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Asymmetric Earth: mechanisms of plate tectonics and earthquakes*

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1. Abstract

The Earth is subject to the secular cooling of a heterogeneous and stratified mantle, and to astronomical tuning. The combination of these two parameters is here inferred as the main controlling factors of plate tectonics. The planet is still hot enough to maintain at about 100 km depth a layer where partial melting determines a low–velocity and low–viscosity layer at the top of the asthenosphere, allowing partial decoupling of the lithosphere with respect to the underlying mantle. The Earth's rotation and the tidal despinning generate a torque acting on the lithosphere, and producing a net westerly directed rotation of the lithosphere with respect to the underlying mantle, being this rotation decoupled in the low–velocity asthenospheric layer. This polarization controls a diffuse asymmetry along plate boundaries, which are shaped by the "eastward" relative mantle flow with respect to the overlying lithosphere. Velocity gradients are the by–product of the lateral viscosity variations in the low–velocity layer, i.e., the decollement plane. The lower the viscosity, the faster westward motion of the overlying lithosphere (e.g., the Pacific plate). Velocity gradients among plates determine tectonics at plate margins and related seismicity. The horizontal component of the solid Earth's tide pushes plates to the "west". When faults arrive at the critical state, the vertical component of tides may trigger the earthquake due to variations of *g*. The brittle–ductile transition is inferred to act as a switch for earthquakes. During the interseismic stage, a dilatational band forms in the brittle upper crust. The stretch-

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ing is partially recovered during the coseismic stage when the fault hangingwall falls down releasing its gravitational energy along a normal fault. On the contrary, for a thrust fault, during the interseismic stage an over compressed band forms above the brittle–ductile transition, which is opposingly dilated during the earthquake, delivering the elastic energy accumulated in the hangingwall. Therefore, energy storage is different and gravity acts with opposed versus as a function of the tectonic style. Fluids react accordingly, pre, syn and post the earthquake.

2. Introduction

The origin of plate tectonics and the mechanisms governing the Earth's geodynamics are still under debate. The most accepted model for the dynamics of the planet is that tectonic plates are the surface expression of a convection system driven by the thermal gradient from the hot inner core (5500–6000 °C) and the surface of the Earth, being the shallowest about 100 km thick the upper thermal boundary layer, a <1300 °C internally not convecting layer called lithosphere. However, during the last decades it has been shown that 1) mantle convection driven from below cannot explain the surface kinematics; 2) plates do not move randomly as required by a simple Rayleigh–Bénard convective system, but rather follow a mainstream of motion; 3) plate boundaries rather show asymmetric characters. Moreover, convection models are generally modelled as deforming a compositionally homogeneous mantle, whereas the Earth is chemically stratified. Based on these assumptions, it is here reviewed an alternative model of plate tectonics, in which the internal heat still maintains possible mantle convection, but this is mostly dominated from above (Anderson, 2001), being the lithosphere active in the process, but sheared by astronomically forces. Moreover, the mantle is assumed not chemically homogeneous, and tomographic images are not indication of cold and hot volumes, but also the result of chemical heterogeneity (Anderson, 2006; Thybo, 2006; Tackley, 2000; Trampert *et al.*, 2004).

3. Tectonic equator and westward drift

The analysis of past and present motion of plates point for a non random distribution, but rather a coherent flow (Fig. 1). Past motion can be reconstructed based on magnetic anomalies in the oceanic crust and paleo–direction of shortening in the orogens associated to subduction zones. Present motions can be described by the space geodesy data, i.e., the GPS, VLBI (Very Long Baseline Interferometry) and SLR (Satellite Laser Ranging). The mainstream can be summarized by the great circle describing plate motions, defined as the "tectonic equator" (Doglioni, 1993; Crespi *et al.*, 2007). It is worth noting that past movements coincide with present day directions of motion (DeMets *et al.*, 2010; Argus *et al.*, 2011). During their journey, some plates may experience a further internal rotation. Since the motion of a plate on a sphere is defined by a rotation, this extra movement has been defined as plate sub–rotation (Cuffaro *et al.*, 2008), which disturbs the first mainstream of plates. Sub–rotations imply that plates have a first

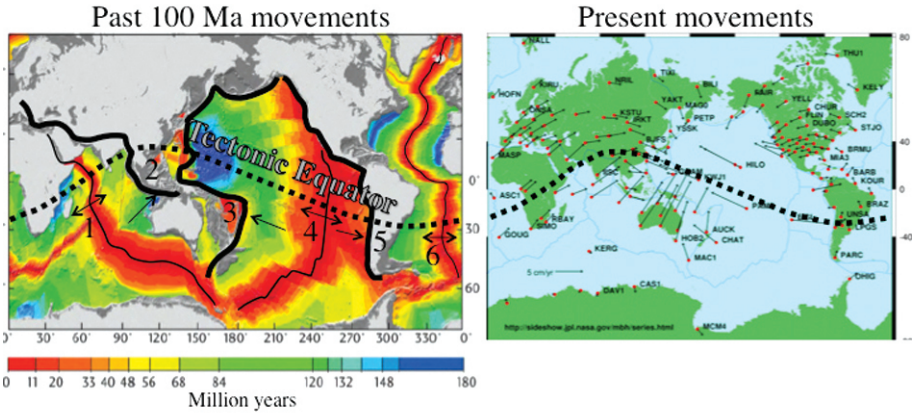

Figure 1. The direction of plate motions across the six major plate boundaries of the Earth (1, Indian ridge; 2, Alpine–Himalayan–Indonesia subduction zones; 3, Western Pacific subduction zone; 4, East Pacific Rise; 5, Andean subduction zone; 6, Mid–Atlantic ridge) shows that lithospheric plates moved along a mainstream which can be summarized by its "tectonic equator". This flow appears persistent both in the geologic past (left) and in the present day motions visible on the GPS Nasa data base (right).

order pole of rotation, plus a second pole of the sub–rotation that is the only point of the plate maintaining the same distance with respect to the first pole. The tectonic equator appears stable through time (at least 100 Myr), and makes an angle of about 30° relative to the geographic equator. The angle of the ecliptic plane plus the angle of the Moon's revolution sum to a very close value. Moreover, plates move faster at low latitudes, as recorded by kinematic data, and by seismicity, which decreases toward the polar areas (Riguzzi *et al.*, 2010; Varga *et al.*, 2012). Therefore, the shape of the tectonic equator and angle suggest an astronomical tuning of plate motions. The tectonic equator could be interpreted as the plane located at interception of the cone described by the precession of the Earth's axis. In fact, the Maxwell time (the time requested for a solid material to flow) of the lithosphere has the same order of magnitude of the precession cycles (20–26 kyr), as proposed in Fig. 2.

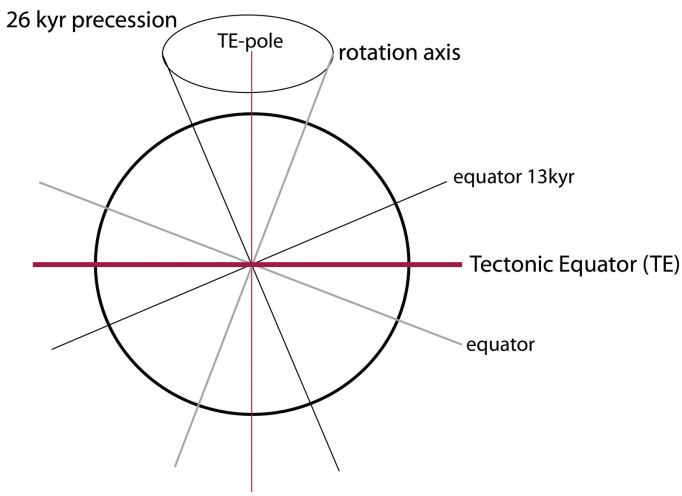

Figure 2. The tectonic equator describes the mean faster motion of the lithospheric plates. The tectonic equator has an angle of about 28–30°, which is here considered as the angle of the ecliptic plane (23°) plus the Moon's revolution plane (5°) with respect to the geographic equator. The location of the tectonic equator is inferred as the mean direction of the Earth's precession axis. The Maxwell relaxation time of the lithosphere (viscosity/rigidity ratio, 10^{22}Pa s/10^{11}Pa=10^{11}s) is of the same order of magnitude of the 26 kyr precession. This could explain the stability of the tectonic equator during the geologic past (Nesi *et al.*, in prep.).

Besides this first simple kinematic description, plates move relative to the mantle. They are decoupled at a depth varying between 60 and 200 km (average 100 km), in the low–velocity layer, where seismic waves slow down because of the presence of some percentage of melt in the mantle peridotites (Green *et al.*, 2010; Hirschmann, 2010; Anderson, 2011; Schmerr, 2012). This layer, named LVZ (Low Velocity Zone), is at the base of the lithosphere, and at the top of the asthenosphere. It is the basic decollement for plate tectonics. The overlying lithosphere acts as an insulator for the heat dissipated by the underlying mantle due to the pristine heat from the Earth's early stage of magma ocean, plus the heat delivered by radiogenic decay from the internal mantle itself. Moreover, in the LVZ has been inferred a larger amount of H_2O delivered by the pargasite mineral when located at pressure greater than 3 GPa, >100 km (Green *et al.*, 2010). Therefore, the LVZ is a layer in which the viscosity is extremely depressed and could reach values as low as 10^{12} Pa s (Jin *et al.*, 1994). Moreover, the top of the asthenosphere is in superadiabatic condition, having a potential temperature possibly higher than the underlying mantle, which has rather been inferred to be subadiabatic and less convecting (Anderson, 2013).

When measuring plate motions relative to the mantle, it results that plates have a mean "westerly" directed component, also defined as westward drift of the lithosphere or net rotation. The resulting velocity of this rotation depends on the used reference frame. If the source of the magmatism describing linear features at the Earth's surface that allow to compute the relative motion of the lithosphere relative to the mantle are located below the decoupling layer, then the net rotation amounts to 0.2–0.4° Myr (Gripp and Gordon, 2002; Conrad and Behn, 2010; Torsvik *et al.*, 2010). However, volcanoes may be sourced from within the decoupling LVZ layer. Therefore, part of the decoupling recorded by the volcanic tracks may be undetected. Since more and more petrologic and geophysical data support a shallow intra–asthenospheric origin of volcanic plumes (e.g., Bonatti, 1990; Foulger *et al.*, 2005; Anderson, 2011; Presnall and Gudfinnsson, 2011), the net rotation of the lithosphere can be much faster, in the order of >1°Myr (Crespi *et al.*, 2007, Cuffaro and Doglioni, 2007) as shown in Fig. 3. In this plate motions reconstruction, all plates move "westerly", along an undulate flow, and the fastest plates are those having at their base a LVZ with lower viscosity (e.g., the Pacific plate, Pollitz *et al.*,

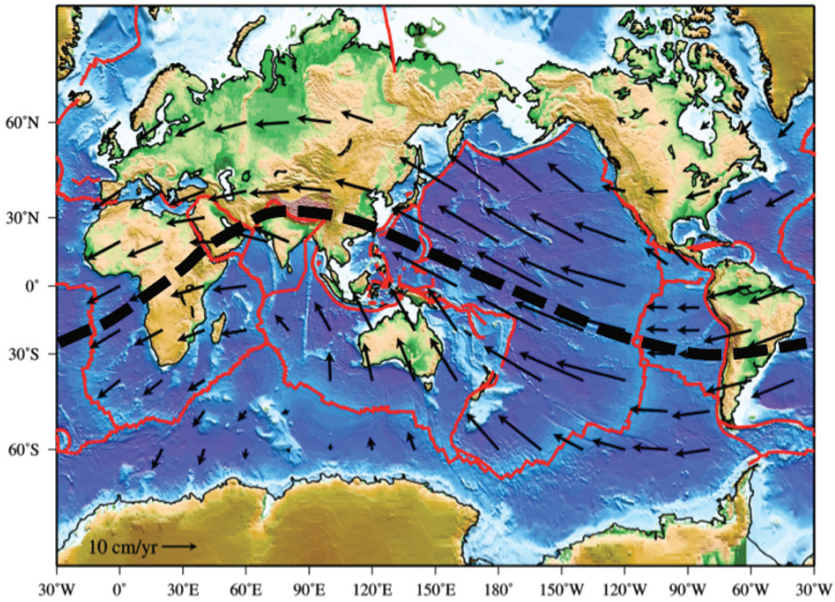

Figure 3. Model of plate motions relative to the mantle assuming an intra–asthenospheric source of volcanic trails (after Cuffaro and Doglioni, 2007). The lithosphere has a net rotation toward the "west", with an equator of this rotation that closely overlaps the tectonic equator.

1998). Therefore, plate motions are polarized toward the west, and the equator of the westward drift (Fig. 3), largely overlaps the shape of the tectonic equator (Fig. 1).

During the last years it was proposed that the misalignment of the tidal bulge relative to the gravitational Earth–Moon alignment due to the anelastic reaction of the Earth with respect to the Moon gravity field (Fig. 4) could be responsible for the westerly directed torque acting on the lithosphere (Scoppola *et al.*, 2006; Riguzzi *et al.*, 2010). A permament oscillation acting on two layers with a difference in viscosity of about 8–10 orders of magnitude would allow the decoupling between the lithosphere and the asthenosphere (Doglioni *et al.*, 2011). In the decoupling layer, the mantle is undergoing shearing and crystals may roll due to the intra–crystalline melt, lowering the viscosity when measured parallel to the horizontal shear (Fig. 5). If a segment of lithosphere is moving westward faster relative to another segment to the east, a lower viscosity can be inferred within the LVZ beneath that

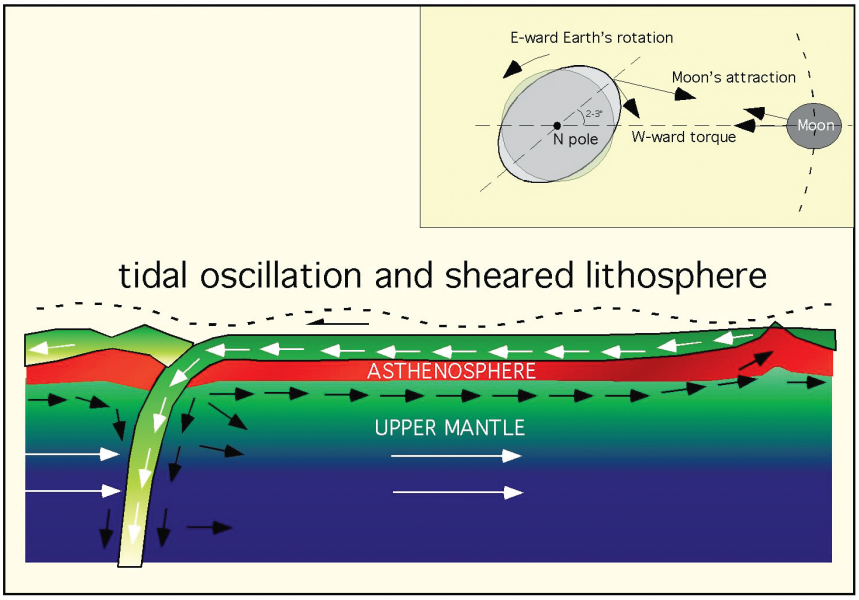

Figure 4. The westward drift of the lithosphere relative to the mantle can be inferred as generated by the torque exerted by the misalignment of the excess of mass of the Earth's bulge with respect to the Earth–Moon gravitational alignment. This momentum should be allowed by the presence of a hot, low–viscosity layer at the lithosphere base, which is detected as the low–velocity layer where seismic waves slow down in the asthenosphere. The tidal oscillation of the lithosphere contributes to activate the decoupling.

plate, generating a rifting in the overlying lithosphere. The spreading is transferred up toward the surface into the brittle regime where most of the faults and related earthquake develop. In case of faster plate to the east relative to a plate to the west, a subduction and related orogeny will occur in between.

The westward drift of the lithosphere (Bostrom, 1971; Knopoff and Leeds, 1972; Moore, 1973) generated a number of asymmetries at plate boundaries.

W–directed subduction zones are steeper (average 65°) than those directed to E or NE (average 27°), and the associated orogens are respectively characterized by lower structural and topographic elevation, occurrence of backarc basins, whereas E or NE–directed subductions show higher structural and morphological elevation and absent backarc basin (Fig. 6). This asymmetry is also marked by the opposite

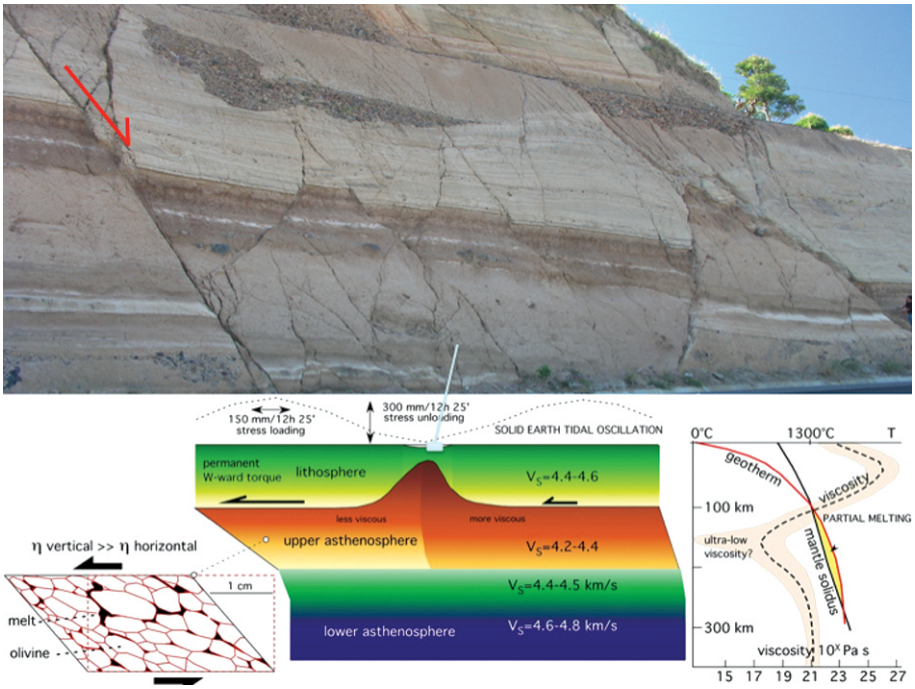

Figure 5. Lateral variations of viscosity in the asthenosphere determine different velocity among plates, hence generating plate boundaries where the lithosphere is either spread apart producing a rift, or converging along subduction zones. Differential velocity is transferred upward generating surface tectonics like the graben of the picture and seismicity.

state of stress of the slabs: down–dip compression and down–dip extension for W and E or NE–directed slabs respectively (Doglioni *et al.*, 2007; Carminati and Petricca, 2010; Riguzzi *et al.*, 2010). These asymmetries are striking when western and eastern Pacific subduction zones are compared. Such differences have usually been interpreted as related to the age of the downgoing oceanic lithosphere (usually older, cooler and denser in the western side). However these differences persist elsewhere, regardless of the age and composition of the downgoing lithosphere, e.g., in the Mediterranean Apennines and Carpathians Vs. Alps and Dinarides, or in the Banda and Sandwich arcs, where even continental or zero–age oceanic (mid–oceanic ridge) lithosphere is almost vertical along W–directed subduction zones (Cruciani *et al.*, 2005; Doglioni *et al.*, 2007). W–directed subductions

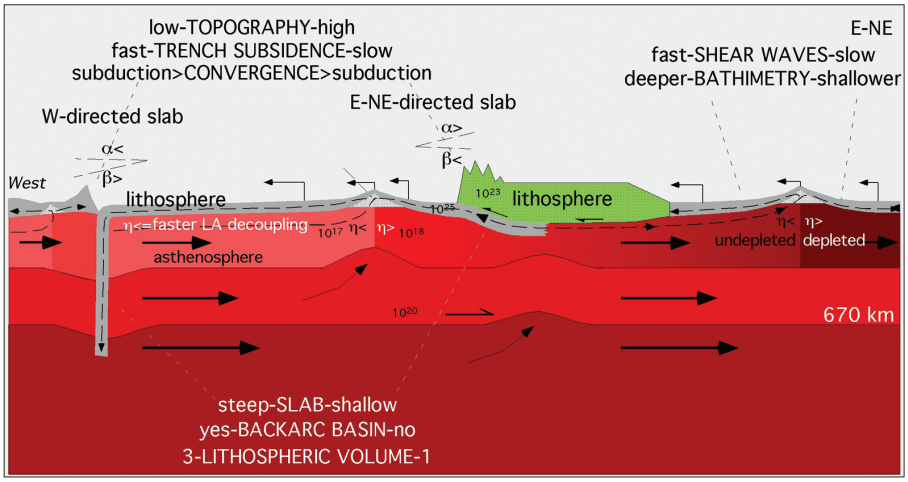

Figure 6. The "westward" drift of the lithosphere relative to the mantle produces a number of asymmetries at plate boundaries, both along subduction and rift zones. For example, slabs are steeper along W–directed subduction zones and are always associated to backarc spreading. E– or NE–directed subduction zones are rather shallower and the overlying orogens show much higher topography and structural elevation with respect to the opposite slabs. Moreover they are double verging orogens, they have shallow foredeeps and wider outcrops of metamorphic rocks. The subduction hinge generally diverges relative to the upper plates along the W–directed slabs, whereas it converges along the opposite E– or NE–directed slabs. Grey layer, oceanic lithosphere; green, continental lithosphere. Legend: η, viscosity values. α and β, dip of the orogens envelope and regional foreland monocline respectively (after Carminati and Doglioni, 2012).

and E or NE–directed have subduction hinges respectively diverging and converging relative to the upper plate, fast versus low subduction rates, low versus high topographic envelopes (α) and high versus low foreland monoclines (β) (Doglioni *et al.*, 2007). Finally the lithospheric volume subducted is three times bigger in W–directed subductions (Doglioni *et al.*, 2007, 2009). The archetypes of the opposite subduction zones can be recognized when comparing Alps (E–SE–directed slab) and Apennines (W–directed), (see Panza *et al.*, 2007; Carminati and Doglioni, 2012; Carminati *et al.* 2012).

Oceanic rifts are also asymmetric, with east flanks more elevated of about 100–300 m worldwide (Doglioni *et al.*, 2003) (Fig. 5). Moreover, along a cross section across the Mid Atlantic Ridge (MAR), Panza *et al.*

(2010) (Fig. 6) showed that shear waves velocities (V_s) suggest a thicker western limb of the MAR and a thinner eastern limb. These authors also emphasized the occurrence of a low velocity zone (LVZ) between the lithospheric mantle (LID) and the upper asthenosphere of about 200 km depth. The asymmetry of the limb thickness at oceanic rift zones was also recently reproduced by Cuffaro and Miglio (2012), modelling mantle upwelling beneath a migrating Mid Atlantic ridge, using absolute velocities, as boundary conditions.

All these asymmetries are compatible with the westward drift of the lithosphere >1°Myr and can be envisaged moving along the flow lines of plate motions described by the tectonic equator. Therefore we need a dynamic model to explain this polarized system. Solid Earth's tides (or body tides) generate significant and often disregarded oscillations in the lithosphere. They uplift the ground of 20–40 cm, swinging back and forth horizontally 10–20 cm at every passage of the Moon and Sun (0.46 of the Moon tide) gravitational waves. Therefore the lithosphere is constantly subject to a vibration, which is westerly oriented due to the misalignment of the Earth's bulge. We infer that after the passage of the body tide, a residual permanent shear remains, like a shifting plastic bottle subject to the wind at sea. The lithosphere appears to behave like a worm, i.e., it is uplifted and sheared horizontally under the effects of the solid Earth's tide (Fig. 7). A hysteresis of about 0.15 mm every Moon passage (12 hours and 25') could explain a lithospheric drift of 5.4 cm/yr relative to the mantle.

4. Seismicity

Most of the energy delivered by the Earth's seismicity (>90%) is dissipated in the first 100 km and by earthquakes with magnitude >6.9. Seismicity is the result of plate tectonics, and the understanding of the forces governing geodynamics may provide fundamental clues for unravelling earthquakes. Therefore, the mechanisms that move plates are of paramount importance for tackling the issue of the seismic hazard. The most energetic earthquakes form along subduction zones because they are cold zones, and breaking rocks under contraction needs more energy than under extension. Recently it has been observed some correlation between seismicity and tides (Wilcock, 2001;

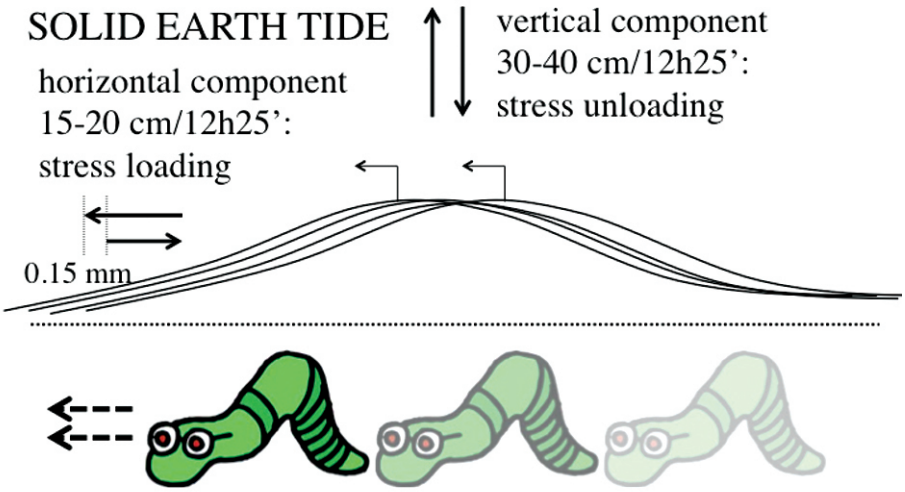

Figure 7. The solid Earth's tide generates a wave, which uplifts and shifts horizontally the lithosphere. A hysteresis of about 0.15 mm (variable with latitude) should accumulate after every solid Earth's passage, due to the westerly–directed torque acting on the lithosphere. At the end of the year, this sums up to the few centimetres of the lithosphere relative to the mantle. The horizontal component of tides is inferred as the pumping system at plate boundaries where faults are gradually loaded of energy over the centuries. The vertical oscillation rather modifies the acceleration of gravity, and is interpreted as the triggering mechanism of earthquakes when a fault has reached its critical state. The model proposes that the westward drift deforms the lithosphere as a worm, which is crawling.

Cochran *et al.*, 2004; Métivier *et al.*, 2009). For example, along the Cascadia subduction zone in the western North America, it has been noted that tremors are correlated during the high tide. Conversely, normal fault–related earthquakes seem more frequent during the low tide (Fig. 8). The tidal forces are too small to generate earthquakes, and for this reason they have been disregarded for long time. However, we have seen that the tide has a horizontal and a vertical component. The horizontal component can pump the system and load of energy a given fault which is locked. When the critical state is reached, even a small change in the lithostatic load can trigger the earthquake. For example, during the high tide, the gravity is at minimum, whereas during the low tide, the Earth's gravity is at its maximum. Since the lithostatic load ϱgz (being ϱ the density of rocks, g, the acceler-

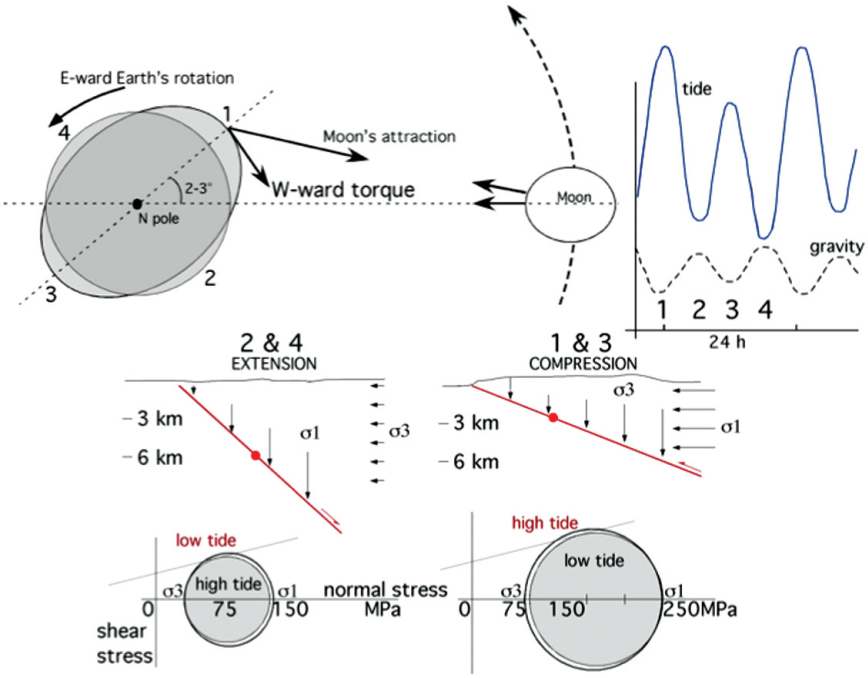

Figure 8. The passage of the solid Earth's tides determines oscillations of the acceleration of gravity (g). High tide corresponds to low gravity, whereas a low tide is associated to a high gravity field. Gravity oscillations generate variations of the lithostatic load (given by ϱgz, i.e., density of rocks, acceleration of gravity and thickness of rocks). In extensional tectonic settings, the increase of g (low tide) increases the differential stress and facilitates failure of the rocks. In compressional tectonic settings instead, the decrease of g (high tide) increases the differential stress, approaching the failure of rocks (modified after Riguzzi *et al.*, 2010).

ation of gravity and z the thickness of the rocks column) acts as the minimum stress tensor in compressional tectonic settings and the maximum stress tensor in extensional tectonic settings, their variation determines opposite effects on the activation of faults. The differential stress increases during the high tide in compressional regimes (i.e., the lithostatic load decreases for a smaller g), thus favouring earthquake nucleation. Alternatively, the differential stress increases during the low tide in extensional regimes (i.e., the lithostatic load increases for a higher g). Therefore, high tide favours fault slip in compressional environment, whereas the low tide is rather positive for normal faulting, which is exactly what observed (Fig. 8). For these reasons, the

two tidal components could be considered as potentially relevant for earthquakes, being the horizontal one providing the energy during the centuries, and the vertical one acting as the final drop that generates the instability of the fault at the coseismic stage.

Within the crust, pressure and temperature increase with depth. However, pressure increases the stability of rocks, whereas temperature has the opposite effect. In a crust of about 30 km thickness, the first 15 km are dominated by the effect of pressure and rocks behave in a brittle way. The deeper 15–30 km are rather controlled by the temperature effect, which is increasing with depth, decreasing the energy required for deforming plastically the rocks. Therefore, the highest energy release by earthquakes occurs near the brittle–ductile transition (BDT), and the main seismic swarm and aftershocks are affecting the brittle upper crust (Fig. 9). In the lower crust, faults (or more exactly shear zones) may be active in a steady state regime, whereas in the upper crust, faults are locked and they move only episodically (stick–slip behaviour).

These basic observations allow us to infer that the BDT plays an important role in earthquakes. In fact, during the interseismic period,

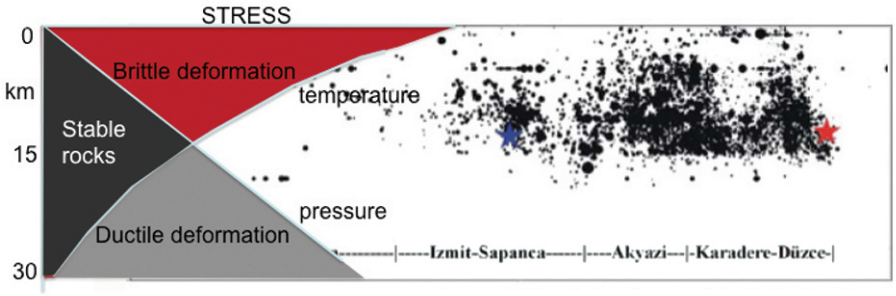

Figure 9. The crust is generally divided into at least two layers, due to the interference of pressure and temperature stability field of rocks. In the upper crust, rocks behave in a brittle regime where pressure dominates, and the deformation is episodic; in the lower crust, the temperature is instead controlling the rheological behaviour of rocks, which rather deform in a steady state plastic–ductile regime. The transition between the upper brittle and lower ductile regimes represents the main basal boundary of seismicity. The stars indicate the hypocentres of the Izmit 1999 earthquake sequence. The black dots are the aftershocks mostly confined in the upper brittle crust.

at the BDT should form a dilated band along extensional fault, and an overcompressed band along a contractional fault (thrust or reverse fault). During the coseismic stage, the bands reverse their behaviour, being the dilated band recompressed, and the overcompressed band rather stretched at the seismic event (Fig. 10). Fluids should react differently as a function of the opposite tectonic styles and in the future they may provide possible seismic precursors.

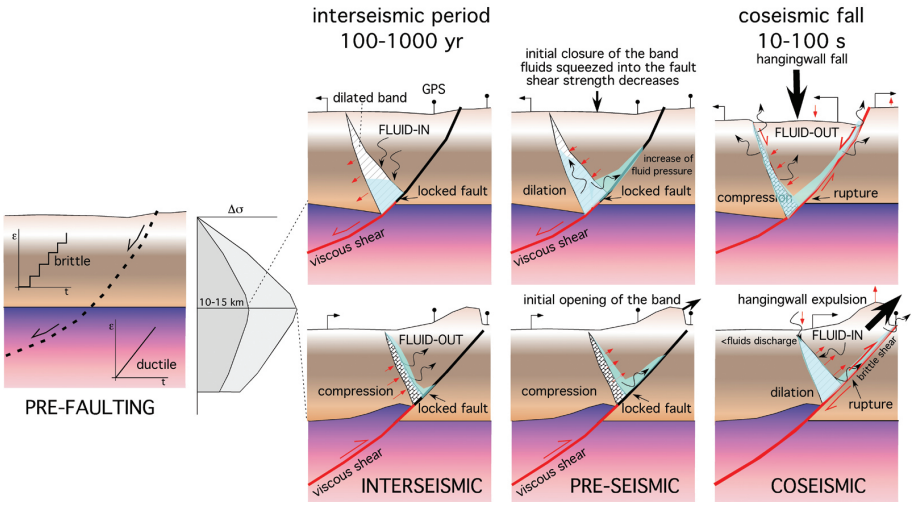

Figure 10. Assuming a simplified two–layer crustal rheology, the ductile lower crust is characterized by a constant strain rate, whereas the brittle upper crust displays episodic locking–unlocking behaviour. Tensional and compressional faults generate opposite kinematics and mechanical evolutions. In the tensional tectonic environment, the triangle of crust above the BDT remains "suspended" while a dilated area forms during the interseismic period. Fluids may enter the fractured volume. Once shear stress along the locked part of the fault becomes larger than fault strength; the hangingwall will begin to collapse, increasing the fluid pressure into the fault gouge. This mechanism is self–supporting because it decreases fault strength, facilitating the final fall of the hangingwall generating the mainshock. Conversely, during the interseismic period, along a thrust plane an over–compressed band separates the ductile shear from the overlying locked fault segment. The hangingwall is eventually expelled during the coseismic period. Fluids discharge behaves differently as a function of the tectonic field (after Doglioni *et al.*, 2013).

5. Conclusions

The Earth's models are generally performed assuming internal dynamics controlled by lateral density variations due to thermal gradients, which should generate both vertical and horizontal movements in the mantle and in the overlying lithospheric plates. The analysis of the geometry at plate boundaries, and the kinematic constraints rather point for a "westerly" polarized flow of plates, which implies a relative opposed movement of the underlying Earth's mantle. This flow determines an asymmetric pattern along subduction zones and two end members of orogens associated to the downgoing slabs, such as the differences between the low topography in the hangingwall of the western Pacific subduction zones, versus the high topography and deep rocks exhumation along the eastern Pacific subduction zones. Rift zones are also asymmetric, having in average the western flank few hundreds meters deeper, and a thicker and faster lithosphere with respect to the eastern flank. All plate boundaries move "west". The decoupling of the lithosphere in the low–velocity zone (LVZ) at about 100–200 km depth allows the "westerly" directed rotation of the lithosphere with respect to the underlying mantle (Fig. 5). This asymmetric movement determines larger volumes of lithosphere returned to the mantle with respect to the opposed subduction zones, allowing some constraints for the reconstruction of the convection pattern in the upper mantle (Fig. 6). The motion of the lithosphere is inferred as the combination of an astronomically driven "westerly" directed horizontal shear acting on plates, combined with the heat dissipation of the planet that maintains the temperature high enough in the LVZ to have the low viscosity required for the lithospheric decoupling. The variations in viscosity within the LVZ control variations in velocity among plates, and eventually seismicity at plate boundaries.

The solid Earth's tide (Fig. 7) may be crucial in understanding geodynamics and earthquakes. The horizontal component, which is westerly polarized due to the misalignment of the Earth's bulge, may provide the source for shearing the lithosphere and loading the faults. The vertical component generates oscillations of the lithostatic load, which may locally trigger earthquakes along fault planes at the critical state (Fig. 8). However, the effect of tides on faults is opposed as a function of the tectonic style. When a fault moves from the interseis-

mic to the coseismic activation, the stress field that characterized the long lasting inactive period, gradually or instantaneously inverts the porosity (either increasing or decreasing it). Therefore, fluids react accordingly, and they might become in the future important precursors of imminent earthquakes (Fig. 10).

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