LITHOSPHERE, ASTHENOSPHERE, AND PERISPHERE

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Abstract. "Lithosphere" is a mechanical concept implying strength and relative permanence. Unfortunately, the term has also been applied to the surface thermal boundary layer (TBL) and a shallow enriched geochemical reservoir, features having nothing to do with strength. The "strong" lithosphere is about one half the thickness of the TBL. The bottom of the elastic plate, the maximum depths of midplate and fracture zone earthquakes, and the top of the low-velocity zone all occur at depths corresponding to the 550°–750°C temperature range. This may also correspond to the base of the coherent plate. Earthquakes in subducted slabs are bounded by isotherms in this temperature interval. Mantle hotter than ~650°C cannot support long-term stresses and does not qualify as lithosphere. There is very little ancient lithosphere (mantle colder than 650 ± 100°C for long periods of time), and this is not a suitable reservoir for continental flood basalts (CFB). A chemical characteristic, that is, "enriched," has been attributed to old lithosphere to distinguish it from the "depleted" upper mantle. The continental lithosphere (CL) is often treated as a viable reservoir for CFB or, when delaminated, for ocean island basalts and enriched mid-ocean ridge basalts (MORB). The CFB reservoir is more likely to be a hot, weak sublithospheric layer which may include the lower part of the TBL. The term "perisphere" has been introduced to accommodate the need for a term for a global, shallow, enriched reservoir or boundary layer. It is physically isolated from the depleted mantle (usually called the "convecting mantle," "asthenosphere" or "upper mantle") not by its strength but by its weakness and buoyancy; it is chemically isolated by its location relative to subduction recycling. It has the chemical characteristics often attributed to continental lithosphere (or plume heads), but it is a permanent part of the sublithospheric shallow mantle and is constantly refreshed by recycling. It is an open system and can also be called the "active layer" or "mixing zone." It is proposed that the depleted reservoir (MORB source or depleted mantle) is below and protected from contamination (chemical isolation) by the filtering action of this boundary layer. Depleted MORB is most prominent at fast spreading ridges, which induce or localize deep, broad upwellings. Melts from enriched mantle (EM) are most evident at new or slowly rifting regions, infant subduction zones, new backarc basins, slab windows, and midplate environments away from spreading induced upwelling. EM is therefore probably shallow. It is not known if volatiles and large-ion lithophiles can recycle much deeper than ~200 km, or into the lower mantle, as is implied by some plume theories. The base of the (strong) lithosphere and plate may correspond to a phase change. If so, the correspondence among the brittle-ductile boundary (the maximum depth of earthquakes), the top of the low-velocity zone, and the elastic plate thickness can be understood. Candidate phase changes include dehydration and clinoenstatite to orthoenstatite since these occur near 600°C. Another rheological boundary may set in near the solidus, but silicates lose most of their strength at absolute temperatures about one half the dry solidus temperature. The region of the subcontinental mantle between ~600°C and the solidus may provide some of the material in continental magmas, but this cannot be considered part of the continental plate. The "continental lithosphere" reservoir of petrologists is actually a weak enriched layer that may spread across the top of the convecting mantle. This is the perisphere. Its existence makes it possible to understand CFB and ocean island chemistry and kinematics without postulating plumes from the lower mantle, plume heads, fossil plume heads, or delaminated CL. The upper mantle is inhomogeneous in chemistry.

INTRODUCTION

Most geologists and geophysicists would define the lithosphere as the "strong outer shell of the Earth." *Barrell* [1914] introduced the idea of a strong outer layer overlying a weak asthenosphere that could flow to maintain isostatic compensation. *Daly* [1940] built on *Barrell*'s [1914] work. The concept has now grown fairly sophisticated, and we know that the thickness of the lithosphere should vary with temperature, stress, strain rate, age, and mineralogy. The thickness of the lithosphere is inferred from topographic and gravity measurements of the flexure near islands, trenches, and mountain belts [*McNutt and Menard*, 1982; *Watts*

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et al., 1980, 1988]. The result is often called the elastic, flexural, or mechanical lithosphere. The effective elastic thickness (EET) of the lithosphere depends on the magnitude of the load and the time since loading [*Anderson and Minster*, 1979]. Lithosphere carries with it a connotation of coldness, high viscosity, permanence, resistance to flow, and a relationship to the overlying crust. A variety of evidence shows that stresses of 100–200 MPa can be supported by the lithosphere [*Wiens and Stein*, 1983; *Zoback*, 1992]. The asthenosphere will flow readily at much lower stress levels. Asthenospheric viscosities are of the order of 10^{21} to 10^{22} Pa s. The lithosphere is generally assumed to have viscosities 2 or 3 orders of magnitude higher. According to *Barrell* [1914], the lithosphere is 100 times stronger than the asthenosphere.

Unfortunately, the term lithosphere has recently been applied to many other concepts. The term is now in use with widely different meanings and implications. For example, the special lithosphere volume of the *Journal of Petrology* [*White*, 1988, p. 1] provided what was called a "natural" definition of the lithosphere: "the outer shell of the Earth where there is a conductive temperature gradient, overlying the well mixed adiabatic interior." This is actually the thermal boundary layer (TBL) which depends on heat flow and conductivity and whose measurement from bathymetry depends on thermal expansion. *McKenzie and Bickle* [1988, p. 625] define lithosphere as "the depth to a constant isotherm." There is no connotation of strength or permanence in these definitions or measurements, and, in fact, the lower portion of this thermal "lithosphere" flows readily. The potential temperature that is commonly used to define the base of the TBL is ~1280°C [*McKenzie and Bickle*, 1988], while the base of the elastic, or strong, lithosphere is closer to 550°–600°C [*De Rito et al.*, 1986; *McNutt*, 1984] or about half the absolute melting temperature. A useful rule of thumb is that the strong, high-viscosity plate is about half the TBL thickness and about half the depth to the dry solidus. Unfortunately, some Earth scientists have confused the concepts of the strong lithosphere and the TBL. *Menzies* [1990] for example, adopts 1280°C as the boundary between the rigid plate and the weak asthenosphere. *Arndt and Christensen* [1992] take 1210°C as the lithosphere-asthenosphere boundary, which they assume to mark the strong-to-weak transition. However, mantle rocks, particularly wet rocks, have little strength above ~650°C. Hence there is much more than harmless semantics involved in lithospheric issues. Lithosphere has become an unnecessarily confusing concept. The various subtleties in defining rheological lithospheres are given by *McNutt* [1984] and *Minster and Anderson* [1981]. In general, we expect strength and viscosity (and seismic velocity) to increase with pressure and to decrease with temperature and water content.

Geochemists and petrologists have given chemical connotations to both the lithosphere and asthenosphere. Thus they speak of a "continental lithosphere signature" in continental and ocean island basalts, and of an "asthenospheric signature" when referring to mid-ocean ridge and other depleted basalts. Their "continental lithosphere" (CL) (actually, the TBL) is viewed as an isolated reservoir that can remain separate from the "convecting" mantle for more than 10^9 years and that can be remobilized or melted to provide continental flood basalts (CFB). Sometimes a depleted (low content of incompatible elements, $^{87}Sr/^{86}Sr$ and so on) signature in ocean island basalts is attributed to melting of the "depleted oceanic lithosphere." Usually, however, a lithosphere signature refers to evolved isotopic ratios and island-arc type chemistry; high large ion lithophile (LIL) contents and LIL/high field strength elements (HFSE) ratios (such as Ta, Hf, Nb). Such a signature can be acquired by mantle that has been infiltrated with slab-derived fluids.

The asthenosphere (weak layer) has taken on the connotation of a homogeneous, well-mixed, convecting, as well as depleted, layer. From a physical point of view the asthenosphere can be enriched or depleted, homogeneous or inhomogeneous, and even stratified. A metasomatized or hydrated region is more likely to be weak than a dry depleted region. Asthenosphere is often used interchangeably with upper mantle, convecting mantle, depleted mantle, and the mid-ocean ridge basalt (MORB) source. *Schilling* [1973a] refers to the seismic low-velocity zone (LVZ) as the "depleted LV," thereby assigning trace element chemical attributes to a seismic zone of the mantle. Ironically, the LVZ is most pronounced under hotspots and continental rifts, regions that provide trace element enriched basalts. The concept of a depleted asthenosphere is now firmly embedded in the geochemical literature. The chemical connotations have preempted the original physical characteristics of lithosphere and asthenosphere. This considerably complicates interdisciplinary communication. Geochemistry provides little information regarding the location or physical state of reservoirs.

GLOSSARY

Asthenosphere: originally, "the weak layer" of the upper mantle. More recent usage, which is not recommended, equates it with the mid-ocean ridge basalt (MORB) source, depleted mantle, convecting mantle and, sometimes, the whole upper mantle.

Backarc basin basalts (BABB): occur above the slab in backarc basins; generally between MORB and hotspot or island arc basalts in chemistry.

Continental flood basalts (CFB): vast outpourings of basalt (~10^6 km^3) that occur on continents, generally next to thick cratonic lithosphere; usually associated with preexisting lithospheric shear zones or sutures.

Chondritic: having the composition of chondritic meteorites, which are thought to be the lowest-temperature, least fractionated meteorites and close to the original condensable fraction of the solar system. The refractory elements in the primitive Earth are thought to occur in chondritic proportions. Some geochemists have argued that there still exists a primitive, undifferentiated chondritic reservoir in the mantle, but strong evidence is lacking.

Continental crust (CC): crust of continents as opposed to oceanic crust.

Continental lithosphere (CL): that part of the continental crust and upper mantle that can support long-term geological loads. This term, as generally used, refers to the thermal boundary layer (TBL) beneath continents and not to the rheologically strong lithosphere, which is a long-lived part of the plate. However, this nonrheological definition is not recommended.

CFB reservoir (CL*): TBL under continents; contaminated layer; perisphere. This is what petrologists mean when they refer to "continental lithosphere."

Core-mantle boundary (CMB): the region between the core and the mantle, usually treated as a sharp discontinuity.

Convecting mantle: a term that geochemists use for DM, the MORB reservoir. Usually equated with asthenosphere and upper mantle. As often used, implies well-stirred and homogeneous. The term has little meaning as generally used in the nonfluid dynamics literature.

Depleted: deficient in the incompatible or mobile trace elements that are extracted in melts or fluids; low in $^{87}Sr/^{86}Sr$, $^{144}Nd/^{143}Nd$, $^{206}Pb/^{204}Pb$, and so on. It should not be confused with infertile, which means deficient in the basalt-forming major elements (Ca, Al, Na, Ti). The processes of IAB and BABB generation and removal lead to depletion of the remaining mantle in LIL.

Depleted mantle (DM): the MORB source. The asthenosphere and the upper mantle are not recommended definitions.

Enriched: the opposite of depleted. Sometimes the reference state is MORB and sometimes bulk silicate Earth or chondrites. Some OIB and CFB come from enriched mantle (EM). "Enriched" refers to trace element and isotopic characteristics. "Fertile" refers to major elements such as Ca, Al, and Na. The MORB source is fertile (can provide basalt by partial melting), yet depleted (these basalts are low in Rb, La, Cs, U, $^{87}Sr/^{86}Sr$, $^{206}Pb/^{204}Pb$, etc.)

Effective elastic thickness (EET): thickness of the uniform beam or plate, which duplicates the flexural shape of the lithosphere when deflected by a geological load.

Enriched mantle (EM): for example, high in LIL, $^{87}Sr/^{86}Sr$ etc., usually relative to MORB.

High-velocity anomaly (HVA): seismic velocities that are higher than average for that depth.

Island arc basalts (IAB): these may be the result of slab-derived fluids fluxing the mantle wedge, the first step in depleting the slab and enriching the shallow mantle or perisphere.

Komatiite: high MgO magmas which are thought to be the highest-temperature terrestrial magmas. As such they should correspond to plumes that are thought to be hotter than normal mantle. Komatiites are often depleted, having isotopic characteristics similar to MORB. Komatiites may evolve to basalts by cooling and crystallizing in the shallow mantle. Komatiites are common in Archean terranes but are rare in modern times.

Lamproite: a class of highly incompatible element-enriched magmas, including kimberlites, which may represent initial small fraction melts, final residual melts, or metasomatic fluids. If MORB represent the depleted extreme of mantle magmas, lamproites represent the enriched extreme. Most magmas may be blends of these or similar endmembers.

LID: the seismic high-velocity region at the top of the mantle. It generally overlays a low-velocity zone, which starts at an average depth of 40 km under oceans and starts as deep as 150 km under cratons.

Large ion lithophile (LIL): DM is depleted and EM is enriched in LIL, which are also called incompatible elements.

Large igneous province (LIP): such as CFB and oceanic plateau provinces; attributed, by some, to plume heads, and by others to the initial magmatic transients resulting from plate extension or lithospheric pull apart.

Lithosphere: the strong outer shell of the Earth. More recently, and not recommended, the enriched layer, contaminated layer (CL*) and the TBL.

Low-velocity anomaly (LVA): seismic velocities that are lower than average for that depth.

Low-velocity zone: a depth interval with seismic velocities lower than the region above or a region of decreasing velocity with depth.

Mantle wedge: the mantle above the slab and below the plate; the site of most mantle recycling.

Metasomatized: enriched in trace elements by fluids that migrate from one place to another. The source of OIB often appears to have been enriched in incompatible elements just prior to eruption. Alternatively, the parent magma of OIB may pick up LIL as it traverses the shallow metasomatized mantle. Subducting slabs and sediments may be the source of these fluids.

Mid-ocean ridge basalts (MORB): most abundant and most depleted magma type. Generally attributed to passive spreading and a shallow (DM) depleted asthenospheric source. The MORB source may be deeper and therefore protected from recycling and slab dehydration by a CL* or perisphere.

Ocean island basalts (OIB): basalts from ocean island that are thought to be due to hotspots.

Perisphere: weak, shallow EM. Source and sink of elements most prone to recycling; probably forms at mantle wedge. May be enriched by residual melts and metasomatizing fluids.

Picrite: a relatively rare high MgO magma; an olivine-rich rock. Picrites probably result from high temperature and extensive melting. Picrites are often depleted and are often attributed to plume stem melts. They may evolve to more common basalt types by crystal fractionation. Depleted picrites often occur at the base of CFB sequences.

Plate: the outer shell of the Earth, which moves coherently and behaves elastically. It may correspond to lithosphere plus the superposed material carried along (see komatiite).

Plume: a hypothetic entity that is supposed to exist independently of mantle convection and plate tectonics and is a narrow (100–200 km radius) hot column that extends to a TBL at the base of a convecting system, usually the CMB. Plumes are assumed to provide magma to hotspots such as ocean islands. Generally considered to be strong, active upwellings in contrast to passive upwellings caused by plate divergence and in contrast to normal large-scale upwellings that exist in any convecting system.

Plume head: a 1000 to 2000 km (radius) pancake-shaped object that forms as plumes approach the upper mantle. Plume heads are assumed by some authors to be responsible for surface uplift, breaking of the lithosphere and LIPs. An alternative to plume heads are large-scale regions of the upper mantle that are hot because they have not been cooled by subduction. There may also exist large-scale regions of the upper mantle that are enriched because of recycling.

Pull-apart: the reverse of suturing. A process that supposes that lithospheric extension is concentrated at boundaries such as fracture zones and sutures, particularly those involving contrasts in strength or thickness. Since cratons involve thick lithosphere, a craton edge is a particularly pronounced discontinuity and a likely site for lithospheric pull-apart, as well as small-scale mantle convection. Pull-apart is an alternate to distributed stretching or thinning.

Siderophile: iron loving. The trace siderophile elements include Os, Ir, and Re, which are among the most depleted elements in the mantle, presumably because they have been removed to the core.

Thermal boundary layer (TBL): occurs at the top of a convecting system and also at the bottom of systems that are heated from below; also called the conductive boundary layer. Recently, it has erroneously been equated to the lithosphere. Thus some petrologists refer to "the continental lithosphere (CL) source" when they mean TBL or upper mantle. A conductive boundary layer (TBL) may be weak and can suffer shearing deformation in its lower portions even while vertical thermal transport is predominantly by conduction rather than advection. Roughly, the top one half of the surface boundary layer has long-term strength.

LITHOSPHERES AND SEMANTICS

The various lithospheres in the literature include the seismic high-velocity layer (LID), the elastic shell, the mechanical boundary layer, the brittle zone, the plate, the slab, and the TBL or thermal lithosphere. Except for the TBL all of these involve strength, viscosity, or elasticity to some extent. Insofar as the physical properties in the TBL depend only on temperature, the thicknesses of these various lithospheres might be expected to covary with each other and with lithospheric age. Actually, stress, strain rate, thermal gradients, mineralogy, composition, phase changes, time since loading, duration of loading, past history, and so on are also involved [*Minster and Anderson*, 1981]. If the thicknesses of the various lithospheres (excluding TBL) agree, then this would suggest that changes in chemistry or mineralogy, or phase changes, including dehydration, might control the transition from strong to weak. If the various transitions depend mainly on temperature then they might approximately correspond, in depth, in a region of high thermal gradient, such as in the TBL.

A TBL exists at the top of a convecting system. The TBL thickens with time until it becomes unstable and falls off. In a homogeneous fluid the distance between downwellings is approximately twice the convecting layer depth. If part of the TBL is very strong or has high viscosity or has some chemical buoyancy it may remain at the surface for longer periods of time than in simpler systems. On the other hand, the TBL may be thickened by plate convergence, or impacted from below by a plume, before it falls off by its own weight. The Earth has all of these complications, but few convection calculations do. In a differentiated planet, buoyant material collects at the surface and part or all of the TBL may be composed of material different than the underlying convecting mantle. In fact, the mantle immediately beneath the lithosphere may also differ from the deeper mantle. Recycling, melt removal and trapping, dehydration, metasomatism and so on, affect the chemistry of the shallow mantle. A thick cold TBL is not necessarily denser than the underlying material, and part of it may remain at the top of the system. This possibility is ignored in lithospheric delamination models of mantle chemistry [*Allègre and Turcotte*, 1985; *McKenzie and O'Nions*, 1983], which treat the mantle as a homogeneous fluid. The TBL should not be referred to as the lithosphere or the plate. TBLs in homogeneous convecting systems are not permanent surface features nor are they necessarily strong. This should be kept in mind when reading petrological papers that refer to the CL reservoir. Since CL, as generally used, is not lithosphere, the enriched reservoir might be designated CL* and would mean "contaminated layer" or CFB reservoir. Since this is also confusing, I eventually propose a new name, "perisphere" for CL*.

The crust is generally weaker than the mantle, particularly the hot crust that occurs below some 10 km, but it is often treated as part of the lithosphere. It is certainly part of the plate, but it is often only able to carry a fraction of the load supported by the lithosphere. Sometimes "crust" is used where lithosphere is meant. Interpretations of crustal deformation (stretching, thinning, shearing, thickening) sometimes ignore the underlying mantle. In high heat flow regions, or in regions metasomatized by slab-derived fluids, the shallow mantle may be weaker than part of the crust. In such regions the crust may be the main load-bearing element and can even transmit stresses for intercontinental distances [*Zoback et al.*, 1993]. The strength of the crust depends on composition, volatile content, fluid pressure, and stress, as well as temperature. Lithosphere is often used to refer to the "crust plus the strong part of the upper mantle." Sometimes magmas or xenoliths are referred to as coming from the continental lithosphere without any evidence that the responsible mantle is cold, strong, or rigidly attached to the overlying crust. Thus petrologists often use lithosphere when they mean "mantle."

What is often wanted is the thickness of the coherent plate, that is, the plate of plate tectonics. The stresses responsible for plate tectonics involve different stress levels and orientations, and timescales, than are available from geophysical measurements, and the thickness of the plate cannot be determined directly. However, the boundary between the plate and the convecting mantle may be at a similar depth as the lithosphere-asthenosphere boundary, at least, over the long term. In most discussions, however, the plate (almost always referred to as the lithosphere in the petrological and geochemical literature) is usually assumed to be the lithosphere plus the thermal boundary layer. The TBL is often referred to as the lithosphere in the mantle convection literature. The lower part of the TBL is sometimes referred to as the "ductile" or "plastic" lithosphere, that is, weak. A moving plate can entrain the hot underlying mantle, but the latter is not a permanent part of the plate.

Seismic Lithosphere (LID)

The LVZ of the upper mantle is sometimes equated with the asthenosphere. The high-velocity LID would then correspond to the lithosphere. The existence of a high-velocity LID, however, implies a different mineralogy, physical state, or chemistry than the underlying mantle (see below). A minor component (hydrous phases, vapor, melt) can reduce seismic velocities appreciably [*Anderson and Sammis*, 1970; *Ito*, 1990; *Lebedev et al.*, 1991]. The LID thickness is often used as a proxy for the lithosphere thickness, but this needs to be justified.

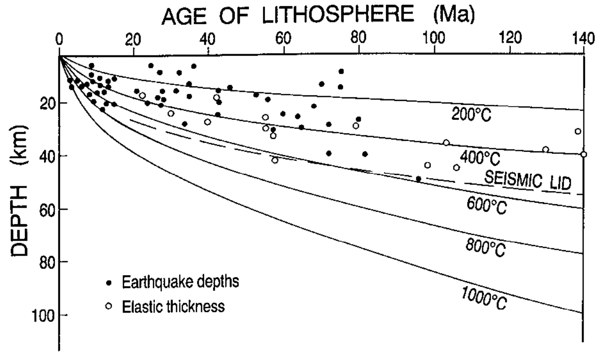

Figure 1. Isotherms versus age of the oceanic thermal boundary layer (TBL). The seismic high-velocity layer (LID) is shown as a dashed line, close to the 600°C isotherm [*Anderson*, 1989; *Regan and Anderson*, 1984]. The depths of earthquakes are solid dots [*Wiens and Stein*, 1983; *Chen and Molnar*, 1983]. Effective elastic thicknesses are shown as open circles [*Watts et al.*, 1980, 1988]. Note that all of these measures of plate rheology indicate that 550°–650°C is an important cutoff temperature. Seismicity cutoffs and high seismic velocities may both define the thickness of the plate. At temperatures higher than ~600°C the mantle may be dragged or entrained by the plate, but it is not a permanent part of the lithosphere. The bottom of the TBL is much deeper than the 600°C isotherm. The thickness of the LID is different if seismic isotropy is assumed. All thicknesses are model dependent.

Seismic velocities increase with pressure and decrease with temperature. The high thermal gradient near the surface of the Earth (TBL) assures that in a homogeneous mantle velocities will decrease gradually with depth down to ~200 km where the pressure takes over [*Anderson and Sammis*, 1969]. Actual seismic velocity profiles show an abrupt decrease in shear velocity at depths ranging from ~10 to 20 km under young oceans to ~150–200 km under cratons. This implies a change in phase, such as dehydration or partial melting, a change in composition or mineralogy, an increase in stress or dislocation density, or a very rapid increase in temperature gradient. Figure 1 shows that the depth to the bottom of the LID corresponds approximately to the 600°C isotherm [*Anderson and Regan*, 1983]. This also applies to the thicker shield LID [*Anderson and Bass*, 1984]. The high-velocity layer under shields extends to ~150 or 200 km [*Anderson*, 1967, 1990; *Grand and Helmberger*, 1984]. Heat flow calculations suggest that in low heat flow areas of cratons the temperature at 150 km depth may be as low as 750°C [*Lesquer and Vasseur*, 1992]. If the seismic and elastic lithospheres coincide in thickness, this suggests a phase change or compositional explanation [*Anderson and Regan*, 1983; *Regan and Anderson*, 1984]. If these depths, in turn, follow an isotherm or an equilibrium phase boundary, one would suspect a phase change such as dehydration. One expects both a weakening and a lowering of seismic velocity for such phase changes [*Lebedev et al.*, 1991; *Raleigh and Paterson*, 1965; *Regan and Anderson*, 1984]. A variety of natural rocks exhibits a substantial reduction in acoustic velocities near 650°C [*Ito*, 1990; *Lebedev et al.*, 1991]. Figure 1 suggests that the common practice of using seismic velocities to measure lithospheric thickness is valid. This in turn implies that the reduction in strength may be abrupt, rather than thermally activated, and may be due to a change in chemistry, mineralogy, or hydration state.

Very high seismic velocities extend to depths greater than 300 km under some cratons [*Anderson and Toksöz*, 1963; *Nataf et al.*, 1986; *Anderson et al.*, 1992; *Jordan*, 1975]. I define the LID, however, as the region above an abrupt decrease of seismic velocity, which usually occurs shallower than 200 km. Whether the actual plate under cratons extends deeper is a matter of current debate. There is a good correlation between Archean cratons and high velocity to depths greater than 200 km but little correlation for younger cratons and mobile belts [*Polet and Anderson*, 1995]. The high-velocity region under the oldest cratons could be a static keel [*Jordan*, 1975] or a dynamic downwelling. The Archean lithosphere and plate are probably at least 150 km thick. The TBL associated with Archean crust and lithosphere is probably much thicker, but it is not necessarily as old.

Elastic Lithosphere

The lithosphere is often treated as an elastic plate overlying a weak, or fluid, half-space. The thickness can be estimated from deformation profiles caused by surface loads (volcanoes, mountains) or bending stresses (trenches). The EET of oceanic lithosphere increases roughly as the square root of age (Figure 1). The base of the elastic lithosphere corresponds approximately to the 450°–650° isotherm [*Furlong*, 1984; *McNutt*, 1990; *Watts*, 1982, 1988; *Wessel*, 1992, 1993]. Cratonic areas have been cooling much longer than oceanic areas, and it is expected that they will be colder to greater depths. They may also differ, however, from other regions in parameters other than temperature (thickness and composition of crust, composition of mantle lithosphere, radioactivity).

Figure 1 summarizes some of the data obtained from flexural studies on oceanic lithosphere. Note that the maximum EET corresponds approximately to the maximum depths of earthquakes and to the seismic thickness of the oceanic LID (i.e., the depth to the LVZ). The 600°C isotherm seems to define the boundary between strong, elastic, brittle, high-velocity material and weak, ductile, low-velocity material. This temperature differs from previous determinations because it is based on a new thermal model of the oceanic lithosphere [*Denlinger*, 1992]. It is of interest that partially hydrated shallow mantle rocks containing some serpentinite show a marked weakening above 500°–600°C [*Raleigh and Paterson*, 1965]. Very dry mantle rocks may be strong to higher temperatures. The Archean plate may be dry as well as cold, and its thickness may help isolate the underlying mantle from revolatilization.

Earthquakes

Earthquakes occur in the brittle part of the lithosphere. The brittle-ductile transition depends on temperature, pressure, stress, strain-rate, tension versus compression, mineralogy, and so on. The timescales involved in the earthquake cycle (hundreds to thousands of years) are so different from the timescales of mantle flow and the duration of geological loads that one expects little correspondence between the depth of the seismogenic zone and the thickness of the plate as measured over geological timescales. For example, the thickness of the lithosphere, as measured by the deformation caused by a load, decreases with time, as stresses relax. Stress-dependent rheology makes the thickness of the lithosphere depend on stress levels. Short-term and low-stress experiments, such as wave propagation or deglaciation or small seamount loads, might be expected to yield thicker lithospheres than deformation associated with mountain belts or oceanic islands. As a matter of fact, all of these experiments give about the same thickness (to within 20–30%) and seem to correspond to similarly low isotherms (550°–650°C). Thus 600° ± 50°C appears to be a "magic temperature," which defines the rheology of the upper mantle. Almost all of the estimates of the temperature of a rheological boundary are encompassed by 650° ± 100°C.

For example, the mantle appears to be unable to store the stresses required for brittle failure (earthquakes) if it is hotter than about 650°C [*Chen and Molnar*, 1983; *Wiens and Stein*, 1983]. The recurrence time of large earthquakes in a given tectonic region is of the order of 100–1000 years so the implication is that stresses are relieved by flow, on this timescale, at higher temperatures. The depths of intraplate oceanic earthquakes (Figure 1) are generally less than the inferred 600°C isotherm but occasionally reach the 750° isotherm [*Chen and Molnar*, 1983; *Wiens and Stein*, 1983]. This corresponds approximately to the thickness of the seismic LID [*Regan and Anderson*, 1984], that is, the depth to the top of the LVZ. This in turn corresponds to the effective elastic thickness of the oceanic plate. We expect, however, that the thickness of the lithosphere is time, frequency, and stress dependent. The existence of a single magic temperature for these various phenomena suggests that the base of the lithosphere is defined by a temperature-dependent change in phase or mineralogy. The maximum depths of earthquakes as a function of the age of the oceanic plate are indicated in Figure 1.

Stresses and strain rates are quite different in downgoing slabs than in midplate environments and along fracture zones. Intermediate depth earthquakes provide a test of the magic temperature concept. Some deep earthquake zones exhibit a double-planed structure down to 200 km. The planes are about 40 km apart initially but converge with depth. *Kawakatsu* [1986] showed that most earthquakes in the slab are confined between the 600°C contours, which define the top and bottom of the seismic zone. The cold, wedge-shaped core of the slab is seismically active but less so than the regions near the 600°C isotherms. This is consistent with phase change control on the seismicity since yield strengths in homogeneous solids are strong functions of strain rates and whether the material is under extension or compression. At depths greater than 200 km we expect pressure-dependent phenomena and other phase changes to be involved and perhaps a departure from the low-pressure 550°–650°C "magic isotherm." The magic temperature concept is not inconsistent with phase change control of intermediate and deep-focus earthquakes.

Phase Changes

Phase changes can be accompanied by a major change in rheological and elastic properties and acoustic emission (microearthquakes). For example, as quartz is heated, a major reduction in compressional velocity at the α–β inversion temperature (573°C) is accompanied by massive emission of sound [*Schmidt-Mumm*, 1991]. The compressional velocity drops by ~20% at the transition and then slowly (with a further increase in temperature) recovers [*Kern*, 1990]. Thus here is an example of a change in velocity associated with a change in seismicity. The thickness of a deforming quartzite tectonic plate could be defined by either seismic velocity or seismicity. Similar effects can be expected for dehydration and other phase changes. This is probably too simple an analogy since not all intraplate earthquakes have a maximum depth defined by a critical temperature. However, few earthquakes, anywhere, occur in material with estimated temperatures in excess of 750°C.

The weakening associated with phase changes can be expected to concentrate deformation at phase change boundaries in the mantle. In the upper mantle the most important such phase changes are α–β quartz, clinoenstatite to orthoenstatite, various dehydration reactions, basalt-eclogite, and partial melting (wet or dry). Many of these occur between 500° and 700°C. These transitions will probably cause abrupt changes in strength and seismic velocity, rather than the gradual weakening associated only with an increase in temperature.

The transformation of low clinoenstatite to orthoenstatite takes place at ~600°C and is a candidate for a phase change induced weakening in dry mantle. This reaction is important because it occurs in a major mantle mineral under anhydrous conditions. This phase change does not exist at depths greater than ~220 km, these phases being replaced by high clinoenstatite [*Angel et al.*, 1992]. This depth is about the maximum thickness of the continental lithosphere. It corresponds, in some geographic areas, to a seismic discontinuity. The maximum temperature of brittle behavior of serpentinite, a possibly important phase in the shallow mantle, is ~600°C [*Raleigh and Paterson*, 1965]. There are also a variety of metamorphic reactions near this temperature. Thus there are mechanisms for subsolidus weakening of mantle rocks under both wet and dry conditions and reasons to believe that rocks will deform readily at temperatures well below the melting point. This is well known to metamorphic petrologists. Using the solidus to define the base of the rigid lithosphere or coherent plate is probably not valid. About half of the continental lithosphere, in the sense used by petrologists, is between the 650°C isotherm and the dry solidus. A wet "lithosphere" is partially molten at these temperatures.

The Plate

The plate of plate tectonics may not have the same thickness as any of the lithospheres defined above. For a homogeneous mantle with no phase changes the various lithospheres will certainly have different thicknesses. On the other hand, if a change in composition, mineralogy, or phase controls the rheological properties, then discontinuities in all physical properties may occur at the same depth. Although the solidus of mantle rock defines an important rheological boundary, the mantle may be very weak, on geological timescales, at much lower temperatures.

We do not know the nature of the coupling between the plate and the interior. The plate could be driven by convection currents in the mantle, in which case there is shear coupling through a shear boundary layer. This boundary layer may or may not be considered part of the plate. In the short term the upper part moves with the plate, but in the long term it gets replaced since it is not rigidly attached. By "long term" I mean the hundreds of million to billions of years required to generate significant isotopic anomalies. It is not yet clear whether TBL delamination occurs for the oceanic plate in the time available to move from ridge to trench or to some other boundary where the stresses may be large. The shallowness of seafloor older than 60–70 Ma may be related to TBL instability and removal. The shear boundary layer, and the lower part of the TBL, do not serve the purpose envisaged by geochemists for a CFB reservoir, that is, an ancient part of the continental plate, genetically related to the overlying crust. The geochemists' lithosphere (i.e., the enriched basalt reservoir, or EM) must be isolated from the convecting mantle (the MORB reservoir) for long periods of time so that it can grow the observed isotopic anomalies. Weak, buoyant, or chemically distinct regions of the mantle can also be isolated. The shallow mantle is an open system. Its chemical and physical state may be maintained by slab-derived fluids and quenched magma trails. A chemically modified TBL may remain buoyant, even as it cools, and can remain subjacent, at least temporarily, to the overlying plate even if weak.

Plates may also drive themselves by "slab pull" and by gravitational forces on the thickening plate ("ridge push") [*Hager*, 1978]. Again, there will be a mechanical boundary layer below the plate, but it may be very thin if the viscosity is low or if melts collect at the base of the plate. There is no need to transmit the motion of the interior to the plate. In fact, the plate may, in principle, be decoupled from the interior by a low-viscosity layer. In any case there will be a layer beneath the plate that is flowing in the same direction as the plate by entrainment or shear coupling, but it is presently unknown how thick this layer is. The boundary layer may differ chemically from the permanent plate and may be part of the perisphere.

To some extent, convection in the mantle is influenced, if not controlled, by plate processes [*Hager and O'Connell*, 1981]. Downwelling slabs generate mantle downwellings by entrainment and cooling. Spreading induces upwellings and adiabatic melting, which induces more upwelling. Horizontally moving plates induce some horizontal motion of the subplate mantle. Discontinuities in lithospheric thickness are associated with lateral changes in temperature, and this induces small-scale convection in the mantle. All of these effects are in addition to instabilities in the lower part of the surface TBL that are a normal part of convection and that exist even in the absence of plates. Mantle convection is strongly influenced by plates, slabs, and lithospheric properties. This may include the locations of hotspots.

Although in the short term (less than about 10^8 years) there may be an ambiguity about the thickness of the plate and the thickness and role of the underlying shear coupled layer, there is less ambiguity about the region that is coherent on long timescales required to generate significant isotopic anomalies if, in fact, these are generated in the plate. We suspect that over the long term timescales appropriate for observed isotopic anomalies to grow the plate is equivalent to the layer with long-term strength and therefore is similar in thickness to the elastic, mechanical or rheological, lithosphere. Its lower boundary is probably defined by the $600° \pm 50°C$ isotherm. Cratonic mantle and low heat flow areas may have temperatures this low at depths as great as 150 km [*Anderson and Bass*, 1984]. If the ancient enriched mantle (EM) reservoir is equivalent to the plate, it must be relatively thin. The question of thick cratonic roots is dealt with later.

The Subplate

Isotopic anomalies can also grow in a weak sublithospheric layer that is isolated from the underlying convecting mantle or depleted mantle (DM), and it matters little if this material is periodically entrained by the plates (or boundary layers) or even incorporated into them. A weak, chemically isolated, subplate, sublithospheric enriched layer serves the same purpose as an enriched lithospheric reservoir from a geochemical point of view. The only question is, can a weak layer remain isolated (chemically or physically) from the deeper mantle for the requisite amount of time? The ability to remain physically isolated depends on the density contrast (buoyancy), viscosity contrast (slipperiness), thickness and convective vigor. A weak continental lithosphere (i.e., the TBL) will flow readily and will not remain confined to the subcontinental mantle (Figure 2). There is abundant evidence that EM is available in a variety of tectonic environments and is not, in fact, confined to continental areas or areas that have been attributed to plume heads. EM, be it weak asthenosphere, strong lithosphere, TBL, or perisphere, is probably not an isolated system. It is inhomogeneous, or at least it provides materials that exhibit a spectrum of geochemical characteristics. It has been treated as "isolated from the convecting mantle" because it differs from DM, and differs from it, in some cases, by amounts that imply isotopic isolation for more than 1 Gyr. However, it can be an open, or active, system, subject to refluxing by various fluids and melts. It may not even be ancient if it is refluxed by material of extreme isotopic composition, such as continental sediments. The various boundary layers discussed above may be thin, at any given time. The EM region, which cycles through the boundary layer, is of unknown extent; it may extend to 200–220 km or even to 400 km. Although a shallow EM layer may be highly engaged in mantle flow, it does not necessarily mix with the DM reservoir if the latter is, on average, deeper and denser. The problem of bringing melts from DM (MORB) through a shallow EM layer also exists for CL and plume head hypotheses. DM may be chemically isolated, or protected, from recycling by the shallow boundary layer or perisphere. This would explain its apparent homogeneity as well as its depletion.

The Slab

When the plate leaves the surface, it is called the slab. In simple convecting systems the surface TBL thickens with time, and after a certain time interval becomes unstable and sinks into the interior where it becomes part of the main convective flow. With homogeneous fluids of uniform viscosity the downwelling is only marginally unstable, and it becomes buoyant again after a time interval similar to its surface lifetime. The Earth's plates differ in several important respects from simple TBLs. Because of the low-den-

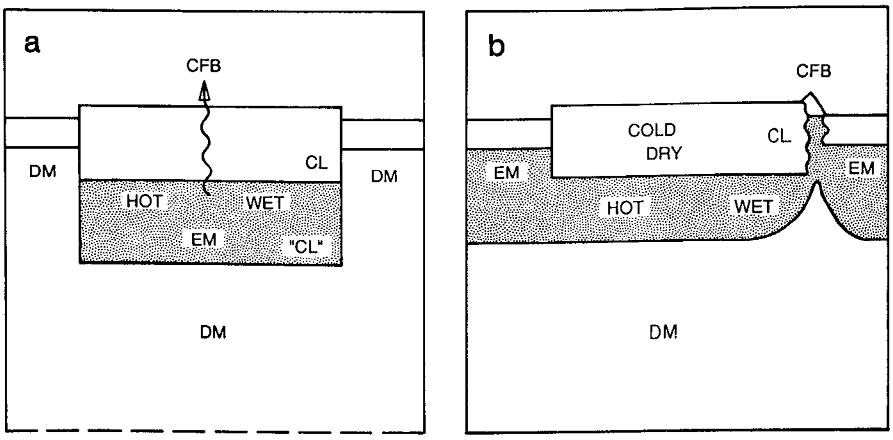

Figure 2. (a) Conventional views of the disposition of depleted mantle (DM) (often called the asthenosphere, upper mantle, convecting mantle, or the mid-ocean rich basalts source) and continental lithosphere (CL). CL is assumed to be enriched (EM) and fertile, and the source of continental flood basalts (CFB) [e.g., *Hawkesworth et al.*, 1987]. (b) The perisphere model. Enriched mantle (EM) is a low-viscosity layer (perisphere) that underlies both continents and oceans, unless thinned or depleted by active or passive spreading. The strong lithosphere extends only to the 600° ± 50°C isotherm. Ultradry mantle may be strong to higher temperatures. This may explain cratonic roots that can extend to >150 km.

sity crust and hydrated upper mantle they are, in part, chemically buoyant and must cool for a longer period of time before becoming unstable. They have strength and high viscosity and cannot spontaneously subduct; they probably need to be pushed down, possibly by an overriding continent, or may require a substantial fault, or even extension, in order to start sinking. Once the sinking process starts phase changes in the slab, including dehydration and basalt eclogite, cause a large nonthermal increase in density. Because of its long residence time at the surface the TBL is thick and strong compared to simpler convective situations. Because it is cold and, possibly, different in chemistry than the surrounding mantle it will undergo phase changes at different depths than ambient mantle. The slab can be thought of as a supercooled TBL rather than one that is just marginally unstable. Because of its high density and viscosity it may sink rapidly to the bottom of the system and may even decouple from the background flow. If so, it will cool the system from below and may prevent active upwellings in the part of the mantle so affected (Figure 3). In fact, ridges and hotspots and the warmer parts of the mantle are mostly confined to regions where there has been little subduction since the breakup of Pangea [*Scrivner and Anderson*, 1992; *Ray and Anderson*, 1994]. In mantle tomographic maps there is a good correlation in the transition region and parts of the lower mantle of high seismic velocities with past positions of subduction zones [*Ray and Anderson*, 1994]. This correlation is better than with ridges or hotspots.

Slabs seem to control the pattern of mantle convection (Figure 4). Since ridges and hotspots, and sites of LIP formation, occur in "uncooled" mantle there may be less distinction between active and passive upwellings than generally assumed. Ridges are mobile and

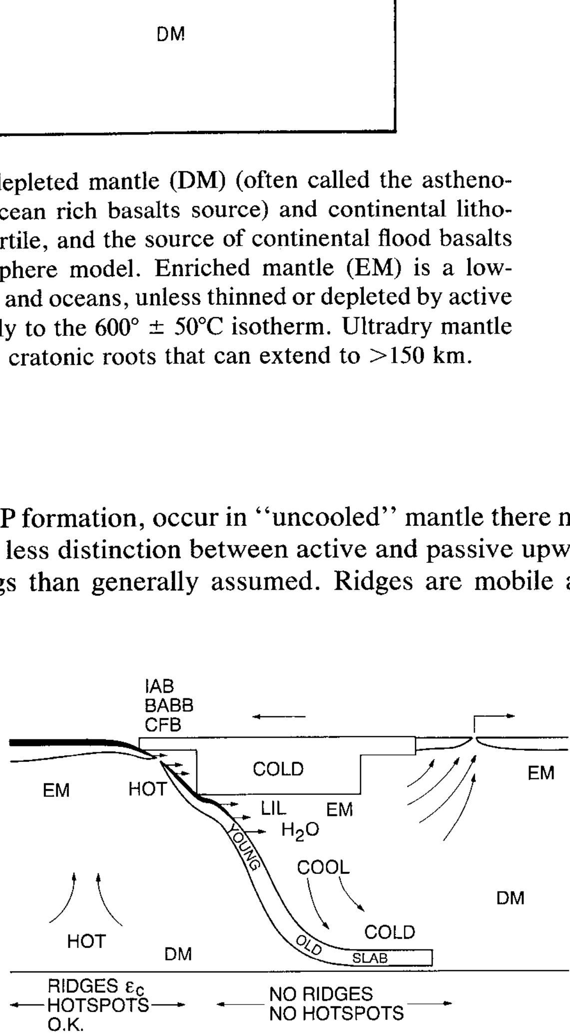

Figure 3. The shallow enriched mantle (EM) model. Subduction introduces sediments and altered oceanic crust into the mantle wedge. This large ion lithophile (LIL), H_2O-rich shallow mantle provides the trace element geochemistry associated with island arc (IAB) basalts, backarc basin basalts (BABB), and continental flood basalts (CFB). A ridge-trench collision provides a slab window and induces extension in the overlying plate and, in some cases, abundant magmatism. The shallow mantle, at a later time, also provides the trace element and isotopic signatures of ocean island basalts (OIB). Depleted basalts occur at mature ridges and other places where the enriched layer has become attenuated or stripped of the low-melting components. Massive magmatism (ridges, hotspots) occur only where the mantle has not been excessively cooled by subduction. EM may be locally absent under regions of divergence or regions of prior melt extraction (e.g., mature ridges).

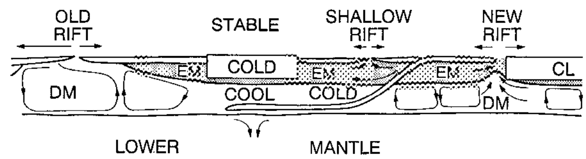

Figure 4. Cartoon illustrating the effects of plates and slabs on mantle convection. EM is generally shallow except at old rifts where it can be removed physically or by melt extraction. EM flows to rifts when they are new or shallow or slowly spreading. Slabs can cool the upper mantle from below and the lower mantle from above. EM is probably formed at subduction zones and depleted by magmatism, including island arc, backarc basin, continental flood, and the initial stages of mid-ocean ridge magmatism.

can migrate readily but they mostly occur over the hotter or "uncooled" parts of the Earth. They may focus the active upwellings. This would be consistent with the proximity of ridges to hotspots. In a convecting system, heated from within, upwellings tend to be broad and convection tends to be unsteady. In a very general sense ridges may be related to these broad upwellings. Just as dead slabs tend to influence the base of the convecting system, the configurations of surface plates, including continents, and boundaries also have a large influence on convection and may impose a long wavelength pattern [*Scrivner and Anderson*, 1992]. A complete up-to-date review of slab matters can be found in the work by *Lay* [1994].

ENRICHED MANTLE

At one time all non-MORB basalts (i.e., "enriched" basalts or basalts from a so-called primordial reservoir) were explained by continental crustal contamination or by "primitive" lower mantle plumes. The upper mantle was regarded as depleted (DM) and the reservoir for MORB, and only MORB, and to have been formed by the removal of the continental crust.

Anderson [1980, 1981a, b, 1982b, c] proposed that most basalts were mixtures of fractionated, or evolved, melts from EM and DM, both reservoirs being in the upper mantle. Various classes of EM have now been proposed (EMI, EMII, HIMU, DUPAL, and so on) [*Hart and Zindler*, 1989]. Some of these represent various kinds of material recycled by subduction (oceanic sediments, continental sediments, oceanic crust, slab-derived fluids), and others may be trapped silicate or carbonatitic melts. "Trapped melts" represent the final residual melt fraction that is difficult to remove because of low porosity, low permeability, neutral density or sealing off of the overlying layers. Even a low-viscosity, mobile region may be inhomogeous. In fact, many theories of petrogenesis invoke a "plum pudding" type structure with low-melting-point inclusions embedded in a more refractory matrix. There are still some advocates of a primitive, undifferentiated, unfractionated, undegassed (chondritic) mantle, usually taken as the lower mantle, but there is little evidence that such a mantle exists [*Anderson*, 1983; *Hart and Zindler*, 1989]. It would be unlikely to survive the high-temperature planetary accretion process or subsequent terrestrial differentiation. The main current argument for the presence of such mantle is the existance of high $^3He/^4He$ ratio basalts, but these may also be tapping recycled ^3He-rich material [*Anderson*, 1993]. High $^3He/^4He$ mantle may reflect low U and Th material, such as harzburgite or dunite, rather than primordial, high ^3He, material.

The physical location of EM has been controversial. In most plume, theories EM is at the core mantle boundary (CMB or D″) or elsewhere in the lower mantle. In the CL theories it is thought to be rigidly attached to ancient continental crust and to be absent in oceanic domains. In other models [*Anderson*, 1982b] it is a global, weak, inhomogeneous sublithospheric layer, formed and maintained by recycling, slab dehydration, and melt trapping (Figure 4). It is enriched in volatiles and LIL, but from a major element point of view it is mainly the refractory and buoyant residue of billions of years of melting and differentiation. It is metasomatized by fluids expelled from the subducted slabs and by residual melts. Under continents it may, in part, be the lower portion of the TBL and therefore the same as the hot and weak "lithosphere" of the geochemists [e.g., *Hawkesworth et al.*, 1988]. Enriched basalts may be mixtures of melts or fluids from EM and from a deeper depleted fertile region.

If EM is at the base of the mantle, next to the molten iron core, it should be depleted in siderophiles and have nonchondritic Os isotopic ratios. As a matter of fact, hotspot magmas are enriched in siderophiles and have near chondritic Os isotopic ratios [*Arndt*, 1991; *Martin*, 1991]. High $^3He/^4He$ ratio basalts are often taken as evidence for a "primitive" lower mantle location for EM. *Anderson* [1993] showed that ^3He-rich interplanetary dust particles (IDP), such as those that are found on the ocean floor, are a plausible source for high $^3He/^4He$ basalts when cycled into the shallow mantle. The "solar" neon of hotspot basalts and xenoliths can be similarly explained. These must represent ancient fluxes, such as comet or meteorite swarms, since the presence IDP flux is too low. On the other hand, the MORB reservoir must have been isolated from the effects of recycling and intramantle fluid transfer for most of the age of the Earth [*Anderson*, 1982a, 1993; *Staudacher and Allégre*, 1988]. It is therefore plausibly below EM but is probably still in the upper mantle. Research on the location of EM has been biased by the widespread assumptions that the asthenosphere is depleted and that the upper mantle can provide only MORB. This has focused attention on CL and deep-mantle plumes as possible EM reservoirs.

Continental Flood Basalts

Continental flood basalts (CFB) occur on continents, generally near preexisting lithospheric discontinuities and at boundaries of Archean cratons. Examples include the Deccan traps of India, the Paraná basalts of Brazil, and the North Atlantic Tertiary Province (Great Britain, Greenland). There may be hotspot trails associated with these CFB. Other CFB such as the Columbia River basalts, Siberian traps, Transantarctic basalts, and the Keweenawan are not associated with hotspot tracks. Estimated volumes and durations are of order of 10^6 km^3 and 10^6 years, respectively. They generally occur at times of rapid motions and plate reorganizations. Recently, they have been attributed to plume heads and stretching of uniform lithosphere, or to melting of TBLs. The plume hypothesis of *White and McKenzie* [1989] and *McKenzie and Bickle* [1988] requires both lithospheric stretching and a hot plume at sites of abundant magmatism. However, many magma-rich margins have no evidence for either a hotspot or stretching [*Bohannon and Eittreim*, 1991; *Hopper et al.*, 1992; *Holbrook and Kelemen*, 1993; *Ellam*, 1991]. Most CFB are on or near a boundary involving thick Archean lithosphere. Only a few CFB can be related to a hotspot or hotspot track.

The massive and rapid magmatism associated with LIPs has generally been attributed to a deeper or hotter source region than other magmatic provinces. In addition to mantle temperature, the thickness of the lithosphere, the rate of plate extension and the fertility and volatile content of the mantle are important parameters. Figure 5 shows schematically how adiabatic ascent from near the base of the TBL affects the thickness of the basalt layer formed by removal of the melt. A thick basaltic sequence can be formed by separation of thick cratonic lithosphere, but conductive cooling is less likely if the craton occurs only on one side of the rift. Rifting of thin lithosphere generally leads to shallower melting and smaller melting intervals.

The chemistry of CFB has been attributed to crustal contamination, continental lithosphere contamination or derivation from primitive, undepleted, or less depleted reservoirs. There are two main current schools of thought (1) CFB are derived from the CL, that is, the shallowest mantle, under continents (actually, the TBL) or (2) CFB are derived from primitive or enriched lower mantle, being entrained into giant plume heads; CFB are viewed as variable mixtures of lowermost and lower mantle [*Hill et al.*, 1991].

One school holds that CFB are the best probes of the CL. Therefore CL is, by definition, the source, or residue, of basaltic volcanism on continents. The CL is, however, cold, and relatively thin and a more suitable reservoir is a deeper, hotter, more voluminous part of the mantle [*Arndt and Christensen*, 1992; *Thompson et al.*, 1984; *Anderson et al.*, 1992]. The CL has also been invoked as a source for OIB (ocean

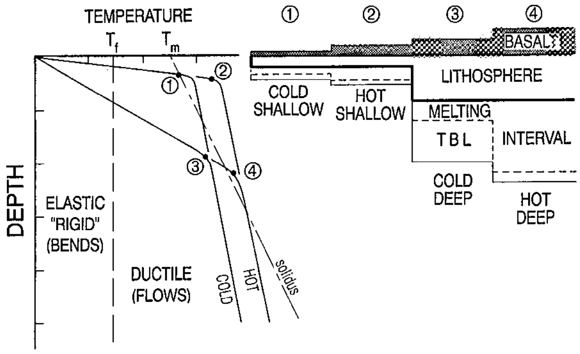

Figure 5. Cartoon illustrating the difference between the elastic or rigid lithosphere, which extends up to T_f and the ductile TBL which extends to roughly the melting temperature. A hot and a cold mantle and a thick and a thin thermal boundary layer are illustrated. Shown schematically are the thicknesses of basaltic layers formed by removing melt adiabatically from the intersections of various geotherms with the solidus. A large melting interval is possible if melts are extracted from beneath thick lithosphere, but conductive cooling is likely unless melts rise at a boundary between thick and thin lithosphere. Pull-apart at structural boundaries may be more important than thinning of uniform lithosphere.

island basalts), and as a contaminant for Atlantic and Indian ocean basalts [*Allègre and Turcotte*, 1985; *Hawkesworth et al.*, 1986; *McKenzie and O'Nions*, 1983; *Peate et al.*, 1990; *Storey et al.*, 1989]. What is usually meant by continental lithosphere is actually the lower part of the thermal boundary layer, or the asthenosphere, as discussed earlier. The difficulties with a lithospheric origin for CFB have been noted many times [e.g., *McDonough and McCulloch*, 1987; *Foland et al.*, 1988; *Halliday et al.*, 1988; *Arndt et al.*, 1993]. The arguments include the thinness and coldness of the lithosphere, the presence of high-temperature melts (picrites) in CFB provinces, the composition of lithospheric xenoliths, and the indifference of the chemistry along hotspot tracks to the age or nature of the overlying lithosphere.

CFB and OIB share many chemical characteristics. The chemical differences between CFB and OIB may be due to processes and depth, rather than source differences. Differences in crustal contamination, ascent temperatures, length of the melting column, degree of partial melting and extent of fractionation as well as the depth, duration, oxidation state, water content, and conditions of sublithospheric evolution prior to eruption can be involved. There may also be an element of timing. The initial magmatism from a section of the mantle affected by subduction may be richer in the very mobile elements (B, Be, Ba, Cs) than later products.

CFB are usually considered to be classic examples of intraplate basalts. However, they often have geochemical characteristics similar to convergent boundary basalts. Sometimes this is taken to suggest influence by a subduction zone or derivation from CL

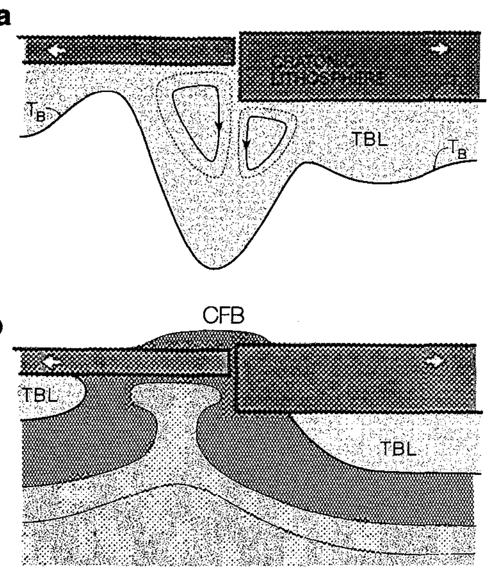

Figure 6. (a) Lithospheric pull-apart at a suture. T_L is the critical temperature defining the base of the plate or the mechanically strong lithosphere. T_B is the temperature defining the base of TBL. TBL can be thickened by the onset of instability, small-scale convection or convergence. Thick, cold continental lithosphere induces downwelling, lithospheric pull-apart generates upwellings. (b) Upwelling mantle replaces the TBL, causing massive magmatism at the cratonic boundary. Plate separation can cause passive, permissive, or opportunistic upwelling of progressively deeper mantle.

which, in turn is assumed to be formed by a subduction related process, for example, accretion of an island arc or a mantle wedge into the CL, or metasomatism of CL by slab-derived fluids. Sometimes a convergence or subducted slab relationship is fairly evident (Columbia River Basalts, Siberian CFB, Keweenawan basalts), while in other cases the nearest contemporaneous subduction zone was 1000 km away (Karoo) or a subduction signature is lacking. Many CFB provinces are in regions affected by convergence, collision or terrane accretion, ridge-trench collision or slab windows [e.g., *Righter and Carmichael*, 1992; *Storey et al.*, 1990]. They usually occur along suture zones at the boundaries of Precambrian cratons. Cratons have particularly thick (>150 km) lithospheres and therefore pronounced lithospheric discontinuities at their boundaries. Vigorous small-scale convection can be expected in the underlying sublithospheric mantle. CFB may be related to this convection (Figure 6).

Subduction can also introduce enriched material into the sublithospheric mantle (what has previously been called the depleted mantle or asthenosphere) that can be tapped through overriding plates, either oceanic or continental. The southern margin of Gondwanaland was enclosed by subduction for hundreds of millions of years and a large amount of recycled sediments, altered oceanic crust and continental debris was dumped into the shallow mantle. Only a fraction of subducted or recycled material reappears in island arc basalts (IAB) or backarc basin basalts (BABB) [*Planck and Langmuir*, 1993] and only a fraction is expected to freeze onto the CL. Most of the LIL, or incompatible, elements recycled into the mantle probably reside in the shallow mantle. In particular, the sublithospheric mantle under southern Africa, South America, India, East Antaractica and western Australia, at the positions they occupied during Gondwana magmatism and Pangea breakup, experienced subduction and recycling, and it is natural to expect that continental magmas from the mantle currently in these areas will exhibit a subduction signature. A southern hemisphere geochemical domain is expected [*Anderson*, 1985]. The geochemical signature usually attributed to CL may well be sublithospheric. In the geophysical fluid dynamics literature, CL is actually the lower part of the TBL and therefore easily removed or delaminated [*White*, 1988]. It is not a long-lived part of the plate, contrary to what is implied in the geochemical literature. The percolation of melts and fluids through the shallow mantle, primarily from subduction related recycling, plus the varying stability depths of hydrous phases, probably means that the shallow mantle is heterogeneous. When a continent is placed in extension by plate tectonic boundary forces the hot metasomatized shallow mantle will probably provide the early melts. Ultimately, the deeper depleted mantle (usually called the asthenosphere in the geochemical literature) will provide magma. This sequence (i.e., enriched basalts under depleted basalts) is common in continental basaltic provinces [*Betton and Civetta*, 1984]. CFB are voluminous and erupt rapidly. This suggests that their ascent is triggered by lithospheric pull apart, over hot mantle or mantle experiencing small-scale convection rather than by the slower processes of crustal stretching, thinning, or erosion. The source mantle is likely to be hot, low viscosity, and may even be partially molten prior to adiabatic decompression. The CL does not seem to be a viable source, but the mechanical response of CL to extension, and its thickness, may control the dynamics of CFB extrusion in a fundamental way. Massive magmatism is sometimes attributed to crustal stretching and thinning. Such crustal response may also be caused by lithospheric pull apart, particularly at sutures and craton boundaries (Figure 6). Rapid small-scale mantle convection and melting are expected at these boundaries [*Mutter et al.*, 1988]. Plumes may cause uplift but they are unlikely to cause breakup [*Hill*, 1991]. Trench roll-back, slab suction, and ridge-trench collisions can also cause extensional stresses on continents. Continental breakup usually occurs along preexisting sutures or other lithospheric discontinuities. CFB are often deposited in depressions (Afar) or subsiding sedimentary basins (Paraná, Karoo, Siberia). Both subsidence and uplift can be caused by changing boundary forces at active continental margins [*Gurnis*, 1993]. Uplift at continental margins is common and is not necessarily the result of plumes.

Many models of massive continental magmatism invoke the heating, stretching, and thinning of uniform lithosphere in order to bring subsolidus mantle into the melting zone by adiabatic ascent. "Pull-apart" is used to describe the separation of lithospheric domains which differ substantially in thickness and age. A lateral contrast in lithospheric or TBL thickness sets up small-scale convection; the boundary is also a stress concentrator. Intrusions can therefore be expected with only minor separation. Uniform lithosphere requires substantial stretching or heating before it is amenable to massive intrusion.

Other mechanisms can also cause continental rifting. For example, the Red Sea and Gulf of Aden are rifts that are propagating toward the Afar depression and are splitting Arabia from Africa. If the Afar were due to a plume, we would expect uplift and rift propagation away from Afar. Other continent splitting rifts also broke toward hotspots (e.g., opening of both the South Atlantic and North Atlantic). The last places to fail are often at boundaries of thick Archean lithosphere.

Some continental flood basalts are underlain by depleted picrites [*Holm et al.*, 1993; *Gill et al.*, 1992]. These provinces are all on the edges of Archean cratons [*Anderson*, 1994]. Lithospheric separation that involves thick (>150 km) Archean lithosphere may induce deeper upwellings and involve melting over a longer column, than initial pull-apart or stretching of uniform lithosphere.

Evidence for pull-apart is indirect. Many LIPs are not accompanied by obvious uplift or crustal stretching, but they are almost all astride significant lithospheric boundaries, usually the edge of thick cratonic lithosphere. Rifts associated with CFB tend to be narrow rather than diffuse. Lithospheric discontinuities can be expected to concentrate stress. Pull-apart has not been investigated theoretically. Most discussions of lithospheric extension start out with homogeneous, constant thickness lithosphere, and the issue does not arise.

Material recycled to the mantle via subduction may initially be confined to the mantle wedge, the probable source of the trace element signatures of IAB and BABB. When subduction ceases, the metasomatized mantle wedge, which may also be basalt depleted, should spread laterally across the top of the sublithospheric mantle. The shallow mantle is an unlikely source for the voluminous and depleted mid-ocean ridge basalts, until it has been cleared out by previous magmatism. It is an open system. Fluids expelled from the slab, probably by dehydration reactions, impart an EM signature to the shallow mantle.

If it were not for sedimentation and reactions with seawater, the oceanic lithosphere would remain depleted all the way from the ridge to the trench. If the incompatible elements introduced by the above processes are removed by processes at the subduction zone, including slab dehydration, then the depleted mantle below the slab dehydration front is essentially a closed system for those elements that distinguish DM from EM. DM may therefore be "isolated" from EM not so much by convective processes as by chemical and petrological processes and location. The shallow mantle acts as a filter for the elements that were added to the crust and lithosphere after its origin at a ridge. Part of these return quickly to the surface in convergent margin magmas. Part are available for later CFB and OIB magmatism.

Continental Lithosphere

Formation. The mantle lithosphere under ancient cratons is cold and thick. It may be of the order of 150–200 km in thickness, as measured both by seismic techniques and by flexure [*Anderson*, 1979; *Anderson et al.*, 1992]. Archean lithosphere is particularly thick and high velocity. It may be the refractory residue of ancient high-temperature differentiation processes, including komatiite extraction. Cratonic lithosphere probably differs from the surrounding mantle in both composition and mineralogy [*Jordan*, 1975; *Anderson*, 1990; *Anderson and Bass*, 1984]. It probably influences mantle convection and entrainment by moving plates.

The ancient continental lithosphere under Archean cratons has been protected from outside influences by a combination of circumstances: it is cold; it is strong; it has high viscosity; it is probably particularly refractory compared to lithosphere elsewhere; it is probably buoyant; it moves away from hot mantle; it is protected from subduction stresses and fluids by its generally central location and surrounding mobile belts; and it may be very dry.

There is evidence that lithosphere splits or pulls apart [*Sykes*, 1978; *Bailey*, 1992], rather than stretching, thinning, eroding, or delaminating. The rheology of the overlying crust and the underlying TBL is quite different, and these elements may stretch and thin. If the lithosphere is relatively easily mobilized, reactivated, left behind, or delaminated, as implied in many current models for CFB, OIB, and asthenospheric contamination, then it could not survive for the two or so billion years required by these models to build up the necessary isotopic anomalies.

On the other hand, a buoyant sublithospheric layer, constantly regenerated by recycling and metasomatism (i.e., an open system) can be both the sink and the source of the geochemical signatures, which characterize "hotspot" magmas. It serves to isolate DM from recycling that evolves as more of a closed system. It matters not that the sublithospheric region is mobile. In fact, in hypotheses that assume that the entire upper mantle is depleted, it must be assumed that CL is remobilized in order to get it to plume sites [e.g., *Allegre and Turcotte*, 1985].

Although CL (as opposed to the continental TBL) may well be a physically isolated reservoir (isolated from the convecting mantle), it is not necessarily a source region for any voluminous basalt type, and, if it is chemically distinct from most of the upper mantle, it is not necessarily denser just because it is cold. A residue from ancient high-temperature melting events will be dry, olivine-rich, and garnet and Fe-poor, which is consistent with the seismic data [*Anderson*, 1990; *Anderson and Bass*, 1984]. This material is an unlikely candidate for CFB genesis.

The lithosphere under continents is often interpreted to be the refractory residue remaining after extraction of basalt [*Jordan*, 1975]. However, some low-temperature garnet peridotites brought up by kimberlites emplaced at Archean cratons are particularly refractory (low in FeO, TiO_2, and Al_2O_3). These may result from the extraction of komatiite (rather than basalt) or they may be residual from magma ocean crystallization [*Anderson*, 1983; *Herzberg*, 1993]. The original material, however, must be more silica-rich than pyrolite, or most peridotites [*Herzberg*, 1993]. In the magma ocean evolutionary scheme [*Anderson*, 1983, 1989] the original composition of the mantle is close to chondritic, and the shallow mantle becomes olivine-rich because olivine floats and eclogite sinks. In this model the refractory material presently in the continental keel may predate the continents. In any case, when the evolution of the Earth is taken into account, there is reason to suspect models that assume that the composition of the continental lithosphere and TBL are identical in composition to the underlying mantle as is commonly assumed in convection modeling and in delamination scenerios. The cratonic keel may differ from continental mantle elsewhere by its dryness [*Polet and Anderson*, 1994], as well as its chemistry.

A cold, ultrathick (>150 km) strong continental keel may not be as invulnerable as generally thought [*Kincaid*, 1990]. It may be easily destroyed by mantle convection. On the other hand, a global buoyant low-viscosity sublithospheric layer can, in effect, remain chemically isolated from the underlying mantle if its density and viscosity make it difficult to entrain. The natural name for such a layer is asthenosphere, but this name has been preempted in the geochemical literature to mean "the depleted layer" or "MORB reservoir." Asthenosphere should not imply depletion or chemical buoyancy, so we should refer to this region by a different name. The asthenosphere could be chemically layered, or zoned, or conceivably could initially provide enriched basalts and then depleted basalts (marble-cake model).

It is perhaps better to think of the depleted (MORB) reservoir as being the isolated reservoir. It is the isotopic differences between MORB and other magmas that indicate the long-term chemical separation of mantle source regions. The shallow mantle (the perisphere) is refreshed by subduction, intramantle fluid transfer, metasomatism and melt trapping, and depleted by melt extraction and is not chemically isolated from all external influences (neither are the so-called continental lithosphere or TBL reservoirs). With regard to trace elements, particularly those carried in hydrous fluids and melts, the shallow mantle is an open system. These postulates are in stark contrast to those in the plume hypotheses that postulate recycling of near-surface material to the core-mantle boundary without contaminating the depleted upper mantle [*Hofmann and White*, 1982].

The deeper, DM is more isolated from recycling and shallow flushing and is protected by the overlying EM that I propose is the sink and source of most of the incompatible element recycling. The homogeneity of the MORB reservoir I attribute to its chemical isolation, not to efficient convective stirring. It evolves as more of a closed system than the shallow mantle. It is probably below the primary dehydration levels of hydrous phases in subducting slabs. Although some hydrous phases and incompatible elements may be carried deeper the bulk of them are in or on the crust, near the top of the slab, and are exposed to high temperatures, high shear stresses, and such phase changes as eclogitization, dehydration, and wet melting. Slab fluids initially enter the mantle wedge (Figure 3), and they are not entirely removed by convergent margin volcanism. Their presence and fate must be dealt with. To ignore the shallow recycled component and to attribute hotspot chemistry entirely to a hypothetical deeply subducted component seems not to be the most sensible approach. Even if deep mantle phases can accommodate water (and incompatible elements), these components must survive the shallow subduction process. Since LIL are introduced into the oceanic plate under rather low-temperature conditions, it seems that they may be removable at the relatively high temperatures that slabs experience at shallow depths. The shallow mantle is unlikely to be a closed system or to be universally depleted.

Lithospheric reservoirs. Continental lithospheric mantle may have suffered numerous trace element and isotopic enrichment events. For this reason it has frequently been proposed as the source region responsible for observed trace element and isotopic characteristics of continental volcanics [e.g., *Hawkesworth et al.*, 1988; *McDonough et al.*, 1985; *Menzies*, 1990]. However, the shallow mantle everywhere, particularly beneath thick plates and especially above subducting slabs, should also receive this enrichment. If it can survive for long periods of time, perhaps by its buoyancy or resistance to entrainment, or is constantly recharged with enriched material, such as subducted continental debris, then it also is a suitable "ancient enriched" reservoir [*Anderson*, 1985]. The "contaminated layer" may, in part, be the lower portion of TBL. It may, in part, be the mechanically weak asthenosphere. It may be involved in the transient shear boundary layers below the plate.

Since lithosphere is cold, an external source of thermal energy is needed, and, if it is refractory, an external source of magma is required as well. The indifference of hotspot chemistry to the nature of the overlying lithosphere (e.g., Cameroon Line, New England hotspot track) suggest that the trace element and isotope chemistry is also derived external to the plate [e.g., *Fitton and Dunlop*, 1985; *Foland*, 1988; *Halliday et al.*, 1988, 1990]. *Menzies* [1992], a previous advocate of the CL reservoir hypothesis, has reappraised the idea and found it wanting on chemical grounds. Xenoliths from the continental mantle do not have the appropriate chemistry to be the source of CFB (see also *Arndt et al.* [1993]).

The CL has been the reservoir of choice for enriched basalts mainly as a last resort. If it is assumed that the asthenosphere (therefore upper mantle, in most stories) and the "convecting" (therefore homogeneous, in most stories) mantle are equivalent to the ancient depleted, or MORB-reservoir, then "the only obvious remaining possibility (for 'less depleted reservoir') is the continental lithosphere" [*McKenzie and O'Nions*, 1983, p. 229]. Actually, asthenosphere, upper mantle, convecting mantle, depleted reservoir, and MORB source are not equivalent concepts and they need not refer to the same part of the mantle. Likewise, convecting mantle does not have to be well mixed or homogeneous. Individual convection cells (or mantle domains) may become chemically homogeneous, but the upper mantle has numerous cells, or domains. The mixing time of the whole mantle is much longer than the mixing time of an individual convection cell [see also *Arndt et al.*, 1993]. Geochemical domains may have dimensions of the order of the convecting layer thickness. Because recycling and melt extraction are continuous processes, acting on different parts of the mantle, one can doubt whether the mantle ever gets well mixed.

The semantic confusion has lead to conceptual confusion. *Perry et al.* [1988] suggest that Rio Grande rift magmas are from EM but that this does not correspond to the lithosphere under this region. *Perry et al.* [1988, p. 432] propose that the lithosphere had been thermally thinned and converted to asthenosphere "while retaining its geochemical properties" and that the thermal conversion of EM lithosphere to EM asthenosphere is followed by erosion of EM and replacement by DM. It would seem much more straightforward to accept the evidence for a hot shallow enriched layer, which is eventually depleted or pushed aside by upwelling DM rather than invoking a local and ancient lithospheric reservoir for which there is no evidence or need. On the other hand, the CL hypothesis does address the need for a shallow enriched layer, an alternative to the deep recycling involved in plume hypotheses.

CL has been viewed as a reservoir physically isolated from the convecting mantle, that can evolve, with time, the isotopic differences that distinguish it from the MORB reservoir. According to *McKenzie and O'Nions* [1983] the continental lithosphere (the coldest part of the mantle) detaches, falls to the base of the uppermantle, heats up, becomes unstable and, as hot rising jets (the hottest part of the mantle) forms oceanic islands. According to this view, isolated enriched material exists at the top and the base of the upper mantle (the convecting mantle) and cycles through it without contaminating the depleted MORB reservoir. *McKenzie and O'Nions* [1983] use the word lithosphere for the TBL and assume that TBL is denser than the underlying mantle because it is cold; that is, the mantle is viewed as a homogeneous fluid.

One must distinguish among the source of magma, the source of heat, and the source of the distinctive geochemistry when discussing the role of CL (or any reservoir) in hotspot volcanism. For example, a depleted but fertile (i.e., basalt-rich) sublithospheric plume (active or passive) could obtain its enriched signature as it rises through, or evolves in, the shallow mantle. Such a plume could provide the heat and the basalt and the shallow mantle could provide the distinctive geochemistry. Some variants of the CL reservoir hypothesis use the CL in this way to provide the CFB geochemical signature. Others use CFB chemistry as evidence for a lower mantle origin, assuming that the basalt, the isotopic signature and the heat source are always together (and that the uppermantle is depleted). The *Hofmann and White* [1982] crustal recycling hypothesis views subducted material as a package. This conflicts with the evidence that slab-derived LIL occur in IAB and BABB and, presumably, therefore enter the shallow mantle at the mantle wedge. The deep recycling hypothesis makes severe demands on processes near the top of the subducted slab, and the interaction between the slab and the depleted reservoir. Alternatively, EM may be a global, shallow layer, underlying the lithosphere and overlying DM. It is probably not a closed system, but it may be chemically isolated from DM by processes already discussed.

Evidence from hotspot traces that cross from ancient continents to the oceans rule out CL as the source of either the magma or the geochemical signatures [*Fitton and Dunlop*, 1985; *Foland*, 1988; *Halliday et al.*, 1990]. Since CL is cold and is not stationary, it is not a suitable source of heat either; it must be activated from below if it is to be involved at all in CFB. Adiabatic ascent of material from below the TBL almost invariably results in melting (Figure 5).

The geochemical characteristics that have been attributed to CL are also found in basalts in the Indian and Atlantic Oceans [*Hawkesworth et al.*, 1986; *Storey et al.*, 1989] and in convergent zone magmas.

One can debate whether the enriched nature generally attributed to CL xenoliths is the cause rather than the result of CFB magmatism. Magmas from fertile mantle rise through, or along-side, the cold lithosphere on their way to the surface. The CL acts as a cold finger, causing crystallization of the rising melt, which imprints some of its geochemical characteristics. Trapped small melt fractions eventually give rise to enriched isotopic signatures. Quite often, however, the properties of lithospheric samples (xenoliths) are used to argue that the chemical characteristics of CFB are inherited from the CL. CFB are themselves often assumed to be probes of the properties of CL. *Menzies* [1992] has ruled out both the CL and the thermal boundary layer as sources of CFB.

In some CFB provinces (e.g., Deccan traps) the purported CL signature is missing [*Storey et al.*, 1989]; in some oceans (e.g., the Indian) the CL signature is present. In this case it has been suggested that the Indian subcontinent lost its CL, which now resides in the Indian ocean mantle. Seismic, heat flow, and flexual data, however, indicate that the elastic CL under India is thick and cold both now and at the time of the Deccan magmatism [*Anderson et al.*, 1992; *Negi et al.*, 1992]. The heat flow from the Indian craton is very low and is not different from other Precambrian shields [*Gupta*, 1993]. The Deccan traps (65 Ma) were apparently fed by a narrow zone near the western edge of the Indian craton, at the intersection of ancient suture zones and a recent (~80 Ma) continental margin, and near a new oceanic spreading center. The rifting and magmatism occurred on a rapidly moving plate that was starting to deform by contact with island arcs to the north [*Courtillot et al.*, 1988; *Klootwijk*, 1992]. Other large igneous provinces on continents are related to ridge-trench annihilation, overridden oceanic ridges, slab windows, ancient sutures, fracture zones, and cratonic boundaries. This suggests that near-surface conditions strongly influence, if not control, location of upwellings and that EM is a widespread, perhaps global, shallow region of the mantle.

The LID velocities in the vicinity of cratons that have been the sites of CFB magmatism and, in some hypotheses, lithospheric heating and stretching, are little different from LID properties elsewhere [*Anderson et al.*, 1992]. This seems to rule out some aspects of both the CL and plume hypotheses for CFB. On the other hand, there is no objection to a relatively shallow, but sublithospheric, source for CFB. Such a source could involve the lower part of the TBL or the shear boundary layers beneath the plate.

Wet lithosphere model. CL is cold and strong and is an unlikely source for the voluminous CFB [*Arndt and Christensen*, 1992]. *Gallagher and Hawkesworth* [1992] propose that CFB are produced from wet lithosphere, one that has a melting point 500° colder than dry mantle. However, wet mantle is also weak mantle and is unlikely to remain attached to the overlying plate for the amount of time required to build up the isotopic anomalies, which have been attributed to the hotspot or plume reservoir. The lower parts of Gallagher and Hawkesworth's lithosphere have temperatures ranging from 1000° to 1500°C. These temperatures, particularly for a wet mantle, yield seismic velocities much lower than those observed beneath Precambrian cratons [*Grand and Helmberger*, 1984; *Anderson*, 1990]. We can estimate the seismic velocities in the CL* of *Gallagher and Hawkesworth* [1992] using data tabulated by *Anderson* [1988]. At 1200°C and 60 kbar (180 km depth) the inferred shear velocity for an olivine-rich mantle is 4.6 km/s. The observed seismic velocity near this depth under cratons is 4.78 km/s [*Grand and Helmberger*, 1984]. A wet lithosphere will have either hydrous minerals or free fluid, and this will reduce the theoretical shear velocity even further. We conclude that the seismic velocities in wet, hot lithosphere will correspond to those in the low-velocity layer or asthenosphere rather than the lithosphere. In order to increase the velocities to observed LID values the temperature must be reduced to about 600°C. This agrees with the more detailed modeling of *Anderson and Bass* [1984].

Gallagher and Hawkesworth [1992] did not discuss the viscosity of their hydrous lithosphere (CL*). Much of their CL* is hotter than 1100°C and will flow readily, particularly if wet. The elastic thickness of a dry olivine lithosphere is roughly defined by the 720–1000 K isotherm; that is, the base of the strong plate is between 450° and 730°C [*De Rito et al.*, 1986]. A wet lithosphere will be even thinner.

For a wet olivine rheology the strain rate for a low stress level of 1 MPa is $\sim 10^{-20}$ s^{-1} for $T = 650°C$ [*De Rito et al.*, 1986]. This can safely be considered elastic or rigid. Lithospheric stress levels are 10 to 100 times larger, and this increases strain rates by 10^3 to 10^6, to 10^{-17} to 10^{-14} s^{-1}. These are still relatively low strain rates. However, strain rate increases exponentially with temperature. A 200° increase in temperature decreases silicate viscosity by about 3 orders of magnitude and increases the strain rate, for a given stress, by the same amount. At 1200°C and 500 MPa the strain rate is 10^{-8} s^{-1}. Thus it is clear that conditions envisaged by *Gallagher and Hawkesworth* [1992] for their basalt reservoir (CL*) continental lithosphere will not be long maintained. The lower part of their hot, wet lithosphere will flow rapidly away (Figure 2), and the stresses and strain rates, at least in peridotite, will approach asthenospheric values. Such material is likely to be buoyant and of low viscosity and will therefore tend to spread along the top of the sublithospheric mantle. The conditions invoked for their CL* are actually appropriate for the TBL or asthenosphere and for what I call the perisphere to avoid this semantic confusion. Hot plume heads will also tend to form a thin layer at the top of the mantle. In some models the asthenosphere is maintained by plumes [*Vink et al.*, 1985]. I have argued above that both the rheology and enriched chemistry of the shallow mantle may be maintained by subduction, including slab-derived fluids. It should also be pointed out that CL beneath CFB provinces does not differ physically from CL elsewhere [*Anderson et al.*, 1992]. There is no geophysical evidence for lithospheric damage at CFB sites.

There is nothing in the geochemistry of the enriched (or nondepleted, or hotspot) magmas that requires that they come from a cold, strong, high-viscosity layer or one that is and has been rigidly attached to the overlying crust for a long period of time. In fact, there is not much encouragement in the geochemistry of mantle xenoliths that suggest that they have the attributes needed for massive volcanism either on continents or in the ocean basins. It is only the belief that the underlying asthenosphere is homogeneous and depleted that gave rise to the concept of a lithospheric reservoir. Lithospheric contamination was also the simplest extension of the crustal contamination model, which was the prevailing explanation of enriched basalts for many years.

Isotopic systematics are the main reason for invoking a CL (i.e., physically isolated) source for CFB. If the radiogenic nature of some of the isotopic systems in CFB are interpreted as due to long-term physical isolation of EM from DM then one requires an isolation mechanism (high strength, buoyancy, location (e.g., continental crust or CMB)). If it were not for the isotopic evidence, and its interpretation in terms of an age constraint, then there would be little reason to associate the CFB reservoir with CL. If all CFB were isotopically similar to the surrounding crust then one would want a mechanism that involved crustal contamination, or local recycling, or contamination by a similar age piece of the mantle. Since the large CFB provinces are found on the margins of Archean cratons (Columbia River—Snake River, Deccan, Paraná, Karoo, Greenland, Keweenawan, etc.) one suspects that thick, or ancient lithosphere is somehow involved in their genesis, either physically or chemically. The ancient rifts and sedimentary basins in these locales may also be involved.

Semantics aside, the evidence is strong that the enriched signature of many basalts is acquired in the shallow mantle (perisphere), and there are many enrichment mechanisms (subduction, residual melts, etc.) [*Anderson*, 1983, 1989; *Hawkesworth et al.*, 1987]. The plume head hypotheses also imply widespread enriched regions in the shallow mantle. There is also evidence that the depleted, or MORB, reservoir acquired its depleted character billions of years ago and that this reservoir is semi-isolated, or not well mixed, with the enriched reservoir or with recycled material. There is no requirement that one of these reservoirs be strong, or rigidly attached to the continental crust. The phrase shallow mantle or uppermost mantle or contaminated layer could be substituted for continental lithosphere in any of the papers that advocate a continental lithosphere source, and all of the same facts would still be satisfied. What would change would be the mechanisms by which enriched basalts are formed and the enriched reservoir isolated from the MORB reservoir. A weak upper mantle EM, however, would not be confined to the vicinity of continents. A complex series of events has been proposed [*McKenzie and O'Nions*, 1983; *Allègre and Turcotte*, 1985] to get cold enriched CL from the continents, to the deep mantle and then to ocean islands in order to form enriched OIB. This must be done without affecting the intervening depleted MORB reservoir. These models imply an enriched (EM) mesosphere and assume a shallow and global DM.

At one time it was believed that CL samples (mantle xenoliths) defined a mantle isochron which was contemporaneous with the continental crust. This concept did not hold up nor was it confirmed by various other isotopic systems. An alternative explanation involved mixing lines. The current concept is that ancient continental lithosphere is complementary (depleted) to the overlying crust but has been metasomatized and reenriched since formation [*Hawkesworth et al.*, 1988]. There is now no reason to believe that the source regions for most CFB provinces are identical in age to the continental crust they are deposited on. *Anderson* [1981c, 1989] suggested that depleted and enriched mantle reservoirs formed at the same time and that the continental crust subsequently formed from the shallow enriched layer. The shallow layer is reenriched by recycling, via subduction, of continental crust and altered oceanic crust, and is contaminated by melts on their way to the crust [*Anderson*, 1981c, 1983]. There may therefore be a genetic chemical relationship between the crust and the underlying mantle, even if they are not rigidly attached. The concept of an EM rather than a primitive mantle, or crustal contamination, has been controversial, but now numerous EM components have been proposed. These have been attributed to various precursors (pelagic sediments, terrigeneous sediments, oceanic crust, dehydration products or residues, trapped silicic or carbonatitic melts, metasomatic fluids).

Continental Roots

Old mountain belts require a strong layer more than 100 km thick to support the load (see *Anderson* [1990] for a review). Stable cratons have higher than average mantle seismic velocities [*Toksöz and Anderson*, 1966; *Nataf et al.*, 1986]. Higher-resolution studies show that correlated high-velocity anomalies (HVA) extend to about 250 km under Archean cratons with only some cratons showing HVA to ~400 km; younger age cratons do not show consistent correlations [*Polet and Anderson*, 1995]. *Jordan* [1975] associated HVA with continental keels while *Anderson* [1990] put the plate boundary at the depth of a decrease in seismic velocity (~150 km under old cratons). If the magic temperature concept holds at high pressure one expects little strength, even under cratons, below some 150–200 km. There is no direct evidence for any load bearing capacity below this depth. On the other hand, conditions during the Archean were unique. Removal of high-temperature magmas, such as komatiites, from the mantle should leave behind a particularly dry (no volatiles (NV)) and refractory residue. Dry rocks may be strong [*Mackwell and Kohdstedt*, 1993]. Venus appears to be strong in spite of high surface temperatures. It may be dry. If the mantle below Archean cratons was formed by komatiite removal and if it was subsequently isolated from slab-derived fluids, perhaps because of its depth relative to dehydration reactions, then it could be ultradry (NV) and could be strong to higher temperatures than normal "wet" mantle (the Venus NV analog).

The deep HVA under Archean cratons, (>150–200 km) could also be transient phenomena, for example, downwellings generated by the lateral change in TBL thickness and temperature or by the motion of the continent relative to surrounding mantle. Even if the cratonic plate is only 150 km thick it should affect mantle flow, particularly the locations of downwelling, which will show up as seismically fast regions.

ASTHENOSPHERE

The asthenosphere, or weak layer, comprises the top part of the upper mantle. According to *Barrell* [1914], the asthenosphere yields readily to large-enduring strains of limited magnitude. It is often equated to the seismic LVZ, a zone that varies substantially laterally but, under oceans, is crudely between the depths of ~20–50 km to about 220 km. where there is a seismic discontinuity or a region of high-velocity gradient. Under cratons the LVZ may start below 150 km [*Grand and Helmberger*, 1984]. In some regions, low velocities extend to depths greater than 400 km. The LVZ may be caused by a partial melt zone [*Anderson and Sammis*, 1970; *Anderson and Spetzler*, 1970]. A partial melt zone can become dynamically unstable and offers an alternate explanation to deep mantle plumes for persistent upwellings [*Tackley and Stevenson*, 1992]. Substantial shear-velocity reductions can also occur under subsolidus conditions due to dislocation relaxation [*Anderson and Minster*, 1981; *Minster and Anderson*, 1980] and dehydration. The lowest velocity regions of the upper mantle have lower velocities than can be achieved by any combination of mantle minerals at reasonable temperatures [*Anderson and Bass*, 1984]. This implies a high-temperature anelastic process, such as a phase change or dislocation relaxation, or partial melting. Mantle silicates lose strength at 650° ± 50°C, about half the absolute melting temperature. The lowest-viscosity regions of the mantle also probably occur in the shallow mantle.

Unfortunately, the geochemical term depleted (low contents of incompatible elements, low $^{87}Sr/^{86}Sr$, and so on) has been applied to the LVZ, giving DLVZ [*Schilling*, 1973b], and subsequently to the asthenosphere and, currently, to the whole upper mantle. In the petrological and geochemical literature the asthenosphere is now almost universally referred to as the depleted asthenosphere, or convecting mantle and is assumed to be the source region for depleted basalts, such as MORB. The shallowest mantle, under the lithosphere, is assumed to be depleted (DM), and this is so widely accepted that little discussion of the assumption is seen in the current literature.

Thus, according to *Weiss et al.* [1987, p. 255], "There is no doubt that the lower mantle is the OIB (Ocean Island Basalt) source and the upper mantle . . . the MORB source . . ." and according to *Hart and Staudigel* [1989, p. 15], "The DMM component (Depleted MORB mantle) comprises the upper mantle . . . this is the only mantle reservoir for which a location and evolutionary process are well established."

A variety of physical and geological processes have indicated the need for a layer of low strength and low viscosity. Most of the sublithospheric shallow mantle is close to, or above, the solidus, and it has therefore been assumed that the most voluminous magma type, mid-ocean ridge basalts (MORB), must originate in the asthenosphere. Since it is weak, it has also been assumed that it is uniform in composition, having been homogenized by convective stirring. The terms asthenosphere, depleted mantle, MORB reservoir, convecting mantle, and upper mantle are used interchangeably by most petrologists and geochemists.

Since the shallow mantle is sampled passively by midocean ridges, it has been assumed that depleted mantle everywhere immediately underlies the lithosphere, or plate. Depleted MORB-like magmas, however, occur only under special circumstances. In the shallow EM models, passive rifting initially draws in the sublithospheric layer, and enriched basalts are expected to dominate at the earliest stages of continental rifting, ridge jumps, and slab-window formation (Figure 4). MORB occurs only at late stages of rift formation, usually just prior to seafloor spreading. If rifting is aborted, or if a ridge is abandoned, one can expect a return to enriched magmas as EM flows back in. An enriched shallow layer can be attenuated or thinned by mechanical processes or depleted of its enriched components by prior melt extraction episodes. In the model being discussed a period of enriched magmatism is a prerequisite to MORB magmatism. It is also important to realize that mid-ocean ridges are not uniform in depth or chemistry and that they are unlikely to be fed by continuous vertical sheets. Convection at ridges is likely to be intrinsically three dimensional with both on-axis and off-axis upwellings and downwellings. Absence of EM melts at some ridge locations may represent either a spatial situation (thinning of EM or upwelling of DM) or a temporal situation (shallow EM has been depleted by prior melt removal). Enriched magmas erupt under a variety of circumstances and seem to originate in a shallow enriched layer. The presence of such a layer (the perisphere) between the plate and the depleted reservoir explains many paradoxes and removes the need for special pleading in the explanation of many hotspot related phenomena. "Plume heads" and the lower part of the thermal boundary layer under continents are weak and spread laterally across the top of the mantle. The net result is a global shallow EM and a deeper isolated MORB source. Basalts from the depleted source are most evident at mature and rapidly spreading ridges. Depleted high-temperature melts (picrites, komatiites) are also components in "hotspot" magmas. These observations support a depleted source that is deeper than EM.

PERISPHERE

It is clear that lithosphere and asthenosphere have lost their original precise geophysical meaning in the modern geological and geochemical literature. It is difficult to discuss the possible locations of the OIB, CFB, and MORB reservoirs because of the preconceptions now associated with lithosphere and asthenosphere and, for that matter, upper mantle and lower mantle.

It is quite likely that enriched, depleted and less depleted magmas all originate in the upper mantle. Even advocates of a depleted upper mantle require plume heads or delaminated CL to locally fill up the asthenosphere on demand when hotspot magmas are required, or to explain CL, plume or lower mantle signatures in magmas in various tectonic environments.

At one time all enrichment of continental basalts was attributed to crustal contamination. It then became clear that some such basalts and some patterns of enrichment were acquired in the shallow mantle prior to passing through the crust. This led to the concept of a continental lithosphere source, a source physically isolated from the depleted mantle, the idea being that if the shallow mantle source was hot or ductile, it could not remain isolated from the depleted, convecting mantle. Although enriched material is found along mid-ocean ridges, at fracture zones, at dying ridges, at new ridges, at propagating ridges, at infant subduction zones and at fault upthrust exposures, it is generally believed that this material must be imported from elsewhere (i.e., from CL or from the CMB). The more obvious explanation of these observations is that there is a widespread enriched shallow layer at the top of the mantle that is either the source of OIB and CFB or, more likely, serves to contaminate the diapirs, or plumes, from greater depth as they traverse or evolve in the shallow mantle. Contamination should be most severe under thick lithosphere, and other locations where direct access of magmas to the surface is inhibited. This assumes that melts that become trapped in the sublithospheric EM become more contaminated as they cool and evolve prior to eruption.

This layer, below the lithosphere, is warm to hot ($T > 550°C$), weak and readily interacts chemically with more depleted plumes or diapirs or melts from a deeper depleted region. It is normally less dense than the underlying mantle but denser than the lithosphere. It is eventually pushed aside or stripped of its low-melting fraction by continual interaction with active or spreading-induced-passive upwellings, thereby permitting uncontaminated melts from the deeper source to rise to the surface. The shallow layer only becomes attenuated or stripped after long sustained spreading or upwelling. MORB-like basalts only emerge after a long history of rifting or spreading. Alkalic and evolved basalts are expected during the initiation and termination of rifting and during the earliest and latest phases of "hotspot" magmatism. Basalts bordering the Atlantic differ from slightly younger basalts that mark the beginning of seafloor spreading.

A similar mechanism is sometimes invoked to explain the temporal sequence from enriched to depleted basalts, common in rifting environments, but it is the cold, strong continental lithosphere that is viewed as the stretching, attenuating, depleting layer rather than a hot, ductile sublithospheric layer. In both cases, EM is shallow, and DM is deep.

Arguments that have been made [*Hawkesworth et al.*, 1987; *Menzies and Hawkesworth*, 1987] to explain the enrichment of the shallow continental mantle (CL or TBL) apply equally well to the shallow mantle everywhere [*Anderson*, 1985]. The shallow layer can be ancient, just as can the CL, but it can be "isolated" from the MORB reservoir by its buoyancy (relative to DM) and weakness, rather than by its high strength.

I propose that this layer be called the perisphere ("peri" for all around, close). The perisphere underlies continents and serves the same purpose, as far as chemistry is concerned, as the continental lithosphere (and, more recently, plume heads) of the petrologists. The perisphere also underlies oceanic environments, although at much shallower depths, and removes the need for delamination, remobilization, or long-distance transport of CL, and plumes or plume heads, to explain enriched basalts or so-called hotspot magmas far from continents. The layer is eventually squeezed, pulled, or pushed out of the way by plate divergence or by active or passive upwellings and therefore is locally absent or thinned or depleted under mature spreading centers (Figure 4). It can be entrained by plates and upwellings. It readily flows back into an area abandoned by spreading or rifting and is available at migrating and propagating ridges and fracture zones. Being buoyant it is not readily dragged down into the mantle and can spread across the top of many convection cells, particularly if entrained by moving plates. On the other hand, it may be dammed by fracture zones and piled up at convergences of convection cells, below the plate but above the convecting mantle. Rifts on drifting continents can sample the shallow enriched layer even many thousands of kilometers away from hotspots or the point of initiation of the rift and the volcanism. The perisphere concept therefore explains many of the paradoxes of CL and plume theories. Midplate volcanoes, such as Hawaii, may obtain their EM signature by plate-entrained flow of the perisphere by the Pacific plate. A deeper upwelling, from DM, would provide the bulk of the magma. Because ridges represent diverging flow there is less chance there for EM contamination than at midplate environments (Figure 4).

I define perisphere as the enriched, weak, shallow mantle beneath the (real) lithosphere. This new word is needed because, in the geochemical literature, the lithosphere is "known" to be enriched (and presumably strong) and the asthenosphere is known to be depleted (and presumably weak). Something that is weak, hot, and enriched (and nonplume) is not in the petrological/geochemical lexicon.

At regions of present and recent convergence the perisphere lies above the subducting/dehydrating/melting slab and the deeper depleted (MORB) reservoir and therefore is constantly reenriched (sometimes with ancient continental crust/lithosphere material), while it protects the depleted reservoir. It also stands between the depleted reservoir and the surface (except when pushed aside by upwellings feeding a mature ridge or entrained by diverging plates) assuring that midplate volcanism is usually a mixture of EM and DM. The difference between the CL model of CFB genesis, for example, *Gallagher and Hawkesworth* [1992], and the perisphere concept is illustrated in Figure 2. Mechanisms for enriching and cooling the mantle are shown in Figure 3.

The perisphere need not be particularly fertile. It may have a basaltic fraction, but it may also serve mainly to provide trace elements to deeper, hotter upwelling material from a global depleted reservoir (DM). The low-melting, enriched fraction of the perisphere may be lamproitic or kimberlitic [*Anderson*, 1982a, c, 1985]. Alkalic basalts are generally intermediate in chemistry between these enriched endmembers and midocean ridge tholeiites.

The trace-element and isotopic inventory of the perisphere may be in rough steady state between the material introduced by subduction and melt trapping and the material removed into OIB, CFB, IAB, and BABB (island, continent, arc, and backarc basalts). The perisphere may be refractory harzburgite or lherzolite, secondarily enriched by small-volume residues of melts or vapors. Large-volume melts from a deeper depleted (but fertile) reservoir may provide the bulk of the basaltic or picritic parent that eventually evolves to a variety of ridge, island and arc magmas. I distinguish between fertile (basalt-rich) and enriched (LIL-rich, radiogenic). The MORB reservoir for example, is fertile and depleted. Metasomatized refractory lithosphere or perisphere is infertile and enriched.

Although the perisphere may be primarily hydrated, metasomatized peridotite it may contain eclogite pods and therefore be somewhat fertile as well as enriched. It seems likely, however, that neither the lithosphere nor the perisphere can melt extensively without an active buoyant upwelling from deeper, or an extending area above, to induce upward flow and adiabatic melting. Lateral variations in lithospheric thickness, and the associated temperature gradients, can also stimulate rapid small-scale convection [*Mutter et al.*, 1988]. The largest flood basalt provinces initiated at boundaries between thick (cratonic) and thin lithosphere, where this small-scale convection can be expected to be most intense.

The depleted reservoir (DM) is likely to be fertile, perhaps more fertile than garnet peridotite or pyrolite. A garnet- and clinopyroxene-rich source has been dubbed "piclogite" [*Bass and Anderson*, 1984]. Such a composition can melt extensively and still retain garnet, which is not possible for pyrolite. Piclogite may be recycled demetasomatized oceanic plate. It may originally have been cumulates from a magma ocean or delaminated eclogitic lithosphere [*Anderson*, 1981a, 1982a]. Upper mantle peridotites may also be buoyant cumulates from an SiO_2-rich (chondritic) magma ocean [*Herzberg*, 1993; *Anderson*, 1983].

Refreshing the Perisphere

The oceanic plate starts its journey at the ridge as virginal LIL-depleted crust and upper mantle. Hydrothermal alteration and sedimentation converts it to a contaminated and enriched package, which, upon recycling, is not a suitable source for the next generation of MORB. The recycled package is often viewed as a potential OIB reservoir [*Hofmann and White*, 1982] after falling to the base of the mantle and being reborn as a narrow cylindrical plume.

Scraping, melting, and dehydration serve to return a large fraction of the upper portions of the slab to the mantle wedge (Figure 3). Dehydration reactions of altered minerals flux the mantle wedge with water and LIL some of which enters island arc and backarc basin basalts, but much of which probably stays in the shallow mantle. Most hydrous minerals are unstable below 100 to 200 km, and it seems unlikely that much LIL is left to accompany the bulk of the slab to greater depths. This, however, is an open question that deserves the attention of petrologists. This process can be called "demetasomatization of the slab," with the implication that the low-temperature metasomatism the plate experiences as it crosses the ocean can be undone at the high temperatures and stresses that the slab encounters upon reentry. Rheologically, most of the perisphere is part of the asthenosphere. The TBL is part of the perisphere. Chemically, it serves the same purpose as the continental lithosphere of the petrologists and the plume head of plume advocates. It is not isothermal. It is hot in regions of continental insulation or absence of subductive cooling. It may partially melt, upon adiabatic ascent, at sites of lithospheric thinning, stretching or pull-apart, or when impacted from below by active upwellings. It differs from previous hypotheses by virtue of its chemical buoyancy. TBLs eventually become gravitationally unstable. The perisphere is a semiconvecting layer.

"Ages" of EM and DM

Material recycled into the perisphere includes isotopically evolved continental sediments and seawater alteration products. Magmas that are produced from this reservoir or that include components from this reservoir will also appear to be isotopically evolved even if the recycling is fairly recent and the residence time of recycled material in the shallow mantle is relatively short. EM may turn over at a rapid rate as long as isotopically evolved material (continental sediments, seawater) are involved in the recycling. The depleted MORB reservoir, however, must be immune from most of this kind of contamination, and it probably has a long history of evolution as a chemically closed system. If migrating ridges are the main mechanism for flushing out the shallow mantle, the mean residence time is calculated to be in excess of 10^9 years [*Anderson*, 1993].

ACTIVE VERSUS PASSIVE UPWELLINGS

The mantle is a convecting system although not "the uniform fluid heated from below" system with only thermal buoyancy that is the basis of most convection cartoons. The mantle is driven from within by radioactive decay, viscous heating, secular cooling and slab cooling. It is differentially cooled from above and below. In part, it is heated from below. Phase changes and partial melting provide heat sources and sinks and volume changes. Plates and slabs help organize the system.

Sinking slabs probably represent the strongest downwellings and the strongest source of cooling. Narrow plumes are generally considered to be a style of convection that is independent of plate tectonics and normal mantle convection. Mid-ocean ridges are generally thought to be the result of mantle passively rising to fill the void left by spreading plates. Where in this scheme are the "normal" mantle upwellings, the broad scale buoyant currents that remove heat from the mantle?

Here we come face to face with the problem of distinguishing between active and passive upwellings. There is a close spatial relationship between ridges and hotspots and hotter than average mantle [*Ray and Anderson*, 1994]. Cold upper mantle (high seismic velocity) correlates with post-Pangea subduction. Parts of the Atlantic and Indian Oceans ridge systems have recently migrated away from hotspots. Many hotspot tracks were built on very young oceanic lithosphere, or at ridges. A long sustained period of passive spreading will eventually control the locations of active upwellings, just as slab locations control locations of cold mantle downwellings [*Anderson et al.*, 1992]. Furthermore, the highest-temperature magmas, picrites, and komatiites, the strongest candidates for melts from hot upwellings, generally have isotopic chemistry identical to "passive basalts" such as MORB [*Anderson*, 1995]. The upper mantle under continents and old oceanic lithosphere has high seismic velocities [*Anderson et al.*, 1992]. All of this suggests that midocean ridges are not as passive as generally assumed. They occur over hotter than average uppermantle and the spreading process itself generates melting and buoyancy. Therefore the shallow perisphere can be dragged away by spreading plates, pushed away by active depleted upwellings, and depleted of its low melting components by prior stages of melt extraction. Since viscosity depends on temperature, pressure, and volatile content, it is not clear whether the perisphere has a much lower viscosity than the rest of the sublithospheric uppermantle.

DISCUSSION AND SUMMARY

Various physical parameters change at about the 550°–650°C isotherm. This temperature apparently defines the long-term thickness of the plate and the lithosphere. The base of the lithosphere is much shallower and much colder than generally assumed by geochemists and petrologists. Mantle that is hotter than ~650°C has no long-term attachment to the overlying plate. On the other hand hot, buoyant and, possibly, low-viscosity mantle can be isolated from the underlying mantle and will not mix in if it differs in major element chemistry. For example, a refractory (olivine-rich, Al-poor) layer, or an H_2O-rich layer may be physically isolated either by its buoyancy or its viscosity from a deeper convecting dry fertile region of the mantle (DM). The perisphere can be described as a barrier or a filter for the more incompatible subducted elements. Otherwise, DM will be subjected to contamination. DM may be chemically isolated, even if it is not physically isolated from the overlying mantle.

I introduced perisphere to accommodate the need for a term for a global, enriched shallow region of the mantle that is relatively isolated from the depleted reservoir that provides mid-ocean ridge basalts and a component of other types of basalts. The use of the expression continental lithosphere for this region introduces semantic confusion. Continental lithosphere has been used as a substitute for thermal boundary layer in some of the geophysics literature, but it has been mistakenly assumed to be "the strong layer" in the petrological literature, and the "ancient layer" in the geochemical literature.

When mantle material becomes hotter than ~650°C, it cannot support the stresses required to cause earthquakes or the stresses associated with trench, mountain belt, and volcanic loads. This temperature also defines the base of the seismic high-velocity layer. Hotter material deforms readily and cannot be a permanent part of the plate or lithosphere. The lower half of the TBL has little strength. Mantle nodules exhibiting long-time enrichment and equilibration temperatures in excess of 1000°C are commonly attributed to the continental lithosphere and the implication is that this part of the mantle has been rigidly attached to the overlying crust for more than 10^9 years. Part of the reason for this assignment is the belief that if the sample is not from CL it would be depleted and MORB-like since it is well known that the asthenosphere is depleted. CL is often described as easily delaminated, reactivated, displaced, replaced, or remobilized in order to explain various aspects of continental magmatism and tectonics and ocean island magmas. I suggest that the perisphere concept explains many elements of continental and oceanic geochemistry. The semantic confusion associated with continental lithosphere and depleted asthenosphere makes a new word necessary.

For prolonged loading the effective elastic thickness of the lithosphere is determined by the 500°C isotherm for a dry olivine rheology [*De Rito et al.*, 1986]. This is similar to the temperature at the base of the elastic plate inferred from flexural profiles [*Furlong*, 1984; *McNutt*, 1984; *Watts et al.*, 1980]. The lower bound on seismicity in oceanic lithosphere lies between the 700° and 800°C isotherms [*Wiens and Stein*, 1983] or between 600° and 700°C using the revised geotherms of *Denlinger* [1992]. The thickness of the high-velocity seismic LID in oceanic and cratonic plates lies roughly along the 600°C isotherm [*Anderson*, 1990; *Regan and Anderson*, 1984]. A variety of phase changes occur in mantle rocks at temperatures of 600° ± 50°. These may explain the coincidence of various rheological phenomena. Mantle hotter than ~650°C does not qualify as lithosphere in any sense of the word that implies strength, permanence, high viscosity, or high seismic velocity.

The viscosity of *Gallagher and Hawkesworth*'s [1992] lithosphere can be estimated from published tables [*Mackwell et al.*, 1990]. The inferred viscosities at temperatures of 1100°–1200° and stresses of 60–200 MPa are 4–10 orders of magnitude less than accepted values of viscosity of the lithosphere. This hot, wet CL* will not remain coherent or attached to the overlying plate. However, by reducing the temperature to 650°C the material reaches acceptable lithospheric viscosities of $>10^{24}$ Pa. s. The sublithospheric layer (perisphere) may be entrained by the moving plate and may provide a component of CFB, but it cannot be considered as a permanent part of the CL. It can remain distinct from the underlying "convecting mantle" if it has a suitable density or viscosity contrast. The perisphere concept removes many of the problems associated with the continental lithosphere school of continental magmatism. It also offers advantages over the enriched, deep-mantle plume hypothesis. The so-called hotspot, plume, enriched (EM), or OIB mantle is probably a global, shallow sublithospheric layer. It is not necessarily fertile. I argue that its enriched chemistry is regenerated from above, by recycling, rather than from below, as in the plume head scenarios.

An unresolved problem is the density of the perisphere. If it is primarily formed as an olivine-rich cumulate from a magma ocean, or as a depleted refractory residue after basalt or komatiite extraction, it will be buoyant relative to fertile or garnet-rich mantle. Metasomatism and hydration will reduce its density even further. The same question is involved in the CL* and plume head theories. Are these proposed reservoirs also buoyant, after extraction of basalt? If so, they may become permanent parts of the perisphere. A weak buoyant layer in a convecting system is sometimes called a "semiconvection" zone.

In conclusion, there would be much less confusion in the literature if the terms lithosphere and asthenosphere were restricted to rheological matters. Many continental basalts have inherited their geochemical characteristics from the shallow mantle but few can be shown to have come from or have been chemically modified by the lithosphere. Basalts obviously come from mantle that is above the melting point and therefore hot and weak and "asthenospherelike." Nothing is gained by attributing one basalt type or another to an "asthenospheric" or a lithospheric source. Many complex models of basalt genesis are based on the false premises that only MORB can be generated in the asthenosphere and the asthenosphere is the same as the upper mantle. The shallow part of the sublithospheric upper mantle is likely to be the source and sink of most of the recycled LIL inventory of the mantle.

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