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# How the delamination and detachment of lower crust can influence basaltic magmatism

Michele Lustrino

*Dipartimento di Scienze della Terra, Università degli Studi di Roma La Sapienza, P.le A. Moro, 5, 00185 Rome, Italy*
*Istituto di Geologia Ambientale e Geoingegneria (IGAG)-CNR, P.le A. Moro, 5, 00185 Rome, Italy*

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**Abstract**

The Earth's lithosphere can focus basaltic magmatism along pre-existing weakness zones or discontinuities. However, apart from the influence on the geochemistry of magmas emplaced in subduction tectonic settings (mantle wedge metasomatism related to dehydration of the subducting plates) the role of lithosphere as a magma source for intra-plate (both oceanic and continental), continental margin, and mid-ocean ridge magmatism is not yet fully understood. In many cases intra-plate magmatism has been explained with the existence of deep thermal anomalies (mantle plumes) whose origin has been placed near the upper–lower mantle transition zone (660 km discontinuity) or even deeper, near the mantle–core boundary (~2900 km). Also in many continental flood basalt provinces (mostly initiated at craton margins) an active role for mantle plumes has been invoked to explain the high melt productivity. In these cases, no active role for melt production has been attributed to the lithospheric mantle. Potential contaminations of asthenospheric or even deeper mantle melts are often considered the only influence of the lithosphere (both crust and mantle) in basalt petrogenesis. In other cases, an active role of the lithospheric mantle has been proposed: the thermal anomalies related to the presence of mantle plumes would trigger partial melting in the lithospheric mantle. At present there is no unequivocal geochemical tracer that reflects the relative role of lithosphere and upper/lower mantle as magma sources. In this paper another role of the lithosphere is proposed.

The new model presented here is based on the role of lower crustal and lithospheric mantle recycling by delamination and detachment. This process can explain at least some geochemical peculiarities of basaltic rocks found in large and small volume igneous provinces, as well as in mid-ocean ridge basalts. Metamorphic reactions occurring in the lower continental crust as a consequence of continent–continent can lead to a density increase (up to 3.8 g/cm$^3$) with the appearance of garnet in the metamorphic assemblage (basalt→amphibolite→garnet clinopyroxenite/eclogite) leading to gravitative instability of the overthickened lithospheric keel (lower crust + lithospheric mantle). This may detach from the uppermost lithosphere and sink into the upper mantle. Accordingly, metasomatic reactions between $SiO_2$-rich lower crust partial melts and the uprising asthenospheric mantle (replacing the volume formerly occupied by the sunken lithospheric mantle and the lower crust) lead to formation of orthopyroxene-rich layers with strong crustal signatures. Such metasomatized mantle volumes may remain untapped also for several Ma before being reactivated by geological processes. Partial melts of such sources would bear strong lower crustal signatures giving rise to Enriched Mantle type 1 (EMI)-like basaltic magmatism. Basaltic magmatism with such a

*E-mail address:* michele.lustrino@uniroma1.it.

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geochemical signature is relatively scarce but in some cases (e.g., Indian Ocean) it can be a geographically widespread and long-lasting phenomenon.


*Keywords:* delamination; detachment; lithosphere; mantle plume; petrology; lower crust; eclogite; pyroxenite

## 1. Introduction

Korenaga (2004) recently proposed an alternative to the mantle plume theory, that can explain voluminous magmatic activity associated with continental break-up. His model suggests that upper mantle heterogeneities are the most important factors for the high melt productivity of the Continental Flood Basalt (CFB) provinces around the Atlantic Ocean. However, according to this model, transport of material from the lower mantle still plays an important role. In particular, the key control responsible for the Atlantic CFB provinces would be the subducted crustal material (the uppermost portion of subducting slabs) stored at the upper–lower mantle transition zone at ~660 km in depth. The model proposed by Korenaga (2004) can be modified and exported in many other cases where CFB activity develops. Many other recent papers also invoke chemical heterogeneity in the shallow mantle to explain features of ocean and continent magmatism (e.g., Meibom and Anderson, 2003). In this paper I emphasize an alternative model for the continental break-up and the associated magmatism that does not require the existence of mantle plumes with deep roots in the lower mantle. The model presented here involves delamination and detachment of lower crust to mantle depths and is not necessarily an alternative to the Korenaga's (2004) model. Recycling of lower crust (coupled with lithospheric mantle) can explain several geochemical peculiarities relatively common in low-volume intra-plate igneous rocks (ocean island basalts and intra-continental rocks), oceanic and continental flood basalts and mid ocean ridge basalts (MORB).

The first part of this paper highlights paradoxes and inconsistencies of the classic mantle plume model, the second part reviews geochemical and geophysical evidence for lower crust/lithospheric mantle delamination and detachment, while a more detailed model is developed in the final section.

## 2. Geochemical expression of mantle convection

### 2.1. Are mantle plumes necessary?

Since at least the 1960s, many studies have invoked strong thermal anomalies as the main cause of large volume igneous activity in both continental and oceanic settings (e.g., Wilson, 1963; Morgan, 1971; Condie, 2001; Marzoli et al., 1999; Kamo et al., 2003; Thompson et al., 2003; Ewart et al., 2004). These plume-like thermal anomalies were considered to be near-cylindrical in shape except for a huge, mushroom-shaped, head, with mean diameters on the order of several hundred to thousand kilometers (Campbell and Griffiths, 1990; Farnetani et al., 2002). Mantle tomography gives contrasting results about the existence of the mantle plumes, either indicating their existence (e.g., Montelli et al., 2004) being ambiguous (e.g., Ritsema and Allen, 2003) or precluding the possibility of deep-rooted origin for these anomalies (e.g., Anderson et al., 1992). The presence of mantle plumes has also been invoked to explain small volume igneous activity both in continental and oceanic settings (e.g., Hoernle et al., 1995; Chauvel et al., 1997; Wilson and Patterson, 2001; Hildenbrand et al., 2004). The source regions of such thermal anomalies are considered to lie at the bottom of the lower mantle (the core–mantle boundary; CMB), the upper–lower mantle transition zone (the 660 km discontinuity) or somewhere at mid-mantle depths (below the 660 km and above the 2900 km discontinuities; i.e., Cserepes and Yuen, 2000; Zhao, 2001; Courtillot et al., 2003; Ritsema and Allen, 2003; Montelli et al., 2004). Since the late 1990s, the classical plume model has been the subject of strong criticism. Many contradictory aspects have been reviewed, among others, by Sheth (1999), Smith and Lewis (1999) and Anderson (2002; reviewed at the web site http://www.mantleplumes.org). The most important problems confronting mantle plume models can be summarized as follows:

a) Experimental studies (e.g., Cordery et al., 1997; Yaxley, 2000) cannot explain the high melt productivity of Large Igneous Provinces (LIPs) without recourse to the existence of a relatively low-temperature melting material resembling crustal rocks stored at various depths in the mantle. This means that the origin itself of the huge volumes of magmas produced in LIPs may ultimately derive not necessarily (or, better, not entirely) from anomalously hot geothermal gradients but, rather, from anomalous fertile sources (fertility being consequence of the presence of Ca–Al–Fe-rich eclogitic slices mixed in a peridotitic medium). Moreover, experimental partial melts of peridotitic composition are distinct from those producing alkaline Ocean Island Basalts (OIB), which appear to require involvement of high pressures (2–5 GPa) partial melts of crustal (silica-deficient garnet pyroxenites) components (e.g., Hirschmann et al., 2003; Kogiso et al., 2003, 2004).
b) The thermal inertia of mantle rocks cannot explain the rapid cessation of most CFB magmatic activity. Indeed, the duration of Large Igneous Provinces (LIPs) activity lasted only a few Ma, with magma productivity peaking confined within 1–3 Ma. The model of a pure thermal anomaly (hot blobs coming from lower mantle) cannot explain the rapid cut-off of this anomaly almost immediately after the magmatic peak.
c) Experimental and numerical simulations used in support of mantle plume theories (e.g., Cserepes and Yuen, 2000; Davaille et al., 2003; Samuel and Farnetani, 2003) are based on unrealistic conditions (e.g., no phase changes, no plate motions, no structural discontinuities, injection of superheated fluid at the base of another fluid, no upper mantle heterogeneities as consequence of melt removal and crustal subduction, no mantle flow, etc.), such that results are strongly model-dependent and may lead to incorrect results.
d) The geographic locations of CFB are not random but invariably associated with ancient mobile belts. This may be considered as evidence for strong structural control of both the crustal and mantle portions of the lithosphere in channeling the CFB feeding systems.
e) There is a paradox between the “pin-point” expression of “plume” thermal anomaly and the planar development of continental rifts. This consideration is closely connected to (d) above, thereby reinforcing the evidence for strong lithospheric control on the localization of LIPs and the development of continental rifting (e.g., Vauchez et al., 1997; Tommasi and Vauchez, 2001).
f) None of the known hotspots meets all five criteria for detecting ultra-deep origins cited by Courtillot et al. (2003): (1) monotonic age progression of linear volcanic chain; (2) flood basalts at the origin of this track; (3) a large buoyancy flux; (4) high $^{3}He/^{4}He$ and $^{21}Ne/^{22}Ne$ isotopic ratios, and (5) low $V_s$ anomalies down to at least the 660 km discontinuity (e.g., Anderson, in press).
g) Heat flow data on one of the most classic type localities where a mantle plume is supposed to meet Earth’s surface (i.e., Iceland) shows no evidence for regional thermal anomalies expected near a mantle plume axis (Stein and Stein, 2003).
h) In some cases tomographic results are inconsistent with the plume theory or observed geodynamic regimes. For example: (1) deep mantle thermal anomalies are recorded down to the lowermost 1000 km of the mantle in places (e.g., beneath southern Africa; Montelli et al., 2004), where the last magmatic activity is dated ~200 Ma (Hawkesworth et al., 1999), this suggesting the presence of anomalous hot mantle without magmatic expression at the surface; (2) while tomography suggests the existence of deep plumes in some places (e.g., beneath Mt. Etna, southern Italy; Montelli et al., 2004) petrological considerations exclude involvement of deep seated sources and indicate shallow magmatic origins (Peccerillo and Lustrino, in press); (3) some models (e.g., Montelli et al., 2004) cite the existence of deep low-velocity anomalies (~1000 km deep) as for St. Helena Island in the Atlantic Ocean, whereas others exclude the presence of shear velocity anomalies in the same area (e.g., Ritsema and Allen, 2003). Moreover, the volumetrically insignificant igneous activity in St. Helena and the nearly amagmatic extension in the equatorial Atlantic Ocean (e.g., Bonatti, 1996) are inconsistent with the existence

of a deep plume, at least in this area; (4) high-resolution tomography precludes the existence of plumes beneath classic localities such as Yellowstone, Guadalupe and the MacDonald islands (Montelli et al., 2004); (5) the size of tomographically imaged mantle plumes is not directly related to the extent of hotspot magma production (e.g., the South Pacific "super Plume" is associated with low-volume ocean island basalts (Marquesas, Tahiti and Cook Islands); (6) there is a lack of correlation between the maximum tomographic depth of a plume image and the associated $^3He/^4He$ ratios of magmatic rocks. Magmas with high $^3He/^4He$ ratios are generally considered to reflect derivation from primitive, undegassed sources, thereby placing constraining origins to the lower mantle (e.g., Dodson et al., 1997; Sumino et al., 2000).

Thus, deep mantle plumes would yield primitive, undegassed magmas. However, suites attributed to deep plumes (e.g., St. Helena, Canary Islands, French Polynesian Islands, Hawaiian Islands) show highly variable $^3He/^4He$ ratios and, in some cases, lower elemental $^3He$ (and lower noble gas abundances in general) compared to MORB (derived from depleted, therefore degassed, upper mantle; Meibom et al., 2003; Stuart et al., 2003; Yamamoto and Burnard, 2005). It has been suggested that high $^3He/^4He$ ratios can be the consequence of high cosmogenic $^3He$ (coming from solar wind to the Earth's surface) rather than being expression of derivation from undegassed (primitive) deep mantle sources (e.g., Meibom et al., 2003; Yokochi et al., 2005). It is important to know that the measurement of $^3He/^4He$ ratios is strongly dependent upon the method used in extracting the gas from the sample, magmatic He being preferentially extracted by crushing minerals under vacuum, whereas cosmogenic and/or radiogenic He is released by mineral melting after prolonged crushing (Scarsi, 2000; Yokochi et al., 2005).

Moreover, the genesis of one of the Earth's largest LIP, the Ontong Java Plateaux in the SW Pacific Ocean, can be explained as a plume product only if several ad hoc poorly constrained assumptions are made (see discussion in Tejada et al., 2004). Geochemists frequently identify the presence of mantle plumes according to the axiom that magma compositions can be linked to specific tectonic settings and mantle structures [e.g., La/Nb ratios < 1 are related to plume-modified asthenospheric mantle (Coulon et al., 2002); high $^3He/^4He$ or $^{21}Ne/^{22}Ne$ ratios record deep mantle plumes (e.g., Courtillot et al., 2003)]. However, magmas attributed to plumes show highly variable compositions described in terms of end-member components such as HIMU, EMI, EMII ± DMM (Zindler and Hart, 1986; Hofmann, 1997, 2004; Lustrino and Dallai, 2003; Lustrino et al., 2004a). Empirically, the geochemical peculiarities of HIMU, EMI and EMII end-members may be explained in terms of crustal recycling in the mantle (e.g., Cordery et al., 1997; Tatsumi, 2000; Yaxley, 2000; Kogiso et al., 2003; Meibom and Anderson, 2003; Lustrino and Dallai, 2003) while DMM melts may also reflect varying degree of crustal interaction at mantle depths (e.g., Hirschmann and Stolper, 1996; Eiler et al., 2000). The most important differences among OIB subgroups characterized by HIMU, EMI and EMII, are in the age and composition of recycled material, its metamorphic history on reaching mantle depths, the time of isolation in the mantle, its depth of storage and style of cratonization (e.g., Hofmann, 1997). Accordingly, the HIMU, EMI and EMII end-members do not necessarily represent discrete reservoirs preserved over time in the deep mantle and successively emplaced as active plume. A more realistic picture was proposed by Meibom and Anderson (2003), according to which such domains may be represented by a strongly heterogeneous sluggishly convecting volume in which isotopic heterogeneities may grow and differentiate.

Another difficulty with the plume hypothesis is the occurrence of HIMU-like compositions in both large and small volume continental basalts. In many cases HIMU-like compositions characterize tectonic settings which are unrelated to suspected mantle plume loci (e.g., lack doming prior to the onset of magmatism, absence of large volume of erupted magma, no evidence for high temperature at the lithosphere/asthenosphere boundary; see Ziegler and Cloetingh, 2004). For example, Miocene to recent volcanism in NE Arabia (Harrat Ash Shaam, Jordan; Shaw et al., 2003) show typical HIMU geochemical isotopic compositions, but is best attributed to simple mantle decompression in the continental Dead Sea

Rift system. Similarly, HIMU-like trace element and Sr–Nd–Pb isotopic features in small volume basaltic rocks (s.l.) from Sardinia (Italy) (Lustrino et al., 2000) are not associated with any mantle plume (see also Lustrino and Wilson, submitted for publication). It is also worth noting that geochemical, geochronological and geophysical data from the type-area for HIMU, EMI and EMII basalts (French Polynesia in South Pacific Ocean) are best easily explained in terms of a strong lithospheric control on petrogenetic processes (e.g., McNutt et al., 1997; Lassiter et al., 2003).

Finally, the negative Clapeyron slope for the post-spinel transition boundary (ringwoodite→perovskite+magnesiowustite; ~0.003 GPa/K; e.g., Gasparik, 2003), marking the upper–lower mantle transition zone (~660 km) appears to preclude the upward passage of mantle plumes rooted in the lower mantle. This transition zone has long been considered an effective barrier against whole mantle convection, including plume-like upwelling from lower mantle. It should be noted, however, that recent results (Fei et al., 2004) indicate a significantly less negative Clapeyron slope for the 660 km discontinuity (~0.0013 GPa/K), therefore enabling (at least from a thermodynamic point of view) flow of matter from deep mantle in form of mantle plumes (see Le Bars and Davaille, 2004 for a more detailed discussion).

In summary, it is suggested that mantle plume models have, in many cases, been applied without justification. Given the present state of knowledge, is not possible to rule out a role for deep mantle convection, although the existence of alternative models for the development of continental and oceanic LIPs and rifting should be vigorously pursued. Among these, should be included a role for the delamination and detachment of lower continental crust and lithospheric mantle in the genesis of voluminous magmatic activity associated with continental break-up. The model proposed here can be applied also for some small-volume magmas not directly linked with continental break-up (e.g., Lustrino et al., 2004b).

### 2.2. Is there a role for the lower crust?

#### 2.2.1. Lithosphere pooling at the 660 km transition?

The rheology of the lower crust is critical for basaltic melts as it acts as a buoyant trap, especially in continental settings. Lower crust contamination of basaltic melts is a relatively common aspect recorded in several igneous provinces (e.g., Baker et al., 1997; Haase et al., 2004), although the process discussed here is source contamination rather than any mixing process (e.g., Lustrino et al., 2004b). Accordingly, an "active" role for the lower crust (or, rather, lower crust-derived partial melts) is proposed as a mantle source contaminant. In this model, the lower crust (coupled to a lithospheric mantle keel) subsides into the upper mantle and is not involved in any subduction process.

A roughly similar model was proposed by Korenaga (2004), based on (1) two-layered convection in the upper and lower mantle divided by the 660 km transition zone, and (2) the existence of subducted slabs remnants that penetrated the upper mantle and accumulated at the 660 km seismic discontinuity. This discontinuity is caused either by the breakdown of ringwoodite (γ-olivine) to perovskite and magnesiowustite/ferropericlase (Fei et al., 2004) or by the reaction garnet + ferropericlase = magnesiowustite + Na-rich phase (Gasparik, 2003, and references therein).

The fate of subducted slabs is a key question. With increasing depth, the basaltic (s.l.) assemblage (essentially plagioclase and augite) of the crustal portion of the slab is transformed to eclogite (mostly garnet and omphacitic pyroxene) its density increasing to 3.4–3.8 g/cm$^3$ (Wolf and Wyllie, 1994; Hacker, 1996; Tatsumi, 2000; Jull and Kelemen, 2001). Ultimately, at pressures between 410 and 660 km, such an assemblage would be transformed to garnetite (> 90% majoritic garnet and minor ferropericlase/magnesiowustite and ringwoodite; Irifune and Ringwood, 1993; Gasparik, 2003). Up to at least the 660 km discontinuity, the eclogitic/garnetitic slab is denser than ambient peridotite. Below this boundary, the density contrast of crustal components with ambient mantle is ~− 6% as consequence of the increase in density of the lower mantle (Hirose et al., 1999; Anderson, 2002; Fei et al., 2004; Korenaga, 2004). At pressures > 27 GPa (~720 km depth) basaltic compositional components are no longer buoyant as consequence of the development of perovskitite lithology (Hirose et al., 1999), suggesting that former basaltic crust will sink into the deep mantle if it accumulates to form a megalith with a thickness > 60 km (i.e., reaching depths > 720 km).

The transition layer between upper and lower mantle is considered to effectively decouple the uppermost basaltic portion of the slab from lower harzburgitic to lherzolitic portions, assuming realistic viscosity and temperature estimates (Karato, 1997). However, on the basis of global tomography, Ritsema et al. (2004) identified high-velocity slabs extending to about 1100 depth beneath several subduction zones (South America, Indonesia, Kermadec), suggesting penetration of the 660 km transition zone.

Korenaga's (2004) model considers that, at the upper–lower mantle transition, the crustal portion of the slab is decoupled from the ultramafic portion, in which case, assuming its viscosity is between ~$10^{21}$ and ~$10^{23}$ Pa s (Karato, 1997), the crustal material is folded and separated from the peridotitic portion. Accordingly, the crustal component can pond at the transition zone, forming a 50–200 km thick garnet-rich layer, whereas the peridotitic portion is recycled into the lower mantle. It is noted that recent estimates indicate that water is extracted from the lithosphere during the subduction process with greater than 92% efficiency (Dixon et al., 2002). If the subducted oceanic lithosphere is not totally dehydrated after subduction, incipient melting could cause the slab to soften, deform and bend as during the process of thermal equilibration with the surrounding mantle.

Evidence of garnetite from the base of the upper mantle includes majorite- (and other high pressure phases) bearing xenoliths from alnoitic (kimberlitic s.l.) pipes and sills from Malaita (Solomon Islands; SW Pacific Ocean; Collerson et al., 2000). Geobarometric estimates for Si-rich majorite (a complex solid solution of pyrope-almandine with orthopyroxene or clinopyroxene) range up to 22 GPa (~570 km), whereas the occurrence of xenoliths bearing Mg–Al-silicate perovskite increase the pressure estimates up to 27 GPa (~700 km; Collerson et al., 2000). These results are consistent with the notion that the upper–lower mantle boundary represents the site of accumulation of subducted oceanic crust and is a volumetrically significant mantle compositional reservoir. In any case, it is worth noting that according to other authors (Neal et al., 2001), the phases described as majorite and perovskite by Collerson et al. (2000) are simply pyroxene and amphibole equilibrated at depth of ~120 km ($<$ 3.6 GPa).

According to Korenaga (2004), crustal material accumulated at the base of the transition zone can be reactivated by vigorous sub-lithospheric convection assisted by plate-driven flow, and decompressing at shallower depths where it starts to melt. The presence of such material would enhance the melt productivity of mantle beneath a ridge axis, producing large-scale magmatic activity that characterize some continental rift systems (e.g., the North Atlantic Igneous Province; Korenaga and Kelemen, 2000) and the geochemical anomalies of some Paraná-Etendeka CFB igneous rocks (the Urubici and Khumib magma types; Peate et al., 1999; Ewart et al., 2004).

#### 2.2.2. Thinning vs. thickening of post-orogenic lithosphere

A potential problem with Korenaga's (2004) model relates to the initiation of sub-lithospheric convection (see his Fig. 3c). His model does not take into account the lithospheric thickening related to the continental collision stage of the Caledonian Orogeny but appeals to sub-lithospheric convection beneath thinned rather than thickened suture zones. Thinning of a suture zone could result both from shallow slab detachment (e.g., Carminati et al., 1998) and from delamination and detachment of a lithospheric keel, a key factor invoked in the model proposed here. Korenaga's (2004) model is also unable to explain within-plate igneous activity that characterizes large oceanic plateaux (e.g., Kerguelen and Ontong Java).

An alternative model for explaining the enhanced melt productivity of Atlantic CFBs and characteristic trace element and Sr–Nd–Pb isotopic compositions of some Paraná-Etendeka igneous rocks is lower continental crust recycling. After continent–continent collision (the Caledonian Orogeny for the North Atlantic and the Panafrican Orogeny for the South Atlantic) a period of tectonic subsidence, accompanied by the formation of intermontane troughs, sudden uplift, development of graben structures and continental rifting would typically lead to formation of oceanic crust. In general, isostatic relaxation follows regional shortening and thrust/nappe formation during continent–continent collision (e.g., Bonin et al., 1998; Lustrino, 2000). The only difference regards the time gap between peak metamorphic conditions and the beginning of crust formation, relatively short for the European Hercynides ($<$ 50 Ma), of medium duration for the Caledonides (~100–200 Ma) and very long for Atlantic Panafrican Orogeny (~300–400 Ma).

The gravitational instability of overthickened lithospheric mélange of the suture zone between two collided cratons can facilitate delamination and detachment of this keel (Fig. 1). Detachment is most likely to occur in the lower crust, due to the more ductile behavior of the latter compared to the upper crust and lithospheric mantle (i.e., the "jelly sandwich model"; e.g., Zuber, 1994; Handy and Brun, 2004). This model can be applied especially in cases of wet mafic (or felsic) lower crust associated to nearly dry lithospheric mantle, whereas cannot be proposed when a dry lower crust lies above a metasomatized, water-rich lithospheric mantle (see discussion in Alfonso and Ranalli, 2004). Thus, lower crust and lithospheric mantle components can delaminate and, eventually, detach given the density increase resulting from basalt/gabbro to granulite–eclogite transition (see summary by Lustrino, 2001). Under these conditions, crustal material may be incorporated into the mantle, without passing through subduction zones. This type of process is supported by: (1) rheological considerations (e.g., Kay and Mahlburg-Kay, 1993; Gao et al., 1998), (2) mathematical modeling (e.g., Schott and Schmeling, 1998; Morency and Doin, 2004), (3) experimental studies on basaltic (s.l.) compositions (Wolf and Wyllie, 1994; Rapp and Watson, 1995; Springer and Seck, 1997; Kogiso et al., 2003), (4) geochemical budget of whole continental crust (Kay and Mahlburg-Kay, 1991; Wedepohl, 1995; Rudnick, 1995; Gao et al., 1998; Rudnick and Gao, 2004), (5) geochemical modeling on basaltic rocks (Lustrino et al., 2000, 2004b; Tatsumi, 2000), (6) metasomatic changes in mantle xenoliths (e.g., Ducea and Saleeby, 1998), (7) thermodynamic constraints (Jull and Kelemen, 2001) and (8) evolutionary models on the formation and stabilization of the earliest continental crust (e.g., Hoffman and Ranalli, 1988; Zegers and van Keken, 2001).

#### 2.2.3. Lithospheric mantle vs. lithospheric mantle + lower crust delamination?

Rheological models favor either delamination and detachment of the entire lithosphere (e.g., lower crust + lithospheric mantle; Marotta et al., 1998) or its mantle portion only (e.g., Morency et al., 2002). Also the duration of delamination following compressive stresses is controversial. Morency et al. (2002) argued that the removal of a 250 km thick lithospheric root takes from 55 to 750 Ma depending on the root width and the viscosity contrast between the root and ambient mantle. Other key parameters include the duration and velocity of lithospheric delamination, the temperature difference between ambient mantle and delaminating material, the types and the kinetics of metamorphic reactions, and the bulk composition of the delaminating keel.

Lower crustal instability is also dependent on the average continental crust thickness which varies between 25 and 40 km, excluding active orogens, where crustal thickness is nearly double. According to Jull and Kelemen (2001), the relatively narrow thickness interval of the continental crust is not accidental and depends on metamorphic reactions occurring between 1 and 1.5 GPa (corresponding to depths of ~30–45 km) at $T$ of 800 °C or less. At these conditions, gabbro and gabbronorite compositions are denser than the underlying mantle and are therefore susceptible to delamination (Jull and Kelemen, 2001). In order to have lower crustal delamination, the crustal portion must be not too much cold (at low temperatures, even with crustal lithologies denser than the underlying mantle, the viscosity is so much high that a convective instability could not occur in geologically relevant times) and not too much warm (at high temperatures the lower crust viscosity is reduced but it might be buoyant as consequence of the reduced stability field of the garnet at temperatures > 800 °C; Kay and Mahlburg-Kay, 1993; Jull and Kelemen, 2001). The "window" in which lower crust is denser than upper mantle but its viscosity remains low enough to permit instability to develop is roughly comprised between 700 and 900 °C (depending on its composition; Jull and Kelemen, 2001).

Lithospheric delamination has been proposed to explain structures and magma geochemistry in several areas, including the Tibetan Plateau (England, 1993), Appalachian Chain (Levin et al., 2000), European Hercynides (Downes, 1993; Schott and Schmeling, 1998), Sierra Nevada batholith (Lee et al., 2000; Zandt et al., 2004), Andean Puna Altiplano (Kay and Mahlburg-Kay, 1993; Kay et al., 1994), Archean Pilbara (Australia) and Kaapval (South Africa) cratons (Zegers and van Keken, 2001), North China (Western Liaoning Province, Gao et al., 2004); also for the

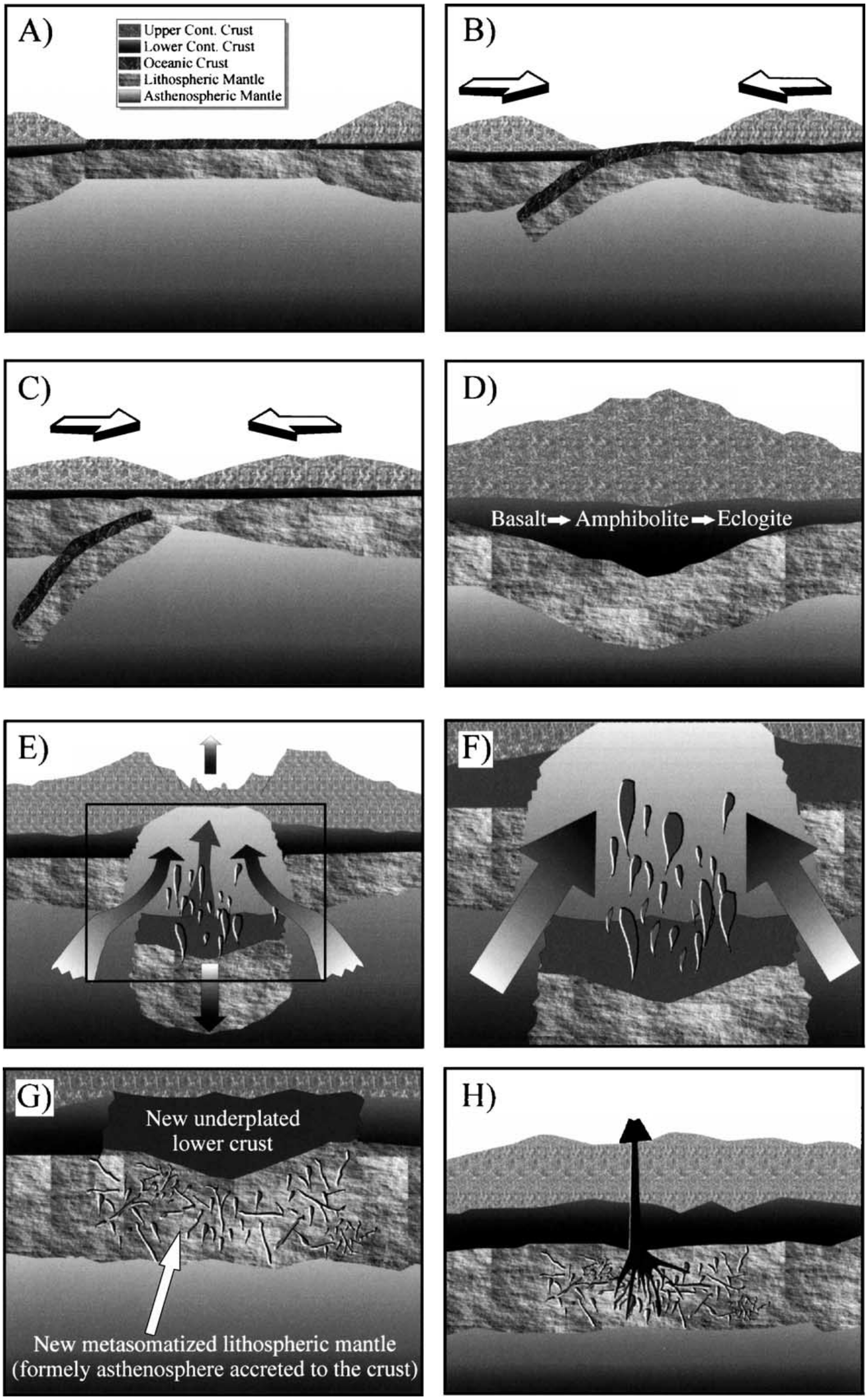
A)
Upper Cont. Crust
Lower Cont. Crust
Oceanic Crust
Lithospheric Mantle
Asthenospheric Mantle
B)
C)
D)
Basalt→ Amphibolite→ Eclogite
E)
F)
G)
New underplated lower crust
New metasomatized lithospheric mantle (formely asthenosphere accreted to the crust)
H)

Fig. 1. (A) Initial situation with thin oceanic lithosphere in between thick continental plates. Continental lithosphere is divided according to rheological behavior: (1) brittle $SiO_2$-rich upper continental crust; (2) ductile mafic lower crust; (3) lithospheric mantle, reaching a depth average of about 80 km, the crustal portion accounting for less than one half. This structure has been referred to as the "jelly sandwich model", weak, viscously deforming lower crust (jelly) intercalated between the overlying brittle upper crust and the underlying, stronger (sometimes brittle), upper mantle (the bread slices; Handy and Brun, 2004). (B) Under compressive stress, the oceanic lithosphere (colder and denser than continental lithosphere) is subducted beneath one of two continental plates, with relatively little deformation of lower continental crust. Limited subduction of the latter may be a consequence of mechanical erosion of the downgoing oceanic slab. (C) Continent–continent collision. Oceanic lithosphere has been completely subducted, its crustal portion having undergone dehydration with volatiles released to the overlying mantle wedge. Metasomatic effects of these modifications are not considered in this model. The density of crustal portions of oceanic lithosphere (a MORB protolith reaches up to 3.8 g/cm$^3$ at $P=14$ GPa) shows a sharp, discontinuous increase at ~ 9 GPa resulting from the coesite–stishovite phase transition (Aoki and Takahashi, 2004). (D) Upper continental crust is tectonically piled and thrusted, leading to thickening of the entire lithosphere, including the lower continental crust. Depending on the extent of pressure increase, original lower crustal basaltic/gabbroic paragenesis may be transformed to amphibolite ($P \leq 1$ GPa) to granulite/eclogite/garnet clinopyroxenite assemblages at higher pressures (~2–3 GPa; e.g., Wolf and Wyllie, 1994; Rapp and Watson, 1995; Jull and Kelemen, 2001). The pressure and temperature intervals of such metamorphic reactions depend mainly on the starting compositions of basaltic (MORB, alkali basalt, etc.) protoliths. The most important lower crustal metamorphic reactions may be summarized as: amphibole + plagioclase = garnet + melt ± plagioclase ± new amphibole (see text for further details), the net effect of which is a density increase by up to 3.5 g/cm$^3$ (e.g., Wolf and Wyllie, 1993). These metamorphic reactions are not strictly isochemical. During the formation of new phases, Rb and U are preferentially concentrated in the melt compared to Sr and Pb, respectively, while Sm and Nd are not strongly fractionated. Restites are therefore characterized by low to very low Rb/Sr and U/Pb and relatively unchanged (low) Sm/Nd ratios. The restite eclogite/garnet-clinopyroxenite thus evolves with low time-integrated $^{87}Sr/^{86}Sr$, $^{204}Pb/^{206}Pb$ and $^{143}Nd/^{144}Nd$. (E) The density increase leads to gravitative instability of an overthickened lithospheric keel. In particular, dense garnet-rich lower crustal restite allows for detachment of the lithospheric mantle from the upper lithosphere levels and its sinking into the asthenosphere. The depletion of lithospheric mantle in compatible elements during partial melting produces a density decrease (the restite has lower Fe/Mg ratio). The lithospheric mantle is thus neutrally buoyant, or buoyant with respect to the warmer, Mg-richer asthenospheric mantle. At first sight, this effect would tend to preclude lithospheric delamination, as required in the proposed model. However, the density difference ($\Delta\rho$) between (e.g.) Kaapvaal spinel- and garnet-bearing peridotites (characterizing lithosphere) and "pyrolite" (representing asthenosphere) is low (< 7% in low-$T$ peridotites and < 5% in high-$T$ peridotites; Kelly et al., 2003; see also Jull and Kelemen, 2001). This density difference is much smaller than the density contrast between garnet-rich restitic lower crust and lithospheric mantle (~15%), assuming $\rho_{\text{lower crust}}=3.8$ g/cm$^3$ and $\rho_{\text{lithospheric mantle}}=3.3$ g/cm$^3$. In conclusion, high densities of average lower crust contrast significantly with those of the lithospheric mantle and asthenosphere, implying strong likelihood of sinking of the overthickened lithospheric keel. According to the model presented here (Fig. 1E), downward motion of the lithospheric mantle is passive, in response to the negatively buoyant lower crustal push. (F) Detail of (E). During sinking, the lower crust is likely to undergo partial melting producing liquids of Tonalitic, Trondhjemitic, Granitic (TTG) and adakitic affinity (e.g., Springer and Seck, 1997; Defant and Kepezhinskas, 2001; Zegers and van Keken, 2001; Xu et al., 2002). Such melts would tend to percolate upwards as represented by $SiO_2$-rich glasses found in mantle xenoliths (from Sierra Nevada, California) interpreted as upper mantle products (Ducea and Saleeby, 1998). During lithospheric detachment, new asthenospheric mantle replaces the region vacated by delaminated lithospheric mantle and lower crust. According to this model, the asthenosphere accretes to the remaining lithosphere and becomes transformed to lithosphere on cooling. Isostatic uplift and formation of intermontane troughs accompany delamination as a consequence of upward impingement of hot, buoyant asthenosphere. Partial melts of the asthenospheric megalith underplate the remaining lithosphere to form the new lower crust and basaltic magmatism at surface. Thus, following development of intermontane trough and isostatic doming at the end of an orogenic cycle, igneous activity with a strong asthenospheric imprint is a common feature (e.g., Bonin et al., 1998), whereby the convecting asthenosphere is transformed into non-convecting lithosphere. This process is not simply a mechanical modification of the uppermost mantle but has significant geochemical implications. Lithospheric mantle formed from former asthenosphere is metasomatized by $SiO_2$-rich melts derived from the coeval sinking of lower crust. Such highly reactive melts would form orthopyroxene-rich zones, yet peridotitic in composition, therefore able to yield $SiO_2$-undersaturated melts at relatively high pressures (see more details in Yaxley, 2000). (G) After delamination of the lower crust and lithospheric mantle, asthenospheric counterflow, the contemporaneous partial melting of lower crust and mechanical accretion of the asthenosphere to the remaining lithosphere, the new mantle structure results in new lithospheric mantle comprising variably depleted peridotite, heterogeneously metasomatized, showing orthopyroxene-rich (lherzolite, olivine-pyroxenite, websterite) lithologies. This mantle source may remain unsampled for several Ma after the end of these processes. In these conditions lithospheric mantle metasomes evolve with peculiar crustal isotopic features, i.e.: (1) elemental Sr originally present in plagioclase is transferred to metasomatic melts during lower crustal partial melting; (2) the presence of residual garnet in sinking lower crust produces partial melts with strongly fractionated LREE/HREE evolving with very low $^{143}Nd/^{144}Nd$ isotopic ratios, and; (3) low $\mu$ ($\mu={}^{238}U/^{204}Pb$) crustal partial melts evolve with low $^{206}Pb/^{204}Pb$ isotopic ratios. (H) Metasomatized lithospheric mantle may be reactivated several Ma after lower crustal delamination occurred. Partial melts of such regions are likely to inherit lower crust-related metasomatic attributes, characterized by typical EMI-like geochemical features [e.g., low uranogenic Pb ratios ($^{206}Pb/^{204}Pb<17$), slightly radiogenic Sr isotopes ($^{87}Sr/^{86}Sr \sim 0.706$) unradiogenic Nd ($^{143}Nd/^{144}Nd \sim 0.5121$), unradiogenic Hf ($^{176}Hf/^{177}Hf \sim 0.2826$), slightly radiogenic Os ($^{187}Os/^{188}Os \sim 0.135$–$0.145$); Lustrino and Dallai, 2003]. Relative mantle-normalized Ba, Pb, Eu or Sr anomalies and variation of Ba/Nb ratios (3.5–47.4), Ce/Pb (1.2–24.6), Nb/U (10.5–71.8), Sr/Nd (6.2–36.4) and Eu/Eu* (0.83–1.25) ratios in EMI-type basalts (Lustrino and Dallai, 2003) reflect the effects of (1) lower crust starting composition, (2) metamorphic paragenesis and $PT$ parameters conditioning lower crust partial melting, (3) metasomatic reactions between $SiO_2$-rich melts and peridotite, (4) cratonization style of asthenosphere, (5) partial melting processes of newly accreted lithospheric mantle, and (6) fractional crystallization (coupled to potential crustal assimilation) of lithospheric melts.

Siberian traps (for which an involvement of a giant mantle plume is the standard accepted model) lithospheric delamination has been considered an important process (Elkins-Tanton and Hager, 2000).

Proposed scenarios include: (1) gravitative lithospheric delamination coeval with rifting stages of the Pangea super-continent (dismembered over a large time span, but generally between 300 and 100 Ma ago), (2) the dispersal of such material (lithospheric slices) within the upper mantle and (3) the subsequent random tapping by magmatic activity in continental and oceanic settings, including mid-ocean ridges (e.g., Mahoney et al., 1996; Hassler and Shimizu, 1998; Peate et al., 1999; Lustrino et al., 2000; Borisova et al., 2001; Frey et al., 2002, and references therein). The emplacement of fertile or volatile-rich material in the mantle, whether by subduction or delamination, clearly promotes melting at normal (rather than anomalous) thermal regimes. Likewise, low-velocity zones identified by mantle tomography may also reflect crustal material heated by ambient mantle to near-solidus temperature (e.g., Anderson, 2003).

In this regard, it is noted that experimental studies of amphibolitic assemblages indicate that metamorphic garnets formed at high P may host small inclusions of unreacted amphibole, thereby retaining up to 0.3 wt.% $H_2O$ (Wolf and Wyllie, 1994). Such garnet can, therefore, contain more structurally bound water than normal mantle minerals. Sinking of such a restite (eclogite or garnet clinopyroxenite) into the mantle could aid the delivery of water to greater depth (Wolf and Wyllie, 1994).

### 2.2.4. Isotopic model and constraints

Lower crustal recycling may be an important process in explaining trace element and isotopic features of the EMI mantle end-member (e.g., Lustrino and Dallai, 2003), despite numerous alternative hypotheses (e.g., Peate et al., 1999; Borisova et al., 2001; Kamenetsky et al., 2001; Thompson et al., 2001; Frey et al., 2002). Most of the latter appeal to lithospheric sources (considered as large fragments delaminated and dispersed into the upper mantle) for EMI (e.g., Mahoney et al., 1996; Douglass et al., 1999) without specifying the respective roles of crustal or lithospheric mantle contributions. It is noted that the LOMU (= low $\mu$, where $\mu$ is $^{238}U/^{204}Pb$ ratio) component of Douglass et al. (1999) is not the same as Zindler and Hart's (1986) EMI composition, being considered to show more radiogenic Sr and higher $SiO_2$ contents. However, in other cases only a minimal involvement of lithospheric mantle has been proposed, with major contribution to the geochemical budget of EMI-like basalts coming from subducted sediments and associated slab (e.g., Eisele et al., 2002; Ewart et al., 2004).

A role for the lower crust in generating EMI is supported by the following observations: (1) the most distinctive isotopic characteristic of the hypothetical EMI mantle end-member is unradiogenic $^{206}Pb/^{204}Pb$ ratio (< 17) plotting to the left of (or close to) the 4.55 Ga geochron (Lustrino and Dallai, 2003); (2) the only known terrestrial reservoir able to evolve to such low $^{206}Pb/^{204}Pb$ ratios is the ancient (Archean/Proterozoic) continental lower crust (e.g., Cohen et al., 1984; Zartman and Haines, 1988; Kempton et al., 1990; Liew et al., 1991; Rudnick and Goldstein, 1990; Halliday et al., 1993; Kramers and Tolstikhin, 1997; Liu et al., 2004). Other important features of EMI are its unradiogenic $^{143}Nd/^{144}Nd$ and only slightly radiogenic $^{87}Sr/^{86}Sr$ character. Compared to upper crust, Sr and Nd isotopic estimates of lower crust are displaced towards similar $^{143}Nd/^{144}Nd$ values but much lower $^{87}Sr/^{86}Sr$ isotopic ratios (e.g., Zartman and Doe, 1981). $^{187}Os/^{188}Os$ are higher than primitive mantle estimates and Depleted MORB Mantle composition (e.g., Shirey and Walker, 1998; Hauri, 2002; Escrig et al., 2004), thus reflecting involvement of crustal lithologies, characterized by several orders higher $^{187}Re/^{188}Os$ ratios than mantle values, due to the higher incompatibility of Re with respect to Os during partial melting processes). While trace element abundances and ratios vary considerably in both EMI- and EMII-like basalts, they are easily distinguishable from MORBs and HIMU-OIBs and clearly require the involvement of crustal material with (e.g.) low Ce/Pb, low Nb/U, relatively high LILE/HFSE ratios. Estimates of these canonical ratios for the lower continental crust are in general close to those for the upper continental crust values (e.g., Rudnick, 1995; Wedepohl, 1995; Gao et al., 1998; Rudnick and Gao, 2004).

It is worth noting that a single origin for EMI is unlikely. In particular, as highlighted by Mahoney et al. (1996), several EMI end-members probably exist, as evidenced by the relatively wide range of $^{207}Pb/^{206}Pb$ isotopic ratios among the most extreme EMI-like basalts (see Lustrino and Dallai, 2003 for further details). This effectively places the lower continental crust as only one of the several potential factors contributing to EMI basalt characters.

The two main alternatives proposed to explain low $^{206}Pb/^{204}Pb$ ratios in EMI basalts (i.e., reservoirs plotting left of the 4.55 Ga Pb geochron) are the Earth's core and garnetite slabs stored at the transition zone between upper and lower mantle. The first model invokes the different partitioning of U and Pb with respect to metal phases forming the core. U is lithophile whereas Pb is siderophile, thereby producing very low $\mu$ (~0) ratios in the metallic core (see discussions in Kramers and Tolstikhin, 1997). Notwithstanding the appeal of mantle plumes initiated at the core–mantle boundary, as suggested by several tomographic studies (e.g., Montelli et al., 2004), recent studies of W isotopes in Hawaiian basalts and South African kimberlites (Schersten et al., 2004) exclude significant contributions from the Earth's core to terrestrial magmas. Moreover, the evidence cited in favor of a core contribution to "plume" magmas (i.e., $^{186}Os$ excess resulting from high $^{190}Pt$ in the core; see Brandon et al., 1999) is readily explained in terms of recycled ferromanganese crust/nodules, strongly qualifying the notion that Re–Os isotope systematics uniquely constrain core–mantle interactions (Baker and Jensen, 2004).

With regard to the second hypothesis, Murphy et al. (2003), explained the first Earth's Pb paradox (the so-called future paradox; Kramers and Tolstikhin, 1997) in terms of a garnetite reservoir (derived from sediment + oceanic crust) stored within the 660 km mantle transition zone. According to this model, this material evolved in two stages: a first stage characterized by high U/Pb (high $\mu$) ratios (typical of the upper crust), and a second stage with low U/Pb, a consequence of U (and Th) depletion with respect to Pb during subduction processes involving the restitic slab. However, this reservoir can explain only part of the Earth's Pb paradox because the most extreme EMI basalts (e.g., those with $^{206}Pb/^{204}Pb < 17.3$) plot to the left of the estimated garnetite composition in $^{206}Pb/^{204}Pb$ vs. $^{207}Pb/^{204}Pb$ diagram, implying the existence of an additional reservoir in the Earth's mantle.

It must be remembered that isotope compositions define neither the location nor size of specific mantle "reservoirs" (actually components) in the shallow mantle, with no a priori requirement for deep recycling.

## 3. A new model for lithosphere delamination

A new model for the mechanism of interaction between lower crustal lithologies and mantle material is outlined here (Fig. 1). All the considerations are based on an equilibrium model hypothesizing immediate and complete transformation of basaltic lithologies to eclogite facies under appropriate $P$ and $T$ regimes. Of course, the kinetics of metamorphic reactions is generally lower and mostly is function of the grain size of the protoliths (e.g., Hacker, 1996). Essential features of the model presented here are as follows:

a) During a continent–continent collision, the lower crust, dominated by either underplated basalts or accretionary slices produced in subduction settings, is forced to greater depths by lithospheric thickening in response to thrusting and folding (Fig. 1A–D);
b) Under these conditions original basaltic or similar assemblages are replaced by amphibolitic assemblages at pressures of ~1 GPa and eclogitic assemblages at pressures of ~1.5–2 GPa, giving rise to a significant increase in density;
c) During such processes of tectonic piling and/or subduction, the lower crust undergoes partial melting (Fig. 1D);
d) The reactions associated with the transition basalt→amphibolite→eclogite/garnet clinopyroxenite are mostly governed by the incongruent (dehydration) melting of hornblende and plagioclase to give garnet, clinopyroxene, liquid ± a new, more aluminous hornblende (Wolf and Wyllie, 1994). If clinopyroxene is Na-rich, the restite becomes eclogite, if Na-poor, the restite becomes garnet-clinopyroxenite, depending on the starting composition (e.g., tholeiitic or alkali basalt) and the metamorphic $P$–$T$ path (e.g., Wolf and Wyllie, 1993). It is noteworthy that the increase of modal garnet (with a density of 3.6–4.0 g/cm$^3$; Hacker, 1996) up to > 40% increases the density of the lower crustal material (up to > 3.5 g/cm$^3$), thus enabling decoupling of this layer from garnet-poor

crust above it and sinking into the mantle (Wolf and Wyllie, 1993; Fig. 1E–F);

e) The contrasts in viscosity and density between lithospheric keel and ambient mantle allow for delamination and detachment of this root (Fig. 1E), such that thermal gradients at the keel margins may enhance convective circulation (e.g., King and Anderson, 1998);
f) Processes of delamination and detachment are probably confined to lower crustal + lithospheric mantle levels, where ductile behavior (as compared to the brittle upper crust) acts as a zone of weakness (e.g., Handy and Brun, 2004, and references therein);
g) Both lithospheric mantle and lower continental crust are thus expected to subside into the warmer asthenospheric mantle (e.g., Kay and Mahlburg-Kay, 1993; Fig. 1F).

There are several critical implications for the geochemical budget of the crust–mantle system. The most important of those are:

1) High-grade metamorphic reactions (granulite to eclogite facies) force crustal material to partially melt creating a low Rb/Sr and U/Pb restite, showing nearly unchanged high Sm/Nd (Fig. 1D).
2) Time-integrated decay within these isotopic systems (showing low U/Pb, low Sm/Nd and relatively low Rb/Sr) therefore produce low $^{206}Pb/^{204}Pb$, low $^{143}Nd/^{144}Nd$ and mildly radiogenic only $^{87}Sr/^{86}Sr$ ratios.
3) Following the delamination and detachment of lithospheric material due to increased density associated with eclogitization (Fig. 1D), the lower crust is forced to increasing depth and temperature. Under these conditions, the lower crust is susceptible to partial melting, producing liquids of dacitic/rhyolitic composition (e.g., Wolf and Wyllie, 1994; Rapp and Watson, 1995; Yaxley, 2000; Fig. 1E–F). The digested lithospheric keel is replaced by asthenospheric mantle which, in turn, may melt adiabatically in response to vertical movements. The latter may resemble "Andersonian" plumes as proposed by Courtillot et al. (2003).
4) $SiO_2$-rich liquids formed by partial melting of (delaminated and/or subducted) crustal material (dacite/rhyolite s.l.; e.g., Zegers and van Keken, 2001; Gao et al., 2004) will react with peridotite to form opx-rich lherzolite or olivine websterite (e.g., Yaxley, 2000) in the asthenosphere replacing the digested lithospheric keel (Fig. 1F–G).
5) Chemical and rheological transformations of the asthenosphere along the locus of detachment allow for adiabatic melting following release of latent heat of fusion, and increased viscosity (i.e., conversion to lithosphere) following conductive cooling from above. Accordingly, asthenospheric material is accreted to the base of the lithosphere (i.e., the remaining from the upper–middle crust) through cooling (Fig. 1G).
6) Orthopyroxene-rich metasomes are frozen into the upper mantle at various depths, infiltrating as dykes and dykelets or via porous flow (Fig. 1G). In these conditions they may be stored, hence tapped, for several Ma after the delamination process occurred by basalt magmatic activity (Fig. 1H).

The proposed contamination of mantle sources by lower crustal materials cannot be confused with magmatic contamination at lower crustal levels. In some cases, EMI signatures are observed in alkaline lavas associated with mantle xenoliths up to 20 cm in diameter, this being consistent with rapid magma velocities and the absence (or relative insignificance) of crustal-depth chambers (e.g., Lustrino et al., 2002, 2004c).

The proposed model requires mechanical decoupling between the lithosphere and asthenosphere (e.g., Doglioni et al., 2003). If, as proposed by Doglioni (1990) and Doglioni et al. (2003) the asthenospheric mantle is displaced eastward in response to the net effect of Earth's rotation, the asthenosphere currently below the Paraná-Etendeka igneous province at ~ 132 Ma is not the same asthenosphere that existed during the Panafrican Orogeny (~500–400 Ma) nor that present now. Consequently, a lithospheric keel that subsided into the asthenospheric mantle at the end of the Panafrican Orogeny cannot be at the same place during the early Cretaceous. The model presented here (lower crustal melts metasomatizing asthenospheric mantle that is replacing delaminated/detached lithosphere) predicts that the lower crust-related metasomatic effects are transferred to the top (to the asthenosphere that is replacing detached lithosphere) and there freeze. This would explain why the locus of stored lower crustal signatures is stored cannot be the asthenospheric mantle but, according to the model presented here, is necessarily the lithospheric mantle. Such metasomatized lithospheric mantle volumes may also be reactivated several Ma after the delamination process occurred.

A somewhat different model invokes the presence of widely dispersed lithospheric slices in the upper mantle (e.g., Mahoney et al., 1996; Borisova et al., 2001; Kamenetsky et al., 2001; Frey et al., 2002; Zhang et al., 2005). In this case, former lower crust remains coupled with lithospheric mantle keels (as garnet clinopyroxenite or eclogite restite) and can be reactivated only when sampled by large-scale convective cells (e.g., below oceanic rift zones) or randomly sampled during the specific tectono-magmatic evolution processes (e.g., Escrig et al., 2004). The presence of such pyroxenitic levels at mantle depths is indicated by model geochemical budgets for OIB (e.g., Hauri, 1996, 2002; Hirschmann et al., 2003, Kogiso et al., 2003, 2004), subduction related (Schiano et al., 2000, 2004), continental intra-plate (Gao et al., 2004, Zandt et al., 2004) and mid-ocean ridge (Hirschmann and Stolper, 1996; Escrig et al., 2004) magmas. The compositions of pyroxenite partial melts vary considerably, ranging from strongly nepheline normative (e.g., basanitic) to strongly quartz normative (e.g., dacitic/rhyolitic; Kogiso et al., 2004 and references therein) character. Mantle regions characterized by the presence of mafic (i.e., pyroxenitic) levels may be reactivated if the relative motions of colliding continental plates diverge and create a fracture zone (e.g., Zhang et al., 2005). Given that the solidi of mafic lithologies are ~20–200 °C lower than peridotitic assemblages (at $P$ ~3 GPa; Yaxley, 2000; Kogiso et al., 2004), partial melts of such mantle regions would show a strong crustal heritage. Since ancient collisional mobile belts represent the loci of continental break-up and CFB magmatism (Tommasi and Vauchez, 2001), the preferred orientation of mantle olivine crystals (representing compressive stresses fabric) may reflect large-scale anisotropy. Thus, attributes of ancient collisional belts may be reactivated during plate-driven continental rifting (Tommasi and Vauchez, 2001), allowing the incorporation of continental material, possibly along with trapped underplated oceanic material) by the shallow convecting mantle.

## 4. Concluding remarks

The lower crust and mantle portions of continental lithosphere exert strong structural and chemical controls on basaltic magmatism. Apart from the obvious controls of subduction on magmatic geochemical budgets and the role of buoyancy in trapping basaltic magmas, an alternative model is advanced here. A combination of lower crust and lithospheric mantle exert control on the composition of both continental and oceanic intra-plate magmatism, and on magmatism associated with continental break-up, related with particularly in regards to basalts of EMI affinity.

The garnetite model of Korenaga (2004), based on rheological considerations, is similar in many ways to that of Murphy et al. (2003), proposed on the basis of Pb isotopic criteria. Both models are able to explain geological–geophysical–geochemical parameters of continental rift evolution and CFB genesis (e.g., Ewart et al., 2004). However, a garnetite layer within the mantle transition zone cannot account for the most extreme compositions of continental and oceanic EMI basalts. Accordingly, a model is proposed that posits a role for lower crustal delamination and detachment in contributing to the upper mantle geochemical budget.

As presented here, the model reconciles the geochemical signature of volumetrically insignificant (but petrologically important) EMI-like magmas. However, the model can be seen as an alternative to crustal recycling unconnected to subduction through gravitative subsidence. Such a process is able to transfer significant amounts of $H_2O$ (in amphibole micro-inclusions in lower crustal metamorphic garnet) without recourse to subduction-related processes (where efficiency of $H_2O$ transfer to the mantle wedge is greater than 92%, resulting in virtually anhydrous crustal slices recycled into the mantle; e.g., Dixon et al., 2002).

Continent–continent collisions force lower crustal rocks to higher pressures and temperatures. Under such conditions, the orogenic lithospheric keel becomes gravitationally unstable and detached, sinking into the asthenospheric mantle (Kay and Mahlburg-Kay, 1991, 1993). The volume formerly occupied by the lithospheric keel is replaced by asthenosphere melts (transformed into new lower crust) and asthenospheric restite (transformed to lithospheric mantle, i.e., the mechanical boundary layer, by cooling from above). The lower crust, forced to great depths, may also succumb to partial melting, yielding liquids of dacitic/rhyolitic composition that react with the newly formed lithospheric mantle, forming orthopyroxene-rich layers. After freezing, such metasomes may also be reactivated several million years after lithospheric delamination occurred. The metasomatized lithosphere may thereby acquire strong crustal geochemical imprints as commonly observed in CFBs, as well as in continental and oceanic intra-plate basalts, and, rarely, mid-ocean ridge magmas (e.g., Kamenetsky et al., 2001).

In summary, sources in the lowermost mantle and anomalously high temperatures are not necessary to geochemically characterize OIB and CFB magmas or to achieve requisite large melt fractions. The alternative model proposed here is predicated on the specific location of lower crustal metasomatic signatures and their isotopic growth prior to partial melting. Crustal material (either continental lower crust or subducted oceanic slab) stored at the mantle transition zone is predisposed to melting at ambient mantle temperatures. Both the crustal and ultramafic parts of the slab sink because they are cold. When heated to ambient temperatures, surrounding mantle effectively being an unlimited heat source, most becomes buoyant. In this case, sinking of subducted lithospheric mantle would be arrested at the 660 km discontinuity, where crustal components begin to melt. Because of their buoyancy, crustal melts trigger plume-like upwelling without recourse to lower mantle convective upwelling (c.f., Korenaga, 2004). The presence of lithospheric material in the upper mantle can explain also the isotopic features of both CFB and their oceanic counterparts.

In conclusion, the role of sinking lower crust and lithospheric mantle is critical in determining the anomalous geochemical attributes in CFB, OIB and LIP magmas, delamination and detachment of which are supported by geophysical, geological, geochemical and petrological considerations.

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