The Westward Drift of the Lithosphere: A rotational drag?

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Abstract

Net westward rotation of the lithosphere relative to the underlying mantle is a controversial phenomenon first attributed to tidal effects, and later to the dynamics of mantle convection. In spite of a number of independent geological and geophysical arguments for westward tectonic drift, this phenomenon has received little recent attention. We suggest that this differential rotation is a combined effect of three processes: 1) Tidal torques act on the lithosphere generating a westerly directed torque decelerating the Earth's spin; 2) The downwelling of the denser material toward the bottom of the mantle and in the core slightly decreases the moment of inertia and speeds up the Earth's rotation, only partly counterbalancing the tidal drag; 3) Thin $(3-30$ km) layers of very low viscosity hydrate channels occur in the asthenosphere. It is suggested that shear heating and the mechanical fatigue self-perpetuate one or more channels of this kind which provide the necessary decoupling zone of the lithosphere.

Keywords: Earth's rotation, westward drift, lithosphere, asthenosphere viscosity, decoupling

1 INTRODUCTION

Since the westward drift of the American blocks described by Wegener (1915), there have been a number of papers proposing a global or net westward drift of the lithosphere relative to the mantle (Rittmann, 1942; Le Pichon, 1968; Bostrom, 1971). This net rotation is indicated by independent kinematic observations such as plate motion within the hotspot reference frame (Ricard et al., 1991; O'Connell et al., 1991; Gordon, 1995; Gripp and Gordon, 2002), plate motion relative to Antarctica (Le Pichon, 1968; Knopoff and Leeds, 1972) and geological asymmetries (Doglioni, 1993).

Tidal or Earth rotation effects were invoked to explain this westward drift (Bostrom, 1971; Knopoff and Leeds, 1972; Moore, 1973). Jordan (1974) and Jeffreys (1975) attacked the theoretical basis of these tidal drag mechanisms, and the model was abandoned. The notion that Earth's rotation influences plate tectonics has been generally discounted due to the requirement of the conservation of the angular momentum of the Earth-Moon system considered as an isolated system. However there is a body of evidence suggesting an astronomical tuning of plate tectonics, such as the distribution of plate velocity and seismicity, which tend to decrease toward the Earth's poles (DeMets et al., 1990; Heflin et al., 2004). Transform faults are longer in the equatorial zones, and mantle thermal minima are also concentrated around the equator suggesting a migration of cooler and heavier material at low latitudes due to centrifugal mass redistribution (Bonatti, 1996). The pole-fleeing force is an example of the rotational component acting on plates, particularly on an oblate planet (Eotvos, 1913; Caputo, 1986a; Gasperini, 1993). Polar wander is also influenced by initiation of subduction zones or internal mass redistributions in the mantle (e.g., Spada et al., 1992).

As Jordan (1974) noted, the idea of tidal drag as the driving mechanism for plate tectonics is particularly intriguing (e.g., Bostrom, 1971; Moore, 1973) because it is energetically feasible. In fact the dissipation of energy by tidal friction is slightly larger ($1.6 \cdot 10^{19}$ J/yr) than the energy released by tectonic activity ($1.3 \cdot 10^{19}$ J/yr), e.g., Denis et al., 2002).

However, Jordan (1974), and later Ranalli (2000) discarded the Earth's rotation as the cause of the westward drift, claiming that the viscosity necessary to allow decoupling between lithosphere and mantle should be about 10^{11} Pa s in the intervening asthenosphere. This value is too low when compared with the present day estimates of the asthenosphere viscosity, ranging between $10^{17} - 10^{20}$ Pa s (Anderson, 1989; Pollitz et al., 1998; Fjeldskaar, 1994; Giunchi et al., 1997; Piersanti, 1999). Therefore the viscosity of the asthenosphere is crucial for understanding the nature of the westward drift of the lithosphere.

In this paper we attempt to revitalize the idea of an astronomical origin of the westward drift, including new ingredients to the Jordan (1974) model, such as the non-linear rheology of the mantle, the mechanical fatigue and the irreversible downwelling of the heavier rocks in the mantle. Moreover we hypothesize the presence of a ultra low-viscosity channel within the asthenosphere (Fig. 1).

2 WESTWARD DRIFT

There are a number of open basic questions regarding the westward drift, including: What is its real speed? What is generating it? Does it affect the entire lithosphere or it is rather only a mean value, with most of the lithosphere moving "west", but part of it still moving "east" in the opposite direction relative to the mantle? Ricard et al. (1991) proposed that the westward drift is only a mean value due to the lower asthenospheric viscosity at the base of the Pacific plate, but geological and geophysical signatures of subduction and rift zones rather show a global signature, supporting a global relative "eastward" motion of the mantle relative to the lithosphere (Fig. 1). Gripp and Gordon (2002) computed an average "westward" speed of the lithosphere relative to the asthenosphere of up to about 49 mm/yr, using the hotspot reference frame.

It is crucial to detect whether hotspots are fixed relative to each other (Molnar and Stock, 1987; Steinberger, 2002) in order to have a reliable hotspot reference frame and to compute the westward drift of the lithosphere. Norton (2000) grouped hotspots into three main families that have very little internal relative motion (Pacific, Indo-Atlantic and Iceland). In his analysis, Pacific hotspots are nearly fixed relative to each other during the last 80 Ma.

We do not have yet a reliable constraint on the source depth of the hot spots (deep mantle or asthenosphere) but, whatever this depth, hotspots indicate relative motion between lithosphere and the underlying mantle.

If hotspots have their source in the asthenosphere (Bonatti, 1990; Doglioni et al., 2005), and we disregard those hotspots located along plate margins do not constitute a reliable reference frame because they are moving relative to each other, then the net rotation of the lithosphere rises to 90 mm/yr. The only assumption made is that the Pacific hotspot tracks are fixed relative to each other (Norton, 2000; Gripp and Gordon 2002) and that they parallel the motion of the underlying mantle relative to the lithosphere.

The most obvious place for decoupling to occur is the asthenosphere where the lowest mantle viscosity values are believed to occur. Kennedy et al. (2002) have shown how mantle xenoliths record a shear possibly achieved at the lithosphere-asthenosphere interface. This supports the notion of a flow in the upper mantle and some decoupling at the base of the lithosphere (Russo and Silver, 1996; Doglioni et al., 1999; Bokelmann and Silver, 2000). A significant radial anisotropy, with horizontally polarized shear waves traveling faster than those that are vertically polarized, is present under continental cratons at 250-400 km depths and under oceanic plates at shallower (80-250 km) depth (Gung et al., 2003). This anisotropy has been related to horizontal shear in the low-viscosity asthenospheric channel, which is thinner below the continents than beneath the oceans (Gung et al., 2003). This is in agreement with a shear in the asthenosphere distributed worldwide. A global shear wave splitting analysis in the asthenosphere (Debayle et al., 2005) show directions consistent with a mantle shear along the undulating pattern of flow suggested by surficial plate motions (Doglioni, 1993). Deviations from this flow occur particularly along subduction zones where the flow is inferred to encroach the slabs.

Horizontal plate speeds range between 1-150 mm/yr, whereas vertical motion (uplift or subsidence) of the lithosphere typically has rates between 0.01-1 mm/yr. Therefore, on average, horizontal velocities are 10 to 100 times larger, suggesting the greater importance of tangential forces acting on the lithosphere, i.e. the toroidal field.

The westward drift can be interpreted as a toroidal field of degree one (Ricard et al., 1991). Bokelmann (2002) suggests that in order to explain the toroidal component, plates and mantle cannot be fully coupled. He also proposed that the mantle is the dominant force in moving North America, based on the dip and orientation of P-wave fast azimuths axes. Holtzman et al. (2003), modeling a decoupling zone, showed that in simple shear experiments on several mantle-like melt-rock systems at high temperature and pressure, melt segregates into distinct melt-rich layers oriented 20° to the shear plane. As an application, in real peridotites, melt-rich bands dipping toward the sense of motion can develop in the mantle near a shear zone. According to the dip of the fast azimuths axes described by Bokelmann (2002), this would indicate that mantle beneath Western North America can move eastward relative to the lithosphere, as also suggested by Silver and Holt (2004). Similar eastward asthenospheric flow has been interpreted by Negredo et al. (2004) for the Caribbean plate. Upper mantle anisotropy parallel to plate motion suggests eastward asthenospheric flow even beneath the East Pacific Rise (Wolfe and Solomon, 1998).

Horizontal shearing across the asthenospheric decoupling zone implies that a force is acting on the lithosphere that opposes to direction of mantle flow. This force can be exerted either by the lithosphere itself (e.g., slab pull or ridge push), or by an external force such as tidal drag. Slab pull is frequently invoked (e.g., Anderson, 2001; Conrad and Lithgow-Bertelloni, 2003) and is notoriously considered one order of magnitude larger than ridge push. Estimates for the magnitude of the slab pull are based on a number of assumptions such as a homogeneous upper mantle geochemistry, and a thermally- and phase changes-driven negative buoyancy. However, the mineralogical constraints of the upper mantle are quite weak, and focal mechanisms indicate that slabs undergo down dip compression beneath depths of 350 km, if not shallower in many subduction zones (Isacks and Molnar, 1971; Frepoli et al., 1996; Doglioni et al., 1999). Assuming an active pull from only that part of a slab between depths of 50 km and 350 km, and considering for example the Marianas slab, the following concerns can be envisaged. The negative buoyancy of about 300 km long slab should be able to pull the 10,000 km long Pacific plate, overcoming the friction at its base without breaking the lithosphere, even though it has very low strength under extension (Turcotte and Schubert, 1982). It should also contemporaneously determine the slab retreat. Based on these counterarguments we are exploring here an astronomical alternative.

Plates move along a sort of mainstream, which is not everywhere oriented E-W, but rather smoothly deviates by up to 60-70 degrees in azimuth, depending on longitude, so as to depict an approximately sinusoidal pattern of flow (Doglioni et al., 1999). Along this flow regime, roughly west-directed subduction zones are steeper than east or northeast-directed zones (Nelson and Temple, 1972; Dickinson, 1978; Uyeda and Kanamori, 1979; Doglioni, 1993; Marotta and Mongelli, 1998), and the associated orogens are respectively characterized by lower structural and topographic elevation, backarc basin, and in the other side by higher structural and morphological elevation and no backarc basin (Doglioni et al., 1999). The asymmetry is striking when comparing western and eastern Pacific subduction zones, and it has usually been interpreted as related to the age of the downgoing oceanic lithosphere, i.e., older, cooler and denser in the western side subducts more steeply. However these differences persist elsewhere regardless the age and composition of the downgoing lithosphere. In fact along the West-directed Apennines, Carpathians, Banda, Barbados and Sandwich subduction zones, where even continental or zero-age oceanic lithosphere subducts, the slab is very steep and the related accretionary prism is small. In the opposite East- or Northeast-directed subduction zones such as the Dinarides-Hellenides, Taurides, Zagros, Himalayas, Indonesia, New Guinea and New Zealand belts, the slab is less inclined and shallower, and the orogen has much larger volumes of involved rocks. Since the asymmetry seems more related to the geographic polarity along the aforementioned sinusoidal flow, rather than to the age and composition of the subducting lithosphere, it appears more connected to an astronomical origin.

Along West-directed subduction zones (e.g., Marianas) the decollements at the base of the accretionary prism are rather shallow, in the upper crust, and therefore they have small orogenic volumes accreted. Along the opposite East- or Northeast-directed subduction-related orogens such as the Andes or the Himalayas, the basal decollements affect the entire crust and upper mantle, and the resulting belt has much larger volumes of rocks (Doglioni et al., 1999). This could explain why the topography of the East-directed subduction-related orogens is higher than the opposite subduction zones. In fact, when using pairs of subduction zones where same age oceanic lithosphere is subducting, the upper plate of West-directed subduction zones regularly shows lower elevation with respect to the opposite subductions (Fig. 2).

Rift zones are also asymmetric, with the eastern side being more elevated of about 100-300 m worldwide. A test of these asymmetries is also in Fig. 2, where the mean topographic and bathymetric elevations are reported for the western and eastern flanks of rift zones. Along rift zones, the asymmetry can be interpreted as related to the depletion of the mantle due to melting along the rift. The residual lighter asthenosphere shifting east- or northeastward might determine a mass deficit in the eastern flank of the ridge, producing a shallower bathymetry (Doglioni et al., 2003).

3 ASTHENOSPHERE VISCOSITY

Apart from experimental studies, the presently known lowest viscosity of the asthenosphere occurs beneath the Pacific plate ($5 \cdot 10^{17}$ Pa s, Pollitz et al., 1998), detected with modeling of earthquake remote triggering. It is an average viscosity (Fig. 3), and therefore we might expect both horizontal and vertical variations of this value. The Pacific asthenosphere is also possibly the most undepleted mantle and so prone to melt. It is noteworthy that the Pacific is the fastest plate in the hotspot reference frame. Therefore there appears to be a relationship between asthenosphere viscosity and plate velocity.

The rheology of the mantle is not entirely understood. The viscosity quantifies the resistance of a fluid to flow, and it is the ratio between the shear stress and the strain rate. Some materials have a viscosity which depends on the time scale of an applied shear stress. The time scale of tidal drag can be considered as infinite. Studies of the mantle's mechanical properties during the last decades have repeatedly pointed out the non linear rheology of the mantle (e.g., Caputo, 1986b; Kornig and Muller, 1989; Ranalli, 1995), which usually implies viscosity in the asthenosphere decreases as the shear stress raises. Viscosity of the asthenosphere computed on the time scale of post-glacial rebound (10 ka) can be significantly different from the one related to long lasting processes (10 Ma). Different models predict different signs on viscosity's time scale (or strain rate) dependence. But there are several classes of models in which viscosity decreases with increasing time scale.

3.1 Asthenosphere viscosity from numerical modeling

Inferences on the asthenosphere viscosity η come from numerical models and from laboratory experiments. Several modelling techniques have been used. Viscosities between $7 \cdot 10^{19}$ and $5 \cdot 10^{20}$ Pa s were inferred from post glacial rebound models (e.g., Fjeldskaar, 1994; Kaufman and Lambeck, 2000). A viscosity of $0.3 - 2 \cdot 10^{19}$ Pa s was proposed using models of post-rifting stress relaxation at divergent plate boundaries (Foulger et al., 1992). Simulations of post-seismic deformation in the Japan region provided viscosity values around $0.93 \cdot 10^{19}$ Pa s (Suito and Hirahara, 1999). Modelling of the oceanic geoid suggests that the viscosity of the asthenosphere is between 3 and 4 orders of magnitude smaller than the viscosity of the lower mantle (Kido et al., 1998). The viscosity of the lower mantle is constrained to about 10^{21} Pa s (Vermeersen et al., 1998). As a consequence, asthenospheric viscosities around $10^{17} - 10^{18}$ Pa s are constrained from geoid modelling. Similarly low ($5 \cdot 10^{17}$ Pa s) viscosities were obtained for the asthenosphere below the Pacific plate using earthquake remote triggering techniques (Pollitz et al., 1998). In summary, asthenospheric viscosities from numerical modelling range reasonably between $10^{17} - 10^{19}$ Pa s. It is here emphasized that such viscosity values are conjectured mean values, averaged over the entire thickness of the asthenosphere. This is mainly due to the limited vertical resolution of such studies.

3.2 Asthenosphere viscosity from laboratory experiments

Laboratory experiments suggest that the viscosity profile of the asthenosphere could be far less homogeneous than envisaged from numerical modelling studies (Fig. 3). Both water and melt content of asthenospheric rocks can significantly influence their viscosity (Hirth and Kohlstedt, 1996; Mei et al., 2002). In deformation experiments on partially molten olivine aggregates, Hirth and Kohlstedt (1995a; 1995b) found that the viscosity of the upper mantle can be reduced by more than one order of magnitude if the retained melt fraction is greater than 0.05. These results were confirmed by experiments on partially molten lherzolite samples (Zimmerman and Kohlstedt, 2004). Using the results of such experiments, several synthetic viscosity profiles were calculated for the mantle. For example, Hirth and Kohlstedt (1996) calculate a variable viscosity profile (with mean viscosity values as low as 10^{18} Pa s) for a melt-free oceanic lithosphere. Mei et al (2002) calculate, for the asthenosphere of a mantle wedge resting above a subducting plate, a very rough viscosity profile. The viscosity of the mantle wedge can vary by about 3 orders of magnitude (between less than 10^{16} Pa s and more than 10^{18} Pa s) over a depth span of 60 km, due to the combined effects of water and melt weakening (Fig. 3).

The distribution of both water and melt could be particularly uneven in the asthenosphere. Larger than expected water content in the upper mantle has recently been envisaged (van der Meijde et al., 2003). In particular the melt content is likely to be controlling deformation. Holtzman et al. (2003) conducted laboratory experiments of large strain deformation and demonstrated that melt can segregate into melt rich regions. These melt rich channels can contain 10 times (or more) the melt with respect to adjacent rocks. As a consequence, melt-rich channels are expected to be significantly weaker than the surrounding material. This process can concentrate deformation into these channels, which are likely to become shear zones. Such shear zones are commonly observed in exhumed mantle rocks of ophiolite complexes. For example, tabular dunite bodies, considered to be the product of fast basaltic melt migration into mantle rocks, are juxtaposed to shear zones in the Josephine peridotites (Kelemen and Dick, 1995). The occurrence of large melt fractions in exposed mantle shear zones has been also proved in the Oman ophiolites (Dijkstra et al., 2002). The same authors, combining such field observations and the experimental results of Hirth and Kohlstedt (1995a; 1995b), suggest that the effective viscosities in mantle peridotite shear zones could range between 10^{15} and 10^{16} Pa s.

A further control on asthenosphere viscosity can be played by lattice preferred orientation of olivine crystals in the mantle. Seismic anisotropy (e.g., Gung et al., 2003) is controlled by deformation-induced lattice-preferred orientations of olivine, which is the most abundant and the weaker mineral of the upper mantle. The direction of polarization of the fast S-wave is parallel to the [100] axis of olivine (e.g., Mainprice and Silver, 1993). The same occurs with the direction of fast propagation of both Rayleigh and P-waves. Experiments on single crystal olivines indicate that this mineral displays a strong mechanical anisotropy. Slip on (010)[100] planes occurs with strain rates one order of magnitude larger than on (010)[001] planes (Bai et al., 1991). In other words, slip parallel to the [100] mineral axis is much easier than slip parallel to the other axes. Numerical modelling suggests that [100] axes of mantle olivine are generally horizontal and parallel to the mantle flow direction (Tommasi et al., 1999). This result is supported by olivine lattice preferred orientation measured in natural peridotites (Tommasi et al., 2001). The same conclusions are reached by seismic anisotropy studies. It can be concluded that in the same rock the viscosity opposing horizontal shear parallel to the flow direction (e.g., due to horizontal plate motion) will approximately be one order of magnitude lower than the viscosity opposing shear in the vertical direction (e.g., induced by glacial loading, Fig. 4). These experimental findings clearly suggest that the asthenospheric viscosities obtained from numerical modelling are to be considered average values. A strong vertical variability, with water- and melt-rich layers characterised by much lower viscosities (down to 10^{15} Pa s), is on the contrary more realistic. The combination of the effects of all the weakening parameters (water content, melt content, shear localization and mechanical anisotropy) has to be still investigated. It cannot be excluded that such experiments could decrease the accepted values of asthenosphere viscosity.

A final remark is to be dedicated to the space and time scales of laboratory experiments. Deformations and strain rates investigated are respectively dramatically smaller and larger compared to mantle processes. Although laboratory results are extrapolated to mantle scale, it cannot be excluded that natural mantle processes can be significantly different at higher degree of deformation and much slower deformation rates.

3.3 Ultra-low viscosity layers in the asthenosphere?

In the asthenosphere, the peridotite with minor carbon + hydrogen (lherzolite-(C+H+O) system) at a temperature of about 1430°C is partially molten (e.g., Schubert et al., 2001; Green and Falloon, 1998; Green et al., 2001). Therefore, inside the asthenosphere there may be one or more layers, with a combined vertical thickness of 530 km, with very high degrees of partial melting and consequently a viscosity 1-4 orders of magnitude lower than normal (Fig. 3). These layers, having lower viscosity, could serve as a decoupling zone between the lithosphere and the underlying mantle.

The theory of the channel flow (Cathles, 1975; Turcotte and Schubert, 1982) shows how in the computation of the post-glacial rebound, a thin (e.g., a few tens of km) low viscosity layer in the asthenosphere remains unsolved or invisible due to the much deeper effect of a 1000 km wide ice cap loading. Consider a circular ice sheet 3000 km wide (radius $R = 1500$ km) that is subsiding at $v = 1$ cm/yr (on average) and all this isostatic motion is accommodated by radial channel flow in an asthenospheric channel only $d = 20$ km thick. Then the volume of material flowing through a circumferential boundary (lying beneath the edge of the ice sheet) per unit time is Av. The area A of this base is πR^2, while the area of the boundary is $2\pi Rd$. So the average velocity of outflow at the boundary is $Av/(2\pi Rd) = (R/2d)v$. So the average velocity v is scaled up by the ratio R/2d which has a numerical value of 37.5. Since the flow rate has to be zero at the bottom and top of the channel, the peak velocity of the channel has to be even higher than this, and so vertical velocity gradient = strain rate in the channel would be very high, of order $v(R/d^2)$. Narrow (in the vertical sense) asthenospheric channels cannot contribute significantly to isostatic subsidence or uplift induced by widely distributed loading or unloading of the lithosphere. It would be much easier to deform the entire upper mantle, even if it has a much higher viscosity.

Theoretical studies of post glacial rebound have recently considered the observable consequences of low viscosity zones within continental crust (Di Donato et al, 2000; van de Wal, 2004). For example, Di Donato et al. (2000) invoked a low viscosity zone 15 km thick in the depth range 25-40 km. This type of low viscosity zone is quite different from that which we invoke: its thickness is comparable to its depth and it is limited to continental crust. We need to invoke a layer that can localize shearing of the lithosphere everywhere (i.e. in the oceanic and the continental realms), and we suppose that the depth of this zone could be 10-50 times greater than its thickness. This depth to thickness ratio explains why this layer is very hard to detect, even given its enormous viscosity contrast.

Since we are not addressing thin shallow low-viscosity earth layers whose depth to thickness ration could be moderate to large, the undetectability of the low viscosity should rather depend on the ratio of the depth of the channel to its vertical thickness. If the channel is thin and deep, as we supposed, it will be extremely hard to detect. Small, isolated loads are extremely difficult to model in practice since they are strongly sensitive to shallow elastic structure, the lithospheric thickness, the presence of faults, etc..

In contrast, the global net rotation of say few cm/yr could be accommodated by a channel with a vertical velocity gradient of just v/d. The (R/d) scale factor does not appear. Thin channels can accommodate horizontal shearing much more easily than they can accommodate vertical motion of the lithosphere over regions much larger than the channel depth. In other words, post glacial rebound will not be sensitive to thin, very low viscosity channels that can contribute significantly to net rotation of the lithosphere.

4 MANTLE CONVECTION AND CORE GROWTH

Before we discuss our model of the astronomic origin of the westward drift, let us point out an underestimated phenomenon, i.e. the downwelling of denser material toward the deepest parts of the planet. This process tends to decrease the moment of inertia and to increase the Earth's spin. In fact, for rotating bodies, if the mass is accumulating closer to the rotation's axis, they must spin faster in order to conserve the angular momentum.

Variations in the Earth's rotational speed (Munk and McDonald, 1960; Lambeck, 1980) arise from i) an external torque associated with tidal friction, and ii) internal variations in the Earth's moment of inertia or in the angular momentum of the core (Munk and Davies, 1964). The observations we have for evaluating these two effects are the variation in the length of the day, which, based on space geodesy data, is presently decreasing at the rate of about $2.3 \times 10^{-5} s$ (Varga et al., 1998), and the recession of the Moon. The increase of the length of the day based on stromatoliths and tidal deposits indicates that the despinning of the Earth is accelerating through time (Denis et al., 2002).

Lunar laser range measurements (Dickey et al., 1994) indicate that the Moon is receding from the Earth at a rate of 38.2 mm/yr (+/-0.7). This value has been related to the lunar tidal torque alone and it has been connected to the despinning of the Earth.

Since the early recognition of mantle convection, it was proposed that down-going currents would tend to leave some of their denser constituents at the base of the mantle while less dense components rose to form the crust (Runcorn, 1962 a,b). Subsequent seismological and petrophysical discoveries refined this idea (Lay et al., 1998; Anderson, 2002). Similar mass redistribution has been recognized in the core and has been related to the nucleation (Jacobs, 1953) and growth (Buffett et al., 1992) of the inner core. There are several indications that the inner core is less than 2.5 Ga old, an age of 1 ± 0.5 Ga being preferred (Labrosse et al., 2001; Labrosse and Macouin, 2003).

Evidence for deep accumulation of denser material in the mantle could be provided by the existence of the 150-300 km thick D" region at the core-mantle boundary (Lay et al., 1998), characterized by 1.5-3% velocity discontinuities for both P and S waves and by shear wave anisotropy. Segregation of dense material from the overlying mantle in the D" region can occur by the descent of liquids or solids.

Irreversible accumulation of denser material at the base of the mantle is predicted by convection, which tends to increase density mantle stratification (Anderson, 2000; 2002). Irreversible mass redistribution within the core is controlled primarily by inner core growth, which has been calculated to occur at rates between 0.2 mm/yr and 0.7 mm/yr (Morse, 2002). Besides accumulating heavier elements in the inner core, this process disturbs the chemical equilibrium between outer core and mantle (Buffett et al., 1992; 2000). This is due to the segregation into the liquid outer core of FeSi and FeO whose concentrations increase above the equilibrium concentration with respect to the mantle. As a consequence, excess FeSi and FeO of the liquid core could be removed by formation of silicate perovskite. This buoyant phase is proposed to accumulate at the bottom of the mantle, producing the ultra-low-velocity zones at the core-mantle boundary (Buffett et al., 2000).

Therefore, irreversible accumulation of heavier elements in the bottom of the mantle and the inner core growth should generate the effect of the rotating ice skater, who spins faster when the arms are drowning, decreasing the moment of inertia.

5 A DISCUSSION OF THE JORDAN'S MODEL

As we have shown in the previous sections, there are several bodies of geological evidence for westward drift of the lithosphere. Should differential rotation between the lithosphere and the mantle have originated by a process that operated in the distant past it would not persist. The difference between the angular velocity of the lithosphere ω_1 and the angular velocity of the mantle ω_2 can not be inherited from their initial different values. The time needed to cancel the transient of a differential rotation of these two bodies is extremely short (e.g., Ranalli, 2000). Therefore we must identify an ongoing process to explain this westward drift. The original idea (Bostrom, 1971; Moore, 1973) to ascribe to the tidal effects the westward drift has been confuted by Jordan (1974). The argument is the following: assume that the system lithosphere-asthenosphere-mantle may be described by a rigid sphere (the mantle) inside a spherical shell (the lithosphere) separated by a viscous fluid (the asthenosphere). Suppose that a torque, due to tidal effects, is applied on the lithosphere. The actual value of the torque is measured by means of the enlarging of the Moon's orbit. The fluid is assumed homogeneous and is described by the linearized Navier-Stokes equation in spherical geometry. In this way the viscosity η of the asthenosphere is estimated to be around 10^{11} Pa s, way too low.

The Jordan's argument is simple and appealing, but it could be unsatisfactory for two reasons. The first reason is the fact that the growth of the Moon's orbital radius is not a good way to measure indirectly the tidal torque. Even in the two body approximation the angular momentum of a body moving along a Keplerian orbit is given by

$$L = C\sqrt{a(1-e^2)}$$

where C is a constant proportional to the square root of the sum of the masses of the two bodies, a is the major semiaxis of the orbit and e is its eccentricity.

Hence the variation of angular momentum has to be computed taking into account also the variation of the eccentricity. On the other side it is well known (e.g., Zahn, 1966) that the tidal drag tends to circularize the orbits, producing a decrease of eccentricity that tends to increase L. The value of the torque measured by Jordan using only the mean radius of Moon's orbit is therefore unreliable.

Moreover one has to take into account the fact that the system Earth-Moon is not isolated: the effect of the Sun should be taken into account. However it is easy to see that we do not have any possibility to measure directly the secular variations of the Earth-Sun orbit with a sufficient degree of approximation, and therefore we can not measure directly the transfer of angular momentum of the Earth to the Earth-Sun system. Wang (1975) also criticized the Jordan (1974) model, claiming how in a non-Newtonian system the tidal drag is effective in moving plates.
Finally, once it is understood that the measure given by Jordan of the tidal torque might be underestimated, one has to face the further problem mentioned in section 3. The secular variations of the Earth's moment of inertia may increase the Westward drift, because the tidal drag might be partially compensated by the tendency of the Earth to increase its angular velocity, due to the fact that its moment of inertia is slowly decreasing.

The second reason why the Jordan's argument is not convincing is the fact that it is based on the assumption that the asthenosphere is a Navier-Stokes fluid of high viscosity, and that this viscosity can be inferred by modeling post glacial rebound. These assuptions are most likely oversimplified as discussed in the previous sections, since the mantle has very likely a non-linear rheology, and the viscosity of the asthenosphere can be far lower in thin undetected layers. If we proceed as in Jordan, and we assume 100 km for the thickness of the intervening asthenosphere atop of the mantle, and we furthermore take into account the effect of the Earth's moment of inertia decrease due to permanent downwelling of heavier material in the lower mantle and in the outer core, we obtain $\eta \approx 10^{12}$ Pa s that is slightly larger than the Jordan's estimate, but still far from the estimates based on the post glacial rebound or, more recently, on the estimates for the internal 50-100 km thick layer of an hydrated asthenosphere as shown in Fig. 3. In conclusion, the tidal torque provides an energetic mechanism to drive the lithosphere, but it requires a decoupling between the lithosphere and the undelying mantle that is incompatible with the present knowledge of asthenospheric viscosity and mathematical models for Navier-Stokes laminar flow as suggested by Jordan (1974).
However, Assuming that the tidal torque acts in an appreciable way on the lithosphere, it is very reasonable that the resulting movement of the plates on the mantle has also a discrete component, due to the sudden breaking of solid bonds between lithosphere and mantle, occurring in the above mentioned less rigid layers. A friction of this kind can be assumed to be similar to the friction due to a fluid interspace only on long time scale. As already discussed, the internal inhomogeneity of the asthenosphere at small scale is almost undetectable with the current measurements, but it could rather play a crucial role. In the next section we present a simple model of this effect.

6 A MODEL FOR LITHOSPHERE-ASTHENOSPHERE DECOUPLING

Let us discuss a natural hypothesis that could help to find a more realistic relation between β and the measurements of the viscosity of the asthenosphere. Assuming that the tidal torque acts in a sensible way on the lithosphere, it is possible that the resulting movement of the plates on the mantle is continuous only when it is measured on a long time scale. On a short time scale the movement has a discrete component, due to the sudden breaking of solid constraints between lithosphere and mantle, occurring in the less rigid layer (the asthenosphere). We present therefore a very simplified model to describe the movement of a solid plate over a solid background

due to the breaking of a set of constraints between the two, due to the continuous action of a constant force.
We suppose that a solid plate and a solid background are constrained by the presence of N constraints, that after a certain (average) time suddenly break. The time necessary to break each constraint is an exponential random variable with average proportional to the strength of the force applied to it which, in its turn, depends on the number of constraints still active. When, at the end, all the constraints are broken, the plate moves and it stops after a small displacement, due to the fact that in its new position a new set of constraints is formed. The velocity of the plate, computed on a long time scale, is therefore inversely proportional to the time needed to break all the N constraints.
In what follows we shall say that the i-th constraint is the one that leaves, when it breaks, $i-1$ constraints intact. Moreover we will assume that the weakest constraints are the first ones to break, so that the constraint N is the weakest and the constraint 1 is the strongest. By the theory of the Poisson processes it is easy to see that this is an upper bound of the actual time needed to break all constraints.
Finally, we will assume that the force applied to the i-th constraint when it breaks is $\frac{1}{i}$ the total force, and that the force applied to the constraint and the breaking time are inversely proportional. Then we can write that

9.1 $$t_i = \bar{t}_i \times i \tag{6.1}$$

where t_i is the time needed to break the i-th constraint, and $\bar{t}_i$ is the time that would be needed to break the i-th constraint if the whole force would be applied on it. The above mentioned fact that the weakest constraint breaks first implies that $\bar{t}_i \geq \bar{t}_j$, if $i < j$.
We start with a homogeneous set of constraints, hence with the choice $\bar{t}_i = \bar{\tau}$ for all i, and we assume that the resulting time $T^{(hom)}$ needed to break all the N constraints is the time that one should associate to a homogeneous asthenosphere, hence to an asthenosphere reasonably described by the Navier-Stokes equation.

By (6.1) the time $T^{(hom)}$ is simply

9.2 $$T^{(hom)} = \sum_{i=1}^{N} \frac{\bar{\tau} \times i}{i} = \bar{\tau}\ N \tag{6.2}$$

where the factor i^{-1} in the left hand side is there because there are i constraints that can be broken when the system is in the state with i intact constraints. This time is inversely proportional to the resulting velocity, and therefore is directly proportional to the average viscosity of the interspace.

Assume now an inhomogeneity in the viscous film between the lithosphere and the underlying mantle such that there is a certain amount of “strong” constraints, that are the one due to the solid part of the asthenosphere, that have the same characteristic time $\bar{t}_i^{(s)} = \bar{\tau}$ of the homogeneous case, and are in fact the constraints responsible for the computations of the viscosity of the asthenosphere due to the response of it to a compression (post glacial rebound). Assume that the rest of the constraints are relative to the melt fraction of the asthenosphere and are therefore much weaker, i.e. $\bar{t}_i^{(w)} = \bar{\tau}^{(w)} \ll \bar{\tau}$. Even if the melt fraction of the asthenosphere is only a few percent, the shear localization of the melting (see Holtzman et al., 2003) and the fact that the small solid lithons between the melt bands have a shape such that the friction is minimized, suggest that shearing is likely to be localized in the liquid phase, with lithons remaining relatively undeformed. For this reason we assume that the fraction of solid constraints $N^{(s)}$ is much smaller than the fraction $N^{(w)}$ of weak ones.

Hence the total breaking time in the inhomogeneous case is

9.3
$$T^{(inhom)} = \sum_{i=1}^{N^{(s)}} \bar{\tau} + \sum_{i=1}^{N^{(w)}} \bar{\tau}^{(w)} \tag{6.3}$$

and this quantity can be extremely smaller than $T^{(hom)}$.
Therefore we have strong indications that the actual friction coefficient between lithosphere and mantle could be much smaller than the one computed by Navier-Stokes homogeneous approximation. The effective viscosity of a layer vertically loaded (η^v) should be much larger than the effective viscosity of the same layer subject to a shear parallel (η^h) to the layer (Fig. 4). This would lead to an average viscosity of the same order of magnitude of the one observed in the thin molten layers of the asthenosphere.
Some more effects may imply that the observed friction is smaller than the one computed by Navier-Stokes equation. In rocks, the finer grain size strongly lowers the mantle viscosity (Solomatov, 2001), and grain size decreases along decoupling surfaces. Even more, decoupling in the asthenosphere should determine shear heating which also decreases viscosity. In other words, the relative motion between lithosphere and underlying mantle could be seen as a self-sustaining process once activated. Moreover we are neglecting in this discussion another important effect, namely the solid Earth tides. Even if the dissipated energy of the solid tides does not change quantitatively the arguments above, the continuous stretching of the bonds due to the rapid semidiurnal few tens of cm up and down movement of the lithosphere under the effect of solid tides may cause a mechanical fatigue of the bonds, contributing to a further decrease of their characteristic time (Fig. 5).
Moreover, the upper mantle, and the asthenosphere in particular, have the highest Rayleigh numbers, i.e., they are the parts of the mantle more vigorously convecting because they have the lowest viscosities and the highest thermal gradients of the whole mantle section. Although it is difficult to predict the velocity and direction of flow of the particles in the asthenosphere, we infer that upper mantle convection can be efficient and chaotic also for the lateral heterogeneities of the lithosphere base and within the asthenosphere itself. This could trigger further edge driven convection (King and Anderson, 1998) that is even more decreasing the strength of the asthenosphere. The presence of convective flow in the mantle or even in the asthenosphere implies that the result (7.3), where laminar flow is assumed, is not realistic. The new arising question is: can the presence of convective cells drastically change the torque applied by the lithosphere to the mantle? In other words: can the presence of convective cells change the effective friction between lithosphere and mantle? Moreover, can the presence of the torque modify the shape and the stability properties of the convective cells (Bostrom, 2000)?
The answer is probably affirmative, but there are no results in literature about this difficult subject.
The enhanced decoupling between lithosphere and mantle might explain why the shape of the plates is different from the shape obtained assuming a simple Bénard convection or other types of convection in the mantle, and why plate margins migrate relative to each other and relative to the mantle, whereas mantle convection models depict stable areas of uprising or downwelling mantle.
Moreover, variations in strength, or viscosity with deformation indicate that the first ingredient of a plate-like flow is that the convecting fluids have non-Newtonian rheology, which means that viscosity reduces with increased deformation rate, or strain rate, which is a self-supporting process (Bercovici, 2003). Analysis of the deformation in fluids has shown how the decoupling

may concentrate in a thinner surface within the originally homogeneous medium. The strain rate is highly concentrated in bandings where the viscosity is drastically non-linearly decreased near one of the planes bordering the flow (e.g., Salmon et al., 2003; Varnik et al., 2003).
Finally, the tidal fluxes and refluxes due to the presence of continents have the effects to invert their action every six hours, with a small westward drift component. This rapidly variating force could have the further effect to decrease the rigidity of the asthenosphere, and hence the strength of the bonds in the model presented in this section, further decreasing the effective friction. It is noteworthy that larger the ocean, the faster the spreading rate at the oceanic ridge, suggesting a possible correlation of larger tidal dissipation in larger oceans.

7 CONCLUDING REMARKS

The essential points of this paper are: i) there is an observed net westward drift of the lithosphere with respect to the hotspot reference frame, of 50-90 mm/year; ii) tidal torque provides a sufficiently energetic mechanism to drive this motion, but requires a mechanical decoupling between the lithosphere and the deeper mantle that is incompatible with current understanding of upper mantle viscosity and mathematical models for Navier-Stokes laminar flow; iii) a thin, low-viscosity layer (shear zone) could accommodate this motion; iv) such a layer might not be detectable using classical analysis of post-glacial rebound.

(Mei et al., 2002) recently estimated viscosity of the asthenosphere lower than 10^{16} Pa s in thin intra-asthenospheric layers. These ultra low viscosity layers could enable decoupling of the lithosphere relative to the mantle induced by rotational drag. The contribution of Earth's rotation to the relative westward motion of the lithosphere could account for the inadequate kinematics of mantle convection on plate tectonics (e.g., Anderson, 1999), and it would provide an explanation on where most of the large amount of energy related to rotation is dissipated and to the balance of forces which are controlling the length of the day (e.g., Lambeck, 1980; Varga et al., 1998; Krasinsky, 1999; Denis et al., 2002). The net rotation of the lithosphere associated with lateral variations of the viscosity-controlled coupling between lithosphere and underlying mantle can determine variable relative velocities between plates, i.e., extension or convergence, or in other words, plate tectonics.

According to this model, shear zones within the asthenosphere should be detected. Recently, based on migrated stacked seismic receiver functions, Zandt et al. (2004) interpreted the occurrence of low-velocity shear zones in the upper mantle underneath the Sierra Nevada.
In this view, plate tectonics would occur with the combination of a rotating planet under tidal torque, efficient internal convection, and lateral viscosity variations at the lithosphere-mantle interface where are supposed to occur thin hydrate layers with very low viscosity, far lower than the average estimates predicted by post-glacial rebound. The permanent, although low, tidal drag and the fatigue could determine the "westward" drift of the lithosphere relative to the mantle (Fig. 1). We do not have a final positive answer to the title question, i.e., if the westward drift IS related to the Earth's rotation. However we consider this likely and worthy of further investigation. It cannot be neglected on the basis of the results of oversimplified mechanical and rheological models.

As an application, this model could explain why unlike the Earth, satellites where stronger gravitational tides operate (e.g., the Moon, the four largest Jupiter's satellites, Io, Europa, Ganymede, Callisto) and the internal convection is low or absent, they moved to the orbital resonance condition (Boccaletti and Pucacco, 2002), or tidal locking where the time of rotation equals the time of revolution around the main planet, and plate tectonics do not occur. On the other hand, moonless planets do not show plate tectonics similar to the Earth.

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Figures caption

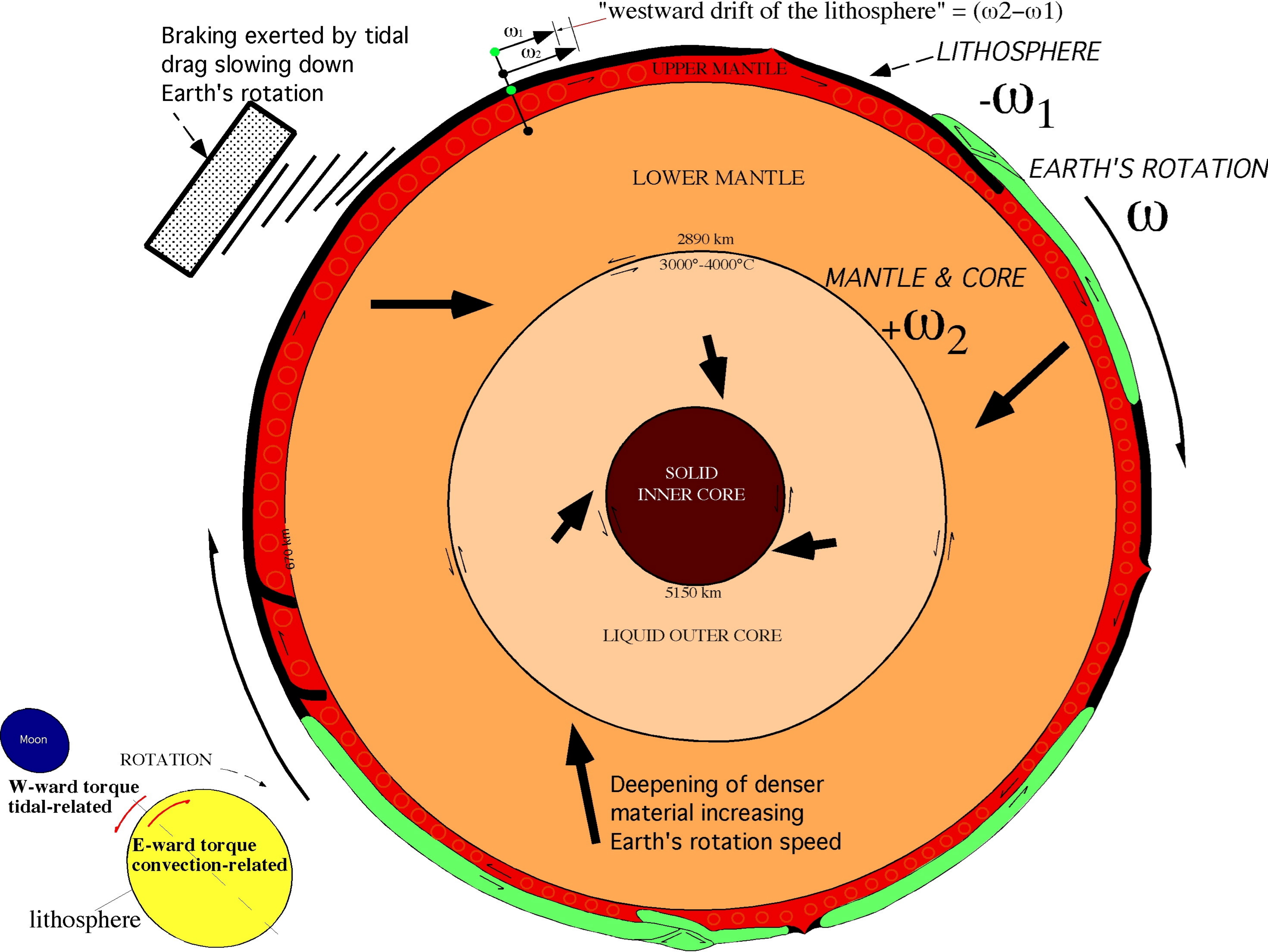
Figure 1. Cartoon showing how Earth, viewed from the southern pole, is undergoing two opposite torques. The tidal torque is opposing Earth's rotation, whereas convection is speeding up the spinning due to the accumulation of heavier material toward the inner parts of the planet, growing the inner core and making the lower mantle denser. Very low viscosity, hydrated layers are inferred in the non-linear rheology asthenosphere where the tides perform mechanical fatigue. The mixing of these different issues could allow decoupling between lithosphere and underlying upper mantle where more vigorous convection takes place.

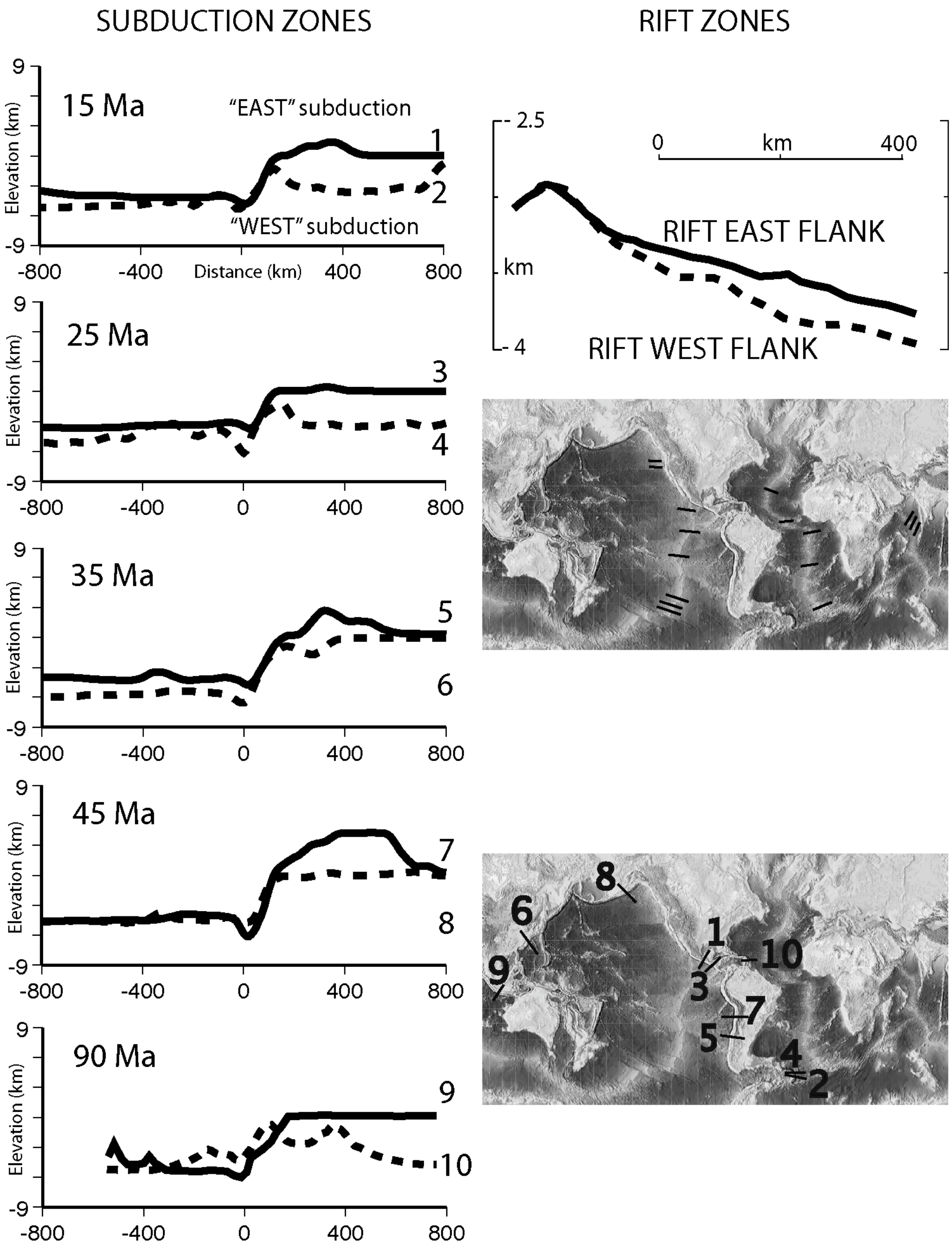
Figure 2. To the left are elevation profiles along pairs of subduction zones with the same age of the oceanic crust at the trench, of 15, 25 (Central America, sections 1 and 3, and Sanwich arc, 2 and 4), 35 (South America and Ryukyu, sections 5 and 6), 45 Ma (South America and east Aleutians, sections 7 and 8) and 90 Ma (Sumatra and Barbados, sections 9 and 10). Numbers of the sections refer to their location in the lower right panel. Note that the West-directed subdution zones with dashed lines have lower upper plate elevation, regardless the age of the subducting lithosphere. To the right are the average bathymetry of the sections crossing oceanic rift zones shown in the middle panel. The western side of rift zones indicated by the dashed line is also less elevated than the opposite flank. Both subduction and rift zones provide evidence for a geographic polarity in global tectonics.

Figure 3. Viscosity profiles of lithospere and asthenospheric mantle. The solid line profile is after Mei et al. (2002) whereas the dashed profile is after Hirth and Kohlstedt (1996). These two curves were calculated using flow laws and information on water content and melt fraction obtained from laboratory experiments. The dotted profile is referred to the work of Pollitz at al. (1998), and is the product of numerical modelling of earthquake remote triggering. Notice the significant variation of viscosity values in the asthenospheric mantle.

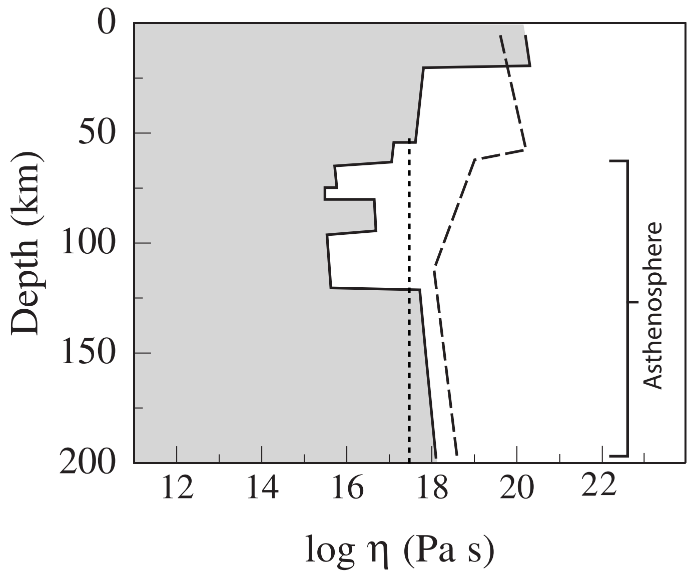
Figure 4. The effective viscosity in a granular layer with intervening melt in the pores is much smaller when measured for a shear parallel to the bedding (e.g. induced by horizontal plate motion) with respect to a vertical load (e.g., induced by ice formation or melting).

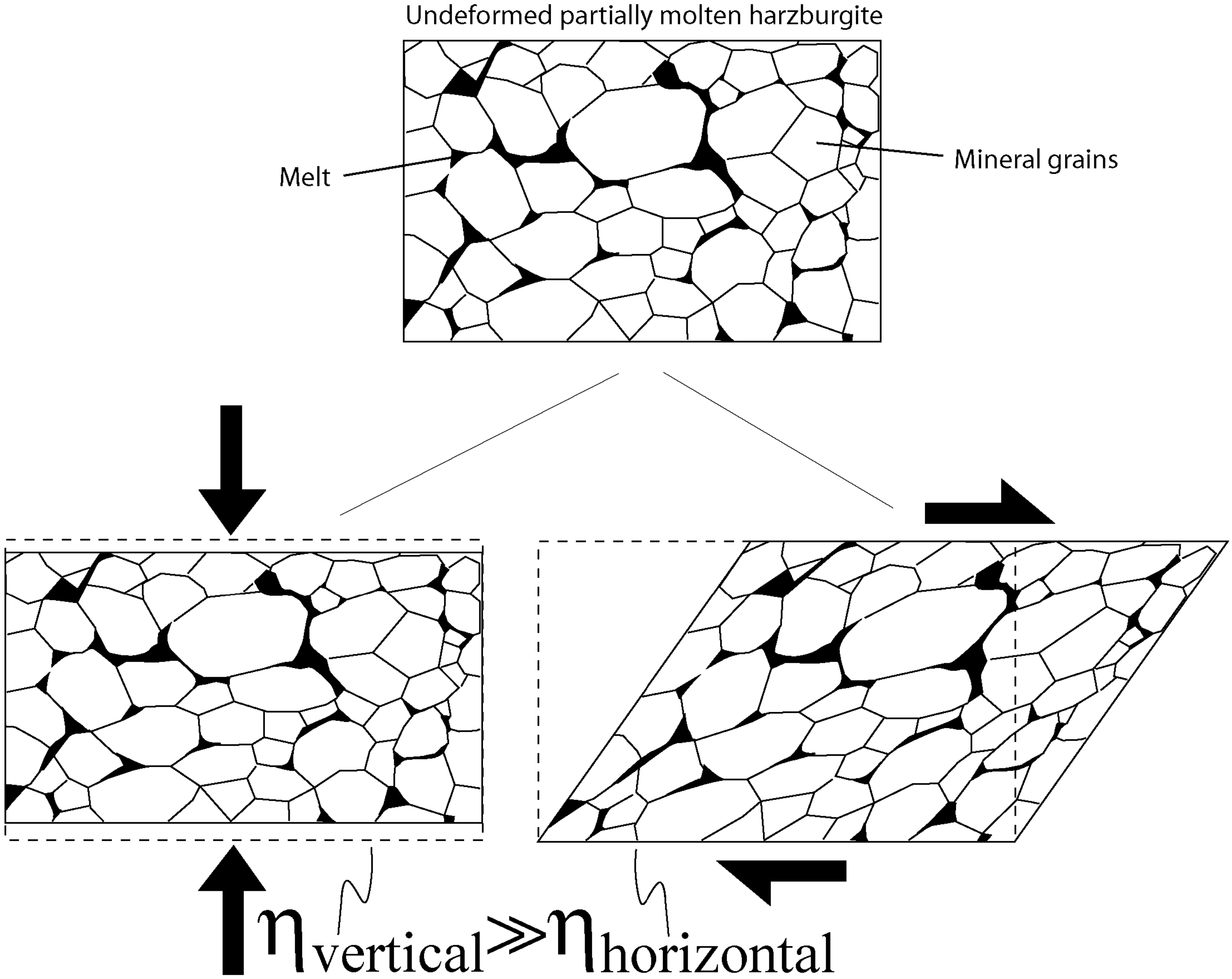
Figure 5. The Earth's outer shells undergo a pulsating semidiurnal oscillation of a few tens of cm due to solid Earth tides. This could result in a mechanical fatigue acting on the asthenosphere, allowing the westerly trending decoupling of the lithosphere.

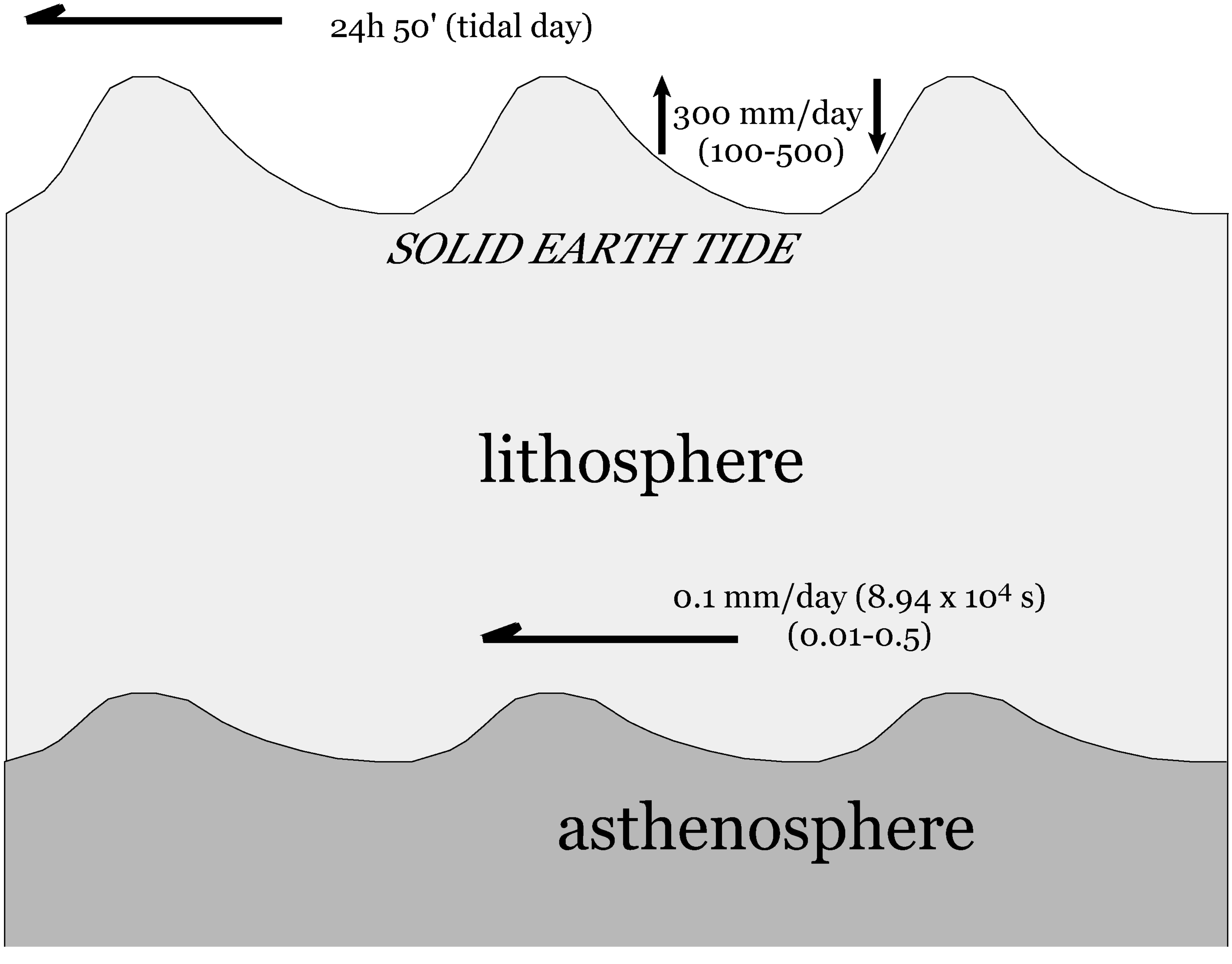

"westward drift of the lithosphere" = (ω2−ω1)
Braking exerted by tidal drag slowing down Earth's rotation
ω1
ω2
LITHOSPHERE
UPPER MANTLE
-ω1
EARTH'S ROTATION
ω
LOWER MANTLE
2890 km
3000°-4000°C
MANTLE & CORE
+ω2
SOLID INNER CORE
670 km
5150 km
LIQUID OUTER CORE
Moon
ROTATION
W-ward torque tidal-related
E-ward torque convection-related
lithosphere
Deepening of denser material increasing Earth's rotation speed



SUBDUCTION ZONES
RIFT ZONES
15 Ma
"EAST" subduction
"WEST" subduction
Elevation (km)
Distance (km)
9
-9
-800
-400
400
800
1
2
25 Ma
3
4
35 Ma
5
6
45 Ma
7
8
90 Ma
9
10
2.5
4
km
0
400
RIFT EAST FLANK
RIFT WEST FLANK
8
6
1
10
3
9
7
5
4
2


0
50
100
150
200
Depth (km)
12
14
16
18
20
22
log η (Pa s)
Asthenosphere


Undeformed partially molten harzburgite
Melt
Mineral grains
$\eta_{vertical} \gg \eta_{horizontal}$


24h 50' (tidal day)
300 mm/day
(100-500)
SOLID EARTH TIDE
lithosphere
0.1 mm/day (8.94 x 10⁴ s)
(0.01-0.5)
asthenosphere