In press on Don L. Anderson Honor GSA-AGU Volume

Top driven asymmetric mantle convection

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Abstract

The role of decoupling in the low-velocity zone is crucial for understanding plate tectonics and mantle convection. Mantle convection models fail to integrate plate kinematics and thermodynamics of the mantle. In a first gross estimate, we computed at about 306 km^3/yr the volume of the plates lost along subduction zones. Mass balance predicts that slabs are compensated by broad passive upwellings beneath oceans and continents, passively emerging at oceanic ridges and backarc basins. These may correspond to the broad low wavespeed regions found in the upper mantle by tomography. However, W-directed slabs enter the mantle more than 3 times faster (about 232 km^3/yr) than the opposite E- or NE-directed subduction zones (about 74 km^3/yr). This difference is consistent with the westward drift of the outer shell relative to the underlying mantle, which accounts for the steep dip of W-directed slabs, the asymmetry between flanks of oceanic ridges and the directions of ridge migration. The larger recycling volumes along W-directed subduction zones imply asymmetry cooling of the underlying mantle and that there is an "easterly" directed component of the upwelling replacement mantle. In this model, mantle convection is tuned by polarized decoupling of the advecting and shearing upper boundary layer. Return mantle flow can result from passive volume balance rather than thermal buoyancy driven upwelling.

Introduction

Mantle kinematics is still poorly constrained due to the limited information available on its composition, thermal state and physical parameters, including the enigmatic toroidal component of global plate motions. Moreover, geophysical, geochemical and tectonic data are still poorly integrated and a coherent interpretation able to satisfy all the different information is missing. In this contribution we combine basic physical and petrological data and related inferences with geological and kinematic signatures at plate boundaries in order to derive an integrated view of mantle internal movements (Foulger et al., 2005; 2013).

The most relevant issues on which the model discussed here is based are the following: 1) the low-velocity zone (LVZ) at the base of the upper thermal boundary layer (BL) is superadiabatic, whereas at least parts of the mantle from the LVZ base to the transition zone and down to the lower BL (D", or core-mantle boundary) is subadiabatic (Moore, 2008; Anderson, 2013; and references therein); 2) the LVZ has a low viscosity (e.g., Jin et al., 1994; Pollitz et al., 1998) that allows decoupling between the crust-lithospheric mantle (LID) system relative to the underlying mantle (Doglioni et al., 2011); 3) plate boundaries show systematic asymmetries in terms of the geological, geophysical and kinematics parameters which appear constrained by the geographic polarity (Doglioni et al., 2007; 2014; Panza et al., 2010); 4) intraplate basalts are derived from the upper part of the LVZ, i.e., within the upper BL, whereas MORBs arise from the deeper transition zone

(Anderson, 2011; Presnall and Gudfinnsson, 2011); 5) volcanic chains are sourced from the shallow mantle rather than from a hotter deep mantle (Anderson and Sammis, 1970; Ligi et al., 2005); 6) a shallow intra-LVZ source determines a hotspot reference frame in which plate kinematics have a faster (>1°/Ma) westerly directed net rotation of the lithosphere relative to the mantle (Crespi et al., 2007).

Recent results confirm early surface-wave studies that indicated that the outer BL of the mantle is about 200 km thick and is strongly heterogeneous and anisotropic. The lithosphere is the shallow, cooler part of the upper BL, whereas the low-velocity zone (LVZ) is the lower, hotter part and acts as the main decoupling surface due to its lower viscosity (Fig. 1). Shearing occurs within the superadiabatic BL, which is characterized by higher potential temperature (Tp) and the local presence of fluids and melts. In this contribution we analyze the volumes recycled into the mantle along oppositely-directed subduction zones and propose a polarized mantle kinematics model.

LLAMA - the surface boundary layer

The surface BL contains laminated lithologies (LL) and aligned melt accumulations (AMA) that fuel within-plate volcanoes, which have their roots in the immediately underlying ambient mantle (Fig. 1). Broad passive upwellings sample the deeper subadiabatic portions of the upper mantle and this is the key to the properties of MORB and near-ridge mantle. Anisotropy under some ridge segments extends to below 300 km (Trampert & van Heijst 2002; Panning and Romanowicz 2006; Anderson, 2007), consistent with a transition zone (TZ) MORB source.

The type of fabric exhibited by a sheared multiphase mélange satisfies the geochemical reasoning that led to zoned-plume, streaky-mantle, marble-cake and plum-pudding type models. The neglect of this fabric, in seismic imaging, and the smoothing of shallow heterogeneity, can generate plume-like mirages and other seismic artifacts. LLAMA accounts for 30% of the upper mantle but it is mainly buoyant refractory harzburgite, containing 2% or less of melt or low-melting components (Anderson, 2011). There is therefore no petrologic or magma-volume issue with Tp in the BL approaching values as high as 1600°C at 200 km depth. The amount of magma, and the bulk chemistry of midplate tholeiites, are consistent with temperatures of ~ 1500°C at 150 km depth under Hawaii (Hirschmann, 2010; Presnall and Gudfinnsson, 2011). The number and thickness of accumulated melt-rich lamellae, not temperature, control wavespeeds, anisotropy, and the fecundity of this part of the mantle. The TZ also accounts for 30% of the upper mantle. It is a slab graveyard but consists largely of older eclogite cumulates that account for most of the basalt/eclogite inventory of the Earth (Anderson, 2007), including that which fuels the global ridge system.

LLAMA is a mélange of harzburgite with minor pyroxenite and magma, the denser lherzolite with more garnet, plus piclogite which is a dense mélange of eclogite and recycled lithosphere. Subducted crust is mainly eclogite and sinks to the base of the TZ. Downwellings are crust plus harzburgite. Upwellings are reheated plates minus their modern crustal ballasts plus entrained ancient cumulates. Therefore LLAMA consists of Lithosphere, Lid, and Ambient Mantle. It is variable in temperature and lithology, with pyroxenite occurring in host peridotite. Intraplate volcanoes tap into melt-accumulation zones in the BL while mid-ocean ridges are fed from deeper via spreading-induced passive upwellings and upside-down drainage below the BL.

LLAMA and the TZ are the sources for OIB and MORB, respectively. Mantle potential temperature, Tp, peaks near the base of LLAMA. Some oceanic crust sinks to the base of the TZ, displacing ancient eclogite cumulates. Although slab fluids are recycled quickly, dense, recycled eclogite piles up in the TZ and need not be recycled from any mass balance point of view.

Subducted olivine-rich lithosphere is buoyant when it warms up and is a temporary resident of the TZ. It can entrain ancient material as it warms and rises. Entrainment, displacement and ridge suction are the mechanisms for levitating mid-ocean ridge source material into the shallow mantle. Passive and slab-induced upwellings may come from the TZ.

Mechanisms of melt extraction from the BL and shearing in the LVZ

Melt-rich networks formed by stress-driven segregation facilitate the continuous reconstitution of such layers, and the formation of volcanic chains when the stress is appropriate (Shaw, 1973; Doglioni et al., 2005; Hirano et al., 2006; Valentine and Hirano, 2010; Conrad et al., 2010, 2011). Shearing in the BL causes melt segregation and stress concentrations that trigger dike formation, which in turn is responsible for intraplate volcanoes. The laminated and metamorphosed nature of LLAMA (Anderson, 2011) decreases the thermal conductivity and increases mantle temperatures relative to passively upwelling regions under thin BLs. The LVZ is the layer with the lowest viscosity values in the mantle (Fig. 1).

Lateral variations of viscosity in the LVZ should determine variations in lithospheric decoupling and lateral velocity gradients among plates. Therefore plate velocity should inversely correlate with viscosity (Doglioni et al., 2011); in fact the fastest-moving Pacific plate has the lowest viscosity in the underlying LVZ (Pollitz et al., 1998). What causes shearing between the lithosphere and the LVZ? Is the negative buoyancy of slabs (Anderson, 2001; 2007; Conrad and Linthgow-Bertelloni, 2003) or the astronomical dynamics such as Earth's rotation and tidal despinning acting as a torque on the outer BL (Scoppola et al., 2006; Riguzzi et al., 2010; Doglioni et al., 2011; Varga et al., 2012)? The shearing and independent motion of the lithosphere relative to the mantle is critical to the understanding of the kinematics and dynamics of the Earth. Petrological and seismological studies indicate a LVZ source of Hawaiian OIB (e.g., Anderson, 2011; Presnall and Gudfinnsson, 2011). Therefore, volcanic chains are fed by the shallow mantle, from within the decoupling layer (Doglioni et al., 2014). This implies a faster westward net rotation of the lithosphere because volcanic tracks at the surface record the motion of the lithosphere relative to the source depth, regardless of the deeper part of the decoupling zone. This is an important new constraint on the depths of the sources of intraplate magmas.

Lithospheric sinking along subduction zones

Subduction zones recycle the lithospheric part of the outer BL into the mantle, which is part of the return flow required by seafloor spreading (e.g., Stern, 2002). We compute the volume of slabs transported into the mantle by each subduction zone. We assume a constant lithospheric thickness of 100 km, a mean convergence rate for each subduction zone and a total subduction length of about 76,400 km. All subduction zone hinges retreat relative to the lower plate. However, relative to the upper plate they can either converge or diverge. The subduction rate, V_S, is computed from the relation $V_S=V_H-V_L$, where V_H and V_L are the velocities of the subduction hinge and of the lower plate respectively with respect to the upper plate (Doglioni et al., 2007). The velocities have been here extracted from the NASA GPS time series data base, computed in the no-net rotation reference frame (ITRF), http://sideshow.jpl.nasa.gov/post/series.html.

The results of the analysis are given in Table 1 and indicate a total subduction rate of about 306 km^3/yr along the world's subduction zones. According to this rate, it would take about 860 Ma to recycle the whole upper mantle down to the 650-km discontinuity. Therefore the majority of the subducted lithosphere volume should have been recycled at least 2 or 3 times as it would take only

170 Ma to consume the entire lithosphere at present rates. This makes sense since the harzburgite part of the plate is intrinsically buoyant at ambient mantle temperatures. However, it is not necessary to recycle oceanic crust, which only constitutes 5-6% of the slab (Anderson, 2007); it can accumulate at the base of the TZ as dense eclogite.

An intriguing result is that the W-directed subduction zones (31,700 km) provide 232 km^3/yr of recycled material, whereas the E- and NE-directed subduction zones (44,700 km) provide only 74 km^3/yr. Therefore, W-directed slabs provide about 3 times as much volume as E- and NE-directed subduction zones. This difference is due to the fact that along the latter, the slab hinge is converging relative to the upper plate, spreading the convergence between shortening of the upper plate and actual subduction, the rate of which is therefore smaller than the convergence rate. The opposite occurs along W-directed subduction zones, where the slab hinge often diverges relative to the upper plate, thus increasing the convergence rate and producing a larger subduction rate. Plates move faster at low latitudes. The present Tonga subduction zone has such fast recycling that it would transfer the entire lithosphere to the bottom of the mantle in 2.5 Ma in whole mantle convection schemes.

There certainly is an error in the precision of these computed amounts, which should be considered only as a first order estimate. However, the 3 to 1 ratio of the volume recycled by oppositely directed subduction zones is consistent with the occurrence of the backarc basins mainly along W-directed slabs. In fact backarc spreading indicates hinge migration away relative to the upper plate, which increases the subduction rate. Moreover, the thickness of the subducting lithosphere is larger in the western Pacific with respect to the eastern side, which would increase even more the volume of subducted lithosphere along W-directed slabs relative to the opposite slabs where younger and thinner lithosphere is recycled.

The volumes of lithospheric loss must be compensated by the same amount of mantle upwelling. If there are 306 km^3/yr of lithospheric loss, we expect the same amount of lithospheric production. New oceanic lithosphere is formed along the about 60,000 km long oceanic ridges and backarc basin systems.

The thermal BL, the upper part of which is the lithosphere, also thickens with time as it cools. The volume balance between slab loss and new lithospheric generation is shown in Fig. 2. Slabs are narrow and sink rapidly, but return flow is slow and diffused over much larger areas (Fig. 2). Broad upwellings can break up into "fingers" that refuel the upper mantle (Marquart, 2001; Anderson, 2013).

Slabs are the coolest part of the mantle and bring about heat loss. Therefore, the western margin of the Pacific mantle should be cooler than the central Pacific mantle. Tomography (e.g., Ritsema, 2005) confirms higher wavespeeds in the western Pacific (Fig. 3). Faster subduction along the western side of the Pacific apparently is not due to older plate. Convergence rate and slab dip are insensitive to the age of the lithosphere (Cruciani et al., 2005). Based on the inversion of gravimetric and shear waves data, slabs may not be denser than the hosting mantle (Brandmayr et al., 2011). Furthermore, W-directed slabs show generally continuous seismicity, which may reach the bottom of the upper mantle, since intra-slab seismicity is mostly down-dip compression. On the other hand, E- or NE-directed slabs are mainly characterized by intra-slab, down-dip extension and they show a seismic gap between 300-500 km depth, suggesting different slab kinematics relative to mantle (Doglioni et al., 2007; Riguzzi et al., 2010). The asymmetry of subduction zones can be ascribed to westward drift of the outer mantle shell (Doglioni et al., 2009).

Subadiabaticity and the geotherm

On the basis of the enigmatic seafloor flattening phenomenon, McKenzie and Bickle (1988) proposed an ad hoc mantle geotherm that has become the canonical reference geotherm for subsequent petrological interpretations (see Anderson, 2012 and references therein). McKenzie and Bickle (1988) assumed that the mantle was homogeneous, fertile, vigorously convecting and adiabatic below 100 km depth, the depth of the 'horizontal isotherm' and that the MORB source defined 'maximum allowable upper mantle temperatures'. They used this geotherm to define 'ambient mantle temperature', 'excess temperature', 'the convecting mantle' and 'lithosphere' in the way that they are still used in petrology (e.g., Herzberg et al. 2007). The geotherm and the underlying assumptions were unconstrained by, and do not agree with, seismological data. Radioactivity, secular cooling–and the consequent subadiabatic geotherm at depth–and effects of temperature and pressure on thermal properties, were not considered. When corrected for these physical and seismological oversights, both the possibility and need for hot and deep mantle plumes disappear. In a sense, the story ends here. LLAMA answers a persistent question "If plumes are not the answer, what is?" (Tackley, 2006). But several new questions arise as a consequence. For example, "Where is the MORB reservoir, and why is it colder than ambient subplate mantle?"

Although simple thermal models of lithospheric cooling predict the first order behavior of seafloor depth with age, there are significant regional variations in seafloor subsidence. Such variations cannot be accounted for by classical cooling models unless implausible variations in mantle temperatures or thermal properties are assumed. An alternative model assumes that the passive upwellings under ridge axes are cooler than ambient temperature elsewhere (Anderson, 2013). Such a model satisfactorily explains the data with temperature variations of ±100°C only. According to Niu and O'Hara (2008), topography along the global ridge system is due primarily to lithological variations that occur at depths >400 km. According to Lago et al. (1990) the anomalous subsidence away from ridges is due to lower average temperatures under ridges than under 'normal' mantle. This view is supported by the increase of seismic wavespeeds that occur below 170 km depth as one approaches a ridge from a plate interior.

Thermodynamically based geotherms in an internally heated mantle exhibit a thermal overshoot, or bump, at the base of the BL (Peltier, 1989; Bunge, 2005; Moore, 2008), followed by a subadiabatic gradient with potential temperatures that decrease with depth. This, plus the effect of pressure on melting, and the upward concentration of radioactive elements (Birch 1952; MacDonald and Ness 1961; Anderson 1965) means that the most likely source of hot midplate magmas is between 150-200 km depth at midplate locations. Thermodynamic modeling of seismic profiles (Xu, 2008) and convection (Schubert et al., 2009) confirm that this is a plausible model for the mantle. The subadiabatic thermal gradient between BLs makes it unlikely that deep active upwellings, even if they exist, will be hotter than at 150-200 km depth; the absence of a surface BL, and the unimpeded passive ascent of deeper mantle explains the properties of ocean-ridge basalts. A temperature inversion is a natural result of BL convection with internal heating and plate tectonics and does not require a plume-fed asthenosphere (e.g., Morgan and Phipps Morgan, 2007).

The adiabaticity assumption is a particular problem in geodynamics and seismology because geotherms derived from realistic fluid dynamic models and from seismic wavespeed gradients are not adiabatic (Birch 1952; Dziewonski and Anderson 1981; Bunge et al. 2001; Schubert et al., 2009; Moore, 2008; Xu et al., 2008). Potential temperatures are 300-500°C colder at the top of deep BLs than the mantle at ~200 km depth, not ~200°C hotter as generally assumed (e.g., Farnetani, 1997). Regions underlain by stagnant cold slabs can also be subadiabatic. Thermodynamic inversion of seismological data show that there is at least a 200°C decrease in temperature between 200 km depth and the base of the TZ. The inferred subadiabaticity amounts to 200-300°C.

The subplate shear BL allows survival of, and access to, distinctive non-MORB components (Kay, 1979; Pilet et al., 2011, Liu & Stegman, 2012). This BL is large enough, hot enough, and variable enough to account for the volumes, temperatures, compositions, fixity and persistence of within-plate magmatism (e.g., Silver et al., 2006; Canon-Tapia 2010; Anderson, 2011). The TZ has about the same volume as 'the convecting mantle' between BLs and is a plausible MORB source (Anderson 1981; Bercovici and Karato 2003). Thermodynamically based geotherms in an internally heated mantle exhibit a thermal overshoot, or bump, at the base of the BL (Peltier 1989; Moore, 2008; Sinha and Butler, 2007), followed by a subadiabatic gradient with potential temperatures that decrease with depth. This favors an upper mantle origin for high Tp magmas, and deeper sources for magmas with MORB-like Tp (Tatsumoto 1978). The fabric and lithology of these distinct sources controls the composition and volumes of the magmas.

The origin of ridge-related passive upwellings

The broad low velocity anomalies (LVAs) associated with ridges appear to be the tops of passive upwellings (Anderson et al., 1992; Ritsema and van Heijst, 1999; Ritzwoller et al., 2004; Kustowski et al., 2008). In the plume hypothesis the global low-velocity zone and localized hotspots are fed by narrow active upwellings that spread out laterally for thousands of km, heating and contaminating the whole upper mantle (Morgan and Phipps Morgan, 2007; Schmerr, 2012). However, within-plate volcanoes can be fueled by shallow shear-driven magma segregation (Shaw, 1973; Doglioni et al., 2005; Conrad et al., 2011) rather than by deep thermally buoyant upwellings and they may tap pre-existing ambient subplate mantle (e.g., Olson et al., 1990). On the other hand, most ridge segments and new ridges such as the Red Sea may be fed by lateral along-axis flow. Lateral transport of upwelling ridge material may be responsible for fueling new ridges such as in the Red Sea and supplying older ridge segments that are not immediately above TZ upwellings. This requires turning canonical geochemical models up-side-down, including the concepts of deeply rooted intraplate volcanoes and long-distance lateral transport of plume material. Dikes and rifts are the result of rifting and seafloor spreading, not the cause. So-called ridge-plume interactions may simply be where TZ passive upwellings intersect the ridge axis. The compositions and temperatures of near-ridge volcanoes and magmas from shallow segments of the ridge system are MORB like. The inflection of the solid adiabat due to melting may decrease the temperature by about 100°C, by the time that the mantle arrives at the surface.

Material displaced by cold slabs and entrained by passive upwellings (Fig. 6) is a plausible source of midocean ridge basalts (e.g., Tatsumoto, 1978; Anderson 1981; Bercovici & Karato 2003) but this material cannot simply be recycled oceanic crust; some of it may never have been at the surface, at least since the magma ocean stage of Earth history. The TZ has three components (Fig. 7); the original eclogite-rich layer, the result of ancient differentiation and cumulate formation (Anderson, 1989), subducted cold harzburgite ('lithosphere') that eventually warms up and returns to the surface, to be reused in the next generation of plate formation (e.g., Presnall and Helsley, 1982), and cold oceanic crust that sinks to 650 km, as eclogite, displacing older material upwards and paving over the base of the TZ (Fig. 7). The olivine-rich portions of slabs are buoyant when they warm up, and are, at most, temporary residents of TZ. We argue that depleted isotope anomalies, plus the ^{4}He excesses in MORB, are grown, and survive, in the more fertile parts of the TZ. On the other hand, low ^{3}He and U contents, and high ^{3}He/^{4}He ratios are preserved in cold refractory strata, which concentrate in the shallow mantle.

The TZ is an ideal place to store ancient residual and depleted materials as well as modern oceanic crust, and to accumulate ^{4}He. It is a more plausible MORB source than the shallow mantle, which is subject to contamination. On the other hand, the shallow location of LLAMA explains the propensity of the OIB source to be more contaminated by recycled crustal materials and slab fluids than the deeper depleted mantle, and more likely to have lost accumulated ^{4}He. Intraplate magmas are mixtures dominated by recycled components. Ironically, the fact that the OIB source could be a mixture of ‘primitive’ and crustal materials, and is therefore not isolated from plate tectonic processes, has been used to argue that it is in the deepest most isolated part of the mantle. This is a hold-over of the view that mantle differentiation and crustal growth were one way processes, with no recycling, and that the lower mantle was undifferentiated undegassed primordial material (e.g., Armstrong, 1968). The ‘purity’ of MORB, compared to hotspot magmas, and the high ^{4}He content, are reasons enough to reverse the canonical arguments about reservoir isolation and the relative locations of MORB and OIB.

Plate boundaries asymmetries

There are several persistent asymmetries in the geological and geophysical characters of plate boundaries as a function of their geographical distribution that may be related to differential rotation of the outer shell over the interior. In particular, the asymmetry is more evident if the concept of tectonic equator is introduced (Doglioni, 1993). In fact, plates move along an undulating pattern when their kinematics are analyzed in the NNR (No-Net Rotation) reference frame, such as the ITRF (International Terrestrial Reference Frame). The tectonic equator makes an angle of about 30° relative to the geographic equator. This angle has been interpreted as the lithospheric plate motion plane that averages the Earth’s precession axis. In fact the Maxwell relaxation time of the lithosphere corresponds to the precession of the Earth’s axis (Doglioni, 2014). Moreover, this pattern becomes polarized toward the “west” (i.e., showing a net rotation) if the plate kinematics are computed relative to the hotspot reference frames (HSRF). The net rotation of the lithosphere emerges from a different reference frame (e.g., Le Pichon, 1968). The inferred net rotation varies considerably, from 0.2°/Ma to more than 1.2°/Ma. It depends on the depths of decoupling and depths assigned to the sources of intraplate magmas. It also depends on whether the HSRF uses Pacific intraplate volcanic chains (Hawaii), or includes wetspots (Bonatti, 1990) located along or close to oceanic ridges (which are by definition unstable relative to the mantle). For a discussion and different views, see Crespi et al., (2007), Torsvik et al. (2010), and Doglioni et al. (2014).

In Fig. 4 different plate kinematics are derived as a function of the source depth of the hotspot reference frame (HSRF), which is dominated by the Hawaii island and seamount chain. The faster net rotation involves a westward drift of the whole outer shell. In this reference frame all plates, albeit with different velocities, shift to the west along the tectonic equator. The velocity of plates appears to be inversely proportional to the viscosity in the LVZ. Therefore plate boundaries and plate tectonics appear controlled by lateral heterogeneities in lithosphere-LVZ decoupling. Shearing in the LVZ is compatible with the strong anisotropy in that layer. Structural anisotropy (sheared lamellae) and lattice preferred crystal orientations are mostly parallel to the motion of plates relative to the mantle (Fig. 5). If the observed anisotropy is due to crystal alignment, essentially 100% orientation of olivine crystals is required (Tanimoto, 1990). If structural anisotropy is responsible for the preferred orientation, the low rigidity layers must be magma rich. The decoupling and shearing in the LVZ part of the BL implies the presence of a torque acting on the lithosphere.

Plates therefore move along the tectonic equator, and they flow “westerly” relative to the underlying mantle. This pattern is in agreement with the observed asymmetry across subduction zones as a function of the slab polarity relative to the mainstream of plates. In fact the westward drift of the outer shell (Nelson and Temple, 1972; Dickinson, 1978; Uyeda and Kanamori, 1979; Doglioni, 1993; 1994) is supported by the steeper slabs (average 65°) and associated backarc spreading of W-directed subduction zones, which are dipping against the relative motion of the subplate mantle (Riguzzi et al., 2010). Accretionary prisms for W-directed subduction zones have low elevation, are single verging with shallow decollements and mostly consist of shallow layers of the lower plate. Along the opposite subduction zones, being sustained by the relative mantle flow, E- or NE-directed slabs are much shallower (average 27°), and the related orogeny is doubly verging, highly elevated, affecting the whole upper plate lithosphere during oceanic subduction and eventually also the lower plate at the collisional stage (Doglioni et al., 2007; Garzanti et al., 2007; Riguzzi et al., 2010).

Asymmetry has been described also along rift zones; the western flanks on average are 100-300 m deeper, and show a seismically higher speed and thicker lithosphere and a lower speed and thicker low-velocity layer relative to the opposite limb (Panza et al., 2010; Doglioni et al., 2014). These observations agree with the notion of the net rotation or westward drift of the lithosphere relative to the mantle (Doglioni, 1993; Crespi et al., 2007; Cuffaro and Doglioni, 2007). The net rotation of the lithosphere or BL implies a decoupling at its base in the LVZ (Doglioni et al., 2011), which has been ascribed to the Earth's rotation and its slowing by tidal friction (Scoppola et al., 2006; Riguzzi et al., 2010; Doglioni, 2014; Doglioni et al., 2014).

Model of mantle convection

An isolated system that is cooling by conduction through the top is characterized by narrow active downwellings and broad athermal upwellings. Internal heating slows down the reduction in temperature. Slabs represent the active downwellings and broad regions of extension and low-wavespeed under ridges and near-ridge volcanoes are the tops of upwellings. Most of the mass flux into the mantle is cold plate, which is primarily refractory harzburgite. An equivalent outward flux occurs at ridges where plates are reconstituted. The plate tectonic cycle is dominated by olivine-rich lithologies, such as harzburgites and dunites, with fluids and basaltic crust playing minor but, for geochemistry, important roles. The subducting package includes sediments, fluids, crust and the dominant subcrustal component, all of which have different fates. If slabs are 120 km thick, the basalt part is 5% of the volume. In a steady state, a constant crustal volume implies that passive upwellings entrain enough TZ eclogite to account for about 5% of their volume. Slabs sink because they are cold and, below some 60 km, are weighed or pushed down by relative mantle flow (Fig. 6). They entrain a small amount of older, warmer eclogite as they rise to keep the ridge system fueled.

The surface BL contains aligned melt accumulations that are the source of within-plate magma. Sediments and some slab fluids exit the slab at shallow depths and are incorporated into LLAMA. The residual depleted slabs enter the TZ, displacing older eclogite cumulates, which are entrained into subridge flow. The harzburgite parts of slabs return to the shallow mantle when they warm up and become part of subsequent plates. Part of the reason for the heterogeneity of the outer 200 km of the mantle is the presence of melt and the accumulation of volatiles and of buoyant recycled products, which include refractory infertile olivine-rich residues and fertile enriched low-melting components. Instead of being homogenized to a fine scale by vigorous convection, these are juxtaposed, but segregated by shear, into a large-scale mélange.

Convection in the interior of a planet can be driven by secular cooling, lateral temperature gradients and by motions of the BLs. Montagner and Romanowicz (1993) and Tanimoto (1990) have shown that lateral heterogeneities are shifted moving across major mantle boundaries. Convection in an internally heated system is composed of narrow active downwellings and broad passive upwellings, the precise opposite of the assumptions of Morgan (1971). The system is driven from above by the cooling and BL motions. The superadiaticity and thermal bump maintains the LVZ viscosity low enough to allow decoupling of the overlying BL.

Earth's rotation, changes in length of day, Chandler wobble, tidal drag and changes in surface loads may also influence the torques acting on the plates, which may explain the motion of the outer shell relative to the mantle. Regardless of whether slabs are pushed down by these torques or pulled down by their negative buoyancy, they cool the mantle and their descent has to be compensated by mantle upwelling (Fig. 7). Broad passive upwellings sample subadiabatic portions of the deep upper mantle; this is the key to understanding the properties of MORB and sub-ridge mantle.

Because of subduction asymmetry, large "easterly" directed (along the tectonic equator) subplate motions are implied (Fig. 6). Differential rotation requires extreme density and/or viscosity heterogeneity either in the outer shell or in the interior, extremely low viscosity in the interior or between shells, or external forces. Is negative buoyancy of the lithosphere enough to drive plate motion (Anderson, 2001) or is this a combined effect of Earth's rotation and slowing by tidal friction with thermal cooling (Riguzzi et al., 2010; Varga et al., 2012)? In both cases the "westerly" directed shear from the top may be modulating convection and its polarization (Fig. 6).

Conclusions

The mantle is highly heterogeneous in density and viscosity both vertically and laterally. Slabs cool off and contaminate the mantle. Mid-ocean ridge basalt compositions are compatible with a source in the TZ, where it is isolated from the major effects of recycling contamination. Volcanic chains can be sourced in the upper part of the LVZ. Ancient enriched components are plausibly isolated in the cold outer shell. The lithosphere is sheared "westward" relative to the mantle along the mainstream of plate motion and therefore W-directed subduction zones contribute to a large fraction of mantle recycling, possibly about three times as much as the opposite E- or NE-directed subduction zones. A similar amount of recycled material has to move to the east. Upward return flow is much more diffuse than the narrow-slab, focused downward flow. Passive upwellings are expected to be slow and may not represent recently subducted lithosphere. Passive upwellings from the TZ are likely to be rather broad–dome or ridge–like, but can become more 3D as they rise (Marquart, 2001) with fingers refueling the LVZ (Fig. 6) everywhere in the upper mantle, not only along oceanic ridges. This essentially passive upwelling refuels the LVZ, which ultimately feeds the ridge system (Fig. 7). Mass balance does not require that the LVZ be plume-fed from the CMB (Morgan and Phipps Morgan, 2007). Deep, passive upwellings from a subadiabatic interior will be colder than the mantle at 200 km depth.

All this occurs in the frame of a generalized relative "westward" drift of the lithosphere relative to the mantle, along the so-called tectonic equator, in which plate boundaries have characters depending on the polarity of the subduction and the rift. The mainstream of the lithosphere relative to the mantle is punctuated by mantle uplift isostatically rebounding, or lithospheric forced descent, which may represent Rayleigh-Taylor instabilities of the convection system, e.g., second order flow patterns. This passive and polarized convection system is allowed by the >1°/Ma "westerly" directed decoupling of the lithosphere (upper BL) over a low-velocity layer Lubricating Layer of Aligned Melt Aggregations (LLAMA) and the underlying mantle. The decoupling occurs at the depth of the undulated thermal bump of the Earth's geotherm (100-220 km), where the mantle viscosity is at local minimum, due to the higher Tp and the presence of fluids and melt.

The effects of compression on physical properties and subadiabaticity of most of the mantle result in sluggish and large-scale lower-mantle convection. It is still controversial whether plate tectonics is the cause or the result of mantle convection. Asymmetric subduction is believed to be the cause of polarized mantle convection rather than the reverse. In this view, mantle convection is primarily driven from the top and sinking of parts of the cold outer shell. Volume balance requires upwelling of an equal volume but this volume need be neither hot nor intrinsically buoyant relative to its surroundings. This is more akin to Archimedes' and Benard's experiments than to Rayleigh convection, driven by thermal gradients alone. Benard-Marangoni convection, for example, is driven from the top.

Acknowledgments
Discussions with Enrico Bonatti, Eugenio Carminati, Françoise Chalot-Prat, Marco Cuffaro, Giuliano Panza and Federica Riguzzi were very much appreciated. Three anonymous referees constructively reviewed the manuscript. Thanks to Gillian Foulger for several helpful suggestions. The research was supported by the Sapienza University and Prin-Miur grants.

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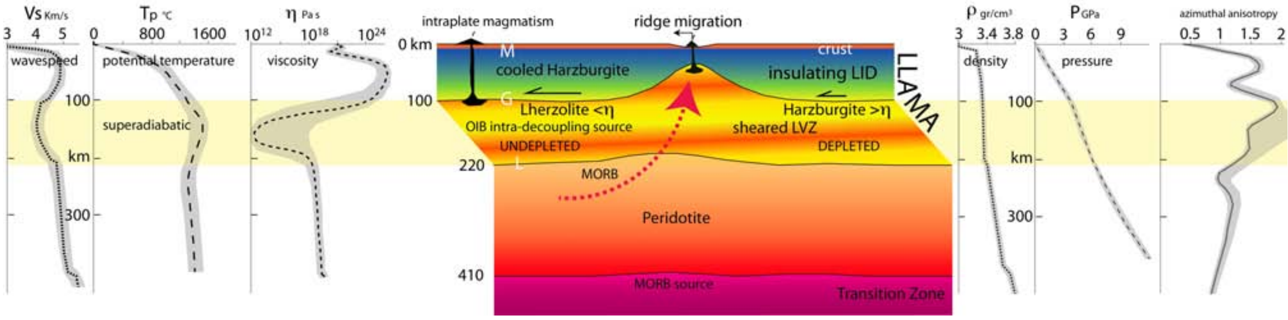

Fig. 1. Main nomenclature and parameters relating to the upper 450 km. Data after Jin et al. (1994), Pollitz et al. (1998), Anderson (1989, 2007, 2013), Doglioni et al. (2011), Tumanian et al. (2012) and references therein. Azimuthal data are the root mean square of relative azimuthal anisotropy amplitude in % with respect to G/L, where G is the anisotropy amplitude, and L (Love) is the elastic parameter, from Yuan and Beghei (2013). The gray bands tentatively indicate the potential variability. The lithosphere and the seismic lid (LL) are underlain by aligned melt accumulations (AMA), a low-velocity anisotropic layer (LVZ) that extends from the Gutenberg (G) to the Lehmann discontinuity (L). The amount of melt in the global LVZ is too small to explain seismic wavespeeds and anisotropy unless the temperature is ~200 K in excess of MORB temperatures, about the same excess as required to explain Hawaiian tholeiites (Hirschmann, 2010) and oceanic heat flow. This suggests that within-plate volcanoes are sampling ambient (local) BL mantle. The lowest wavespeeds and the highest temperature magmas are associated with the well-known thermal overshoot. The most likely place to find magmas hotter than MORB is at this depth under mature plates, rather than in the subadiabatic interior. Most of the delay & lateral variability of teleseismic travel times occur in the upper 220 km. If this plus anisotropy is ignored, the result in relative tomography can be a plume-like artifact in the deep mantle. M, Moho.

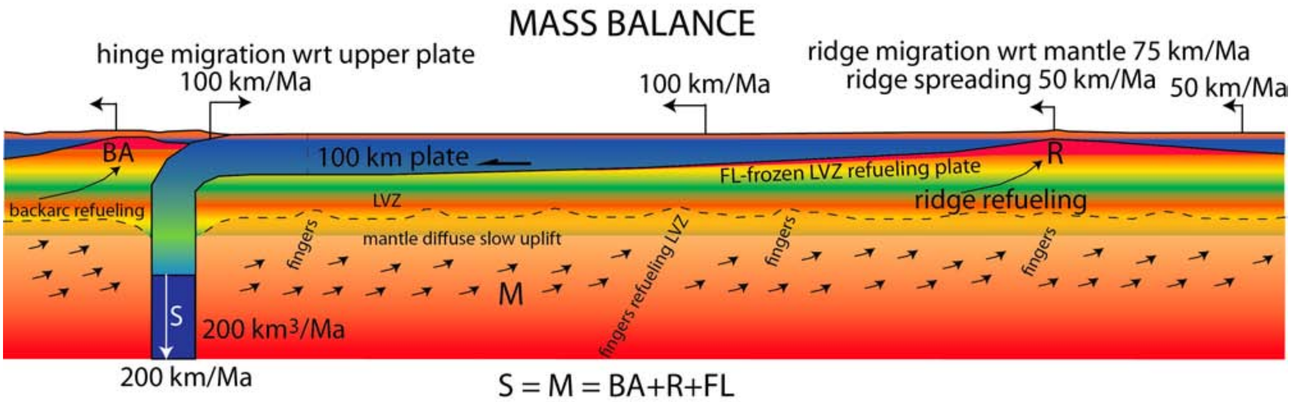

Fig. 2. Slabs provide focused recycling into the mantle, which must be compensated by an equivalent volume of diffuse mantle upflow which may include shallow fingers. W-directed slabs contribute three times the lithospheric feeding of the mantle than oppositely directed slabs (see Table 1). Assuming a portion of lithosphere 100 km thick and 1 km along strike, the slab volume entering the mantle every million years is given by the velocity of the subduction hinge relative to the upper plate, minus the convergence rate of the lower plate relative to the upper plate. In the example, 200 km^3 of slab (S) enter the mantle every million years. The mass balance predicts a similar amount of mantle (M) should rise to compensate the slab flux. Mantle uplift is accommodated by ridge spreading (R), BL cooling and freezing (FL), plus mantle uplift along backarc basins (BA) that accommodates slab rollback. LVZ, low velocity zone.

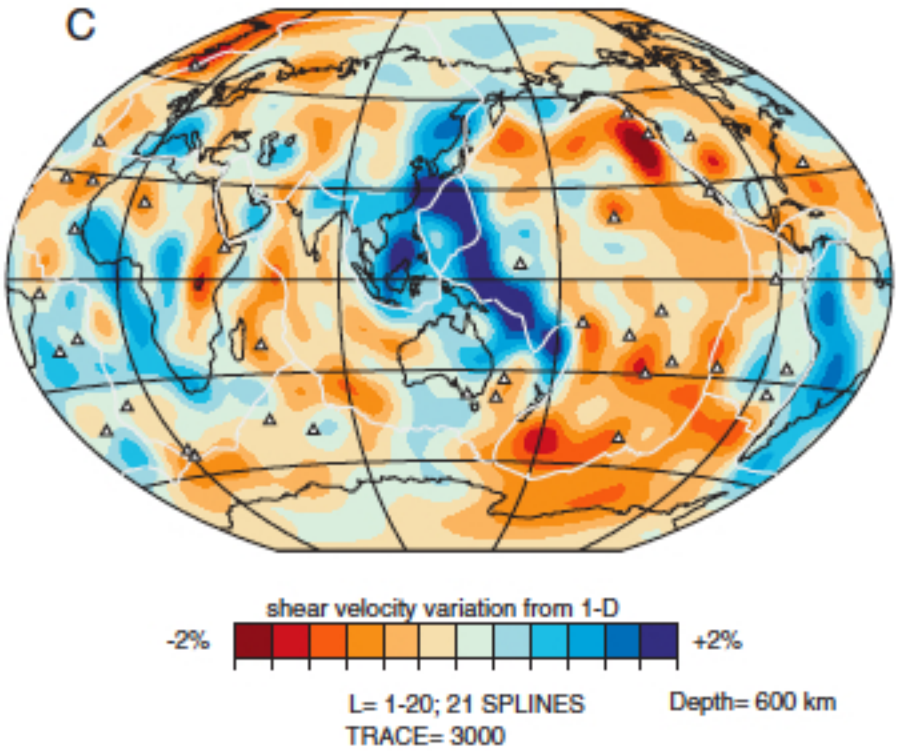

Fig. 3. Shear-wave velocity anomalies at 600 km depth, after Ritsema (2005). Note the higher velocities along the western side of the Pacific, confirming the larger and deeper volumes of cooler lithosphere along the W-directed slabs, as indicated in Table 1 and Fig. 2.

	E-directed	convergence	hinge wrt	subduction rate	volume rate		W-directed	convergence	hinge wrt	subduction rate	volume rate
	length	mm/yr	upper plate mm/yr	mm/yr	km3/yr		length	mm/yr	upper plate mm/yr	mm/yr	km3/yr
N-America	2573	37	-18	19	4,9	Barbados	2273	0	20	20	4,5
S-America	9900	66	-35	31	30,7	Sandwich	1582	0	18	18	2,8
South Ameri	1641	22	-14	12	1,9	Aleutians	3674	65	15	80	37,9
Antarctica	1793	8	-5	3	0,5	Kurile-Kamchatka	2141	80	14	66	13,7
Macquarie-N	2563	6	-4	2	0,5	Carpathians	1110	0	2	2	0,3
Papua-NewG	2511	5	-3	2	0,5	N-Japan	1039	80	-19	61	6,5
Indonesia	5127	75	-50	25	12,8	Nankai-Ryukyu-Philippi	4320	70	23	93	39,3
Himalaya	2995	36	-30	6	1,8	Izu Bonin-Marianas	4703	50	30	80	37,6
Zagros	3539	24	-14	10	3,5	N-NewZealand-Tonga-K	3233	80	100	180	58,2
Dinarides	1114	4	-3	1	0,1	Bismarck-Solomon	3457	60	20	80	27,7
Hellenides	952	36	-15	21	2	Banda	2555	0	10	10	2,5
Cyprus	731	10	-5	5	0,4	Apennines	1700	3	2	5	1
Taiwan-Born	2860	60	-35	25	7,1	Total length	31787				232
Alps	850	1	-0,5	0.5	0,04						
New Hebride	3098	60	-40	20	6,2						
Caucasus	2481	12	8	6	1,5						
Total length	44728				74,44						

Table 1. Subduction volumes at subduction zones as a function of their polarity, i.e., W-directed versus E- and NE-directed along the tectonic equator. Assuming the upper plate is fixed, the velocity of subduction V_S is given by the velocity of the subduction hinge V_H relative to the upper plate minus the convergence rate V_L between the upper and lower plates ($V_S=V_H-V_L$; Doglioni et al., 2007). For each single subduction zone the volume of the slab transported into the mantle is computed, assuming a constant lithospheric thickness of 100 km, and a mean convergence rate for each subduction zone. Velocities are computed on the ITRF-NASA data base (http://sideshow.jpl.nasa.gov/post/series.html). W-directed subduction zones appear to deliver into the mantle at least three times the volume of oppositely directed subduction zones. This is a preliminary estimation of the total recycled volumes and the differences between the oppositely directed subduction zones.

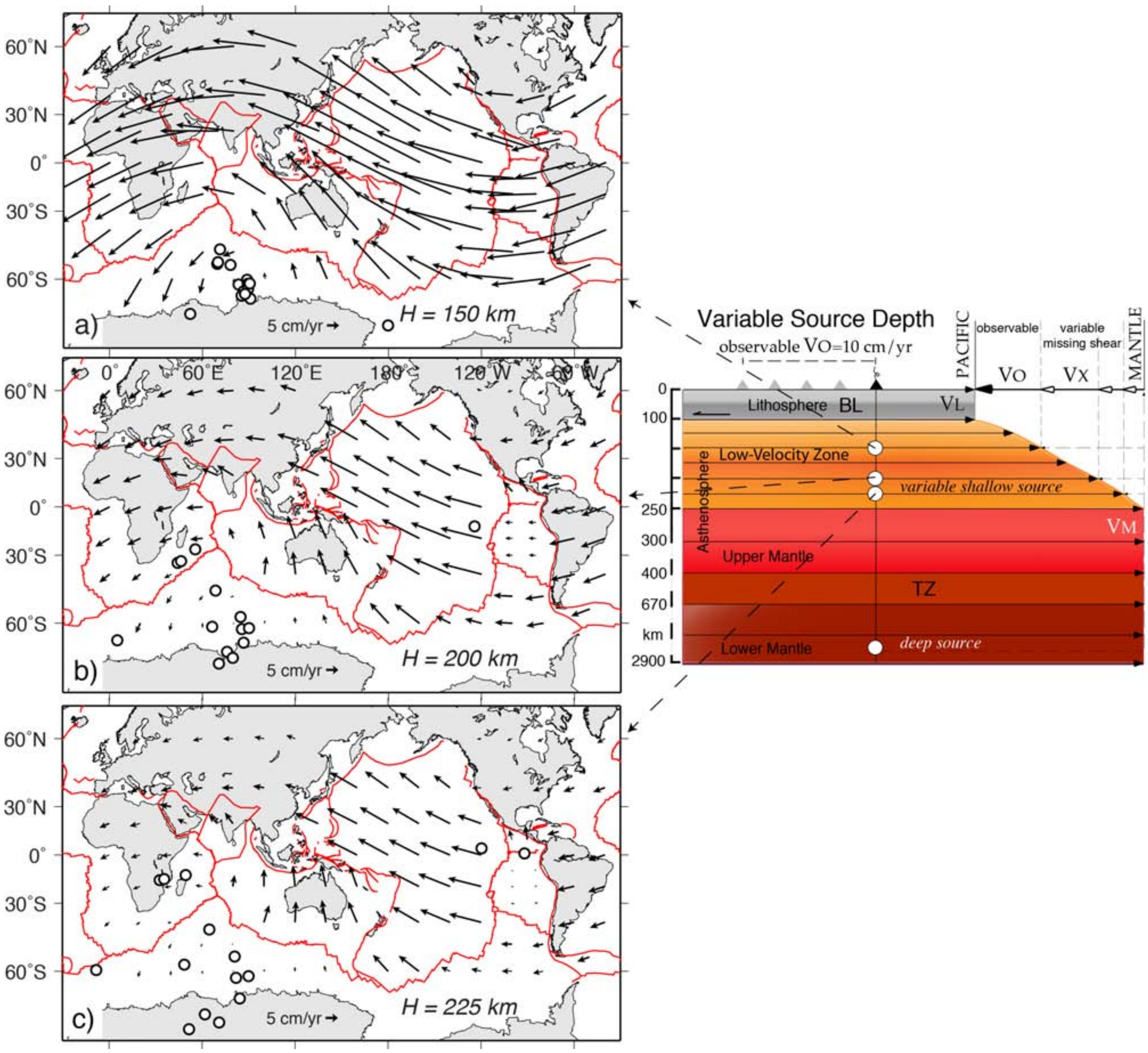

Fig. 4. Plate kinematics computed assuming a hotspot reference frame "anchored" in the asthenosphere, i.e., within the decoupling zone. Note that the faster decoupling and "westward" drift of the lithosphere (>1°/Ma) occurs if the volcanic tracks are fed from a depth of about 150 km a), within the LVZ since the surficial age-progressive volcanic track does not record the entire decoupling between lithosphere and the mantle beneath the LVZ. The two other deeper magmatic sources at 200 and 225 km, b) and c), would result in slower westward drift. An even deeper source (2800 km) would make the net rotation very slow, in the order of 0.2-0.4°/Ma (slightly modified after Doglioni et al., 2014).

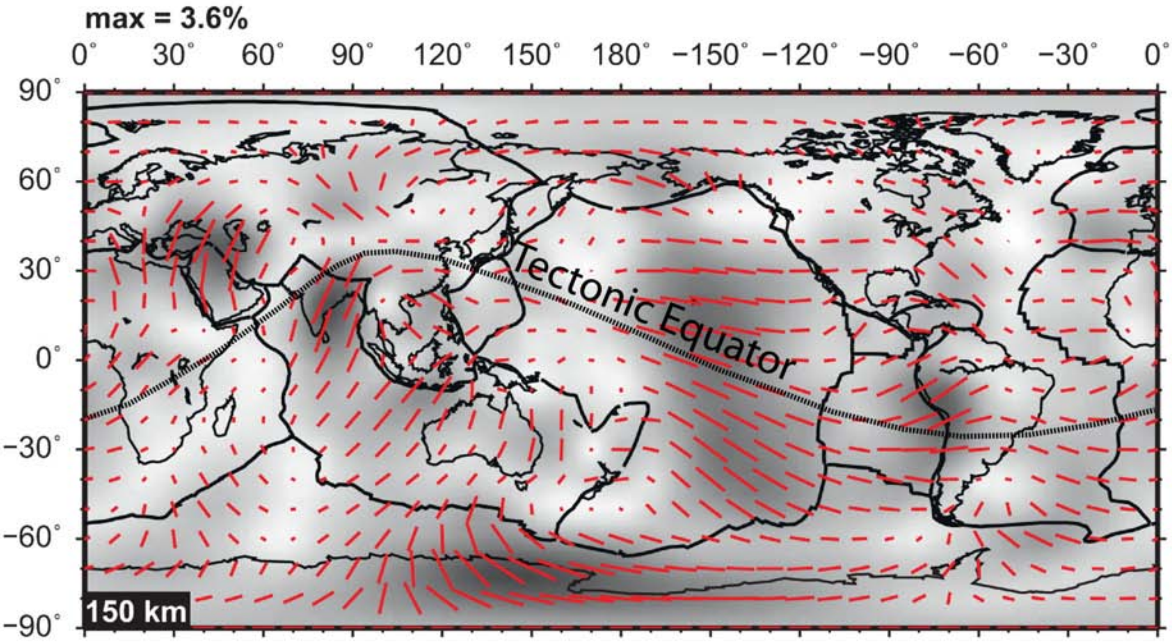

Fig. 5. Azimuthal anisotropy at 150 km depth according to Yuan and Beghei (2013). The red bars represent the fast direction of propagation for vertically polarized shear-waves and their length is proportional to the amplitude of the anisotropy. The "tectonic equator" is added on the map, showing a significant correlation in terms of overall direction. This is interpreted as the preferred trend of LPO (lattice preferred orientation) of crystals within the decoupling layer (LVZ) plus the structural anisotropy due to the tilted laminated structure of LLAMA, generated by the shear of the plate moving "westward" relative to the underlying mantle along the undulated trend. Compare this map with Fig. 4.

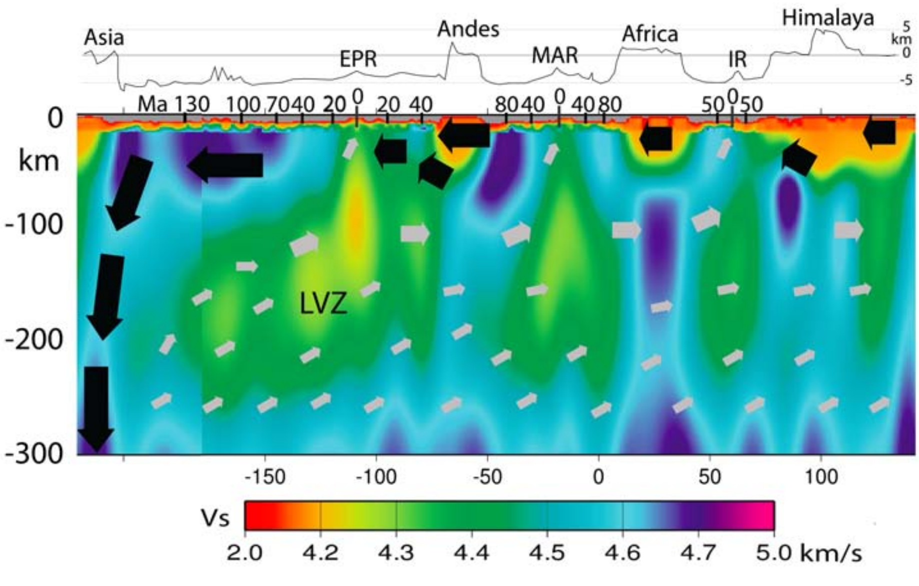

Fig. 6. Shear-wave absolute tomography after Panza et al. (2010) along the tectonic equator in which are inferred and schematized the westerly polarized vectors of the lithosphere (black arrows) and the mantle "eastward" relative motion (gray arrows). Notice the concentrated larger volumes of lithospheric recycling along the western Pacific slabs. The mantle compensates the lithospheric loss with a diffuse, relatively easterly-directed upwelling, up to the sheared LVZ, where the lithosphere is decoupled and shifting to the west. Numbers above the tomographic section indicate the age of the oceanic crust.

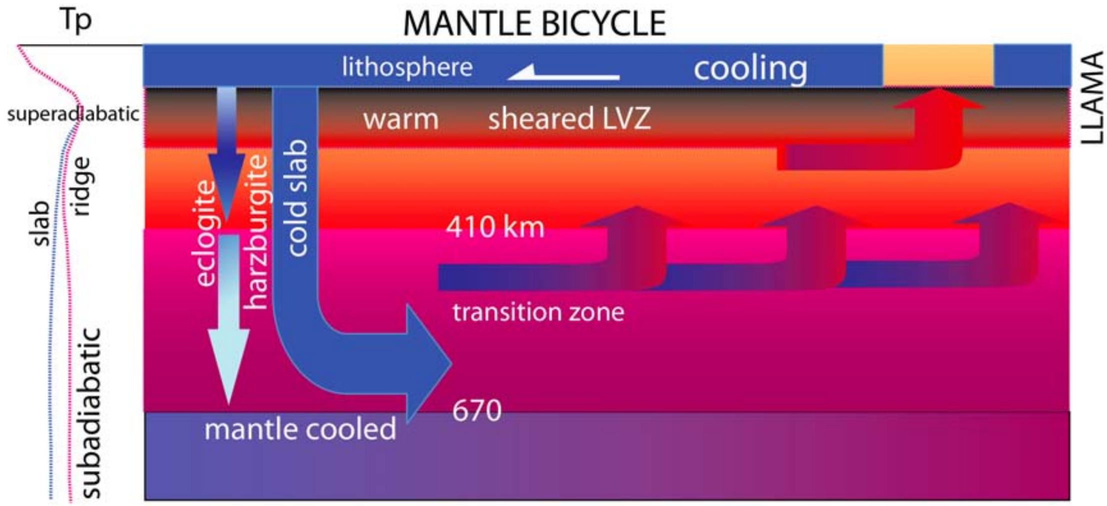

Fig. 7. Counterclockwise convection results in lateral variations of mantle potential temperature. Along subduction zones the host mantle is inferred to be cooler than elsewhere. The thermal buffer exerted by the cold lithosphere, the radiogenic decay, and the shear heating in the LVZ, all indicate a superadiabatic upper asthenosphere and a subadiabatic lower upper mantle. The "bicycle" is polarized by the westward drift of the lithosphere and the relative easterly directed compensating mantle, balancing slab loss along W-directed subduction zones. The intrinsically buoyant harzburgite component of slabs contributes to the upward return flow.