Large Igneous Provinces, Delamination, and Fertile Mantle

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When continental crust gets too thick, the dense eclogitic bottom detaches, causing uplift, asthenospheric upwelling, and pressure-release melting. Delamination introduces warm blocks of lower crust with a low melting point into the mantle; these eventually heat up, ascend, decompress, and melt. The mantle below 100 km depth is mainly below the melting point of dry peridotite, but its temperature will be above the melting point of recycled fertile (basaltic or eclogitic) components, obviating the need for excess temperature to form 'hotspots' or 'melting anomalies'. When plates pull apart or delaminate, the mantle upwells; entrained crustal fragments of various ages are fertile and create melting anomalies along developing mid-ocean ridges, fracture zones and old suture zones. Eclogites associated with delamination are warmer and less dense than subducted oceanic crust and more susceptible to melting and entrainment.

KEYWORDS: LIPs, delamination, plumes, hotspots, lithosphere, eclogite

INTRODUCTION

Large igneous provinces (LIPs) are generally attributed to hotter-than-normal mantle. It is important, therefore, to know the normal range of mantle temperatures. Convection calculations for a fluid with mantle-like properties that is heated internally and cooled from above predict temperature fluctuations of at least ± 100°C (Anderson 2000). Geophysical evidence suggests that the mantle temperature under most LIPs was in this normal range while the LIPs were erupting (Clift 2005; Korenaga et al. 2002), and, where measured, the present heatflow is also normal (i.e. appropriate for the age of the underlying crust). The sedimentary records from a range of swells and plateaus of various ages from all major ocean basins (North Atlantic, Mid-Pacific Mountains, Shatsky Rise, Ninetyeast Ridge, Ontong Java Plateau) are compatible with eruption of magma from mantle in the normal range of temperature, followed by conductive cooling of the type associated with regular oceanic crust. The North Atlantic Ocean is particularly shallow, but modeling by Clift (2005) indicates that the depth is consistent with a temperature anomaly of +100°C or less. Furthermore, he shows many LIP sites where the depth to the base of the sediments implies colder than average mantle temperatures.

The largest LIPs (Siberia Traps, Ontong Java Plateau) were arguably erupted at, below, or near sea level rather than at an elevation of one or two kilometres above the surrounding terrain, as predicted by the plume hypothesis. LIP uplift, if it occurs, appears to be syn- or post-volcanism, rather than millions of years prior to volcanism, as predicted by thermal models. There is no indication from uplift, heatflow, or seismic tomography for mantle temperatures more than 100°C above the mean under these LIPs (Czamanske et al. 1998; Korenaga 2005; Roberge et al. 2005; Gomer and Okal, 2003).

The crust under continental LIPs is thinner than average continental crust (Mooney et al. 1998). Many LIPs occur atop deep sedimentary basins, in back arcs, or on old convergent margins. The chemistry of LIPs is highly variable, with compositions ranging between mid-ocean ridge basalt (MORB) and ocean island basalt; many continental LIPs show clear evidence of input from continental crust or lithospheric mantle. Picritic melts are rare. These observations are all enigmatic in the context of the usual high-temperature or plume explanations for LIPs. The alternative explanation is that the mantle is hotter than generally assumed (Anderson 2000) or is compositionally heterogeneous on a large scale, or both.

GONDWANA, ATLANTIC, AND INDIAN OCEAN LIPs

The breakup of Gondwana was preceded by extensive volcanism along the future Atlantic and Indian ocean margins (FIG. 1). Volcanism continued during breakup, along the continental margins, and on the separated fragments (in and around the North Atlantic and on and near Madagascar, Kerguelen, and the Rio Grande Rise, see FIG. 1). Plate reconstructions (Müller et al. 1993) show that the currently continent-hugging plateaus (~1000 km offshore) were mainly formed at ridges and triple junctions in the newly opened Atlantic and Indian oceans, some tens of millions of years after breakup of the supercontinent. This can be illustrated by calculating the approximate half-width of the ocean at the time of peak LIP volcanism, based on magnetic anomaly maps and ages obtained from the plateaus. Some of these plateaus (and half-widths of the ocean, in kilometres) are as follows: Azores (1300), Bermuda (1100), Cape Verde (900), Crozet (1000), Discovery (2000), Iceland (1000), Jan Mayen (900), Madagascar (1200), Mozambique (800), Rio Grande Rise (1200), Broken Ridge (1100), Kerguelen (1100), and Walvis Ridge (1200). The delay between continental breakup and the main stage of plateau formation is usually about 20–50 Myr (FIG. 1). Although some LIPs are currently intraplate, they formed, without known exception, at plate or craton borders, at triple junctions, or

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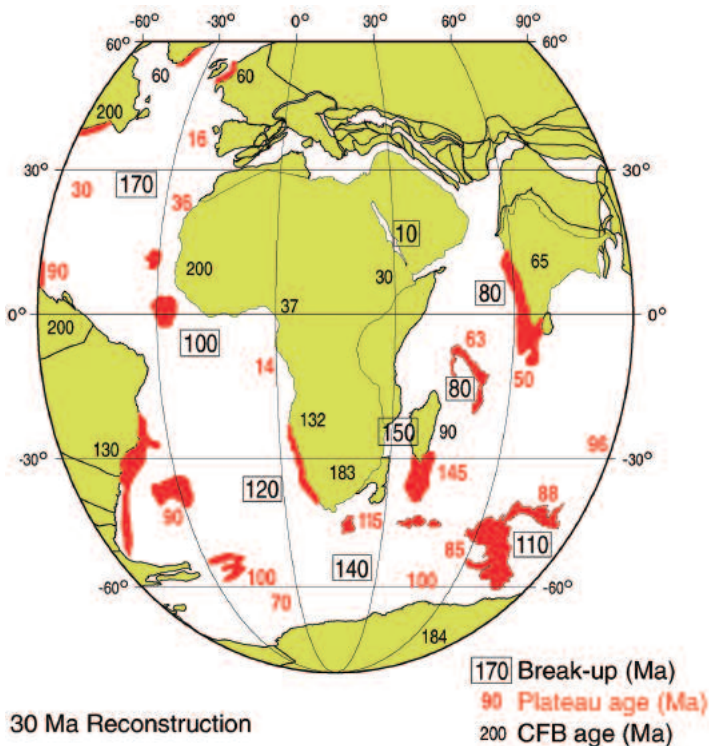   Pasadena, CA 91125, USA
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FIGURE 1 Distribution and ages of LIPs in the Gondwana hemisphere. Also shown are the ages of continental breakup and uplift. This figure illustrates the geometry and relative timing of supercontinent breakup and magmatism, and LIPs in the newly opened ocean. Delamination of lower crust and consequent asthenospheric upwelling may be responsible for the LIPs in the continents (ages are given in small black numbers), which usually occur in mobile belts, arcs, suture zones and accreted terranes. If the continents move away from the delamination sites, it may be possible to see the reemergence of fertile delaminated material in the newly formed ocean basins (e.g. oceanic plateaus; red patches), particularly where spreading ridges allow asthenospheric upwellings. CFB: continental flood basalt. Continents are a mosaic of fragments, some of which are shown with thin and thick lines as borders. The names of the CFB and oceanic LIPs are left off for clarity of the figure (see http://www.largeigneousprovinces.org/ and http://www.mantleplumes.org/).

on a spreading ridge about 1000 km offshore from newly separated continents.

THE PACIFIC PLATEAUS

The Pacific plate originated as a roughly triangular microplate, antipodal to Pangea and surrounded by ridges (Natland and Winterer 2005). It grew by the outward migration of ridges and triple junctions. The growing Pacific plate may have been stationary because of the absence of bounding trenches. The great oceanic plateaus in the Pacific were being constructed at the boundaries of the expanding Pacific plate between the times of Pangea breakup and the construction of the large igneous plateaus in Africa and South America and in the wake of the drifting continents; they are not connected to chains of volcanic islands. An unstable stress regime, plate reorganizations, and complex triple junction jumps may be responsible for the formation of Pacific seamount fields and plateaus (Natland and Winterer 2005). The implication is that the mantle is close to the melting point and is variable in composition and melting point (FIG. 2) and that magmatism is focused by lithospheric architecture and stress.

CONSTRAINTS FROM SEISMIC TOMOGRAPHY

Some LIPs have small-diameter, seismic low-velocity zones (LVZ), in the upper 200–350 km, rarely deeper, of the underlying mantle (e.g. Christiansen et al. 2002; Allen and Tromp 2005). LVZs are regions in the mantle where seismic waves are slowed. This happens because the mantle is hot and/or partially molten or has a mineralogy different from that of the normal ambient mantle. For example, eclogite in the mantle has a low melting point (FIG. 2) and can have slow seismic wave speeds (e.g. FIG. 3 in Anderson 1989a). It is usually assumed that these LVZ are related to the overlying volcanism and that they are thermal in nature. But some ancient LIPs, such as Paraná and the Ontong Java Plateau (OJP) retain seismic low-velocity zones in the upper 200–300 km of the mantle (Vandecar et al. 1995; Gomer and Okal 2003) even though they have traveled thousands of kilometers away from the putative source and the mantle has had more than 120 Myr to cool. The OJP low-velocity zone slows seismic waves down but does not attenuate their amplitude significantly, as would be the case if the mantle were hot or partially molten (Gomer and Okal 2003). Compositional, rather than thermal, effects are implied (e.g. as in FIG. 3). Both upwelling asthenosphere and sinking eclogite can have low seismic velocities. Sinking eclogite, however, melts at lower temperatures than peridotite (FIG. 2) (Note that a decrease in seismic velocity is not the same thing as attenuation. Cold material, such as cold eclogite, can have a low seismic velocity but can be transparent to seismic waves, i.e. have low attenuation. A LVZ with low attenuation implies that it is compositional rather than thermal in origin).

CONTINENTAL CRUST IN THE MANTLE?

Fragments of continental crust have been found along the Mid-Atlantic Ridge (e.g. Bonatti et al. 1996) and in hotspot and LIP magmas. Schaltegger et al. (2002) found continental zircon xenocrysts in basalts from Iceland and Mauritius. Continental crust is inferred to exist at the Seychelles, the Faeroes, Rockall Bank, Jan Mayen, Kerguelen, the Ontong
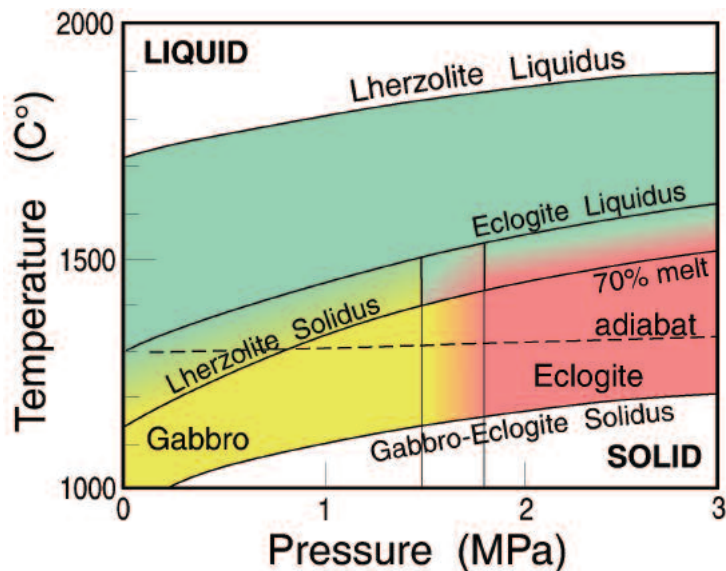

FIGURE 2 Melting relations in dry lherzolite and eclogite based on laboratory experiments. The dashed line is a reference 1300°C mantle adiabat. Eclogite will melt as it sinks into normal-temperature mantle, and upwellings from the shallow mantle will extensively melt any entrained gabbro and eclogite. Eclogite will be about 70% molten before dry lherzolite starts to melt (compiled by J. Natland, personal communication). The two vertical lines separate the gabbro field from the eclogite field (in between is the mixed phase region).

Java Plateau, Cape Verde, and the Cameroon Line (e.g. Frey et al. 2002; Ishikawa and Nakamura 2003). The widespread isotopic characteristics of Indian Ocean basalts have been attributed to the presence of "lower continental crust entrained during Gondwana rifting" (Hanan et al. 2004) or to "delamination of lower continental crust" (Escrig et al. 2004).

THE DELAMINATION MECHANISM

The challenge is to find mechanisms that can explain the volume of basalt, the uplift history, and the ubiquitous evidence for involvement of both continental and mid-ocean ridge material in LIP magmas. Some igneous provinces are built on top of rafted pieces of microcontinents or abandoned island arcs, but is there any mechanism for putting large chunks of continental material into the source regions of LIPs? Lower crustal delamination is such a mechanism, although it has been basically unexplored in this context.

The lower continental crust thickens by tectonic and igneous processes (Kay and Kay 1993; Rudnick 1995), including magmatic underplating. Presumably the same thing can happen at intra-oceanic arcs. Below about 50 km, mafic crust (basalt, dolerite, gabbro) transforms to dense garnet pyroxenite (eclogite, see glossary. Arc-eclogites in FIG. 3). Histograms of the thickness of the continental crust show a sharp drop-off at a thickness of 50 km (Mooney et al. 1998). I suggest that this is controlled by delamination. Once a sufficiently thick eclogite layer forms, it will detach and founder because its density is 3 to 10% greater than that of normal mantle peridotite (FIG. 3). Delamination of a 10 km thick eclogite layer can lead to 2 km of uplift and massive melt production within 10 to 20 Myr (Vlaar et al. 1994; Zegers and van Keken 2001). Density contrasts of 1% are enough to drive downwelling instabilities (Elkins-Tanton 2005). Thus, delamination is a very effective and non thermal way of thinning the lithosphere, extending the melting column, and creating massive melting and uplift. In contrast to thermal models, uplift occurs during and after volcanism, and crustal thinning is rapid.

Lee et al. (2005) estimate that it takes 10–30 Myr for a lower-crustal mafic layer to reach critical negative buoyancy and for foundering to take place. The thickness of the mafic layer at the time of foundering ranges between 10 and 35 km, resulting in significant size heterogeneities in the mantle. When the lower crust is removed, the underlying mantle upwells to fill the gap and melts because of the effect of pressure on the melting point. This results in a pulse of magmatism and an episode of rapid uplift. The lower crust then rebuilds itself and cools, and the cycle repeats. The delamination mechanism creates multiple pulses of magmatism, separated by tens of millions of years, a characteristic of some LIPs. If the crustal thickening is due to compressional tectonics, the time scales will be dictated by convergence rates. In a typical convergent belt, thickening and delamination may take 25–35 Myr.

There are several ways to generate massive melting: one is to bring up hot material adiabatically from depth until it melts; the other is to insert fertile material with a low melting point —delaminated lower arc crust, for example—into the mantle from above and allow the mantle to heat it up by conduction. Eclogite that was subsolidus at lower crustal depths can melt extensively when placed into ambient mantle (FIG. 2). Both mechanisms may be involved in LIP formation. The time scale for heating and recycling lower-crust material is much less than for subducted oceanic crust because the former starts out much hotter and does not sink as deep (FIG. 3). The total recycle time, including reheating, may take 30 to 75 Myr. If delamination occurs
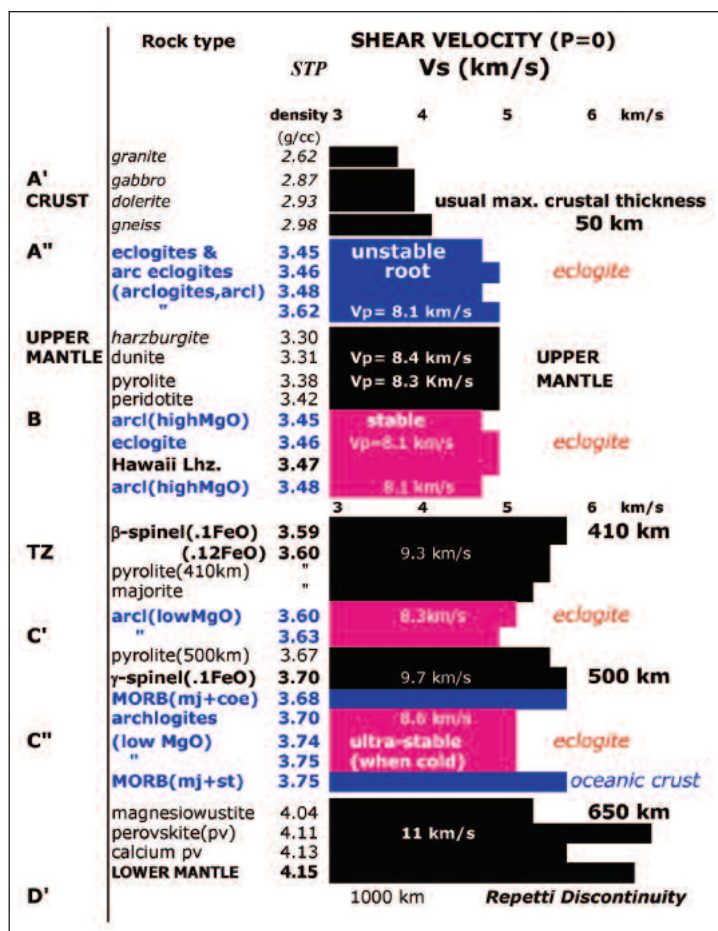
FIGURE 3 The neutral density profile of the crust and mantle. The materials of the crust and mantle are arranged mostly in order of increasing density. **A"** is the region of over-thickened crust that can transform to eclogite and become denser than the underlying mantle; **C'** is the top part of the transition region. Eclogites having STP densities between 3.6 and 3.7 g/cc may be trapped in C'. Eclogites and peridotites have similar densities in this interval but different seismic velocities. In **C"**, MORB-eclogite and low-MgO arc-eclogites reach density equilibration. Trapped eclogite and dense eclogite sinkers can be low-velocity zones. The order approximates the situation in an ideally chemically stratified mantle, except for the region just below the continental Moho where potentially unstable lower crustal cumulate material is formed. In such a stably stratified structure the temperature gradient will be greater than the adiabatic gradient.

near the edge of a continent, say along a suture belt (Foulger et al. 2005), and the continent moves off at 3.3 cm per year (the average opening velocity of the Atlantic ocean), the delamination site will have moved 1000 km to 2500 km away from a vertically sinking root in the time since delamination.

BROAD DOMAL UPLIFT

Broad domal uplift is a characteristic of delamination (Kay and Kay 1993). The magnitude is related to the density and thickness of the delaminating column. Crustal domes of ~1000 km in lateral extent and elevations of ~2 km above background, with no heat flow anomaly, can be explained with such shallow processes (Petit et al. 2002). The Mongolian dome, for example, is underlain by a small anomaly (5% seismic velocity reduction and a density reduction of only 0.01 g/cc) of limited vertical extent, 100–200 km deep, yet it has the same kind of domal uplift that has been assumed to require a large and deep thermal perturbation. The removal of a dense eclogitic root and its replacement by

upwelling peridotite can create regional uplift and a shallow low-velocity zone; the eclogite sinker also has low seismic velocities. This process is essentially a top-down athermal process.

One of the best-documented examples of delamination, uplift, and volcanism is the Eastern Anatolia region (Keskin 2003), which was below sea level between ~50 and ~13 Ma. It was then rapidly elevated above sea level. Uplift was followed by widespread volcanic activity at 7–8 Ma, and the region acquired a regional domal shape comparable to that of the Ethiopian High Plateau. Geophysical, geological, and geochemical studies support the view that domal uplift and extensive magma generation were linked to the mechanical removal of the lower crust accompanied by upwelling of normal-temperature asthenospheric mantle to a depth of 50 km. The above examples are important in showing that well-understood shallow processes can generate regional domal structures and large volumes of magma. The LVZ under some LIPs, including ancient ones, and domal structures may be associated with cold eclogite rather than hot upwellings (FIG. 3).

DO WE NEED TO RECYCLE OCEANIC CRUST?

Recycled oceanic crust is often considered to be a component of ocean-island and LIP magma, although this view is disputed. Once in the mantle, subducted MORB-eclogite reaches neutral buoyancy at depths of 500–650 km (Anderson 1989b; Hirose et al. 1999) (FIG. 3). Very cold MORB may sink deeper (Litasov et al. 2004). If current rates of oceanic crust recycling operated for 1 Gyr (Stern 2005), the total subducted oceanic crust would account for 2% of the mantle, and it could be stored in a layer 70 km thick only. The surprising result is that most subducted oceanic crust need not be recycled or sink into the lower mantle in order to satisfy any mass-balance constraints (see also Anderson 1989a). The MORB-like component in some LIPs may simply be due to passive asthenospheric upwelling, as in the delamination model.

The recycling rate of lower-crustal mafic rocks (Lee et al. 2005) implies that about half of the continental crust is recycled every 0.6 to 2.5 billion years. In contrast to oceanic crust, one can make a case that eroded and delaminated arc and continental material is not stored permanently, or over the long term, or very deep in the mantle; it is reused and must play an important role in continental mass balance, global magmatism, and shallow-mantle heterogeneity.

MELTING OF ECLOGITE

Geophysical estimates of the potential temperature (see glossary) of the mantle are about 1350–1400°C (Anderson 2000), with statistical and geographic variations of at least 100°C. These temperatures are about 100 degrees higher than generally assumed by petrological modelling. This range permits partial melting of peridotite, the formation (in places) of high-MgO magmas, and extensive melting of eclogite. Melting experiments (e.g. Yaxley 2000) suggest that 60–80% melting of eclogite is required to reproduce compositions of some LIP basalts (Natland, personal communication). FIG. 2 suggests that this is plausible and that lherzolite will start to melt under these conditions. The interaction of melts from eclogite and lherzolite is implied. The model discussed here does not imply low mantle temperatures. In fact, an internally heated (i.e. by radioactive decay), chemically stratified mantle achieves higher temperatures than a uniform mantle heated from below. One must explain, then, not the existence but the rarity of picrites. The eruption of high-temperature MgO-rich magmas may require special circumstances because of their high density. This special circumstance may be the rapid upwelling of asthenosphere following a delamination event and edge-driven convection (King and Anderson 1998).

DISCUSSION

Some LIPs may simply be due to passive upwelling of inhomogeneous asthenosphere as continental fragments diverge (McHone 2000). Some are the result of reactivated suture zones or other weaknesses of the lithosphere, combined with the variable fertility and melting point of the underlying mantle (Foulger et al. 2005). In this paper, I have focused on a new mechanism that augments these other processes. Delaminated lower crust sinks into the mantle as eclogite, where it has a relatively low seismic velocity and melting point compared to normal mantle peridotite. Although delaminated continental crust enters the mantle at much lower rates than oceanic crust, the rates are comparable to LIP production rates. I speculate that the large melting anomalies that form on or near ridges and triple junctions may be due to the resurfacing of large fertile blobs, including delaminated continental crust. Ponded melts may contribute at new ridges and triple junctions. Delaminated eclogites may form a unique component of hotspot and ridge magmas (Lee et al. 2005), but I suggest that lower continental crust is not just a contaminating agent. Blocks of it are responsible for the melting anomalies themselves, including the Kerguelen Plateau and other features in the Indian Ocean. The massive plateaus, such as the Ontong Java Plateau, may be due to a combination of delamination (Korenaga 2005), excess mantle fertility, slightly higher average mantle temperatures than usually assumed (~100°C), slightly lower melting temperatures (~100°C), and focusing of magma at a triple junction.

The crustal delamination, variable mantle fertility model, combined with passive asthenospheric upwelling, has the potential to explain the tectonics and compositions of LIPs, including heatflow and uplift histories. Apparently, no other model explains the formation of LIPs and uplifted domes so elegantly, with so few contradictions. But the model needs to be tested further and quantified. ◾

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(Current information on the subject matter of this article can be found at www.mantleplumes.org)

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