TOPSIDE TECTONICS

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TABLE 1: Convection and Plate Tectonics

Style of Convection	Successful ?	Problems
Isoviscous	NO	Many
Boussinesq	NO	Many
Constant properties	NO	Many
Pressure dependent properties	YES	
No phase changes	NO	Tomography, physics
Two dimensional	NO	Plate varticity
Cartesian—narrow box	NO	Degrees of freedom
Whole mantle	NO	Tomography, topography
Layered convection		
Shear coupled	NO	Tomography
“Primordial” lower layer	NO	Temperature, bathymetry
Boundary at 650 km.	NO	Geoid
Heated from below	NO	Bathymetry
“Thermal” coupled	YES	
Boundary at ~1000 km.	YES	
Multi-layered	Probably	
Boundary conditions		
Free slip	NO	Scale
Bottom heating	NO	Unphysical
Cooling from above	YES	
Internal heating	YES	
Plate-like	YES	
Assigned plate velocities	NO	Singularities, tomography
Spherical	YES	Resolution
Viscosity (T, P)	Essential	
á (T, P)	Essential	
Three-dimensional	Essential	Vorticity, Sprouts
Small-scale convection	TBD	
Self-forming plates	TBD	
Freely-evolving plates	TBD	

Abstract

Plate tectonics is a far-from-equilibrium self-organized system powered by heat and gravity from the mantle and organized by dissipation in and between the plates. Without plate tectonics and with simple boundary conditions the mantle would be the self-organizing dissipative system. Plates, slabs and continents control the pattern and style of mantle convection, not *vice versa*. Plate buoyancy and dissipation control plate motions, stresses, locations of plate boundaries, intraplate extensional zones and volcanic chains. The cold stiff outer shell of Earth is the active element and the template; the underlying convective mantle is passive. Active upwellings from the core-mantle boundary are not an important element in plate tectonics and volcanism. Shallow dynamical mechanisms are responsible for features attributed to *hotspots*, *plumes* or *ridge-plume* interactions. Fluctuations in stress permit *magmafracture, diking, linear volcanic chains*, and new plates, and can trigger global reorganization of plate tectonics and mantle convection.

Terminology

Passive – upwellings that are caused by divergence of plates, as at midocean ridges, or displacement by slabs; Convection that is driven by the upper boundary layers in fluids that are cooled from above.

Active – downwellings and upwelling that are driven by thermal buoyancy; the lithosphere is the active element in plate tectonics and mantle convection.

Dynamic – convection driven by lateral temperature gradients and lithospheric architecture, as at at edges and slots.

Plume – narrow stationary hot active upwellings from a lower thermal boundary layer; in plume theories it is high temperature rather than active convection in the plume or plume head that is emphasized.

Megaplume – the low heat flow from the core, the low coefficient of thermal expansion in the deep mantle and the high viscosity of the lower mantle requires very large features in order to build up buoyancy. In the lower mantle active upwellings are large and sluggish.

Boussinesq – an approximation commonly made in convection simulations; the effect of pressure and temperature on most material properties is ignored. The neglect of compression effects gave rise to the mantle plume hypothesis.

Introduction

The traditional view of mantle geodynamics and geochemistry is that observed structures such as volcanic chains and large igneous provinces, and phenomena such as continental breakup and plate reorganization, are due to convection currents, hotspots and hotlines in the mantle, and the importation of core heat, via plumes, into the upper mantle. Convection in a system with uniform boundaries can spontaneously organize itself into a pattern of convection cells. The pattern is created and maintained via dissipation, in the fluid, of the free energy provided from the underlying energy source. Dissipative structures result from amplification of temperature fluctuations that would die out in a close-to-equilibrium situation. When a plate tectonic and continental template is placed on top of the convecting system, it organizes the convective flow and the plates themselves become the dissipative self-organized system.

In this paper I suggest that mantle dynamics is almost entirely a top-down system and that mantle convection of various scales is controlled by plate tectonics, not *vice versa*. The surface boundary layer is the active element, the "convecting mantle" is the passive element. Furthermore, I propose that the surface plates are a self-organized dissipative system. Upwellings induced by lithospheric architecture are intrinsically unsteady, 3D, localized and variable in temperature, features which have been thought to require deep thermal anomalies and active upwellings.

I argue that the deep mantle communicates with the surface only by Newton's and Fourier's laws, not by advection of material, and that excess magmatism and volcanic chains are expected features of plate tectonics with real (non-rigid, non-uniform) plates. Because of pressure, and low heat-flow from the core, the lower thermal boundary plays a completely different role than assigned to it in current geodynamic and geochemical models(1).

Plate Tectonics and Mantle Convection

The Earth's surface is covered by about twelve large interlocking plates which move relative to each other and to the underlying mantle. The oceanic lithosphere is often considered to be part of mantle convection but the plates drive themselves, and mantle convection, through a combination of the buoyancy forces associated with cooling lithosphere and subducting oceanic lithosphere.

To date, attention of geodynamic modelers has focused on the rheology and forces of single plates or on the convection of the whole mantle with simple boundary conditions. The view that convection drives plate tectonics or that plates are simply the upper boundary layer of mantle convection has not progressed very far. Although plate tectonics and mantle convection are related, it has not been possible to successfully explain plates and plate tectonics from a thermal convection point of view. Viewing mantle convection as a consequence the result of plate tectonics and lithospheric conditions has been much more successful (Hager, O'Connell, Conrad, Christianson, Gurnis) (2-5).

It is commonly supposed that plate tectonic rates and mantle convection are controlled by the viscosity of the Earth's deep interior. However, it is dissipation in and between the plates that limit plate velocities and controls the sizes and shapes of plates. Surface plates determine the scale of thermal structures in the mantle, and the rate at which the mantle cools (4,5). A fluid cooled from above develops an unstable surface layer which drives convection in the underlying fluid and introduces thermal, chemical and density inhomogeneities into it. The plates and the downwellings are the active parts of the system. There are no active upwellings. The "hot" parts of the fluid are simply regions uncooled by subduction. Upwellings are passive responses to the active downwellings (Tackley, 6). A mantle heated from within by long-lived radioactive elements and cooled from above also has no active upwellings. The system is driven from the top. This is the opposite point of view than that adopted by Morgan ( ) in which narrow active upwellings drive the system. It also contrasts with the standard view (e.g. 1) that plates drive the "plate-scale" flow but narrow active upwellings from deep in the mantle control the locations of island chains and the bursts of magmatism that accompany plate reorganizations.

Mantle convection is extremely sensitive to changes in the surface boundary (Gurnis)(2). The introduction of a single continent, crack or a bend in the lithosphere, completely changes the style of convection. The collision of two continents can reverse mantle flow (Lowman, 7). Both plate tectonics and mantle convection represent balances between driving and resisting forces and the change in any force is immediately transmitted throughout the system and causes a global reorganization.

Variations in stress, radioactivity, thickness, motions and conductivity of the lithosphere strongly affect underlying convection (Perkeris, Hager, Kaula, Gurnis, Lowman, Anderson)(8-13). Lateral variations in the thickness, conductivity and radioactivity of the surface layer sets up lateral temperature gradients which drive small-scale convection. These convective velocities can reach 10 to 15 cm/yr., comparable or greater than velocities achieved by radial temperature gradients and Bénard convection driven by bottom heating(8).

Plates not only drive themselves but they can break themselves. Plate motions are controlled by the integral of body forces and resisting forces. Plate stresses are controlled by local buoyancy, boundary and thermal stresses. Interior stresses in plates are mainly lateral compression; that is why they hold together as plates. Stresses are extensional at plate boundaries and incipient plate boundaries (volcanic chains). Buoyant magma helps to break the plates in these regions (magmafracture) . Rocks have little strength in tension and lithosphere overlying partially molten mantle has even less strength.

Nevertheless, so-called midplate volcanoes and volcanic chains are usually attributed to narrow active upwellings from deep in the mantle rather than to the stress-state of the plate and lithospheric architecture. Although the surface boundary layer can generate narrow dense downwellings it is not evident that deep boundary layers behave in the same way ( ). There is symmetry in a uniform incompressible fluid heated from below but the mantle is far from this situation.

Although it is generally agreed (e.g. Davies, 1) that the top thermal boundary layer drives upper mantle convection, it is not agreed as to the relative importance of stress and fractures in the lithosphere (2,14) or of the lower thermal boundary layer (6,13). A self-consistent model of plate tectonics, seafloor spreading, continental breakup and drift, global heat-flow, episodic volcanism and midplate tectonics and magmatism is being developed (Keen, Korenaga, King, Lowman, Jarvis, Leonardic, Gurnis, Hager)(7,17) which is controlled from the top, with an entirely passive mantle. There is no need for, or evidence for, advection of heat or material from the deep mantle or the CMB.

The strong variation of radioactivity and material properties with depth creates drastic differences in the way the surface and lower thermal boundary layers behave (6). In particular, the outer surface frequently ejects narrow cold plumes into the interior. The lower boundary slowly grows a few giant instabilities which may last a large fraction of the age of the Earth. The mantle cools more rapidly than the core so the weak thermal boundary layer at the base of the mantle evolves gradually.

Early views of plate tectonics treated plates as responding passively to mantle convection. Ridges were the upwellings and slabs were the downwellings. It was then recognized that plates could drive themselves and could organize mantle convection. Plates provide the template for "hotspot" magmatism and tectonics, phenomena not obviously related to plate tectonics or plate boundaries. Narrow hot upwellings are held responsible for "hotspots", linear upwellings for "hotlines", giant upwellings for large igneous provinces and continental breakup and for influencing plate motions. These ideas developed from the tacit assumptions that the lithosphere is rigid and uniform, the upper mantle is isothermal and homogeneous and that locations of volcanoes are controlled by mantle temperature and convection, not the stress state of the lithosphere.

Non-Equilibrium

A far-from-equilibrium system can act as a whole, in spite of the short-range character of the interactions. Plate tectonics can be considered as a non-equilibrium process in the sense that a small perturbation in the system can cause a complete reorganization in the configuration and sizes of plates and the planform of mantle convection. Such perturbations include stress changes in a plate, and ridge--trench and continent--continent collision and separation. Changes in boundary conditions are transmitted essentially instantaneously throughout the system and new plates and new boundaries form. These topside changes cause mantle convection to reorganize (Gurnis) or reverse (Lowman). The lithosphere is not *rigid*, as generally assumed, but is fractured and is weak in extension. It reassembles itself into new plates, held together by lateral compression. Regions of extension become volcanic chains and plate boundaries. The sign of the horizontal stress field determines what becomes a plate and what becomes a plate boundary.

Plate tectonics is, therefore, an example far-from-equilibrium of *self-organization*. If the convecting mantle was bounded by isothermal free-slip surfaces it would be free to self-organize and to organize the surface motions, but it is not and cannot.

Near equilibrium, systems respond to a fluctuations by returning to equilibrium. Systems far from equilibrium may respond by evolving to new states. The driving forces of plate tectonics are thermal and gravitational and therefore change slowly. Dissipation forces, such as friction and continent-continent collision, however, change rapidly as do normal stresses across subplate boundaries and tensile stresses in plate interiors. These stress-related fluctuations can be large and rapid and can convert the slow steady thermal and gravitational stresses into episodic 'catastrophes'. This is the essence of *self-organization*. This is the reverse of the currently popular view that it is mantle convection that causes breakup and reorganization of plates and giant igneous events, i.e. a rapidly changing and localized event in the convecting mantle or at the core mantle boundary.

Plates are not permanent; they are temporary alliances of subplates. Global plate reorganization processes episodically change the directions and speeds of plates and redefine the plates. Plates annex and lose territory to adjacent plates and they breakup or coalesce. New plate boundaries do not form all at once but evolve as age progressive chains of volcanoes. Volcanic chains can also be extinguished if lateral compression takes over from local extension. Volcanism can be turned on and off by changing stress but it is not so easy to turn off plume volcanism, or to suddenly reduce the temperature of the mantle. Most island chains do not have the orientations or histories of uplift and volcanism predicted by the plume hypothesis ( ). This plus the absence of heat flow and deep tomographic signatures favors lithospheric and stress, rather than thermal, control for all classes of volcanism.

In plate tectonics and mantle convection, as in other slow viscous flow problems, there is a balance between buoyancy forces and dissipation. Plates and slabs are driven by gravity and resisted by mantle viscosity and friction between plate subunits. Resisting and dissipating forces in and between plates can change rapidly, and are communicated essentially instantaneously throughout the system. Episodic global plate reorganizations are inevitable. Even slow steady changes can reverse the normal stresses across fracture zones and can start or shut off volcanic chains. Construction and erosion of volcanoes change the local stress field and can generate self-perpetuating volcanic chains ( ). All of these phenomena are controlled by the lithosphere itself, not by a hot convective template from the underlying mantle or introduction of core plumes into the shallow mantle.

Self-Organization

Self-organized systems evolve via dissipation when a large external source of energy is available, and the systems are far from equilibrium. This self-ordering is in apparent violation of the second law of thermodynamics (Prigogone). This process is called “dissipation controlled self-organization.” A fluid heated uniformly from below is an example of far-from-equilibrium self-organization. The static fluid spontaneously organizes itself into convection cells but one cannot predict when, or where the cells will be, or their sense of motion. In contrast, if the heating or cooling is not uniform, the boundary conditions control the planform of convection. The fluid is no longer free to self-organize. In most mantle convection simulations the boundary conditions are uniform and various patterns and styles evolve which do not replicate conditions inside the Earth or at the surface. In many simulations the effect of pressure on material properties is ignored. The few simulations that have been done with more realistic surface boundary conditions (Gurnis) and pressure-dependent properties (Tackley) are more Earth-like, but plate-like behavior and realistic plate tectonics do not evolve naturally.

Because of the inordinate sensitivity of mantle convection and lithospheric stress to surface conditions it is necessary to exhaust the various plate tectonic explanations for episodicity, plate reorganization, volcanic chains, periods of massive magmatism, continental breakup etc. before one invokes unstable deep thermal layers and narrow active upwellings ("plumes").

The difficulty in accounting for plate tectonics with computer convection simulations may be explained if plates are a self-organized system and *they* control mantle convection rather than *vice versa*. Since the plates exert so much control on convective motions, velocities, interior temperatures and cooling (or heating) rates, and scale of convective structures it is probably not reasonable to suppose that convection calculations can reproduce plate tectonics. Plates cannot be treated as a small perturbation to a purely hydrodynamic situation or as an imposed boundary condition.

Far from equilibrium systems do strange things. They become inordinately sensitive to external or internal influences. Small changes can yield huge, startling effects up to and including reorganization of the entire system. We expect self-organization in slowly driven interaction-dominated systems. The resulting patterns do not involve templates or tuning. The dynamics is dominated by mutual interactions, not by individual degrees of freedom. Periods of gradual change or calm quiescence are interrupted by period of hectic activity. I suggest that such changes in the geologic record are due to plate interactions and plate reorganization, rather than events triggered by mantle convection.

Dissipation

Although plate-driving forces are now well understood, the resisting or dissipative forces are the source of self-organization. Surprises are still in store regarding their sources and effects (Conrad, Hager). Bending forces in plates, for example, may control the cooling rate of the mantle. Sources of dissipation include the viscosity of the mantle, friction along faults internal plate deformation, continent-continent collision, and so on. These forces also generate local sources of heating(Yuen). Dissipative forces and local stress conditions can change rapidly.

An important attribute of plate tectonics is the large amount of energy associated with toroidal (strike-slip) motions. This does not arise directly from buoyancy forces involved in normal convection. In a convecting system the buoyancy potential energy is balanced by viscous dissipation in the fluid. In plate tectonics both the buoyancy and the dissipation are generated by the plates. The slab provides the driving buoyancy and this is balanced almost entirely by slab bending (Conrad, Hager) and by transform fault resistance as characterized by the toroidal/poloidal energy partitioning of plate motions (O'Connell et al). This further confirms the passive role of the mantle. Cooling of the Earth is not currently governed by the viscosity of the mantle but by deformation and sliding resistance of the plates.

The system reorganizes itself to minimize dissipation. It can do this most effectively by changing the lengths, directions and normal stresses across transform faults, reducing the toroidal component of plate motions, localizing deformational heating, increasing the sizes of plates and so on. If plate dynamics is a far-from-equilibrium system, sensitive to initial conditions or fluctuations, then we expect each planet to have its own style of behavior, and the Earth to have behaved differently in the distant past and, perhaps, during different supercontinent cycles.

Dissipation is also involved in the style of small-scale convection. Tabular or linear upwelling evolve as they rise to 3D plume-like upwellings. Longitudinal and transverse rolls develop under moving plates and in Hele-Shaw (slot convection) cells to minimize dissipation. Two-dimensional convection simulations do not capture these 3D effects. This is one reason why an independent style of convection, and source of heat, is after invoked to explain "hotspots." Plate tectonics on a sphere has additional out-of-plane sources of lithospheric stress which may account for linear volcanic chains (Turcott, Oxburgh, Sandwell).

Plate tectonics is a slowly driven, interaction-dominated system where many degrees of freedom are interacting. The dynamics is dominated by mutual interactions, rather than by the intrinsic dynamics of individual degrees of freedom. Stresses in plate interiors can change much more rapidly than plate motions since the latter are controlled by the integral of slowly varying buoyancy forces. Apparent changes in orientation and activity of volcanic chains are therefore more likely to be due to changes in stress than to changes in plate motions (Jackson). The stress-crack hypothesis seems to apply to most volcanic chains (McNutt, Dickinson, Favela) while the core-plume hypothesis accounts for only a fraction of intraplate magmatism (Wessell). Dramatic changes in plate configurations in part of the world (e.g. Pangea breakup) are likely to be accompanied by a global reorganization (e.g. formation of new plate boundaries and triple junctions in the Pacific and Farallon plates) and creation of new plates. Minimization of dissipation appears to be a useful concept in global tectonics (Sleep, O'Connell). Obviously, formation of a new crack, ridge or suture may reduce TF or collisional resistance, and may change or reversal mantle convection.

Plates, like other dynamical systems, organize themselves into structures which interact. They are provided with energy from the mantle and they evolve under the influence of driving forces and interaction forces. There is no single characteristic plate size or characteristic time scale. Such systems organize without any significant tuning from the outside.

Topside Tectonics

Although the mantle is slowly heated from within and below, it is the rapid cooling from above that dominates Earth's thermal history and dynamics. Because of continents, cratons and super-continents this cooling is not uniform in space and time. It is the instability of a large part of the outer shell that drives plate tectonics and cools the interior. The recycling of water into the upper mantle lowers the melting point and viscosity and helps maintain the process. From this point of view the surface temperature of a planet and the nature of the atmosphere may control the style of mantle dynamics.

Plate tectonics creates chemical heterogeneity and temperature and melting variations in the upper mantle. Convection is not an efficient stirring or homogenization agent but the Central Limit Theorem assures us that large volume sampling, such as at the mid-ocean ridge system, will yield products (basalts) that are much more homogeneous than at oceanic islands, seamounts and fracture zones, and slow-spreading ridges.

The top-down school of mantle dynamics and convection has a long history, with major contributions made by Perkeris, Elsasser, Elder, Knopoff, Hager and O'Connell. It has only recently been appreciated that the lithosphere can regulate the heating and cooling of the whole planet (Gurnis, Lowman, Christensen, Hager, Conrad) and can reorganize or reverse mantle flow (Jarvis, Lowman).

Elsasser, in particular, argued that convective circulation in the mantle is largely controlled by irregularities at the top. He showed that horizontal sliding of the top layer was mechanically easier than circulation of the material below. He assumed, along with Birch and Patterson that the lower mantle had been drained of its radioactive elements and lighter constituents at an early stage. This is confirmed by mass-balance calculations (Anderson).

Elsasser specifically ruled out thunderstorm like convection, with concentrated vertical updrafts and slow distributed downdrafts, as proposed by Morgan ( ).

Temperature variations and lithospheric architecture at the top of the system can create upwellings of vertical velocity, 10-15 cm/yr., comparable or greater than rifting and plate velocities (Korenaga, 2000). No excess temperature is required to generate thick igneous crust at the edges of cratons or continents, or along fracture zones and sutures.

Although the mantle can be treated as a viscous fluid for long-term processes, its behavior is quite different from normal pot-on-a-store Rayleigh-Bénard convection. Because of pressure and internal heating the upper and lower thermal boundary layers do not play equivalent roles. Although narrow cold downwellings (e.g. slabs) can peel off of the upper boundary layer, the deep mantle is characterized by buoyant structures orders of magnitude layer ( ). This is confirmed by seismic tomography, theoretical considerations and numerical calculations. Narrow upwellings are not a normal part of a convecting mantle with realistic parameters. Plume simulations involve injecting hot fluids into a static tank of fluid, heating a small patch at the base of the system, or numerically, by inserting a thermal singularity in the system before turning on gravity. These simulations ignore effects of pressure, phase changes and background convection. No known physical process can create these singular initial or boundary conditions.

The hypothetical deep narrow upwellings of the core plume hypothesis are beyond the resolution of geophysical imaging but giant horizontal features are predicted as these plumes spread out under chemical, viscosity and phase--change barriers. These easy to image large-scale features are not evident under oceanic islands or recent continental flood basalt provinces. Instead, volcanoes of all kinds, including midocean ridges, are concentrated over large scale hot regions of the upper mantle. These regions correlate with supercontinent locations and with the absence of subduction cooling.

Rayleigh-Bénard Convection

The most familiar form of convection is the uniform fluid heated form below. The vigor and style of this convection is controlled by the Rayleigh number

$$Ra = q\acute{a}h^{3}\ddot{A}T/ \acute{\imath}k$$

where í, k, á , q, h and ÄT are the viscosity of the fluid, the thermal diffusivity, the thermal expansion coefficient, the thickness of the fluid layer and the imposed temperature difference between top and bottom, respectively. Below the critical Rayleigh number $(Ra)_{cr}$ of about 10^3 the fluid is stable and the heat is conducted. Above $(Ra)_{cr}$ the stable fluid is sufficiently far from equilibrium that small fluctuations cause it to self-organize into a regular hexagonal pattern of convection cells. For a uniform fluid, with uniform isothermal boundaries, the pattern is extraordinarily regular. As the Rayleigh number is varied, the pattern can change from regular cells, to spokes, to rolls and ultimately to chaotic convection.

For a fluid heated from within and cooled from above the upwellings become broad and unsteady, since all parts of the fluid must rise to get rid of their heat, and the cold downwellings are narrow slab-like features, evolving to drips at depth. A chemically layered mantle will also involve broad upwellings in the deeper layers because the coefficient of thermal expansion is suppressed by pressure, and even small intrinsic density changes across boundaries can trap the hot upwelling material ( ). Seismic tomography confirms the general picture of narrow upper mantle downwellings and broad (thousands of kilometers) lower mantle upwellings.

Volcanoes and volcanic chains are very small scale (10^2 km) compared to theoretical and observed (tomography, geoid) thermal anomalies in the mantle (10^4 km). One needs to “cascade” the thermal anomaly down in characteristic scale by several orders of magnitude. This can be done by small scale convection or by lithospheric processes involving focusing or magma fracture ( ).

Temperature Fluctuations

The temporal and lateral temperature fluctuations that, are of necessity, associated with a convecting system are a weak function of Rayleigh number, Ra;

$$\ddot{a}T / \ddot{A}T \sim Ra^{-1/6} = 0.37\ Ra^{-0.145} \qquad \text{(Niemela et al, 2000)}$$

where äT and ÄT are, respectively, the rms temperature fluctuations and the total temperature rise across the system. For the mantle, the fluctuating part translates to about ±200 °C, which encompasses the whole range of temperatures inferred from petrology and geophysics along “normal” ridge segments plus temperatures inferred for “hotspots” and “coldspots” (Anderson, 2000). The importance of this result cannot be overstated. The mantle is often assumed to be almost isothermal, with all excess temperature fluctuations due to the lower thermal boundary layer.

Convection

Convection in the mantle need not involve upwellings. The active, or driving, elements may be cold dense downwellings with complementary passive upwellings.

I need to distinguish between *passive*, *active*, and *dynamic*. Ridges and dense downwellings can induce upwelling. These are termed *passive*. It is generally agreed that upper mantle convection is passive; the plates and the slabs are the active elements (Davies). In the plume hypothesis the narrow hot upwelling are *active*; they uplift and break the plate, spread out in the asthenosphere, and drive plate tectonics (Morgan). Although it is no longer believed, by most, that plumes drive plate tectonics it is an important part of the theory. As Morgan states, if plumes are not strong, the ridges would close up, and the hypothesis would fail. It is also important to the plume hypothesis that plumes provide about half the mantle heat flow. This is off by an order of magnitude. Today, plumes are not treated as *active* but simply as *hot*. Regions of excess volcanism are considered to be hotter and more buoyant than normal mantle. This doesn't explain why some 'hotspot' provinces have low elevation and heat flow (Czaminski, McNutt), and doesn't take into account the "normal" fluctuations in temperature.

Small-scale convection is set up by lateral temperature gradients associated with plate architecture; corners, edges, boundaries, rifts and slots. Convection driven by these gradients, and by diverging plates, can be quite vigorous. By rapidly circulating mantle through the shallow high-melting regime, more melt can be generated than by simple passive upwellings even with normal temperature mantle. I call this a *dynamic* process to contrast it with melting that is controlled by temperature alone or by passive upwellings. Calculations show that this kind of dynamic process can readily provide the rates and volumes of magma required for "hotspot" magmatism (King, Keen, Korenaga).

Lower Thermal Boundary Layer

If the core is hotter than the base of the mantle a thermal boundary layer will form, allowing heat to flow out of the core, Since the core is almost isothermal while the mantle can support large temperature variations most of the core heat loss is to the colder parts of the lower mantle. There is probably very little radioactivity in the core so it is primarily heat from growth of the inner core and cooling of the outer core that is conducted across the CMB. These heat losses are controlled by the mantle, not by the core. The core will cool only as fast as the heat can be conducted across the variable TBL at the base of the mantle.

The cooling (or heating) rate of the core is quite uncertain but the energy required to sustain the dynamo can be maintained by a heat loss of less that 1% to 10% of the Earth's total heat flow. Heating of the base of the mantle by heat leaving the core if better described as differential cooling of the core by the mantle since the low-viscosity core is a passive element.

Convection in the mantle is dominated by radioactive heating and secular cooling, not by core heat. There is likely to be small-scale convection in D½ and this apparently has been observed (Helmberger). Interestingly, these small dome-like hot feature appear mostly at the edges of cold regions, where core heat may be diverted.

A chemically distinct layer at the base of the mantle need have an intrinsic density excess of only 1 to 2% in order to be isolated, since the coefficient of thermal expansion is so low. However, this implies very large thickness changes for this layer. There is likely to be large scale relief on any internal density boundary in the mantle, and small-scale convection, inside the layers. In the upper mantle, intrinsic density differences of order 5% can be breached. This means that chemical boundaries, as between fertile and residual peridotite or eclogite, can be breached by slabs or asthenospheric upwellings of quite different chemistry and intrinsic density. The resulting upper mantle stratigraphy is reversible. For the same reason, the Earth is easily stratified during accretion (high-T, low-P) but not easily unstratified at depth, today (high-P). The hot lower thermal boundary layer may be stabilized by chemistry and pressure dependent properties. It is certainly convecting and highly variable in properties and thickness but it probably has nothing to do with plate tectonics and volcanism which seem to be entirely controlled by the cold surface boundary layer and the hot underlying mantle. Midplate volcanism is likely also controlled by properties of the lithosphere and asthenosphere, not by the core--mantle boundary.

The local Rayleigh number of a thermal boundary layer depends on heat flow, conductivity, viscosity and coefficient of thermal expansion. A boundary layer will go unstable when its local Rayleigh number exceeds a critical valud (Howard, Elder). Because of the low heat flow, low á and high-viscosity at the base of the mantle the bottom boundary layer must be an order of magnitude or more thicker than the upper TBL before it goes unstable. According to Tuckley (2000) these enormous lower TBL instabilities may rise most of the way into the lower mantle. Because of the low heat-flow from the core (0.01 to 0.1 of surface heat-flow) it takes a long time for thermal buoyancy to accumulate at the base of the mantle. The low á and high viscosity means that rise time is also long. Furthermore, the mantle must cool substantially before there is any temperature difference across the CMB. The lower TBL grows with time.

A density difference as small as 1% is adequate to trap hot upwellings in the lower mantle.

These considerations suggest that the large-scale features in the deep mantle, revealed by seismic tomography, may be ancient, perhaps billions of years old. Even if they are trapped in the lower mantle and can communicate with the shallow mantle only by conduction (and gravity) they will influence upper mantle convection including, possibly, the locations of downwellings and continental aggregations. In most studies it is presumed that features in the high-viscosity deep mantle are controlled by descending cold currents in the upper mantle, but the reverse is more likely. The analogy is a broad flat pan of water sitting atop a stove with two burners on.

Deep Active Upwellings

Tomography is able to resolve the differences between a deep buoyant upwelling and a passive upwelling, even if the upper mantle stem of a plume cannot be resolved.

An active upwelling, in contrast to a passive upwelling, will spread out laterally under the lithosphere and at endothermic phase changes (e.g. the 650 km. discontinuity) because of its excess buoyancy and flow rates greater than passive spreading rates. The upwelling will be broad in high-viscosity regions of the mantle and narrow as it enters the low-viscosity asthenosphere. None of these characteristics are evident under Iceland, the only deep mantle tomographic structure that has been attributed to a through–going plume ( ). On the other hand the detailed seismic studies of the Iceland upper mantle are consistent with dynamic flow, driven by opening and cooling of the North Atlantic mantle and confined by the bounding cratons and the 650 km. discontinuity. The numerous plateaus in the Atlantic and Indian oceans all formed at about the same distance from continents and in the middle of 2000 km. wide oceans. They are probably all EDGE effects (Vogt, King, Anderson), i.e. complementary upwellings to the cold downwellings at cratonic edges ( ).

Volcanic Chains

If much of the upper mantle is at the melting point, as now seems unavoidable, then there is a simple explanation for island chains, extensive diking at onset, or prior to, rifting and large igneous provinces. The criterion for diking is that the least compressive axis in the lithosphere is horizontal and that melt buoyancy can overcome the strength of the plate. Magma can break the lithosphere at lower deviatoric stresses than it would otherwise take. Dikes propagate both vertically and along strike, often driven by the hydrostatic head of an elevated region (Fialko). Dikes, like cracks, can propagate laterally. Typical crack propagation velocities are 2 to 10 cm/yr, comparable to plate velocities ( ). These velocities are known from the study of ridge "propagators" and the penetration of ridges into continents. These propagating cracks are volcanic so they are probably examples of propagating magma-filled cracks or dikes. These velocities place constraints on the viscosity of the asthenosphere ( ) which turns out to be appropriate for small-scale convection and 3D upwellings ("sprouts") from the shallow mantle, common along spreading ridges (Parmentier).

Slot Convection

Supercontinents cause heat to accumulate in the underlying mantle and they are broken up by far-yield stresses created by subducting slabs augmented by the flow driven by plate tectonic generated lateral temperature gradients (Lowman). The most dramatic temperature gradients are set up when a craton splits, allowing hot asthenosphere to rise into the "slot." The newly exposed mantle cools rapidly from the top and the sides, setting up vigorous small-scale convection which removes cold material, replacing it with hot asthenosphere. Similar small-scale convection is set up at all lithospheric "jumps" including transform faults (both active and dead), rifts, continental margins, abandoned ridges and so on. It is no accident that most, if not all, volcanic chains and large igneous provinces are associated with this type of convection was discussed by Nussult and Elder and is referred to as the Hale–Shaw configuration. Excess volcanism is often attributed to excess temperature but vigorous small-scale convection can also increase the local melting rate.

Thermal convection of a viscous fluid contained between two vertical isothermal walls is a classical problem in fluid mechanics and is an important but ignored problem in geodynamics. In this configuration, in contrast to the Rayleigh-Bénard case, the fluid cannot remain at rest. There is spontaneous convection, no matter how small the temperatures difference.

Convection is controlled by the Elder number

$$E = \frac{g \alpha \theta L^3}{\nu k}$$

where È is the temperature difference, either between the walls or between the wall and the fluid, and L is the wall separation, or the width of the slot.

A low E there is a single elongate vertical cell with cold fluid sinking along the cold wall and rising along the hot wall. If both walls are colder than the fluid then the hot fluid rises in the center and there are two counter-rotating elongate cells. At higher E there are roll-like secondary cells, parallel to the walls. The form of convection is controlled by buoyancy vs. shearing forces, and heat conduction into the side-walls

This configuration mimics the initial opening of an ocean between two cratons, e.g. the early opening of the North Atlantic. The extra complications in that case are the extra cooling at the top of the “slot” and the resupply of hot mantle from below.

For a long narrow slot there eventually develops a 3D instability (Korenaga). The sheet-like upwellings develop a “sprout” in the center, or a series of equally spaced “sprouts”, stubby plume-like projections off of the linear upwelling. For and open topped configuration the central “ridge”. As the slot widens, as in the case of an opening ocean, the hot material is eventually swept to the side by the linear upwelling and cools and descends at the wall.

For $E < 10^3$ there is weak unicellular flow, for $10^3 < E < 10^5$ there is stronger flow and at $E > 10^5$ secondary roll convection sets in. For $E > 10^6$ the secondary flow is vigorous. At $E > 10^7$ the flow becomes unsteady and travelling waves, or solitons, are permanent. The flow is “turbulent” or chaotic at $E > 10^7$.

The Hale-Shaw-Elder configuration can be used to analyze the flow of magma up dikes at the initial stage of rifting, and the upwelling of asthenosphere between two separating cratons, at a later stage of drift.

An analysis by Korenaga (   ) of the early Atlantic opening problem gives vertical velocities of >10cm/yr., for asthenospheric viscosity of 10^{18} Pa.s.

With typical mantle parameters the Elder-type convection goes from conduction dominated slow unicellular flow at 100 km width to 3D turbulent flow at a width of 1000 km, with the horizontal roll, vertical "sprout" and unsteady – flow regimes setting in at in intermediate widths. The preponderance of oceanic plateaus that formed when the proto-Atlantic and adjacent Arctic ocean are in this stage now (the Iceland and, perhaps, Van Mayen plateaus). The Bermuda, Cape Verde, Rio Grande (S. Atlantic) and Kerquelen Plateaus may have formed as a consequence of this EDGE, or slot, type convection (King and Anderson), when the cold edges of the oceans were in "resonance".

A very early analysis of upper mantle convection driven by horizontal temperature gradients, as between cratons and oceans, showed that temperature differences of 500   C over 800 km distance could drive convection at 10cm/yr even for high mantle viscosity's (10^{21} cm ^{2}/sec) (Allen et al). Temperature gradients across new rifts and oceans and at edges of continents also generate motions of about 10cm/yr (Korenaga). Upwellings mantle melts extensively as it rises above about 60 km. depth so rapid advection creates large igneous provinces with no temperature excess. New oceans that form as a result of supercontinents breakup, however, can by expected to have mantles that have excess temperatures but these are generated at the top of the convecting system.

Sprouts

Shallow plume-like upwellings ("sprouts") are a natural part of convection in the mantle between separating plates. They are ephemeral and unsteady; they form at intermediate stages of a newly forming and widening ocean, and at slow-spreading ridges. They are preceded by sheet-like upwellings at an earlier stage (continental margin dipping reflectors) and dikes at the incipient stage. After an ocean is >2000 wide, normal plate-driven midocean ridge volcanism takes over. These "sprouts" are not imposed artificially from outside (by injecting a fluid at the base of the system, for example) or by an *ad hoc* singularity (Cordery), as in the plume hypothesis.

Sprouts are stubby, aspect ratio ~1, and are not parallel to the driving ÄT gradient. They are shallowly rooted and driven from the top of the system. They are due to cooling from above and by the side-walls, and side-wall friction, *not* to heating from below.

The convection associated with cratonic separation and Elder-flow extends deep into the mantle but is terminated by endothermic phase changes or jumps in mantle viscosity (King). This type of flow is consistent with tomographic images of the North Atlantic (Ritsema, Foulger). It is similar to the edge-effect proposed by Vogt( ).

The controlling parameters for this type of explanation for Iceland (and other Atlantic and Indian ocean plateaus) are width of the ocean, depth of the "solid" walls, thickness and viscosity of asthenosphere, and the imposed separation velocity.

We expect tabular upwellings below the projected depth of craton bottoms, and 3D-plume-like-flow at shallow depths, as observed tomographically (Foulger). The 2D tabular upwelling also dominates at the beginning of separation and evolves with time to a shallow sprout in the upper part of the system. Jha et al (1994) show that 3D upwellings are preferred at slow spreading ridges if the asthenospheric viscosity is low enough, $\sim 10^{19}$ Pa.s.

The North Atlantic is not only a slow-spreading ridge but the mantle is probably hot, because of continental insulation, and has a low-melting point (because of volatile-content). Thus, this is an ideal situation for both shallow stalk-like convection and EDGE convection.

The preference of 3D convection in a fluid bounded by sidewalls is only for stationary state convection – it does not apply to the early stages of transient convection (Korenaga). Thus, linear ridges, dikes and continental margin igneous features (Seaward Dipping Reflectors) will precede the formation of oceanic plateaus and islands which form at a later stage of new ocean opening.

Hele-Shaw (Slot) Convection

The Iceland Plateau is centered in the recently (65 M.y.) opened north Atlantic, about 1000 km. from ancient cratons in Greenland and Fennoscandia. It may be a modern example of the oceanic swells and plateaus that occur about 1000 km. offshore of the conasts of the Americas and Africa (Bermuda, Rio Grande Rise, Cape Verde…) and were volcanically active when they were centered in a new 2000 km. wide ocean. If continental edge effects cause asthenospheric upwellings about 1000 km. from the edge, ( ), then the Atlantic-Indian plateaus may be "resonance" effects. Korenaga ( ) has shown that isolated 3D structures are the expected result of instabilities at the top of linear upwellings. Even long linear midocean ridges have variable elevation along strike, a combination of normal temperature variations and the intrinsic 3D nature of initially ridge-like upwellings. Iceland, and the Iceland Plateau, are ideal testing grounds for the idea that diverging cratons can generate episodic and transient bursts of magmatism without involving core-heat (King, Anderson, Korenaga).

The Korenaga (2000) theory explains the rapid fluxing of the upper mantle through the melting zone, the episodic or pulsating nature of the volcanism and the shallow plume-like structure that forms sometime after the initiation of the rift. The absence of a continuous ridge from Iceland to Greenland and the termination of the low-velocity tomographic anomaly at 650 km. are unexplained by the core-plume hypothesis but are expected features of the topside hypothesis ( ).

Pulsation's

All of the top-, edge- and slot-driven convective flows discussed in this paper are transient unsteady or episodic. It is likely that variability in magma production rates is mainly controlled by lithospheric conditions, such as stress, the intrinsic episodicity of topside driven flows may be responsible for the large fluctuations in magmatism seen at the edges and centers of new oceans, as in the Greenland-Iceland provinces.

Hotspot Tomography

The best imaged hotspots are Yellowstone (Salzer and Humphreys, 1997) and Iceland (Walfe, Foulger, Keller). Neither is consistent with a plume model. Interestingly, both are bounded by thick Archean lithosphere, as are most continental flood basalt provinces (Anderson). The small seismic anomalies extend down to about 200 km. about the depths of the adjacent Archean lithospheric keels. The anomalies narrow upwards, consistent with focused passive upwellings rather than broadening active upwellings. It has previously been noted that there is no evidence for sublithospheric spreading of hot material in other hotspot areas (Anderson) and no evidence for uplift (Czamanke) or heat-flow anomalies (McNutt). Passive or EDGE upwelling is indicated. In the casee of Iceland the anomaly grades downward into a ridge-like anomaly. Global tomography (Zhou, Ritsema) shows that the low-velocities under Iceland and the north Atlantic terminate at 650 km., as expected for a top-down process (King and Ritsema).

Hawaii

The Hawaiian chain is the most often quoted example of an obvious plume hotspot. The age progression and the large swell surrounding the southeastern part of the chainare usually taken as sufficient evidence for a plume origin. However, heat flow and seismic measurements are inconsistent with the thermal swell hypothesis, and the modulation of volume stress, and geochemistry by fracture zones suggest an important control by the lithosphere ( ). Age progression is consistent with a propagating crack, volcanic flexure or propagating dikes. The inferred temperatures of the mantle under Hawaii are consistent with the normal range of upper mantle temperatures.

The Evolution of a Young Ocean; From Rift to Drift

At the onset of rifting the newly exposed mantle is cooled from above and by the sidewalls. When the rift is narrow, hydrothermal fluids preheat the sidewalls, cool the mantle and cause wallrock alteration. Magma buoyancy drives sills and dikes to shallow depths but magma viscosity and cooling limit the volumes. Further extension of the lithosphere is accompanied by normal faulting, extensive diking and sedimentary basin formation. Eventually, the rift is wide enough to allow asthenosphere to intrude the space. The thermal gradient across widening rifts drives vigorous dynamic convection which supplements passive buoyancy driven passive upwelling. Sidewall friction and lateral cooling resists the vertical flow; lateral and vertical temperature gradients and sidewall erosion accelerate the flow. Wide rifts are characterized by rapid, pulsating flow, with extensive melting of asthenosphere that is driven through the melting zone ( ). Longitudinal rolls may form between the sidewalls and eventually fully 3D flow develops, with sprout-like upwellings at the top. This is the time of midocean plateau formation (e.g. Kerguelen, Bermuda, Iceland). Flow is unsteady but rapid.

As the ocean widens, and time goes on, the horizontal temperature gradient becomes less important and plate scale flow takes over. The plateaus become inactive and are stranded about 1000 km. offshore (e.g. Bermuda, Rio Grande Rise, Cape Verde).

The mantle in a narrow ocean cools and convects rapidly because the asthenosphere is loosing heat both out the sides and out the top. As the boundary layers thicken the heat lose decreases. The vertical TBL may ablate or fall off. In any case, the convection at a new rift and narrow ocean is episodic, reaching velocities of 15cm/yr. (Korenaga).

The 650 km Discontinuity

The mantle transition zone (Bullen's region C) extends from 400 km to 1000 km depth. In the 1960's it was discovered that C contained two major discontinuities near 400 and 650 km (Anderson, Toksoz, Niazi, Johnson, Kovach, Archambeau). These were interpreted as phase changes in mantle minerals, primarily the olivine -"spinal" phase change at 400 km and the "spinal" – "post-spinal" phase change at 650 km (Anderson, 1967). The latter one was inferred to have a negative Clapeyron slope and therefore a potential barrier to mantle convection (Anderson, 1967). Some evidence for a discontinuity near 1000 km was also found (Repetti, Archambeau, Whitcomb) which corresponded to the top of the lower mantle as defined as the basis of velocity gradients by Bullen ( ) and by ratios of elastic constants by Anderson ( ).

Although there may be a slight change in chemistry across 650 km (Anderson, ,), perhaps due to slab trapping there is little doubt that the seismic discontinuity itself represents primarily a change in crystal structure, i.e. a phase change (Anderson, 1967). Whether it is an *equilibrium* phase change or the same phase change, at all locations is not completely clear. The variation in topography is consistent with an equilibrium phase change (Shearer).

Geochemical models have long associated this boundary with a major compositional boundary, usually between “depleted” upper mantle, and “undegassed primordial” lower mantle (Wasserburg). Many geodynamicists followed suite, with hypothetical chemical changes associated with this boundary rather than elsewhere. Geodynamic arguments against layered convection are usually arguments against this particular depth for the chemical change (Hager). Wen and Anderson showed that the geoid and dynamic topography were consistent with a small deeper (~1000 km) chemical discontinuity.

Anderson (1989) showed that the geochemical mass balance calculations which appeared to support a depleted upper mantle were more consistent with depletion of most of the mantle, presumably during accretion, to form the crust and upper mantle. This agreed with the ^{40}Ar in the atmosphere which demanded efficient differentation and degassing of the mantle. This produced a model with a lower mantle depleted in radioactive, and other incompatible, elements, and with a major element stratification created by magma ocean separation of materials with different densities and melting points. Olivine, for example, would be concentrated in the upper mantle and the dense residue would have lower Mg/Si ratios, as do chondrites.

Chemical separation of an accreting planet is primarily near the surface (low-P, high-(accretional)-T). As the planet grows and cools the deep material, which may have participated in mantle convection at earlier stages, is trapped by high-P, and loew-T, condidtions which reduce the chemical buoyancy. During accretion most of the lange-ion-lithiphile (LIL) elements are taken to the shallow mantle by buoyant ascending melts. At some pressure melts may become denser than the residual mantle (Stolper) but this reverses again at phase change boundaries. So the LIL, including K,U and Th have been concentrated in the crust and upper

(<400 km) mantle since the beginning. These elements are recycled down to the dehydration boundary of slabs (200-300 km).

Mass balance calculations show that all of the geochemical reservoirs for various kinds of basalts – island, ridge, arc – can exist in the upper mantle (Anderson,) which is diluted at mature ridges but shows up at the onset of extension.

Chemical Stratification

The effect of pressure on the coefficient of thermal expansion, á, is such that chemical discontinuities are easy to maintain in the deep mantle (Anderson, Tackley). Intrinsic density differences of 1 to 2% can irreversibly stratify the mantle (Tackley). The belated reconition of the importance of compressibility has led to the widespread belief that chemical straitificaiton is difficult or impossible.

Quantative analysis of tomogrphic models (as opposed to visual inspection of a few selected cross-sections) strongly supports a barrier near 1000 km depth, which may be chemical. There is also a large increase in viccosity near this depth. Viscosity variations alone may stratify mantle convection ( ).

A small chemical discontinuity near 1000 km explains the geoid and the dynamic topography ( ). It also explains the correlation of tomography with past subduction ( ). This is the boundary between Bullen's regions C and D, the top of the lower mantle. Although some slabs may penetrate as deep as ~1000 km in the mantle they will certainly be slowed down by the negative Clapeyron slope at 650 km and a possible buoyancy barrier below 600 km where eclogite and depleted harzburgite are less dense than normal mantle.

The early arguments in favor of whole mantle convection involved the length scale of tectonic plates and the perceived homogeneity of the mantle in viscosity and mean atomic weight

(Elsasser et al,). It is now evident that plate scales are different from convective scales, that viscosity varies by orders of magnitude with depth, and that mean atomic weight is not a useful measure of mantle homogeneity. The components that vary with depath include CaO, Al_2O_3, MgO and SiO_2, all of which have similar mean atomic weights. The mantle can be stratified with small changes of density that cannot be resolved from seismology.

The present argument for whole mantle convection is primarily a single tomographic cross-section which has been claimed to indicate a slab traversing the mantle from top to bottom although, ironically, it looks more like cross-sections of strongly stratified laboratory flows.

The early arguments for whole mantle convection were based on the sizes of plates, thought to correspond to the depth of convection, the presumed high coefficient of thermal expansion of lower mantle minerals and the evidence against variations in viscosity and FeO-content of the mantle (Olson et al). With a better understanding of the interaction of plates with convection, and of the effect of pressure on physical properties and intrinsic density variations at constant FeO-content the above arguments are no longer valid. The Boussinesq approximation, widely used in geodynamic modeling, ignores the effect of compressibility and, hence, pressure dependent á and $K_{L,}$ which turns out to be the essence of the question regarding stratified mantle convection. Minor density differences between layers, 1 to 2%, can irreversibly stratify the mantle. In the Boussinesq approximation density differences of 6% can be overcome by thermal buoyancy. The Earth is gravitationally stratified during accretion and minor differences in MgO, Al_2O_3, and SiO_2 content, for example, of the various layered can prevent mixing of these layers.

Much of our intuition regarding convection in the mantle are based on calculations and experiments that ignore the effects of pressure, compression and the surface plates. Viscous dissipation, absent in the incompressible approximation, can be comparable, locally, to rediactive heating (Yuen) and can facilitate "thermal coupling" in stratified mantle flow (Nataf). Transformational superplasticity due to recrystallization may also introduce low-viscosity regimes at 650 km and 1000 km.

The arguments against chemical stratification of the mantle are not general but reger to very specific models. Models with chemical boundaries below 670 km, or with low radioactivity in the deep mantle or with low viscosity or thermal coupling or with large viscosity contrasts between layers cannot be ruled out. If convection in the mantle is stratified we know that the layers are not entirely mechanically coupled, that the boundaries must be deep and that the deeper layers must be low in heat producing elements.

Some tomographic cross-sections apparently show slabs penetrating from shallow depths to the core-mantle boundary and it has been argued that these support whole-mantle convection ( ). This view has been widely quoted and adopted by non-specialists, particularly in geochemistry ( ). These images actually are not representative and they look more like simulation of layered mantle convection than whole mantle convection ( ). Visual inspection of a few color-saturated cross-sections cannot replace quantatative analyse of the tomographic results, which support the concept of barriers in the mantle and stratified flow. Most tomographic cross-sections show slab flattening near 650 km, eith possible local protrusions to 1000 km, and little suggestion of continuous hot upwellings from the core-mantle boundary. Simulation of layered flow also show some regions where there is apparent continuity across the mantle. Deep slab penetration and whole mantle convection is inconsistent with the highly siderophile content of mantle rocks (Aland et al, 2000) and the density-depth relations in slabs (Anderson, Kesson et al).

The conclusion that whole mantle convection prevails in the Earth is not robust. It is based on the assumptions that the upper and lower layers are mechanically coupled (Hager), that the deeper layer is highly radioactive (Davies), that the main chemical interface is at 650 km, and that a few selected whole mantle cross-sections are representative and can be interpreted visually (Van der Hildst).

As a matter of fact superposed fluids of different viscosities become mechanically decoupled and exhibit "thermal" coupling (i.e. downwellings under downwellings) (Nataf). Thus, there appear to be continuous cold downwellings from top to bottom, in some places, but there is a change in planform and spectra across the boundary, as is evidence in global tomogrphy ( ). Mantle tomogrphy, including the apparently continuous (but rare) blue (Van der Hildst) and red (Bjawaard) features resemble layered mantle convection experiments (Nataf, Cz) more than whole layer convection experiments.

Wen and Anderson ( ) showed that any chemical interface in the mantle should be deeper than about 900 km. Such a model explains the geoid, dynamic topography and plate motions. Anderson ( ) showed that the lower mantle was depleted in radioactive elements. Most of these are in the crust and upper mantle.

Although the geophysical evidence favors a stratified mantle, this is probably not relevant to current mantle geochemistry. Recycling, and a heterogeneous upper mantle can explain the chemistry of midocean ridge and ocean island basalts, and the evolution of their reservoirs (Anderson, Allègre, Turcotte, Smith, Lewis). The important "layers" in mantle geochemistry are all small and shallow (crust, lithosphere, perisphere, sediments, mantle wedge, asthenosphere) and the important processes are recycling, contamination, dehydration, and sampling.

Convection in a homogeneous fluid heated from below with no radioactivity or pressure dependent properties is certainly easier to compute than a layered system with internal heating and effects of compression taken into account but this does not endow the calculation with the virtues of Occam's razor.

Conjecture

The mantle below 1000 km. depth contributes nothing to the surface of the Earth (except gravity and heat). Upper mantle convection is passive, following the dictates of the plates and, to some extent, lateral heat flow variations in the high-viscosity lower mantle. Lower mantle convection is sluggish because of high viscosity, high thermal conductivity, low coefficient of thermal expansion and, possibly, low radioactivity and isolation from subduction cooling.

The assumption that the lower mantle is primordial and undeqassed is, after all, just an assumption. The assumptions that it is both isolated and accessible are incompatible. These assumptions, in fact, are responsible for most of the paradoxes that plague mantle geochemistry. All predictions of this hypothesis have been shown to be false but auxiliary hypotheses have proliferated to rationalize the unexpected, surprising, enigmatic, counter-intuitive results which characterize research based on this paradigm.

Unification

The Standard Model assigns a distinct and important role to both the upper and lower thermal boundaries. The lower (hot) TBL is assumed to be responsible for regions of excess magmatism and elevation along the global ridge system, for continental breakup and for volcanic islands and plateaus, continental flood basalts and anomalous geochemistry. Key concepts in the standard model are temperature, homogeneity and mixing. The key concepts being discussed in this paper are lithospheric architecture and stress, sampling, differentiation, inhomogeneity and passive convection. Thus, there is little overlap or communication between these hypothesis.

Once the assumptions of isothermal and homogeneous upper mantle are dropped we recognize that the range of temperatures and chemistries exhibited by mantle basalts are available in the upper mantle. Lithospheric architecture and spreading introduce various scales and styles of small-scale convection including shallow plume-like structures. Once the steady-state or uniformitarianism assumption is dropped we can recognize a temporal and transient component in the life cycle of a ridge with early products representing either a shallow or easily removed enriched component that is lost or overwhelmed at more mature stages of rifting.

The outer shell is not just a thermal boundary layer. It is, in part, the accumulated buoyant residue of mantle differentiation including the on-going process of seafloor spreading. This part can be treated as isolated from the low-viscosity fertile interior but available to contaminate initial magmas at new ridges—rifts.

The range of normal mantle temperatures encompasses the solidus of candidate mantle materials but melts pond at density strength and stress barriers until the conditions for diking come into play.

The plate is not *rigid* or *uniform*, and the underlying mantle is not isothermal. Flow up ridges and at new rifts is intrinsically 3D and gives insolated regions of high elevation, heat-flow and magma productivity.

Earthquakes and volcanoes not only mark plate boundaries but they anticipate new plate boundaries, changes in boundary conditions and dying plate boundaries.

Although the lower mantle, in particular D½, is an important and interesting place, and generates features that appear to rise high into the lower mantle, there is no evidence that it communicates with the surface except through Newton's force and, at the interface, Fouier's Law.

If the shallow mantle is partially molten, volcanoes have a simple cause, stress. Mantle convection, triggered from above, can rapidly flux material through the melting zone at edges, corners and slots, of which there is an abundance in the outer shell.

Discussion

Standard models of geodynamics often assume uniformitarianism, uniform rigid plates, isothermal subsolidus mantle, and a well-mixed homogenous asthenosphere. Plates, plate boundaries and plate motions are often imposed, rather than resulting from the calculation. In such models midplate volcanoes, island chains, large igneous provinces and the chemistry of mantle basalts are unexplained. An active lower thermal boundary, powered by core heat, and independent of plate tectonics and upper mantle convection, is invoked to explain hotspot magmatism and chemistry.

However, plate tectonics is intrinsically episodic, with constantly changing plate boundaries and evolving plates. Convection is driven by horizontal temperature differences which for mantle Rayleigh numbers amount to ±200 °C ( ). Lithospheric architecture provides these gradients at the top of the mantle. The base of the mantle is bounded by an isothermal core, so temperature fluctuations at this boundary are stochastic and ephemeral.

Large plates insulate the mantle and isolate it from subduction cooling. These processes generate transient excess temperatures of about 100 °C ( ). This insulation/isolation does not occur for normal Rayleigh-Bénard convection with uniform boundary conditions and boundary layers with no intrinsic buoyancy or strength. Normal convection and normal plate tectonics cause “excesses” in mantle temperatures, some of which are transient, that cover the entire range inferred from petrology and geophysics (,). With these, normal, fluctuations of temperature, the upper mantle cannot be entirely subsolidus. The buoyancy of magma can fracture the overlying plate where the stress state is horizontal tension.

The topography at the bottom of the plate drives large-scale flow, as from continents to oceans or from cratons to ridges, and small-scale eddies and gyres at edges of cratons, continental margins, triple junctions and developing rifts. This kind of flow is episodic and is most vigorous for a new narrow ocean when hot upwelling asthenosphere is adjacent to deep cratonic lithosphere. Dynamic melting, excess magmatism due to advection of mantle through the “melting zone”, can account for large igneous provinces and their transient nature.

Summary

The physical properties and heat flow change dramatically with depth in the mantle. This results in a gross asymmetry between the upper and lower thermal boundary layers. The upper mantle is characterized by narrow dense downwellings and broad passive upwelling. For the thermal convection component, and various scales of small-scale convection driven by lithospheric architecture.

The lower thermal boundary layer is characterized by giant “megaplumes” which rise slowly because of high-viscosity and low expansivity in the deep mantle. These megaplumes cannot cross even modest density barriers in the mantle. There may be small-scale convection inside of the megaplumes. Although megaplumes affect the geoid,surface topography and lithospheric stress, and differentially heat the upper manlte by conduction, there is no evidence that they rise to the surface.

Episodicity in plate tectonics is unlikely to be the result of massive mantle overturns. Episodicity is an expected result of plate tectonics on a sphere and plate dynamics in a far-from-equilibrium system. From all indication the convective pattern in the lower mantle has long-term stability (Tarduno).

As plates move over deep mantle upwellings they are uplifted and experience tensile membrane stresses. This is a generalization of the equatorial bulge hypothesis of Turcotte and Oxburgh ( ). Stresses can be very large.

Volcanoes and dikes are recorders of the stress-state of the lithosphere. The stress-dike hypothesis explains bilateral volcanic chains, repeated volcanism at the same place (lithospheric coordinates), non-parallel island chains and hotlines. The controlling factor is the stress state of the plates, not the geometry of hot regions in the underlying mantle.

Since plates cannot change their motions rapidly but local stresses can it is suggested that the apparently chaotic motion of volcanoes on the Pacific plate is actually due to local changes in the stress state. The orientations of volcanic chains are a combination of lithospheric fabric and the nature of the stress field.

Plates are not completely rigid. They are held together by external and boundary forces. When the local or regional stresses satisfy the failure or diking condition then rifts and/or volcanic chains ensue. Thermal contraction and local volcanic loads contribute to conditions that allow ascent of magma. The fabric of the plate– fracture zones, transform faults, old plate boundaries modulate the far-field and regional stresses, causing volcanic chains to be on such boundaries and, to a large extent, explaining their parallelism. Numerous studies and conference have been devoted to the subject of ridge-plume interactions. The underlying assumption is that ridges and plumes are separate phenomena, and plumes are the active element.

This approach ignores the three-dimensionality of passive upwelling and the normal temperature fluctuations that occur in the mantle. A planet with internal heating and plate tectonics is characterized by broad passive upwellings. Large scale flow is controlled by craton and slab separations. Small scale, intrinsically 3D, convection occurs at lithospheric discontinuities and plate boundaries. Regions of “excess” elevation and magmatism are a normal part of plate tectonics and topside induced upwelling.

Figure Captions

Figure 1 A tomographic cross-section from the South Pole across Africa to Europe. The shallow parts of the blue regions are old cratons. They cool off the underlying mantle by conduction and deformation. The deep mantle under Africa is characterized by broad low-velocity regions, perhaps buoyant upwellings. The style of convection appears to change at 1000 km (dotted line), the top of Bullen's Region D´ (lower mantle).

Figure 2 A cross-section from Australia to Kamchatka. The dark blue regions are thick cold lithosphere, thickest under continents. The pale blue regions are probably cold downwellings under cratons and subduction zones. The shallow red regions are hot, probably partially molten regions above down-going slabs.

Figure 3 A cross-section from the mid-American trench , across Hawaii, to the old part of the Pacific plate. Green dots are earthquakes Hawaii if offset from a hot part of the upper mantle which extends to 250 km.

Figure 4 A cross-section across Iceland and the Arctic ocean plate boundaries. The Canadian-Greenland and Fennoscandian shields are the blue regions. The Gulf of California ridge system shows up as a shallow red region to the left. Iceland is a more subdued hut region extended to 650 km, possibly related to EDGE driven convection between the cold cratons.

Figure 5 A horizontal step in lithospheric thickness generates convective currents at the edge which fluxed material through the upper mantle melting zone, delivering large amounts of melt to continental edge and craton edge environments.

Figure 6 A tomographic cross-section along the mid-Atlantic ridge system (green line), relative to a global Earth model (PREM) and an oceanic reference model (ORM). The ridge system is characterized by low seismic velocities to about 100 km depth. The ridge opened from the South to the North and the North was covered by insulating continent until 60 million years ago. Low seismic velocities extend to greater depth (650 km) under the far North Atlantic.

Figure 7 Mantle upwellng in a "slot" is intrinsically 3D, with sprout-, or plume-like upwellings peeling off from the tops of deeper ridge-like upwellings.

Figure 8 Systems heated from below and cooled from the top develop a symmetric system of independent downwellings and upwelling. Systems mainly heated from within (radioactivity) have narrow downwellings and broad internal upwellings.

Figure 9 Schematic diagram illustrating the elements of a top-down system, convection driven by lithospheric architecture and volcanism controlled by lithospheric stress.