. Steady-state convection calculations may be inappropriate if the response time of the mantle is long. It is more plausible than generally thought that the mantle is chemically stratified, but not with a radioactive-rich lower layers or zones, or a well stirred upper layer.

The kind of chemical stratification that seems to be most consistent with all geochemical and geophysical evidence is essentially the inverse of the standard model and recent modifications of it (e.g. Albarede and van der Hilst, 2004). It involves a refractory barren inaccessible lower mantle (not primordial, undegassed or highly radioactive), with irregular chemical boundaries near 1000 and 2000 km, and upper-mantle circulation closed between average depths of 650 and 1000 km [Anderson, 2002; Anderson, 1989; Hamilton, 2002]. The lower mantle (below 1000 km) is probably depleted in U, Th and K. The upper mantle is heterogeneous, both radially and laterally, (e..g. Meibom and Anderson, 2003) and in both isotope geochemistry and fertility

(Foulger et al., 2005). This kind of layered model removes the objections that have been raised against layered models (e.g. Davies, 1988; Helffrick and Wood, 2001; Hager et al., 1985; Coltice and Ricard, 1999; Schubert et al., 2001; Wen and Anderson, 1997). [see Appendix 1].

Most of the mantle is both depleted in LILE and barren, but only parts of the upper mantle are depleted or fertile. The deeper layers of the mantle may be inaccessible. The upper mantle is probably heterogeneous but magmas are homogenized at ridges where large volumes of mantle are processed continuously for long periods of time (Meibom and Anderson, 2003). This contrasts with the situation away from mature ridges.

This view of mantle structure and evolution contrasts with current competing models, but does not involve the conflicts and paradoxes that are intrinsic in those models. These are mainly the result of assumptions, not data. The conclusions concerning permanent stratification are tentative since they require knowledge of temperature variations in the deep mantle, and this requires a self-consistent convection calculation that allows for self-gravity, depletion of radioactivity in the deep layers, multiple layers, and self-organized motions and boundary deformations instead of imposed ones. Convection calculations that have been used as evidence against layered convection have assumed uniform radioactivity, used the Boussinesq or extended Boussinesq approximations, enforced flat boundaries or shear coupling, and have not explored a range in possible boundary depths. On the other hand, models that attempt to model layered convection, or convective stirring, often use unreasonable Rayleigh numbers or distributions of radioactive elements. Moving beyond this state of affairs is a challenge to convection modelers.

With regard to the assumptions underlying the standard models; one can reverse all these assumptions (e.g. well-stirred homogenous upper mantle; primordial, undegassed or less degassed lower mantle; high U and ^{3}He lower mantle and so on) and remove most of the paradoxes and conflicts between 'geophysical data' (e.g. van der Hilst et al., 1997) and 'geochemical data' (e.g. DePaolo and Wasserburg, 1976). This is one sign of a non-robust model.

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APPENDIX 1

THE GEOPHYSICAL DATA

The evidence against the standard model is not evidence against chemical stratification in general. However, because of the perceived crisis, some investigators have argued for a return to one-layer mantle models, ignoring a large body of other geophysical evidence. Geophysical data that have been cited in support of whole mantle convection include;

1. The long-wavelength tomographic structure of the lower mantle and whole-mantle convection models with plausible velocity-density scalings and viscosity models, successfully explain the geoid (Hager et al., 1985; Hager and Clayton, 1989)..

2. The bathymetry of the seafloor is explained by conductive cooling of the plate and whole mantle convection (Davies, 1988).

3. Selected tomographic cross-sections show a few high-velocity features in the mantle below 650 km. Intuitive scaling relations suggest that these may be cold, dense slabs (Grand et al., 1997; van der Hilst et al., 1997).

4. Scattering of high-frequency seismic energy is thought to be consistent with slab fragments in the lower mantle and whole mantle convection (Helffrich and Wood, 2000).

5. The imbalance between heat flow and heat production is thought to require a deep mantle rich in U, Th, and K.

6. The seismic properties of the lower mantle are consistent with pyrolite, the prototype upper-mantle rock.

Several of the interpretations, including the assumption that deep slabs are efficient scatterers, rely on scaling relations and assumed mineral properties at high pressure. Problems with these interpretations are;

1. The geoid represents the combined effects of density variations in the interior of the mantle and the accompanying distortion of the boundaries (termed "dynamic topography") including the surface, the core-mantle boundary and any internal interfaces (e.g. Hager et al., 1985). Therefore, both geoid and dynamic topography must be explained by the same model. Layered convection with a chemical boundary near 900 km, can explain both datasets (Wen and Anderson, 1997).

2. The inability of some layered models to explain bathymetry is due to the assumption that most surface heat flow is from radioactivity in the lower mantle. A chemically stratified mantle with most of the radioactive elements in the crust and upper mantle (TOE) does not suffer from this problem, or a problem with lower mantle overheating. On the other hand, the large dynamic topography associated with whole-mantle convection affects the square-root age bathymetry relation. In the model of Wen and Anderson (1997), dynamic topography is generated by density variations in the upper mantle and is of low amplitude. Ocean-floor bathymetry is dominated by cooling of the plate. Anomalous bathymetry is primarily due to shallow variations in density, not necessarily high temperature.

3. If the mantle is layered, tomography and the geoid can rule out shear coupling between layers, but thermal and topographic coupling, or random correlations, cannot be ruled out. The nature of the coupling depends on viscosity and density contrasts, layer thicknesses and Rayleigh numbers. In layered convection simulations thermal coupling induces structures that visually resemble downwellings that penetrate the interface, but are not (e.g. Cizkova, 1999; Cizkova and Matyska, 2004). Quantatative analysis of tomographic models and the history of plate subduction confirm the importance of a barrier near 900 km (Wen and Anderson, 1995). Most tomographic cross-sections show slab flattening near 650 km with a few deeper, diffuse features extending to 1000-1300 km (Fukao et al., 1992, 2001). Even for a chemically and irreversibly stratified mantle there are regions in deeper layers with higher seismic velocities than average, and there will be downwellings from internal boundary layers.

4. More recent seismic scattering studies have been able to isolate better the source of the

scattering. Most of the scattered energy comes from the upper mantle (Shearer and Earle, 2004; Baig and Dahlen, 2004) and can be attributed to slab fragments (e.g Meibom and Anderson, 2003). Deep-mantle chemical discontinuities are elusive because acoustic impedance is predicted to be low and topographic relief is expected to be high. Although coherent reflections may be difficult to detect with standard seismological techniques they may appear as high-frequency scattered energy, an alternative to the lower mantle slab interpretation.

5. The imbalance between heat productivity and heat flow is a result of secular cooling and time lags associated with heat transport to the surface. Layered mantle models have larger time lags. Mass balance calculations are consistent with most of the radioactivity, and other large-ion lithophile elements, being in the crust and upper mantle (Rudnick and Fountain, 1995; TOE, Chapter 8).

6. The seismic properties of the lower mantle are consistent with rocks quite different from pyrolite (Lee et al., 2004; Mattern et al., 2004), including rocks that are similar in major element chemistry to meteorites, and to meteorite compositions with the crust and upper mantle removed.

A strong case for whole-mantle convection has not been made. In addition to the arguments presented above, the geochemical models can also be criticized on both geochemical and physical grounds [e.g. TOE]. Can a primordial undegassed mantle survive accretion? Is the standard model gravitationally stable? Can the upper mantle be vigorously stirred and homogeneized? Are all regions of the mantle accessible to surface volcanoes? What mass balance or isotope evidence requires material transport from the lower mantle? Might all the components of oceanic and continental basalts reside in the upper mantle (e.g. Meibom and Anderson, 2003)? These are also problems in whole mantle models. Some of these questions and the assumptions behind them involve mineral physics or self-consistent convection simulations.

APPENDIX 2

VISCOSITY

Viscosity ν can be written;

$$\nu = \nu_0 (V) \exp G^*(V)/RT$$

where $G^*=E^*+PV^*$, G^* is Gibbs free energy, R is the gas constant and E^* and V^* are the activation energy and activation volume, respectively. A typical value for G^*/RT is 30 which means that V^* decreases from 4.3 to 2.3 cm^3/mole through the lower mantle (Anderson, 1989). In contrast to the anharmonic properties the effect of temperature on viscosity at high pressure is high. Viscosity is predicted to increase with depth throughout most of the lower mantle. The pressure gradient is high and uniform and the temperature gradient is low, compared to thermal boundary layers. It may decrease with depth in thermal boundary layers, however, if the effects of composition, grain size and phase changes are not dominant. In the model of Forte and Mitrovica (2001) viscosity decreases strongly over the depth intervals 1000-1400 km and at 2000-2500 km. The former interval also has anomalous thermal and FeO-gradients (Mattern et al., 2004) and seems to be a fundamental boundary in the mantle (e.g. Wen and Anderson, 1997).

Theoretical estimates of viscosity in the mantle rely on estimates of the activation parameters or the melting point and these also can be estimated from lattice dynamics and seismic estimates of compression (Poirier, 2000; TOE, Chapters 7 and 14). Viscosity may depend on other parameters, such as stress, grain size, anisotropy, history and impurities but pressure still tends to increase it.

A lattice dynamic estimate of the melting temperature can be written;

$$d\ ln\ T_m / d\ ln\ V = 1 + (d\ ln\ K / d\ ln\ V)$$

where K is the bulk modulus and T_m is the melting point.

In the homologous temperature scaling viscosity scales as;

$$\exp(-gT_m/RT)$$

Since the mantle is closest to the melting point near the bottom of thermal boundary layers (e.g. the asthenosphere and D”) we expect the mid-mantle to have the highest viscosities except possibly at internal boundary layers.

In the outer parts of the Earth viscosity is strongly affected by water content, as well as temperature. The rapid drop in viscosity from the lithosphere to the asthenosphere is therefore not entirely a result of temperature rise. It is not obvious that a similar drop would occur across D”. This plus the lower temperature rise across the bottom boundary layer of the mantle that results from internal heating and other factors (e.g. Tackley, 1998) suggests, but does not require, a smaller viscosity drop across the bottom TBL.

If the seismic quality factor Q is due to the motions of dislocations the distribution of lengths and relaxation times can be estimated [TOE, Chapter 14]. If viscosity is due to the glide and climb of these dislocations then, in principle, there is a relation between Q and viscosity.