The Eclogite Engine: Chemical geodynamics as a Galileo Thermometer

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A Galileo thermometer is a liquid-filled container with a number of suspended objects. As the liquid changes temperature, the suspended objects rise and fall to stay at the position where their density is equal to that of the surrounding liquid, or until they hit a barrier. The controlling parameter is buoyancy; the object's mass relative to the mass of the liquid displaced. If the object's mass is greater than the mass of liquid displaced, the object will sink. If the object's mass is less than the mass of liquid displaced, it will rise. Different blobs settle at different depths, for a while.

ABSTRACT

Migrating and incipient ridges and triple junctions sample the heterogeneous mantle created by plate tectonics and crustal stoping. The result is a yo-yo vertical convection mode that fertilizes, cools and removes heat from the mantle. This mode of mantle convection is similar to the operation of a Galileo Thermometer (GT). The GT mode of small-scale convection, as applied to the mantle, differs from the Rayleigh-Taylor (RT) instability of a homogeneous fluid in a thermal boundary layer. It involves stoping of over-thickened continental crust and the differences in density and melting behavior of eclogites and peridotites in the mantle. The fates of subducted and delaminated crust, underplated basalt, and peridotite, differ because of differences in scale, age, temperature, melting temperature, chemistry, thermal properties and density. Cold subducted oceanic crust–as eclogite–although denser than ambient mantle at shallow depths, may become less dense or neutrally buoyant somewhere in the upper mantle and transition zone (TZ), and be gravitationally trapped to form mafic eclogite-rich blobs or layers (Anderson, 1989a,b, 2005a,b, 2006). Detached lower continental crust starts out warmer; it thermally and gravitationally equilibrates at shallower depths than slabs of cold mature lithosphere. The density jumps between 400 and 650-km depth act as barriers. Trapped eclogite is heated by conduction from the surrounding mantle and its own radioactivity. It is displaced, entrained and melted as it warms up to ambient mantle temperature. Both the foundering and the re-emergence of mafic and fertile ultramafic blobs create midplate magmatism and uplift. Mantle upwellings and partially molten blobs need not be hotter than ambient mantle or from a deep thermal boundary layer. The fertile blobs drift slowly in the opposite direction to plate motions–the counterflow model–thereby maintaining age progressions and small relative motions between hotspots. Large-scale midplate volcanism is due to mantle fertility anomalies, such as large chunks of delaminated crust or subducted seamount chains, or to the release of accumulated underplate when the plate experiences flexure or pre-breakup extension. Eclogite can have lower shear velocities than volatile-free peridotite and will show up in seismic tomograms as low-velocity, or red, regions, even when cold and dense. This model removes the paradoxes associated with deep thermal RT instabilities and alternatives such as propagating cracks and small-scale thermal convection. It explains such observations as relative fixity of melting spots, even though the fertile blobs are shallow.

Keywords: delamination, small-scale convection, plumes, blobs, LIPs, eclogite

INTRODUCTION

The subduction of oceanic lithosphere drives convection and creates chemical heterogeneity in the mantle. Subduction and conduction of heat through the surface boundary layer are the main mechanisms for cooling the mantle. The delamination or stoping of lower continental crust also affects the dynamics, temperature and composition of the mantle. This is a completely different mechanism of cooling and convection than is traditionally considered (Appendix). Most of the volcanism of the world is associated with plate boundaries; about 24 km^3 of magma is extracted from the mantle each year at plate margins, mostly at midocean ridges. Only minor amounts of magma (1 to 2 km^3/yr) are removed by intraplate (also called midplate, hotspot or anomalous) volcanism. The fact that magmatism is geographically widespread implies that partial melting of the mantle is a common phenomenon, even though the melting temperature of dry pyrolite is very high. The volume of anomalous volcanism is much less than the amount of oceanic crust that is recycled back into the mantle. Melting in the asthenosphere is widespread, but magma may not always be able to rise to the surface. Minor fluctuations in temperature and fertility (eclogite content) are expected to give large variations in melting for a mantle that is close to the average melting temperature.

Most, if not all, volcanic regions lie in areas of extensional lithospheric stress. Melt is extracted from the mantle through dikes that form in such regions, often by magma fracture and shattering of the country rock (stoping). A necessary but not sufficient condition for magmatism is extensional stress, modulated by lithospheric fabric and architecture. A more controversial issue is the required condition for large amounts of melting in so-called hotspot areas. It may be mantle fertility rather than high absolute temperature or water content and may involve magma ponding and episodic release of melt. It may also involve the stability of over-thickened continental crust. All extending regions and rift zones are not expected to be equally productive of magma.

DEFINITIONS

Eclogite, as used here, is a garnet- and clinopyroxene-rich rock of unspecified origin. It includes jadeite-poor materials usually referred to as *garnet clinopyroxenites*. Mantle eclogites and pyroxenites have a variety of origins and can be metamorphic, igneous or reaction products (Lee et al., 2006). Eclogites may be high-pressure crystallized melts, remnants of subducted oceanic crust, or the residue of partial melting of such crust. Some mantle eclogites may be products of metamorphism of mafic lower continental crust, i.e., gabbroic to anorthositic protoliths and some may be high-pressure garnet-pyroxene cumulates. Secondary eclogites can form by the interaction of eclogite partial melts with peridotites, or the intrusion of basalt into lower crustal cumulates.

Piclogite is an assemblage that includes eclogites and peridotites, and low- melting and high-melting temperature components. The components may be kilometers in dimension and separations. Piclogite does not necessarily come in hand-specimen-sized samples. The term *piclogite* was coined because eclogites have a specific composition and origin in the metamorphic literature. Piclogites may be thought of as assemblages that have less olivine and orthopyroxene than peridotites, and are of variable fertility and density. It is unlikely that fertile blobs in the mantle are pure eclogite (Figure 2); they are likely to be attached or surrounded by peridotite and therefore form piclogite packages that can become buoyant as they warm up.

Stoping: A process by which magma intrudes; blocks of wall rock break off and sink.

Large-scale Magmatic Stoping (Daly, 1933); magma rises and shatters, but does not melt the surrounding rocks which sink, making room for more magma to rise. This theory was instrumental in explaining the structure of many igneous rock formations.

THE MANTLE AS AN ENGINE

The type of convection proposed in this chapter differs so dramatically from the type presently being pursued by convection modelers (Appendix) that a different metaphor, or conceptual model, is needed. The pot-on-the-stove metaphor is not appropriate. Large changes in properties due to compression, phase changes, heat exchangers, condensers, pistons and repetitious cycles are essential to the operation of an engine. Thermodynamic consistency, material properties that depend on both temperature and pressure, low efficiency, a way to expel waste heat, and obedience to the Second Law distinguish a real engine from a perpetual motion machine. Mantle convection is often simulated with fluids that have constant properties or properties that depend only on temperature, no large volume or material property changes, no phase changes, inconsistent themodynamic approximations and, sometimes, violations of the Second Law. It is useful to think of mantle convection as a self-consistent thermodynamic cycle or an engine, rather than as an *ad hoc* cycle that is driven by thermal expansion alone at constant pressure (Appendix).

The mantle is a machine that converts energy into mechanical forces and motions. It is therefore an engine. In the eclogite engine the working fluids are not water and steam but basalt, eclogite and magma, and packages of the same. At the top of the system, magma and basalt are converted to dense eclogite by cooling and compression. This 'fluid' is further compressed–adiabatically–by sinking into the mantle, which is, in effect, an infinite heat source. The eclogite package is heated by conduction from the heat bath, and heated by 'internal combustion' (radioactivity). It expands and changes phase causing it to further decompress and rise. At the end of the stroke it discharges some heat and magma through the surface heat sink and the cycle repeats. This process cools the mantle and brings magma to the surface.

As the surface layer thickens by cooling, underplating, intrusion, extrusion and compression, it can become unstable. If the upper part is buoyant, or stiff, the lower portion can peel off or founder (stope), and be replaced by asthenosphere or magma. This stage is accompanied by magmatism. Delaminated material will melt as it warms up by conduction from ambient mantle. This reverses the density contrast and it rises back to the surface; ascent increases the melting and further lowers the density. A second stage of magmatism is the result. This is a form of convection; it can be called *yo-yo tectonics*, an up-down motion controlled from the top and by gravity. Relative buoyancy is mainly controlled by phase changes, gabbro to eclogite and melting. Density variations associated with these phase changes are an order of magnitude greater than can be achieved by thermal expansion, and they can be more abrupt than normal thermal expansion. Industrial engines do not work well without the large volume changes associated with pressure and phase changes.

Normal thermal convection is driven by thermal expansion, which gives a gradual decrease of density with a rise in temperature. The eclogite engine is driven by more abrupt changes in density at the basalt-eclogite-melt phase boundaries and the rapid decrease in density of the melt phase with decreasing pressure. The basalt-eclogite transition occurs over a finite interval (Ringwood and Green, 1966) of temperature and pressure but it is still abrupt, and large, compared to thermal expansion-induced density changes, and it is added to other thermal effects. Sinking eclogite trapped at a chemical or phase boundary will be denser than the surrounding mantle. It therefore takes some time to heat up to neutral buoyancy and more time before the accumulated buoyancy causes it to rise. Parts of a deep layer can also be displaced upwards by subsequent subduction and delamination, rather than local buoyancy and an RT instability. As a displaced blob rises, it may melt. This kind of convection does not require a superadiabatic gradient or a deep thermal boundary layer. The ascent is terminated by density or strength barriers. The further step of intrusion and eruption requires the cooperation of the outer shell. Underplating, or ponding, may lead to crustal thickening and future instabilities and melt release.

UNDERLYING ASSUMPTIONS

A basic premise of this chapter is that the large volume changes associated with partial melting, and other phase changes, and the large density differences between eclogites and peridotites may be as important as thermal expansion in mantle dynamics. Another premise is that melt volumes and melt compositions are more a function of mantle composition and lithology than of absolute temperature, and that the mantle is lithologically heterogeneous. This contrasts with the homogeneous mantle and melting models of Langmuir et al. (1992) and McKenzie and Bickle (1988) that dominate the current petrological literature (Appendix and other papers in this volume; for a complete guide to the literature of mantle geodynamics and geochemistry see http://www.mantleplumes.org/). An important but often over-looked result of petrology and phase relations is that melt compositions can be similar for a wide variety of starting material (Presnall, 1969; Jaques and Green, 1979). In contrast to standard models, absolute temperature variations may have almost nothing to do with the so-called global arrays of geochemical parameters, the volumes of basalts erupted at melting anomalies, or the driving forces of mantle convection. The assumptions underlying conventional models of mantle convection and mantle geochemistry are given in Tackley et al. (2005), Anderson (2005a, b) and in the Appendix.

Large magma fluxes can be produced from eclogite-rich regions of the mantle without high absolute temperatures (Takahashi et al., 1998; Yasuda & Fujii, 1998). Takahashi & Nakajima (2002) suggested that some Hawaiian basalts were produced from large eclogite blocks with an inferred excess potential temperature of about 100°C. On the other hand, high potential temperatures for Hawaii are implied with the standard homogeneous peridotitic source assumption, and the assumption that a given magma is a unique product of a given composition and temperature.

It is mantle heterogeneity that mainly explains the diversity of midocean ridge basalts, and melt volume and elevation anomalies. There are numerous ways to introduce heterogeneity into the mantle but I concentrate on the basalt and eclogite blobs that get into the mantle by processes other than subduction. Convective homogenization, another theme of standard models, is of secondary importance (e.g. Meibom and Anderson, 2003). Plate tectonics, recycling, crustal stoping, and differentiation serve to dehomogenize the mantle, and these are ongoing processes.

RAYLEIGH-TAYLOR (RT) INSTABILITIES

In a cooling fluid the surface thermal boundary layer becomes unstable at relatively short times and thicknesses. Part of the outer boundary layer of the mantle is compositionally buoyant and this extends the conduction gradient to greater depths. When the temperature dependence of viscosity is taken into account the boundary layer becomes thicker still. The buoyancy and extra stiffness requires more force, and more cooling time, to remove it. The average thickness of the boundary layer of the mantle has been estimated to be 280 km (Kaula, 1983), whereas the maximum thickness for an homogenous fluid having mantle properties is more like 100 km. Kaula (1983) also estimated that the potential temperature at the fully convective depth was about 1410 ± 180 ^{o}C, which is higher than petrological estimates derived from magmas–which may come from shallower depths–but is consistent with the thicker conduction layer. Under these conditions the lower part of the boundary layer may peel off, delaminate or founder, leaving the colder, lower density and stiffer part behind.

With modern estimates of mantle and melting temperatures [http://www.mantleplumes.org/Temperature.html], and their variations, it is difficult to avoid localized melting and the partial melting explanation of the asthenosphere (e.g. Anderson and Sammis, 1970, Lambert and Wyllie, 1970). The mantle is hotter, and solidi can be lower and more variable, than generally assumed. The shallow mantle is replenished from above, in part by the insertion of low-melting mafic material lost from the base of continents. A mass balance calculation can be done for this mechanism. A homogeneous partially molten peridotitic asthenosphere is expected to drain rapidly, but a mantle with variable melting temperatures and unconnected fertile regions can develop melt retention buoyancy. The partial melt explanation for the lowest seismic velocity regions of the asthenosphere does not violate global tomographic studies since these average out the extremes of velocity variations.

MASS BALANCE

Oceanic crust is created and destroyed at about the same rates. It is not clear if this represents a cycle or if oceanic crust is lost from the system over periods of billions of years or longer, to be replaced by melts from the displaced upper mantle. One Ga of subduction is equivalent to a 70 km thick layer of eclogite. This may be in the transition region, and there is no mass balance calculation or tomographic image that prohibits this, or that requires oceanic crustal recycling. Material is also removed from and added to the continental crust, apparently at similar rates.

The total subduction rate of oceanic crust is ~20 km^3/yr. Midplate magmatism is roughly 1-2 km^3/yr, comparable to the rates of crustal delamination and seamount subduction. Scholl (2006) estimates 2.5 km^3/yr for the underside erosion of continents at marine margins and 2-3 km^3/yr of erosion plus delamination at continental collisions [http://gsa.confex.com/gsa/2006AM/finalprogram/abstract_110054.htm]. Some of this may be dragged beneath the continent and underplated, but there is a much closer match between rates of midplate volcanism and continental crustal recycling than there is with rates of oceanic crustal recycling.

Larger and hotter chunks may be involved in the stoping part of the process than would be the case with subduction of normal oceanic crust. This material may be responsible for fertile melting anomalies, in addition to contributing trace element and isotopic inhomogeneity to the mantle. This eclogite cycle serves to fertilize and cool the mantle.

LAMINATING THE MANTLE

If the upper mantle is compositionally variable, convection and mantle temperatures will be quite different from standard pictures involving a single surface thermal boundary layer and a vigorously stirred, high Rayleigh number, mantle. Cooling of the mantle, top-down convection, and plate tectonics can still occur. Small-scale features in the mantle are below the resolution of global tomographic and geoid data, but are resolvable with high-resolution geophysical data (Anderson, 2005a,b, 2006). For example, many robust mantle reflectors occur above 1300-km depth (Deuss and Woodhouse, 2002), implying some sort of stratification. Scatterers and less robust (less reproducible) reflectors imply, in addition, a blobby mantle. All of this detail is volume averaged in global tomographic images.

The density and shear-wave seismic velocity of crustal and mantle minerals and rocks at standard temperature and pressure (STP) are arranged according to increasing density in Figure 1. This approximates the situation in an ideally chemically stratified mantle. P and T effects may change the ordering and the velocity and density jumps but do not change the overall picture. Eclogite can settle to various levels, depending on composition; the eclogite bodies that can sink to greater depths because of their density and size have low seismic velocities compared to other rocks with similar density. MORB-eclogite contains stishovite at high pressure and may sink deeper than other eclogites. Subducted and delaminated material eventually heats up and may rise, even if it does not lie in a TBL, creating a gigantic Galileo thermometer (Figure 2).

The Earth–and the mantle–is stratified by density, composition and phase. It is also stratified by viscosity. A low viscosity zone in the upper mantle–a natural result of the depth variation of P and T–was prominent in early discussions of the structure of mantle flow, particularly plate-driven shallow counterflow. It is important in the present discussion since it explains why hotspots show such small relative motions, and mimic the behavior of fixed thermal anomalies (see Counterflow section in the Appendix).

DENSITY VS. DEPTH

The possible density cross-over, at depth, between eclogite and peridotite (e.g. Anderson, 1979, 1989a,b) is well known but continues to be controversial (Ringwood, 1975, 1976; Hirose et al., 1999; Litasov et al., 2004; Cammarano et al., 2005; Irifune and Ringwood, 1993). The variation of in-situ density of eclogite and peridotite, with depth, is discussed or plotted in some of these papers, which come to contradictory conclusions about the fates of eclogite and slabs in the mantle. A density crossover may lead to the formation of garnetite-rich layers or blobs in the transition zone.

What will a chemically stratified mantle look like to a seismologist? I have tabulated the measured or inferred zero pressure densities and shear velocities of mantle minerals and mineral assemblages in Figure 1. This gives the approximate ordering in an ideally density stratified mantle. If the density contrast between layers is greater than the thermal expansivity times the temperature fluctuations, then the stratification is stable (see Apppendix). Figure 1 shows that eclogites can have lower shear velocities than some peridotites of the same density. In a chemically stratified mantle an eclogite blob can have similar or lower shear velocities than the peridotite that it is in density equilibrium with. A profound chemical or density boundary can have a small, or even a negative, velocity jump. The perceived absence of large seismic velocity jumps in global tomographic models of the mantle, except at 400 and 650-km depths, has led many to suppose that the mantle is homogeneous (e.g. Albarède and van der Hilst, 2002a,b; Bercovici and Karato, 2003). The detection of low shear velocity regions in the mantle might also lead one to speculate that these are high-temperature or partially molten regions, but these could also be compositional. Shear velocity does not correlate well with density.

PROBLEMS WITH ECLOGITE

Can eclogite or eclogite-rich material be an important magma source in the mantle? Two assumptions about eclogite sources are responsible for the perception that it cannot. The first is that eclogite is a well-defined rock type that is denser than the mantle, at all depths (e.g. Ringwood, 1975); it therefore sinks readily and rapidly into the lower mantle and removed from the upper mantle. The second problem with eclogite is that almost complete melting is required in order to yield a basalt. A corollary of these concerns is the belief that partial melts of eclogite will rapidly drain away, as they are predicted to do in a homogeneous peridotite mantle; extensive melting is therefore impossible. A contradictory assumption is that the fertile eclogitic portion of the mantle is distributed as small-scale lamellae or veins, the result of vigorous mantle convection and stirring (Allegre and Turcotte, 1986). Meibom and Anderson (2003) and Anderson (2006) argued that eclogite exists as discrete large blobs in the mantle, and that the homogeneity of MORB is a result of magma blending, and sampling of melt aggregations (SUMA) not chaotic stirring in the solid state. Fertile veins in a shallow peridotite can form a permeable network but isolated fertile blobs need not [see also http://www.mantleplumes.org/LowerCrust.html]. A partially molten diapir may also rise faster than melt can escape (Marsh et al., 1981).

Eclogites and basalts–which probably constitute no more than about 7 % of the mantle (Anderson, 1989a)–enter the mantle as large chunks of oceanic crust and lower continental crustal cumulates and metamorphics. The oceanic crust is cold, thin and initially buoyant; the lower crust is warm and can be thick and dense. In both cases the eclogite, once in the mantle, will be surrounded by warmer refractory peridotite that, in general, has higher melting temperatures and low subsolidus permeabilities. Piclogites are *assemblage* of eclogites and peridotites but the eclogite may occur in dispersed blobs rather than as grains or veins; it can be a chunky stew or gumbo[1] instead of a plum pudding or marble cake, or a rock type. If the chunks are small they are not likely to rapidly sink through the mantle. Under these conditions a fertile diapir can melt extensively as it warms up at depth and as it rises into the shallow mantle (Marsh et al., 1981; Ghods and Arkani-Hamed, 2002). This is sometimes called 'melt-retention buoyancy'. The main difference between this model and melting of peridotite with fertile streaks or grains is the scale and the starting temperature of the mafic components. Extensive melting of eclogite at depth is possible because melts are much less mobile and more dense at high pressure and because the dispersed eclogite is surrounded by the more abundant peridotite. A garnet peridotite or pyrolite source region has centimeter sized fertile streaks while a piclogite source region may have km to tens of km scale fertile patches and these need not be interconnected. Melting reduces the density of

[1] A New Orleans version of a celebrated seafood stew from Provence, made with an assortment of fish and shellfish, onions, tomatoes, white wine, olive oil, garlic, saffron and herbs. The stew is ladled over thick slices of French bread.

eclogite and this density reduction increases rapidly as the blob starts to rise, partly because magma is very compressible. At low pressure, magmas are 10-20 % less dense than their source rock but at depth of orders 400 km the density differential can be a few percent or less (Sakamaki et al., 2006).

A third problem with the eclogite source hypothesis is that high magnesium melts, such as non-cumulate picrites and komatiites, may require an ultramafic and high-temperature source. Highly magnesian melts (high-Mg basalts, picrites, and komatiites) are rare but they are found in several large igneous provinces and in Hawaii. Their rarity implies that the conditions for high-temperature melting and extraction of dense melts are also rare. Mantle temperatures are higher and more variable than usually assumed (Kaula, 1983; Anderson, 2000); melting temperatures are also variable, and can be lower than dry peridotite. Melts derived from eclogite may react with the surrounding peridotite to form secondary pyroxenitic sources (Yaxley and Green, 1998; Sobolov et al., 2005), olivine being consumed by this reaction. The low-melting 'eclogite-component' of the mantle may actually be refertilized peridotite. The 'eclogite' imprint is transferred by melts from eclogite partial melts reacting with and re-fertilizing peridotite, which is then the source of the picrites. Clearly, the diversity and volumes of mantle magmas depends on more than absolute temperature.

GALILEO THERMOMETER

Delamination and buoyant decompression melting mimic the behavior of a Galileo thermometer. A Galileo thermometer (GT) is a sealed liquid-filled container, with a number of suspended objects. It is heated or cooled from the sides by ambient air. The fluid changes temperature as the outside temperature changes, and the suspended objects rise and fall to stay at the position where their density is equal to that of the surrounding liquid. The factor that determines this is buoyancy; the object's mass relative to the mass of the liquid displaced. If the object's mass is greater than the mass of liquid displaced, the object will sink. When the object's mass becomes less than the mass of liquid displaced, it will rise. If the blobs absorb sunlight then they will behave as if internally heated. The relative buoyancies of the blobs change with time because the fluid temperature changes with time; the opposite is the case for mantle blobs.

The masses in the mantle are recycled–initially cold–blobs, inserted from the top, that have different properties from the surrounding mantle. They change density rapidly as they warm up because of thermal expansion, partial melting and other phase changes (Figure 2). The mantle is cooled from the top and heated from within. The blobs are cooled from the top, sink and are heated by the mantle. A blob sinks because it is intrinsically denser than the top of the mantle and because it is cold. It rises as it warms if it becomes buoyant at temperatures lower than ambient mantle temperatures. This will be the case if it is intrinsically less dense than ambient mantle, or if it melts or undergoes phase changes at temperatures lower than ambient mantle. At any given time, there are blobs at different levels, just as in a Galileo thermometer. Basalt and eclogite blobs are expected to contain more radioactive elements than peridotite blobs and therefore to heat up faster by internal heating than cold harzburgite blobs.

There are several differences between the mode of convection being discussed here and GT and Rayleigh-Taylor instabilities. The heating and cooling cycles of eclogite near its solidus are not symmetrical, and the launching of eclogite into the mantle may be caused by tectonic rather than buoyancy forces. Because of adiabatic decompression melting and viscous heating the upwelling may be a runaway process.

The viability of the GT mode of convection and cooling of the mantle depends on the relative densities of the mantle and the blobs. The various parts of the cycle are classical physics problems; growth of instabilities at a cooling surface, sinking and rising of spherical blobs in a viscous fluid, heating of a sphere by conduction and internal radioactivity, and melt segregation in a rising diapir. I concentrate instead on the conditions that will allow this mode of small-scale convection rather than rapid removal of eclogite to the lower mantle, or rapid homogenization by stretching and folding, and the thermal implications.

THERMAL IMPLICATIONS

In a chemically stratified internally heated mantle, with periodic injections of cold slabs and warm delaminated lower continental crust, the temperature gradient will be complex; it will not be a simple adiabat or a conduction geotherm. Deep mantle temperatures can be higher than in a homogenous convecting mantle, and there will be subadiabatic temperature gradients at depth (Jeanloz and Morris, 1986, 1987). Mattern et al. (2005) inferred a subadiatic temperature gradient and a decreasing iron content from 660 to 1300-km depth and significant heterogeneity from 800 to 1300-km depth. This can be understood if this region, not D”, is the graveyard–temporary or permanent–of cold dense MORB-eclogite sinkers (Wen and Anderson, 1997). Cold slabs will displace the surrounding mantle, which may melt (Figure 2).

FATE OF ECLOGITE IN THE MANTLE

A series of papers (see below), starting with Ringwood (1975) argue that eclogite will sink deep into the mantle, rather than being trapped above 650 km depth or at the 1000 km discontinuity, or other shallow levels, as I have argued. These papers assume that the top of the oceanic crust, MORB, is representative of all eclogites and that if eclogite is denser than pyrolite it will sink. In other words, local density differences alone control the fate of eclogite, and pyrolite provides a good density model for the mantle. The fate of eclogite in the mantle depends, in part, on the relative densities between eclogite and the surrounding mantle, and also on the relative sizes of eclogite blobs. The mantle, particularly the transition region, may be denser than pyrolite (Cammarano et al., 2005), and eclogite-rich (Anderson, 1975). The upper mantle contains fertile peridotite, residual peridotite, dunite and various kinds of mafic materials and melts (Figure 1); the dominant lithology probably changes with depth. Plausible end-member lithologies range in density from 3.3 to 3.65 g/cm^3. The upper mantle may involve mixtures or layers of these; the denser lithologies will tend to lie deeper. Convection can still occur in the presence of a stabilizing density gradient, with an accompanying non-adiabatic temperature gradient, but it is unlike any kind of mantle convection that has been treated to date. It is Top-Down convection, similar to a Galileo thermometer, composed of sinking and rising blobs. It is not the whole-mantle vigorous convection that homogenizes the mantle, as is often envisaged, but it can cool the mantle, and can deliver melt to the surface.

Hirose et al. (1999) suggest that *if* MORB could penetrate the buoyancy barrier between 650 and 720 km depths (in a pyrolite mantle) it could sink further into the lower mantle. This assumes that eclogite is unmodified as it sinks into the mantle (i.e. it stays unaltered and does not lose a silica-rich melt), and that some mechanism exists for dragging it through higher-density mantle. It also assumes that pyrolite is a good model for the TZ and lower mantle. Lee et al. (2004) found that the lower mantle is 2 – 4% denser that pyrolite, which would reverse the conclusion of Hirose et al. (1999) regarding the ability of eclogite to sink into the lower mantle.

Nishihara et al. (2005) conclude that MORB-eclogite is denser than pyrolite in the transition region; other eclogites are less MgO- and SiO_2-rich and should be less dense. Pyrolite may be a good model for the shallow mantle but it is not a good match to the seismic velocities in the TZ and lower mantle (Lee et al., 2004; Cammarano et al., 2005). A pyrolite lithology has higher seismic velocities in the TZ and below, than are obtained from seismology. This suggests more garnet, and perhaps more FeO, in the TZ than in pyrolite. The seismic data also suggest a gradient with depth in the TZ, with, possibly, more high-pressure olivine and FeO-rich phases at depth (Mattern et al., 2005). The TZ and the phase change discontinuities vary laterally (Ishii & Tromp, 2004; Deuss et al., 2006), as expected if the TZ is the dumping ground, even temporarily, of eclogite. Other evidence for chemical stratification and barriers near 650- and 1000-km depths is summarized in Anderson (2005a,b) [see also the Appendix and http://www.mantleplumes.org/Eclogite.html].

RELATIVE DENSITIES

The viability of the GT and fertile blob models depends on density contrasts between eclogite and ambient mantle (e.g. Yasuda & Fujii, 1998), the size and viscosity of the eclogite components (e.g. Meibom and Anderson, 2003), and the temperature at which a density reduction sets in due to partial melting. The above papers focus on the densities and fates of subducted oceanic upper crust. Other eclogites in the mantle have deep crustal protoliths and some are secondary, forming from reactions of eclogite melts with peridotites. Some eclogites may be cumulates that have been invaded by basaltic melts. Subsolidus eclogite is denser than most peridotites but may have a significantly lower solidus temperature; some eclogites–restites–have had the lower melting components removed. Eclogites are therefore variable in density and in the temperature of initial melting. Eclogite is commonly assumed to have a density of 3.5 g/cm^3 (Leitch & Davies, 2001), or even greater (Sleep, this volume) but this is too high to be a good average. Eclogites have measured or inferred densities in the range 3.30 to 3.65 g/cm^3. Packages of eclogite and peridotite–piclogites–will have even lower densities.

The fate of eclogite also depends on ambient mantle density and viscosity. Estimates of upper mantle densities based on abyssal and obducted peridotites and xenoliths from the continental lithosphere range from 3.30 to 3.40 g/cm^3. Dunites have densities of 3.35 to 3.50 gm/cm^3. These may not be representative of the density of the deeper mantle. An intrinsic density of 3.3 g/cm^3 may be appropriate for the depleted residue that accumulates at the top of the mantle but is probably too low to represent the bulk of the mantle below some 200 km. This range of intrinsic densities corresponds to a temperature variation of 2000 ^{o}C and strongly implies that the mantle must be chemically stratified. Available samples of mantle rocks may be biased toward the more buoyant ones, and the shallow mantle.

Kogiso et al. (2003, 2004) and Petermann & Hirschmann (2003) calculate an STP solidus density of eclogite that is only 2–4 % denser than many garnet peridotites and pyrolite and substantially less than the 4·5–6 % contrast that is commonly assumed. Additionally, it is not just the local density contrast that controls the fate of eclogite in the mantle. It is the integrated density and the volume of the sinking or rising mass, and the surrounding viscosity. There is also a spread of densities of about 3 % among various types of lherzolites and dunites. It is therefore not obvious that eclogites are substantially denser than the upper mantle or transition region. In fact, the transition region may be eclogite-rich (Anderson, 1979, 1989a). Eclogites are also likely to contain more radioactive elements than upper mantle peridotites.

An upper-mantle blob with 10-20% subsolidus eclogite of density 3.44 g/cm^3 must be 50–300°C hotter than an ambient pyrolite mantle to have neutral buoyancy; this would imply partial, or even complete, melting of the eclogite. Smaller excess temperatures are required if ambient mantle is more fertile and iron-rich, or if the eclogite is garnet-poor. No excess temperature is required if eclogite is sufficiently above its solidus. Eclogite can be less dense than commonly assumed and the excess temperature required for a buoyant blob to support a given eclogite load is markedly less than usually assumed. An eclogite bearing blob may conceivably become buoyant with no excess temperature, e.g. having potential temperatures in the range inferred from midocean ridge basalts and from geophysical data. But it needs to be stressed that fertile streaks or blobs in the mantle need not be intrinsically buoyant in their own right in order to be involved in mantle melting.

CONSISTENCY WITH HIGH-RESOLUTION SEISMOLOGY

Global tomographic images have little resolving power but higher resolution and spectral domain and correlation techniques are available. The mantle contains many seismic discontinuities above 1300 km depth (Deuss and Woodhouse, 2002). Deep reflection seismology shows a complicated and variable structure with single and double reflections ranging from 640 to 720 km depth near the base of the transition region (Deuss et al., 2006). The spectral amplitude of the lateral density variations of the mantle exhibits a maximum around 600 km depth. The density of the mantle transition zone has negative or zero correlation with shear- and compressional-wave velocities; TZ velocity-velocity and velocity-density correlations are completely different from the overlying and underlying regions (Ishii and Tromp, 2004). Gu et al. (2001) showed that there is a distinct change in seismic velocity variations, near 650 km depth, with large variations above and small variations below. All of this is inconsistent with thermal variations alone or with whole-mantle convection. Subduction of cold eclogite into the TZ will lower the shear-wave seismic velocity there (Figure 1), but will have little effect on the density until it warms up (the eclogite displaces similar density material, as in a Galileo thermometer). An eclogite-rich TZ may also explain the velocity jumps at 410 km and 650 km, which are too small and too large, respectively, to be entirely due to phase changes in pyrolite. Garnet does not undergo a phase change near 400 km so it dilutes the jumps created by phase transitions in olivine and orthopyroxene at depths of 400 and 500 km. The low shear-wave seismic velocity of eclogite and its low melting temperature will create large tomographic variations in the transition zone, particularly since the distribution of eclogite is not expected to be uniform.

Seismic profiles calculated for a peridotite mantle, compared to PREM and other seismic models, have lower velocities above 400 km, larger jumps near 410 km, lower transition zone gradients, lower jumps around 650 km, and strong velocity gradients directly below (Cammarano et al, 2005). The lower mantle is denser than pyrolite (Lee et al., 2004). The upper mantle is full of seismic scatterers (Anderson, 2006) and sometimes there is a low-velocity zone atop the 410 km discontinuity (Song et al., 2004). The nature of the upper mantle discontinuities is not constant from place to place (Deuss et al., 2006) but there is little evidence that these variations are due to large variations in temperature. These observations are all consistent with upper mantle with variable amounts of eclogite

(Anderson, 1979, 1989a; Anderson and Bass 1986). The seismic observations are inconsistent with a homogeneous pyrolite mantle with recycled material sinking readily to the core-mantle boundary.

The evidence summarized above and in my earlier papers, supports the idea that the transition zone is a density filter, and is a barrier to subduction. This evidence is much stronger than the hints of through-going convection based on visual inspection of a few selected low-resolution tomographic cross-sections. The proposal that eclogite, volatiles and radioactive elements are trapped in and above the transition region is the opposite of the standard model and the model of Bercovici and Karato (2003), which assumes that water and LIL, including U, Th and K, are primarily in the lower mantle and that ascending–rather than descending–plumes are filtered by the TZ. In those models the upper mantle is heated strongly from below, while in the model proposed here, convection is driven by secular cooling of the mantle, slabs and upper mantle radioactivity. The mantle is fertilized from above, not from below. The fertile parts of the TZ are predicted to start out cool, and then warm up. High-resolution seismic studies, and those that measure impedance rather than just shear velocity, are consistent with a stratified and blobby mantle and with the predictions of the GT mode of convection.

DISCUSSION

A mantle with phase changes at various depths, and with variably melting, large-sized constituents, exhibits behavior that is completely different from a homogenous or layered fluid or a well-mixed solid. It is not well approximated by any of the reservoir and box models or convection calculations that have been done to date (Appendix). Fertile blobs rise or fall, depending on their size and the time since they were inserted into the mantle. Upwelling that involves melting is a runaway process because of adiabatic decompression melting and the high compressibility of the melt phase. Upwellings do not have to have high absolute temperatures and they do not have to start in a deep thermal boundary layer heated from below. Their isotopic signatures do not evolve from a depleted-MORB or primordial starting signature. The predicted lithologic variability of this kind of mantle is in agreement with various kinds of petrological and geochemical data (Anderson, 2005; Agranier et al., 2005; Jull and Kelemen, 2001; Natland, 1989 and this volume). The GT model replaces the isolated reservoir models of mantle geochemistry and the one- and two-layer models of geodynamics, which are incompatible (Albared and van der Hilst, 2002a,b).

The upper mantle may be composed of transient blobs as well as being density stratified. I have concentrated on the density differences between various eclogites and between eclogite and ambient mantle (which is not necessarily pyrolite). There are equally important density and fertility differences among natural peridotites; the density differences correspond to a temperature range of at least 300˚C ΔT in a lherzolite mantle (David Green, personal communication). The shallow mantle may be the refractory residue after melting out of basaltic crust, and therefore intrinsically buoyant (the perisphere). Low density peridotites (depleted or iron-poor) may never get cold enough to sink very far, or for very long, in a mantle composed of denser peridotites or piclogites. At depth, it would be difficult to make subsolidus eclogites buoyant within depleted lherzolite or pyrolite, but there is no reason to suspect that these lithologies represent ambient mantle at all depths. If the mantle is stratified by intrinsic density, then denser material than pyrolite may occupy the deeper upper mantle and the transition zone, as well as the lower mantle. In the extreme case the mantle is continuously stratified by density, melting point and volatility throughout the accretional, differention and cooling processes. Some of this may be irreversible.

The shallow fertile blob source for melting anomalies has many interesting consequencies and satisfies data regarding relative motions of hotspots (e.g. Cuffaro & Doglioni, 2007, this volume). Shallow sources and shallow return flow imply faster plate motions and migration of ridges over fertile mantle. This mode of mantle convection solves several long-standing problems, including how ridges can be continuously productive and how the asthenosphere survives if it is constantly leaking. This problem has been used as evidence against the partially molten asthenosphere hypothesis and in favor of the plume hypothesis [http://www.bioedonline.org/news/news.cfm?art=2687]. However, ridge migration and a supply of material from internally cycled eclogite prevents the asthenosphere from slowly wasting away. The fact that all rifts and margins are not equally magmatic has also been used as an argument against non-plume models (see papers in this volume). A blobby mantle solves this problem. It also addresses issues recently raised against all current models of mantle dynamics (Tackley, 2006). An unresolved issue is the melting behavior, and therefore density, of a plausible range of eclogite compositions as a function of pressure. The type of convection and heterogeneity discussed here provide challenges to geodynamic modelers and seismologists. It may require a fundamental change in the assumptions and ground rules behind current geodynamic modeling (Appendix).

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Jim Natland reviewed the preliminary manuscript and stimulated a complete rewrite, as did the reviewers of the present chapter. My interest in delamination was started by a series of discussion seminars at Rice University and by conversations with Alan Levander, Adrian Lenardic, Richard Gordon, Gene Humphreys and Cin-Ty Lee. The manuscript benefited enormously from comments and reprints by Dave Green, John Hernlund and Jim Natland. The information on density borrows heavily from papers by Cin-Ty Lee and Dave Green.

APPENDIX

COMPARISON WITH OTHER PROPOSED MECHANISMS

The reviewers of this chapter suggested that I compare the Galileo Thermometer mode (GT) of mantle convection with previous models, stressing that which is new and different. Since this does not fit in well with the main text I collect this information, plus other background material, in this appendix. I also review the PLATE model.

Until recently, most convection simulations have ignored melting, gravitational differentiation, delamination, heterogeneous fluids and density stratification; these form the essence of the present chapter. GM convection has some similarities with previous studies in the areas of planetology and petrology where differentiation and gravitational stratification are well known concepts. The present mantle is often considered to be homogeneous, or, at most, to consist of only two layers, either separated at 650-km or at the D" boundary. GT assumes that material inserted into the mantle as large blocks will tend to settle to a depth of neutral buoyancy, usually at a phase boundary of ambient mantle but also at chemical boundaries. On average, the intrinsic density of the mantle increases with depth. This contrasts with models that assume vigorous stirring and rapid homogenization.

Tackley and Stevenson (1998) discuss a mechanism for spontaneously creating buoyant upwellings by perturbations in the melt content of a homogeneous layer of the upper mantle. This process has been termed a "buoyant decompression melting instability" (Raddick et al., 2002), a potentially important source of intra-plate volcanism. A similar mechanism, but at lower temperature, will operate if the melting temperature of the mantle varies, as in GT. John Hernlund discusses a "drippy mode" of convection that is related to the RT mode of delamination [http://geodyn.ess.ucla.edu/~hernlund/smscconv.html].

The melt instability discussed here differs from those treated above, in the following ways;

1. the instability is triggered/nucleated by cold low-melting blobs heating up toward ambient temperature, not by a statistical variation in temperature or melt content
2. much more melt is created by a given rise in temperature than in a uniform peridotite layer
3. the fertile blob is surrounded by subsolidus impermeable low homologous temperature peridotite
4. there is a natural way to replenish the upwelled materials; the melting zone is replenished by downwelling instabilities starting within and below the overlying plate
5. the instability can operate in an isothermal mantle
6. the instability can initiate at lower than ambient temperature
7. the scale is set by the scale of the heterogeneity

Many authors have suggested that material from a deep, dense layer can be displaced upwards by sinking slabs, usually thinking of the D" region above the core. If the upper mantle and TZ are chemically stratified then the upward displacement may result in melting–and more buoyancy–which is unlikely for deeper dense layers.

A geological analogue to the GT and 'yo-yo' mechanism is Dunk Tectonics (Brueckner et al., 2004), used to explain the cycle that takes crustal material to ~150 km depth and then returns it to the surface as Ultra-High Pressure (UHP) rocks containing coesite and eclogite. Dunk Tectonics (DT), in fact, may be a minor side effect of GM. In both cases, shearing, thrusting and entrainment may be more important that strictly fluid dynamic instabilities, although both have been analyzed in terms of RT and density instabilities.

The reheated blob–and yo-yo–mechanisms (Figure 2) superficially resemble the thermochemical oscillatory dome mechanism of (Davaille 1999a, b) that occurs in chemically stratified fluids. The system she studied experimentally consists of superposed layers of homogeneous fluids with different viscosities and densities, which convect in response to heating from below. The effects of pressure, melting and phase changes on material properties, and internal heating, are not simulated; these are serious shortcomings. Upwellings from the deeper denser layer are caused by very high temperatures, and are a form of thermochemical convection. This differs from the current model in that temperatures must be even higher than in models of thermal convection since the intrinsic density contrast must be overcome. The most important parameter controlling the dynamics of layered convection, and accretional differentiation, is the "buoyancy number" **B**, defined as the ratio of the chemical buoyancy (the intrinsic density difference between layers or blobs) to the total thermal buoyancy due to the temperature contrast across the tank. This geometry can produce oscillatory domes, large thermochemical plumes, or families of small secondary plumes, depending on the density contrast. When the lower layer is thin and potentially buoyant, thermochemical plumes rise from the interface and entrain filaments of chemically dense material from the lower layer. A second mode consists of large "domes" of fluid that oscillate over the whole height of the tank. A third mode consists of small plumes that are generated at the top of a rising dome.

In the *eclogite engine*, intrinsic density contrasts are overcome by phase changes and **B** varies with time and depth. For certain parameter ranges, and compositions of the differentiation products, the layering can be irreversible (Anderson, 2002a). In GT the fertile blobs start out cold, but they are much warmer than subducted oceanic crust. Heating is by a combination of heat conducted into the blob from the surrounding fluid, and, over a longer time scale, by internally generated heat in the blob (Figure 2). Mantle heterogeneities are introduced from the top, not the bottom. The GT mode owes its existence to smaller scale chemical heterogeneity and variations in intrinsic density, thermal expansion, viscosity and melting intervals. Because of the effects of pressure on the coefficient of thermal expansion and melt density, the buoyancy parameter decreases with depth, making irreversible chemical stratification more likely than reversible stratification, except in the upper parts of the mantle. Secular cooling can also reverse the direction of net mass flux; for example, from upward to downward transport of mafic material. **B** is the most important parameter in GT and in convection calculations that include differentiation and the possibility of mantle layering–reversible and irreversible. The homologous temperature replaces the absolute temperature as the fundamental temperature scale. In most simulations, the Rayleigh number and the absolute temperature are the dominant parameters.

A layered mantle can convect and can remove mantle and core heat but this convection does not resemble the broad cellular convection, with steady 2D convection cells, that is depicted in textbooks, and that is perceived to be the source of conflict between geochemistry and geophysics (e.g. Bercovici and Karato, 2003; Anderson, 2005a,b; Albarede and van der Hilst, 2002). Superimposed on large scale convection and subduction driven flow are various smaller scales of convection. In the present model, this scale is set by petrology and tectonics, not by fluid dynamic considerations, as in most treatments of small-scale thermal or melt-induced convection.

Bercovici and Karato (2003) argue that the mantle is chemically homogeneous since they see no evidence for chemical boundaries and thermal boundary layers. They rely on selected low-resolution global seismic tomographic cross-sections and qualitative visual tomographic interpretations, and ignore the high-resolution seismic evidence for a complex mantle, with multiple reflections and scatterers (e.g. Anderson, 2005b). They do not define how they would recognize layered convection if it existed. They state that layered mantle models "suffer from dynamic inconsistencies…most importantly, because any enriched lower layer has most of the heat producing elements, the depleted overlying layer is heated almost entirely along its base…" and this kind of convection can be ruled out. But there is no evidence for an enriched lower mantle; this is purely an assumption, and violates mass balance calculations (Anderson, 1989a). A radioactive-rich lower mantle will overheat and overturn, resulting in most of the radioactivity being transferred to the upper layers. But it is more likely that the large-ion radioactive elements were zone-refined into the shallow mantle during the planetary accretion stage. If most of the radioactivity is in the crust and upper mantle, then layered models cannot be ruled out so casually. The arguments of Bercovic and Karato (2003) attack artificial and unphysical models and do not discuss thermodynamically self-consistent models, plausible heating modes and quantitative tomographic interpretations.

Yo-Yo Tectonics

Eclogite that has settled into the upper mantle will be trapped at various depths above the 650-km discontinuity (Anderson, 1989b; 2005; Hirose et al, 1999); molten eclogite will density-equilibrate near 400-km (see Sakamaki et al., 2006 and Figure 1). Only the coldest, thickest, driest and most SiO_2-rich eclogites are candidates for sinking deeper than 650-km. To do so, they must first decouple from any attached depleted harzburgite.

The large density jump at 400-km depth in ambient mantle is mainly due to phase changes in olivine and orthopyroxene. Eclogite does not undergo phase changes at the same depths. Some eclogites and magmas are therefore likely to be trapped at 400-km depth, causing a low shear velocity anomaly. Large eclogite blobs are more likely to control their own fate than are the fertile veins in peridotites. Any chemical stratification involving cold eclogite and other low melting point blobs is temporary since the melting associated with heating involves large density changes. Hot or partially melted eclogite will equilibrate at shallower depths, and may rise and spread out beneath the lithosphere (Figure 1).

Dry peridotitic melts and komatiites may be denser than the mantle between 250 and 410-km. Eclogite-rich blobs warming up in a peridotite subsolidus mantle probably can retain their melts, although they will react with the peridotite. There is a density reduction as garnet in eclogite is converted to magma but initially the melts have little buoyancy (Figure 1). The complete cycle is shown schematically in Figure 2.

The Plate Process

A generalization of plate tectonics that emphasizes the role of near surface phenomena and features such as Plates, Lithosphere, Asthenosphere, Tectonics and Eclogite blobs, is called the PLATE paradigm; cooling of the surface, extensional stress, recycling, homologous temperature, lithospheric architecture, counterflow, delamination and the buoyancy parameter are the important parameters and processes. Plate motions drive a shallow mantle counterflow–counter to plate motions; the thickness and viscosity of the counterflow channel are important parameters. In the thermal–or standard–models the main parameter–essentially the only parameter–is absolute temperature; plate motions and mantle convection are driven from below. The present chapter elaborates on one part of the Top-Down (Anderson, 2001, 2005a), or PLATE, paradigm. A key element is that the volume of melt produced from a given volume of mantle is controlled by the fraction of low-melting material, not by the absolute temperature. The temperatures in D" and the conditions at the core-mantle boundary are interesting but they have little to do with surface magmatism, including large igneous provinces (LIPs) (Anderson, 2005a). Other elements of the PLATE process have been developed in Foulger et al. (2005), Foulger (this volume) and Anderson

(2001, 2002a,b, 2005a,b, 2006). Counterflow calculations are given in Harper (1978) and Chase (1979); these predict small relative motions of entrained blobs and therefore of surface hotspots. Earlier work on the stability and composition of near-surface layers (stoping or delamination potential) is given in Daly (1933), Ringwood and Green (1966), Green and Ringwood (1967) and Ringwood (1975), which give detailed discussions of these themes. What emerges is a novel form of convection that is not captured by laboratory and computer simulations of convection in a homogeneous or layered fluid, mainly heated from below or uniformly heated from within. It is more akin to a GT. Key aspects of the Top-Down process, some extracted from the above papers, are described below.
[see also http://www.mantleplumes.org/TopPages/SelfOrganisedTop.html, http://www.mantleplumes.org/Convection.html].

High pressure stiffens the deep parts of a silicate planet and low temperature stiffens the outer shell. If there are continents and stiff and buoyant regions at the surface, they control the pattern of convection, the cooling rate, the aspect ratio of convection, and the lag between heat generation and surface heatflow. The cooling surface layer and the surface boundary condition is the active and self-organizing part of the system and it drives and organizes the motions in the passive interior, including the shallow counterflow, which is opposite to plate motions. The average thermal gradient in the interior will be subadiabatic under these conditions.

If the 'fluid' is a mixture of several components with different intrinsic densities, melting behavior, phase changes and other properties it can evolve to a chemically layered system. The deeper part of the stratification will be irreversible if the intrinsic density contrasts or pressure are great enough. If there are a large number of layers, cooling is primarily by conduction, unless part of the cool surface layer can detach (subduct, delaminate, founder, stope) and cool off the interior; these latter methods for cooling the interior and providing melt to the surface are seldom addressed.

The processes of planetary accretion, melt extraction and accretional differentiation placed most of the radioactive elements into the materials that became the crust and upper mantle. The processes that heat and cool the mantle, and create and dissipate buoyancy, are most effective at the top.

Counterflow

Arguments against plume-alternative mechanisms are usually based on the propagating crack idea, or on the perceived fixity of hotspots. Paul Tackley (2006), in reviewing mechanisms for hotspots, recently posed the following questions: Why do the Pacific hotspots exhibit little relative motion as the plate moves over them? If they are caused by propagating cracks, why do all the cracks propagate at the same rate? Plate boundaries tend to be linear; why then are flood basalts, and hotspots that occur at spreading centers, not linear, following the plate boundary?

Shallow counterflow mechanisms explain the small relative motions of the Pacific hotspots, and the measurable relative motions between groups of hotspots on different plates. Heterogeneities in the upper mantle flow in the opposite direction to the overlying plate and at a small fraction of the plate velocity. This mechanism gives tracks, on a given plate, that are parallel. Statistical compatibility tests show that the ages and trends of volcanic chains are incompatible with fixed hotspots (Wang and Liu, 2006). Furthermore, hotspot rates are consistently lower than those predicted by best-fitting absolute plate motion models. This may be explained by fertile spots in the mantle moving systematically opposite to the plate motion, as predicted by counterflow models of mantle convection (Harper, 1978; Chase, 1979). The locations, trends and relative motions of melting anomalies are then due to asthenospheric flow that is almost exactly opposite to plate motion, as in the lubrication models of plate tectonics (Harper, 1978) and in recent models of relative motions (Wang and Liu, 2006). Dense sinkers that bottom out at upper mantle phase changes may also exhibit little relative motion, particularly if they return to the surface rather quickly. Cuffaro & Doglioni (2007, this volume) and Norton(2007, this volume) have shown that shallow sources satisfy plate reconstruction data.

Plate boundaries and fracture zones tend to be linear, as do so-called hotspot tracks. Many flood basalts, and hotspots are not linear because they occur at triple junctions, and are transient in nature. Fertile streaks in the upper mantle do not need to follow plate boundaries nor do they need to be linear. The intersections of fertile blobs with ridges, triple junctions reactivated sutures, and with new plate boundaries, and the configurations of tensile stress domains, are what control the location and shape of melting anomalies. Nevertheless, much of the fabric of the seafloor is parallel to plate motions, and linear volcanic chains are expected for incipient–and dying–plate boundaries. It is not obvious why the linearity of long-lived volcanic constructs, and the non-linearity of short-lived constructs and those at triple junctions are arguments against plate tectonic control and shallow origins of 'midplate' phenomena (e.g. Tackley, 2006). Likewise, it is not clear why the absence of LIPs along some extending margins is an argument for plumes.

Ground rules

Many concepts and scalings of mantle dynamics are based on small-scale laboratory experiments and simplified theory motivated by experiments; these have become so engrained in the collective consciousness of geodynamicists that they are taken for granted. Tackley et al. (2005) summarize these concepts as 14 rules that dictate the current directions of fluid dynamic research. These rules do not include the issues raised in this chapter and in other recent papers that stress the top-down nature of mantle convection, self-consistent thermodynamics, and the onset of phase-change induced instabilities (Anderson, 2005a,b). These new concepts can be formalized into 14 alternative rules;

(i) The basic structure of the mantle may not involve steady flow, or convection cells with narrow boundary layers and an adiabatic interior
(ii) The volume- and pressure-dependence of thermal properties are as important as temperature dependence
(iii) The planform of three-dimensional convection may be controlled by lithospheric architecture and mantle chemical heterogeneity rather than by fluid dynamic instabilities of a homogeneous fluid; plates and slabs may organize or even control mantle convection
(iv) The volume-dependence of properties such as viscosity, conductivity and thermal expansivity makes deep mantle upwellings broader, increases the horizontal wavelength and time constants of deep flow, and contributes to irreversible chemical stratification; pressure can suppress the importance of lower mantle convection in determining surface processes
(v) The radioactive elements are not uniformly distributed throughout the mantle nor in the various components
(vi) Plate tectonic forces, including magma fracture, can break the lithosphere; plates are

not a rigid lid nor a uniform or permanent boundary condition

(vii) Phase transitions, including eclogite-basalt-magma, in a multicomponent mantle can induce instabilities and permanent or episodic chemical layering and blobby-type heterogeneity

(viii) Continents, including their motions and the inherent instability of thick continental crust cannot be ignored

(ix) Lithological variations, and variations in heat productivity and viscous dissipation can cause large heating locally.

(x) Convection models must be thermodynamically self-consistent.

(xi) Lithologic variations affect seismic velocities; current scaling parameters between temperature, density and seismic velocity are inadequate and often misleading.

(xii) The locations and scales of mantle instabilities may be set by the locations and scales of lithologic heterogeneity rather than by fluid dynamics of a homogeneous fluid.

(xiii) The scales of chemical heterogeneities in the mantle control their fate and their geochemical signatures

(xiv) The upper mantle is close to or above the melting temperature of some of its components

Mantle convection is a non-linear process with outcomes that depend strongly on initial and boundary conditions, and other assumptions; actual mantle dynamics may violate the collective consciousnesses and conventional wisdoms of geodynamicists operating by the rules of the current paradigm.

Geochemical Models

Geochemistry cannot constrain the depth or locations of the sources of basalts. They are usually cast as boxes or reservoirs, connected by arrows, and other information is used to suggest a location. In the standard models of mantle geochemistry, the long-wavelength, or dispersed, parts of mantle heterogeneity (melting anomalies) are “best accounted for by the presence of mantle hot spots” (Agranier et al., 2005). These in turn are attributed to the return of deeply subducted ancient oceanic crust from the lower mantle back into the upper mantle (Hofmann and White, 1982). The injection of lithospheric plates continuously introduces differentiated material into the mantle that either sinks to the core-mantle boundary (Hofmann and White, 1982) or is mixed into the shallow mantle (Christensen and Hofmann, 1994). A common assumption is that, upon subduction, 5–10 km thick sequences typical of oceanic crust will be folded within thicker layers of oceanic lithosphere and ambient mantle (Christensen and Hofmann, 1994), thereby creating a homogeneous source. It is considered inescapable that recycled material is multiply stretched and refolded by mantle convection (Agranier et al., 2005), and that old plates find their way back to the asthenospheric upper mantle, where they become an intrinsic part of the well-mixed MORB source (the so-called convecting upper mantle). Delamination of subcontinental lithosphere and metasomatism at subduction zones, are also assumed to create heterogeneities that mantle convection repeatedly folds and stretches as in the popular marble cake mantle model (Allegre and Turcotte, 1986).

Isotopic mantle heterogeneities along mid-ocean ridges that cannot be attributed to stretching and folding of heterogeneities by convection are attributed to overprinting of the mid-ocean ridges by injections of deep mantle (Hamelin et al., 1984); melting anomalies represent hot material from the deep mantle.

In contrast to this type of model is the Statistical Upper Mantle Assemblage model (SUMA) in which the mantle is intrinsically heterogeneous at all scales, but small-scale heterogeneity is averaged out by the melting and sampling process at ridges and is mainly evident only in smaller scale samples–ocean island volcanoes, seamounts and xenoliths–and dense sampling programs (Meibom and Anderson, 2003). Larger scale heterogeneities, such as chunks of delaminated crust, are not averaged out or mixed back into 'the convecting mantle'. They can be resampled along the global spreading ridge system and as midplate melting anomalies, without being particularly hot (Anderson, 2005a,b). Chemical heteorogeneity and magma volume anomalies are due to the heterogenous nature of the mantle, not primarily to temperature variations. Relatively fixed hotspots are due to the slow counterflow of the asthenosphere (Harper, 1978; Chase, 1979).

Usual treatments of terrestrial magmatism also consider that there are only three possible causes for melting in the mantle: decompression, heating and insertion of volatiles. If parts of the upper mantle are already above the solidus, then adiabatic ascent or an increase in temperature can increase the amount of melting. The presence of water may significantly reduce the melting temperature but the degree of melting is small between the wet and dry solidi due to the high solubility of water in melt. A partially molten asthenosphere may develop melting instabilites, driven by thermal, melt and depletion buoyancy. Asthenosphere can also passively upwell in response to spreading, and to the delamination of continental crust or lithosphere. Volcanism associated with melting instabilities and shallow passive upwelling of fertile mantle are two mechanisms for creating melting anomalies. It is almost universally assumed, however, that large amounts of melting require much higher than average mantle temperatures. Attempts are then made to rationalize the major element chemistry of 'hotspot' magmas in terms of larger degrees and depths of melting due to increased temperatures (Langmuir et al., 1999),

There are other mechanisms for increasing the melt content of a given region of the mantle, and for initiating a melt or buoyancy instability. Variations in melting temperature–or homologous temperature–and lithology can cause instabilities. These instabilities are triggered by cold fertile blobs that are warming up to ambient temperature, rather than by regions that are hotter than ambient mantle. They are buoyant because they contain a large melt fraction, or a large depleted peridotite component, or both.

Fertile blobs vs. hot plumes

LIPs are produced at high eruption rates but there is no evidence that they are produced from particularly hot mantle (Clift, 2005). Hot upwellings would be expected to produce highly magnesian lavas but these are rather rare in LIPs. The locations of LIPs along mobile belts, sutures and triple junctions suggests that shallow mantle fertility, release of ponded melts, edge effects and focusing may be responsible for the large volumes and short durations (King and Anderson, 1998; Foulger and Anderson, 2005; Foulger et al., 2005). These locations also suggest that crustal stoping may be involved (Anderson, 2005; see also Daly, 1933). Delamination produces asthenospheric upwelling and decompression melting as the root is detached, and produces magmatism again as it returns to the shallow mantle.

A number of workers have proposed that large magma fluxes can be produced from eclogite-bearing mantle, without high potential temperatures (Takahashi et al., 1998; Yasuda & Fujii, 1998; Leitch & Davies, 2001). Takahashi et al. (1998) argued that LIPs could result from partial melting of eclogite with a potential temperature not greatly in excess of the

MORB adiabat. Takahashi & Nakajima (2002) suggested that the Koolau component in Hawaii can be produced from large eclogite blocks but the inferred excess potential temperature is only about 100°C. This means that high absolute temperatures can be replaced by high homologous temperatures. Partial melts of eclogite can explain the silica-enriched compositions of some LIPs (Takahashi et al., 1998) and extensive melting of eclogite can explain magmas with no garnet residue signature. The cited papers all consider that the eclogite is delivered by deep-seated thermal plumes. The present chapter argues that such plumes are unnecessary.

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FIGURE CAPTIONS

FIGURE 1: Crustal and mantle minerals arranged in order of increasing density. The STP densities and shear velocities are given. The depth scale is approximate. Along the mantle geotherm the order will change because eclogite will be above the solidus. The densities of mafic and ultramafic melts are given in terms of the density deficit at 1600^{o}C and high pressure with respect to PREM (Dziewonski and Anderson, 1981). Because of the high compressibility of magma, the density deficit increases rapidly with decreasing depth. Partially molten eclogite diapirs become increasingly buoyant as they rise. Some magmas are neutrally buoyant at depth in the mantle. Multiple discontinuities and low-velocity zones (eclogite or magma) are predicted in a chemical stratified mantle. These features may not be obvious in global tomography but they shown up in high-resolution reflection, receiver function and scattering experiments. The layering shown is a snap-shot. Eclogite becomes buoyant as it collects heat from the adjacent mantle. Delaminated eclogite will be 400-700^{o}C colder than ambient mantle while oceanic crust can be 1300^{o}C colder. Deeper layers can be displaced upwards by later generations of subduction and delamination, and by ridge spreading. In a layered mantle the temperature gradient is superadiabatic (a conductive geotherm) unless there is radioactive heating or insertion of cold dense material. Subadiabatic gradients may occur at depth in this situation. In general, one cannot infer temperature simply by assuming an adiabatic gradient below some 200 km depth. Convection in this kind of system is similar to a lava lamp, and is quite different from a pot-on-a-stove or thermal plume convection, or convection in a homogeneous fluid. The system is driven by cooling from above. Magma densities are from Sakamaki et al. (2006). Rock and mineral densities are from standard sources and databases and Lee et al. (2006).

FIGURE 2: Schematic illustration of the operation of the eclogite cycle; cold slabs (yellow) sink to the base of the transition region, displacing warmer or less dense material upwards. They are not treated in this chapter. 1) over-thickened continental crust or underplated basalt cools and converts to dense eclogite, which is unstable in the shallow mantle; it may take 10 Ma for a sufficiently thick eclogite-rich (piclogite) root to form and detach (all times are approximate). 2) the cold piclogite assemblage sinks rapidly to a depth of neutral buoyancy or is trapped by a phase change density jump in ambient mantle. The blob is surrounded by hot ambient mantle and also has higher radioactive content than ambient mantle; it warms up gradually. 3) liftoff occurs as a result of heating, melting or displacement by subsequent downwellings. 4) upwelling may be accompanied by partial melting and a further density reduction; melts drain to the top of the blob and may react with the surrounding peridotite. 5) the blob delivers melt to the shallow mantle and surface. The whole cycle may take 40 to 80 Myr. This is much faster than the plume cycle that relies on the small amount of heat leaving the core. The relative densities (ρ) are shown. The approximate time scales of the various processes (cooling, delamination, heating, melt percolation) are given in Jull and Kelemen (2001), Ghods and Arkani-Hamed (2002), Anderson (2005a), and Raddick et al. (2002), and in standard texts. Also see http://www.mantleplumes.org/TopPages/LithThinTop.html & http://www.mantleplumes.org/LithGravInstab.html.

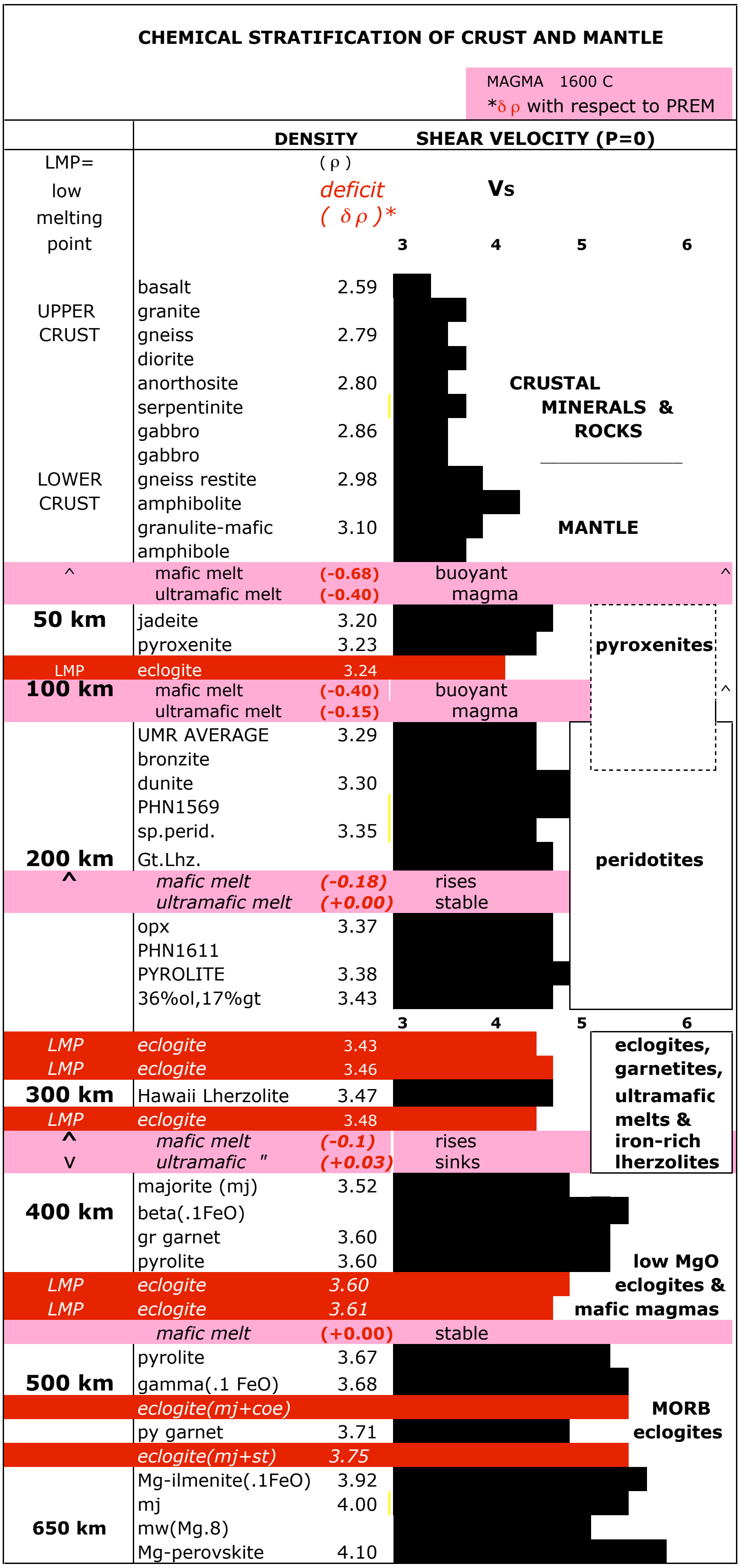
CHEMICAL STRATIFICATION OF CRUST AND MANTLE
MAGMA 1600 C
*δ ρ with respect to PREM
DENSITY
SHEAR VELOCITY (P=0)
LMP=
low
melting
point
(ρ)
deficit
(δ ρ)*
Vs
3
4
5
6
UPPER
CRUST
basalt 2.59
granite
gneiss 2.79
diorite
anorthosite 2.80
serpentinite
gabbro 2.86
gabbro
LOWER
CRUST
gneiss restite 2.98
amphibolite
granulite-mafic 3.10
amphibole
CRUSTAL
MINERALS &
ROCKS
MANTLE
^
mafic melt (-0.68)
ultramafic melt (-0.40)
buoyant
magma
^
50 km
jadeite 3.20
pyroxenite 3.23
pyroxenites
LMP
eclogite 3.24
100 km
mafic melt (-0.40)
ultramafic melt (-0.15)
buoyant
magma
^
UMR AVERAGE 3.29
bronzite
dunite 3.30
PHN1569
sp.perid. 3.35
200 km
Gt.Lhz.
peridotites
^
mafic melt (-0.18)
ultramafic melt (+0.00)
rises
stable
opx 3.37
PHN1611
PYROLITE 3.38
36%ol,17%gt 3.43
3
4
5
6
LMP eclogite 3.43
LMP eclogite 3.46
300 km
Hawaii Lherzolite 3.47
LMP eclogite 3.48
^
mafic melt (-0.1)
V
ultramafic " (+0.03)
rises
sinks
eclogites,
garnetites,
ultramafic
melts &
iron-rich
lherzolites
400 km
majorite (mj) 3.52
beta(.1FeO)
gr garnet 3.60
pyrolite 3.60
low MgO
eclogites &
mafic magmas
LMP eclogite 3.60
LMP eclogite 3.61
mafic melt (+0.00)
stable
500 km
pyrolite 3.67
gamma(.1 FeO) 3.68
eclogite(mj+coe)
MORB
eclogites
py garnet 3.71
eclogite(mj+st) 3.75
650 km
Mg-ilmenite(.1FeO) 3.92
mj 4.00
mw(Mg.8)
Mg-perovskite 4.10

Figure 1

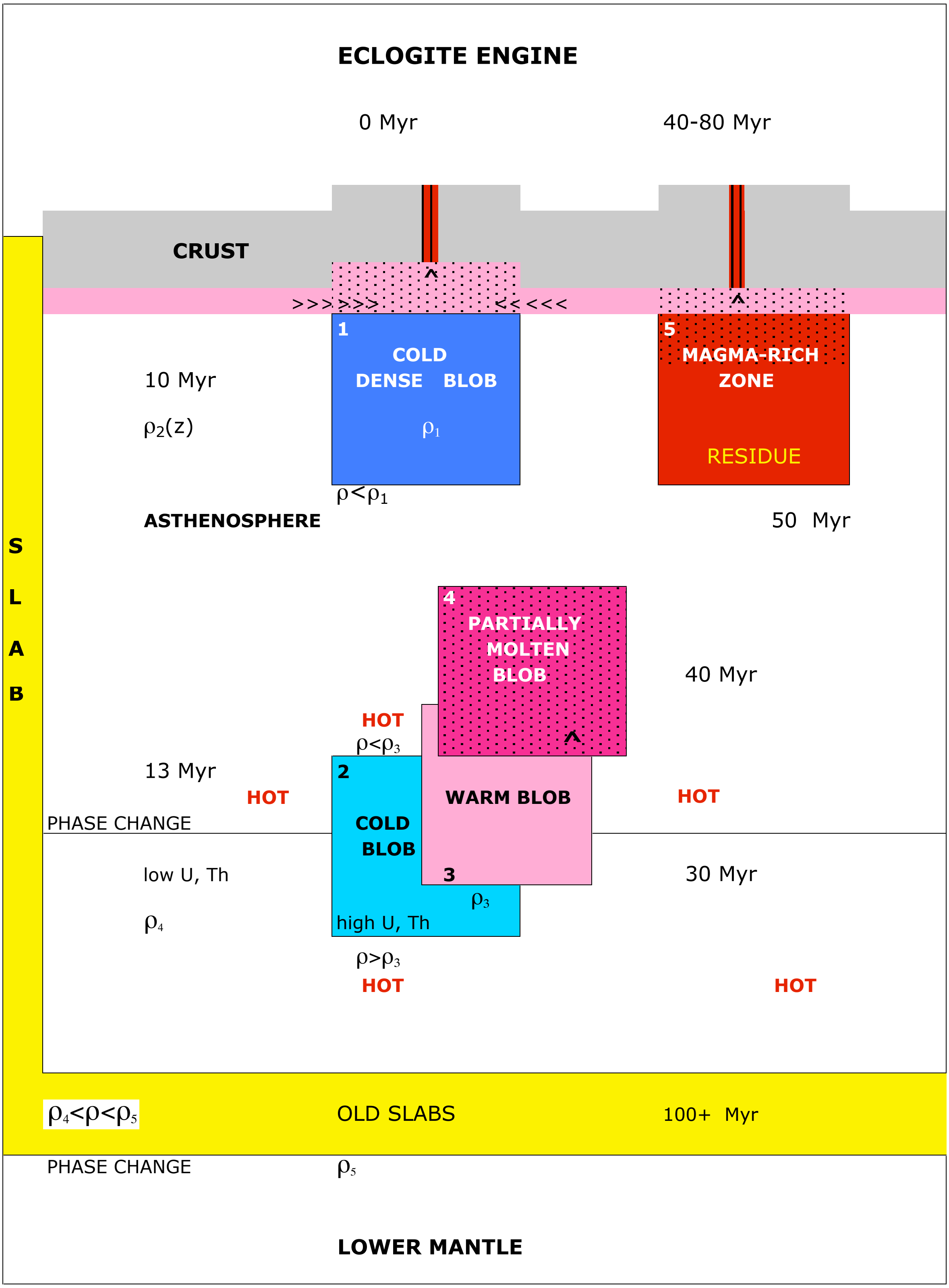
ECLOGITE ENGINE
0 Myr
40-80 Myr
CRUST
>>>>>>
<<<<<
1
COLD
DENSE BLOB
ρ1
ρ<ρ1
5
MAGMA-RICH
ZONE
RESIDUE
10 Myr
ρ2(z)
ASTHENOSPHERE
50 Myr
S
L
A
B
4
PARTIALLY
MOLTEN
BLOB
40 Myr
HOT
ρ<ρ3
13 Myr
HOT
2
WARM BLOB
HOT
PHASE CHANGE
COLD
BLOB
low U, Th
3
30 Myr
ρ3
ρ4
high U, Th
ρ>ρ3
HOT
HOT
ρ4<ρ<ρ5
OLD SLABS
100+ Myr
PHASE CHANGE
ρ5
LOWER MANTLE

Figure 2