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Westward migration of oceanic ridges and related asymmetric upper mantle differentiation

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Highlights:

1. Westward ridge migration induces higher mantle refertilisation of the western plate ...88
2. Upwelling residual mantle is abandoned eastward consistent with seismological asymmetry ...91
3. Ridge migration provides permanently MORB mantle source from the west ...73

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 refertilisation) 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 refertilisation 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 parasitic amphibole producing a sharp depression of the water-undersaturated solidus, 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 water-undersaturated 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

44 from the lower asthenosphere through the upper asthenosphere are oblique, upward and eastward.
45 MORB are sourced from an eastward and oblique, near-adiabatic mantle upwelling from the lower
46 asthenosphere. This unidirectional mantle transfer is induced by isostatic suction of the migrating
47 spreading ridge.
48

49 **Key words**

50 Asymmetric oceanic lithosphere, Westward drift of the lithosphere, Migrating ridge, Mantle melting,
51 Percolation-reaction and refertilization, Detachment fault and mantle core complex, Secondary
52 lherzolite and upper mantle differentiation

53 **1. Introduction**

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

69 Since the 1960s, plate tectonics is conceived as driven either by active mantle upwelling ('bottom
70 up' or alternatively 'ridge push') or alternatively by the negative buoyancy of slabs ('top down' or
71 alternatively 'slab pull'). It is clear that creation of oceanic lithosphere and its subsequent subduction
72 are fundamental mechanisms in governing the cooling of the Earth. In other words, the surface
73 dissipation of both the Earth's internal primordial and radiogenic heat is tied with the re-entrance of
74 the cold lithosphere along subduction zones to determine the cooling of the planet. However, the
75 mechanisms determining mantle drag and mantle convection remain a matter of debate. The
76 asymmetry of subduction zones (those directed to the west are steeper and faster, e.g., Doglioni et al.,
77 2007; Riguzzi et al., 2010) and plate motion reconstructions relative to the mantle reference frame
78 support a westerly-directed drift of the lithosphere relative to the underlying mantle, also called net
79 rotation (as a mean value) or westward drift (Wegener, 1915; Holmes, 1944; Carey, 1958; Le Pichon,
80 1968; Bostrom, 1971; Moore, 1973; Ricard et al., 1991; Gripp and Gordon, 2002; Crespi et al., 2007;
81 Cuffaro and Doglioni, 2007). The net rotation or the westward drift of the lithosphere requires both a
82 decoupling at the lithosphere-asthenosphere interface and a mechanism driving this rotation (Doglioni
83 et al., 2011; Doglioni and Anderson, 2015; Doglioni and Panza, 2015). The motion is not east-west,
84 but along an undulated flow that varies from WNW to E-W to ENE trends, along the so-called
85 'tectonic equator' (Crespi et al., 2007). Proposed mechanisms involve either the negative buoyancy of
86 the lithosphere (Ricard et al., 1991), or the astronomical drag induced by the Earth's rotation

87 combined with tidal friction (Riguzzi et al., 2010). Potentially astronomical forces may interact with
 88 mantle convection resulting from the cooling of the Earth (Holmes, 1944; Carey, 1958; Scoppola et
 89 al., 2006).

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

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

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

104 Since the recognition of magnetic anomalies on both sides of oceanic ridges (Vine and Matthews,
 105 1963), oceanic rift zones were considered as having symmetric features, and plate spreading as a
 106 symmetric geological process. As shown by Doglioni et al. (2003), observations on the bathymetry of
 107 ridge flanks (shallower eastward) of Pacific and Atlantic oceans and on variable rates of oceanic
 108 spreading (Muller et al., 1997; 2008; Mallows and Searle, 2012) contradict the concept of full
 109 symmetry. Panza et al. (2010) and Panza and Romanelli (2014) provided evidence for asymmetry in
 110 both shear wave velocity and thickness within mantle lithosphere and asthenosphere when comparing
 111 the two sides of most spreading ridges. In general, the old (>60 Ma) western lithospheric plates have a
 112 faster shear wave velocity and are thicker (≈ 100 km versus ≈ 80 km in the eastern flank). Seismic
 113 tomography (Thybo, 2006; Panza et al., 2010; Schmerr, 2012; Panza and Romanelli, 2014) provides
 114 robust evidence of an asymmetric LVZ at depths from about 80-100 km down to 180-225 km. The
 115 average V_s within the western LVZ is slower than the eastern one. The LVZ shape is itself
 116 asymmetric, being thicker and wider on the western side of ridges. Also passive continental margins
 117 and the related continent-ocean transition support a systematic asymmetry of rift zones (Lavier and
 118 Manatschal, 2006; Brune et al., 2014).

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

124 Panza et al. (2010) explain these observations as a result of the depletion of the upper
 125 asthenosphere, that evolves from pre-melting lherzolite into post-melting harzburgite composition
 126 below the oceanic ridge, while the ridge is moving westward relative to the asthenospheric mantle,
 127 i.e., in the hot-spot reference frame (Crespi et al., 2007). The role of ridge migration (Scheirer et al.,
 128 1998; Small and Danyushevsky, 2003; Doglioni and Panza, 2015) in producing an oceanic plate
 129 asymmetry is a key parameter of the model. To the east of the ridge, the lithospheric mantle (named
 130 also LID) would represent the residual product from partial melting of upwelling asthenosphere,

131 abandoned and in the process of cooling after the ridge migration to the “west”. Since ridges are
 132 shifting above a fertile asthenospheric mantle, there is a continuous supply of MORB source mantle
 133 beneath the migrating spreading ridge (Panza et al., 2010). This model, deduced from geophysical and
 134 kinematic data (Fig. 1), suggests that the westward drift of the lithosphere relative to the underlying
 135 asthenosphere is a global phenomenon, a concept that implies decoupling between lithosphere and
 136 asthenosphere occurring within the LVZ, i.e., in the upper asthenosphere.

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

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

150 **3. Deciphering the compositional and physical evolution of the oceanic lithospheric mantle on** 151 **both sides of a spreading ocean ridge**

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

159 *3.1. General physical features and composition of oceanic lithosphere*

160 The oceanic lithosphere is deeply fractured at the ridge axis. This major fracture network reaches
 161 the top of the ductile layer in the lithosphere, and two plates, A and B, form and spread apart. The
 162 western plate A always moves westerly faster than the eastern plate B (Fig. 2) as shown by Crespi et
 163 al. (2007). In our model, due to the westward drift of the lithosphere (Le Pichon, 1968; Doglioni et al.,
 164 2003), both plates move to the west, the western plate A at V_A and the eastern plate B at V_B , being
 165 $V_A > V_B$ (Fig. 2). The oceanic ridge moves west at $V_r = (V_A + V_B)/2$. Plate growth rate is due to the
 166 difference between V_A and V_B . Thus the separation between points on each plate increases at the half
 167 spreading rate (hS) which is given by $(V_A - V_B)/2$. V_r is always faster than hS .

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

171 Geophysical data show lithospheric thickness decreasing towards ridge axes above shallowing
 172 asthenosphere. However we do not think that the lithosphere-asthenosphere boundary (LAB) reaches
 173 the base of the newly formed basaltic crust at fast spreading ridges, nor approaches the sea floor at

174 magma-starved slow-spreading ridges. This view is based on the petrology (phase assemblages),
175 chemical compositions and isotopic compositions of peridotites sampled from rifted margins,
176 ridge/transform intersections and slow-spreading ridges. Only lithospheric (secondary harzburgite or
177 secondary lherzolite; see Piccardo et al., 2007b), but not asthenospheric mantle compositions have
178 been identified either at the bottom of rift axis in slow-spreading active and fossil oceans (e.g., Seyler
179 and Bonatti, 1997; Dijkstra et al., 2001; Takahashi, 2001; Tartarotti et al., 2002; Müntener et al.,
180 2004; Piccardo et al., 2004, 2007a b, 2014; Kaczmarek and Müntener, 2008, 2010; Rampone et al.,
181 1997, 2005, 2008; Rampone and Borghini, 2008; Warren and Shimizu, 2010; Müntener et al., 2010;
182 Piccardo and Garnieri, 2010; Rampone and Hofmann, 2012; Warren, 2016), or below oceanic crust in
183 fast-spreading active and fossil oceans (Dick and Natland, 1996; Abily and Ceuleneer, 2013). The
184 depth of the LAB under the ridge may be inferred in two ways: 1. estimating the depth of melt
185 segregation for parental MOR picrites or basalts and 2. by the P, T conditions recorded in ridge
186 peridotites (including their P, T decompression paths in some cases). Indeed experimental results (Fig.
187 3) demonstrate that parental MOR picrite segregation from residual peridotite occurs mostly between
188 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
189 to 30 km), around 1400°C (up to 1430°C) from a trace element depleted lherzolite (Eggins, 1992a,b;
190 Falloon et al., 2007a,b; Green et al., 2001; Green and Falloon, 2005; Green et al., 2014; Green and
191 Falloon, 2015). Magma genesis produces mantle residual peridotite less dense than the asthenospheric
192 protolith, accreting above its asthenospheric source and potentially subject to percolation-reaction of
193 new magma ascents, giving birth to a refertilized constantly growing oceanic lithospheric mantle.
194 From experimental results on liquid compositions in equilibrium with spinel and/or plagioclase
195 lherzolite assemblages (Falloon et al., 2007a, b; Chalot-Prat et al., 2010; 2013), analogous to natural
196 oceanic mantle samples, the oceanic lithospheric mantle thickness at the ridge is at least 20 km and up
197 to 40 km in thickness. Note that density differences are due to chemical/mineralogical differences but
198 also to temperature differences (e.g., residual mantle temperature has been reduced by the lost of the
199 latent heat of melting). Thus determining density differences is subject to significant uncertainties.

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

213 As far as it concerns the depth/temperature of the base of the oceanic lithosphere outside the ridge
214 (>500 km), the model of Afonso et al. (2007) predicts 1300°C at 105 km depth. This temperature is
215 somewhat too high according to experimental data of Green et al. (2010, 2014), showing that a “wet”
216 (>200 ppm H₂O) oceanic lithospheric mantle is above its solidus below 80-100 km if temperature is
217 higher than 1000 °C (Fig. 3). Below this depth, this is the LVZ, thus the upper asthenosphere. An
218 incipient melt is present, the melt fraction being determined primarily by the water and CO₂ contents.

219 This melt has low permeability (i.e. can only move slowly by porous flow), but has a large effect on
 220 seismic and rheological properties.

221 *3.2. How oceanic mantle lithosphere of the western limb of the ridge axis becomes thicker and denser*
 222 *compared to the eastern limb?*

223 *3.2.1 Intra-lithospheric mantle decoupling*

224 In this section we will focus on the modalities of mantle mass transfer upwards and laterally within
 225 both spreading plates in the framework of the westward drift of the lithosphere. Oceanic mantle
 226 lithosphere is created at the ridge through accretion of residues above the asthenospheric partial
 227 melting area. These residues are successively transferred upwards and laterally at hS rate within the
 228 mantle lithosphere on each side of the ridge (Fig. 2). As V_r is higher than hS , the transfer rate of
 229 successive residues above the melting area is slower than the westward velocity of the melting area
 230 itself. Accordingly, mantle exhumed at the bottom of any active axial ridge is always older than the
 231 basaltic volcanoes emplaced at the axial ridge (Bonatti et al., 2003). It is the same for intrusive gabbro
 232 bodies which move upwards and laterally, driven by and with the upwelling of the host-mantle: for
 233 instance, for $hS=2\text{cm/yr}$, $V_r=4\text{cm/y}$ and a mantle residue initial depth of $D_i=40\text{ km}$, the mantle residue
 234 takes $t_1\ 2\text{ Ma}$ ($t_1=D_i/hS$) to ascend and to be exhumed at the surface. Thus active volcanoes at the
 235 ridge axis are about 2 Ma younger than the mantle on which they are formed. Besides, while this
 236 mantle residue is rising up to the ocean floor, the ridge itself with the uppermost part of the
 237 lithosphere will have moved (t_1*V_r) 80 km westward. However it may be possible that due to
 238 ‘second-stage melting’, the extra buoyancy added to the residue may accelerate the upwelling – in
 239 which case some residue may be younger than 2Ma.

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

248 Normal faults are the natural feature associated with rift zones and as indicated by rock mechanics,
 249 their mean dip is around 60° . Lower dips may naturally result from a decrease of friction due to the
 250 high heat flow and moving into shear zones within the underlying ductile layer. Besides faults and
 251 fractures may dip predominantly in one direction if spreading is asymmetric. Indeed the half spreading
 252 rate is often slightly asymmetric (Mueller et al., 2008), but this is related not to the velocity of plates
 253 relative to the mantle, but to the velocity of the ridge which sometimes is not exactly $(V_A+V_B)/2$; it
 254 may be $(V_A+V_B)/1.95$ or $(V_A+V_B)/2.05$ etc. Therefore faster and larger spreading may occur in one
 255 flank without varying the relative velocity of the two plates with respect to the mantle. This can be
 256 explained by the lateral viscosity variations in the flanks right beneath the ridge. Therefore, ridge-
 257 related faults and associated fractures would dip mainly “eastward” and be convex upwards since the
 258 westward drift is faster upwards, but their dip may also be opposite due to the half-spreading process.
 259 This is the geometry of detachment faults, i.e., large offset low-angle faults capping the eastern (rarely
 260 western) side of oceanic mantle core complexes (OCC ; Smith et al., 2008, 2012; MacLeod et al.,
 261 2009; Reston and Ranero, 2011; Mallows and Searle, 2012; Whitney et al., 2013; Maffione et al.,
 262 2013) and extending along steep faults probably rooted in a melt-rich zone. They are responsible for

263 exhumation of mantle and plutonic rocks onto the seafloor in the footwall of normal faults, whereas
 264 extrusion of basalt occurs on the hanging wall.

265 This intra-mantle decoupling within the lithosphere means also that even a mid-oceanic ridge and
 266 its related fracture network shift probably westward within the lithosphere, moving itself westward
 267 relative to the asthenosphere. This cannot be visible from the surface, and this is why general
 268 symmetry of magnetic stripes on the oceanic sea floor is preserved. However Carbotte et al. (2004)
 269 and Mallows and Searle (2012) note a westward shift of volcanic activity with time along the active
 270 Mid-Atlantic ridge axis. The geological maps of Mallows and Searle (Fig. 4; 2012) and Smith et al.
 271 (Fig. 4; 2012) show that most OCC outcrops on the western side of the ridge and that spreading rates
 272 are asymmetric with faster spreading on western plates. According to Whitney et al. (2013 and
 273 references therein), OCC account for 60% to 100% of the total plate spreading, showing that tectonic
 274 process is the main driver of mantle exhumation in the oceans and of heat and mass transfer in the
 275 Earth. Furthermore, Maffione et al. (2013) introduce the concept of detachment-mode of seafloor
 276 spreading. Moreover, the usual sill shape of gabbro intrusive bodies both at multimetric and
 277 multikilometric scale (Henstock et al., 1993 and references therein; Chalot-Prat, 2005) testifies sub-
 278 horizontal magma injections within weakness zones of lithospheric mantle, supporting intra-mantle
 279 decoupling during ocean spreading. Volcanoes grow on the topmost part of the hangingwall of these
 280 large detachment faults, structuring hangingwall rider blocks (Whitney et al., 2013; Maffione et al.,
 281 2013 and references therein). Small-scale decoupling occurs between the base of volcanoes just
 282 emplaced and underlying mantle or/and gabbro bodies in the process of exhumation. This decoupling
 283 occurs on both sides of relief where eruptions take place, underlining a pseudo-symmetric spreading
 284 (Chalot-Prat, 2005).

285 *3.2.2. Asymmetric evolution of lithospheric mantle compositions and physical parameters on each* 286 *side of the ridge*

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

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

304 In addition, if the westward drift is producing intra-lithospheric deformation preferentially on the
 305 western plate – then we might expect easier and more abundant pathways for off-axis magmatism and
 306 hence off-axis melt impregnation and channelling. Some of the upwelling diapirs may lose part of
 307 their melt fraction before arriving directly under the ridge axis. Evidence for this could be an

308 asymmetry in abundance of off-axis seamounts (i.e. volcanic chains on western plate as in Shen et al.,
 309 1993, White et al., 1998). If this off-axis process occurs – then there will be less melt arriving directly
 310 under ridge axis – and hence the eastern flank will receive long-term less fertilised residue than the
 311 western flank.

465 is higher than ρ_B (previous paragraph), μ_A must be much higher than μ_B . Since the modulus of
 466 rigidity of Fe is around 3 to 4 times higher than that of Mg and Si, the rigidity of rocks is expected to
 467 be enhanced if they are richer in Fe-rich minerals (Fyfe, 1960; Jordan, 1979). Thus higher μ_A
 468 compared to μ_B is an expected natural consequence of enhanced melt refertilization under the western
 469 plate A during westward migration of the spreading ridge.

470 **4. Deciphering the physical and compositional evolution of the asthenosphere below a spreading** 471 **oceanic lithosphere**

472 *4.1. Lower asthenosphere*

473 The lower asthenosphere is subsolidus garnet lherzolite, about 180 km thick (230 to 410 km depth)
 474 overlying the Transitional Zone beginning at ~ 410 km (Anderson, 2007; 2010; Schmerr, 2012). From
 475 the top to the bottom of this layer, seismic waves gradually increase their speed with respect to the
 476 LVZ. V_s is much higher than in the upper asthenosphere and more or less similar to that of the
 477 lithosphere (Panza and Romanelli 2014). Its mineralogical and chemical composition is inferred from
 478 experimental petrology on MORB genesis (Green et al., 2014; Green, 2015), being the N-MORB
 479 mantle source located below 230 km. It is a fertile garnet lherzolite, depleted in the most incompatible
 480 trace elements, with ≥ 200 ppm H_2O and with incipient melt at ≤ 7 GPa/ ~ 230 km, $1490^\circ C$. As shown
 481 in Fig. 3, an oceanic intraplate geotherm joins the mantle adiabat at about 230 km depth defining the
 482 base of the upper asthenosphere. The lower asthenosphere may migrate upwards as diapirs within the
 483 230-100 km including very small, relatively immobile, near-solidus melts (0.05–0.1 % of melting).

484 *4.2. Upper asthenosphere*

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

498 The LVZ has been seismically recognized below all oceans, both beneath the “western” and the
 499 “eastern” plates. It has a very small aspect ratio (thickness: 130 km; length in spreading direction:
 500 $6 \cdot 10^3$ km below the Atlantic to $11 \cdot 10^3$ km below the Pacific). Experimental results (Fig. 3) show that
 501 the upper asthenosphere corresponds petrologically to a mantle layer including small interstitial melt
 502 fractions (i.e. 0.05 to 0.1 %). According to Doglioni et al. (2005), the degree of partial melting would

503 be increased (up to 1.5 %) by shear heating (more than 100°C) generated by decoupling between
 504 lithosphere and lower asthenosphere (Fig. 3) and with largest contribution i.e. largest E-W relative
 505 movement at ~230 km depth. It may be alternatively due to higher H₂O content in the mantle (Bonatti,
 506 1990; Ligi et al., 2005).

507 Moreover, seismic images from Panza and Romanelli (2014) show an asymmetric shape of the
 508 LVZ, thicker and more elongated westward (2/3 in volume) than eastward (1/3 in volume) of ridge
 509 axis in all oceans. This spectacular and permanent asymmetry of the LVZ shape raises the question
 510 not only of the origin of partial melting, but also of its much greater development below the western
 511 part of the oceans. From experimental results (Fig. 3), the presence of these very small melt fractions
 512 have two distinct origins depending on the H₂O content of the mantle sources: 1- below the
 513 lithosphere-asthenosphere boundary (LAB) and far away from the ridge (>500 km): because of
 514 destabilization of amphibole, below 2.5-3 GPa/ 80-100 km/ 1000-1100°C, when "wet" (>200 ppm
 515 H₂O) oceanic mantle lithosphere, i.e. a subsolidus pargasite-bearing garnet lherzolite cools, subsides
 516 and becomes asthenosphere, i.e. a pargasite-out garnet lherzolite with incipient melting. 2- above the
 517 lower-upper asthenosphere boundary: during adiabatic (≈1490-1450°C) "dry" (<200 ppm H₂O) ascent
 518 of the lower asthenosphere from ≥230 km/ ≥7 GPa towards the MORB genesis area, i.e. 65-45 km/ 2-
 519 1.6 GPa below the ridge axis. Let us note that if the first partial melting origin corresponds to a rather
 520 static process linked only to the permanent thickening of the lithospheric mantle, the second is
 521 synchronous with a dynamic process of partially molten transfer of matter, both being imaged by the
 522 LVZ.

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

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

549 “Eastward” of the ridge, because of the “westward” drift of the lithosphere, the suction effect fades
 550 and mantle melting progressively mitigates top to bottom to become quite insignificant at more than
 551 1000 km from the ridge. Accordingly in this model, off-axis melt lenses on the eastern side of the East
 552 Pacific Ridge (Canales et al., 2012; Toomey, 2002, 2012) witness the asthenospheric partial melting
 553 zone underlying the ridge just before eastward abandonment due to the westward migration of the
 554 ridge.

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

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

568 All these data confirm that the asthenosphere forms a whole, 300-320 km thick, with a lower
 569 “nearly dry” (~200 ppm H₂O) solid part (≈185 km thick; 2/3 in volume) and an upper “relatively wet
 570 (> 200 ppm H₂O) solid + incipient melt” part (≈ 125 km thick; 1/3 in volume), the latter interacting
 571 with the overlying lower lithosphere. The existence of the asymmetric LVZ below the oceans
 572 indicates that below divergent plates, upwards and “eastward” transfer of lower asthenosphere is more
 573 active than elsewhere, accelerating its replacement by eastward lateral mantle flow coming from
 574 below the adjoining continents and related passive continental margin located to the “west” of a given
 575 ocean. So before flowing below and interacting with oceanic lithosphere, the upper asthenosphere
 576 underwent an earlier history of interaction with continental lithosphere, hence adding further chemical
 577 heterogeneity to intraplate mantle sources.

578 **5. How to understand, within the framework of plate tectonics, the coexistence of mid-oceanic** 579 **basalt and oceanic intraplate basalt mantle sources within the asthenosphere?**

580 *5.1. Location of mid-oceanic basalt mantle sources*

581 Experimentally (Fig. 3), MORB are shown to be sourced from a roughly adiabatic (1490 to
 582 1430°C) upwelling of the uppermost part (250-230 km) of lower asthenosphere with melt fraction
 583 increasing dramatically above its anhydrous solidus (3 GPa/~100 km; 1450°C). MORB melt
 584 segregation occurs at 15-20% melting between 2 and 1.6 GPa (65 to 45 km; Green and Falloon, 2005;
 585 Falloon et al., 2007b) within the so called “melting area” below the ridge axis in our model (Fig. 2).

586 The MORB mantle source composition is lherzolitic and fertile in terms of mineral assemblage and
 587 major element contents (Green and Falloon, 1998; 2005). However both trace element contents and
 588 Nd-Sr-Pb isotopic signatures are those of a mantle depleted in the most incompatible trace elements,
 589 *interpreted* as representative of a residual mantle composition. This decoupling between major and
 590 trace element composition interpretations is explained considering that the MORB source is itself a
 591 secondary lherzolite coming from the refertilization of a mantle residue by melts extracted from an
 592 already trace element depleted lherzolitic mantle (Frey and Green, 1974; Piccardo et al., 2007), or an

593 eclogitic magmatic crust (Eller et al., 2000), or recycled components (e.g. oceanic crust, Eiler et al.,
 594 2000; or metasomatized oceanic lithosphere, Niu et al. 2002; Shimizu et al., 2015), or a mix
 595 (Rosenthal et al., 2014). So our model assumes that the uppermost part of the lower asthenosphere is a
 596 secondary lherzolite, being the refertilization mostly provided from the recycling of W-directed
 597 subducting oceanic lithospheric slabs (Rosenthal et al., 2014; Doglioni and Anderson, 2015; Green,
 598 2015).

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

610 *5.2. Location of oceanic intraplate basalt mantle sources*

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

633 Besides the Nd-Sr-Pb isotopic signatures of intraplate basalts attest for mantle sources also
 634 metasomatized at some point in their long history by continental lithospheric products (Pilet et al.,
 635 2005; 2008; 2011). Notice that most Pacific, but also Indian "hot spots" basalts (dominant alkali
 636 intraplate basalts) are on the western or south-western plates relative to their respective axial ridges,
 637 thus are sourced within the upper asthenosphere having previously interacted with western adjoining

638 and overlying continental plates. Such an interaction may be a candidate for explaining this
639 continental imprint on intraplate basalt mantle sources, coherently with the relative “eastward” mantle
640 flow.

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

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

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

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

660 From natural and experimental data, three processes, two just below and one outside the spreading
661 ridge, are known to generate compositional heterogeneity of oceanic mantle lithosphere of both plates:
662 1) sub-ridge accretion of partial melting residues formed by melt segregation from mantle
663 asthenosphere upwelling; mantle lithosphere thickness at the axial ridge is estimated, via experimental
664 results on mantle phase equilibrium, between 30 to 80 km; 2) melt-rock reactions or mantle
665 metasomatism by basaltic melt percolation through previously accreted residues below the ridge,
666 inducing refertilisation and secondary lherzolite formation; 3) capture and addition of asthenospheric
667 mantle to the base of mantle lithosphere, because of cooling of the uppermost asthenosphere far from
668 the ridge. This last process occurs when the geotherm drops below 1050°C at 3 GPa which yields a
669 LAB at ~90km below which neither pargasite or hydrous carbonate-bearing silicate melt could exist.
670 It gives a thickened lithosphere with the growing layer of subsolidus phlogopite-bearing garnet-
671 lherzolite with carbonatite, or graphite+ H₂O+CH₄ fluid.

672 Structural data released by mapping numerous mantle core complexes, mostly immediately
673 westward of active slow-spreading ridges, give strong structural constraints for understanding
674 lithospheric mantle dynamics during plate spreading. Indeed whether in the west or east, these core
675 complexes represent outcrops of mantle and related intrusive gabbro bodies exhumed along the
676 footwall of convex up-wards detachment faults dipping eastward and rooted down to 7-8 km. We
677 interpret these detachment faults as effects of top to bottom asymmetric lithospheric mantle shear,
678 linked to the eastward flowing of asthenosphere delaying the base of the lithosphere moving
679 westward. The usual sill shape of gabbro intrusive bodies is another observation supporting intra-
680 mantle decoupling during ocean spreading. Furthermore, at the topmost part of the hangingwall of
681 these large detachment faults, a small-scale decoupling occurs between the base of volcanoes and
682 underlying mantle or/and gabbro bodies in process of exhumation, underlining this time a pseudo-

683 symmetric spreading. It follows that whatever the scale, the top to bottom decoupling dynamics
684 enables the exhumation of deep rocks (mantle with or without intrusive gabbro), thus the creation of
685 new surfaces, giving all its meaning to the concept of "detachment-mode of seafloor spreading"
686 developed by Maffione et al (2013). *This top to bottom mantle decoupling preserves totally magnetic*
687 *anomalies recorded at the surface on each side of the axis.*

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

699 The petrology of asthenospheric mantle is known experimentally since asthenosphere is the mantle
700 reservoir, at distinct P-T-depths, of MORB (lower asthenosphere) and intraplate basalts (middle to
701 upper asthenosphere). An asymmetric "incipient melt + solid" mantle zone, the LVZ, exists between
702 230 and 100 km, being much more developed and more fertile westward. So MORB are sourced
703 from a lherzolitic mantle in equilibrium at $\approx 250\text{-}230$ km / 7 GPa, the ascent of which up to 65-45 km
704 / 2-1.6 GPa just below the ridge is roughly adiabatic, thus rather fast and without any significant
705 interaction with surrounding crosscut upper asthenosphere. According to the relative "eastward"
706 direction of the asthenospheric mantle flow lines, the diapiric rise of MORB mantle source from lower
707 asthenosphere is significantly oblique and eastward. This mantle transfer is triggered and checked by
708 the suction effect of the spreading ridge, and the westward ridge migration determines a self-
709 perpetuating mechanism for permanently renewing the MORB source mantle, which explains in
710 return the great homogeneity of MORB through time. Synchronously, major mantle transfers within
711 middle to upper asthenosphere come from the lower asthenosphere farther westward and follow the
712 eastward mantle flow lines, slightly slantwise but barely deviated by the suction effect of the
713 spreading ridge. Mantle undergoes incipient melting, except that the interstitial melts percolate
714 laterally and upwards, react with and enrich the upper asthenosphere in water and the most
715 incompatible elements, a possible mantle reservoir for intraplate basalts. This incipient melting zone
716 represents also the lithosphere-asthenosphere decoupling zone affected by shear heating, reducing the
717 P/depth of intraplate magma genesis. Also the higher shear wave velocity eastward means a density
718 decrease in relation with a somewhat residual feature of the eastern asthenospheric mantle and
719 reflecting a mixing with residual mantle coming from the high degree melting area below the ridge
720 and not accreted to the lithosphere.

721 **7. Conclusions**

722 The proposed oceanic plate spreading model shows that numerous asymmetric features observed
723 with geophysics and structural analysis in the first 300 km below the oceans are consistent with
724 petrological data on upper mantle and basaltic compositions. Our multi-disciplinary study emphasizes
725 the major role of physical and chemical dynamics of the upper mantle for understanding oceanic plate
726 tectonics, and in turn how evolution of both the composition and internal structure of oceanic plates
727 strongly depends at all scales on plate kinematics. Our model suggests that both plate composition and

728 kinematics are interdependent on asthenosphere compositional evolution. The spreading of an ocean,
 729 synchronous with the migration of the mid-oceanic ridge, induces a complete and permanent material
 730 renewal, by mass transfer at all scales, particularly in the shallow 300 km of the Earth's mantle below
 731 the whole width of the oceans. Another major concept is that all petrological processes, occurring
 732 during different types of solid or/and liquid mantle mass transfers, ultimately lead to the genesis of a
 733 more or less fertile mantle composition, i.e. a lherzolite but secondary in nature (residual harzburgite
 734 refertilized by melt interaction), which is the main composition of the first 300 km of the Earth below
 735 oceans and most likely the continents as well. Incompatible trace elements abundances and isotopic
 736 values of mantle sources will necessarily vary from one specific mantle site to another reflecting the
 737 detailed individual geological histories of 'lherzolites'.

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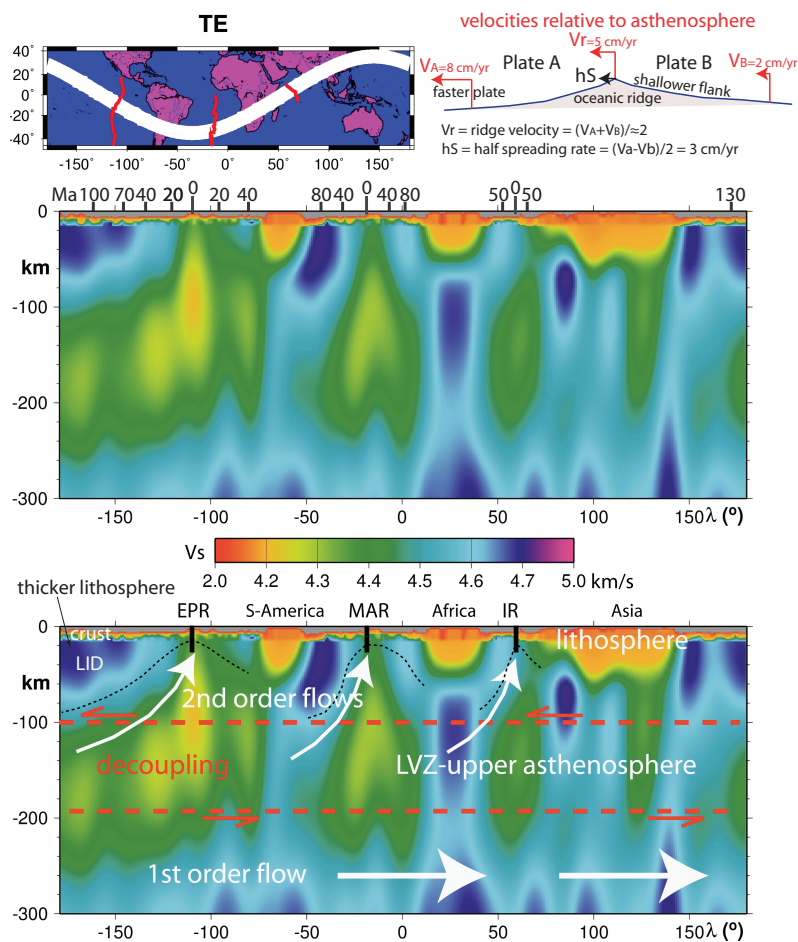
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1004

1005 **Figure captions**

1006 **Figure 1.** Uninterpreted (above) and interpreted (below). Shear wave velocity (V_s) sections along
 1007 the tectonic equator (TE) for the Earth's first 300 km. The mantle is faster and the lithosphere is
 1008 thicker in the western side with respect to the eastern side of the three major oceanic ridges, EPR—
 1009 Eastern Pacific Ridge, MAR—Mid-Atlantic Ridge, IR—Indian Ridge. Upper asthenosphere
 1010 superposes to low-velocity layer (LVZ), i.e., which is the main decoupling surface between
 1011 lithosphere and mantle, allowing net rotation of lithosphere, i.e., a first-order eastward relative
 1012 mantle flow, or westward drift of lithosphere. Secondary flow should be related to mantle obliquely
 1013 upraised along oceanic ridges. Asymmetry between two sides of ridges is independent from age of
 1014 oceanic lithosphere, shown at top in million years (modified after Panza et al., 2010).
 1015

Figure 1

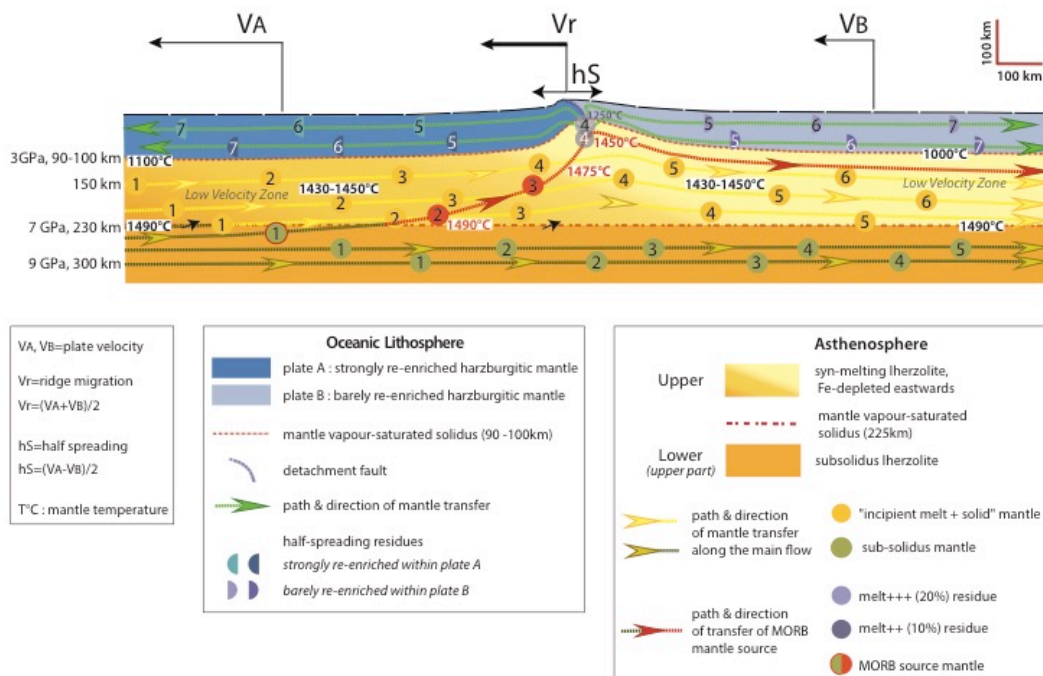


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1018 **Figure 2.** Inferred kinematics of different particles of mantle rocks along a migrating mid-oceanic
 1019 ridge, and related asymmetric upper mantle differentiation. Numbers from 1 to n indicate
 1020 progressive age evolution of the rock markers within lithosphere and asthenosphere. Plates A and B
 1021 move with velocity V_A and V_B , respectively. V_r is the velocity of the ridge; hS is the half-spreading
 1022 rate. Since the ridge is moving westward, new sections of asthenospheric mantle migrate “eastward”
 1023 relative to the lithosphere, permanently renewing mantle source of isostatically upwelling magma
 1024 generated by depressurization below the ridge. During this mass transfer, the eastern part of upper
 1025 asthenosphere becomes somewhat Fe-depleted, while a top to bottom mantle path decoupling occurs
 1026 due to the relative “westward” drift of overlying lithosphere. The upper asthenosphere (LVZ)
 1027 corresponds to the oblique ascent of partially molten mantle from lower asthenosphere. Two
 1028 different paths of transfer exist: the main one (yellow paths) barely deviated in passing below the
 1029 ridge, the other (red path) monitored by the suction effect from the strongly depressurized area
 1030 below the spreading ridge and giving MORB (see Fig. 3 for details). Mantle lithosphere is created at
 1031 the ridge through accretion of residues above the asthenospheric partial melting area, residues
 1032 successively transferred upwards and laterally at hS rate within the mantle lithosphere on each side
 1033 of the ridge. A top to bottom decoupling exists within the mantle lithosphere linked to the slow-
 1034 down by the eastward underlying flow of asthenosphere, and is responsible for detachment faults
 1035 and mantle exhumation at the spreading ridge. Besides the mantle lithosphere composition, different
 1036 on each side of the ridge, comes from a permanent refertilisation of the western plate whereas the
 1037 eastern plate just preserves the barely refertilized feature of a harzburgitic residue generated during
 1038 asthenosphere partial melting below axial ridge. See text for more details.

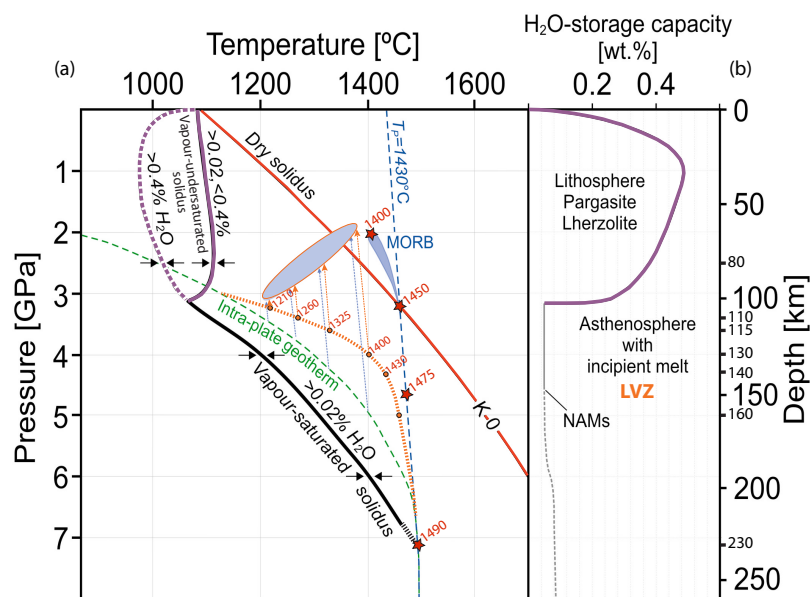
Figure 2



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1041 **Figure 3.** P-T-H₂O diagrams, modified from Fig. 6 in *Green et al.* 2014. a) Experimentally
 1042 determined solidi for hydrous silicate melt in fertile lherzolite for different water contents; *blue*
 1043 *dashed line*: mantle adiabat; *red star*: MORB source mantle upwelling potential temperature at 230,
 1044 150, 105 and 65 km; *green dashed line*: intraplate geotherm; *orange dotted line*: perturbation of
 1045 oceanic intraplate geotherm by shear heating, adapted from Fig. 4 in *Dogliani et al.* 2005; *orange*
 1046 *dot*: intraplate magma genesis at shallower depths (105 to 125 km vs. 110 to 160 km) taking into
 1047 account shear heating within the lithosphere - asthenosphere decoupling zone. MORB are sourced
 1048 from upwelling trace element depleted lherzolite from lower asthenosphere. Intraplate basalts,
 1049 including OIB, are the products of upwelling of trace element enriched lherzolite from the middle
 1050 and upper asthenosphere. See text for more details. b) Water storage capacity of lherzolitic mantle as
 1051 a function of depth along the vapour-saturated solidus. Pargasite becomes unstable at >3 GPa, and
 1052 water storage capacity drops to that (~200 ppm) which can be retained in NAMs (nominally
 1053 anhydrous minerals) in lherzolite, inducing incipient melting of upper asthenosphere (LVZ). See text
 1054 for more details.

Figure 3



1055