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ScienceDirect

Earth-Science Reviews xx (2007) xxx–xxx

EARTH-SCIENCE REVIEWS

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Subduction kinematics and dynamic constraints

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Received 29 August 2006; accepted 19 April 2007

Abstract

The kinematics of subduction zones shows a variety of settings that can provide clues for dynamic understandings. Two reference frames are used here to describe the simple 2D kinematics of subduction zones. In the first, the upper plate is assumed fixed, whereas in the second frame upper and lower plates move relative to the mantle.

Relative to a fixed point in the upper plate U, the transient subduction hinge H can converge, diverge, or be stationary. Similarly, the lower plate L can converge, diverge or be stationary. The subduction rate V_S is given by the velocity of the hinge H minus the velocity of the lower plate L ($V_S = V_H - V_L$). The subduction rate 1) increases when H diverges, and 2) decreases when H converges.

Combining the different movements, at least 14 kinematic settings can be distinguished along the subduction zones. Variable settings can coexist even along a single subduction zone, as shown for the 5 different cases occurring along the Apennines subduction zone. Apart from few exceptions, the subduction hinge converges toward the upper plate more frequently along E- or NE-directed subduction zone, whereas it mainly diverges from the upper plate along W-directed subduction zones accompanying backarc extension.

Before collision, orogen growth occurs mostly at the expenses of the upper plate shortening along E–NE-directed subduction zones, whereas the accretionary prism of W-directed subduction zones increases at the expenses of the shallow layers of the lower plate.

Backarc spreading, forms in two settings: along the W-directed subduction zones it is determined by the hinge divergence relative to the upper plate, minus the volume of the accretionary prism, or, in case of scarce or no accretion, minus the volume of the asthenospheric intrusion at the subduction hinge. Since the volume of the accretionary prism is proportional to the depth of the decollement plane, the backarc rifting is inversely proportional to the depth of the decollement. On the other hand, along E- or NE-directed subduction zones, few backarc basins form (e.g., Aegean, Andaman) and can be explained by the velocity gradient within the hangingwall lithosphere, separated into two plates.

When referring to the mantle, the kinematics of subduction zones can be computed either in the deep or in the shallow hotspot reference frames. The subduction hinge is mostly stationary being the slab anchored to the mantle along W-directed subduction zones, whereas it moves W- or SW-ward along E- or NE-directed subduction zones. Surprisingly, along E- or NE-directed subduction zones, the slab moves “out” of the mantle, i.e., the slab slips relative to the mantle opposite to the subduction direction. Kinematically, this subduction occurs because the upper plate overrides the lower plate, pushing it down into the mantle. As an example, the Hellenic slab moves out relative to the mantle, i.e., SW-ward, opposite to its subduction direction, both in the deep and shallow hotspot reference frames. In the shallow hotspot reference frame, upper and lower plates move “westward” relative to the mantle along all subduction zones.

This kinematic observation casts serious doubts on the slab negative buoyancy as the primary driving mechanism of subduction and plate motions.

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doi:10.1016/j.earscirev.2007.04.001

W-directed subduction zones rather provide about 2–3 times larger volumes of lithosphere re-entering into the mantle, and the slab is pushed down. This opposite behavior is consistent with the down-dip extension seismicity along E–NE-directed subduction zones, and the frequent down-dip compression along the W-directed subduction zones.

Subduction kinematics shows that plate velocity is not dictated by the rate of subduction. Along the W-directed subduction zones, the rate of subduction is rather controlled i) by the hinge migration due to the slab interaction with the “easterly” trending horizontal mantle wind along the global tectonic mainstream, ii) by the far field plate velocities, and, iii) by the value of negative buoyancy of the slab relative to the country mantle.

Alternatively, E–NE–NNE-directed subduction zones have rates of sinking chiefly determined i) by the far field velocity of plates, and ii) by the value of negative buoyancy of the slab relative to the country mantle. Along this type of subduction, the subduction hinge generally advances E–NE-ward toward the upper plate decreasing the subduction rate, but it moves W–SW-ward relative to the mantle.

All this indicates that subduction zones have different origin as a function of their geographic polarity, and the subduction process is more a passive feature rather than being the driving mechanism of plate motions. A rotational component combined with mantle density and viscosity anisotropies seems more plausible for generating the global tuning in the asymmetry of subduction zones.

Keywords: subduction; kinematics; hinge migration; remounting slab; westward drift

1. Introduction

The subduction process (Fig. 1) is a vital ingredient of plate tectonics (Anderson, 1989, 2001; Stern, 2002; Conrad and Lithgow-Bertelloni, 2003; Anderson, 2006). By definition, along subduction zones, the lithosphere enters into the underlying mantle (e.g., Jarrard, 1986; Turcotte and Schubert, 2002). Most of the oceanic lithosphere is recycled into the mantle, but also some portions of continental lithosphere can subduct (Ampferer, 1906; Panza and Mueller, 1978; Ranalli et al., 2000; Hermann, 2002), usually termed as the collisional stage of the process. There still are debates on the initiation of the subduction process (Spence, 1987; Niu et al., 2003), but once activated it may work for several tens of million years. Subduction zones have a related accretionary prism, or orogen, a subduction hinge that can migrate (Garfunkel et al., 1986), and a large variety of geophysical, magmatological and geological signatures (Isacks and Molnar, 1971; Jarrard, 1986; Royden and Burchfiel, 1989; Tatsumi and Eggins, 1995; Hacker et al., 2003; Carminati et al., 2004a,b; Ernst, 2005).

Subduction zones may initiate when two basic parameters concur: i) the two plates have at least an initial convergent component of relative plate motion and ii) one of the two plates is sufficiently denser, thinner, stronger and wider to be overridden. As a counterexample, when an oceanic rift opens to a width equal or smaller than the thickness of the adjacent continental lithosphere, a complete subduction cannot develop (Fig. 2). In other words, small oceans (150–200 km wide) cannot generate normal, steady state, subduction systems. This setting can rather evolve to obduction, where ophiolitic slices of the oceanic realm are buckled and squeezed on top of the continental lithosphere. One example could be the Oman ophiolite complex (Nicolas et al., 2000; Garzanti et al., 2002), even if it is unclear whether this ocean was a narrow independent basin, or rather part of a wider Tethyan branch (Stampfli and Borel, 2002).

Regardless of the reason why the lithosphere moves, e.g., slab pull (Forsyth and Uyeda, 1975) or mantle drag (Bokelmann, 2002), the slab is assumed to sink at a speed linearly correlated to the convergence rate (Jarrard, 1986) and its dip (Fig. 1). It is usually assumed that the convergence rate between two plates equals the subduction rate (e.g., Ruff and Kanamori, 1980; Stein and Wysession, 2003; Fowler, 2005). However, the behavior of the subduction hinge has been shown to be a crucial parameter in subduction zones (e.g., Karig, 1971; Brooks et al., 1984; Carlson and Melia, 1984; Winter, 2001; Hamilton, 2003). For example, a subduction hinge migrating away from the upper plate, i.e. accompanying the slab retreat or slab rollback, generates backarc spreading (e.g., Brooks et al., 1984; Waschbusch and

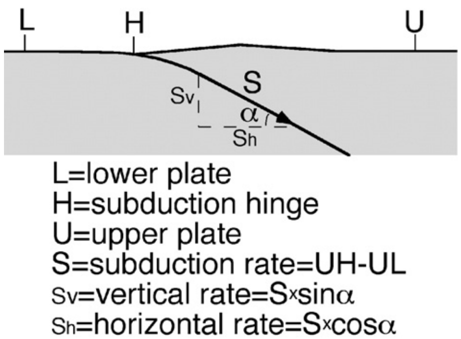

Fig. 1. Basic kinematic terminology for a subduction zone used in the text. Steeper the slab, faster will be its penetration component into the mantle.

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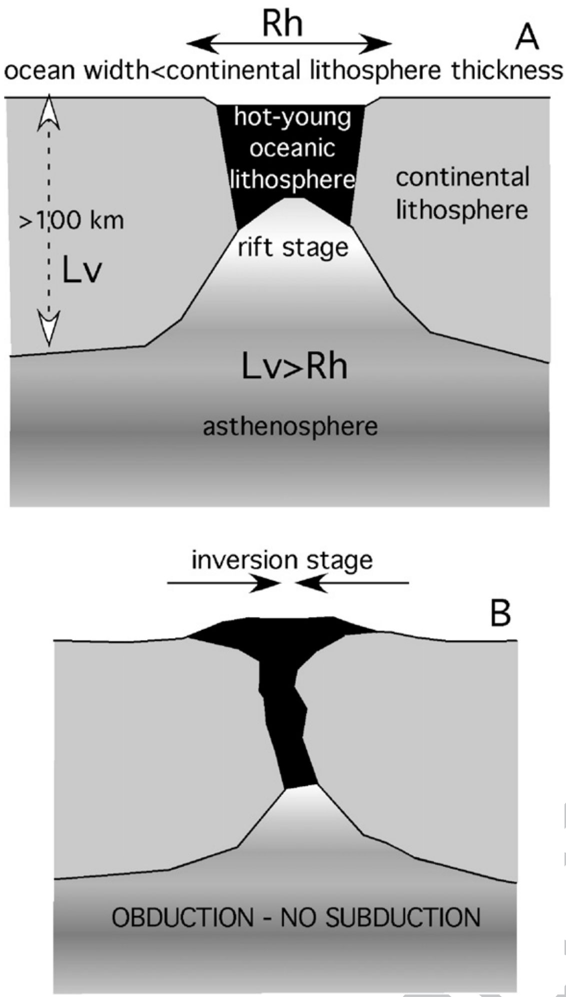

Fig. 2. An ocean basin that has a width (Rh) smaller than the thickness of the continental lithosphere (Lv) at its margins as in the upper panel A cannot evolve to a complete subduction, lower panel B. In this setting the ophiolites can be more easily obducted.

Beaumont, 1996; Heuret and Lallemand, 2005; Lallemand et al., 2005). Backarc basins prevail along the western Pacific supporting a geographic asymmetry of subduction zones (Dickinson, 1978; Uyeda and Kanamori, 1979; Doglioni et al., 1999a).

We contribute here with a simple 2D kinematic analysis of subduction zones moving the lower plate and the subduction hinge relative to the upper plate or, alternatively, moving all three points relative to the mantle. It will be shown that the subduction rate is strongly dependent on the rate and migration direction of the subduction hinge that can either converge or diverge relative to the upper plate, decelerating or accelerating the speed of subduction respectively, and a variety of different settings can be identified.

The model is tested with the GPS NASA data set (Heflin et al., 2005) since space geodesy data of plate motions have been shown to be equivalent to the velocities computed over the geologic time, e.g., through magnetic anomalies in the NUVEL-1 (Gordon and Stein, 1992; Robbins et al., 1993).

Plate kinematics is also analyzed with respect to the deep or shallow hotspot reference frames (Gripp and Gordon, 2002; Doglioni et al., 2005b; Crespi et al., 2007; Cuffaro and Doglioni, in press), supporting the notion of a westerly-polarized undulated lithospheric flow relative to the mantle, or tectonic mainstream (Doglioni, 1990; Crespi et al., 2007). In our analysis, unexpectedly against their definition, along some subduction zones directed E- or NE-ward, the slab appears to "escape" from the mantle, moving in the opposite direction to the subduction, and claiming for alternative dynamics of the subduction process.

On the basis of the kinematic inferences, the paradigm of the slab pull is seriously questioned. A number of counterarguments to the slab pull efficiency will be posed and critically disputed, and an alternative mechanical hypothesis is forwarded. Some basic dynamic speculations on the forces determining the subduction process and plate tectonics in general will be presented.

2. Subduction zones asymmetry

Subduction zones and related orogens show significant differences as a function of their polarity. It is referred here as polarity the direction of subduction with respect to the tectonic mainstream, which is not E–W, but undulates around the Earth (Doglioni, 1990; Riguzzi et al., 2006; Crespi et al., 2007).

Therefore, the asymmetry can be recognized not simply comparing W-directed vs. E-directed subduction zones, but subductions along the sinusoidal flow of absolute plate motions that undulates from WNW in the Pacific, E–W in the Atlantic, and NE to NNE from eastern Africa, Indian Ocean and Himalayas. In the hotspot reference frame a "westward" rotation of the lithosphere can be observed (Gripp and Gordon, 2002; Cuffaro and Jurdy, 2006). The origin of this net rotation of the lithosphere (Bostrom, 1971) is still under debate (Scoppola et al., 2006), but it should range between 4.9 cm/yr (Gripp and Gordon, 2002) and 13.4 cm/yr (Crespi et al., 2007) at its equator. This implies that the plate motions flow is significantly polarized toward the "west" (Doglioni et al., 1999a), and subduction zones follow or oppose the relative "eastward" relative mantle flow. Subduction zones following the flow are: North and South America cordilleras, Dinarides, Hellenides, Caucasus, Zagros, Makran, Himalayas, Indonesia–Sunda arc, Taiwan, New Guinea, New Hebrides, southern New Zealand. Subduction zones opposing the flow are: Barbados, Sandwich, Apennines, Carpathians, Banda, Molucca, Tonga, Kermadec, Marianas, Izu–Bonin, Nankai, Philippine, Kurili, Aleutians. The Japan subduction appears as a transitional subduction zone.

Since many subduction zones have undulations or arcs along strike, their dip and strike can be oblique or parallel to the proposed tectonic mainstream. For example, in the transfer zone between Makran and Himalayas NNE-directed subduction zones, along the Chaman left-lateral transpressive system, the tectonic mainstream is about parallel to the belt. Another emblematic case is the Aleutians arc, here considered as a subduction opposing the flow (W-directed). Along their western termination, they almost parallel the Pacific subduction direction (WNW), where the slab is a right-lateral ramp of the subduction, and it dips to the NNE.

The W-directed slabs are generally very steep (up to 90°) and deep, apart few cases as Japan. They have a co-genetic backarc basin, and the related single verging accretionary prism has low elevation (e.g., Barbados, Sandwich, Nankai), is mostly composed of shallow rocks, and has a frontal deep trench or foredeep (Doglioni et al., 1999a). The E- or NE-directed subduction zones are less inclined (15–70°), and the seismicity generally dies at about 300 km, apart some deeper clusters close to the upper-lower mantle transition (Omori et al., 2004). The related orogens have high morphological and structural elevation (e.g., Andes, Himalayas, Alps), wide outcrops of basement rocks, and two shallower trenches or foredeeps at the fronts of the double verging belt, i.e., the forebelt and the retrobelt. The retrobelt decreases its development when the upper plate is subject to extension (e.g., Central America, Aegena and Andaman Seas).

Lallemand et al. (2005) argue that the slab dip is not significantly influenced by the polarity of the subduction. However their analysis is different from what was suggested in a number of alternative articles where the slab dip is measured not simply comparing E- vs. W-directed subduction zones, but is measured along the undulated flow of absolute plate motions (e.g., Doglioni et al., 1999a,b, and references therein), and the definition of W- vs. E- or NE-directed is rather related to subductions following or contrasting this flow. Moreover their analysis subdivides the slab into a shallow (<125 km) and a deeper part (>125 km). This subdivision is ambiguous for a number of reasons. The E- or NE-directed subduction zones have mostly continental lithosphere in the upper plate and the dip of the first 125 km is mostly constrained by the thickness and shape of the upper plate. Moreover, oblique or lateral subduction zones such as the Cocos plate underneath Central America are, by geometrical constraint, steeper (>50°) than frontal subduction zones (e.g., Chile), like the oblique or lateral ramp of a thrust at shallow levels in an accretionary prism is steeper than the frontal ramp. Cruciani et al. (2005) reached similar conclusions of Lallemand et al. (2005) stating no correlation between slab age and dip of the slab, but their analysis is stopped to about 250 km depth because E- or NE-directed subduction zones do not have systematic seismicity at deeper depth, apart few areas where seismicity appears concentrated between 630 and 670 km depth, close the lower boundary of the upper mantle. The origin of these deep isolated earthquakes remains obscure (e.g., mineral phase change, blob of detached slab or higher shear stress) and therefore they could not represent the simple geometric prolongation of the shallow part of the slab. Therefore, at deep levels (>250 km) the dip of the slab based on seismicity cannot be compared between W- and E–NE-directed subduction zones, simply because most of the E- or NE-directed slabs do not show continuous seismicity deeper than 250 km. High velocity bodies suggesting the presence of slabs in tomographic images often do not match slab seismicity.

Moreover, Lallemand et al. (2005) note that steeper slabs occur where the upper plate is oceanic, while shallower slabs occur where the upper plate is continental. However, the majority of E- or NE-directed subduction zones worldwide have continental lithosphere in the upper plate, confirming the asymmetry. Therefore any statistical analysis on the slab dip based on seismicity cannot be done simply comparing W- and E-directed slabs because E- or NE-directed subduction zones do not have comparably deep seismicity. Moreover subduction zones juxtapose plates of different thickness and composition generating variation in the dip. The variability of the angle of obliquity of the subduction strike with respect to the convergence direction determines a further change in dip of the slab. Removing these issues, the W-directed subduction zones when compared to E- or NE-directed slabs still maintain a number of differences, such as they are steeper, they are deeper or at least they present a more coherent slab-related seismicity from the surface down to the 670 discontinuity, and they show opposite down-dip seismicity as it will be discussed later. But, even more striking, surface geology and topography of the orogens contrast dramatically with a number of differences between the opposite subduction zones such as shallow vs. deep rocks, steep vs. shallow dip of the foreland monocline, low vs. higher inclination of the topographic and structural envelop, small vs. wider area above

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sea-level, etc., respectively for W- vs. E- or NE-directed subduction zones (e.g., Lenci and Doglioni, in press).

Moreover backarc spreading occurs mostly in W-directed subduction zones. Two counterexamples are proposed as proofs that this statement is not correct, i.e., the Aegean and the Andaman Seas, which are related to NE- to NNE-directed subduction zones. However these two cases have a different kinematics and geodynamic setting with respect to the W-directed subduction zones, where upper plate extension is concomitant to subduction hinge migrating away with respect to the upper plate, being the lithospheric deficit due to subduction, compensated by mantle replacement (e.g., Doglioni, 1991). Along the Aegean and Andaman rift zones the extension rather accommodates only the differential velocity within the upper lithospheric plate, which is split into two plates overriding the subduction (e.g., Doglioni et al., 2002). Along the Andaman–Sunda–Indonesia arc for example, the flow of plates is NE- to NNE-directed (Heflin et al., 2005). Extension or "backarc" spreading is not diffused along the entire arc, but rather concentrated at the western margin, along the transfer zone between the SW-ward faster advancing of Indonesia upper plate over the lower oceanic plate (about 60 mm/yr) with respect to the slower velocity of Eurasia upper plate in overriding the continental Indian lithosphere along the Himalayas belt (about 40 mm/yr). Where upper plate extension occurs along these settings, the retrobelt of the orogen is poorly developed and narrower. Another example seems the Central America subduction zone (e.g. Guatemala, El Salvador, Nicaragua, Costa Rica), in which the upper plate extension accommodates the faster westward motion of North America relative to South America.

Royden (1993) proposed that the asymmetry of subduction zones is related to a convergence rate faster or slower than the slab retreat. Doglioni et al. (2006a) rather suggested that the subduction system is primarily sensitive to the behavior of the subduction hinge, i.e., moving toward or away from the upper plate, regardless the convergence rate. The resulting subduction rate is faster than the convergence rate where the hinge migrates away from the upper plate, whereas it is slower than the convergence where the hinge converges or advances toward the upper plate. The two end members appear sensitive to the geographic polarity of the subduction zones, being the asymmetry not among E–W subduction zones, but following or opposing an undulated "westerly" directed lithospheric flow (Doglioni, 1993), recently validated by space geodesy and statistical analysis (Crespi et al., 2007). This tectonic mainstream implies a relative mantle undulated counterflow, "easterly" directed.

The result of the asymmetry among opposite subduction zones is that along the W-directed subduction zones only the shallow rocks on the subduction hinge are accreted (sedimentary cover and slices of the topmost basement) because the accretionary prism basal decollement is located at the top of the subduction, without an actively thrusting upper plate. In the opposite E- or NE-directed subduction hinge, the lithosphere basal decollement is ramping upwards, providing a mechanism for uplifting deep-seated rocks (Doglioni, 1992).

As a consequence, the shortening is concentrated in the shallow layers of the lower plate along W-directed subduction zones, both during the oceanic and the continental stages of subduction, providing small volumes to the orogen and maintaining a low elevation of the critical taper (e.g., Bigi et al., 2003; Lenci et al., 2004).

The shortening is rather mostly concentrated in the upper continental lithosphere in E- or NE-directed oceanic subduction zones. Andean type trenches are in fact often characterized by tectonic erosion (e.g., von Huene and Lallemand, 1990; Ranero and von Huene, 2000; Kukowski et al., 2001; Clift et al., 2003), rather than accretion (Moore et al., 1990; Kukowski et al., 1994). It is unclear how deep it is transported the material offscraped by the tectonic erosion. It could eventually be re-transferred to the upper plate in the lower crust or upper mantle, and later re-exhumed during shortening progression.

However, accretion is documented along the Cascadia, Indonesia and several other segments of this type of subduction zones (e.g., Hyndman et al., 1993), thus providing a transfer of shallow rocks from the lower to the upper plate. But it is only at a collisional stage that the lower plate is eventually entirely involved and shortened. During this final stage, the lower plate contributes significantly to the volume increase of the orogen because shear zones and decollement planes enter much deeper in the continental crust of the footwall plate.

No tectonic erosion has seismically been observed along most of W-directed subduction zones. It has been proposed along the Tonga and other western Pacific subduction zones only on the basis of the forearc subsidence (e.g., Clift and MacLeod, 1999). However such subsidence can also be alternatively interpreted as generated by the hinge retreat. In the Apennines, documented accretion occurs while subduction hinge retreat-related subsidence exceeds the prism uplift (Doglioni et al., 1999a,c).

The term accretionary prism is often confused with the term forearc, which is the region between the subduction trench and the volcanic arc, having not always a unique tectonic meaning. The arc is also a misleading term used both to define a volcanic belt or a structural undulation of a thrust system. Moreover in the so-called forearc, extension can be widespread, without active accretion.

The shortening in prisms of W-directed subductions can be larger than the convergence rate, whereas it is in general smaller than the convergence in orogens of E- or NE-related subductions (Doglioni et al., 2005a, 2006a). This occurs because along W-directed slabs the shortening is confined in the lower plate and directly equivalent to the amount of subduction, which is larger than the convergence rate. In the E- or NE-directed slabs, the shortening is smaller than the convergence because the convergence is partitioned partly in the subduction rate, and partly in the contraction of the upper plate. While W-subduction-related prisms have a forebelt and a backarc basin, E–NE-subduction-related orogens have a forebelt and a widely developed retrobelt since the very early stages.

The different rock exhumation in the different subduction settings is highlighted by peculiar petrography of transported sediments (Garzanti et al., in press).

Regional foreland monoclines at the accretionary prism fronts also show steep dip and fast subsidence rates along W-directed subduction zones, and shallow dip and slow subsidence rates for E- or NE-directed subduction zones (Doglioni, 1994; Mariotti and Doglioni, 2000). This is a paradox because if the foredeep subsidence is generated only by the mountain load, along the highest mountains as the Andes and Himalayas we should expect the deepest trenches and foredeeps, but we rather observe these features along the opposite subduction zones where there is not a significant lithostatic load, like the Marianas trench, the Apennines and Carpathians foredeep. The relative “eastward” mantle flow could favor the bending of the slab and the foredeep fast subsidence along the hinge of W-directed subduction zones. The mantle counter flow should rather contrast the subsidence along the opposite subduction zones, providing a force sustaining the slab (Doglioni, 1994). While W-directed subduction-related prisms have low-grade metamorphism, the E- or NE-subduction-related orogens at the collisional stage may exhibit UHP rocks (Doglioni et al., 1999a). Metallogenesis appears to be also controlled by subduction style (Mitchell and Garson, 1981). The Mariana type subduction is characterized by Kuroko or similar volcanogenic sulphide deposits (Nishiwaki and Uyeda, 1983). Porphyry copper deposits are instead concentrated in collisional settings and Andean type subduction zones.

The seismicity of the slabs (Isacks and Molnar, 1971; Seno and Yoshida, 2004) is mostly characterized by down-dip compression along W-directed subduction zones, whereas it is quite often down-dip extensional along E- or NE-directed subduction zones. Japan subduction shows a different case, having two separate layers of slab seismicity, i.e., an upper one characterized by down-dip compression, and a lower one where down-dip extension prevails. However, Japan seems an intermediate case of subduction, where the Neogene W-directed subduction system is presently initiating to flip, having the backarc basin started to shrink (Mazzotti et al., 2001; Doglioni et al., 2006a). The seismicity is frequently down-dip extensional all along the subduction hinges of both subduction polarities, possibly related to the bending of the lithosphere.

Within the two opposite end-members, a number of different settings can occur. For example along the E–NE-directed subduction zones there are oceanic slabs under continental lithosphere as in the Andes, which may or may not evolve to continent–continent collision such as the Alps or Himalayas (Ernst, 2005). Along W-directed subduction zones there also are variable compositions of the lower plate (both oceanic and continental lithosphere, e.g., Marianas and Apennines, Banda arc) and variable depth of the basal decollement plane, determining variations in the volume of the related accretionary prism (e.g., Bigi et al., 2003; Lenci et al., 2004).

The vertical and lateral growth of orogens is strongly asymmetric. In fact prism or orogens related to W-directed subduction zones mostly generate an E-ward migrating wave, never reaching high structural and morphologic elevation, where the volumes and elevation rates of the prism are constrained by the subduction rate and the depth of the basal decollement plane of the accretionary prism. On the other hand, orogens related to E- or NE-directed subduction zones are much more elevated and their growth occurs both vertically and horizontally (Doglioni et al., 1999a).

In this article we only discuss the most striking asymmetries among subduction zones and their kinematics, and we speculate on how these interpretations allow alternative point of view on the dynamics of plate tectonics.

As prototypes of different subduction zones, the Alps and the Apennines (Fig. 3) are orogens formed above opposite slabs (Fig. 4). Along the first belt, since the Cretaceous, the European plate subducted SE-ward underneath the Adriatic plate, whereas along the second belt, since the Late Oligocene, the Adriatic plate subducted W-ward below the European plate, with fast slab rollback and backarc spreading in the western Mediterranean (Rehault et al., 1984; Rehault et al., 1985; Doglioni et al., 1999b; Frizon de Lamotte et al., 2000; Faccenna et al., 2004). The two belts show distinct characters (Fig. 4).

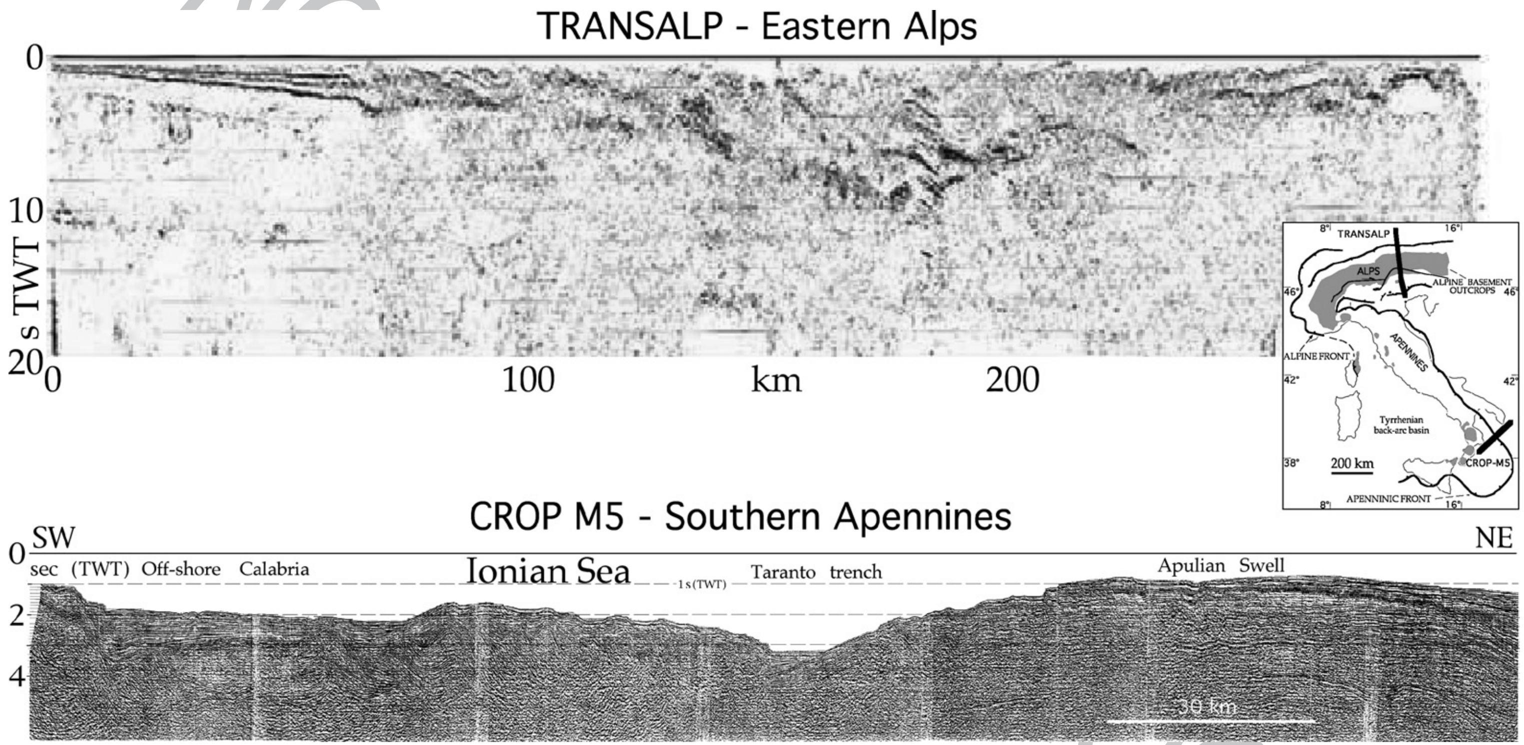

Fig. 3. Seismic sections of the Alps (Transalp, 2002) and of the Apennines (Scrocca et al., 2003).

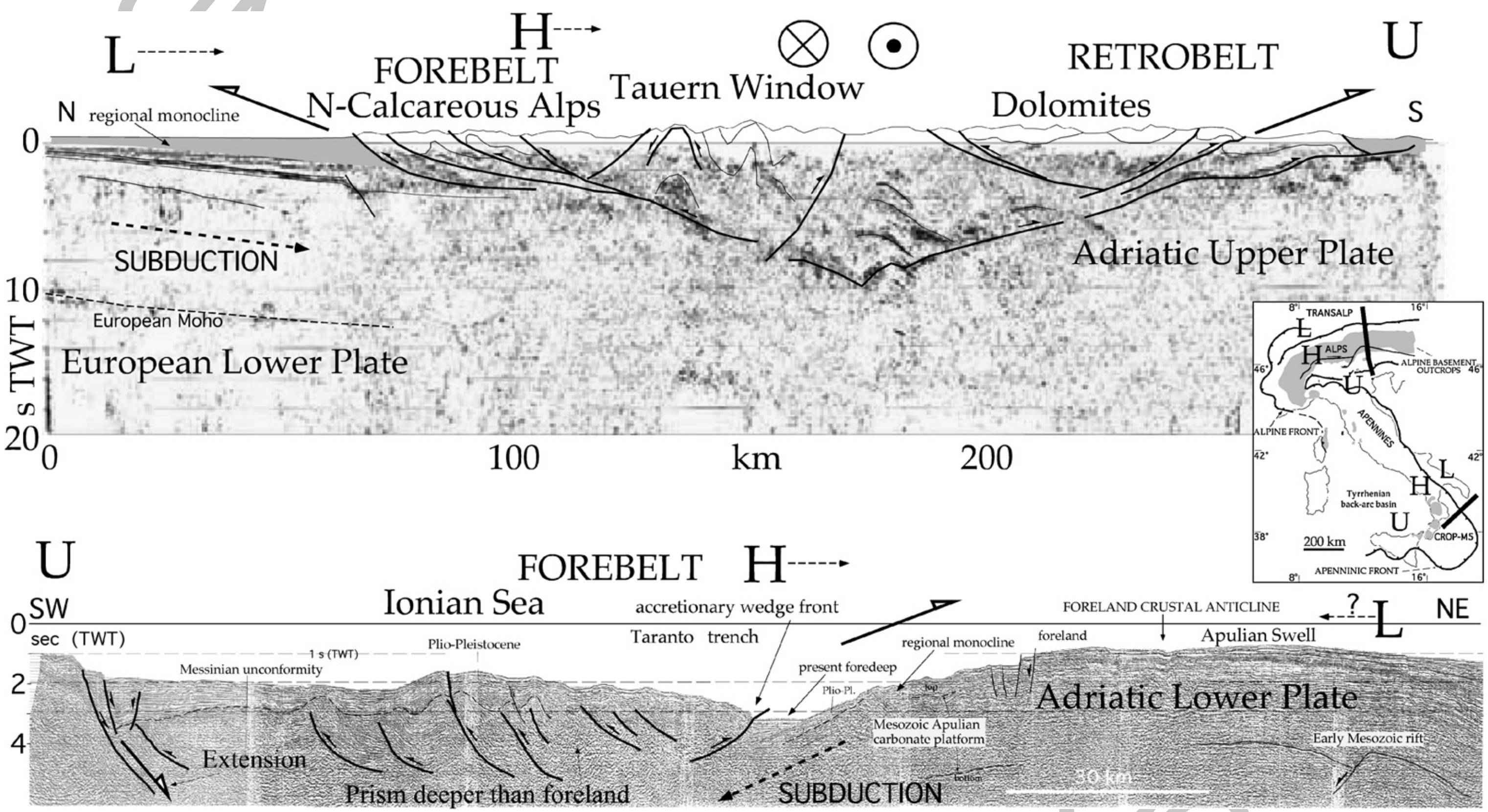

Fig. 4. Same sections as in Fig. 1 with some interpretation (modified after Doglioni et al., 1999c; Transalp, 2002). Note the Alpine double vergence vs. the single Apennines vergence, where the prism is even deeper than the foreland and followed by extension to the left. L, lower plate, H, subduction hinge, U, upper plate. In the Alps H migrated toward the upper plate, whereas it moves toward the lower plate in the Apennines. The Adriatic plate is the upper plate U in the Alps, while it is the lower plate L in the Apennines. The E–W segment of the Alps formed under right-lateral transpression.

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The Alps, unlike the Apennines, have double vergence, high structural and morphologic elevation, and no backarc basin. The Alps have two shallow foreland monoclines at the base of two foredeep basins, whereas the Apennines have a single deep foredeep, with steeper monocline and faster subsidence rates (Doglioni et al., 1999a). Beneath the Alps the crust is about doubled, and the lithosphere base is deeper than 100 km. Below the Apennines, a new shallow Moho formed in the western backarc side, and the asthenosphere is very shallow (30–40 km, Panza et al., 2003). The new Moho is kinematically required by the replacement of the original Moho, now subducted, on which the present surface thrust sheets were originally lying, being part of a pre-existing passive margin section. A new mantle section from west has replaced the subducted crustal section, the pre-subduction Moho and the lithospheric mantle of the lower plate, since the slab is eastward retreating (Doglioni, 1991).

The Alps have widespread outcrops of basement rocks. In fact along the Alpine belt, thrusts are deeply rooted, cross-cutting the whole crust and upper mantle (Dal Piaz et al., 2003). Seismic tomography shows large involvement of continental lithosphere in the subduction system (Mueller and Panza, 1986). In the Apenninic belt, the prism is rather mostly composed of shallow (mostly sedimentary) crustal rocks of the lower plate because the decoupling surface is traveling atop the downgoing lithosphere (Bally et al., 1986).

The different structural settings of Alps and Apennines suggest different dynamic and consequently kinematic evolution of the related subduction zones. These asymmetries are diffused worldwide in the orogens, and appear controlled by the polarity of the subduction, i.e., following or opposing the polarized undulated flow of plate motions.

A number of exceptions occur to these two end members. For example, Japan has low elevation and other characteristics of W-directed subduction zones, such as the backarc basin and an accretionary prism mostly composed of shallow rocks (von Huene, 1986). However the slab has low dip of the subduction plane (Jarrard, 1986) and there is no active hangingwall extension, in spite of an onshore morphology suggesting widely distributed structural depressions. The GPS data confirm that the Japan Sea backarc is not opening anymore, but it is rather closing (Mazzotti et al., 2001). This indicates that the system possibly arrived at an end, and its inversion started. Unlike other living W-directed subduction zones, the slab hinge in Northern Japan is moving toward the upper plate.

W-directed subduction zones formed mostly along the retrobelt of pre-existing E–NE-directed subduction zones, provided that oceanic lithosphere was present in the foreland to the east (Doglioni et al., 1999b). This would explain why, for example, the Barbados and Sandwich arcs formed only where the Northern and Southern America continental lithospheres narrow, and slices of Cordillera type are boudinaged and scattered in the backarc setting. Similarly, the Japanese W-directed subduction, developed during the Neogene, to the east of a retrobelt of an earlier Cretaceous Andean type orogen, generated by an E-directed subduction (Cadet and Charvet, 1983). This is also supported by the outcrops of Andean type co-genetic porphyry copper deposits in Japan and Chorea (Sillitoe, 1977; Mitchell and Garson, 1981).

W-directed subduction zones on Earth appear as short lived. If the Marianas or Japan subductions were active since the Cretaceous with a steady state E-ward hinge retreat, they should be now positioned in the middle of the Pacific. Backarc spreading in the hangingwall of W-directed subduction zones are mostly Cenozoic, pointing out that the related subduction should have a similar age. Backarc basins probably arrive to a critical opening stage, and then they become closed, until a new subduction starts. The non-standard Japan subduction in fact shows scattered intermediate deep seismicity, unlike it typically occurs along similar slabs (e.g., Omori et al., 2004).

Another apparent discrepancy to the W-directed vs. E–NE-directed slabs asymmetry is the occurrence of backarc basins also in the hangingwall of E- or NE-directed subduction zones. However these types of rifts rarely arrive to oceanic spreading as W-directed subduction do in their hangingwall. Moreover they may be post-subduction (e.g., Basin and Range, Doglioni, 1995), or sin-subduction but accommodating differential advancement of the upper plate over the lower plate. Examples of this type are the aforementioned Aegean and the Andaman rifts. The extension in western Turkey, Aegean sea, Greece and Bulgaria can be interpreted as a result of the differential convergence rates between the NE-ward directed subduction of Africa relative to the hangingwall disrupted Eurasian lithosphere (Doglioni et al., 2002). Considering fixed Africa, the faster SW-ward motion of Greece relative to Cyprus–Anatolia determines the Aegean extension. The differences in velocity can be ascribed to differential decoupling with the asthenosphere. Unlike west-Pacific backarc basins, where the asthenosphere replaces a subducted and retreated slab, the Aegean rift represents a different type of extension associated to a subduction zone, where the hangingwall plate overrode the slab at different velocities, implying internal deformation. While W-directed subductions occur with the rollback of the lower plate relative to the upper plate, the Aegean setting needs three plates, i.e., a common lower plate, and two plates overriding at different velocities. Analogously, assuming only few mm/yr of relative motion between the Indian and Australian plates (e.g., Bird, 2003), along the Himalayas collision, the Indo-Australian plates converge relative to Asia at about 36 mm/yr, while the same Indo-Australian plates along their oceanic northern part are overridden by the Sumatra–Burma plate at around 64 mm/yr (Cuffaro et al., 2006). Therefore, assuming a relatively coherent lower plate, the hanging-wall plates move SSW-ward at different velocity, this gradient being responsible for the extension between Asia and Sumatra–Burma, and generating the Andaman rift. In this type of geodynamic setting the subduction hinge still moves toward the upper plate, as in the normal E- or NE-directed subduction zones.

There are belts that are apparently not following the global trend of plate motions, like orogens that are E–W trending, and related to N–S convergence (e.g., the Pyrenees, Venezuela–Colombia belt). They have Alpine character and may kinematically be explained as related to subrotation of plates (Cuffaro et al., 2004). Other obliquely oriented belts are the Atlas and the Maghrebides in northwest Africa. The Atlas is not directly related to a subduction zone, but has been interpreted as an intraplate inversion structure (Laville and Petit, 1984; Warme, 1988; Doglioni, 1989; Frizon de Lamotte et al., 2000; Teixell et al., 2003; Frizon De Lamotte, 2005), while the E–W-trending Maghrebides (Frizon de Lamotte et al., 2000; Roca et al., 2004) are rather the right-lateral prolongation in northwestern Africa of the arcuate Apennines–Maghrebides subduction system. Along this segment of the belt, the right-lateral transpression coexisted with an about 5 times slower roughly N–S component of Africa–Europe convergence (Gueguen et al., 1998). The arcuate geometry of the Moroccan Riff implies local left-lateral transpression along its eastern margin, where the Maghrebides–Tell–Riff belt interferes with the inverted Middle Atlas (Doglioni, 1989). The E–W-trending Himalaya is instead almost perpendicular to the global tectonic mainstream (Doglioni, 1990).

3. Basic kinematics in the upper plate reference frame

When the subduction hinge behavior is added to plate motion, the following three end members can occur. If the hinge H is stationary, the subduction will equal the convergence (Fig. 5A). If the hinge advances as the convergence, the subduction rate will be zero (Fig. 5B), whereas in case the hinge retreats, the subduction rate will be added to the convergence rate. For example, it will be twice the convergence rate if the hinge retreats at the same speed of the convergence (Fig. 5C).

The rate of vertical descent Sv and horizontal penetration Sh into the mantle will be Ssinα, and Scosα, respectively, where α is the slab dip (Fig. 1). The steeper the slab, the faster will be its descent into the mantle and viceversa.

Let us then consider, for a subduction zone, a reference frame with three points, one attached to the upper plate U, a second attached to the lower plate L, and the third on the transient subduction hinge H (Fig. 1). The point U located on the upper plate is taken as fixed, i.e., its velocity $V_U = 0$. The motion of the two remaining points is considered relative to the fixed upper plate, so that V_L is the velocity of the lower plate and V_H is the velocity of the hinge. If lower plate and subduction hinge move towards the fixed point in the upper plate, then V_H and V_L are assumed to be

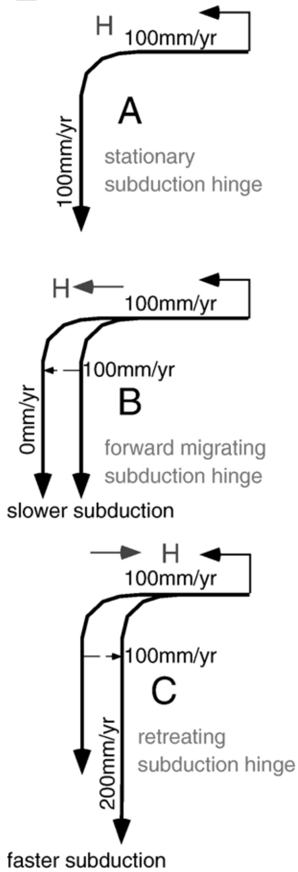

Fig. 5. Along a subduction zone, say with 100 mm/yr convergence, if the subduction hinge H is stationary (A), the subduction rate will be the same of the convergence. If H migrates left 100 mm/yr (B), than the subduction would be null. If H migrates right of the same amount, the subduction will be doubled (C).

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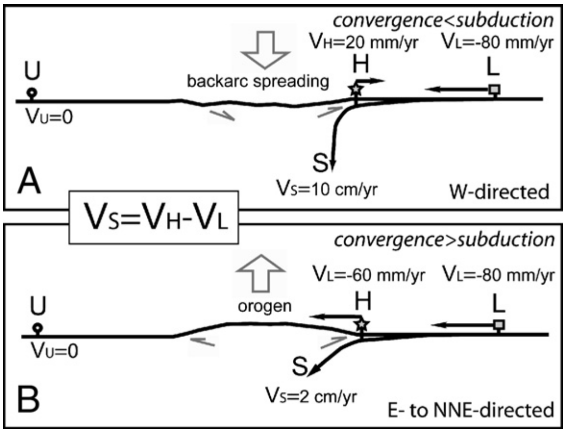

Fig. 6. Basic kinematics of subduction zones, assuming fixed the upper plate U, a converging lower plate L, and a transient subduction hinge, H. The subduction rate S is given by $V_S = V_H - V_L$. Values are only as an example. S increases when H diverges relative to the upper plate (a), whereas S decreases if H converges (b). The movements diverging from the upper plate are positive, whereas they are negative when converging. The case a) is accompanied by backarc spreading, a low prism and is typical of W-directed subduction zones, whereas in case b) double verging and elevated orogens form and is more frequent along E- to NNE-directed subduction zones. Note that in both W- and E–NE-directed subduction zones, the hinge migrates eastward relative to the upper plate, suggesting a global tuning in subduction processes.

negative. If lower plate and subduction hinge move away, then V_H and V_L are positive. In the following, the term convergence (of lower plate and hinge with respect to upper plate) will be adopted for negative V_H and V_L, whereas divergence will refer to positive V_H and V_L. V_L is easily derived from plate motion models. The exact determination of the hinge location is not that easy due to the wide rounded area of dip change from the horizontal lower plate lithosphere to the inclined downgoing slab. For sake of simplicity, the motion of the hinge zone can tentatively be assumed to be close or coincident with that of the subduction trench (this approximation has been used also in other studies of subduction zones (Heuret and Lallemand, 2005).

Particular attention will be given to the effects of the kinematics of upper, lower plate and hinge on the subduction velocity (V_S). This is defined as the velocity with which the subducting lithosphere enters the subduction zone and can be calculated with the simple equation: ($V_S = V_H - V_L$). For example, in the particular case of the lower plate converging with the upper plate, the subduction rate is a function of the behavior of the subduction hinge.

In this section we speculate on all the possible combinations of U, H and L, irrespective of the fact that the analyzed cases can be found in nature. For each case we describe the expected subduction velocity and the predicted deformation of the upper plate. Generally speaking, the upper plate is predicted to stretch (with the opening of backarc basins) for diverging hinges (Figs. 6, 7, 8 and 9), or it is expected to shorten (or shrink) for

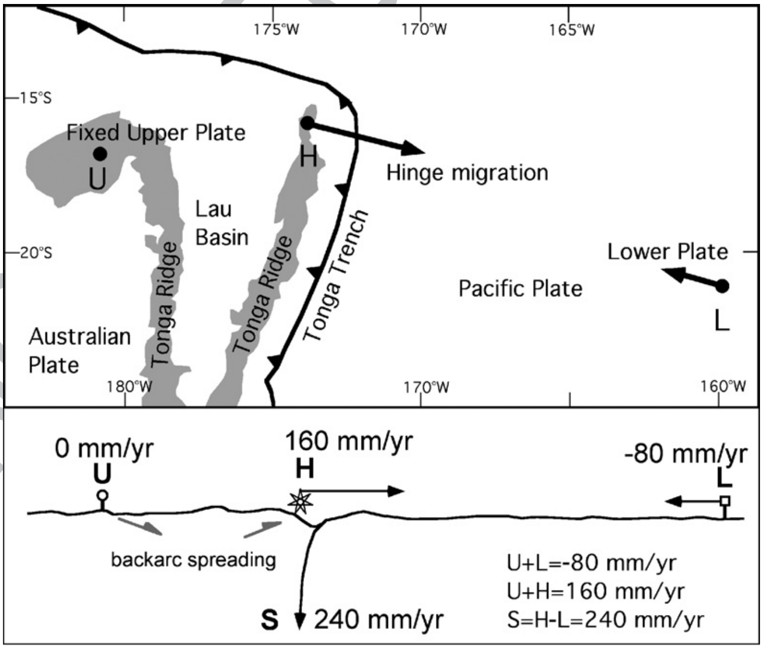

Fig. 7. GPS data of Bevis et al. (1995) along the Tonga subduction zone, taken as fixed upper plate. Note that the subduction rate is the sum of convergence between U and L, plus the motion of H. This is the fastest subduction in the world, where more than 700 km of lithosphere should sink in about 3 Ma.

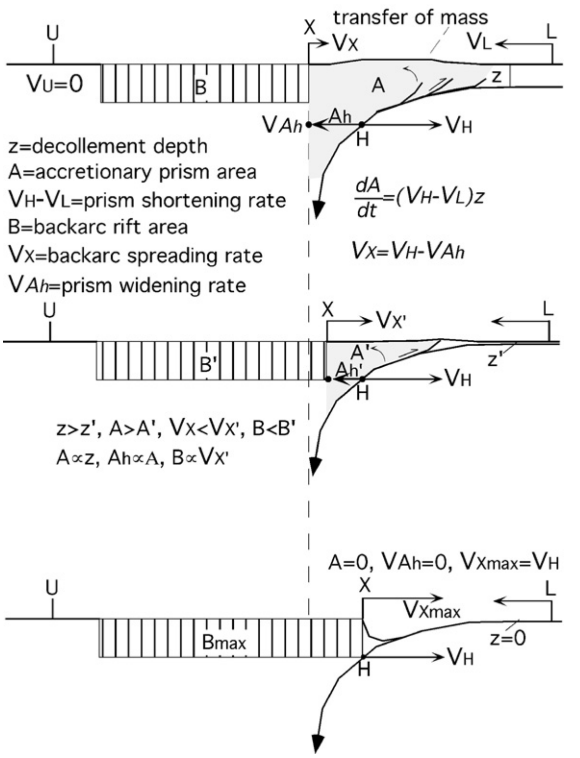

Fig. 8. In case the accretionary prism is entirely formed at the expenses of the lower plate as it occurs along W-directed subduction zones, the shortening is equal to the subduction rate ($V_S = V_H - V_L$). With given convergence (V_L) and subduction hinge (V_H) migration rates, three different cases are illustrated. The deepest position of the decollement plane depth (z) of the accretionary prism determines a larger transfer of mass per unit time from the lower to the upper plate (upper panel). This decreases the width of the backarc rift and its velocity of spreading (V_X). As a consequence, the reference point X migrates toward the foreland slower as the decollement depth and the accretionary prism width (Ah) increase. U, fixed upper plate. In case no accretion occurs at the prism front (i.e., $z = 0$ and the lower plate is entirely subducted, lower panel), the backarc spreading should equal the velocity of the hinge ($V_X = V_H$). The middle panel is an intermediate case. Therefore, the reference point X might not be a reliable kinematic indicator of the subduction hinge migration rate when accretion occurs, besides all the other problems discussed in the text.

hinges converging toward the upper plate (Figs. 10, 11, 12 and 13). These behaviors occur regardless the relative motion of the lower plate. Such subduction kinematics are summarized in Table 1. When available, we will briefly describe natural examples of the analyzed cases. We will show that some kinematics are typical of "W"-directed slabs and some others of "E- or NE"-directed subduction zones.

It is finally emphasized that, in this section, the analysis is purely kinematic. As a consequence, the reader could be confronted with subduction kinematics contrasting commonly accepted ideas on subduction dynamics (controlled by slab pull). We show that such contrasts vanish if other plate driving forces are considered.

In Table 1 three major groups of kinematic settings are listed: for converging, for stationary and for diverging lower plates. For each of these three groups all the combinations with the relative motion of the subduction hinge with respect to the upper plate are considered. The behavior of H is variable along subduction zones worldwide. Convergence rate is also changing in the different settings and even along a single subduction zone. The different kinematics are reported as a function of the main subduction polarity in Figs. 14 and 15, and the different cases are associated to regional examples. Generally speaking, along steady W-directed subduction zones, the hinge mostly diverges relative to the upper plate, and backarc basin forms. On the contrary, along E- or NE-directed subduction zones the most commonly the subduction hinge converges relative to the upper plate, and a double verging orogen develops, without backarc extension. A synoptic view is in Fig. 16, where the aforementioned fourteen kinematics are listed.

3.1. *For converging lower plates (L<0), six potential kinematics can occur*

3.1.1. *Kinematics #1*

If the hinge diverges (H>0), then backarc extension occurs in the upper plate and subduction is faster than

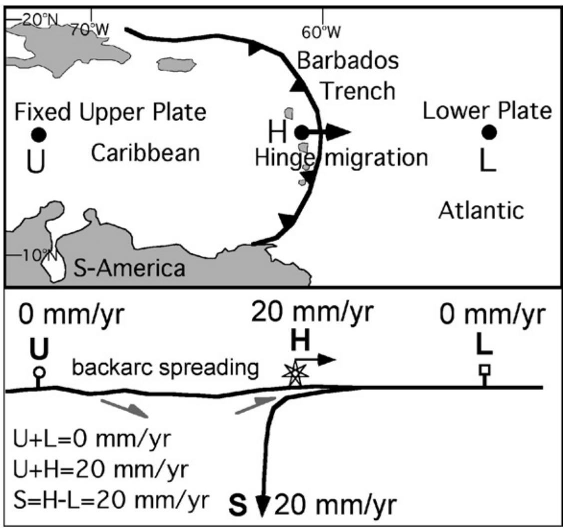

Fig. 9. Along the Caribbean subduction zone, the hinge migrates E-ward while no convergence should occur between the central parts of the backarc basin to the west and the Atlantic Ocean to the east, although no data are yet available for the interior of the backarc basin. GPS data after Weber et al. (2001).

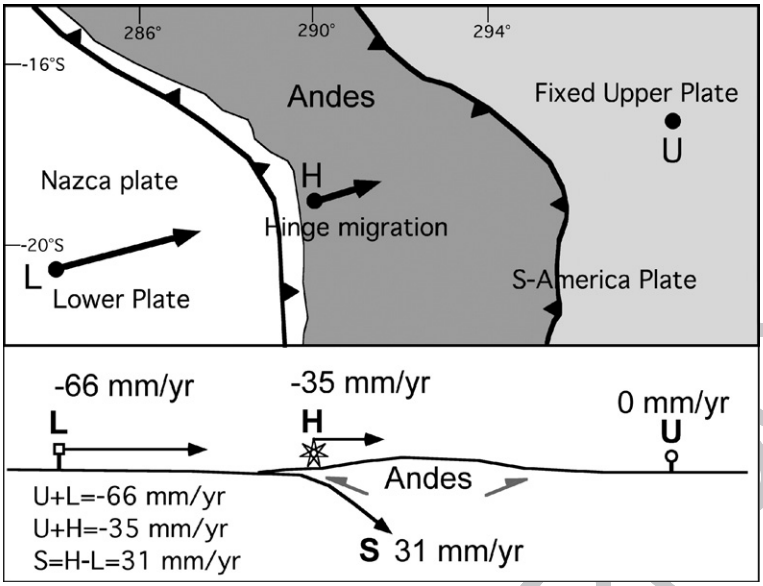

Fig. 10. The convergence between Nazca and South America plates is faster than the shortening in the Andes. The upper plate shortening decreases the subduction rate. The convergence/shortening ratio is about 1.88. GPS data after Liu et al. (2000).

convergence (fast sinking slab). This kinematics is typical of the western Pacific subductions, such as Tonga and Marianas. The present-day motion of the Tonga subduction zone (Fig. 7) has been investigated by Bevis et al. (1995). In this subduction, the worldwide fastest plate velocities have been described. Converting their results in the fixed upper plate reference frame, the Pacific plate (L) is converging at 80 mm/yr, and the Tonga islands, fairly representing H, are diverging at about 160 mm/yr. Therefore the subduction rate S should be 240 mm/yr (Fig. 7).

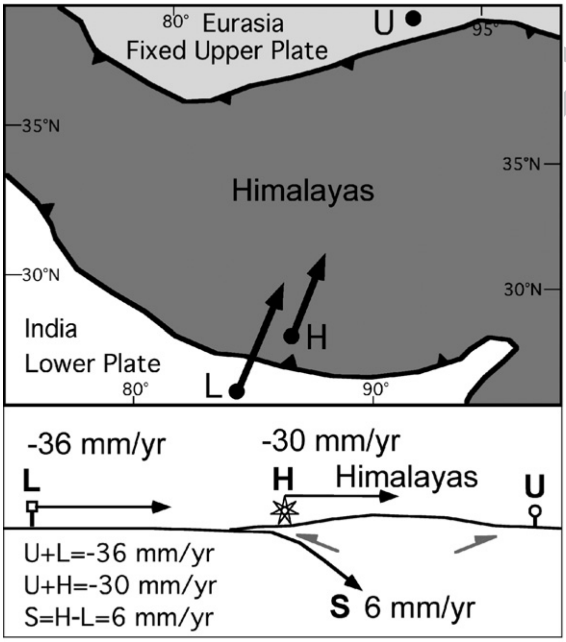

Fig. 11. The convergence between India and Eurasia plates is slightly faster than shortening in the Himalayas. The difference should be the continental subduction rate. The convergence/shortening ratio is about 1.2. GPS data after Jouanne et al. (2004) and Zhang et al. (2004).

In the case of the Marianas, we assume that variable rates of backarc spreading are related to different rates of hinge (H) migration from the upper plate. In the northern Mariana backarc, rates ranging from 60 to 20 mm/yr have been described from Pliocene to present times (Yamazaki et al., 2003). The related variations of H velocity should have determined different subduction rates, since H is summed to the convergent rate of the lower plate L to compute the slab sinking (Fig. 14, 1W). Convergence rates calculated using the NUVEL1A rotation poles (DeMets et al., 1994) vary between 90 and 95 mm/yr (Cruciani et al., 2005). As a consequence, subduction rates along the Marianas subduction zone vary between 110 mm/yr and 150 mm/yr. Variable backarc spreading rates have been reported also in the Lau backarc basin, in the hangingwall of the Tonga subduction zone (Taylor et al., 1996). Other possible theoretical cases are where the hinge diverges in absolute value faster or slower than the convergence rate of the lower plate. The sum of hinge divergence and lower plate convergence in both cases determines the subduction rate. However, faster hinge divergence generates larger backarc spreading, whereas larger lower plate convergence rate should produce larger shortening rate in the accretionary prism. However we do not distinguish these cases considering them as variations of the kinematics #1.

3.1.2. Kinematics #2

If the hinge is stationary (H=0), then neither stretching nor shortening occur in the upper plate and subduction velocity equals the lower plate convergence (S=|L|) (sinking slab). It is most likely typical of the onset stages of both W-ward and E–NE-directed subduction zones. No present-day examples are available.

3.1.3. Kinematics #3

If the hinge converges (H<0) slower than the lower plate (|H|<|L|), then subduction is slower than lower plate convergence (slowly sinking slab). Shortening occurs normally along E- or NE-directed subductions (e.g., Andes, Fig. 10). In W-directed subductions, this kinematics normally follows a stage of kinematics#1, when a backarc basin opened. In this case, such as for Japan, inversion tectonics occurs in the inactive (now closing) backarc basin of the upper plate. The South Japan subduction zone (Fig. 13) is a peculiar case of W-directed subduction zone where unlike typical similar settings, H is converging rather than diverging (Fig. 13). In fact, the backarc spreading apparently ended according to seismicity and space geodesy data (Mazzotti et al., 2001). At latitude 35°N, along the Japan subduction, the lower plate converges at 92 mm/yr. The backarc basin is closing at rates of 25 mm/yr (Cruciani et al., 2005) and such rates can be taken as representative of the hinge convergence. As a consequence, subduction rates are

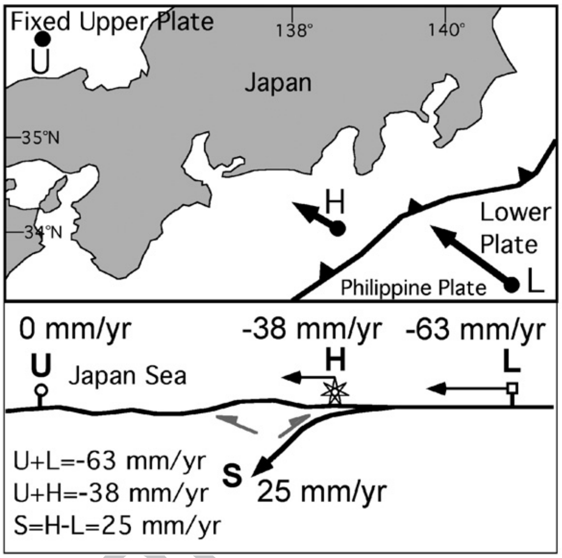

Fig. 13. The Japan subduction zone appears as a setting where a W-directed subduction zone is inverting. The hinge is not migrating away from the upper plate, but rather converging. The backarc basin is starting to shrink. GPS data of Japan stations relative to fixed Amuria plate (Eurasia), after Mazzotti et al. (2001).

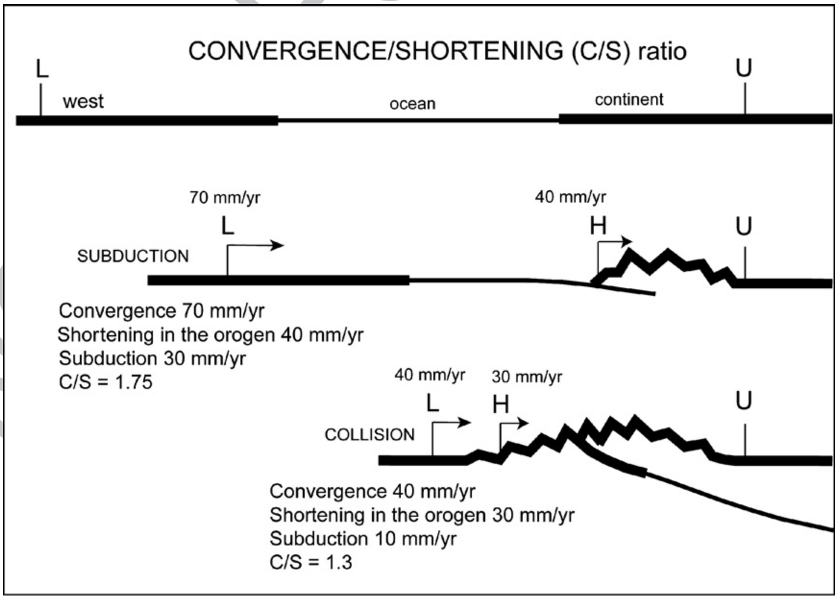

Fig. 12. During oceanic subduction, the shortening is mostly confined in the continental upper lithosphere due to its lower strength and of the coupling between upper and lower plates. At the collisional stage also the lower plate is widely involved. Larger the shortening, smaller the subduction rate, and lower the upper plate strength+greater the plate coupling. The convergence/shortening ratio decreases at the collisional stage.

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t1.1 Table 1

t1.2 Main kinematic relationships of subduction zones, assuming fixed the upper plate, and moving the lower plate L and the subduction hinge H. For readability sake, S = V_S, H = V_H, L = V_L

	Case # and kind of subduction	Lower plate (L) and subduction hinge (H) motions and their velocities (L and H)	Subduction velocity	Kind of deformation of the upper plate	Does it occur in E–NE-directed subductions?	Does it occur in W-directed subductions?	Natural examples
t1.3							
t1.4	1 Fast-sinking slab	L converging (L<0) H diverging (H>0)	Subduction S faster than L convergence (S>|L|)	Backarc extension	No	Yes	Tonga, Marianas
t1.5	2 Sinking slab	L converging (L<0) H stationary (H=0)	Subduction velocity equal to L convergence (S=|L|)	Neither stretching nor shortening	Yes	Yes	Onset of subduction. No present day examples
t1.6	3 Slowly sinking slab	L converging (L<0) H converging (H<0) slower than L (|H|<|L|)	Subduction slower than L convergence (S<|L|)	Shortening (Andes) or inversion tectonics (Japan)	Yes	Yes	Andes Japan
t1.7	4 Hanging slab	L converging (L<0) *H* converging (H<0) as fast as *L* (|H|=|L|)	Subduction rate is null (S=0)	Fast shortening	Yes	No	Final stages of continental collision
t1.8	5 "Emerging" or detaching slab	L converging (L<0) H converging (H<0) faster than L (|H|>|L|)	Subduction rate is negative (S<0)	Fast shortening	No	No	/
t1.9	6 Sinking slab and three plates	L converging (L<0) Two upper plates, U–U"	Subduction faster than L convergence (S>|L|)	extension	Yes	No	Aegean Sea Andaman
t1.10	7 Sinking slab	L stationary (L=0) *H* diverging (H>0)	Subduction rate equals *H* retreat (S=|H|)	Backarc extension	No	Yes	Barbados
t1.11	8 Hanging slab	L and H stationary (L=H=0)	Subduction rate is null (S=0)	Neither stretching nor shortening	Yes	Yes (?)	Urals and Carpathians (?)
t1.12	9 "Emerging" or detaching slab	L stationary (L=0) H converging (H<0)	Subduction rate is negative (S<0)	Shortening	No	No	/
t1.13	10 "Emerging" or detaching slab	L diverging (L>0) H converging (H<0)	Subduction rate is negative (S<0)	Shortening	No	No	/
t1.14	11 "Emerging" or detaching slab	L diverging (L>0) H stationary (H=0)	Subduction rate is negative (S<0)	Neither stretching nor shortening	No	No	/
t1.15	12 "Emerging" or detaching slab	L diverging (L>0) H diverging (H>0) slower than L (|H|<|L|)	Subduction rate is negative (S<0)	Post-subduction extension	Yes	Yes	Southern Apennines (?) California (Basin and Range)
t1.16	13 No sinking-laterally moving slab	L diverging (L>0) H diverging (H>0) as fast as L (|H|=|L|)	Subduction rate is null (S=0)	Backarc extension	No	Yes	Southern Apennines
t1.17	14 Slow-sinking slab	L diverging (L>0) H diverging (H>0) faster than L (|H|>|L|)	Subduction rate is slower than H retreat (S<|H|)	Backarc extension	No	Yes	Banda

expected to be of 67 mm/yr. No direct measurements are available for the motion of the Andean subduction hinge (Fig. 10). However, along the Andean belt the calculated shortening in the orogen is about half of the convergence rate between the Nazca plate and the South America plate (Liu et al., 2000). The unavailability of direct measurements of hinge convergence velocities can be overcome assuming that the shortening is approximately equal to the hinge convergence. Along the Andes subduction zone, at about 27° S, the lower plate convergence L calculated using the NUVEL1A rotation poles is about 66 mm/yr. Using the data by Liu et al. (2000), the subduction hinge converges toward the upper plate at about 35 mm/yr. As a consequence the subduction rate S will be 31 mm/yr (Fig. 10). The hinge convergence relative to the upper plate indicates the amount of the upper plate shortening and is faster as the strength of the upper plate is lower (Doglioni et al.,

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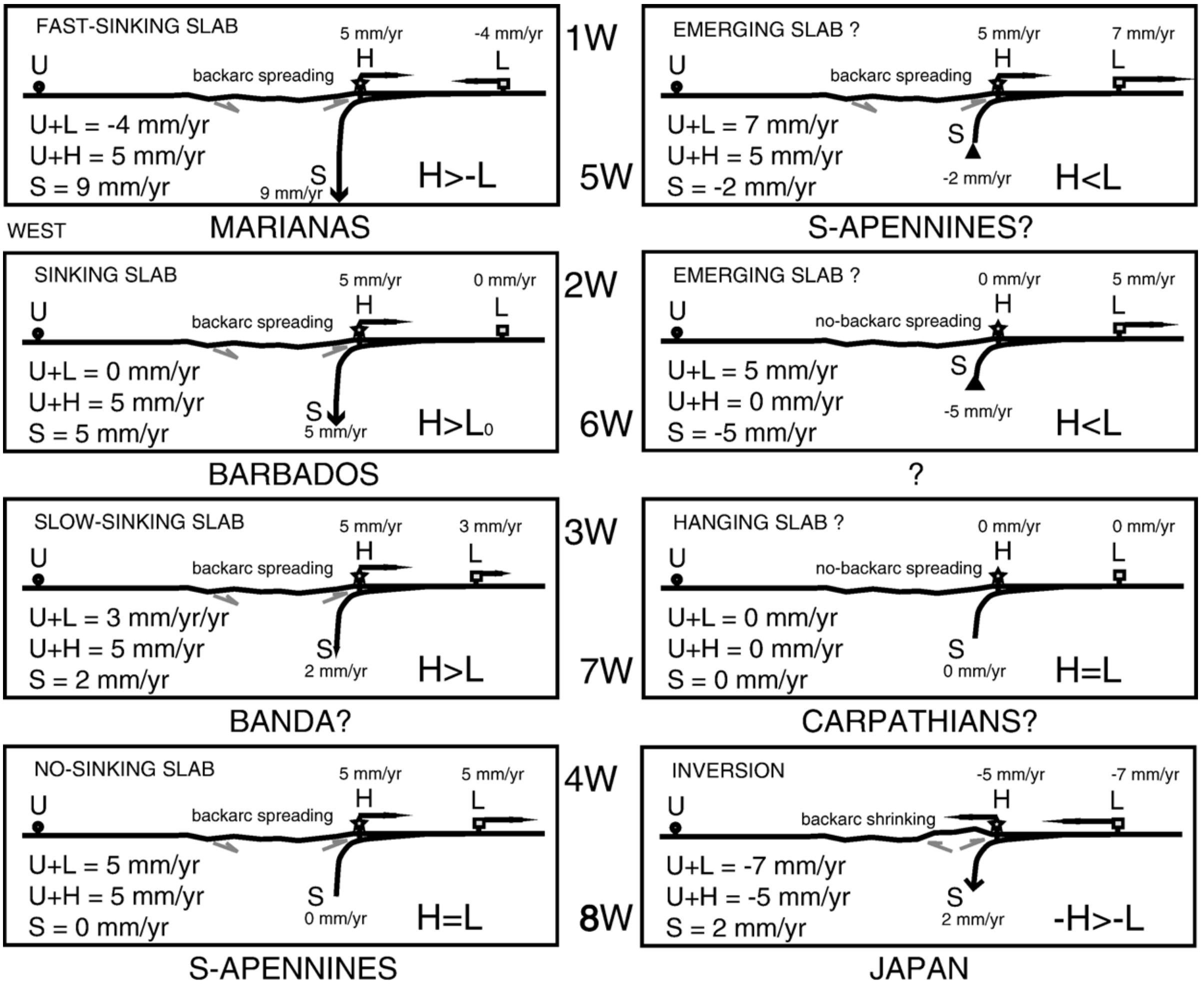

Fig. 14. Different kinematic settings of W-directed subduction zones. In the 1 to 5 sections, H is moving away from the U, in 6 and 7 H is fixed, and in section 8 is moving toward U. The site L is converging relative to U in sections 1 and 8. It is fixed to U in sections 2 and 7. The other cases have L moving away from U, but with different relationship with H, i.e., faster, slower or fixed. The different regions are interpreted as examples of the variable settings. The different regions are interpreted as examples of the variable settings. Velocities are only for relative comparison and do not apply to the example areas.

2006a). The convergence/shortening (C/S) ratio could be a useful parameter for describing the strength of the upper plate. The strength of the lithosphere is controlled by active deformation mechanisms (either brittle or plastic) and varies consistently with depth. Although on the long period the rheological behavior of the lithosphere is best described by power-law equations, the strength of the entire lithosphere is sometimes described in terms of viscosity. For example, in the section of Fig. 10, the C/S ratio is 1.88, and moving southward along the Andes it decreases, suggesting relative lower strength of the upper plate. Assuming stress balance, the viscosity of the upper plate continental lithosphere in the Andes has been computed as low as 3×10^{21} Pa s (Husson and Ricard, 2004). The viscosity of oceanic lithosphere tends to be generally larger, and it decreases with depth. For example an oceanic lithosphere at 25 km depth might have a range of viscosity between 10^{22} and 10^{27} Pa s, increasing with its age (Watts and Zhong, 2000). This may also explain why the upper plate made by softer continental lithosphere absorbs most of the deformation.

Higher the convergence/shortening ratio (C/S), higher the strength of the upper plate (Doglioni et al., 2006a). A further parameter controlling the C/S ratio is the coupling between upper and lower plates. A strong coupling may be achieved when the friction of rocks at the interface between subducting and overriding plates is high. Strong friction will transform a consistent part of the motion of the subducting plate into deformation of

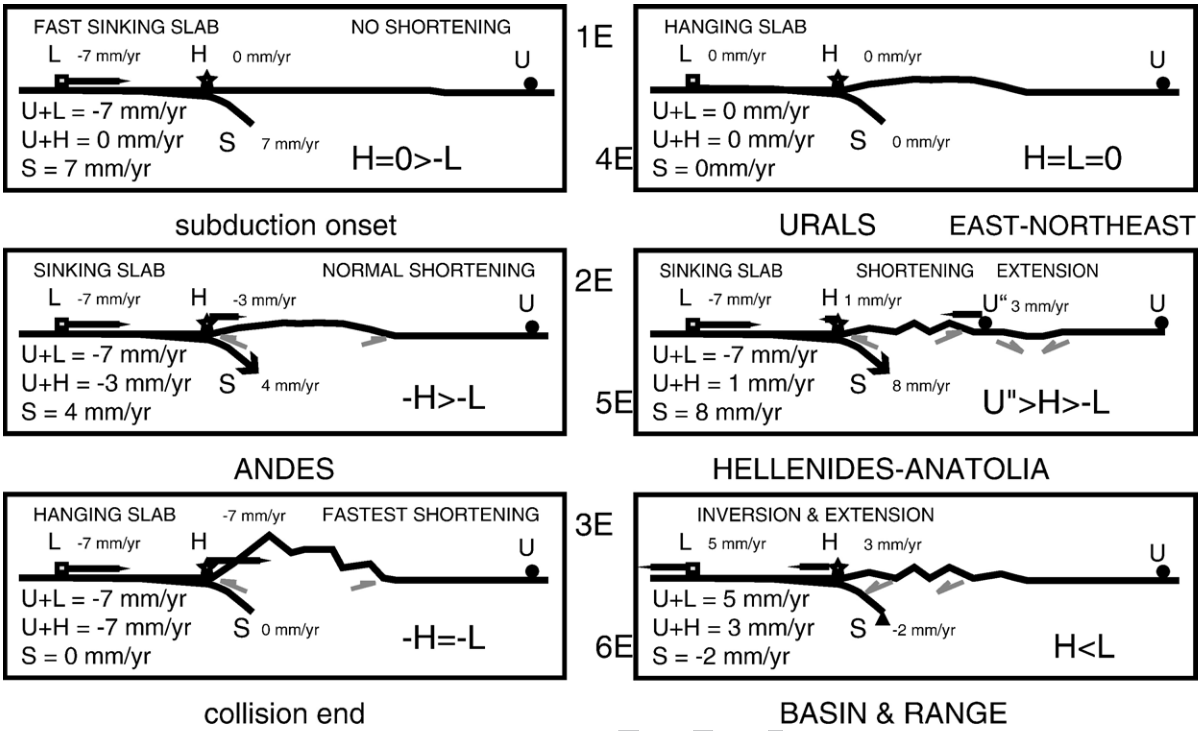

Fig. 15. Different kinematic settings of E- or NE-directed subduction zones. In the sections 1, 2, 3 and 5, L is converging relative to U, whereas it is fixed in section 4 and diverging in section 6. The hinge H is fixed relative to U in sections 1 and 4. It is rather converging in sections 2, 3 and 5, whereas it is diverging in section 6. The different regions are interpreted as examples of the variable settings. Velocities are only for relative comparison.

the upper plate and thus enlarge the C/S ratio. The minimum value of this ratio for E–NE-directed subduction zones should be 1, where the amount of convergence equals the amount of shortening in the belt, indicating very low strength and large coupling, and virtually no subduction. Subduction is inversely proportional to the amount of shortening in the upper plate, and the convergence/shortening ratio increases exponentially with the decrease of shortening, and the increase of the subduction rate. During collision, convergence rate slows down, and the shortening is propagating more pervasively into the lower plate (Fig. 11). Therefore the C/S ratio decreases during the final stages of the orogen life (Fig. 12).

The C/S ratio is 1 if convergence and shortening were equal. Therefore, as the shortening increases due to lower strength of the upper plate, the C/S ratio becomes higher.

The Himalayas range (Fig. 11) had an intense geodetic investigation during the last years (e.g., Banerjee and Bürgmann, 2002; Jade et al., 2004; Jouanne et al., 2004; Zhang et al., 2004). The convergence between the India and Eurasia plates is around 36 mm/yr, and the orogenic shortening is lower (30 mm/yr) along the selected section as in the Andes, but also the C/S ratio is lower (1.2), since the shortening is relatively higher with respect to the convergence rate, possibly due to the involvement of the softer lower plate (Fig. 11).

As demonstrated by Zhang et al. (2004), the deformation of the Tibetan Plateau cannot be accommodated by single thrusts, but it is rather a diffuse deformation across the entire lithosphere. This observation can be applied to most of the orogens worldwide, and it is consistent with a viscous dissipation of the shortening within the whole deeper part of the orogen, accounting for a large part of the convergence rate, but not all of it. The remaining part is expressed by the subduction rate.

The C/S ratio is > 1 along E- or NE-directed subduction zones, whereas it is < 1 along accretionary prisms of W-directed subduction zones where the subduction rate and the related shortening are faster than the convergence rate.

The geodynamic setting where the H migrates toward the upper plate generates a double verging Andersonian belt, no backarc basin, and high structural and morphologic elevation (Figs. 6, 10, 11 and 15).

Another example of convergence faster than hinge convergence is the E-directed Cascadia subduction zone (Dragert et al., 2001). Convergence rate between L and fixed U is about 35–40 mm/yr. The shortening relative

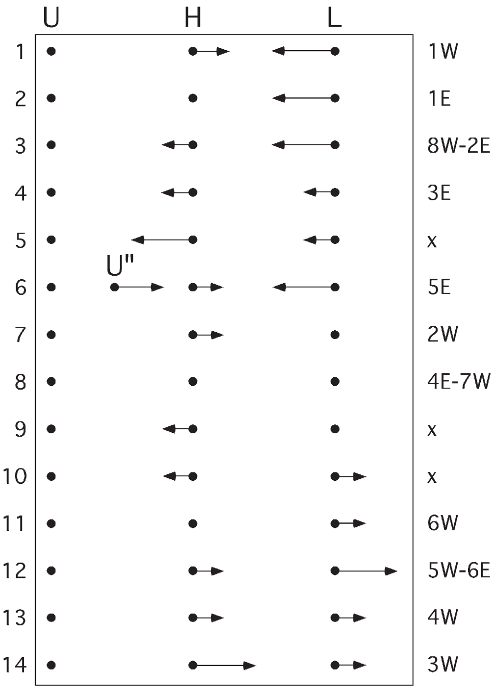

Fig. 16. Geodynamic settings where the upper plate U is considered fixed, the subduction hinge H can diverge, be fixed or converge, and the lower plate can either converge, be fixed or diverge. Case 6 occurs when the upper plate is in the third dimension split into two plates, U and U″. Few cases are potentially nor realistic, i.e., 5, 9 and 10 (*X*) because the subduction hinge cannot converge in case of absence of faster or equal lower plate convergence. Numbers on the left refer to the kinematics examples discussed in the text, and numbers and letters to the right refer to the cases of Figs. 14 and 15.

to fixed upper plate gradually decreases by means of deep ductile deformation and upper crustal thrusting from about 20 mm/yr at the trench, considered as the hinge H, to zero in the stable plate interior.

3.1.4. Kinematics #4

If the hinge converges (H<0) as fast as the lower plate (|H|=|L|), then subduction rate is null (S=0; hanging slab) and the upper plate is subject to shortening. This kinematics is typical of final stages of continental collision.

3.1.5. Kinematics #5

If the hinge converges (H<0) faster than the lower plate (|H|>|L|), then subduction rate is negative. In other words, the plate subducted during previous stages characterized by other kinematics would tend to re-emerge from the mantle. Since the negative buoyancy of the subducted slab would prevent this, the pull-down forces due to negative buoyancy and the pull-out forces due to plate kinematics would induce a strong tensional stress field in the slab. The slab could respond to such a stress field breaking off (slab detachment, e.g., Wortel and Spakman, 2000). The same observations will be applicable to later cases characterized by similar negative subduction rates. No present day examples of this kinematics are known, and this setting seems to be unrealistic because the hinge can converge only if the lower plate converges as well, unless it is broken into two diverging subplates.

3.1.6. Kinematics #6

A particular case occurs when, along a subduction zone, the hangingwall is not a coherent single plate, but differential velocities occur. Then the upper plate has to be considered as composed by two upper plates (Fig. 15, 5E). Assuming fixed the far field upper plate (U=0), a segment of the upper plate closer to the subduction zone diverging from the far field upper plate with velocity U″ (as an example we assume U″=3 mm/yr), a lower plate converging for example at rates L=−7 mm/yr, and a hinge diverging at H=1 mm/yr, than the subduction rate will be 8 mm/yr. In other words the subduction rate is calculated with respect to the part of the upper plate closer to the subduction zone. The upper plate U″ will suffer a shortening of 2 mm/yr, and will be rifted from the upper plate U at 3 mm/yr. These kinematics are more complex than the other cases here analyzed, since they include three plates (two upper plates and a lower plate). Although not easily described by the simple relation of the two-plate cases, this kinematic setting is particularly important since it explains the occurrence of extension in the hanging-wall of E- or NE-directed subduction zones such as the Aegean and the Andaman seas, where the rift is sin-subduction, but not related to typical slab retreat. In other words, the two upper plates override the lower plate at different velocities, and the velocity gradient between the two upper plates control the rift. For example, along the Hellenic and Cyprus–Anatolia subduction zone, Greece is SW-ward overriding Africa faster than Anatolia (Doglioni et al., 2002). The faster separation of Greece relative to Anatolia in the hangingwall of the subduction generates the Aegean extension.

3.2. For stationary lower plates (L=0), three potential kinematics can occur

3.2.1. Kinematics #7

If the hinge diverges (H>0) then backarc extension occurs in the upper plate and subduction rate equals the hinge retreat (sinking slab). This case is observed in the W-directed Caribbean (Barbados) subduction zone (Fig. 9). The kinematics of the Barbados–Caribbean arc is well constrained by present day data (Weber et al., 2001), showing 20 mm/yr of E-ward migration of H. The Atlantic side, i.e., the lower plate L can be considered fixed to South America. On the other hand backarc spreading should increase the distance between the Caribbean arc and the Central America Cordillera to the west. Unfortunately there are no GPS sites in the center of the Caribbean basin to validate this statement, but assuming no relative eastward motion between the center of the backarc basin and the Atlantic, the convergence between lower and upper plate should be null (L = 0). As a consequence the subduction rate would coincide with the H velocity and be equal to 20 mm/yr (Fig. 9). This interpretation differs from most of the published kinematic models that assume the Caribbean plate as a single coherent block (e.g., Weber et al., 2001).

3.2.2. Kinematics #8

If the hinge is stationary (H = L = 0) then no deformation in the upper plate is expected to occur and the subduction rate is null as well (hanging slab). Post-collisional settings such as the Urals or other more recent presently inactive alpine segments could be similar cases. Another potential instance could be the northern Carpathians.

3.2.3. Kinematics #9

If the hinge is converging (H < 0), then shortening should occur in the upper plate and the subduction rate should be negative (emerging or detaching slab), as in case #5. No present-day examples are known.

3.3. For divergent lower plates (L > 0), five potential kinematics can occur

3.3.1. Kinematics #10

If the hinge converges (H < 0) shortening should occur in the upper plate and the subduction rate should be negative (emerging or detaching slab). No present-day examples are known, and this case seems kinematically impossible, since hinge convergence occurs only in the lower plate converges as well.

3.3.2. Kinematics #11

If the hinge is stationary (H = 0), neither stretching nor shortening should occur in the upper plate and the subduction rate would be negative (emerging or detaching slab). No present-day examples are known.

3.3.3. Kinematics #12

If the hinge diverges (H > 0) slower than lower plate (|H| < |L|), stretching should occur in the upper plate and the subduction rate would be negative (emerging or detaching slab). The Southern Apennines and the California (Basin and Range) subduction zones are present-day possible examples of this case. The Southern Apennines front does not show active compression, and the belt is dominated by extension. Deep seismicity is also very limited. In the second case the subduction of the Farallon plate was stopped when the oceanic ridge bordering its western margin was subducted. Then the WNW-ward faster moving Pacific plate west of the North America plate switched the lower plate convergence into divergence relative to U, and the North America Cordillera in the upper plate started to collapse (Doglioni, 1995). Therefore the Basin and Range is a post-subduction rift, and related to far field velocities of the involved plates. It is not a typical backarc basin.

3.3.4. Kinematics #13

If the hinge diverges (H > 0) as fast as the lower plate (|H| = |L|), then backarc extension is predicted to occur in then upper plate. Subduction rates should be null, i.e. the slab will not sink and will move laterally.

3.3.5. Kinematics #14

If the hinge diverges (H > 0) faster than the lower plate (|H| > |L|) then the subduction rate will be slower than the subduction hinge retreat (S < |H|) and extension is still predicted to occur in the upper plate (Fig. 14, 3W). This situation might be typical of some W-directed subductions and the northern Apennines and the Banda arc could be present-day examples.

In summary (Table 1), in seven cases the upper plate stretching is predicted theoretically, five along W-directed subduction zones (Fig. 14, 1W, 2W, 3W, 4W, 5W,), and two along E–NE-directed subduction zones (Fig. 15, 5E, 6E). Among these, two occur in post-subduction stages, i.e., 5W and 6E. Upper plate shortening is predicted in four cases. Three real cases are observed for real E- or NE-directed subductions (Fig. 15, 2E, 3E, 5E), and one for W-directed subduction (Fig. 14, 8W). In this latter case, the hinge zone that usually diverges from the upper plate, at later stages may slow down and converge (e.g., subduction flip, N-Japan), or due to external boundary condition (e.g., plate sub-rotation, such as Africa moving relatively N-ward and deforming the southern Tyrrhenian Sea, or South America moving relatively N-ward and deforming the southern Caribbean sea).

A striking feature of the above kinematic analysis is the theoretical prediction of kinematics of subduction zones in which the plate subducted at previous stages tends to move out from the mantle. However, only one potential case of emerging or detaching slab has been recognized.

Such subduction zone kinematic prediction is unexpected and contrasts one of the main paradigms on plate driving forces, i.e., the idea that density contrasts, in particular the negative buoyancy of subducting plates, provides the force (slab pull) to move horizontally the plates. In such a scenario, the escape of a subducted plate from the mantle seems absurd. However, if an alternative view is considered, this inconsistency can be eliminated. In such a scenario, subduction is a consequence of plate motions rather than their cause and the negative buoyancy of subducted slabs is a secondary effect of subduction and not the primary cause.

In five other cases the slab will be sinking in the mantle. This is the far more frequently occurring case, and examples for all five theoretical cases occur worldwide. In the three remaining theoretical cases the slab will be hanging and only for two cases real examples were recognized.

4. Accretionary prism and backarc spreading

By definition, the tectonic accretion is the transfer of mass from the lower to the upper plate, whereas the tectonic erosion is the transfer of mass from the upper to the lower plate. In case the accretionary prism is entirely formed at the expenses of the lower plate as it occurs along W-directed subduction zones, the shortening rate is equal to the subduction rate ($V_S = V_H - V_L$). A backarc basin forms when the subduction hinge retreats relative to the upper plate. Let us assume that there is a steady state convergence between upper (U=0) and lower plate (L=V_L), and steady state divergence of the subduction hinge (H=V_H), (Fig. 8).

When the decollement depth (z) of the accretionary prism is deep (Fig. 8, upper section), the transfer of mass to the area (A) of the accretionary prism is larger (e.g., Bigi et al., 2003) than for shallow z (Fig. 8, lower section). We may quantify this increase as a vector (V_{Ah}) indicating the rate of widening of A per unit time, being Ah the distance between H and the backarc border. As a consequence, the increase of width of the prism should generate a decrease of the backarc rifting widening rate (V_X), and a reference point (X) at the margin of the backarc rift should move more slowly toward the foreland. This would generate a smaller backarc area (B) (Fig. 8, upper section). We can express $A \propto \zeta \cdot B \propto V_X \cdot \delta A/\delta t = (V_H - V_L)z$. Therefore the speed of X is inversely proportional to V_{Ah} (prism widening rate), i.e., $V_X = (V_H - V_L)/V_{Ah}$, whereas Ah is in turn proportional to the depth of the decollement. The relation between A and Ah is variable as a function of the internal friction, the friction on the decollement (Davis et al., 1983), and the diffuse extension affecting the prism at the transition with the backarc.

Therefore, the backarc spreading V_X can be quantified as the velocity of H moving away from the upper plate, minus the prism widening rate, i.e., $V_X = V_H - V_{Ah}$ (Fig. 8). Zero accretion in the prism means $V_X = V_H$, and maximum backarc rifting. In conclusion, this kinematic analysis casts doubts on simplistic interpretations of GPS data as direct indicators of deep movements, i.e., between slab hinge and mantle; if there is accretion in the prism, surface movements cannot represent the hinge rollback. Moreover, evidences of mantle penetrating between the upper and lower plate have been interpreted along Pacific subduction zones where accretion is very low or absent. Suyehiro et al. (1996) show the presence of a serpentinitic diapir just in front of the forearc along the trench of the Izu–Bonin subduction zone. This volume prevents an equivalent spreading in the backarc as well.

The decrease in speed in backarc rifting settings might then have different complementary origins: i) related to a real slow down of the subduction hinge retreat relative to the upper plate; ii) related to the deepening of the decollement plane, or to a faster convergence rate and related increase of mass transfer to the upper plate; iii) in case of scarce or no accretion, the asthenospheric wedge may intrude above the subduction hinge. Normal faulting in the prism may further enlarge the upper plate close to the subduction hinge, inhibiting the widening of the backarc spreading.

In fact, in this model, among others, three main issues are neglected for sake of simplicity: a) the isostatic subsidence due to the load exerted by the wedge itself, b) the along strike lengthening generated by the arc growth typical of W-directed subduction zones, and c) the diffuse extension affecting the inner part of the accretionary prism.

As an application, a decrease or even a stop of the rifting has been proposed in the Tyrrhenian Sea backarc basin (D'Agostino and Selvaggi, 2004) in spite of normal fault-related seismicity (Pondrelli et al., 2004) and active convergence rate between the Ionian Sea and the upper plate (e.g., Sardinia). Also in the Marianas backarc basin the spreading rate has slowed down in the Quaternary (Yamazaki et al., 2003), but an alternative interpretation considers the transfer of mass from the lower to the upper plate or an asthenospheric intrusion along the subduction hinge.

5. The Apennines subduction

As an example, the kinematic analysis developed in the previous section is applied to the present movements

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of the Apennines arc. From the analysis of GPS data, along the Apennines subduction zone at least five different kinematic settings coexist, showing how a single subduction can have internal variable velocities as a function of the combination V_H and V_L. The late Oligocene to present Apennines–Tyrrhenian system formed respectively as the accretionary prism (e.g., Bally et al., 1986) and the backarc basin (e.g., Kastens et al., 1988) in the hangingwall of an arcuate W-directed subduction system. The downgoing lithosphere has variable thickness and composition (Adriatic continental, Ionian oceanic (?) and African continental, e.g., Calcagnile and Panza, 1981; Farrugia and Panza, 1981; IESG-IGETH, 1981; Calcagnile et al., 1982; Panza et al., 1982; Doglioni et al., 1999b; Catalano et al., 2001; Panza et al., 2003; Faccenna et al., 2004).

Moving along strike, due to the relevant anisotropies of the downgoing lithosphere (e.g., Calcagnile and Panza, 1981), the slab had and still has different behaviors, both in terms of seismicity, surface geology, and kinematics.

We analyzed the motions of selected GPS stations of Italy, using different database. We have chosen five main sections, characterizing most of the country area (Fig. 17). We used data from D'Agostino and Selvaggi (2004) for the sections 1, 3, 5, whereas for section 2 data come from the EUREF database of July 2006 (http://www.epncb.oma.be/). In section 4 (Fig. 17) we used data from Hollenstein et al. (2003) for the PORO station, whereas the velocity of point A in the Ionian Sea (18.50°E 38.3°N) is computed using Africa–Eurasia relative motion parameters from the REVEL plate kinematic model (Sella et al., 2002) (Fig. 17A). For a discussion on errors, see the related papers. We are aware that using different data sets may lead to inappropriate kinematic solutions. However for each single cross-section the data are extracted from the same solutions (apart section 4), and again, our goal is not yet to arrive to the exact determination of the velocity, but to show a methodological approach and to highlight how different kinematic and tectonic settings coexist along a single subduction zone. Error ellipses in the GPS data can be close to the estimated velocity, and we understand that these values might be minimum rates with respect to the real velocity field. Moreover, the location of H in our computation is ambiguous and could not represent the real velocity of the hinge, but the value we use

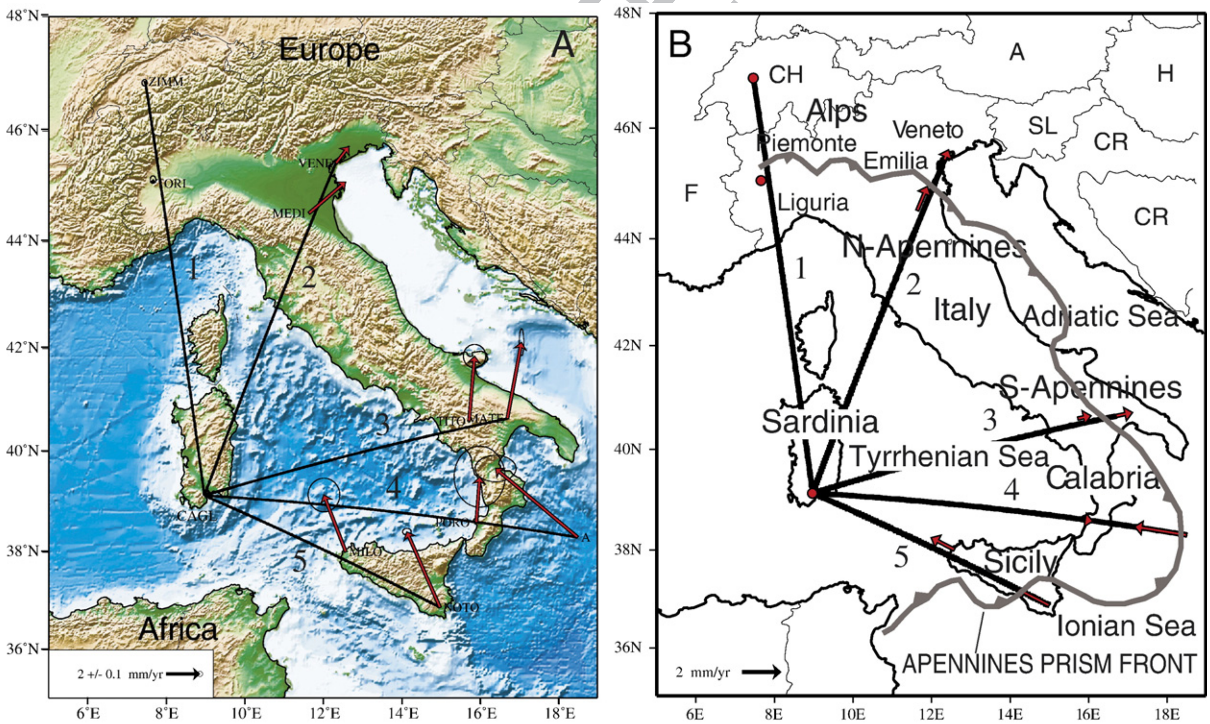

Fig. 17. A) Directions of selected GPS stations relative to Eurasia. Data are from three different databases. See text for explanation. The site of Cagliari (CAGL), which is on the Sardinia–Corsica continental boudin, is considered fixed with respect to Eurasia. B) Projections of velocity vectors on the sections connecting the stable CAGL with the other GPS stations across the Apennines subduction zone, reported in the following figure. U, upper plate; H, subduction hinge; L, lower plate; MEDI, Medicina; VENE, Venice; TITO, Tito; MATE, Matera; PORO, southwest Calabria; C, Ionian Sea, assumed fixed to Noto; MILO, Trapani; NOTO, Noto; TORI, Torino; ZIMM, Zimmerman.

can be considered as a minimum rate at which H moves. This analysis aims mainly at a methodological aspect.

In Fig. 17B are reported the projections on the sections of the velocity vectors. In this way, it is possible to compare relative components of the motion on the trajectories connecting the three GPS stations involved in the analysis. From the GPS data of Devoti et al. (2002), Oldow et al. (2002), Hollenstein et al. (2003), Battaglia et al. (2004), Serpelloni et al. (2005), the Sardinia plate (e.g., the Cagliari site) can be considered practically fixed relative to Eurasia. Cagliari is located in the Sardinia–Corsica micro-continent, a remnant of the upper plate boudinage in the backarc setting of the Apennines subduction (Gueguen et al., 1997). Therefore Cagliari–Sardinia can be used as reference for the upper plate of the subduction system.

From Sardinia to Liguria and south Piemonte in the northern Apennines, there are no significant movements, and both H and L can be considered fixed relative to U (Fig. 17, and Fig. 14-7W). Moving from Sardinia to the northeast, through Emilia and Veneto regions, the site of Medicina (e.g., Battaglia et al., 2004), that is assumed to be coherent with the subduction hinge, is moving away from the upper plate faster than sites located more to the northeast in the foreland, so that $V_H > V_L$ (Figs. 17 and 18, section 2). Ongoing extension is documented in the central-northern Apennines (Hunstad et al., 2003), is consistent with the positive value of V_H (i.e., the hinge migrates toward the foreland) that explains the active spreading of the Tyrrhenian backarc basin.

In the Southern Apennines the setting changes and, although the determination of the velocity of H it is not precise, the site used here would suggest a slower velocity of the diverging hinge with respect to the foreland, i.e., both H and L move away from the upper plate, but the lower plate L seems faster (Fig. 18-3). This setting would correspond to Fig. 14-5W, where the slab is paradoxically moving away with respect to the upper plate and subduction would result as negative (Figs. 17 and 18, section 3). This is in agreement with the paucity of geological and geophysical observations supporting active compression at the Southern Apennines front, while extension is widespread in the belt (Scrocca et al., 2005).

In a cross-section from Sardinia to Calabria and Ionian basin (Fig. 18-4), the setting changes again becoming similar to that of Fig. 14-1W. The Ionian (L), unlike the previous sections, is converging relative to U because it is moving with the Pelagian shelf south of Sicily, which is converging relative to Sardinia (Figs. 17 and 18, section 5).

During present times, it is doubtful if and how fast Calabria (H) is still moving eastward relative to Sardinia (U), in the frame of still active extension of the Tyrrhenian (Goes et al., 2004; D'Agostino and Selvaggi, 2004; Pondrelli et al., 2004). However seismicity and seismic reflection profiles suggest active, although slow,

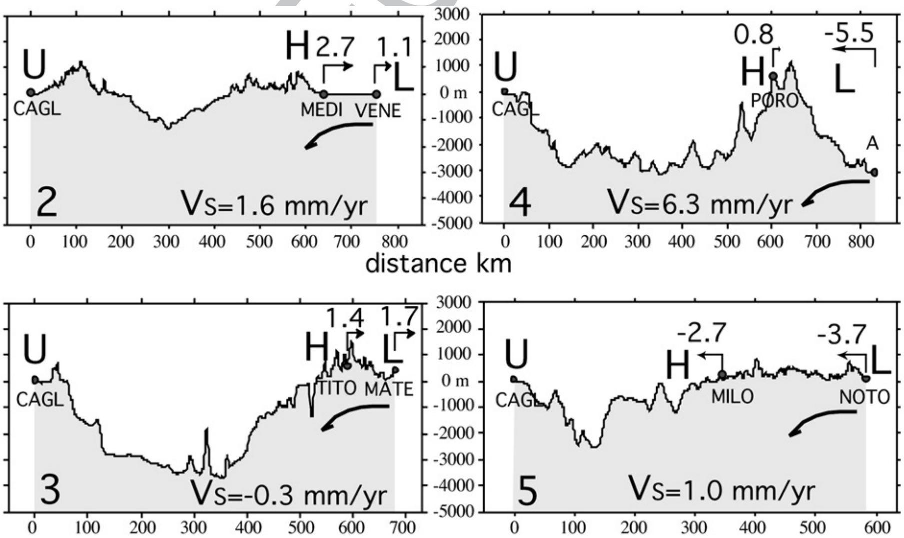

Fig. 18. Based on the data of the previous figure, along the Apenninic arc different relationships between U, H and L occur. The upper plate U is considered fixed (Sardinia–Corsica). The computed subduction rates are minimum estimates based on the approximation of a valid location of the subduction hinge H. Note that each section has different geodynamic settings and variable subduction rate, if any. The rates of H and lower plate L are in mm/yr. Section 1 is omitted since movements are close to null. Note that in section 3 the subduction is paradoxically negative, pointing for a detachment of the slab but from the surface. The variable kinematic settings are consistent with the seismicity of the area. Site and section locations in the previous figure. The fastest subduction rate is along the Calabrian arc (section 4).

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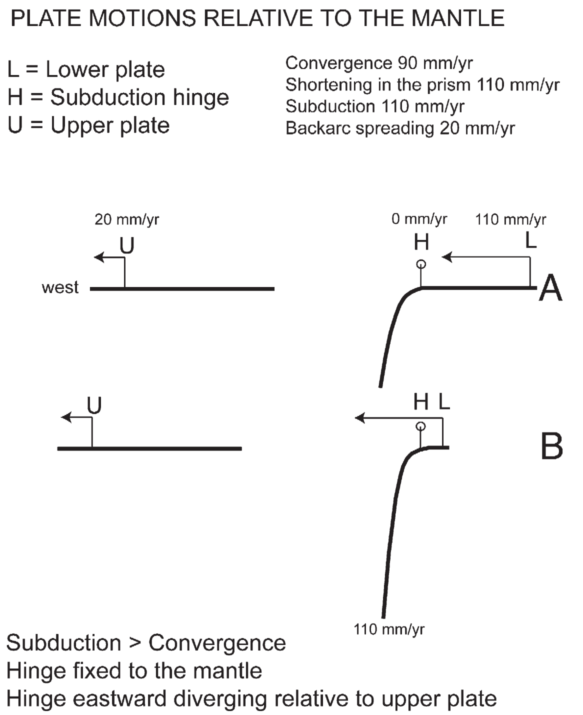

Fig. 19. Cartoon showing a two stages (A, B) absolute kinematics of a W-directed subduction zone. The slab is anchored in the mantle, and the subduction rate is faster than the convergence rate. The subduction hinge H is fixed relative to the mantle but is moving east relative to the upper plate.

extension in the Tyrrhenian Sea (Scrocca et al., 2003; Doglioni et al., 2004; Pondrelli et al., 2004; Chiarabba et al., 2005). The Ionian (L) is converging relative to U because it is moving with the Pelagian shelf south of Sicily, which is converging relative to Sardinia.

An increase of accretion in the Ionian prism of the Apennines should provide an upper plate crustal thickening and widening, partly preventing the backarc extension, as previously discussed. Along this section there is the fastest subduction rate (Figs. 8 and 18, section 4).

From Sardinia to Sicily, and from Sicily foreland to the south, the African vectors (V_L) move toward Cagliari (Devoti et al., 2002; Goes et al., 2004), but in northern Sicily (H) the movement is slower than in the foreland to the south (Battaglia et al., 2004). This is the setting of Fig. 18, section 5, and Fig. 14, 8W, where H is rather approaching U, with a slower velocity than the L, i.e., $-V_H < -V_L$. This is consistent with the compressive seismicity both south and north of Sicily (Chiarabba et al., 2005). The backarc setting along this section is then shrinking.

However, in the southern Tyrrhenian Sea, different settings may coexist in a single area due to the 3D nature of the subduction. For example, in northern Sicily, convergence of the hinge relative to the upper plate in a NW–SE section concurs with an E–W extension related to the divergence of the hinge in Calabria. Therefore compressive–transpressive tectonic features overlap with extensional–transtensional faults in the southern Tyrrhenian Sea. This is consistent with seismic section interpretation (Pepe et al., 2005) and seismicity of the area (Pondrelli et al., 2004; Chiarabba et al., 2005).

6. Basic kinematics in the mantle reference frame

We can now use the mantle as a reference for the motion of the three points located in the upper plate, lower plate and subduction hinge (Figs. 19 and 20). Using the mantle as a reference needs either to adopt a hotspot reference frame, or to infer other fixed mantle frames from geological indicators. The dilemma of choosing the most acceptable reference frame, if any, is out of the scope of this paper. After this preamble, we use a classic hotspot reference frame such as the one proposed by Gripp and Gordon (2002). In this reference, the lithosphere is moving relative to deep hotspots, and

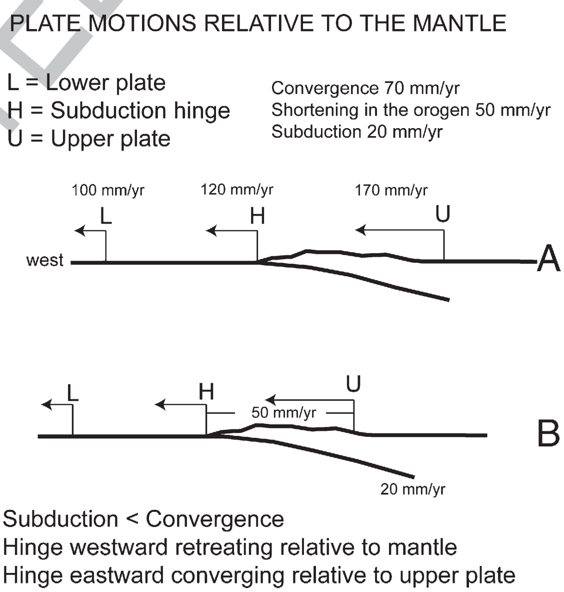

Fig. 20. Cartoon showing two stages (A, B) absolute kinematics of an E- NE-directed subduction, which has a slower subduction rate with respect to the convergence rate. The trench or subduction hinge H is moving west relative to the mantle but is moving east relative to the upper plate. Therefore in both W- and E- NE-directed subduction zones, the hinge migrates eastward relative to the upper plate. Larger the shortening in the orogen, lower the strength of the upper plate and lower the coupling between upper and lower plates. The convergence/shortening ratio in this example is 1.4 and is function of the lithospheric strength. From this analysis, plate motions are not controlled by subduction rate, but vice versa.

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it has a net “westward” rotation of about 0.44 (±0.11) deg Myr^{-1}, the so-called westward drift of the lithosphere (Bostrom, 1971; Moore, 1973). Plate motions in the no-net-rotation frame such as Heflin et al. (2005), can be transferred into the net rotation frame indicated by the motion of the lithosphere relative to hotspots. Recently a debate arose regarding whether hotspots are deep or shallow features (Foulger et al., 2005), and if they can represent a fixed reference frame (e.g., Norton, 2000). In case it will be proven they are shallow sourced, the net rotation of the lithosphere can be much faster than presently estimated (Doglioni et al., 2005b). Both in the deep and shallow source hypothesis, the W-directed subduction zones have hinges fixed or anchored to the mantle (Fig. 19). Along the opposite E- or NE-directed subduction zones, due to the global polarization, the hinge is moving west or southwest (Fig. 20). For example, the Chile trench, along the Andean subduction zone, shows a hinge moving west relative to the mantle, while it moves eastward relative to the upper plate.

When analyzing the motion of the Nazca and South America plate in the two reference frames (i.e., deep and shallow hotspots), Nazca moves eastward relative to the mantle in the deep hypothesis, whereas it would move westward in the shallow interpretation (Cuffaro and Doglioni, in press, Fig. 21).

Along the Hellenic subduction zone, the former speculation seems to be valid regardless the deep and shallow models (Fig. 22). In this last case, the slab would move out of the mantle, suggesting that it is escaping from the mantle, but it subducts because the upper plate is moving westward or southwestward faster than the lower plate (Fig. 23). This kinematic constraint is in a way indicating a paradox, i.e., along subduction zones the slab can even move upward relative to the mantle. This could be the rule along E- NE-directed subduction zones in the shallow hotspot reference frame hypothesis. But, more importantly, in the meantime it shows that the slab pull cannot be the main cause for subduction and the energy source of plate tectonics.

The subduction almost equals the relative motion of the upper plate over the lower plate at the trench. In fact, in the example of Fig. 23, the upper plate overrides the lower plate at only 10 mm/yr at the trench, in spite of a convergence of 80 mm/yr. Moving from the base of the slab upwards, the relative motion between upper and lower plates decreases because part of the converge is adsorbed by the upper plate shortening (Fig. 23). Note that in these shallow settings there should occur the slowest subduction rates. It is interesting that the tectonic erosion (e.g., von Huene and Lallemand, 1990; Ranero and von Huene, 2000) has been described more frequently along this type of geodynamic environment. Tectonic erosion means that the basal decoupling surface of the orogen is in the upper plate rather than in the lower plate when accretion occurs. This is also logical during oceanic subduction because of the higher strength of the oceanic rocks with respect to the continental crust. In these settings, most of the shortening is in fact concentrated in the upper plate.

7. The Hellenic subduction

Along the Hellenic trench the Africa plate subducts NE-ward underneath Greece. The Hellenic subduction zone is active as evidenced by seismicity, volcanism, space geodesy data, etc. (Innocenti et al., 1982, 1984; Christova and Nikolova, 1993; Clément et al., 2000; Doglioni et al., 2002). Along slab, seismicity shows down-dip extension (Papazachos et al., 2005). Using the hotspot reference frame of Gripp and Gordon (2002), Africa is moving west slower than Greece. This is even more evident in the shallow hotspot reference frame, but the directions are more southwestward directed (Fig. 22). Therefore, regardless the mantle reference frame, Africa moves relatively west or southwest, in the opposite direction with respect to the direction of subduction under the Hellenic trench. Therefore, the mantle should move eastward with respect to the slab, i.e., the slab would move west or southwestward relative to the mantle. The deep seismicity of the slabs under the Apennines and the Hellenides shows opposite behavior, i.e., the slab is under down-dip compression along the W-directed subduction zone (Frepoli et al., 1996), and down-dip extension along the Hellenides (Papazachos et al., 2005). These kinematics (Fig. 24) suggest that the slab under the Apennines resists the sinking, whereas the Hellenic slab is pulled either from below (slab pull) or from above (W-ward drift of the lithosphere). The relative eastward mantle flow would be consistent with these observations, pushing down the Apennines slab, and moving eastward with respect to the Hellenic slab, attached to the Africa plate which is due west.

A mantle flowing to the east or northeast in the Mediterranean is consistent with a number of observations, such as the shear-wave splitting analysis (Margheriti et al., 2003; Barruol et al., 2004), S-wave mantle tomography (Panza et al., 2007), the fast eastward Neogene to present retreat of the Apennines–Maghrebides subduction zone, the shallow dip and depth of the Hellenic slab vs. the steeper and deeper W-ward directed Apennines subduction, plus another number of geological asymmetries such as the shallow vs. the deep trench or foreland basin, the shallow vs. the steep regional monocline, etc. (Doglioni et al., 1999b).

In this interpretation, the Hellenic subduction is moving out of the mantle (Fig. 23), no matter the hotspot reference frame. Subduction occurs because the upper plate (Greece) is overriding the lower plate Africa faster. This setting appears as a prototype of a subduction not driven by the negative buoyancy of the slab because the kinematic observation is not biased by the reference frame, and seems consistent with the corollary of geological and geophysical observations.

8. What drives subduction?

One paradigm of plate tectonics relates the dip of the slab to the buoyancy of the downgoing lithosphere along subduction zones, being the negative buoyancy proportional to the age of the oceanic lithosphere, length of the slab and length of the trench (Forsyth and Uyeda, 1975; Jarrard, 1986; Anderson, 2001; Conrad and Lithgow-Bertelloni, 2003). Carlson et al. (1983) found an important correlation between the absolute velocity of plates and the age of the oceanic crust, pointing out for a

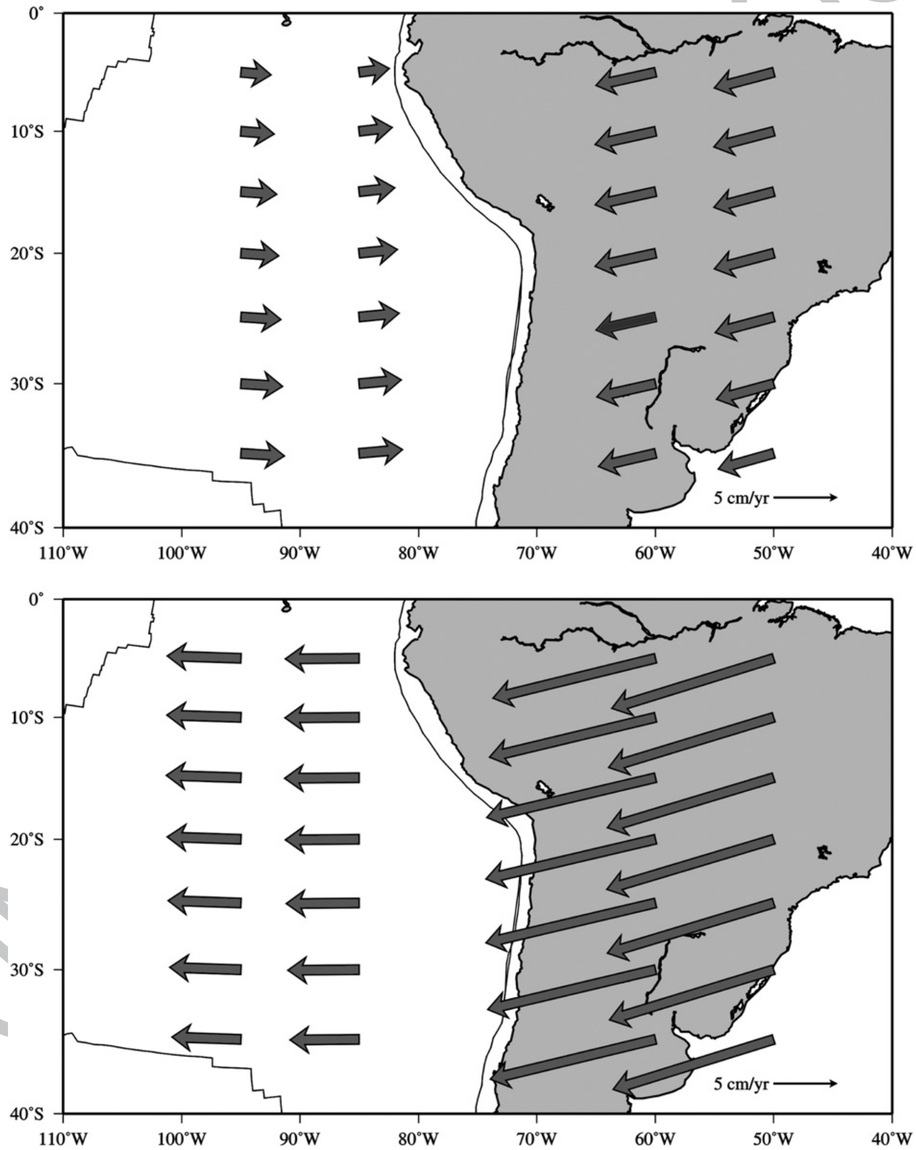

Fig. 21. Plate motions of South America and Nazca relative to the mantle in the deep (above) and in the shallow hotspot reference frames (below). Note that in the shallow hotspot reference frame, the Nazca plate is moving westward faster than the underlying mantle, escaping from the subduction.

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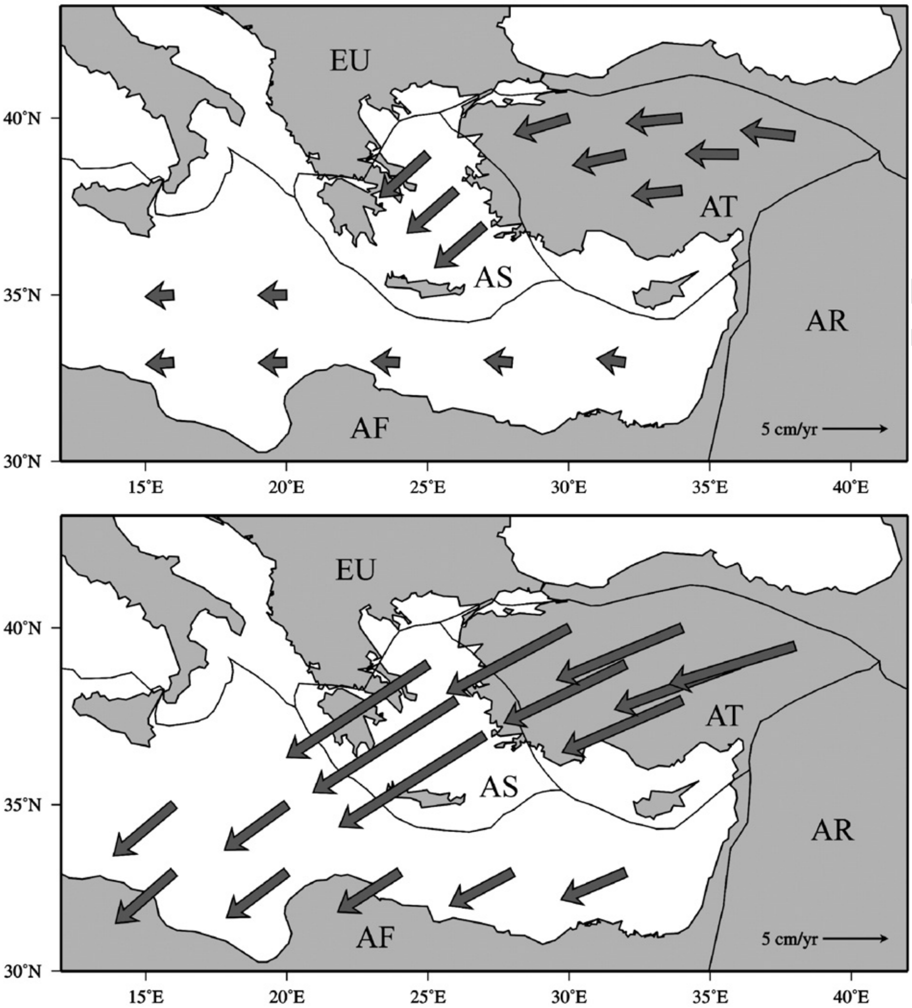

Fig. 22. Plate motions of Africa, Greece and Anatolia relative to the mantle in the deep (above) and in the shallow hotspot reference frames (below). Note that in both reference frames, Africa is moving westward faster than the underlying mantle, escaping from the subduction. This setting refers to the case of the frame 5E of Fig. 15.

fundamental role of the slab pull as the driving mechanism: the older and cooler oceanic lithosphere should result in a denser and faster sink along the subduction zone driven only by its negative buoyancy. However there is no correlation between convergence rates and age of the oceanic crust (Fig. 25). Moreover, when plate motions are computed in the deep and in the shallow hotspot reference frame, few plates move out of the trench along E–NE-directed subduction zones (Figs. 21 and 22), indicating only the negative buoyancy of the slab does not drive that subduction. In their analysis, Carlson et al. (1983) have a strong weight of the Pacific plate which is the fastest plate surrounded by a number of subduction zones (Marianas, Izu–Bonin, Japan, Kurili, Tonga, Kermadec). But is the plate fast because driven by the slab pull at the margin, or are there other forces acting on the plate and the subduction is a consequence of this speed? It is worth noting that the Pacific plate has the lowest asthenosphere viscosity values (Pollitz et al., 1998), i.e., it is the most decoupled plate with respect to the asthenosphere. Moreover the Pacific plate has a single angular velocity, it has no relevant internal strain, but nevertheless is treated by Carlson et al. (1983) as several independent plates to which different slabs are attached. Along the regression line computed in their research, W-directed subduction zones are generally faster than the opposite E to NNE-directed slabs. This is predicted by the simple aforementioned kinematics. Moreover, due to the "westerly" directed rotation of the lithosphere, the plates located to the west of a rift (e.g., the Pacific plate to west of the East Pacific Rise) tend to be preserved longer. For example

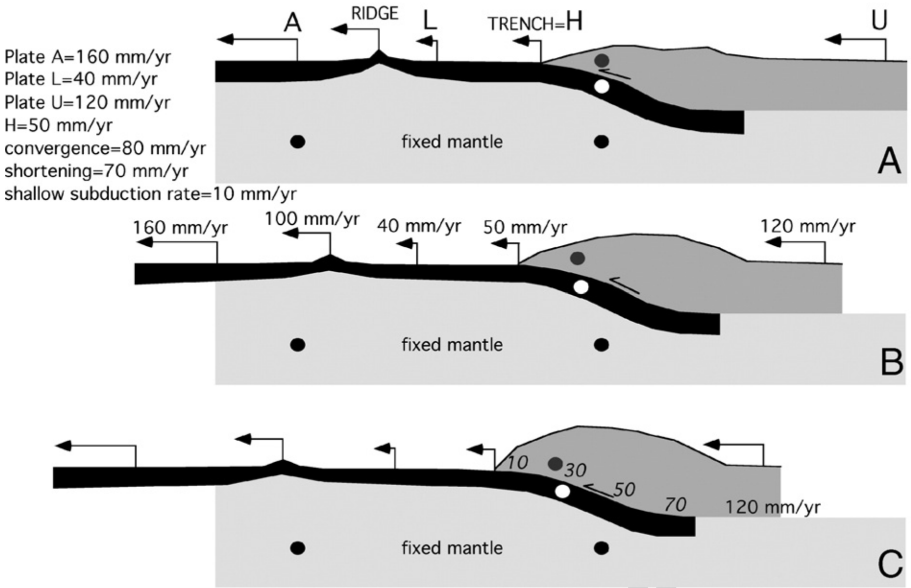

Fig. 23. When plate motions are considered relative to the hotspot reference frame, i.e., assuming fixed the mantle, the slabs of E- or NE-directed subduction zones may move out of the mantle. In the three stages sketch (a, b, c), the white circle moves leftward relative to the underlying black circle in the mantle. Subduction occurs because the upper plate dark gray circle moves leftward faster than the white circle in the slab. In this model, the slab moves west at 40 mm/yr relative to mantle. The subduction rate is the convergence minus the orogenic shortening. With different velocities, this seems to apply to the Hellenic subduction and, in the shallow hotspot reference frame, also to the Andean subduction. This kinematic evidence of slabs moving out of the mantle casts doubts on the slab pull as the driving mechanism of plate motions. In the lower section, the numbers in italic from 10 through 70 indicate the relative velocity in mm/yr between the upper and the lower plate. Note that the subduction rate should increase with depth, where the upper plate shortening is decreasing. This is consistent with the down-dip tension seismicity of this type of subduction.

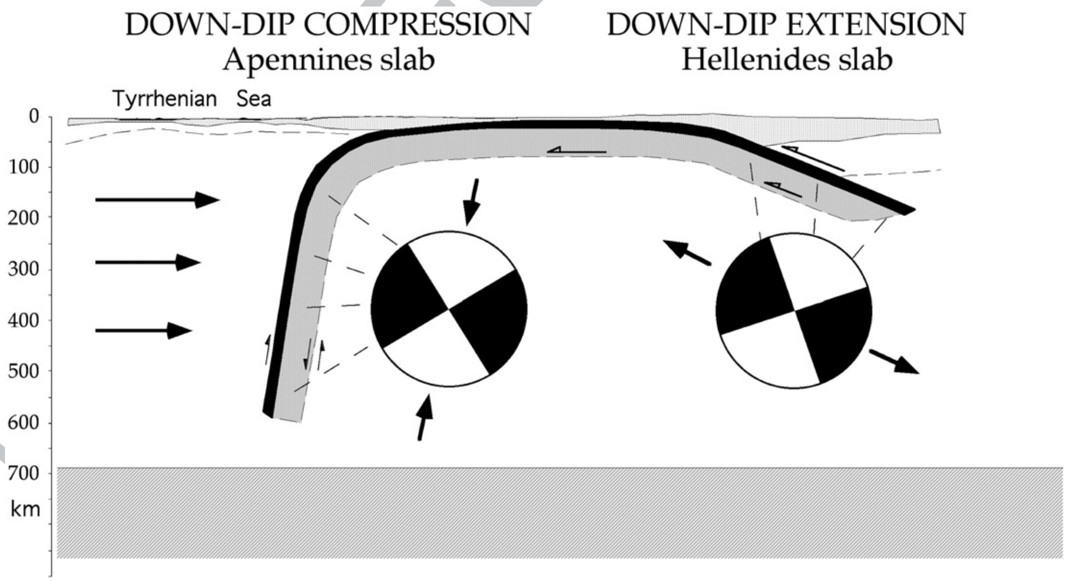

Fig. 24. The deep seismicity along the Apennines and Hellenides slab shows opposite behavior, being steeper and deeper vs. shallower and less inclined respectively. Moreover the Apennines slab is undergoing down-dip compression (Frepoli et al., 1996), whereas the Hellenic slab suffers down-dip extension (Papazachos et al., 2005). This opposite behavior mostly occurs also comparing the western and the eastern margins of the Pacific subduction zones. This asymmetry is consistent with the W-ward drift of the lithosphere relative to the mantle.

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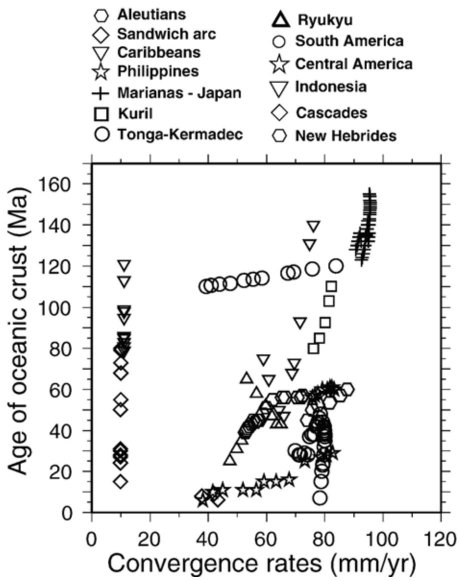

Fig. 25. Age of oceanic lithosphere entering the trench (after Mueller et al., 1997) vs. velocity of convergence calculated using the NUVEL1A (DeMets et al., 1994) rotation poles. The diagram shows a plot obtained for 13 subduction zones. Note the absence of correlation suggesting no significant relation between plate motions speed and negative slab buoyancy. Data taken after Cruciani et al. (2005).

the plates in the hangingwall of the western margin of the Pacific ocean also move west, but in the eastern margin, since again both upper and lower plates move west, the lower plate will be gradually overridden by the upper plate. This could explain why the age of the Pacific Ocean is older in the western side, and the lithosphere is characterized by the oldest ages along the western trench.

We have seen in the previous sections that the subduction rate depends both on the absolute velocity of the lower and upper plates, plus the velocity and direction migration of the subduction hinge. Therefore the absolute velocity of the lower plate does not provide direct information on the subduction rate, which is considered in the slab pull model only constrained by the negative buoyancy.

Several analogue and finite element models have been carried out in order to reproduce subduction mechanisms (e.g., Sheena, 1993; Regard et al., 2003). However all these models a priori impose a denser lithosphere with respect to the underlying mantle. This assumption is questionable because there are no direct density constraints on the hosting mantle, which is considered denser than the lithosphere in all models at least below the asthenosphere (Dziewonski and Anderson, 1981; Kennett et al., 1995).

The larger negative buoyancy has been invoked to explain the steeper dip of the western Pacific subduction zones because the subducting western Pacific oceanic lithosphere is older, cooler and therefore denser. However, the real dip of the slabs worldwide down to depths of 250 km shows no relation with the age of the downgoing lithosphere (Cruciani et al., 2005; Lallemand et al., 2005). In fact there are slabs where moving along strike the age of the downgoing lithosphere varies, but the dip remains the same (Barbados), or vice versa, the age remains constant while the dip varies (Philippines). There are cases where the age decreases and the dip increases (W-Indonesia), and other subduction zones where the age increases and the dip decreases (Sandwich). This shows that there is not a first order relationship between slab dip and lithospheric age. This suggests that supplemental forces or constraints have to be accounted for, such as thickness and shape of the hangingwall plate, relative and absolute plate velocity, presence of lateral density variations in the hosting upper mantle, effects of accretion/erosion, subduction of oceanic plateaus and slab deformation due to the motion of the mantle relative to the subducting plate (Cruciani et al., 2005). Seismicity illuminating the slab geometry is strongly influenced by the composition, thermal state and velocity of the downgoing plate (e.g., Carminati et al., 2005).

In the following sections we will analyze the alternative mechanisms that have been proposed to drive plate tectonics, namely slab pull, mantle convection and the forces generated by the Earth's rotation.

8.1. On the efficiency of the slab pull

The negative buoyancy of slabs should determine the pull of plates, but it has been shown that the dip of the subduction zones is not correlated with the age and the thermal state of the down going plates (Cruciani et al., 2005). Moreover relative convergence rates at subduction zones do not correlate with age of oceanic lithosphere at the trench (Fig. 25). One statement used to corroborate the slab pull is the trench length with respect to the plate velocity (Forsyth and Uyeda, 1975). On the basis of a similar observation it could be argued that fast spreading ridges generate fast plate motions. However these relations may be a circular reasoning, e.g., longer subduction zones and faster ridges form when plates move faster (Doglioni et al., 2006b). The relationship between trench length and plate velocity is also questionable for other reasons; for example the absolute plate velocity can be recomputed either in the deep hotspot (Fig. 26, A) or in the shallow hotspot reference frame (Fig. 26, B), and the different results do not support a correlation between slab length percentage (length of the trench with respect to the length of the boundary surrounding the plate) and plate velocity.

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The relationship between the area of plates (Schettino, 1999) and the angular velocity of plates in the deep hotspot reference frame (HS3-NUVEL1A, Gripp and Gordon, 2002) shows no correlation (Fig. 26, C), as already observed by Forsyth and Uyeda (1975). However, when plotting the area vs. the absolute angular velocities of plates in the shallow hotspot reference frame, a correlation seems to be present, being bigger plates generally faster (Fig. 26, D). When comparing similar size of plates, the oceanic or continental plus oceanic, oceanic plates travel faster than purely continental.

Both analyses do not support a significant correlation casting more doubts on the importance of the slab pull, which has a number of further counterarguments. For example, the assumption that the slab is heavier than the country mantle remains debatable, particularly because there are not constraints on the composition of both slab and mantle at variable depth (e.g., the amount of Fe in the lower asthenosphere and the lower upper mantle). Is the slab pull the energetic source for plate motions? Is it large enough? Is it correctly calculated? Are the assumptions reliable? Most of the literature indicates that the slab pull is about 3.3×10^{13} N m^{-1} (e.g., Turcotte and Schubert, 2002). This is a force per unit length parallel to the trench. However this value is very small when compared to other energetic sources for Earth, such the energy dissipated by tidal friction, heat flow emission, and Earth's rotation (e.g., Denis et al., 2002). Moreover the slab pull would be even smaller if chemical and mineralogical stratification are introduced in the upper mantle. Most of the Earth's volcanism is sourced from above 200 km: the subduction zones release magmatism at about 100–150 km depth (Tatsumi and Eggins, 1995); mid-oceanic ridges are sourced by even shallower asthenosphere melting (100–30 km, e.g., Bonatti et al., 2003); hotspots are also debated as potentially very shallow, and sourced by the asthenosphere (Bonatti, 1990; Smith and Lewis, 1999; Doglioni et al., 2005b; Foulger et al., 2005). Since even xenoliths in general and kimberlite chimneys originated at depth not deeper than the asthenosphere, we have no direct sampling of the composition of the standard lower part of the upper mantle. Therefore we cannot exclude for example a more Fe-rich fayalitic composition of the olivine, heavier and more compacted than the Mg-rich olivine (forsterite, which is presently assumed as the more abundant mineral of the upper mantle. In case more iron is present in the upper mantle olivine, the density of the ambient mantle would be slightly higher, making the slab pull smaller, if any. The slab pull concept is based on the hypothesis of a homogeneous composition of the upper mantle, with the lithosphere sinking only because it is cooler (e.g., Turcotte and Schubert, 2002). However, the oceanic lithosphere is frozen shallow asthenosphere, previously depleted beneath a mid-oceanic ridge. Depleted asthenosphere is lighter than the "normal" deeper undepleted asthenosphere (see Oxburgh and Parmentier, 1977; Doglioni et al., 2003, 2005a,b for a discussion). Therefore the assumption that the lithosphere is heavier only because it is cooler might not be entirely true, and the slab pull could be overestimated. Phase transitions within the subducting lithospheric mantle would enhance the slab pull in the transition zone (300–400 km; Stern, 2002; Poli and Schmidt, 2002), but again, the occurrence of higher density country rocks due to chemical and not only phase transitions could make the effect of the slab pull smaller and smaller. Moreover, the occurrence of metastable olivine wedges in fast subducting oceanic lithosphere is considered to create positive density anomalies that should counteract the effects of slab pull (Bina, 1996). A further density anomaly that is suggested to drive slab pull is expected to come from the eclogitization of the subducting oceanic crust. This process involves only a thin layer (5–8 km thick) and not the entire downgoing lithosphere (70–90 km thick). Nevertheless, this type of metamorphic transition is often assumed to be able to determine the slab pull. The eclogites reach densities of about 3440–3460 kg m^3 only at depths of about 100 km (Hacker et al., 2003; Pertermann and Hirschmann, 2003). The density of the country mantle at comparable depths according to the PREM model is 3370 kg m^3 (Anderson, 1989), i.e., only slightly lighter than the eclogitized oceanic crust. Both eclogite and mantle densities are quite speculative. The small density contrast between subducting crust and country mantle casts doubts on the potential effect of the negative buoyancy of oceanic crust. Therefore we do not have hard constraints on the depth at which the slab pull should turn on and at what depth it should turn off since the mineralogy of the slab and the hosting mantle is still largely unknown. Why then a slab should maintain its shape and coherence down to the 670 km discontinuity? The easiest explanation would be its higher stiffness. Since seismic wave velocity is inversely proportional to density, the high velocity of the slab detected by tomography could be related not to its higher density, but to its higher rigidity. Certainly the slab becomes heavier during sinking for phase transformations, but is it a priori denser, or does it become heavier on the way down? Is it continuously reaching density equilibrium while moving down (Doglioni et al., 2006b)?

Trampert et al. (2004) have recently demonstrated that low velocity volumes of the mantle detected by

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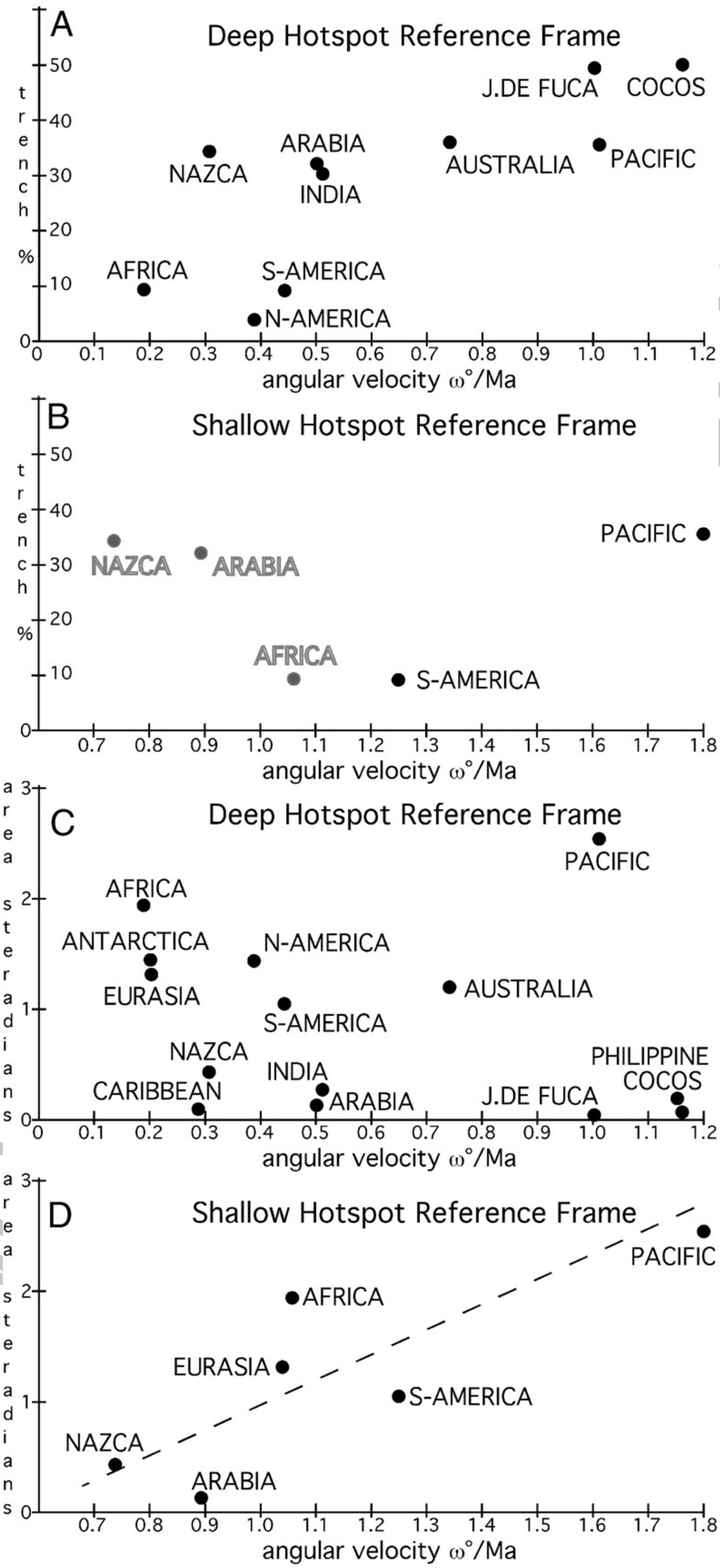

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tomography can be due to lateral variations in composition rather than in temperature, i.e., they can be even higher density areas rather than hotter lighter buoyant material as so far interpreted. In fact, considering the main low velocity zones in the mantle such as the asthenosphere or the liquid core, their decrease in speed of the P waves is related to their lower rigidity (e.g., Secco, 1995) either generated by CO_2 content in the asthenosphere, or higher density — low viscosity iron alloys in the liquid core. As extreme examples, gold or lead have high density but low seismic velocity. Therefore the interpretation of tomographic images of the mantle where the red (lower velocity) areas are assumed as lighter and hotter rocks can simply be wrong, i.e., they may even be cooler and denser (Van der Hilst, 2004). With the same reasoning, blue (higher velocity) areas, which are assumed as denser and cooler rocks may even be warmer and lighter.

Trampert et al. (2004) also suggest that the low velocity in the lower mantle could for example be due to higher concentrations in iron. Minerals containing more iron are more conductive, and at that depth the coefficient of thermal expansion must be very low. Both factors decrease the Rayleigh number, making the convection very sluggish (e.g., Anderson, 2002). The onion structure of the Earth with shells compositionally homogeneous (e.g., the PREM, see Anderson, 1989) is a misleading oversimplification, since the occurrence of lateral heterogeneities in the whole Earth layers has been widely demonstrated.

The main geometric, kinematic and mechanical counterarguments on the slab pull as the primary mechanism for moving plates and for triggering subduction are listed:

1) The dip of the slab is independent from the age of the oceanic lithosphere (Cruciani et al., 2005). Therefore, the supposed larger negative buoyancy determined by the cooler oceanic lithosphere does not control the slab dip.
2) We have no hard constraints of the real composition of the upper mantle: there could be more fayalite, making the upper mantle more dense and the slab negative buoyancy smaller.
3) Subduction processes involve also continental lithosphere descending to depths deeper than 100–150 km (Ampferer, 1906; Dal Piaz et al., 1972; Trümpy, 1975; Ranalli et al., 2000; van Hinsbergen et al., 2005), although subducted average continental crust is most probably buoyant with respect to mantle rocks (Hermann, 2002).
4) The oceanic lithosphere is frozen shallow (30–100 km deep) asthenosphere, previously depleted below ridges. Therefore the oceanic lithosphere is the differentiated lighter upper part of the mantle. Then why should it be a priori heavier than the undepleted deeper (100–300 km) asthenosphere? A pyrolite density of 3400 kg m^3 in the asthenosphere lying beneath the old oceanic lithosphere has been inferred (Jordan, 1988; Kelly et al., 2003). Moreover, hydrothermal activity generates serpentinization of the mantle along the ridge that decreases even more the density.
5) If oceanic lithosphere is heavier than the underlying mantle, why are there no blobs of lithospheric mantle (LID) falling in the upper mantle below the western older side of the Pacific plate?
6) Within a slab, eclogitization is assumed to make the lithosphere denser. However, eclogitization is concentrated in the 6–8 km thick oceanic crust, whereas the remaining 60–80 km thick lithospheric mantle does not undergo the same transformation. Therefore only 1/10 of the slab is apparently increasing density, but the main mass of the slab (90%) does not change significantly.
7) The density increase due to eclogitization is in contrast with the exhumation of the eclogitic prism that is usually detached with respect to the "lighter" lithospheric mantle (G.V. Dal Piaz, pers. comm.).
8) Why the lithosphere should start to subduct? This crucial point arises particularly when considering an oceanic hydrated and serpentinized lithosphere that has not yet been metamorphosed by the

Fig. 26. A) Relationship between absolute plate motions angular velocity vs. trench percent in the deep hotspot reference frame. There is not evident correlation between the two values. For example, the Nazca and Pacific plates have about the same percentage of trench length with respect to the plate circumference, but the Pacific is much faster. Angular velocities after Gripp and Gordon (2002). B) Relationship between absolute plate motions angular velocity vs. trench percent in the shallow hotspot reference frame. The absence of correlation between the two values is even more evident, but the gray points on the left show a negative motion of the plates, i.e., away from the trench. Therefore, in this reference frame, plates cannot be moved by the slab pull. Angular velocities after Crespi et al. (2007). C) Plot of plate areas and absolute angular velocities in the deep hotspot reference frame. Areas of plates after Schettino (1999). As already shown by Forsyth and Uyeda (1975), no relation is observable. D) Plot of plate areas and absolute angular velocities in the shallow hotspot reference frame. Angular velocities after Crespi et al. (2007). Unlike the previous figure, a correlation seems to exist, i.e., larger plates move faster, even if oceanic plates still move relatively faster than continental plates for comparable areas.

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subduction process, and consequently it is still less dense (G.V. Dal Piaz, pers. comm.).

9) Down-dip compression affects most of the slabs, all below 300 km (Isacks and Molnar, 1971), most of them even at shallower depth (e.g., Frepoli et al., 1996), pointing out for a slab forced to sink rather than actively sinking.
10) The 700 km long W-Pacific slab, where only the upper 300 km show some potential down-dip extension seismicity (but it could be generated also by horizontal shear in the mantle, Giardini and Woodhouse, 1986) should pull and carry the 10,000 km wide Pacific plate, 33 times bigger, winning the shear resistance at the plate base, and the opposing basal drag induced by the relative eastward mantle flow inferred from the hotspots migration (Doglioni, 2006b).
11) Kinematically, subduction rollback implies that the volumes left in the hangingwall of the slab have to be replaced by horizontal mantle flow, whether this is a consequence or the cause of the retreat (Doglioni et al., 1999b). However, in order to allow the slab to move back, the slab retreat needs that also the mantle in the footwall of the slab moves away in the direction of the slab retreat. This is true regardless this motion is generated by the slab pull or it is an independent mantle horizontal flow. But the energy required to push forward the mantle is much greater than the slab pull can effort. Where there is no convergence or rather divergence occurs between upper and lower plates, the slab pull has been postulated as the only possible driving mechanism. However the slab pull has not the energy to push back eastward the whole section of mantle located east of the slab, in order to allow the slab rollback. A relative eastward motion of the mantle would be much more efficient in terms of scale of the process and mass involved, to generate the eastward slab hinge retreat, determining active subduction without plates convergence (e.g., Apennines, Barbados).
12) Are plates surrounded by long slabs and trenches faster? It might be a circular reasoning because long subduction zones might be a consequence of fast movements of plates. Moreover plates are considered fast in the no-net-rotation (NNR) reference frame (Conrad and Lithgow-Bertelloni, 2003). For example, measuring plate motions in the hotspot reference frame, i.e., relative to the mantle, Nazca is very slow relative to mantle, so the relation between plate velocity, slab age and length of a subduction zone is not that simple.
13) Some plates in the hotspot reference frame move without any slab pulling them, e.g., the westward movements of North America, Africa and South America (Gripp and Gordon, 2002). Trench suction has been proposed to explain these movements, but beneath both North and South America the mantle is relatively moving eastward, opposite to the kinematics required by the trench suction model.
14) Plate velocities in the hotspot reference frame seem to be inversely proportional to the viscosity of the asthenosphere rather than to the length of the subduction zones and the age of the downgoing lithosphere. In fact the Pacific, which is the fastest westerly moving plate (Gripp and Gordon, 2002), has the lowest viscosity values (Pollitz et al., 1998).
15) The horizontal velocity of plates is 10–100 times faster than vertical velocity (subduction related uplift or subsidence) suggesting that vertical motions are rather passive movements. Moreover, the kinematic analysis of section 3 shows that subduction rates appear controlled by rather than controlling horizontal plate motions.
16) The energy for shortening an orogen is probably larger than the one supposed for the slab pull.
17) When describing the plate motions relative to the mantle, e.g. in the hotspots reference frame, along E- or NE-directed subduction zones the slab might move out of the mantle, e.g., in the opposite direction of the subduction. It is sinking because the faster upper plate overrides it.
18) There are rift zones formed between plates not surrounded by oceanic subduction to which the pull for moving the lithosphere can be attributed (e.g., the Red Sea).
19) Although the knowledge of the rheological behavior of subducted lithosphere is very poor, it can be conjectured that the downgoing slab, being progressively heated, could potentially lose strength, diminishing the possibility to mechanically transfer the pull (Mantovani et al., 2002).
20) The folding and unfolding of the lithosphere at the subduction hinge makes the slab even weaker for supporting the slab pull.
21) Slab pull has been calculated to be potentially efficient only at a certain depth (e.g. 180 km, McKenzie, 1977); and shallower than that? How does subduction initiate?
22) At the Earth's surface, oceanic lithosphere has low strength under extension (e.g., 8×10^{12} N m^{-1}, Liu et al., 2004) and is able to resist a force smaller than that requested by slab pull (3.3×10^{13} N m^{-1},

Turcotte and Schubert, 2002). If the slab pull is the cause for the motion of the Pacific plate, this observation argues for a stretching of the Pacific lithosphere before slab pull being able to move the plate. In other words, the plate cannot sustain the tensional stresses eventually due to slab pull. The low lithospheric strength problem could be, however, partly counterbalanced by the mantle flow and viscous tractions acting on the plates induced by slab sinking (e.g., Lithgow-Bertelloni and Richards, 1998). Due to low temperature and high pressure, the strength of subducted oceanic lithosphere rises to some 2×10^{13}–6×10^{13} N m^{-1} (Wong A Ton and Wortel, 1997) and would make sustainable the eventual pull induced by density anomalies related to phase changes at depth. In summary the subducted slab is probably able to sustain the load induced by slab pull but probably this load cannot be transmitted to the unsubducted portion of the plate without breaking it apart.

This long list casts doubts on the possibility that the slab pull can actually trigger subduction, slab rollback, and drive plate motions. Density anomalies due to phase changes occurring at depth within the slab could enhance the sinking of the slab. However, the slab pull alone, even if efficient at some depth, is apparently unable to explain the initiation of the subduction, and the mechanism perpetuating plate motions in general.

The slab detachment model is conceived as a consequence of the negative buoyancy of the slab and it has been invoked many times to explain the supposed rupture of the slab in tomographic images (e.g., Wortel and Spakman, 2000) and to fit the geochemistry of magmatism (e.g., Lustrino, 2005). However, tomographic images are based on velocity models that often overestimate the velocity of the asthenosphere where usually the detachment is modeled. Therefore the detachment disappears when using slower velocity for the asthenosphere in the reference velocity model, or generating regional tomographic images with better accuracy (e.g., Piromallo and Morelli, 2003). Recently, Rychert et al. (2005) have shown how the base of the lithosphere — top of the asthenosphere (LVZ, e.g., Panza, 1980) is characterized by unexpected, few km thick, extremely low velocities beneath northwestern North America, far from subduction zones. This implies a revision of the velocity models used for mantle tomography, particularly in areas characterized by strong lateral variations in composition of the subducting lithosphere (e.g., continental vs. oceanic) that cannot be 3D modeled with a 1D velocity model.

8.2. Mantle convection

It is obvious that convection occurs in the mantle, not only from modeling, but also from the kinematics of plate boundaries, where mantle upraises along ridges and lithosphere sinks along subduction zones. It is also evident that oceanic lithosphere circulates in the mantle much more easily than the continental lithosphere, since only relatively young (180–0 Ma) oceans cover the Earth's surface, comparing to the much older cratons (>3000 Ma), being the thick continental lithosphere buoyant over the mantle. Convection is required to cool the Earth. But convection models are necessarily oversimplified and possibly overvaluated. The mantle is considered compositionally quite homogeneous, but is not, having both vertical and lateral significant heterogeneities. The whole Earth is intensely stratified both in density and chemistry from the topmost atmosphere down to the core. The supposed convection cells should be made of an uprising warmer buoyant mantle, laterally accompanied by downwelling cooler currents. In the view of convection modelers, the surface expression of cells should be the plates. But the Atlantic, E-Africa and Indian rifts have no intervening subductions; there are also several cases of paired subduction zones without rifts in between: this shows the inapplicability of the convection cells to the simple superficial plate tectonics kinematics.

In most of the convection models, uprising and downwelling mantle currents are stationary, but we know that all plate margins rather migrate. Convection styles frequently generate polygonal shapes for cells, but plate margins can be very linear e.g., the Atlantic ridge, in contrast with the typical mushroom shape of mantle plumes.

The fastest W-ward moving plate relative to the mantle (the Pacific plate) has the lowest asthenosphere viscosity value (Pollitz et al., 1998), and it is the most decoupled plate, but mantle convection should rather predict that faster moving plates are more coupled (higher viscosity) with the mantle.

The Hawaii hotspot volcanic chain indicates that the underlying mantle is moving E–SE-ward. Beneath the East Pacific Rise, an eastward migrating mantle has been modeled by Doglioni et al. (2003) and Hammond and Toomey (2003). An eastward migrating mantle has been suggested also beneath the Nazca plate by Russo and Silver (1994) through shear wave splitting analysis. An eastward relative mantle flow beneath the South America plate is imposed by the hotspot reference frame (Van Hunen et al., 2002). A relatively moving eastward mantle flow has been proposed also beneath North America (Silver and Holt, 2002) and beneath the Caribbean plate (Negredo et al., 2004). Beneath the

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Tyrrhenian Sea a similar west to east flow of the mantle can be inferred from mantle anisotropy (Margheriti et al., 2003). A global reconstruction of the anisotropy in the asthenosphere (Debayle et al., 2005) fits quite well the sinusoidal flow of plate motions (e.g., Doglioni et al., 1999a), apart along subduction zones where the shear wave splitting anisotropy shows orthogonal trend compatible with the re-orientation of a flow encroaching an obstacle.

8.3. Earth's rotation

The lithosphere is decoupled relative to the mantle as indicated for example by the hotspots tracks. The anisotropy detected by shear wave splitting supports a shear zone active in the asthenosphere (Gung et al., 2003). Sheared asthenospheric xenoliths confirm decoupling at that depth (Kennedy et al., 2002), and the migration of plate boundaries in general (Garfunkel et al., 1986; Doglioni et al., 2003). But what is forcing the lithosphere relative to the mantle? The decoupling is polarized toward the west (Rittmann, 1942; Le Pichon, 1968; Bostrom, 1971; Wang, 1975), although along a sinusoidal flow (Doglioni et al., 1999a; Crespi et al., 2007).

Scoppola et al. (2006) recently proposed a combined model where the net westward rotation of the lithosphere relative to the underlying mantle 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 decreasing the moment of inertia and speeding up the Earth's rotation, only partly counterbalancing the tidal drag; 3) The development of thin (3–30 km) layers of very low viscosity hydrate melt rich channels in the asthenosphere. Scoppola et al. (2006) 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. This can account for the geological and geophysical asymmetry characterizing W-directed vs. E- or NE-directed subduction zones and related orogens (Marotta and Mongelli, 1998; Doglioni et al., 1999a). The fastest westerly moving plate (Pacific) is the slowest toward the east possibly due to the more effective decoupling in the asthenosphere generated by the Earth's rotation.

9. Horizontal vs. vertical movements at subduction zones

The comparison between horizontal speed of plates and their vertical rate at plate boundaries could provide some insight on plate dynamics. Slab pull is widely considered the engine of plate motions (e.g., Anderson, 2001; Conrad and Lithgow-Bertelloni, 2003; Sabadini and Vermeersen, 2004). Most numerical models for vertical motions in subduction zones use this assumption, although the excess mass of subducted slabs predicted by slab pull models is greater than that predicted by geoid models (Chase, 1979). Such models might be able to reproduce surface topography and subsidence rates of trenches (e.g., Zhong and Gurnis, 1992; Giunchi et al., 1996; Sabadini and Vermeersen, 2004) and suggest that topography could be a dynamic feature depending on the balance between tectonic and buoyancy forces (Melosh and Raefsky, 1980; Wdowinski, 1992). However, a sensitivity analysis of the effects of slab buoyancy showed that typical trench bathymetries are obtained with both positive and negative density anomalies of the slab (Hassani et al., 1997). According to this modeling slab buoyancy controls overriding plate topography, but overriding plate topography is dramatically influenced by parameters not included in calculations, such as the accretional (Karig and Sharman, 1975) or erosional (von Huene and Lallemand, 1990) nature of the subduction, the amount of shortening and the depth of the decollement (Woodward et al., 1989), the deformation partitioning between brittle and ductile levels and erosion (Willett and Brandon, 2002).

The present (Fig. 27) and past motions along subduction zones, as in any other plate boundary (Cuffaro et al., 2006), have the horizontal component in average 10 to 100 times faster (10–100 mm/yr) than the vertical component (0.01–1 mm/yr; Fig. 28).

Vertical movements along subduction zones such as uplift in the overriding plate and subsidence in the subducting plate accompany respectively the growth of an orogen and the deepening of a trench or foreland basin. Bernet et al. (2001) used apatite fission-track grain–age distributions for detrital zircons to infer a steady-state exhumation in the Alps at rates of 0.4–0.7 mm/yr since at least 15 Ma. Subsidence rates in the alpine foredeep are in the order of 0.1–0.3 mm/yr (Doglioni, 1994). Rates along the Andean subduction zone are of the order of 1–4 mm/yr for uplift and less than 0.5 mm/yr for subsidence. Fission-track analysis in the Peruvian Andes suggests 1.1 mm/yr uplift (Montario, 2001). Convergence rates along the same subduction zone are in the order of 30–100 mm/yr.

In Alaska, exhumation rates of about 3 mm/yr have been suggested (Spotila et al., 2004). Faster (5–10 mm/yr) uplift rates have been computed in Taiwan and Papua New Guinea (Liu, 1982; Dadson et al., 2003; Baldwin et al., 2004).

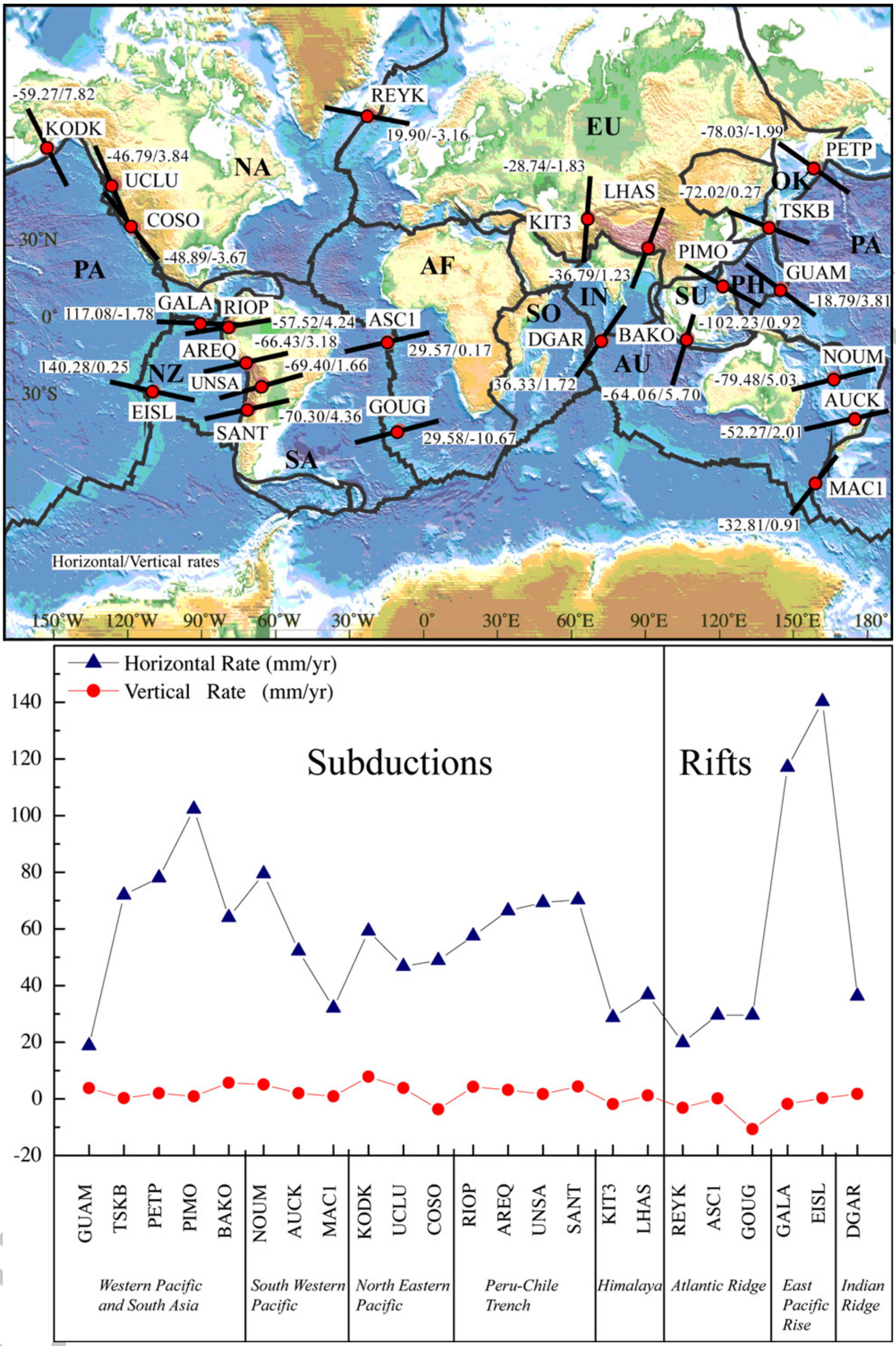

Fig. 27. Comparisons between relative horizontal and vertical motions of selected GPS stations. The oriented segments in the upper panel show relative motion directions. Units are in mm/yr, and show horizontal and vertical velocities respectively. The dot indicates the location of the vertical motion on the plate boundary. About horizontal velocities, negative signs show contraction and positive signs show extension. About vertical velocities, negative signs show subsidence and positive signs show uplift. Site localities are from Heflin et al. (2005). PA, Pacific plate; PH, Philippine plate; AU, Australia plate; IN, Indian plate; SO, Somalia plate; EU, Eurasia plate; SU, Sunda plate; OK, Okhotsk plate; NZ, Nazca plate; NA, North America plate; SA, South America plate; AF, Africa–Nubia plate (after Cuffaro et al., 2006).

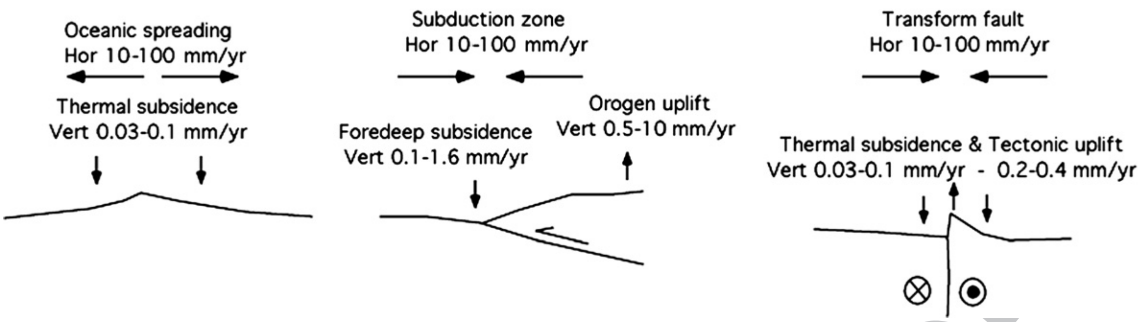

Fig. 28. Present and past horizontal relative plate motions are about 10–100 times faster with respect to vertical movements in all geodynamic settings.

Foredeeps and trenches during the last 100 Ma have worldwide subsidence rates spanning on average from 0.1 to 1.6 mm/yr (Doglioni, 1994), with the fastest rates located along the west-directed subduction zones. Along the Marianas subduction zone, where the slab pull is theoretically the highest on Earth, the Pacific plate moves WNW-ward faster than 100 mm/yr, whereas the subsidence in the trench is in the order of few mm/yr maximum.

The slow uplift of an orogen is partitioned throughout the belt, while the two convergent plates move 10–100 times faster. Plate boundaries can in fact be several hundreds km wide (e.g., Gordon, 2000). Moreover, the subduction rate can be even smaller than the convergence rate (Doglioni et al., 2006a).

If plate velocity is controlled by the plate boundary itself (i.e., slab pull or trench suction), how is it possible for the fastest down going plate to have such relatively smaller uplift in the belt and smaller subsidence rate in the trench, or to have a plate moving faster than the energetic source? This would rather support a passive role of plate boundaries with respect to far field forces determining the velocity of plates. The faster horizontal velocity of the lithosphere with respect to the upward or downward velocities at plate boundaries supports dominating tangential forces acting on plates. These forces acting on the lithosphere can be subdivided in coupled and uncoupled, as a function of the shear at the lithosphere base. The higher the asthenosphere viscosity, the more significant should be the coupled forces, i.e., the mantle drag and the trench suction. The lower the asthenosphere viscosity, the more the effects of uncoupled forces might result determinant, i.e., the ridge push, the slab pull and the tidal drag. Although a combination of all forces acting on the lithosphere is likely, the decoupling between lithosphere and mantle suggests that a torque acts on the lithosphere independently of the mantle drag. The Earth's rotation might rather have a primary role if the viscosity of the upper asthenosphere is sufficiently low.

Slab pull and ridge push are candidates for generating a torque on the lithosphere, but few counterarguments are presented. For example, no significant correlation exists between trench percentage and plate velocity (Fig. 26, A, B). Moreover, unlike these boundary forces, the advantage of the tidal drag is to be a volume force, acting simultaneously on the whole of the plate, and being the decoupling at the lithosphere base controlled by lateral variations in viscosity of the low-velocity layer.

10. Upper mantle circulation

Subduction zones represent part of the mantle convection. Hamilton (2003) proposes a restricted upper mantle circulation driven from the cooler plates. Regardless the slabs penetrate or not into the lower mantle, we observe that W-directed subductions enter in the mantle on average at least 2–3 times faster than the opposite E–NE-directed subduction zones. There are two reasons for this kinematic asymmetry: 1) most of the W-directed subductions have the subduction rate S increased by the hinge subduction retreat; 2) they are steeper and have therefore a higher $Sv = S \sin\alpha$.

Therefore W-directed subduction zones contribute more efficiently to mantle recycling; for example along the northern Tonga subduction zone the subduction rate can be as high as 240 mm/yr (Bevis et al., 1995), whereas the central Andes subduction rate can be of about 31 mm/yr (Fig. 10).

Along oceanic ridges, melting produces a residual depleted asthenosphere that is also more viscous that the undepleted one. Braun et al. (2000) have shown that water extraction during melting leads to higher viscosity in the residual mantle up to 2 orders of magnitude. The mantle, once depleted along the transit beneath the ridge should be cooler, less dense and more viscous. The most

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shallow, depleted asthenosphere will eventually become lithospheric mantle with the progressive cooling, shifting away from the ridge. In this view, moving horizontally within the asthenosphere, the viscosity distribution could be due more to lateral compositional variations rather than temperature gradients.

Since the Pacific plate has the lowest viscosity of the underlying asthenosphere and it is the fastest plate, there seems to be a positive correlation between asthenospheric viscosity, lithospheric decoupling and plate velocities.

Plate variable velocities seem controlled by the lithosphere–asthenospheric decoupling, which is a function of the asthenospheric viscosity. Subduction zones and oceanic ridges contribute to upper mantle convection. The W-directed subductions clearly reach the lower boundary of the upper mantle. The E- or NE-directed subduction zones are rather shallower and less inclined. Most of their seismicity terminates into the asthenosphere, apart fewer deep earthquakes at the upper–lower mantle transition. These earthquakes are considered either related to phase transition, or to remnants of cold detached slabs. An alternative interpretation would be that they represent the faster shear and related strain rate between upper and lower mantle, triggered by the Bernoulli principle, for the narrower upper mantle section beneath the E–NE-directed subduction zones.

Most of the Earth's subduction-related magmatism is sourced from a top slab depth range of 65–130 km (England et al., 2004). As shown in the previous chapters, the Andean subduction rate should be slower than the convergence rate. Moreover, the slab along some segments of the Andes becomes almost flat, indicating the insertion of the slab into the asthenosphere. At that depth, the slab shares the pressure and temperature that allow partial melting of the asthenosphere. Therefore the oceanic lithosphere might re-enter the source mantle, becoming ductile, and gradually melting, refertilizing the mantle that generated it (Fig. 29). Note that in the mantle reference frame, assuming shallow hotspots and the faster W-ward drift of the lithosphere, the Nazca plate is coming out of the mantle, but subduction occurs because the South America plate moves westward faster than Nazca (Fig. 21).

The W-directed subduction zones are penetrating and geochemically modifying the lower upper mantle (Fig. 29), whereas the E- or NE-directed subduction zones are more suitable for regenerating the shallow upper mantle that is the most favorable candidate for sourcing MORB along mid-oceanic ridges (e.g., Bonatti et al., 2003).

For the aforementioned kinematic constraints, the rate of subduction is expected to be generally 2–3 times higher along W-directed subduction zones than along the opposite settings (Fig. 30). The different kinematics also generate different geometries and depth of the 1300 °C isotherm assumed to mark the lithosphere–asthenosphere boundary. Therefore, the asthenosphere is much shallower along W-directed subduction zones than along the opposite slabs (Fig. 31). Moreover, in the hangingwall of the slab, the asthenospheric section cross cut by the fluids released by the slab at about 100–130 km (Syracuse and Abers, 2006) to generate the magmatic arc, is thicker along W-directed subduction zones than again along the opposite subduction zones (AW>AE, Fig. 30), possibly providing geochemical differences among the two subduction end members.

11. Dynamic speculations

As discussed earlier, orogens and related features show marked asymmetries (Fig. 32). The topography and the foreland monocline are lower and steeper for the

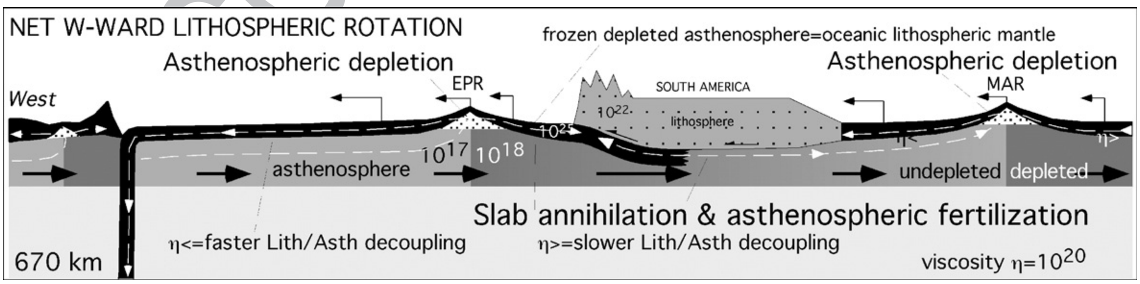

Fig. 29. Model for the upper mantle cycle. Lower the asthenospheric viscosity, faster the W-ward displacement of the overlying plate. The asthenospheric depletion at oceanic ridges makes the layer more viscous and decreases the lithosphere/asthenospheric decoupling, and the plate to the east is then slower. The oceanic lithosphere subducting E-ward enters the asthenosphere where it is molten again to refertilize the asthenosphere. W-directed subductions provide deeper circulation. Note that the E-directed subduction (the Andes) tends to escape out of the mantle, but it is overridden by the upper plate (South America, after Doglioni et al., 2006a).

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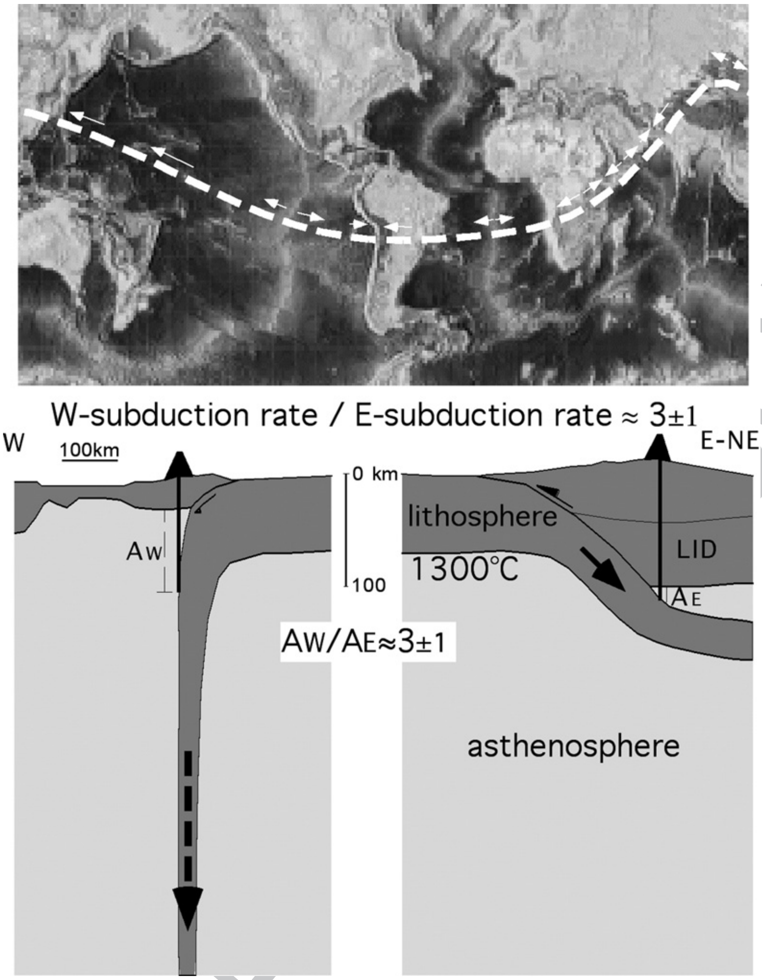

Fig. 30. The main differences between orogens are a function of the subduction polarity along the tectonic mainstream (Doglioni, 1993; Crespi et al., 2007). The volumes recycled along W-directed subduction zones is about 2–3 times higher than along the opposite settings due to the aforementioned kinematic constraints. Moreover, the asthenospheric wedge above slabs is thicker along W-directed subduction zones (AW) with respect to the E–NE-directed subductions (AE).

W-directed subduction zones (Fig. 33) and the area above sea level is remarkably higher for the opposite E- to NNE-directed subduction zones (Fig. 34). The aforementioned kinematic and geometric observations allow us to make a few dynamic considerations. All types of tectonic-geodynamic settings at plate boundaries show 10–100 times faster horizontal velocity with respect to the vertical motion (Cuffaro et al., 2006). Is this a trivial observation, or is it rather telling us something fundamental on the dynamics of plate tectonics? Does slower vertical motion imply strain partitioning and passive role of plate boundaries, as suggested, for example, by the gradual decrease in shortening from the subduction hinge to the fixed upper plate (Fig. 10)?

The mechanisms driving plate motion, e.g., plates driven by 'the boundary forces', slab pull and ridge push (Forsyth and Uyeda, 1975), vs. plates actively dragged by the asthenosphere flow (e.g., Bokelmann, 2002) seems not relevant to the preceding discussion of horizontal vs. vertical motion rates, because the rates themselves do not provide evidence for or against any particular mechanism. Both 'active plates and passive asthenosphere' and 'an active asthenospheric flow dragging passive plates' may be consistent with faster horizontal motions. The inertia of plates is negligible, and each plate must be in dynamic equilibrium, so the sum of the torques acting on a plate must be zero (Forsyth and Uyeda, 1975).

The main forces acting on the lithosphere can be subdivided into coupled and uncoupled forces (Fig. 35). Mantle drag and trench suction (e.g., Bercovici, 1998; Conrad and Lithgow-Bertelloni, 2003) need high coupling (higher viscosity) between the lithosphere

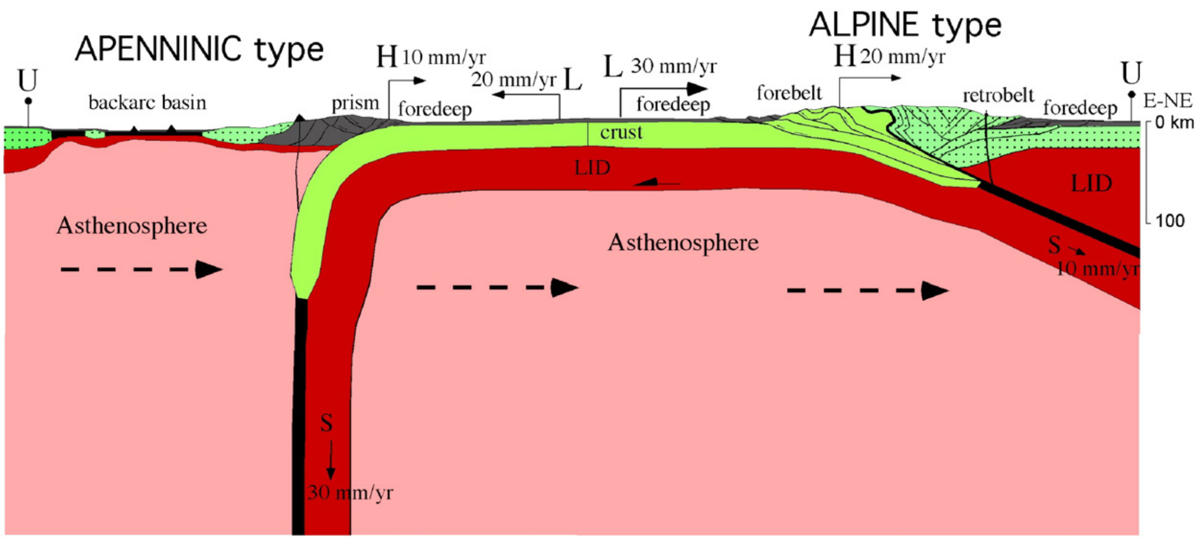

Fig. 31. Schematic sections showing how in an Alpine setting, the subduction rate is decreased by the migration of the hinge *H* toward the upper plate U, and the orogen in the final collisional stage is composed both by the upper and lower plate *L* rocks. In the opposed Apenninic setting, the subduction rate is rather increased by the migration of *H* away from U, and the accretionary prism is made of shallow rocks of the lower plate. Note also the shallower asthenosphere in the hangingwall, which is typical of W-directed subduction zones.

and the asthenosphere to be more effective. The ridge push, the slab pull and the tidal drag should rather need low coupling (lower viscosity) to be efficient (Fig. 35).

The down-dip extension along E–NE-directed slab can be generated either by the slab pull from below, or by the tidal drag acting on the surface plate.

Among the uncoupled forces, the ridge push is at least one order of magnitude lower than the slab pull (e.g., Ranalli, 1995). The dissipation of energy by tidal friction is even larger (1.6×10^{19} J/yr) than the energy released by tectonic activity (1.3×10^{19} J/yr, Denis et al., 2002). The tidal drag can effectively move plates only if very low viscosity intra-astenospheric layers occur (Scoppola et al., 2006). In this case, tidal forces, combined with mantle convection, could trigger plate tectonics. A very low velocity layer at the very top of the asthenosphere (100–150 km) has been recently demonstrated (Panza et al., 2007).

Therefore the viscosity of the upper layers of the asthenosphere plays a crucial role in controlling plate

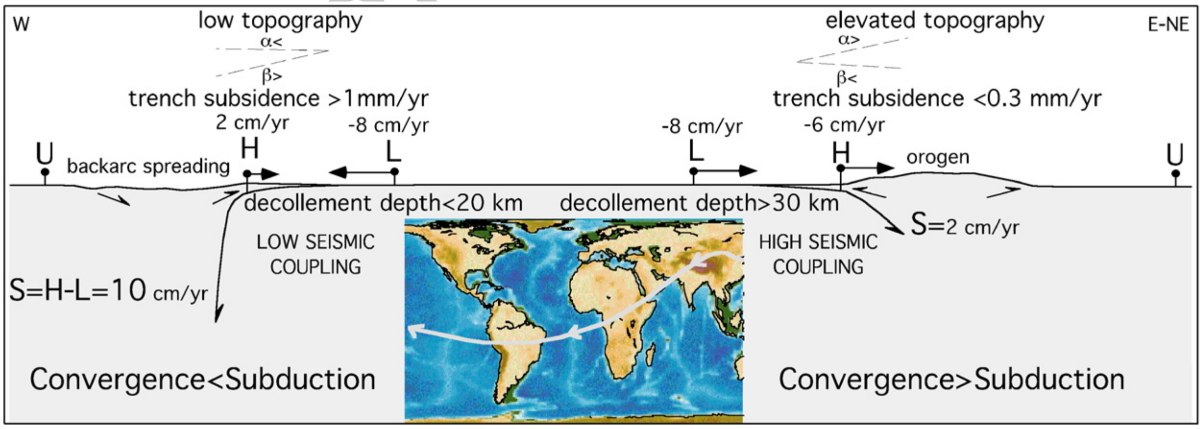

Fig. 32. Assuming fixed the upper plate U, along west-directed subduction zones the subduction hinge H frequently diverges relative to U, whereas it converges along the opposite subduction zones. L, lower plate. Note that the subduction S is larger than the convergence along W-directed slabs, whereas S is smaller in the opposite case. The two end-members of hinge behavior are respectively accompanied in average by low and high topography, steep and shallow foreland monocline, fast and slower subsidence rates in the trench or foreland basin, single vs. double verging orogens, etc., highlighting a worldwide subduction asymmetry along the flow lines of plate motions indicated in the insert (after Lenci and Doglioni, in press).

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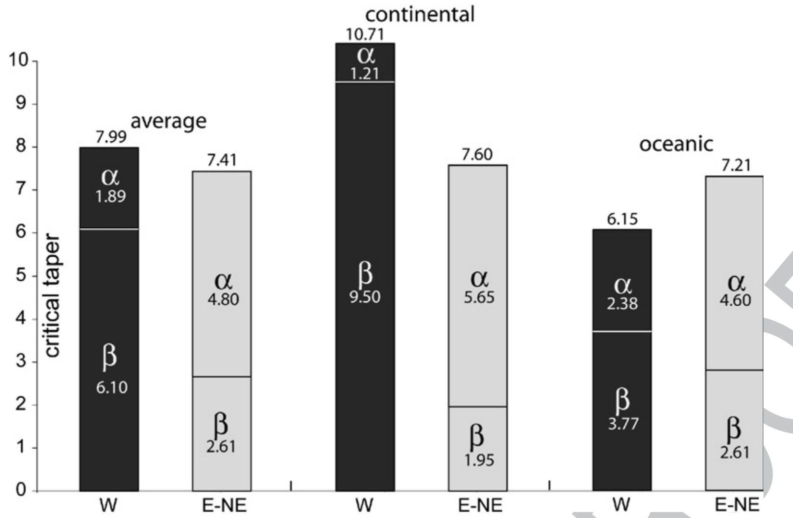

Fig. 33. Average values of the topographic envelope (α), dip of the foreland monocline (β), and critical taper ($=\alpha+\beta$) for the two classes of subduction zones, i.e., W-directed and E- or NE-directed. Note that the "western" classes show lower values α and steeper values of β.

tectonics. Moving from the highest viscosity (10^{19-20} Pa s) to the lowest (10^{12-14} Pa s), the most likely mechanisms able to move plates are in order: the mantle drag, the trench suction, the slab pull, the ridge push and the tidal drag.

Relatively small forces can move a floating plate fast horizontally, because no work has to be done against gravity, whereas non-isostatic vertical motions require work to be done against gravity. However this can be true when at the base of the lithosphere there is a very low viscosity in the decoupling layer, i.e., the weaker low velocity zone in the upper asthenosphere. Increasing the asthenosphere viscosity, larger forces are required to decouple the lithosphere. On the other hand, if the lithosphere is not moved by lateral forces such as the slab pull, but rather passively dragged by the mantle, the higher viscosity will enable a better coupling. Then, what is generating the decoupling of the lithosphere (Fig. 36)? Are there external tangential forces acting on the lithosphere?

There are lines of evidence that the lithosphere is partly decoupled from the mantle as suggested for example by

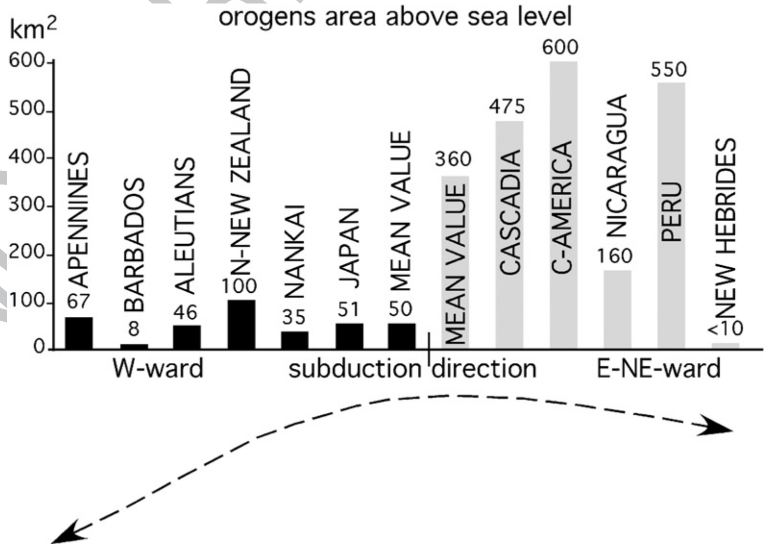

Fig. 34. Average values of the area above sea-level of the main subduction zones, showing how orogens above E- or NE-directed subduction zones are about 6–8 times larger than the W-directed subduction zones-related accretionary prisms. After Lenci and Doglioni (in press).

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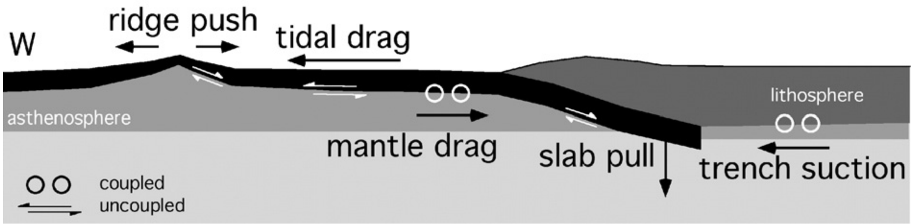

Fig. 35. Main forces acting on the lithosphere. Mantle drag and trench suction need high coupling (higher viscosity) between lithosphere and asthenosphere to be more effective. Ridge push, slab pull and tidal drag should rather need low coupling (lower viscosity) to be efficient. Since the lithosphere is decoupled with respect to the asthenosphere, possibly more than one force is actively forcing plate motions. Circles indicate coupled forces, white half arrows show the uncoupled forces. See text.

the hotspot tracks and by the asthenosphere anisotropy (e.g., Silver and Holt, 2002). A super fast net rotation of the lithosphere relative to the mantle has been proposed by Crespi et al. (2007), assuming a shallow origin of the Pacific plumes used as reference frame (Fig. 36). If so, where does the energy providing this torque come from? What is moving plates relative to the mantle? The net westerly directed rotation of the lithosphere has been attributed either to lateral variations in asthenosphere viscosity (Ricard et al., 1991), or to the Earth's rotation (Scoppola et al., 2006). The westward drift (Le Pichon, 1968) is consistent with the asymmetry of subduction and rift zones worldwide along an undulated plate motions flow (Doglioni et al., 2006a). A number of authors (e.g., Dickinson, 1978; Uyeda and Kanamori, 1979; Doglioni, 1990) proposed a shear at the lithosphere base driven by mantle drag or relative mantle flow.

Plate motions are driven either by coupled or uncoupled forces. A comparison between horizontal and vertical motions does not allow to state which plate tectonics driving mechanism prevails. However, the steady 1 or 2 order of magnitude faster horizontal over vertical motion at plate boundaries points to a stronger tangential component in plate tectonics.

However, if ridges and subduction zones trigger convection, but are nevertheless still passive features, what does move the plates? Whatever the mantle convection works, it cannot explain the lithosphere decoupling alone. Therefore the uncoupled forces appear to dominate, but we cannot exclude that possibly more than one force, both

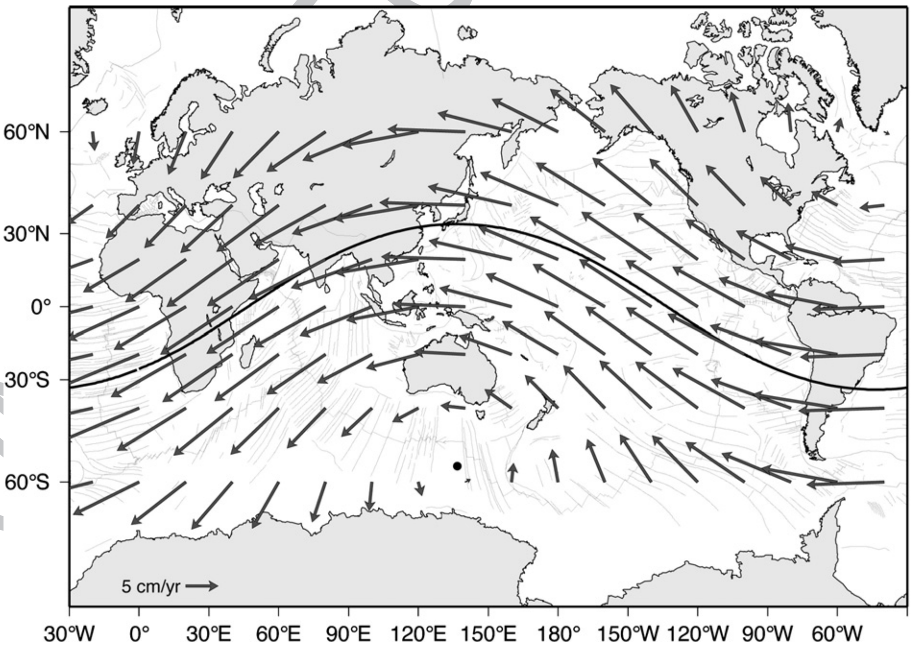

Fig. 36. Global lithospheric net rotation relative to the mantle assuming a mid-asthenospheric source of the Pacific plumes, after Crespi et al., 2007. Subduction asymmetries should not be considered as E–W related, but following or opposing the absolute tectonic mainstream.

coupled and uncoupled, energize plate motions. Earth's rotation cannot work alone because mantle convection is required to maintain the mantle fertile and the low viscosity in the asthenosphere. Moreover density gradients (e.g., slab pull) allow differential sinking of plates at convergent margins.

Are plates dragged horizontally by mantle convection (e.g., Bercovici, 1998)? Are they dragged and sheared at the base by a faster moving mantle (Bokelmann, 2002)? Are they rather pulled by slab pull forces (Forsyth and Uyeda, 1975; Anderson, 2001)? Could they be driven by Earth's rotation and tidal drag (Scoppola et al., 2006)? We do not have a final word, but further studies on the composition, water content and viscosity of the asthenosphere might significantly contribute to answer these basic questions.

12. Discussion and conclusions

A subduction zone should be analyzed considering at least three points, i.e., two located in stable areas of both the upper and the lower plate, and one located at the plate boundary, along the subduction hinge. Two main types of subduction zones can be distinguished, i.e., 1) those where the subduction hinge migrates away from the upper plate, and 2) those in which the subduction hinge migrates toward the upper plate (Fig. 31). This distinction recalls what Laubscher (1988) defined as pull arc and push arc respectively. Apart few exceptions, this distinction seems to apply particularly for W-directed and E- or NE-directed subduction zones respectively (e.g., Apennines, Marianas, Tonga and Carpathians for the W-directed, Andes, Alps, Dinarides and Hellenides for the opposite case). In either W- and E–NE-directed subduction zones, the hinge migrates eastward or northeastward relative to the upper plate (Table 2).

In the literature it is often referred to retreating or advancing slab. However this terminology might generate confusion because a retreating hinge or slab retreat in the upper plate reference frame might become a fixed hinge in the mantle reference frame (e.g., Barbados). On the other hand, an advancing hinge relative to the upper plate is a retreating hinge relative to the mantle (e.g., Andes).

The rate of subduction is generally larger than the convergence rate along W-directed subduction zones, whereas it is smaller along E- or NE-directed subduction zones. Therefore the subduction rate is decreased or increased as a function of whether the subduction hinge converges or diverges relative to the upper plate. Along W-directed slabs, the subduction rate is the convergence rate plus the slab retreat rate, being the latter close to the backarc extension rate, if no accretion occurs in the upper plate, which is a rare case.

As a result, in eastern Pacific subduction zones, and in the E- or NE-directed subduction zones in general such as the Alps or Himalayas, the subduction rate should be lower than the convergence rate. On the other hand, along the western Pacific subduction zones, and the W-directed subduction zones in general such as the Apennines, the subduction rate has to be faster than the convergence rate since it is incremented by the hinge retreat.

In this interpretation, the far field velocities of the upper and lower plates control the subduction rate, and the subduction is a passive process. In fact, the rates of subduction do not determine plate velocities, but are rather a consequence of them.

The convergence/shortening ratio along W-directed subduction zones is instead generally lower than 1. Along E- or NE-directed slabs, the shortening in the upper plate decreases the subduction rate, and typically no backarc basin forms. The convergence/shortening ratio in this type of orogens is higher than 1 and is inversely proportional to the strength of the upper plate and is directly proportional to the coupling between upper and lower plates. Higher the strength and lower the coupling, smaller the shortening, and faster is the subduction rate.

Along both W- and E- NE-directed subduction zones, the hinge migrates eastward relative to the upper plate, apart few exceptions like Japan. Therefore, most frequently along the W-directed subduction zones the hinge migrates away with respect to the upper plate, whereas the hinge migrates toward the upper plate along E- NE-directed subduction zones. In this interpretation, the far field velocities of the upper and lower plates control the subduction rate, and the subduction is a passive process.

Along E- or NE-directed subduction zones, the convergence rate is partitioned between upper plate shortening and subduction. The shortening is mainly concentrated in the upper plate until it is continental and less viscous than the lower oceanic plate. At the collisional stage even the lower plate is extensively shortened. However, the observation that, say 80 mm/yr convergence are transferred to 60 mm/yr shortening in the upper plate and 20 mm/yr only are reserved to subduction, point out a fundamental result, i.e., the lower values of shortening and subduction rates with respect to the convergence rate are hierarchically a consequence of the far field plate motion. This means that the driving primary source of energy for determining the convergence is neither with the slab, nor in the related

t2.1 Table 2
Few main geometric, kinematic and dynamic differences between orogens and subduction zones following or opposing the tectonic mainstream.
t2.2 Subduction zones parallel to the mainstream (e.g., Pyrenees) have similar characters as the subduction zones following it

		Subductions opposing the tectonic mainstream (W-directed)	Subductions following the tectonic mainstream (E–NE–NNE-directed)
t2.3			
t2.4	Elevation average	−1250 m	+1200 m
t2.5	Foreland monocline average dip	6.1°	2.6°
t2.6	Trench or foredeep subsidence rate	>1 mm/yr	<0.3 mm/yr
t2.7	Prism envelope average dip	1.9°	4.8°
t2.8	Orogen-prism vergence	Single verging	Mostly Double verging
t2.9	Type of prism rocks	Mostly sedimentary cover & volcanics	Largely basement, sedimentary cover & volcanics
t2.10	Prism decollement depth	0–10 km; (rarely up to 20 km) offscraping the top of the lower plate	>30 km; Oceanic subduction, affecting mostly the whole section of the upper plate; continental subduction affecting also the lower plate
t2.11	Seismic coupling	Mainly low	Mainly high
t2.12	Moho	Shallow (<30 km) new upper plate Moho	Deep (>40 km) doubled old Mohos
t2.13	Asthenosphere depth	Shallow (<20–50 km) beneath the arc	Deep (>70–100 km) beneath the arc
t2.14	Seismicity	0–670 km; intra-slab mostly down-dip compression and horizontal shear	0–250 km and scattered 630–670 km; intra-slab mostly down-dip extension
t2.15	Slab dip	25°–90°	15°–50°, steeper up to 70° along oblique subductions and thicker upper plate
t2.16	Subduction hinge motion relative to upper plate	Mainly diverging Eastward (except Japan where subduction started to flip)	Mainly converging Eastward or northeastward
t2.17	Subduction hinge motion relative to mantle	Fixed	Westward or southwestward
t2.18	Subduction rate	S=H−L, faster than convergence rate; mainly slab retreating and entering the mantle	S=H−L, slower than convergence rate; slab “escaping” relative to the mantle, overridden by the upper plate
t2.19	Backarc spreading rate	H(>0), — prism accretion, — hinge asthenospheric intrusion	Differential velocity between two hangingwall plates
t2.20	Slab/mantle recycling	About 3 times higher than opposite	About 3 times lower than opposite
t2.21	Subduction mechanism	Slab-mantle wind interaction+far field plate velocities+slab density gradient relative to the country mantle	Far field plate velocities+slab density gradient relative to the country mantle

orogen. A paradigm of plate tectonics is that the negative buoyancy of slabs drives plate motions (e.g., Conrad and Lithgow-Bertelloni, 2003), as suggested by the steeper dip of the slab bearing old oceanic crust (Forsyth and Uyeda, 1975), and the convergence rate at subduction zones related to the age of the oceanic crust at the trench (e.g., Carlson et al., 1983). However a number of aforementioned counterarguments make the slab pull weaker than so far accepted in the literature. For example the energy required to pull the plates is far higher than the strength that plates can afford under extension. Moreover the asymmetry which is evident comparing the western and the eastern Pacific subduction zones occurs also in the Mediterranean subductions, regardless the age and composition of the downgoing lithosphere (Doglioni et al., 1999a). Moreover, Lallemand et al. (2005) and Cruciani et al. (2005) have demonstrated that there is no correlation between the slab dip and the age of the subducting lithosphere. To this discussion we add here the observation that there is no correlation between convergence rate and age of the oceanic lithosphere at the trench (Fig. 25), suggesting that the negative buoyancy cannot be the primary driving force of plate tectonics. In the hotspot reference frame, the Africa plate moves westward (Gripp and Gordon, 2002) without any slab in its western side. Moreover, it moves opposite to a hypothetic Atlantic ridge push. The only slabs attached to Africa are in its northern margin, i.e., the Hellenic–Cyprus and Apennines subduction zones. Although problematic, another small, finger-like, east-dipping slab has been supposed beneath the Gibraltar arc (Gutscher et al., 2002). The Hellenic–Cyprus slab is also dipping/directed NE-ward, opposite to the direction of motion of the Africa plate providing a kinematic evidence of no dynamic relationship. The Apennines slab is retreating and westward directed. These northern Africa subduction zones are a small percentage of the plate boundaries surrounding Africa, and they dip in opposite directions with respect to the absolute motion of the plate. Therefore they cannot be the cause of its motion.

Please cite this article as: Doglioni, C. et al. Subduction kinematics and dynamic constraints. Earth-sci. Rev. (2007), doi:10.1016/j.earscirev.2007.04.001

The Apennines–Marianas and the Alps–Andes (continental and oceanic subduction zones) are representative of the two major classes of orogens where the subduction hinge migrates away from, and toward the upper plate respectively (Fig. 32). However, it has been shown that in the Apennines, a number of sub-settings can be described, and in the southern side of the arc the hinge migrates toward the upper plate, while in the rest of the arc the hinge migrates away (Fig. 18).

The orogens of Alpine–Andean type have therefore slower subduction rates than the Apenninic–Marianas type when convergence is constant. The double verging Alpine–Andean type belt is composed mostly by upper plate rocks during the oceanic subduction, being the lower plate more extensively involved during the later collisional stage. The single verging Apennines–Marianas type belt is rather composed primarily by shallow rocks of the lower plate (Fig. 31). A wide variety of different geophysical, geological and volcanological signatures mark the two end members of orogens and subduction zones (Doglioni et al., 1999a). The two end-members where the subduction hinge migrates away or toward the upper plate largely match the two opposite cases of seismic decoupling or seismic coupling (e.g. Scholz and Campos, 1995). However, along the Apennines subduction zone at least five different kinematic settings coexist, showing how a single subduction can have internal deviations from the standard model and variable velocities as a function of the combination V_H and V_L.

Relative to the mantle, the W-directed slab hinges are almost fixed, whereas they move west or southwest along E- or NE-directed subduction zones. When describing the plate motions relative to the mantle, e.g. in the hotspot reference frame, both Africa and Greece move southwestward with respect to the mantle (Greece faster). This implies that the slab is moving in the opposite direction of the subduction when studied relative to the mantle. The slab sinks because it is overridden by the faster upper plate (Fig. 23). This observation indicates that the slab pull cannot be the only driving force of neither the Hellenic subduction, nor of E- or NE-directed subduction zones in general, because it is moving SW-ward or W-ward relative to the mantle. Along this subduction zone, again, plate motions are not controlled by subduction rates, but vice-versa.

Unlike E–NE-directed subduction zones, along W-directed subduction zones, the slab generally sinks faster than the convergence rate. However, the lower plate can converge or diverge from the upper plate. With respect to the mantle, W-directed slab should in general have a fixed, or anchored, hinge, whereas the hinge migrates eastward along the opposite subduction zones. In case the lower plate diverges faster than the hinge, the subduction rate is negative, i.e., the lower plate is escaping from the subduction (Fig. 14, 5W and 6W). This setting might represent a final evolution of the subduction zone and could be an alternative cause for the detachment of slab that is generally inferred as related to the sinking of the supposed denser lower part (e.g., Wortel and Spakman, 2000). Alternatively, since once penetrated into the mantle, the slab cannot be re-exhumed, the detachment of the slab could rather occur because the top part of the lower plate L is moving away from the deeper (>100–150 km?) segment (Fig. 14, 5W and 6W).

W-directed subduction zones bring larger volumes of lithosphere back into the mantle than the opposite subduction zones.

W-directed subduction zones have the rate of sinking controlled by the slab-mantle horizontal "easterly"-directed wind interaction, which mostly determines the retreat of the subduction hinge, plus the far field velocities of the plates, and the value of the negative slab buoyancy.

Alternatively, E–NE–NNE-directed subduction zones have rather rates chiefly determined by the far field velocity of plates, being the subduction hinge generally advancing toward the upper plate and decreasing the subduction rate (Table 2).

The analyzed kinematics frames suggest that subduction zones have rates of sinking controlled by far field plate velocities, hinge migration direction, and subduction polarity, claiming for a passive behavior of the slabs. This is more reasonable if the net "westward" rotation of the lithosphere is a global phenomenon rather than the simple average of plate motions (Scoppola et al., 2006). Tidal drag (Bostrom, 1971; Moore, 1973) combined with Earth's rotation, mantle convection, and an ultra-low viscosity layer in the asthenosphere could trigger plate motions (Scoppola et al., 2006).

The ridge push, related to the topographic excess, should be higher along elevated orogens, where on the contrary, plates converge rather than diverge. Boundary forces such as slab pull and ridge push could in principle generate a deceleration moving away from the energy source, but plates rather show internal homogeneous velocity (Fig. 34). Mantle convection could satisfy a steady state speed of the overlying lithosphere, assuming low or no decoupling at the asthenosphere interface. However, mantle convection is kinematically problematic in explaining the migration of plate boundaries and the occurrence of a decoupling surface at the lithosphere base. Although a combination of all forces acting on the lithosphere is likely, the decoupling between lithosphere and mantle suggests that a torque acts on the lithosphere independently of the mantle drag. Slab pull and ridge push

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are candidates for generating this torque, but, unlike these boundary forces, the advantage of the Earth's rotation and related tidal drag is to be a volume force, acting simultaneously and tangentially on the whole plates. Tidal drag maintains the lithosphere under a permanent high frequency vibration, polarized and sheared toward the "west" (Fig. 37). Earth's rotation and the break exerted by the lag of the tidal bulge (Bostrom, 1971) can be efficient only if very low viscosity occur at the lithosphere–asthenosphere transition (Jordan, 1974), and growing evidences are emerging on the presence of an ultra-low viscosity layer at the very top of the asthenosphere (e.g.,

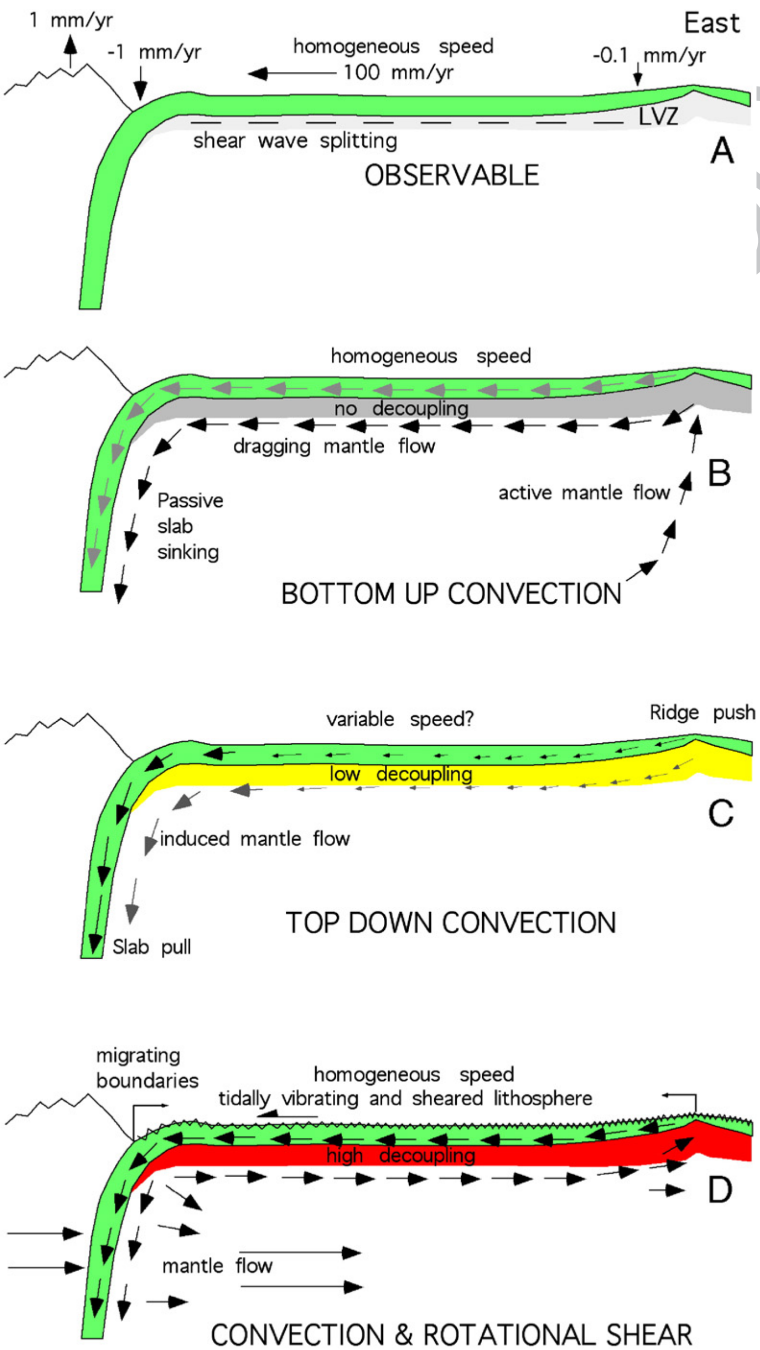

Fig. 37. The surface observables (A) are compared with three models of plate dynamics, where plate motion is induced by classic mantle convection (B), boundary forces (C), or the combination of the aforementioned mechanism plus the rotational drag (D).

Rychert et al., 2005), possibly related also to higher fluids concentration in the mantle. Lateral variations in the low-velocity layer viscosity could control the different velocity of plates. The Earth's rotation contribution to move the lithosphere could account for i) the homogeneous internal velocity of each plate, ii) the decoupling occurring at the lithosphere base, and iii) the westerly polarized migration of the lithosphere and the plate boundaries, consistent with the geological asymmetries of subduction and rift zones as a function of the geographic polarity. In this view, plate dynamics could be a combination of mantle convection and the shear induced by the tidal drag (Fig. 37).

13. Uncited references

Oliver, 1972

Shemenda, 1993

Acknowledgments

The article benefited of the critical reading by D. Frizon de Lamotte and N. Kukowski. Discussions with M. Battaglia, C. Chiarabba, M. Crespi, G. Dal Piaz, C. Faccenna, E. Garzanti, F. Innocenti, S. Lallemand, G. Panza, and F. Riguzzi were very stimulating. Research was supported by the University La Sapienza, Roma.

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Please cite this article as: Doglioni, C. et al. Subduction kinematics and dynamic constraints. Earth-sci. Rev. (2007), doi:10.1016/j.earscirev.2007.04.001