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Westward migration of oceanic ridges and related asymmetric upper mantle differentiation

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ABSTRACT

Combining geophysical, petrological and structural data on oceanic mantle lithosphere, underlying asthenosphere and oceanic basalts, an alternative oceanic plate spreading model is proposed in the framework of the westward migration of oceanic spreading ridges relative to the underlying asthenosphere. This model suggests that evolution of both the composition and internal structure of oceanic plates and underlying upper mantle strongly depends at all scales on plate kinematics. We show that the asymmetric features of lithospheric plates and underlying upper asthenosphere on both sides of oceanic spreading ridges, as shown by geophysical data (seismic velocities, density, thickness, and plate geometry), reflect somewhat different mantle compositions, themselves related to various mantle differentiation processes (incipient to high partial melting degree, percolation/reaction and refertilization) at different depths (down to 300 km) below and laterally to the ridge axis. The fundamental difference between western and eastern plates is linked to the westward ridge migration inducing continuing mantle refertilization of the western plate by percolation-reaction with ascending melts, whereas the eastern plate preserves a barely refertilized harzburgitic residue. Plate thickness on both sides of the ridge is controlled both by cooling of the asthenospheric residue and by the instability of pargasitic amphibole producing a sharp depression in the mantle solidus as it changes from vapour-undersaturated to vapour-saturated conditions, its intersection with the geotherm at ~90 km, and incipient melt production right underneath the lithosphere-asthenosphere boundary (LAB). Thus the intersection of the geotherm with the vapour-saturated lherzolite solidus explains the existence of a low-velocity zone (LVZ). As oceanic lithosphere is moving westward relative to asthenospheric mantle, this partially molten upper asthenosphere facilitates the decoupling between lower asthenosphere and lithosphere. Thereby the westward drift of the lithosphere is necessarily slowed down, top to down, inducing a progressive decoupling within the mantle lithosphere itself. This intra-mantle decoupling could be at the origin of asymmetric detachment faults allowing mantle exhumation along slow-spreading ridges. Taking into account the asymmetric features of the LVZ, migration of incipient melt fractions and upwelling paths from the lower asthenosphere through the upper asthenosphere are oblique, upward and eastward. MORB are sourced from an eastward and oblique, near-adiabatic mantle upwelling from the lower asthenosphere. This unidirectional mantle transfer is induced by isostatic suction of the migrating spreading ridge.

1. Introduction

The theory of plate tectonics, the post-1960s version of continental drift (Wegener, 1915), proposes that the lithosphere (crust + uppermost mantle) is divided into a number of plates that can move over, as well as descend into, the underlying asthenosphere and even deeper. However, since the development of plate tectonics theory, the link between geodynamics and the compositional differentiation of lithospheric and asthenospheric mantles has received relatively little attention. Most commonly, both plates and underlying asthenosphere are treated in geophysical modelling as closed systems evolving separately with little mutual interaction. Petrological and structural studies of exposed mantle lithologies and experimental petrology demonstrate significant interactions between lithospheric and asthenospheric mantles at mid-oceanic spreading ridges. Moreover in recent years, geophysical observations of oceanic plates have demonstrated a significant asymmetry of the oceanic lithosphere on both sides of oceanic spreading ridge axis (Doglioni et al., 2003; Müller et al., 2008; Panza et al., 2010 and references therein) in terms of seismic velocities, density, thickness, and plate geometry (Fig. 1). These observations are fundamental and require an explanation that integrates the whole data in order to improve our understanding of how the lithosphere and asthenosphere interact at and near spreading ridges.

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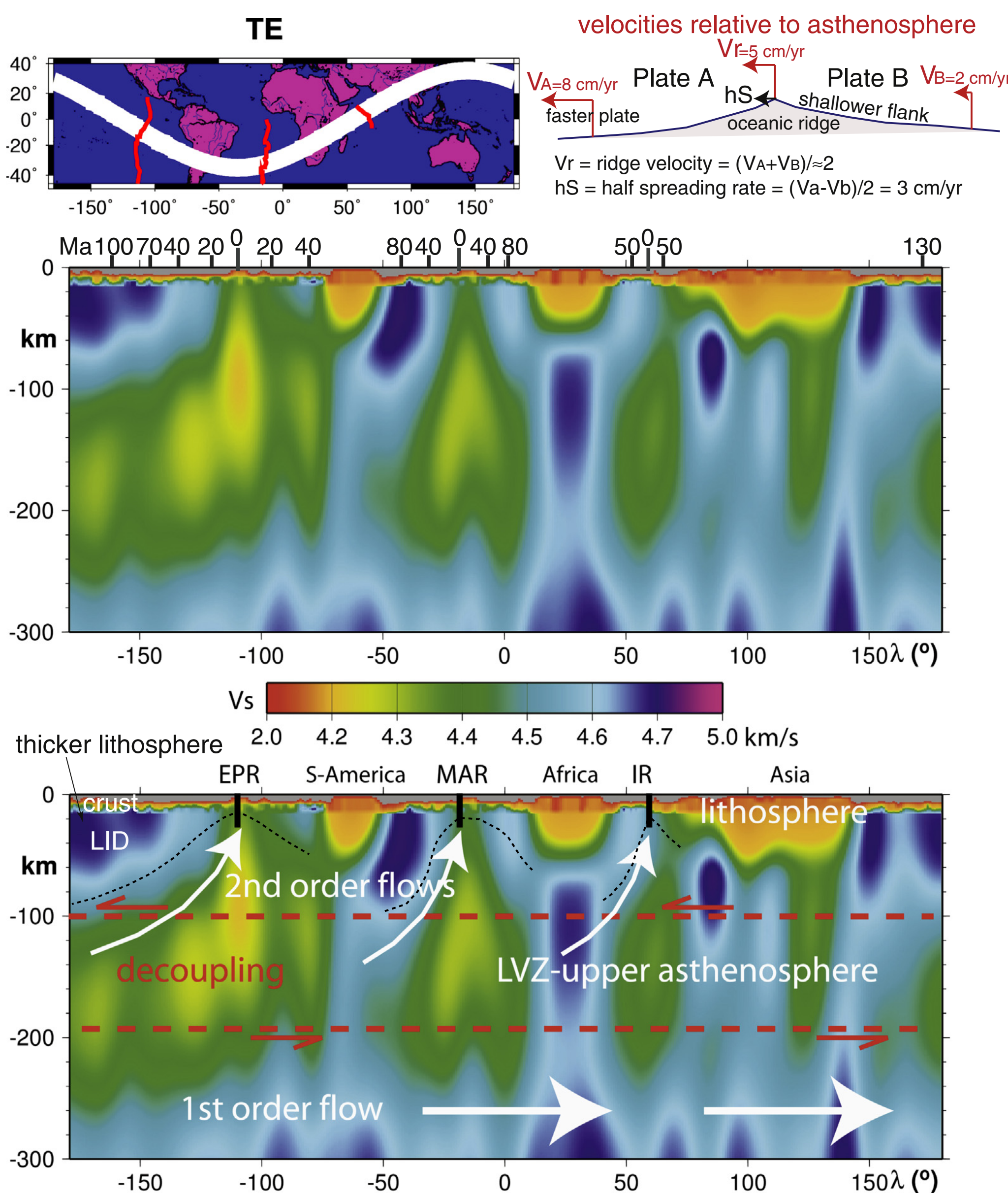

Fig. 1. Uninterpreted (above) and interpreted (below). Shear wave velocity (Vs) sections along the tectonic equator (TE) for the Earth's first 300 km. The lithosphere (LID) is faster and thicker in the western side with respect to the eastern side of the three major oceanic ridges, EPR—Eastern Pacific Ridge, MAR—Mid-Atlantic Ridge, IR—Indian Ridge. Upper asthenosphere superposes to low-velocity layer (LVZ), i.e., which is the main decoupling surface between lithosphere and lower asthenosphere, allowing net rotation of lithosphere, i.e., a first-order eastward relative mantle flow, or westward drift of lithosphere. Secondary flow should be related to mantle obliquely upraised along oceanic ridges. Asymmetry between two sides of ridges is independent from age of oceanic lithosphere, shown at top in million years.
(Modified after Panza et al., 2010.)

Since the 1960s, plate tectonics is conceived as driven either by active mantle upwelling ('bottom up' or alternatively 'ridge push') or alternatively by the negative buoyancy of slabs ('top down' or alternatively 'slab pull'). It is clear that creation of oceanic lithosphere and its subsequent subduction are fundamental mechanisms in governing the cooling of the Earth. In other words, the surface dissipation of both the Earth's internal primordial and radiogenic heat is tied with the re-entrance of the cold lithosphere along subduction zones to determine the cooling of the planet. However, the mechanisms determining mantle drag and mantle convection remain a matter of debate. The asymmetry of subduction zones (those directed to the west are steeper and faster, e.g., Doglioni et al., 2007; Riguzzi et al., 2010) and plate motion reconstructions relative to the mantle reference frame support a westerly-directed drift of the lithosphere relative to the underlying mantle, also called net rotation (as a mean value) or westward drift (Bostrom, 1971; Carey, 1958; Crespi et al., 2007; Cuffaro and Doglioni, 2007; Gripp and Gordon, 2002; Holmes, 1944; Le Pichon, 1968; Moore, 1973; Ricard et al., 1991; Wegener, 1915). The net rotation or the westward drift of the lithosphere requires both a decoupling at the lithosphere-asthenosphere interface and a mechanism driving this rotation (Doglioni and Anderson, 2015; Doglioni and Panza, 2015; Doglioni et al., 2011). The motion is not east–west, but along an undulated flow that varies from WNW to E-W to ENE trends, along the so-called 'tectonic equator' (Crespi et al., 2007). Proposed mechanisms involve either the negative buoyancy of the lithosphere (Ricard et al., 1991), or the astronomical drag induced by the Earth's rotation combined with tidal friction (Riguzzi et al., 2010). Potentially astronomical forces may interact with mantle convection resulting from the cooling of the Earth (Carey, 1958; Holmes, 1944; Scoppola et al., 2006).

The aim of our paper is to build an oceanic spreading model integrating existing compositional, structural and geophysical data across oceanic ridges and the underlying upper mantle. We show the interdependence between the oceanic spreading process, the upper mantle differentiation and plate kinematics driven by the westward drift of the lithosphere. Our model enables to integrate, in space and time, natural and experimental petrology results on upper mantle differenciation (down to 300 km depth) and oceanic, tholeiitic and alkalic, basalt genesis (Green, 2015; Green and Falloon, 2015; Green et al., 2014 and references therein), and provides a mechanism for upper mantle geophysical stratification with lateral heterogeneity as evidenced by Thybo (2006).

We will use the well accepted terminology of geophysicists, i.e., upper asthenosphere as the equivalent of "petrological' asthenosphere" corresponding to the Low-Velocity Zone (LVZ), and lower asthenosphere as the equivalent of "petrological' sub-asthenosphere" that reaches about 410 km of depth (Green and Falloon, 1998). They are defined in terms of depth, thickness and composition later in this paper.

2. Asymmetry on both sides of oceanic spreading axis and preliminary model

Since the recognition of magnetic anomalies on both sides of oceanic ridges (Vine and Matthews, 1963), oceanic rift zones were considered as having symmetric features, and plate spreading as a symmetric geological process. As shown by Doglioni et al. (2003), observations on the bathymetry of ridge flanks (shallower eastward) of Pacific and Atlantic oceans and on variable rates of oceanic spreading (Mallows and Searle, 2012; Müller et al., 1997, 2008) contradict the concept of full symmetry. Panza and Romanelli (2014) and Panza et al. (2010) provided evidence for asymmetry in both shear wave velocity and thickness within mantle lithosphere and asthenosphere when comparing the two sides of most spreading ridges. In general, the old (>60 Ma) western lithospheric plates have a faster shear wave velocity and are thicker ($\approx$100 km versus $\approx$80 km in the eastern flank). Seismic tomography (Panza and Romanelli, 2014; Panza et al., 2010; Schmerr, 2012; Thybo, 2006) provides robust evidence of an asymmetric LVZ at depths from about 80–100 km down to 180–225 km. The average Vs within the western LVZ is slower than the eastern one. The LVZ shape is itself asymmetric, being thicker and wider on the western side of ridges. Also passive continental margins and the related continent-ocean transition support a systematic asymmetry of rift zones (Brune et al., 2014; Lavier and Manatschal, 2006).

All these asymmetric features appear to be independent from the age of the oceanic lithosphere, although they become more prominent moving towards older lithospheric ages (Panza et al., 2010). These observations suggest that mantle compositions and related physical parameters such as rigidity μ, density ρ and their ratio μ/ρ, are systematically different from one side to the other of a spreading ridge within both the lithosphere and the upper asthenosphere, i.e., the LVZ.

Panza et al. (2010) explain these observations as a result of the depletion of the upper asthenosphere, that evolves from pre-melting lherzolite into post-melting harzburgite composition below the oceanic ridge, while the ridge is moving westward relative to the asthenospheric mantle, i.e., in the hot-spot reference frame (Crespi et al., 2007). The role of ridge migration (Doglioni and Panza, 2015; Scheirer et al., 1998; Small and Danyushevsky, 2003) in producing an oceanic plate asymmetry is a key parameter of the model. To the east of the ridge, the lithospheric mantle (named also LID) would represent the residual product from partial melting of upwelling asthenosphere, abandoned and in the process of cooling after the ridge migration to the "west". Since ridges are shifting above a fertile asthenospheric mantle, there is a continuous supply of MORB source mantle beneath the migrating spreading ridge (Panza et al., 2010). This model, deduced from geophysical and kinematic data (Fig. 1), suggests that the westward drift of the lithosphere relative to the underlying asthenosphere is a global phenomenon, a concept that implies decoupling between lithosphere and asthenosphere occurring within the LVZ, i.e., in the upper asthenosphere.

Important questions that arise from the above observations are as follows:

i. How does oceanic lithospheric mantle form?
ii. How can the mantle lithosphere composition evolve differently on each side of the ridge?
iii. How to combine oceanic spreading and westward drift of the lithosphere? Could the detachment-mode of seafloor spreading (Maffione et al., 2013; Whitney et al., 2013 and references therein), be the visible major effect in the field?
iv. How does asthenosphere evolve in relation with oceanic lithosphere during plate spreading? What controls the differentiation between lower and upper asthenosphere and the rather homogeneous thickness ($\approx$125 km) of the upper part?
v. How to combine the ascent of MORB source mantle from the lower asthenosphere according to experimental P-T-depth conditions (Green et al., 2014) with oceanic spreading and westward drift of the lithosphere?
vi. Is the oceanic lithosphere denser than the underlying asthenosphere?

3. Deciphering the compositional and physical evolution of the oceanic lithospheric mantle on both sides of a spreading ocean ridge

Lithospheric plates of slow spreading oceans (up to 4 cm/yr, e.g., Atlantic, Arctic, southwest Indian and Antarctic) are known to be mostly composed of mantle rocks. Volumes of extrusive and intrusive magmatic rocks, representative of the magmatic oceanic crust, remain minor (Cannat, 1993; Chalot-Prat, 2005; Smith et al., 2008, 2012). Lithospheric mantle is thus the main component of the oceanic plate, which reinforces the major role of oceanic mantle for understanding the nature and construction of the ocean floor. For this reason the following discussion (below) will focus firstly on the physical then secondly on the petrological features of the lithospheric mantle.

3.1. General physical features and composition of oceanic lithosphere

The oceanic lithosphere is deeply fractured at the ridge axis. This major fracture network reaches the top of the ductile layer in the lithosphere, and two plates, A and B, form and spread apart. The western plate A always moves westerly faster than the eastern plate B (Fig. 2) as shown by Crespi et al. (2007). In our model, due to the westward drift of the lithosphere (Doglioni et al., 2003; Le Pichon, 1968), both plates move to the west, the western plate A at VA and the eastern plate B at VB, being VA > VB (Fig. 2). The oceanic ridge moves west at Vr = (VA + VB) / 2. Plate growth rate is due to the difference between VA and VB. Thus the separation between points on each plate increases at the half spreading rate (hS) which is given by (VA − VB) / 2. Vr is always faster than hS.

The thickness of the oceanic lithosphere can be estimated from geophysical data including seismic, gravity, and electrical conductivity. It can also be inferred from geotherm/lherzolite solidus intersection based on experimental studies at controlled P, T and water contents.

Geophysical data show lithospheric thickness decreasing towards ridge axes above shallowing asthenosphere. However we do not think that the lithosphere-asthenosphere boundary (LAB) reaches the base of the newly formed basaltic crust at fast spreading ridges, nor approaches the sea floor at magma-starved slow-spreading ridges. This view is based on the petrology (phase assemblages), chemical compositions and isotopic compositions of peridotites sampled from rifted margins, ridge/transform intersections and slow-spreading ridges. Only lithospheric (secondary harzburgite or secondary lherzolite; see Piccardo et al., 2007b), but not asthenospheric mantle compositions have been identified either at the bottom of rift axis in slow-spreading

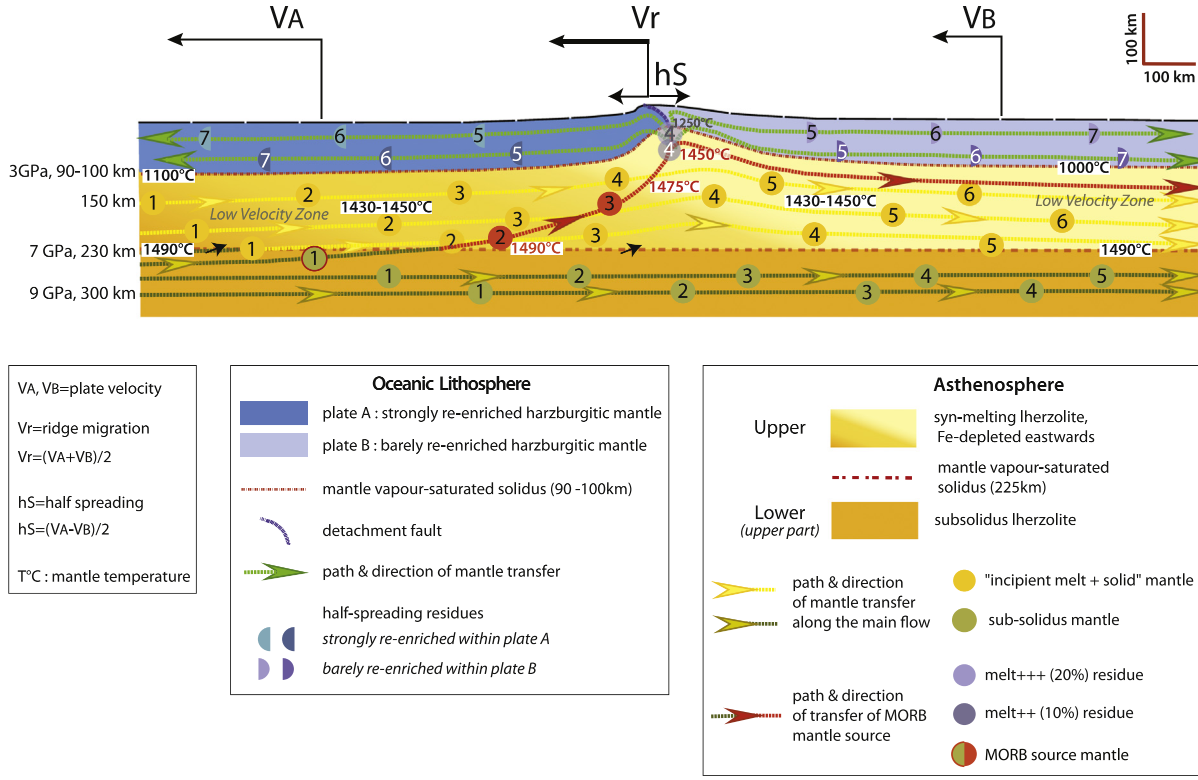

Fig. 2. Inferred kinematics of different particles of mantle rocks along a migrating mid-oceanic ridge, and related asymmetric upper mantle differentiation. Numbers from 1 to n indicate progressive age evolution of the rock markers within lithosphere and asthenosphere. Plates A and B move with velocity VA and VB, respectively. Vr is the velocity of the ridge; hS is the half-spreading rate. Since the ridge is moving westward, new sections of asthenospheric mantle migrate "eastward" relative to the lithosphere, permanently renewing mantle source of isostatically upwelling magma generated by depressurization below the ridge. During this mass transfer, the eastern part of upper asthenosphere becomes somewhat Fe-depleted, while a top to bottom mantle path decoupling occurs due to the relative "westward" drift of overlying lithosphere. The upper asthenosphere (LVZ) corresponds to the oblique ascent of partially molten mantle from lower asthenosphere. Two different paths of transfer exist: the main one (yellow paths) barely deviated in passing below the ridge, the other (red path) monitored by the suction effect from the strongly depressurized area below the spreading ridge and giving MORB (see Fig. 3 for details). Mantle lithosphere is created at the ridge through accretion of residues above the asthenospheric partial melting area, residues successively transferred upwards and laterally at hS rate within the mantle lithosphere on each side of the ridge. A top to bottom decoupling exists within the mantle lithosphere linked to the slow-down by the eastward underlying flow of asthenosphere, and is responsible for detachment faults and mantle exhumation at the spreading ridge. Besides the mantle lithosphere composition, different on each side of the ridge, comes from a permanent refertilization of the western plate whereas the eastern plate just preserves the barely refertilized feature of a harzburgitic residue generated during asthenosphere partial melting below axial ridge. See text for more details.

active and fossil oceans (e.g., Dijkstra et al., 2001; Kaczmarek and Müntener, 2008, 2010; Muntener et al., 2010; Müntener et al., 2004; Piccardo and Guarnieri, 2010; Piccardo et al., 2004, 2007a, 2007b, 2014; Rampone and Borghini, 2008; Rampone and Hofmann, 2012; Rampone et al., 1997, 2005, 2008; Seyler and Bonatti, 1997; Takahashi, 2001; Tartarotti et al., 2002; Warren, 2016; Warren and Shimizu, 2010), or below oceanic crust in fast-spreading active and fossil oceans (Abily and Ceuleneer, 2013; Dick and Natland, 1996). The depth of the LAB under the ridge may be inferred in two ways: 1. estimating the depth of melt segregation for parental MOR picrites or basalts and 2. by the P, T conditions recorded in ridge peridotites (including their P, T decompression paths in some cases). Indeed experimental results (Fig. 3) demonstrate that parental MOR picrite segregation from residual peridotite occurs mostly between 2 to 1.6 GPa (up to 2.2 GPa; down to 1.1 GPa), thus between 55 to 45 km depth (down to 65 km; up to 30 km), around 1400 °C (up to 1430 °C) from a trace element depleted lherzolite (Eggins, 1992a, 1992b; Falloon et al., 2007a, 2007b; Green et al., 2001; Green and Falloon, 2005; Green et al., 2014; Green and Falloon, 2015). Magma genesis produces mantle residual peridotite less dense than the asthenospheric protolith, accreting above its asthenospheric source and potentially subject to percolation-reaction of new magma ascents, giving birth to a refertilized constantly growing oceanic lithospheric mantle. From experimental results on liquid compositions in equilibrium with spinel and/or plagioclase lherzolite assemblages (Chalot-Prat et al., 2010, 2013; Falloon et al., 2007a, 2007b), analogous to natural oceanic mantle samples, the oceanic lithospheric mantle thickness at the ridge is at least 20 km and up to 40 km in thickness. Note that density differences are due to chemical/mineralogical differences but also to temperature differences (e.g., residual mantle temperature has been reduced by the loss of the latent heat of melting). Thus determining density differences is subject to significant uncertainties.

Laterally on both sides of the ridge axis, seismic data show a progressive thickening of oceanic mantle lithosphere until around 80–100 km maximum, equivalent to 3 GPa of pressure. This thickening is assumed to derive first from the addition of the cooled uppermost part of asthenosphere (below 1100 °C; Green and Falloon, 2005; Green et al., 2014). Green and Falloon (2005) and Green et al. (2010, 2011, 2014) have shown experimentally that the lithospheric mantle thickness depends on its water storage capacity (Fig. 3). Indeed at pressures greater than 2.5 to 3 GPa (80 to 100 km depth), instability of pargasite causes a sharp drop in the water storage capacity of fertile lherzolite, and for water content greater than ~200 ppm in the lithosphere, the mantle solidus is depressed at 2.5–3 GPa along the geotherm. In both cases partial melting can explain both the thickness of the oceanic lithosphere and the occurrence of a LVZ just below the LAB. Therefore from a petrological point of view, the LAB detected by the slowing of seismic waves is the intersection of the geotherm with the vapour-saturated lherzolite solidus, and the LVZ is a layer in which there is incipient melt (probably <1–2% melt).

As far as it concerns the depth/temperature of the base of the oceanic lithosphere outside the ridge (>500 km), the model of Afonso et al.

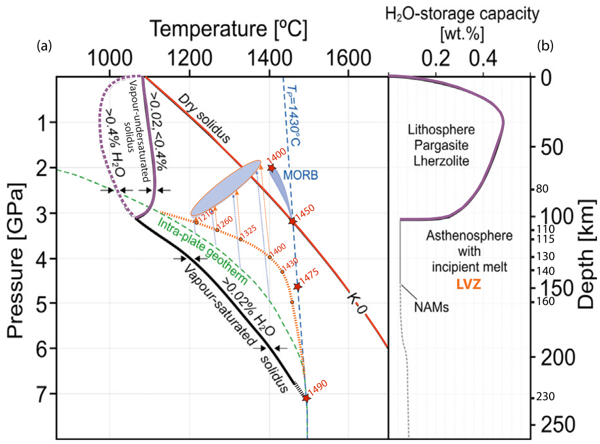

Fig. 3. P-T-H_2O diagrams, modified from Fig. 6 in Green et al. (2014). a) Experimentally determined solidi for hydrous silicate melt in fertile lherzolite for different water contents; *blue dashed line*: mantle adiabat; *red star*: MORB source mantle upwelling potential temperature at 230, 150, 105 and 65 km; *green dashed line*: intraplate geotherm; *orange dotted line:* perturbation of oceanic intraplate geotherm by shear heating, adapted from Fig. 4 in Doglioni et al. (2005); *orange dot*: intraplate magma genesis at shallower depths (105 to 125 km vs. 110 to 160 km) taking into account shear heating within the lithosphere–asthenosphere decoupling zone. MORB are sourced from upwelling trace element depleted lherzolite from lower asthenosphere. Intraplate basalts, including OIB, are the products of upwelling of trace element enriched lherzolite from the middle and upper asthenosphere. See text for more details. b) Water storage capacity of lherzolitic mantle as a function of depth along the vapour-undersaturated solidus. Pargasite becomes unstable at >3 GPa, and water storage capacity drops to that (~200 ppm) which can be retained in NAMs (nominally anhydrous minerals) in lherzolite, inducing incipient melting of upper asthenosphere (LVZ). See text for more details.

(2008) predicts 1300 °C at 105 km depth. This temperature is somewhat too high according to experimental data of Green et al. (2010, 2014), showing that a “wet” (>200 ppm H_2O) oceanic lithospheric mantle is above its solidus below 80–100 km if temperature is higher than 1000 °C (Fig. 3). Below this depth, this is the LVZ, thus the upper asthenosphere. An incipient melt is present, the melt fraction being determined primarily by the water and CO_2 contents. This melt has low permeability (i.e. can only move slowly by porous flow), but has a large effect on seismic and rheological properties.

3.2. How oceanic mantle lithosphere of the western limb of the ridge axis becomes thicker and denser compared to the eastern limb?

3.2.1. Intra-lithospheric mantle decoupling

In this section we will focus on the modalities of mantle mass transfer upwards and laterally within both spreading plates in the framework of the westward drift of the lithosphere. Oceanic mantle lithosphere is created at the ridge through accretion of residues above the asthenospheric partial melting area. These residues are successively transferred upwards and laterally at hS rate within the mantle lithosphere on each side of the ridge (Fig. 2). As Vr is higher than hS, the transfer rate of successive residues above the melting area is slower than the westward velocity of the melting area itself. Accordingly, mantle exhumed at the bottom of any active axial ridge is always older than the basaltic volcanoes emplaced at the axial ridge (Bonatti et al., 2003). It is the same for intrusive gabbro bodies which move upwards and laterally, driven by and with the upwelling of the host-mantle: for instance, for hS = 2 cm/yr, Vr = 4 cm/y and a mantle residue initial depth of Di = 40 km, the mantle residue takes t1 2 Ma (t1 = Di/hS) to ascend and to be exhumed at the surface. Thus active volcanoes at the ridge axis are about 2 Ma younger than the mantle on which they are formed. Besides, while this mantle residue is rising up to the ocean floor, the ridge itself with the uppermost part of the lithosphere will have moved (t1 ∗ Vr) 80 km westward. However it may be possible that due to ‘second-stage melting’, the extra buoyancy added to the residue may accelerate the upwelling – in which case some residue may be younger than 2 Ma.

Therefore, besides the notion that asthenosphere is moving “eastward” relative to the lithosphere (or more precisely along the flow lines parallel to the undulate tectonic equator of Crespi et al., 2007 and Cuffaro and Doglioni, 2007), the westward drift of the lithosphere is necessarily slowed down, top to down, inducing a decoupling within the mantle lithosphere itself. This is visualized on the model (Fig. 2) with the progressive top-down eastward shifting of the “half-spreading residues” relative to each other within both plates. This decoupling would be progressive such that the top of the melting area would be somewhat delayed relative to the top of the axial ridge, even if they are interdependent.

Normal faults are the natural feature associated with rift zones and as indicated by rock mechanics, their mean dip is around 60°. Lower dips may naturally result from a decrease of friction due to the high heat flow and moving into shear zones within the underlying ductile layer. Besides faults and fractures may dip predominantly in one direction if spreading is asymmetric. Indeed the half spreading rate is often slightly asymmetric (Müller et al., 2008), but this is related not to the velocity of plates relative to the mantle, but to the velocity of the ridge which sometimes is not exactly (VA + VB) / 2; it may be (VA + VB) / 1.95 or (VA + VB) / 2.05, etc. Therefore faster and larger spreading may occur in one flank without varying the relative velocity of the two plates with respect to the mantle. This can be explained by the lateral viscosity variations in the flanks right beneath the ridge. Therefore, ridge-related faults and associated fractures would dip mainly “eastward” and be convex upwards since the westward drift is faster upwards, but their dip may also be opposite due to the half-spreading process. This is the geometry of detachment faults, i.e., large offset low-angle faults capping the eastern (rarely western) side of oceanic mantle core complexes (OCC; MacLeod et al., 2009; Maffione et al., 2013; Mallows and Searle, 2012; Reston and Ranero, 2011; Smith et al., 2008, 2012; Whitney et al., 2013) and extending along steep faults probably rooted in a melt-rich zone. They are responsible for exhumation of mantle and plutonic rocks onto the seafloor in the footwall of normal faults, whereas extrusion of basalt occurs on the hanging wall.

This intra-mantle decoupling within the lithosphere means also that even a mid-oceanic ridge and its related fracture network shift probably westward within the lithosphere, moving itself westward relative to the asthenosphere. This cannot be visible from the surface, and this is why general symmetry of magnetic stripes on the oceanic sea floor is preserved. However Carbotte et al. (2004) and Mallows and Searle (2012) note a westward shift of volcanic activity with time along the active Mid-Atlantic ridge axis. The geological maps of Mallows and Searle (Fig. 4; 2012) and Smith et al. (Fig. 4; 2012) show that most OCC outcrops on the western side of the ridge and that spreading rates are asymmetric with faster spreading on western plates. According to Whitney et al. (2013 and references therein), OCC account for 60% to 100% of the total plate spreading, showing that tectonic process is the main driver of mantle exhumation in the oceans and of heat and mass transfer in the Earth. Furthermore, Maffione et al. (2013) introduce the concept of detachment-mode of seafloor spreading. Moreover, the usual sill shape of gabbro intrusive bodies both at multimetric and multikilometric scale (Henstock et al., 1993 and references therein; Chalot-Prat, 2005) testifies sub-horizontal magma injections within weakness zones of lithospheric mantle, supporting intra-mantle decoupling during ocean spreading. Volcanoes grow on the topmost part of the hangingwall of these large detachment faults, structuring hangingwall rider blocks (Maffione et al., 2013; Whitney et al., 2013 and references therein). Small-scale decoupling occurs between the base of volcanoes just emplaced and underlying mantle or/and gabbro bodies in the process of exhumation. This decoupling occurs on both sides of relief where eruptions take place, underlining a pseudo-symmetric spreading (Chalot-Prat, 2005).

3.2.2. Asymmetric evolution of lithospheric mantle compositions and physical parameters on each side of the ridge

The variation of Vs between the two sides of the oceanic ridge should have an explanation in terms of composition related to the different kinematics. As the melting area moves westward with the migrating spreading axis, the eastern plate B incorporates successively the abandoned depleted eastern parts of the ridge. It undergoes a relatively brief time of refertilization by ascending MORB and its composition remains close to a harzburgitic residue and is relatively Fe-poor. At the same time, harzburgitic residues are also continuously added to the western plate A, but, in contrast to the eastern plate B, they are also continuously percolated by MORB as the ridge is moving westward above the underlying melting area linked to the ridge location. Thus the composition of the western plate A is that of a strongly refertilized harzburgite, thus a secondary lherzolite. An outcropping example could be the lherzolites from the Lanzo oceanic mantle area (Piccardo et al., 2007a, 2007b, and references therein). Therefore lithospheric mantle of the plate A is continuously re-enriched in Fe and Ca, whereas lithospheric mantle of the plate B remains a harzburgitic residue barely re-enriched in Fe and Ca.

Nevertheless as Vr > hS, the axial ridge is moving westward within the lithosphere itself, and the western plate A is progressively incorporated within the eastern plate B. This means that the plate B composition will become somewhat more refertilized than presented above. However the overall consequence is that the western plate A is thicker, always richer in Fe and thus denser than the eastern plate B, resulting in a first order asymmetry in composition between the two plates.

In addition, if the westward drift is producing intra-lithospheric deformation preferentially on the western plate – then we might expect easier and more abundant pathways for off-axis magmatism and hence off-axis melt impregnation and channelling. Some of the upwelling diapirs may lose part of their melt fraction before arriving directly under the ridge axis. Evidence for this could be an asymmetry in abundance of off-axis seamounts (i.e. volcanic chains on western plate as in Shen et al., 1993, White et al., 1998). If this off-axis process occurs – then there will be less melt arriving directly under ridge axis – and hence the eastern flank will receive long-term less fertilized residue than the western flank.

From seismic data of Panza et al. (2010), Vs ($=\sqrt{\mu/\rho}$; μ: rigidity; ρ: density) of the western plate A is on average faster than the eastern plate B. As ρA is higher than ρB (previous paragraph), μA must be much higher than μB. Since the modulus of rigidity of Fe is around 3 to 4 times higher than that of Mg and Si, the rigidity of rocks is expected to be enhanced if they are richer in Fe-rich minerals (Fyfe, 1960; Jordan, 1979). Thus higher μA compared to μB is an expected natural consequence of enhanced melt refertilization of the western plate A during westward migration of the spreading ridge.

4. Deciphering the physical and compositional evolution of the asthenosphere below a spreading oceanic lithosphere

4.1. Lower asthenosphere

The lower asthenosphere is subsolidus garnet lherzolite, about 180 km thick (230 to 410 km depth) overlying the Transitional Zone beginning at ~410 km (Anderson, 2007, 2010; Schmerr, 2012). From the top to the bottom of this layer, seismic waves gradually increase their speed with respect to the LVZ. Vs is much higher than in the upper asthenosphere and more or less similar to that of the lithosphere (Panza and Romanelli, 2014). Its mineralogical and chemical composition is inferred from experimental petrology on MORB genesis (Green, 2015; Green et al., 2014), being the N-MORB mantle source located below 230 km. It is a fertile garnet lherzolite, depleted in the most incompatible trace elements, with ≥200 ppm H_2O and with incipient melt at ≤7 GPa/~230 km, 1490 °C. As shown in Fig. 3, an oceanic intraplate geotherm joins the mantle adiabat at about 230 km depth defining the base of the upper asthenosphere. The lower asthenosphere may migrate upwards as diapirs within the 230–100 km including very small, relatively immobile, near-solidus melts (0.05–0.1% of melting).

4.2. Upper asthenosphere

Within the upper part (80–220 km depth) of the asthenosphere, i.e., the LVZ, Vs is slower in the western than in the eastern side of the ridge, the reverse of what it is observed in the oceanic lithosphere (Panza et al., 2010). We infer that during its transfer from west to east below the oceanic lithosphere, the eastern part of upper asthenosphere becomes somewhat Fe-depleted by incipient melting (Fig. 2) inducing a density decrease for a potentially similar value on rigidity from one side to the other of the axis. Moreover as shown by experiments of Conder and Wiens (2006), Green et al. (2014), and Hammond and Toomey (2003), the western part of the upper asthenosphere is hotter and/or richer in $H_2O + CO_2$ than its eastern part and includes higher melt fractions inducing a rigidity decrease in the western side. It follows that to the “west” of a spreading ridge, both the oceanic lithospheric mantle (down to 90–100 km) and the upper asthenosphere (down to 230 km) should be more fertile and denser than the “eastern” part. Let us remember that in both mantle layers, density is determined not only by mantle fertility and Fe-enrichment, but also by the pargasite content in the lithospheric mantle and by percentage of melt within the upper asthenospheric mantle.

The LVZ has been seismically recognized below all oceans, both beneath the “western” and the “eastern” plates. It has a very small aspect ratio (thickness: 130 km; length in spreading direction: $6 * 10^3$ km below the Atlantic to $11 * 10^3$ km below the Pacific). Experimental results (Fig. 3) show that the upper asthenosphere corresponds petrologically to a mantle layer including small interstitial melt fractions (i.e. 0.05 to 0.1%). According to Doglioni et al. (2005), the degree of partial melting would be increased (up to 1.5%) by shear heating (more than 100 °C) generated by decoupling between lithosphere and lower asthenosphere (Fig. 3) and with largest contribution i.e. largest E-W relative movement at ~230 km depth. It may be alternatively due to higher H_2O content in the mantle (Bonatti, 1990; Ligi et al., 2005).

Moreover, seismic images from Panza and Romanelli (2014) show an asymmetric shape of the LVZ, thicker and more elongated westward (2/3 in volume) than eastward (1/3 in volume) of ridge axis in all oceans. This spectacular and permanent asymmetry of the LVZ shape raises the question not only of the origin of partial melting, but also of its much greater development below the western part of the oceans. From experimental results (Fig. 3), the presence of these very small melt fractions has two distinct origins depending on the H_2O content of the mantle sources: 1- below the lithosphere-asthenosphere boundary (LAB) and far away from the ridge (>500 km): because of destabilization of amphibole, below 2.5–3 GPa/80–100 km/1000–1100 °C, when “wet” (>200 ppm H_2O) oceanic mantle lithosphere, i.e. a subsolidus pargasite-bearing garnet lherzolite cools, subsides and becomes asthenosphere, i.e. a pargasite-out garnet lherzolite with incipient melting; 2- above the lower-upper asthenosphere boundary: during adiabatic (≈1490–1450 °C) “dry” (<200 ppm H_2O) ascent of the lower asthenosphere from ≥230 km/≥7 GPa towards the MORB genesis area, i.e. 65–45 km/2–1.6 GPa below the ridge axis. Let us note that if the first partial melting origin corresponds to a rather static process linked only to the permanent thickening of the lithospheric mantle, the second is synchronous with a dynamic process of partially molten transfer of matter, both being imaged by the LVZ.

In parallel, the LVZ asymmetry suggests that the on-going ascent of partially molten mantle from lower asthenosphere is oblique, and not vertical, as already proposed by Carbotte et al. (2004). Taking into account its dimension, the ascent would be initiated around some thousands kilometres “west” from the ridge (Fig. 2), and driven by a suction effect towards the shallow area below the spreading ridge moving "westward", due to the removal of the overlying lithosphere and the consequent asthenospheric isostatic compensation.

Nevertheless according to the entrance angle of lower asthenosphere into upper asthenosphere, this lateral and upward mass transfer is more or less counteracted by the relative "eastward" horizontal mantle flow linked to the net rotation of the lithosphere, whose origin can be ascribed either to lateral viscosity variation (e.g., Ricard et al., 1991) or to the Earth's rotation (Riguzzi et al., 2010; Scoppola et al., 2006). Two different paths of transfer would exist. The main path involving the largest mantle volume follows a very low angle trajectory, barely deviated in passing below the spreading ridge; it would concern the mantle flux emerging rather far laterally (more than 500 km?) from the spreading centre. From experimental results (Fig. 3), this mantle flux crosscuts the intraplate geotherm, modified by shear heating, and undergoes incipient to low degrees of partial melting (up to 4% down to 80–100 km deep/2.5–3 GPa, Green and Falloon, 2005). The other path is single inasmuch as, all things (composition, flowing velocity, temperature, pressure) being equal, a more opened entrance angle of mantle flux at a shorter distance from the ridge (500 km or less?) enables it to be mainly monitored by the suction effect coming from the strongly depressurized shallow area below the spreading ridge. This single path is assumed to be the one producing MORB after 15–20% melting between 65 and 45 km depth (2 to 1.6 GPa) (see Section 5.1 below). Indeed at small-scale, we suggest that convective eddies form by peeling off the upper surface of the lower asthenosphere and ascend driven by buoyancy and rheology contrast with the upper asthenosphere. These ascending eddies would become 'isolated' or encapsulated by a cooling rind/skin as the temperature contrast increases between eddy core and ambient mantle, forming diapirs. These eddies/diapirs move obliquely towards the shallow area below the spreading ridge where they are tapped by fracturing at the rift.

"Eastward" of the ridge, because of the "westward" drift of the lithosphere, the suction effect fades and mantle melting progressively mitigates top to bottom to become quite insignificant at more than 1000 km from the ridge. Accordingly in this model, off-axis melt lenses on the eastern side of the East Pacific Ridge (Canales et al., 2012; Toomey, 2012; Toomey et al., 2002) witness the asthenospheric partial melting zone underlying the ridge just before eastward abandonment due to the westward migration of the ridge.

According to this new scenario, the westward drift of the lithosphere and the related oceanic plate spreading have a strong mechanical effect on lateral (several thousand kilometres long) and upward (from 230 to 45 km depth) mantle mass transfer below the western plate A. This effect appears as extremely relevant in terms of mass transfer and is synchronous with partial melting and percolation/reaction processes, and thus mantle differentiation.

Therefore, as already inferred by Green (2015) and Green et al. (2014), the upper asthenosphere mantle composition is heterogeneous at a small scale, being influenced from below by the ascending lower asthenosphere and from above by the oceanic lithosphere during its journey and related process of thickening. In other terms, from 230 km to 90–100 km, the upper asthenosphere should be a mantle layer including incipient melt fractions coming from two distinct mantle sources: the lower asthenosphere, the "dry" lherzolitic source of MORB on one hand; the lowermost part of lithosphere, made of harzburgitic residues produced below the ridge axis, more (westward) or less (eastward) refertilized and "wet" once accreted to the lithosphere.

All these data confirm that the asthenosphere forms a whole, 300–320 km thick, with a lower "nearly dry" (~200 ppm H_2O) solid part ($\approx$185 km thick; 2/3 in volume) and an upper "relatively wet ($>$200 ppm H_2O) solid + incipient melt" part ($\approx$125 km thick; 1/3 in volume), the latter interacting with the overlying lower lithosphere. The existence of the asymmetric LVZ below the oceans indicates that below divergent plates, upwards and "eastward" transfer of lower asthenosphere is more active than elsewhere, accelerating its replacement by eastward lateral mantle flow coming from below the adjoining continents and related passive continental margin located to the "west" of a given ocean. So before flowing below and interacting with oceanic lithosphere, the upper asthenosphere underwent an earlier history of interaction with continental lithosphere, hence adding further chemical heterogeneity to intraplate mantle sources.

5. How to understand, within the framework of plate tectonics, the coexistence of mid-oceanic basalt and oceanic intraplate basalt mantle sources within the asthenosphere?

5.1. Location of mid-oceanic basalt mantle sources

Experimentally (Fig. 3), MORB are shown to be sourced from a roughly adiabatic (1490 to 1430 °C) upwelling of the uppermost part (250–230 km) of lower asthenosphere with melt fraction increasing dramatically above its anhydrous solidus (3 GPa/~100 km; 1450 °C). MORB melt segregation occurs at 15–20% melting between 2 and 1.6 GPa (65 to 45 km; Green and Falloon, 2005; Falloon et al., 2007b) within the so called "melting area" below the ridge axis in our model (Fig. 2).

The MORB mantle source composition is lherzolitic and fertile in terms of mineral assemblage and major element contents (Green and Falloon, 1998, 2005). However both trace element contents and Nd-Sr-Pb isotopic signatures are those of a mantle depleted in the most incompatible trace elements, *interpreted* as representative of a residual mantle composition. This decoupling between major and trace element composition interpretations is explained considering that the MORB source is itself a secondary lherzolite coming from the refertilization of a mantle residue by melts extracted from an already trace element depleted lherzolitic mantle (Frey and Green, 1974; Piccardo et al., 2007a, 2007b), or an eclogitic magmatic crust (Eiler et al., 2000), or recycled components (e.g. oceanic crust, Eiler et al., 2000; or metasomatized oceanic lithosphere, Niu et al., 2002; Shimizu et al., 2016), or a mix (Rosenthal et al., 2014). So our model assumes that the uppermost part of the lower asthenosphere is a secondary lherzolite, being the refertilization mostly provided from the recycling of W-directed subducting oceanic lithospheric slabs (Doglioni and Anderson, 2015; Green, 2015; Rosenthal et al., 2014).

As shown by our model, the unique and unidirectional path of transfer of a mantle with a MORB source signature, combined with the westward drift of the lithosphere and of the spreading ridge itself, highlights the systematic and permanent identical replacement of the MORB mantle source within the same P-T conditions of ascent and partial melting, which fits with the observed relative compositional homogeneity of MORB (dominant N-MORB) through geological times. This compositional homogeneity testifies itself for an absence of significant interactions of this oblique succession of mantle diapirs (Green, 1971) crosscutting the surrounding upper asthenosphere, and therefore for the relatively high velocity of such mantle diapirs driven by the suction effect of the spreading ridge. Diapir buoyancy driven by density, rheology and isolation from wall-rock reaction, is probably more a consequence of increasing temperature contrast if the diapir is near adiabatic i.e. garnet lherzolite with incipient melt into a rind of spinel lherzolite to plagioclase lherzolite (foliated to mylonitic).

5.2. Location of oceanic intraplate basalt mantle sources

As summarized by Green (2015), Green and Falloon (1998, 2005), Green and Falloon (2015), and Green et al. (2014), intraplate magmas, including oceanic intraplate basalts (or "hot spots" basalts), range from olivine melilitites and nephelinites, to olivine-rich basanites and finally to alkali basalts and olivine tholeiites. Experimentally, these basalts represent low degree (2 to 10%) melts from mantle sources located within the uppermost asthenosphere on 50 km thickness ($\approx$110–160 km depth; $\approx$3.3 to 5 GPa) or only 20 km thickness ($\approx$105–125 km depth; $\approx$3 to 4 GPa) depending on the considered geotherm ("intraplate" or "intraplate + shear heating effect"; Fig. 3). This upper asthenospheric source mantle is lherzolitic as the MORB mantle source and thus fertile in terms of mineral assemblage and major element contents, but above all rather enriched, and not depleted as the MORB source, in water (>0.02%) and in the most incompatible trace elements. As inferred in the previous chapter, they are secondary lherzolites in which refertilization of mantle residues can have two distinct origins. On one hand the lherzolites could derive from destabilization (≥90–100 km depth) of the lowermost part of the subsiding oceanic lithosphere far from the axial ridge. This result validates the hypothesis of an origin of some intraplate basalts from the lower lithospheric mantle in terms of composition (Anderson, 2010, and references therein). In that case, mantle metasomatism occurred below the spreading ridge by percolation/reaction of lithospheric mantle by MORB a rather long time ago. On the other hand, these secondary lherzolites could derive from metasomatism of lower asthenospheric material ascending throughout the upper asthenosphere along the main path with a very low angle trajectory (Fig. 2; see Section 4.2 for explanations). In that case, metasomatism results from successive percolation/reaction of incipient melts with mantle constantly on the move laterally. Nevertheless the Vs variations in the LVZ from one side to another of the ridge axis (cf. Section 4.2) suggests that the uppermost mantle fertility is somewhat reduced eastward.

Besides the Nd-Sr-Pb isotopic signatures of intraplate basalts attest for mantle sources also metasomatized at some point in their long history by continental lithospheric products (Pilet et al., 2005, 2008, 2011). Notice that most Pacific, but also Indian "hot spots" basalts (dominant alkali intraplate basalts) are on the western or southwestern plates relative to their respective axial ridges, thus are sourced within the upper asthenosphere having previously interacted with western adjoining and overlying continental plates. Such an interaction may be a candidate for explaining this continental imprint on intraplate basalt mantle sources, coherently with the relative "eastward" mantle flow.

6. Towards an alternative oceanic plate spreading model for a polarized plate tectonics

To date, a number of geophysical data at different scales demonstrate asymmetry of oceanic plates, whereas mantle petrologists still conceive mantle differentiation as a symmetric process on both sides of divergent plate boundaries and the underlying asthenosphere. To resolve this inconsistency, combination of updated geophysical, structural and petrological data on oceanic upper mantle (down to about 300 km) and also related concepts on mantle and magma genesis and plate kinematics, leads to the development of an oceanic spreading model where both the growing oceanic lithospheric mantle and the underlying upper asthenospheric mantle are to varying degrees asymmetric in composition. Our model takes place in the frame of the westward drift of the lithosphere, which means that oceanic ridges migrate laterally to the "west" relative to the asthenospheric mantle, and thus are always moving over a fertile mantle.

From the geophysical data, the asymmetry concerns bathymetry of the ridge flanks (less steep eastward), plate velocity rates (slower for the eastern one), thickness of plates (thinner eastward), and faster shear wave velocity in the western lithosphere. It comes that plates on each side of the ridge differ in density as well as rigidity, thus in composition, and thereby in growing conditions of mantle lithosphere at the axial ridge.

To understand from a petrological but also kinematic point of view the aforementioned asymmetry, mantle lithosphere and mantle asthenosphere were examined separately, even if they are interdependent.

From natural and experimental data, three processes, two just below and one outside the spreading ridge, are known to generate compositional heterogeneity of oceanic mantle lithosphere of both plates: 1) sub-ridge accretion of partial melting residues formed by melt segregation from mantle asthenosphere upwelling; mantle lithosphere thickness at the axial ridge is estimated, via experimental results on mantle phase equilibrium, between 20 to 40 km; 2) melt-rock reactions or mantle metasomatism by basaltic melt percolation through previously accreted residues below the ridge, inducing refertilization and secondary lherzolite formation; 3) capture and addition of asthenospheric mantle to the base of mantle lithosphere, because of cooling of the uppermost asthenosphere far from the ridge. This last process occurs when the geotherm drops below 1050 °C at 3 GPa which yields a LAB at ~90 km below which neither pargasite or hydrous carbonate-bearing silicate melt could exist. It gives a thickened lithosphere with the growing layer of subsolidus phlogopite-bearing garnet-lherzolite with carbonatite, or graphite + H_2O + CH_4 fluid.

Structural data released by mapping numerous mantle core complexes, mostly immediately westward of active slow-spreading ridges, give strong structural constraints for understanding lithospheric mantle dynamics during plate spreading. Indeed whether in the west or east, these core complexes represent outcrops of mantle and related intrusive gabbro bodies exhumed along the footwall of convex up-wards detachment faults dipping eastward and rooted down to 7–8 km. We interpret these detachment faults as effects of top to bottom asymmetric lithospheric mantle shear, linked to the eastward flowing of asthenosphere delaying the base of the lithosphere moving westward. The usual sill shape of gabbro intrusive bodies is another observation supporting intra-mantle decoupling during ocean spreading. Furthermore, at the topmost part of the hangingwall of these large detachment faults, a small-scale decoupling occurs between the base of volcanoes and underlying mantle or/and gabbro bodies in process of exhumation, underlining this time a pseudo-symmetric spreading. It follows that whatever the scale, the top to bottom decoupling dynamics enables the exhumation of deep rocks (mantle with or without intrusive gabbro), thus the creation of new surfaces, giving all its meaning to the concept of "detachment-mode of seafloor spreading" developed by Maffione et al. (2013). *This top to bottom mantle decoupling preserves totally magnetic anomalies recorded at the surface on each side of the axis.*

Considering all these results, the mantle lithosphere composition can evolve differently on each side of the ridge, coming from a permanent refertilization of the western plate whereas the eastern plate just preserves the barely refertilized feature of a harzburgitic residue generated during asthenosphere partial melting below axial ridge. So rigidity, density to a lesser extent and thickness of the western plate become much higher, whereas the shallower bathymetry of the east flank of the ridge is explained by isostatic adjustment because of a somewhat lower density. Also as the eastern plate mantle is lighter and more viscous, there is a higher coupling at the lithosphere-asthenosphere boundary, determining a slower velocity of the eastern plate during ridge migration. The top to bottom and west to east internal decoupling of the mantle lithosphere, linked to its westward drift above the asthenospheric mantle flowing eastward, explains detachment faults and related mantle core complex exhumation.

The petrology of asthenospheric mantle is known experimentally since asthenosphere is the mantle reservoir, at distinct P-T-depths, of MORB (lower asthenosphere) and intraplate basalts (middle to upper asthenosphere). An asymmetric "incipient melt + solid" mantle zone, the LVZ, exists between 230 and 100 km, being much more developed and more fertile westward. So MORB are sourced from a lherzolitic mantle in equilibrium at ≈250–230 km/7 GPa, the ascent of which up to 65–45 km/2–1.6 GPa just below the ridge is roughly adiabatic, thus rather fast and without any significant interaction with surrounding crosscut upper asthenosphere. According to the relative "eastward" direction of the asthenospheric mantle flow lines, the diapiric rise of MORB mantle source from lower asthenosphere is significantly oblique and eastward. This mantle transfer is triggered and checked by the suction effect of the spreading ridge, and the westward ridge migration determines a self-perpetuating mechanism for permanently renewing the MORB source mantle, which explains in return the great homogeneity of MORB through time. Synchronously, major mantle transfers within middle to upper asthenosphere come from the lower asthenosphere farther westward and follow the eastward mantle flow lines, slightly slantwise but barely deviated by the suction effect of the spreading ridge. Mantle undergoes incipient melting, except that the interstitial melts percolate laterally and upwards, react with and enrich the upper asthenosphere in water and the most incompatible elements, a possible mantle reservoir for intraplate basalts. This incipient melting zone represents also the lithosphere-asthenosphere decoupling zone affected by shear heating, reducing the P/depth of intraplate magma genesis. Also the higher shear wave velocity eastward means a density decrease in relation with a somewhat residual feature of the eastern asthenospheric mantle and reflecting a mixing with residual mantle coming from the high degree melting area below the ridge and not accreted to the lithosphere.

7. Conclusions

The proposed oceanic plate spreading model shows that numerous asymmetric features observed with geophysics and structural analysis in the first 300 km below the oceans are consistent with petrological data on upper mantle and basaltic compositions. Our multi-disciplinary study emphasizes the major role of physical and chemical dynamics of the upper mantle for understanding oceanic plate tectonics, and in turn how evolution of both the composition and internal structure of oceanic plates strongly depends at all scales on plate kinematics. Our model suggests that both plate composition and kinematics are interdependent on asthenosphere compositional evolution. The spreading of an ocean, synchronous with the migration of the mid-oceanic ridge, induces a complete and permanent material renewal, by mass transfer at all scales, particularly in the shallow 300 km of the Earth's mantle below the whole width of the oceans. Another major concept is that all petrological processes, occurring during different types of solid or/and liquid mantle mass transfers, ultimately lead to the genesis of a more or less fertile mantle composition, i.e. a lherzolite but secondary in nature (residual harzburgite refertilized by melt interaction), which is the main composition of the first 300 km of the Earth below oceans and most likely the continents as well. Incompatible trace elements abundances and isotopic values of mantle sources will necessarily vary from one specific mantle site to another reflecting the detailed individual geological histories of 'lherzolites'.

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References

Abily, B., Ceuleneer, G., 2013. The dunitic mantle–crust transition zone in the Oman ophiolite: residue of melt-rock interaction, cumulates from high-MgO melts, or both? Geology 41:67–70. http://dx.doi.org/10.1130/G33351.1.

Afonso, J.C., Fernandez, M., Ranalli, G., Griffin, W.L., Connolly, J.A.D., 2008. Integrated geophysical-petrological modeling of the lithosphere and sublithospheric upper mantle: methodology and applications. Geochemistry, Geophysics, Geosystems 9, Q05008. http://dx.doi.org/10.1029/2007GC001834.

Anderson, D.L., 2007. New Theory of the Earth. Cambridge University Press.

Anderson, D.L., 2010. Hawaii, boundary layers and ambient mantle-geophysical constraints. Journal of Petrology http://dx.doi.org/10.1093/petrology/egq068.

Bonatti, E., 1990. Not So Hot "Hot Spots" in the Oceanic Mantle. Science 250, 107–111.

Bonatti, E., Ligi, M., Brunelli, D., Cipriani, A., Fabretti, P., Ferrante, V., Ottolini, L., 2003. Mantle thermal pulses below the Mid Atlantic Ridge and temporal variations in the oceanic lithosphere. Nature 423, 499–505.

Bostrom, R.C., 1971. Westward displacement of the lithosphere. Nature 234, 536–538.

Brune, S., Heine, C., Pérez-Gussinye, M., Sobolev, S.V., 2014. Rift migration explains continental margin asymmetry and crustal hyper-extension. Nature Communications 5: 4014. http://dx.doi.org/10.1038/ncomms5014.

Canales, J.P., Carton, H., Carbotte, S.M., Mutter, J.C., Nedimovi'c, M.R., Xu, M., Aghaei, O., Marjanovi'c, M., Newman, K., 2012. Network of off-axis melt bodies at the East Pacific Rise. Nature Geoscience 5, 279–283.

Cannat, M., 1993. Emplacement of mantle rocks in the seafloor at mid-ocean ridges. Journal of Geophysical Research, Solid Earth 98 (B3), 4163–4172.

Carbotte, S.M., Small, C., Donnelly, K., 2004. The influence of ridge migration on the magmatic segmentation of mid-ocean ridges. Nature 429, 743–746.

Carey, S.W., 1958. Continental Drift—A Symposium. Univ. of Tasmania, Hobart, pp. 177–363 (March 1956).

Chalot-Prat, F., 2005. An undeformed ophiolite in the Alps: field and geochemical evidences for a link between volcanism and shallow plate tectonic processes. In: Foulger, G.R., Anderson, D.L., Natland, J.H., Presnall, D.C. (Eds.), Plates, Plumes & Paradigms. Geological Society of America, Special Paper 388, pp. 751–780.

Chalot-Prat, F., Falloon, T.J., Green, D.H., Hibberson, W.O., 2010. An experimental study of liquid compositions in equilibrium with plagioclase + spinel lherzolite at low pressures (0.75 GPa). Journal of Petrology 51 (11), 2349–2376.

Chalot-Prat, F., Falloon, T.J., Green, D.H., Hibberson, W.O., 2013. Melting of plagioclase + spinel lherzolite at low pressures (0.5 GPa): an experimental approach to the evolution of basaltic melt during mantle refertilisation at shallow depths. Lithos 172-173, 61–80.

Conder, J.A., Wiens, D.A., 2006. Seismic structure beneath the Tonga arc and Lau backarc basin determined from joint Vp, Vp/Vs tomography. Geochemistry, Geophysics, Geosystems 7, Q03018.

Crespi, M., Cuffaro, M., Doglioni, Giannone, C., Riguzzi, F., 2007. Space geodesy validation of the global lithospheric flow. Geophysical Journal International 168:491–506. http://dx.doi.org/10.1111/j.1365-246X.2006.03226.x.

Cuffaro, M., Doglioni, C., 2007. Global kinematics in deep versus shallow hotspot reference frames. In: Foulger, G.R., Jurdy, D.M. (Eds.), Plates, Plumes, and Planetary Processes. Geological Society of America Special Paper 430:pp. 359–374. http://dx.doi.org/10.1130/2007.2430(18).

Dick, H.J.B., Natland, J.H., 1996. Late-stage melt evolution and transport in the shallow mantle beneath the East Pacific Rise. In: Mével, C., et al. (Eds.), Proceedings of the Ocean Drilling Program, Scientific Results 147. Ocean Drilling Program, College Station, Texas, pp. 103–134.

Dijkstra, A.H., Drury, M.R., Visser, L.R.M., 2001. Structural petrology of plagioclase peridotite in the West Othris mountains (Grece): melt impregnation in mantle lithosphere. Journal of Petrology 42, 5–24.

Doglioni, C., Anderson, D.L., 2015. Top driven asymmetric mantle convection. The Interdisciplinary Earth: A Volume in Honor of Don L. Anderson. Geological Society of America Special Paper 514, American Geophysical Union Special Publication 71 http://dx.doi.org/10.1130/2015.2514(05) .

Doglioni, C., Panza, G.F., 2015. Polarized plate tectonics. Advances in Geophysics 56 (3): 1–167. http://dx.doi.org/10.1016/bs.agph.2014.12.001.

Doglioni, C., Carminati, E., Bonatti, E., 2003. Rift asymmetry and continental uplift. Tectonics 22:1024–1037. http://dx.doi.org/10.1029/2002TC001459.

Doglioni, C., Green, D.H., Mongelli, F., 2005. On the shallow origin of hotspots and the westward drift of the lithosphere. In: Foulger, G.R., Natland, J.H., Presnall, D.C., Anderson, D.L. (Eds.), Plates, Plumes, and Paradigms. Geological Society of America Special Paper 388:pp. 735–749. http://dx.doi.org/10.1130/2005.2388(42).

Doglioni, C., Carminati, E., Cuffaro, M., Scrocca, D., 2007. Subduction kinematics and dynamic constraints. Earth Science Reviews 83:125–175. http://dx.doi.org/10.1016/j.earscirev.2007.04.001.

Doglioni, C., Ismail-Zadeh, A., Panza, G., Riguzzi, F., 2011. Lithosphere-asthenosphere viscosity contrast and decoupling. Physics of the Earth and Planetary Interiors 189, 1–8.

Eggins, S.M., 1992a. Petrogenesis of Hawaiian tholeiites: 1. Phase equilibria constraints. Contributions to Mineralogy and Petrology 110, 387–397.

Eggins, S.M., 1992b. Petrogenesis of Hawaiian tholeiites: 2. Aspects of dynamic melt segregation. Contributions to Mineralogy and Petrology 110, 398–410.

Eiler, J.M., Schiano, P., Kitchen, N., Stolper, E.M., 2000. Oxygen-isotope evidence for recycled crust in the sources of mid-ocean-ridge basalts. nature 403, 530–534.

Falloon, T.J., Danyushevsky, L.V., Ariskin, A., Green, D.H., 2007a. The application of olivine geothermometry to infer crystallization temperatures of parental liquids: implications for the temperature of MORB magmas. Chemical Geology 241, 207–233.

Falloon, T.J., Green, D.H., Danyushevsky, L.V., 2007b. Crystallization temperatures of tholeiite parental liquids: implications for the existence of thermally driven mantle plumes. In: Foulger, G.R., Jurdy, D.M. (Eds.), Plates, Plumes, and Planetary Processes. Geological Society of America Special Paper 430:pp. 235–260. http://dx.doi.org/10.1130/2007.2430(12).

Frey, F.A., Green, D.H., 1974. The mineralogy, geochemistry and origin of lherzolite inclusions in Victorian basanites. Geochimica et Cosmochimica Acta 38, 1023–1059.

Fyfe, W.S., 1960. The possibility of *d*-electron coupling in olivine at high pressures. Geochimica et Cosmochimica Acta 19, 141–143.
Green, D.H., 1971. Composition of basaltic magmas as indicators of conditions of origin: application to oceanic volcanism. Philosophical Transactions of the Royal Society of London, A 268, 707–725.
Green, D.H., 2015. Experimental petrology of peridotites, including effects of water and carbon on melting in the Earth's uppermantle. Physics and Chemistry of Minerals 42:95–102. http://dx.doi.org/10.1007/s00269-014-0729-2.
Green, D.H., Falloon, T.J., 1998. Pyrolite: a ringwood concept and its current expression. In: Jackson, I.N.S. (Ed.), The Earth's Mantle: Composition, Structure and Evolution. Cambridge University Press, Cambridge, pp. 311–380.
Green, D.H., Falloon, T.J., 2005. Primary magmas at mid-ocean ridges, 'hotspots', and other intraplate settings: constraints on mantle potential temperature. In: Foulger, G.R., Natland, J.H., Presnall, D.C., Anderson, D.L. (Eds.), Plates, Plumes, and Paradigms. Geological Society of America, Special Papers Vol. 388, pp. 217–247.
Green, D.H., Falloon, T.J., 2015. Mantle-derived magmas: intraplate, hot-spots and mid-ocean ridges. Scientific Bulletin. http://dx.doi.org/10.1007/s11434-015-0920-y.
Green, D.H., Falloon, T.J., Eggins, S.M., Yaxley, G.M., 2001. Primary magmas and mantle temperatures. European Journal of Mineralogy 13 (3), 437–451.
Green, D.H., Hibberson, W.O., Kovacs, I., Rosenthal, A., 2010. Water and its influence on the lithosphere – asthenosphere boundary. Nature 467:448–451. http://dx.doi.org/10.1038/nature09369.
Green, D.H., Hibberson, W.O., Kovacs, I., Rosenthal, A., 2011. Water and its influence on the lithosphere-asthenosphere boundary, 467, 448–451, 2010. ([Addendum.]). Nature 472 (7344), 504.
Green, D.H., Hibberson, W.O., Rosenthal, A., Kovacs, I., Yaxley, G.M., Falloon, T.J., Brink, F., 2014. Experimental study of the influence of water on melting and phase assemblages in the upper mantle. Journal of Petrology 55 (10):2067–2096. http://dx.doi.org/10.1093/petrology/egu050.
Gripp, A.E., Gordon, R.G., 2002. Young tracks of hotspots and current plate velocities. Geophysical Journal International 150:321–364. http://dx.doi.org/10.1046/j.1365-246X.2002.01627.x.
Hammond, W.C., Toomey, D.R., 2003. Seismic velocity anisotropy and heterogeneity beneath the Mantle Electromagnetic and Tomography Experiment (MELT) region of the East Pacific Rise from analysis of P and S body waves. Journal of Geophysical Research 108 (B4):2176. http://dx.doi.org/10.1029/2002JB001789.
Henstock, T.J., Woods, A.W., White, R.S., 1993. The accretion of oceanic crust by episodic sill intrusion. Journal of Geophysical Research, Solid Earth 98, 4143–4161.
Holmes, A., 1944. Principles of Physical Geology. Thomas Nelson and Sons LTD, Edinburgh.
Jordan, T.H., 1979. Mineralogies, densities and seismic velocities of garnet lherzolites and their geophysical implications. In: Boyd, F.R., Meyer, H.O.A. (Eds.), The Mantle Sample: Inclusion in Kimberlites and Other Volcanics. American Geophysical Union, Washington D.C.:pp. 1–13. http://dx.doi.org/10.1029/SP016p0001.
Kaczmarek, M.A., Müntener, O., 2008. Juxtaposition of melt impregnation and high temperature shear zones in the upper mantle; field and petrological constraints from the Lanzo peridotite (Northern Italy). Journal of Petrology 49, 2187–2220.
Kaczmarek, M.A., Müntener, O., 2010. The variability of peridotite composition across a mantle shear zone (Lanzo massif, Italy): interplay of melt focusing and deformation. Contribution to Mineralogy and Petrology 160, 663–679.
Lavier, L.L., Manatschal, G., 2006. A mechanism to thin the continental lithosphere at magma-poor margins. Nature 440, 324–328.
Le Pichon, X., 1968. Sea-floor spreading and continental drift. Journal of Geophysical Research 73 (12), 3661–3697.
Ligi, M., Bonatti, E., Cipriani, A., Ottolini, L., 2005. Water-rich basalts at mid-ocean-ridge cold spots. Nature 434 (7029), 66–69.
MacLeod, C.J., Searle, R.C., Murton, B.J., Casey, J.F., Mallows, C., Unsworth, S.C., Achenbach, K.L., Harris, M., 2009. Life cycle of oceanic core complexes. Earth and Planetary Science Letters 287, 333–344.
Maffione, M., Morris, A., Anderson, M.W., 2013. Recognizing detachment-mode seafloor spreading in the deep geological past. Scientific Reports 3:2336. http://dx.doi.org/10.1038/srep02336.
Mallows, C., Searle, R.C., 2012. A geophysical study of oceanic core complexes and surrounding terrain, Mid-Atlantic Ridge 13_N–14_N. Geochemistry, Geophysics, Geosystems 13:Q0AG08. http://dx.doi.org/10.1029/2012GC004075.
Moore, G.W., 1973. Westward tidal lag as the driving force of plate tectonics. Geology 1, 99–100.
Müller, R.D., Roest, W.R., Royer, J.Y., Gahagan, L.M., Sclater, J.G., 1997. Digital isochrons of the world's ocean floor. Journal of Geophysical Research 102, 3211–3214.
Müller, R.D., Sdrolias, M., Gaina, C., Roest, W.R., 2008. Age, spreading rates, and spreading asymmetry of the world's ocean crust. Geochemistry, Geophysics, Geosystems 9, Q04006. http://dx.doi.org/10.1029/2007GC001743.
Müntener, O., Pettke, T., Desmurs, L., Meier, M., Schaltegger, U., 2004. Refertilization of mantle peridotite in embryonic ocean basins: trace element and Nd-isotope evidence and implications for crust–mantle relationships. Earth and Planetary Science Letters 221, 293–308.
Muntener, O., Manatschal, G., Desmurs, L., Pettke, T., 2010. Plagioclase peridotites in ocean-continent transitions: refertilized mantle domains generated by melt stagnation in the shallow mantle lithosphere. Journal of Petrology 51, 255–294.
Niu, Y.L., Regelous, M., Wendt, I.J., Batiza, R., O'Hara, M.J., 2002. Geochemistry of near-EPR seamounts: importance of source vs. process and the origin of enriched mantle component. Earth and Planetary Science Letters 199, 327–345.
Panza, G.F., Romanelli, F., 2014. Seismic waves in 3-D: from mantle asymmetries to reliable seismic hazard assessment. Earthquake Science 27 (5):567–576. http://dx.doi.org/10.1007/s11589-014-0091-y.
Panza, G.F., Doglioni, C., Levshin, A., 2010. Asymmetric ocean basins. Geology 38 (1): 59–62. http://dx.doi.org/10.1130/G30570.1.
Piccardo, G.B., Guarnieri, L., 2010. The Monte Maggiore peridotite (Corsica, France): a case study of mantle evolution in the Ligurian Tethys. Geological Society, London, Special Publications 337:7–45. http://dx.doi.org/10.1144/SP337.2.
Piccardo, G.B., Müntener, O., Zanetti, A., Pettke, T., 2004. Ophiolitic peridotites of the Alpine-Apennine system: mantle processes and geodynamic relevance. International Geology Review 46, 1119–1159.
Piccardo, G.B., Poggi, E., Vissers, R.L.M., 2007a. The pre-oceanic evolution of the Erro–Tobbio peridotite (Voltri Massif, Ligurian Alps, Italy). Journal of Geodynamics 43, 417–449.
Piccardo, G.B., Zanetti, A., Müntener, O., 2007b. Melt/peridotite interaction in the Southern Lanzo peridotite: field, textural and geochemical evidence. Lithos 94, 181–209.
Piccardo, G.B., Padovano, M., Guarnieri, L., 2014. The ligurian tethys: mantle processes and geodynamics. Earth-Science Reviews 138:409–434. http://dx.doi.org/10.1016/j.earscirev.2014.07.002.
Pilet, S., Hernandez, J., Sylvester, P., Poujol, M., 2005. The metasomatic alternative for ocean island basalt chemical heterogeneity. Earth and Planetary Science Letters 236, 148–166.
Pilet, S., Baker, M.B., Stolper, E.M., 2008. Metasomatized lithosphere and the origin of alkaline lavas. Science 320, 916–919.
Pilet, S., Baker, M.B., Muntener, O., Stolper, E.M., 2011. Monte Carlo simulations of metasomatic enrichment in the lithosphere and implications for the source of alkaline basalts. Journal of Petrology 52, 1415–1442.
Rampone, E., Borghini, G., 2008. The melt intrusion/interaction history of the Erro–Tobbio peridotites (Ligurian Alps, Italy): insights on mantle processes at nonvolcanic passive margins. European Journal of Mineralogy 20, 573–585.
Rampone, E., Hofmann, A.W., 2012. A global overview of isotopic heterogeneities in the oceanic mantle. Lithos 148, 247–261.
Rampone, E., Piccardo, G.B., Vannucci, R., Bottazzi, P., 1997. Chemistry and origin of trapped melts in ophiolitic peridotites. Geochimica et Cosmochimica Acta 61, 4557–4569.
Rampone, E., Romairone, A., Abouchami, W., Piccardo, G.B., Hofmann, W., 2005. Chronology, petrology and isotope geochemistry of the Erro–Tobbio peridotites (Ligurian Alps, Italy): records of late Palaeozoic lithospheric extension. Journal of Petrology 46, 799–827.
Rampone, E., Piccardo, G.B., Hofmann, A.W., 2008. Multi-stage melt-rock interaction in the Mt. Maggiore (Corsica, France) ophiolitic peridotites: microstructural and geochemical records. Contribution to Mineralogy and Petrology 156, 453–475.
Reston, T.J., Ranero, C.R., 2011. The 3-D geometry of detachment faulting at mid-ocean ridges. Geochemistry, Geophysics, Geosystems 12:Q0AG05. http://dx.doi.org/10.1029/2011GC003666.
Ricard, Y., Doglioni, C., Sabadini, R., 1991. Differential rotation between lithosphere and mantle: a consequence of lateral mantle viscosity variations. Journal of Geophysical Research 96 (B5):8407–8415. http://dx.doi.org/10.1029/91JB00204.
Riguzzi, F., Panza, G., Varga, P., Doglioni, C., 2010. Can Earth's rotation and tidal despinning drive plate tectonics? Tectonophysics 484:60–73. http://dx.doi.org/10.1016/j.tecto.2009.06.012.
Rosenthal, A., Yaxley, G.M., Green, D.H., Hermann, J., Kovacs, I., Spandler, C., 2014. Continuous eclogite melting and variable refertilisation in upwelling heterogeneous mantle. Scientific Reports 4:6099. http://dx.doi.org/10.1038/srep06099.
Scheirer, D.S., Forsyth, D.W., Cormier, M.H., Macdonald, K.C., 1998. Shipboard geophysical indications of asymmetry and melt production beneath the East Pacific Rise near the MELT experiment. Science 280:1221–1224. http://dx.doi.org/10.1126/science.280.5367.1221.
Schmerr, N., 2012. The Gutenberg discontinuity: melt at the lithosphere-asthenosphere boundary. Science 335, 14801483.
Scoppola, B., Boccaletti, D., Bevis, M., Carminati, E., Doglioni, C., 2006. The westward drift of the lithosphere: a rotational drag? Geological Society of America Bulletin 118: 199–209. http://dx.doi.org/10.1029/2004TC001634.
Seyler, M., Bonatti, E., 1997. Regional-scale melt-rock interaction in lherzolitic mantle in the Romanche fracture zone (Atlantic Ocean). Earth and Planetary Science Letters 146, 273–287.
Shen, Y., Forsyth, D.W., Scheirer, D.S., Macdonald, K.C., 1993. Two forms of volcanism: implications for mantle flow and off-axis crustal production on the west flank of the southern east pacific rise. Journal of Geophysical Research 98 (B10), 17,875–17,889.
Shimizu, K., Saal, A.E., Myers, C.E., Nagle, A.N., Hauri, E.H., Forsyth, D.W., Kamenetsky, V.S., Niu, Y., 2016. Two-component mantle melting-mixing model for the generation of mid-ocean ridge basalts: implications for the volatile content of the Pacific upper mantle. Geochimica et Cosmochimica Acta 176, 44–80.
Small, C., Danyushevsky, L., 2003. A plate kinematic explanation for mid-ocean ridge depth discontinuities. Geology 31, 399–402.
Smith, D.K., Escartín, J., Schouten, H., Cann, J.R., 2008. Fault rotation and core complex formation: significant processes in seafloor formation at slow-spreading mid-ocean ridges (Mid-Atlantic Ridge, 13_–15_N). Geochemistry, Geophysics, Geosystems 9 (3). http://dx.doi.org/10.1029/2007GC001699.
Smith, D.K., Escartín, J., Schouten, H., Cann, J.R., 2012. Active long-lived faults emerging along slow-spreading mid-ocean ridges. Oceanography 25 (1):94–99. http://dx.doi.org/10.5670/oceanog.2012.07.
Takahashi, N., 2001. Origin of plagioclase lherzolite from the Nikanbetsu peridotite complex, Hokkaido, northern Japan: implications for incipient melt migration and segregation in the partially molten upper mantle. Journal of Petrology 42, 39–54.
Tartarotti, P., Susini, S., Nimis, P., Ottolini, L., 2002. Melt migration in the upper mantle along the Romanche fracture zone (equatorial Atlantic). Lithos 63, 125–149.
Thybo, H., 2006. The heterogeneous upper mantle low velocity zone. Tectonophysics 416, 53–79.
Toomey, D.R., 2012. Piecing together rifts. Nature Geoscience 5, 235–236.
Toomey, D.R., Wilcock, W.S.D., Conder, J.A., Forsyth, D.W., Blundy, J.D., Parmentier, E.M., Hammond, W.C., 2002. Asymmetric mantle dynamics in the MELT region of the East Pacific Rise. Earth and Planetary Science Letters 200, 287–295.

Vine, F.J., Matthews, D.H., 1963. Magnetic anomalies over oceanic ridges. Nature 199 (4897):947–949. http://dx.doi.org/10.1038/199947a0.
Warren, J.M., 2016. Global variations in abyssal peridotite compositions. Lithos 248-251, 193–219.
Warren, J.M., Shimizu, N., 2010. Cryptic variations in abyssal peridotite compositions: evidence for shallow-level melt infiltration in the oceanic lithosphere. Journal of Petrology 51, 395–423.
Wegener, A., 1915. Die Entstehung der Kontinente und der Ozeae: Samml. Vieweg, Braunschweig 23, 1–94.
White, S.C., Macdonald, K.C., Scheirer, D.S., Cormier, M.H., 1998. Distribution of isolated volcanoes on the flanks of the East Pacific Rise, 15.3°S-20°S. Journal of Geophysical Research 103 (B12), 30,371–30,384.
Whitney, D.L., Teyssier, C., Rey, P., Buck, W.R., 2013. Continental and oceanic core complexes. GSA Bulletin 125 (3/4):273–298. http://dx.doi.org/10.1130/B30754.1.