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Geological evidence for a global tectonic polarity

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Abstract: Inferences about plate motions over the past 40 Ma, based on structural and geodetic data, suggest that there is a coherent movement direction on a global scale. Moreover a 'westward' drift of the lithosphere relative to the asthenosphere is indicated by plate motions in the hot-spot reference frame. Independent geological observations such as the different styles of deformation in thrust belts associated with subduction following or opposing the relative 'eastward or northeastward' mantle flow also support this global tectonic polarity. Unfortunately most hot-spot reference frames are based on assumptions which strongly limit their validity. For example, many combine magmatic sources originating from different depths in the mantle, including hot-spots located along oceanic ridges which are not fixed relative to one another. Using a hot-spot reference frame filtered to exclude shallow hot-spots and anomalous 'wet-spots', we would predict a much higher relative 'westward' drift of the lithosphere. Plate motions are not random, rather that they are controlled by the relative motion between lithosphere and asthenosphere, and by gradients in lithospheric thickness and composition.Triangular shaped plates may experience rotations resulting from differential drag at the base of the lithosphere with respect to the asthenosphere (e.g. the clockwise rotation of South America). Plates appear to follow an undulatory pattern of motion on a global scale and their poles of rotation fall into a main cone located near the Earth's rotation poles. Plate rotations induced by triangular shape, differential basal drag or lithospheric anisotropies along subduction zones may generate the scattered distribution of Euler poles. This might have inhibited the recognition of a coherent global tectonic pattern.

The main aim of this paper is to discuss the evidence for the global tectonic polarity (Doglioni 1990, 1991*a*), and to discuss the geodynamic implications of such a model. The basic assumption is that the lithosphere is detached with respect to the mantle, as indicated, for example, by the age progression of volcanism along the Hawaiian chain, which demonstrates that the lithosphere is moving relative to the mantle (Wilson 1963). This decollement may be distributed in the asthenosphere where a strong viscosity contrast exists relative to the lithosphere (Sabadini & Yuen 1990) and it may be variable from point to point. Based on this simple but often forgotten statement, the influence of lateral lithospheric heterogeneities is discussed. Excellent reviews of the driving mechanisms of plate tectonics are presented in Ziegler (1992, this volume), and these will not be reiterated here.

The flow lines of plate motion

Linking vectors of plate motion, which can be inferred from structural data on rift zones, transform zones and orogens of the last 40 Ma, allows us to plot a series of flow lines depicting the direction of plate motion (Fig. 1). These lines represent the average direction of motion of plates in geographic coordinates (Doglioni 1990, 1991*a*), independent from their relative motion. The data have been filtered to take into account oblique motions (transtensive or transpressive zones) or other factors deviating the main stress with respect to the plate motion vectors. It has been demonstrated that the stress field is not necessarily a direct indication of plate motion. For example transform faults with transtensive or transpressive components are not parallel to the plate motion vectors and they should be used carefully. The San Andreas fault is a good example of an 85° of deviation of the main stress with respect to the plate motion vector (Argus & Gordon 1990; Zoback 1991; Zoback & Beroza 1993). As an example of the estimation of plate movement directions, the motion of Arabia may be inferred by considering the Red Sea as a left lateral transtensive rift and the Gulf of Aden as right lateral transtensive rift. Clearly only one line can represent the average main direction of motion of Arabia in present-day geographic coordinates (Fig. 1). Another example is the undulation of the southern cordillera in Peru where left lateral transpression occurs and the local σ_1 deviates from the direction of westward motion of South America, which is very well constrained by the E–W-trending Atlantic transform faults.

The flow lines (Fig.1) generate ellipses and not circles and thus a unique pole of rotation is insufficient to describe plate motions; instead we require a cone located near the geographic pole. A few deviations from the main trends occur. The Cocos plate, for example, along the Galapagos ridge, is opening at about 90° with respect to the global flow. However the length of this feature is about one or two orders of magnitude less important with respect to the general trends over the Earth's surface. An attempt to explain this deviation will be proposed later. The flow pattern becomes less clear in the polar regions where the plates are moving more slowly.

The main flow lines shown in Fig. 1 are valid for at least the last 40 Ma, from the beginning of the linear hot spot track of the Hawaiian chain. However, for South America, the Atlantic Ocean, Africa, India and Eurasia, the mainstream flow direction may have been stable for the past 250 Ma, based on the orientation of rifts, thrust belts and transform fault trends. The flow pattern describes a gradual undulation from eastern Africa to Asia, where it

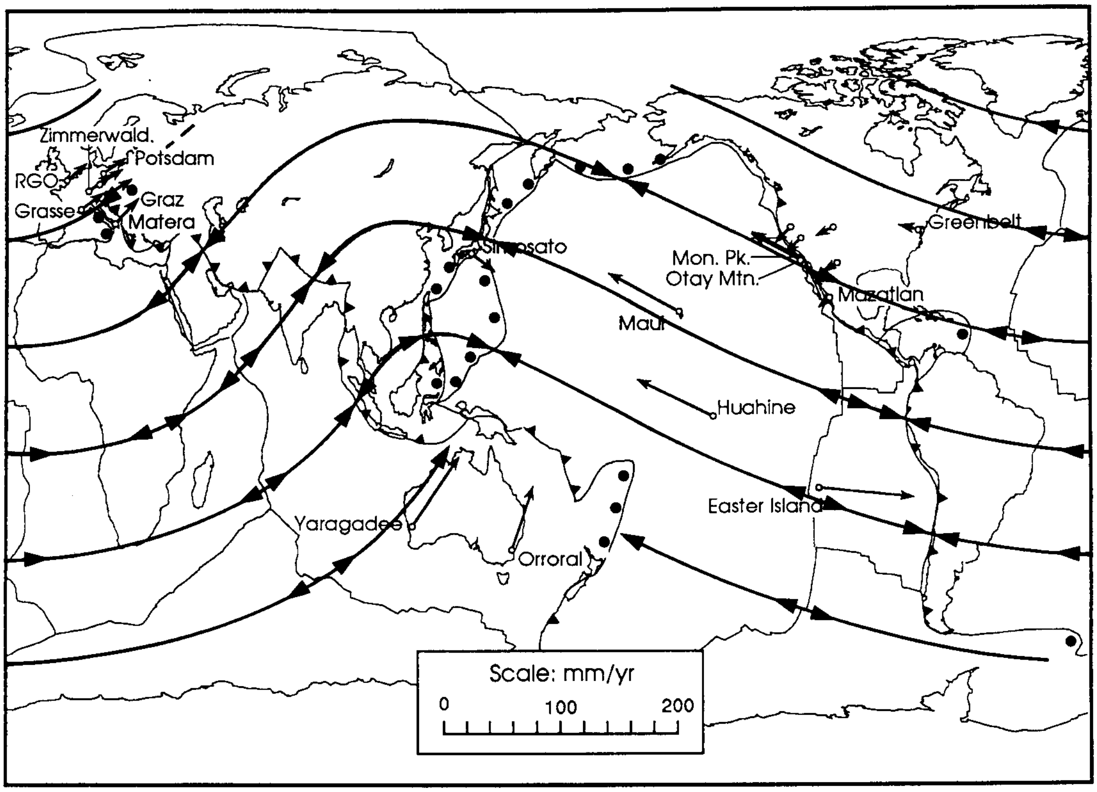

Fig. 1. Connecting the directions of plate motions that we can infer from large scale rift zones or convergent belts of the past 40 Ma, we may observe a coherent global flow pattern along which plates appear to move at different relative velocities in geographic coordinates (after Doglioni 1990). Note the coincidence of the flow with the isolated arrows indicating the directions of plate motions according to the NUVEL 1 NNR model (DeMets *et al.* 1990) which are very similar to the Lageos satellite data (after Smith *et al.* 1990). Misfits may occur along transpressive plate margins (e.g. west North America). Note that the main flow of plate motions does not directly imply the 'westward' drift of the lithosphere. 'West'-dipping subduction zones are marked by black dots and 'east or NE'-dipping subduction zones by small triangles.

rapidly turns through 90° toward the Pacific. The area of this bending runs from the Barents Sea across Russia, China, Indonesia and north of Australia (Fig. 1). It is interesting to note this occurs in a region dominated by continental lithosphere, which may be significant. This change in direction of plate motions and the nature of motions in the polar regions, are clearly areas which require more detailed study in any future analysis of global flow patterns.

In Fig. 1 the directions of plate motion based on the model NUVEL 1 (DeMets *et al.* 1990) are also plotted, which are very similar to the geodetic data obtained from laser satellite ranging between 1980 and 1990 by Smith *et al.* (1990, figs 15 & 19). These data are consistent with the proposed flow lines based on independent tectonic data. Gordon & Stein (1992) have shown that recent space geodetic data are consistent with the main directions which have been proposed on the basis of other plate motion indicators (e.g. magnetic anomalies, focal mechanisms; Minster & Jordan 1978; DeMets *et al.* 1990).

Figure 1 may be useful in studies of the Mediterranean region, a controversial area for which many authors have proposed different geodynamic interpretations mainly based on the relative motion between Africa and Europe. One classic problem in Mediterranean geology and geophysics is that Africa appears to be moving northward or northwestward with respect to Europe, on the basis of focal mechanisms and one of the several interpretations that we can get from the analysis of magnetic anomalies in the Atlantic. This intepretation of plate motion directions may be a cultural inheritance from the beginning of this century, when Argand (1924) demonstrated that the Alps were the result of N–S collision between the African promontory (the Apulia Plate) and Europe, in spite of E–W plate motion indicators including regularly orientated E–W transform faults in the Atlantic or NE–SW transform faults in the Indian Ocean. Since then most of the reconstructions of the Mediterranean have tried to unravel the plate kinematics forgetting the thousands of kilometres of N–S or NW–SE-trending thrust belts such as the Apennines or the Dinarides, or the eastward very fast (10–15 Ma) opening of back-arc basins (i.e. Tyrrhenian and Pannonian Basins). In addition, seismic data suggest that Africa was diverging from Europe towards the SW during the Neogene; the NE–SW-trending extension in the Sicily channel and on the Pelagian Shelf is an example. Doglioni *et al.* (1991) proposed a mainstream direction of plate motions in the Mediterranean (Fig. 1) which can account for the eastward opening directions of western Mediterranean basins, the trend of the inversion in the North Sea, the Rhine and Rhone grabens, the main E–W convergence in the Western Alps and dextral transpression in the Central-eastern Alps and in the entire Rif, Tell, Maghrebides and Sicilian thrust belt, the NE-dipping Dinaric–Hellenic–Cyprus subduction, and the WSW-dipping Carpathians subduction. This

Mediterranean–European undulation from E–W, NE–SW to NNE–SSW mimics the E–W trend of opening in the Atlantic to the NE–SW and NNE–SSW directions of motion in the Red Sea and the Indian Ocean, or similar directions of convergence in the Zagros and Himalayas. This trend is valid at least for the last 40 Ma in the Mediterranean and it is in very good agreement with the most recent geodetic data for Europe (Smith *et al.* 1990). N–S compression in the central-western Mediterranean is mainly a result of dextral transpression (northern Africa, central eastern Alps) or of the counterclockwise rotation of microplates (Apulia, Spain, Channell *et al.* 1979). Along transpressive or transtensive areas, the maximum stress may strongly deviate by up to 90° with respect to the plate vector (Zoback 1991). Therefore focal mechanisms which record the displacement due to those deviations of the stress field are not direct indicators of plate motion. The flow lines depicting plate motions in Fig. 1 do not necessarily imply the westward drift of the lithosphere, they simply link together the main directions of movement. A comparison between the average westward drift of the lithosphere in the hot spot reference frame and the tectonic flow is shown in fig. 1 of Doglioni (1992).

The hot spot reference frame and the 'westward' drift of the lithosphere

The existence of a hot-spot reference frame has been proposed by many authors (Morgan 1972; Burke *et al.* 1973; Wilson 1973; Morgan 1981; Crough 1983; Olson & Nam 1986; Davies 1988; Sleep 1990; Duncan 1991; and references therein), in order to characterize plate motions and constrain true polar wander. However there remain many controversial aspects about hot spots. What is their origin? From which depth in the mantle do they come from? Are they related to thermal or chemical anomalies? Are they fixed relative to one another? How can they kinematically relate to mantle convection? Recent discussions of these problems are given by Anderson (1989) and Wilson (1989).

Rejuvenating volcanic tracks at the Earth's surface may be a result of deep mantle plumes (e.g. Hawaii), retrogradation of a subducting slab, migration of a back-arc spreading centre, along strike propagation of a continental rift (e.g. East Africa), or propagation of transform faults with a transtensive component (e.g. Chagos?). These volcanic trails have different mantle source depths and therefore should be differentiated (Fig. 2). Indeed many of them should not be considered hot spot trails at all. Clearly propagating rifts (hot-lines, etc.) are only shallow phenomena which are not fixed to the deeper mantle. Recently, Bonatti (1990) has demonstrated that many so-called hot-spots may actually be 'wet' spots in which magma production is facilitated by higher water contents in the shallow upper mantle, which may be able to produce larger amounts of magma, from relatively colder mantle, without any particular need to involve a deep hot mantle plume. According to Bonatti (1990) some of these 'wet-spots' (e.g. Tristan da Cunha, Ascension, St Peter and Paul) might have sources coming from the upper mantle and not from the core–mantle boundary as supposed for Hawaii.

Differential rotation between the asthenosphere and the

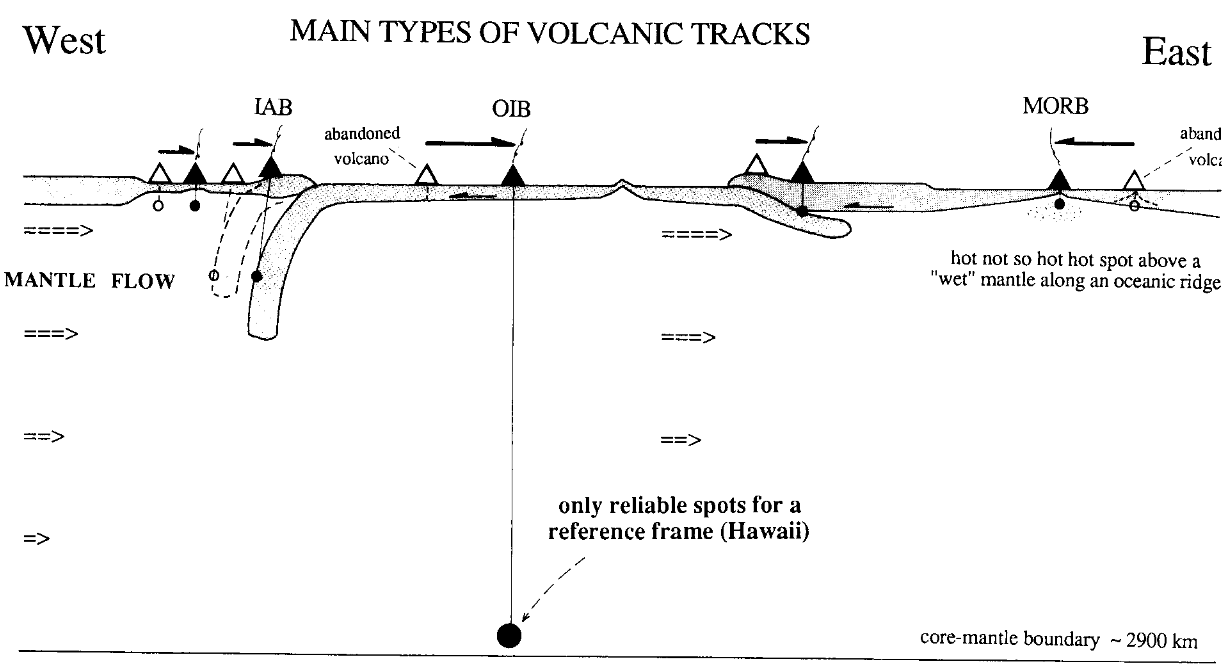

Fig. 2. The main volcanic trails at the Earth's surface have different source depths. The solid half arrows indicate the direction of migration of volcanism with time. Filled triangles represent the youngest volcanic products. Superficial hot spots such as some of those located along mid-oceanic ridges are not fixed because oceanic ridges are continuously moving with respect to one another, and thus the relative positions of the volcanic tracks are not reliable indicators of a fixed frame of reference. Other volcanic trails may be identified in a map view due to the propagation of a rift zone, without any implication for the motion of the lithosphere relative to the underlying mantle. If the hot spot reference frame is filtered, by removing such misleading tracks, the apparent westward drift of the lithosphere would increase by several centimetres per year. Light stipple, lithosphere; IAB, island arc basalts; OIB, ocean island basalts; MORB, mid-ocean ridge basalts.

lower mantle may be expected due to their different viscosities (Sabadini & Yuen 1989). Consequently, magmatic sources coming from different levels of the mantle cannot be used to define a unique hot-spot reference frame (Fig. 2) and the only hot spots which should be used are those which are plume-related, originating by upwelling from the same thermal boundary layer in the mantle (e.g. the core–mantle boundary). Burke *et al.* (1973) and Molnar & Francheteau (1975) have shown that the Atlantic and Indian hot-spots are not fixed with respect to each other, but that they may be fixed within a self frame or for short periods of time. It is evident that the oceanic ridges surrounding Africa have moved away from the craton during the opening of the Atlantic and Indian oceans. This indicates that any hot-spots associated with these oceanic ridges cannot be considered as being fixed with respect to any global reference frame. In a few cases, a double track on both sides of the oceanic ridge is present, which is sometimes 'V' shaped (e.g. Ascension, Tristan Da Cunha), indicating that the plume may remain stable beneath the moving ridge for long periods of time. Therefore the mantle source of such plumes might not be fixed relative to a global hot-spot reference frame.

Using the classically accepted hot spot reference frames, which include several hot-spots which may be wet-spots rather than hot spots, located along or near oceanic ridges (i.e. Galapagos, Easter Island, Ascension, Reunion) several authors (e.g. Le Pichon 1968; O'Connell *et al.* 1991; Ricard *et al.* 1991; Cadek & Ricard 1992) have demonstrated a westward component in the motion of the lithosphere relative to the asthenosphere of a few centimetres per year. However, if we disregard those hot-spot tracks located along oceanic ridges, the apparent relative westward drift of the lithosphere may increase by 3–4 cm a^{-1}. Contrary to the above, some of the shallow suspect hot-spots (wet-spots) appear to indicate an apparent westward motion of the mantle with respect to the lithosphere. This raises the question as to whether there really are places where the asthenosphere is moving westward with respect to the lithosphere, or is it simply that the lithosphere is moving westward at different velocities with respect to the asthenospheric mantle? If the lithosphere is entirely moving 'westward', any magma source in the asthenosphere must be moving relatively 'eastward' and passing below different sections of lithosphere. This question is still open and is clearly one of the most important problems to solve in geodynamics. The hot-spot reference frames have, thus far, been used somewhat uncritically, mixing together magmatic tracks at the surface which are supplied from different depths of the mantle. If this is true, the relative 'eastward–northeastward' mantle flow should be better defined. From this perspective the oceanic ridges around Africa may be seen as detached passive plate boundaries above an independent common flow (Pavoni, this volume).

The 'westward' drift implies that the lithospheric plates have a general sense of motion and that they are not moving randomly. If we accept this postulate, then the plates must be moving along this trend at different velocities toward the 'west' relative to the mantle. Rather than exactly west it would be better to say moving generally 'westward' (SW, WNW, etc.) along the flow lines of Fig. 1, which undulate and are not E–W parallel. On this basis the plates are more or less detached with respect to the mantle. The degree of decoupling is controlled by lithospheric thickness and composition, the thickness and viscosity of the asthenosphere, the nature of mantle convection (Liu *et al.* 1991) and the lateral variations of these parameters. When a plate moves faster towards the west with respect to an adjacent plate to the east, the resulting plate margin is extensional; when a plate moves faster to the west with respect to the adjacent plate to the west, their common margin will be of the convergent type (Fig. 3).

Subduction zones and thrust belts

It is common knowledge that the west-dipping subduction zones of the western Pacific margin are steeper and deeper, compared to the east-dipping shallower and flatter eastern Pacific slabs (Fig. 2): (Bostrom 1971; Nelson & Temple 1972; Dickinson 1978; Uyeda & Kanamori 1979; Uyeda 1981; Jarrard 1986; Ricard *et al.* 1991). A similar asymmetry is evident in the geometry of subduction zones associated with intracontinental collision belts (e.g. in the Mediterranean). This suggests the presence of a tectonic polarity on a global scale and supports the interpretation of an 'eastward' mantle flow. Different ore deposits are also associated with the opposite subduction styles (Mitchell & Garson 1981).

The differences between thrust belts are the most geologically impressive evidence for such a global polarity (Fig. 4). Thrust belts associated with subduction zones dipping westward with respect to the flow lines of Fig. 1 have low elevation (0–4000 m) and they have a shallower Moho beneath the thrust belt with respect to the foreland. Moreover they are composed of rocks scraped off the top of the subducting plate and basement rocks from pre-existing structural highs or orogenic belts. In some sections of such belts the crust may be missing. Associated with a west-dipping subduction zone there is always a back-arc basin, showing fast eastward progradation (50 km Ma^{-1}, e.g. the Tyrrhenian Sea). The west-dipping subduction zones (i.e. most of the West Pacific subduction zones, Barbados, South Sandwich Islands, Carpathians, Apennines) are considered to oppose the relative eastward–northeastward mantle counterflow that results from the 'westward' drift of the lithosphere. In fact they all form arcs verging to the east, as if they are obstacles to a mantle flow in the opposite direction. On the other hand, thrust belts associated with subduction zones dipping eastward or northeastward along the flow lines of Fig. 1 have high elevation (0–8000 m) and the Moho is deeper under the orogenic belt than the foreland. They involve lower crustal rocks and generally have a linear (rather than arcuate) trend, except where subduction has occurred along undulating continental margins. The E–NE–NNE-dipping subduction zones may be considered to follow the mantle counterflow related to the 'westward' drift of the lithosphere (i.e. East Pacific subduction zones, Andes, Alps, Dinarides, Taurides, Zagros, Himalayas, Indonesia). No 'classic' back-arc extension occurs in these geodynamic settings. In typical back-arc settings we observe fast opening and oceanization (within 10–20 Ma) toward the east, coeval with frontal accretion. This does not occur for example in the Indonesian or Aegean 'back-arcs' where there still is a thick continental crust in spite of long standing NNE- or NE-dipping

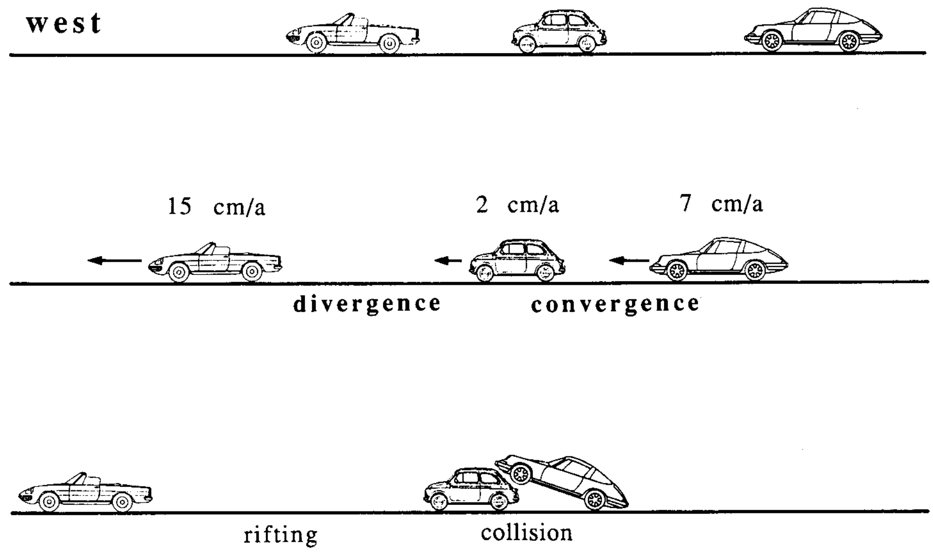

Fig. 3. Cartoon illustrating that plates (cars) are moving along a common trail (e.g. the flow lines of Fig. 1) but with different velocities toward the west, due to the general 'westward' drift of the lithosphere detected from the hot spot reference frame. These differential velocities control the tectonic environment. When the western plate is moving westward faster with respect to the plate to the east there is extension, while convergence occurs when the plate to the east is moving westward faster with respect to the plate to the west. Note that when the car in the middle is 'subducted', the regime will be in extension because the car to the west is moving faster. Changing the hypothetical velocities, the car on the left could be compared to the West Pacific plate, the smaller one in between would be the Nazca plate and the car on the right the South American plate. The reference frame could also be inverted if we take the relative eastward–northeastward motion of the mantle: the small car in the centre would become the fastest toward the 'east'.

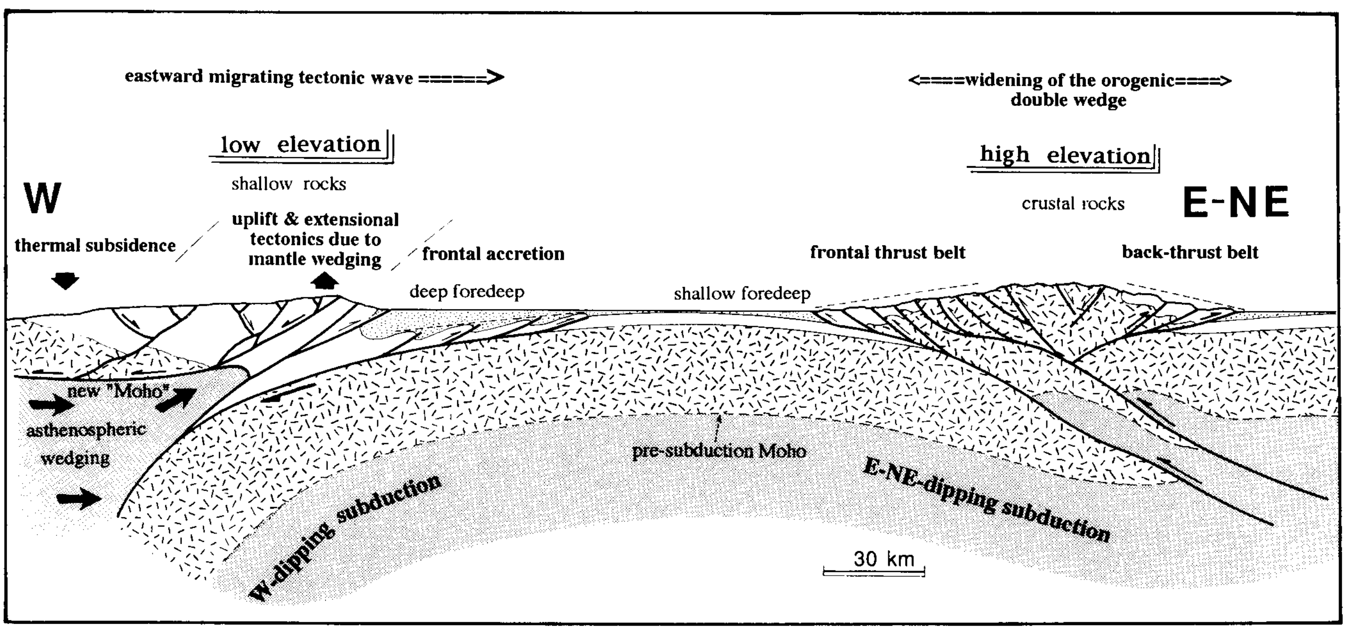

Fig. 4. The difference between thrust belts related to west-dipping subduction (left) and east- or NE-dipping subduction (right) provides striking evidence for the existence of a global tectonic polarity. The thrust belts related to west-dipping subduction (West Pacific accretionary wedges, Barbados, South Sandwich Islands, Carpathians and Apennines) have low elevation. In contrast, thrust belts related to east- or NNE-dipping subduction (Andes, Alps, Dinarides, Taurides, Zagros, Himalayas) are characterized by high elevations. On the left mainly shallow rocks are accreted in the frontal accretionary wedge (apart from remnants of crustal rocks from earlier east-dipping subduction); there is a deep foredeep with high subsidence rates and the uplifted zone is in extension, controlled by the vertical push of the underlying asthenospheric wedge. On the right hand side the mountain range is made by crustal rocks and there are shallow foredeeps in front of the frontal and back-thrust belts. This polarity may be explained in terms of the different behavior of the basal plate decollements in the two subduction systems, opposing (west-dipping) or following (E–NNE-dipping) the eastward–northeastward mantle flow. Any magmatism has been omitted for sake of simplicity. Modified after Doglioni (1992).

subduction (e.g. Meulenkamp *et al.* 1988) which was active without active 'back-arc' extension. In these back-arc basins extension propagates southwestward and at slower rates, and is not always contemporary with the NE-dipping subduction.

The main differences between the two types of thrust belt (Fig. 4) may be interpreted as resulting from the polarity induced by the different behavior of the decollement planes associated with subduction opposing or following the mantle flow. In the first case the decollement at the base of the lithosphere is subducted and only shallow upper layers of the lithosphere are accreted. In the second case, the decollement at the base of the lithosphere ramps upward toward the surface and allows the uplift of deep seated rocks (Fig. 5). For a more detailed description of this see Doglioni (1992). The gravity and heat flow profiles across the two end-member types of subduction are also very different. In the Apennines (west-dipping subduction zone), for example, the gravity minima are located more toward the foredeep, rather than exactly below the mountain range (Mongelli *et al.* 1975), while they are more coupled with the topography in the Alps (Nicolas *et al.* 1990). In the past decade, seismic data have provided a powerful tool in comparing continental margins all over the world (e.g. Bally 1983; von Huene 1986). Such data may be used to demonstrate that the envelope to the fold crests may dip toward the trench in the frontal accretion zone of west-dipping subduction systems, while it rises toward the hinterland in thrust belts related to E–NE-dipping subduction (Fig. 4).

Foredeeps are also very different depending upon whether they are associated with thrust belts characterized by west-dipping subduction zones or E–NE-dipping subduction zones. Along west-dipping subduction zones the subsidence rate of the foredeep may be of the order of $1600\,\mathrm{m\,Ma^{-1}}$ (e.g. Apennines and Carpathians), while along E- or NE-dipping subduction zones they are of the order of $300\,\mathrm{m\,Ma^{-1}}$ or less (e.g. Alps and Dinarides, Doglioni 1992). This observation enables us to interpret the slow filling of foredeeps of the first type with huge flysch deposits (Apennines) or limited deposition (Mariana trough), and the much faster filling of foredeeps of the second type which are quickly filled by the classic flysch-molasse sequences and then by-passed. In the first case the origin of the foredeep appears to be controlled by the eastward roll-back of the subduction hinge resulting from the eastward mantle push. In the second case the westward or southwestward roll-back of the subduction hinge is probably controlled by the lithospheric load and by the downward component of the advancing upper plate, contrasting with the upward component of the eastward–northeastward mantle flow. The cross-sectional area of thrust belts associated with west-dipping subduction is smaller in comparison with the area of the foredeep. The area of thrust belts associated with E–NE-dipping subduction zone is instead always greater than the area of the foredeep (Doglioni in press). All these observations appear to maintain their general validity both in the case of subduction of oceanic lithosphere and of thin continental lithosphere.

Most of these differences have been so far explained in terms of slab pull and the age of the subducting oceanic lithosphere, or rates of convergence between plates (e.g. Royden & Burchfiel 1989). However, in the Mediterranean, the Adriatic continental lithosphere and the Ionian oceanic lithosphere are subducting both under the Apennines (steep west-dipping subduction) and under the Dinarides–Hellenides (shallow NE-dipping subduction). The two related thrust belts follow the east and west Pacific rules, without age and thickness variations of the subducting lithosphere. In the Pacific itself, the west-dipping subduction systems are the fastest in the world and the slab is steep, while the Andean subduction zone has been active since the Mesozoic and the slab is shallow. Those examples are clearly in contradiction with the age of the subduction and the age and thickness of the subducting lithosphere being first order controls of the tectonic style. The westward drift of the lithosphere could explain these asymmetries.

Lateral heterogeneities

In this section it is proposed that physical anisotropies within and beneath the lithosphere may exert an important control on plate motions. These include: (a) thickness variations of the lithosphere (ranging between 20 and 300 km); (b) density variations of the lithosphere (i.e. continental and oceanic); (c) the lateral viscosity variations in the asthenosphere at the base of the plates. A plate may be considered as a part of lithosphere with a different angular velocity with respect to the neighbouring lithosphere, the velocity being controlled by the amount of basal drag. Sabadini *et al.* (1992) have shown that the drag at the base of the lithosphere is direct function of the dimensions and composition (i.e. continental versus oceanic) of a lithospheric plate. In general the continental lithosphere is thicker than the oceanic lithosphere and is possibly more firmly anchored to the underlying mantle.

The preceding section discussed the evidence that the average direction of motion of the lithospheric plates is westward directed, confirming the notion of a general 'eastward' flow of the asthenospheric mantle with respect to the lithosphere (Bostrom 1971; Nelson & Temple 1972; Doglioni 1990, 1991*a*). In this section some examples of the effect of lateral lithospheric anisotropies on plate motions are presented. A continental root may induce the onset of extension at its western margin (Sabadini *et al.* 1992), due to the bigger drag effect of a thicker lithosphere anchored to the mantle. Consequently the overthickening induced by collisional events (orogens) might automatically generate extension, and we do observe this in the repetition of the 'Wilson cycle'. After a compressive event (e.g. the Caledonian and Variscan orogens) a new rifting cycle may start (e.g. the Atlantic and Tethys rifts; Bernoulli & Lemoine 1980; Dercourt *et al.* 1986; Ziegler 1988, 1990) if the lithospheric overthickening is accompanied by the relative westward drift of the lithosphere with respect to the mantle and lateral asthenospheric anisotropies induce velocity variations in the overlying lithospheric plates. Unfortunately, there are no indications of the direction of relative mantle flow in pre-Mesozoic time. However, indirect assumptions could be made on the basis of the lithologies and morphologies of the Palaeozoic thrust belts.

Linear rift zones may occur in continental or oceanic environments. In contrast back-arc basins (Brooks *et al.* 1984) tend to be semi-circular (Caribbean, Scotia Sea, Parece Vela Basin, Japan Sea, Pannonian and Tyrrhenian Basins, etc.) with stretching taking the form of a migrating wave forming semicircles toward the east, in contrast to the linear rifts which have very different styles of extension. Back-arc extension related to west-dipping subduction zones is characterized by high subsidence rates which may reach 500-700 m Ma^{-1} (e.g. Tyrrhenian and Pannonian basins), while linear intra-continental rifts and passive margins have much lower values (50–300 m Ma^{-1}, e.g. Atlantic and Tethyan margins). In other words, the margins of a back-arc basin (particularly the western one) undergo the fastest stretching and are clearly asymmetric, because extension is eastward propagating, due to the fast (3–7 cm a^{-1}) roll-back of the subducting slab. 'Linear' rifts such as the Atlantic rift or the East Africa rift tend to follow a N–S trend which has been stable since the Mesozoic. In contrast extensional or convergent zones of Mesozoic age in the Tethyan domain follow the present undulation of the global flow (Fig. 1), from E–W directions of motion in the Atlantic to NE–SW toward Asia, as suggested by synsedimentary normal faults and Cimmeride convergence. Inherited Palaeozoic and Mesozoic lateral heterogeneities of the lithosphere, due to rift or collision zones, may have controlled the location of later tectonic activity.

The biggest stress accumulations occur at the strongest lithospheric anisotropies (i.e. the continental margins) where thickness and compositional gradients are maximum. In fact we observe that, in the past, subduction zones began at continental margins and were controlled by their geographical distribution. An east-dipping subduction zone developes only if oceanic lithosphere is located to the west of thicker continental lithosphere, e.g. the Andean subduction system, which has been active for several hundred million years because of the existence of thinner oceanic lithosphere to the west of it (i.e. Nazca plate). Conversely, we observe that west-dipping subduction zones are initiated only where a thin lithosphere is positioned to the east with respect to a thicker continental lithosphere, i.e. all the west-dipping subduction zones of the western Pacific margin occur to the east of the thick Asiatic continental lithosphere (Doglioni 1991*a*, 1992). When relatively thick lithosphere arrives at a subduction zone, the subduction may flip in the other direction. For example, at Timor the NNE-dipping subduction zone carrying the Australian continent appears to have switched to a SSW-dipping subduction system with the arrival of the thick continental lithosphere at the trench (Price & Audley-Charles 1987), contrasting with the relatively thinner lithosphere to the north.

During west-dipping subduction, there is compensating mantle flow into the back-arc area, explaining the high heat flow values and the generation of new oceanic MORB-type crust. The evolution of a back-arc automatically determines the generation of lithospheric anisotropies at its western and eastern margins. These thickness and composition gradients can control the development of new subduction zones at those new margins: west-dipping to the west (e.g. the Nankai trench) and east-dipping to the east (e.g. the Taiwan–Manila–Borneo trench). In general, marginal basins related to west-dipping subduction systems usually have a short life (20–40 Ma?) because once formed, sooner or later they close due to the development of an east-dipping subduction zone (Fig. 5). Kinematically, during west-dipping subduction, asthenospheric mantle wedges at the top of the subduction hinge determine rates of uplift and extension (Fig. 5, upper section). Only the superficial layers of the subducting plate are accreted in the compressive sector. These are delaminated from the underlying crust and uplifted by the mantle wedging, lying above a newly forming Moho (Doglioni 1991*b*). When an east-dipping subduction later cross-cuts the structure, it may uplift deep seated rocks, and the former mantle wedge may become a relict within the new orogen (Fig. 5).

An east-dipping subduction zone remains active until the lithostatic load or vertical shear becomes greater with respect to the horizontal shear. This value decreases as a consequence of the subaerial erosion of the upper plate and results in minor collision reactivations (Fig. 6). If the system locks due to the lithospheric overthickening, a new west-dipping subduction zone may be initiated if there is a thinner lithosphere to the east and the eastern plate is still moving faster toward the west. This seems to be the case in the Neogene Tyrrhenian–Apennines system related to west-dipping subduction which developed after the Alpine closure to the west; or the Pannonian–Carpathian system also associated with west-dipping subduction, developed between Early Miocene and Pliocene after the Alpine and Dinaric closure to the west. Both systems had the opportunity to grow due to the presence of a thinner lithosphere to the east of the collided area. A similar pattern of evolution may be traced in Japan, where a Late Tertiary west-dipping subduction zone overprinted an earlier east-dipping subduction related orogen; back-arc extension disrupted and disconnected the former thrust belt from the Asiatic continent.

Doglioni (1991*b*) proposed that the asymmetry of the Apennines–Tyrrhenian system is related to lithospheric longitudinal variations along the subduction zone. The Apenninic subduction system is highly irregular in map view, and the asymmetry appears to be controlled by the composition and thickness of the subducting plate; subduction reaches a maximum in the south, where, to the east of the trench, the Ionian oceanic lithosphere is present. The amount of subduction decreases northward, particularly at the transition of the Ionian lithosphere and the Adriatic lithosphere, which is continental in origin and thicker.

All these observations suggest that physical anisotropies in the lithosphere are a major controlling factor in plate tectonics. Lithospheric overthickening may control the subsequent location of rift zones while lithospheric thinning may control the location of subduction zones. A 200–300 km thick continental lithosphere surrounding the entire planet would inhibit any kind of relative motions. Basically, plate tectonics occurs because lateral lithospheric anisotropies allow rift and subduction zones to develop.

The 'triangle' effect

The model becomes more complete if we observe plate motions not only in cross-section, but also in a map view. Continental plates very often have irregular shapes in map view (Fig. 7). Along lines of longitude, perpendicular to the interpreted westward flow of the lithosphere, plates may be wider and thicker with respect to other extreme parts of the plate. If we consider the continental plate schematically as a triangle, the smaller part of the plate should have a relatively minor coupling with the mantle. So the thinnest and smaller areas of the plate should move westward faster because they are the parts of lithosphere most decoupled from the underlying mantle. However in these computations we have also to consider the decreasing velocity of plates toward the polar regions.

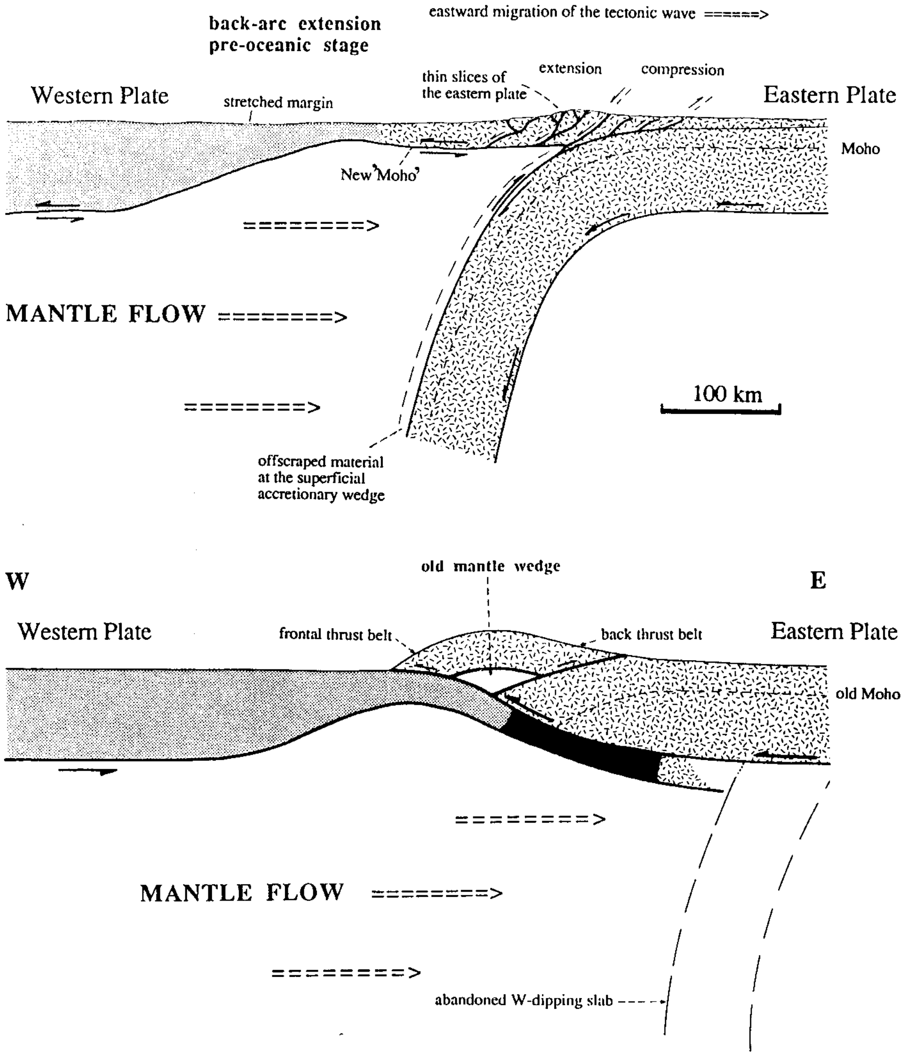

Fig. 5. A west-dipping subduction zone is associated with a back-arc basin, due to the 'loss' of subducted lithosphere, which is compensated for by the 'eastward' mantle flow (upper section). The asthenospheric wedging at the top of the subduction hinge is responsible for uplift of the arc and for the frontal accretion of upper layers of the crust. Thermal subsidence occurs in the back-arc while the extension-compression wave propagates 'eastward'. The back-arc extension generates two new thickness and composition discontinuities, at the western and at the eastern margins, which may control the onset of new subduction systems, west-dipping at the western margin, or east-dipping at the eastern margin. The lower section shows the case of an east-dipping subduction zone superimposed upon an earlier west-dipping one. This type of subduction may cause uplift of deep-seated rocks and in a case like this it will close the back-arc basin (black). Note that kinematically we might expect a part of the mantle wedge of the earlier subduction system to be trapped in the hangingwall. The basal detachment of the eastern plate is subducted in the upper case and only shallow rocks are offscraped from the top of the lithosphere. In the lower section the detachment at the base of the eastern plate is instead ramping upward towards the surface and may uplift deep crustal rocks.

If we take the shape of the South American continental lithosphere which is tapering toward the southern tip, we may consider it as an irregular triangle (Figs 7 & 8). This fact should induce a different coupling of the lithosphere to the mantle between the northern and southern parts of the South America: a thicker lithosphere to the north generates a greater anchorage to the mantle counterflow with respect to the southern parts which should move westward faster because of the smaller drag at the base of the lithosphere, generated by the smaller area and thickness of the continental lithosphere. This produces different westward velocities of the plate from north to south which in turn generates rotation of the plate. If we apply this simple model to the South American continent we would predict clockwise rotation due to the southward decrease in the dimension of the continental lithosphere with respect to the 'westward' flow of the lithosphere or 'eastward' mantle counterflow (Fig. 8). This is in agreement with the known clockwise rotation of South America (Rabinowitz & La Brecque 1979), the wider opening of the southern Atlantic since Late Cretaceous and the coeval N–S compression at the northern margin of the South American continent, i.e. the South Caribbean–Venezuela thrust belt, and the bending of the Romanche fracture zone. The Barbados arc is in fact deformed along its southern arm by this northward directed protrusion. The wider opening of the Southern Atlantic is not a simple effect of the cartographic projection. Moreover toward the poles rifts and subduction zones tend to decrease their rate of movement. The Southern Atlantic does not follow this rule.

A similar, but reverse triangular shape may be applied to the Indian continent, although this is rather more subjective. During its NNE motion, India had its longest segment with respect to the direction of motion in the eastern part (Fig. 8), the shorter segment along the western margin has now collided along the Chaman left lateral transpressive belt. This general triangular shape could have controlled the counterclockwise rotation of India (e.g. Dewey *et al.* 1989). The African craton has a triangular shape, but this is not related to the global flow of plate motions: sections traced parallel to the proposed flow do not show significant length variations. Instead Africa is moving counterclockwise passing throughout the major undulation of the global tectonic flow (Fig. 8). However during this motion, a minor clockwise rotation might have been superimposed and may be indicated by the northward decreasing divergence of the Red Sea rift.

This 'triangle' effect works for continental roots, but it

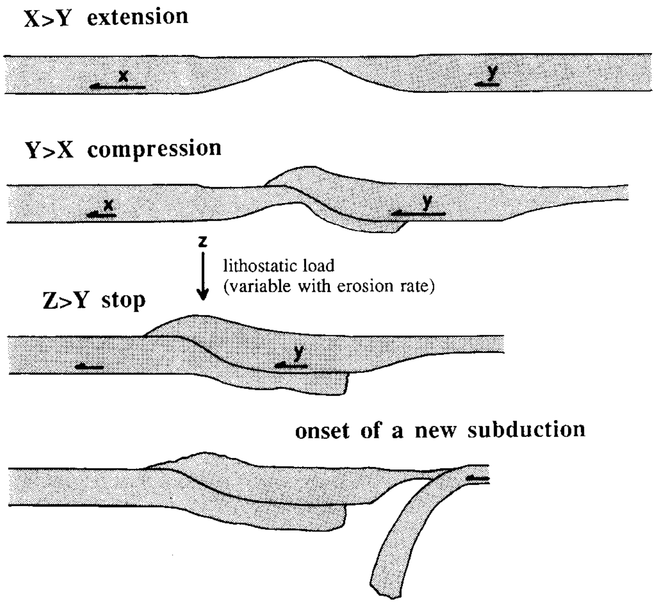

Fig. 6. Subduction zones are also controlled by the distribution of lithospheric gradients in thickness and composition. An east- or NE-dipping subduction will continue to act until the lithosphere to the west (or SW) is relatively thinner and the eastern plate moves westward faster ($Y > X$). When the collision generates a lithostatic load greater with respect to the horizontal shear ($Z > Y$), the subduction will stop. Small reactivations may occur due to depressurization induced by the erosion. Once the system is closed, a new west-dipping subduction may start if there is a thinner lithosphere to the east moving faster westwards.

seems to be valid also for oceanic plates. The Cocos plate for example is wider in its southern part. This should enable a counterclockwise rotation of the plate which is evident by the spreading along the Galapagos ridge (Searle & Hey 1983), which decreases to zero westwards (Lonsdale 1988). Coeval transpression occurs along the central American margin. Along the East Pacific rise there is no compression due to the Cocos plate rotation because the Pacific plate is moving westward too fast. Other triangular plates may have undergone similar rotations (e.g. North America?). This is something which clearly requires further investigation. The geological evidence for the rotation of triangular plates also indirectly supports the notion of a global flow of plate motion.

In the case of E- or NE-dipping subduction, when there are thickness and composition variations along strike in the footwall (but they could also be in the hanging wall) of the subduction zone, the section in which continental collision occurs earlier will slow down, while along sections in which there still is thinner oceanic lithosphere in the footwall, the convergence rate will continue to be relatively high, rotating the hanging-wall plate (Fig. 9). These kinematics might be applied to the Dinaric–Hellenic subduction system; here the northern sector of the accretionary wedge reached a high lithospheric thickness during the Early Miocene and the subduction has effectively ceased in Friuli and Slovenia since that time. Thrusting and subduction are still active from Albania southward because thin crust in the footwall to the southwest (Southern Adriatic and Ionian Seas) remaining to be subducted underneath Greece.

Conclusions

There are geological and geophysical lines of evidence for a global tectonic polarity both in the direction of plate motions (Fig. 1) and in their sense of motion which supports the concept of 'westward' drift of the lithosphere relative to the mantle (Figs 2 & 3). Geodetic satellite data (Smith *et al.* 1990) are in very good agreement with the direction of plate motions that we can infer from tectonic data (Doglioni 1990). These are particularly unexpected for the European and Mediterranean areas for which a main direction trending NE–SW has been proposed, about 90° from the predictions of classic plate tectonic models for the Alpine collision zone. Although it has not yet been entirely physically demonstrated, there seems to be some control on this global tectonic pattern by the Earth's rotation.

Many plate tectonic models have been based on the hot-spot reference frame. However this frame may be biased by an inappropriate mixing of data from hot-spots coming from different levels of the mantle (Fig. 2). Some well-known plumes may originate within the upper parts of the mantle whereas others may be sourced at the core-mantle boundary. If this is true, the apparent rate of 'westward' drift and the true polar wander should be recomputed. Nevertheless, the westward drift of the lithosphere appears to be confirmed even using the most unfavourable hot-spot reference frames.

The differences between eastward- and westward-dipping subduction zones and related thrust belts are internally consistent with a globally orientated lithospheric flow pattern (Fig. 4). The importance of lateral thickness and compositional variations within the lithosphere are stressed as first order factors in controlling plate geodynamics because they directly determine the lateral variations of decoupling with the underlying asthenosphere (Figs 5 & 6). A west-dipping subduction system may generate a back-arc basin which introduces two new lithospheric gradients at its margins. A west-dipping subduction may then start at the western margin of the back-arc basin, or an east-dipping subduction may develop at the eastern margin. In this second case, the mantle wedge at the top hinge of the earlier west-dipping subduction may be cut and passively transported up by thrusting.

The shape of plates can directly control rotations and deviations of plate motions with respect to the main global flow. In fact the coupling between the basal lithosphere and the asthenosphere is greater when the lithospheric root is wider and deeper, here referred to as the 'triangle effect' (Fig. 7). This appears to be particularly true for plates containing deep continental roots, but may also be applied for oceanic plates (Fig. 8). Along strike variations in thickness and composition of the plate (perpendicular to the main direction of motion) induce differential degrees of decoupling and as a consequence rotation takes place. This may explain for example, the clockwise rotation of South America (with the wider opening of the southern Atlantic and coeval N–S compression in the southern Caribbean–Venezuela thrust belt, Fig. 8). Another case is the variation along strike of the thickness and composition of a subducting plate. Subduction will terminate when the lithostatic load becomes greater with respect to the horizontal shear, but if along strike there still is thin lithosphere to subduct, the convergence rate may be greater, inducing rotation of the upper plate (Fig. 9).

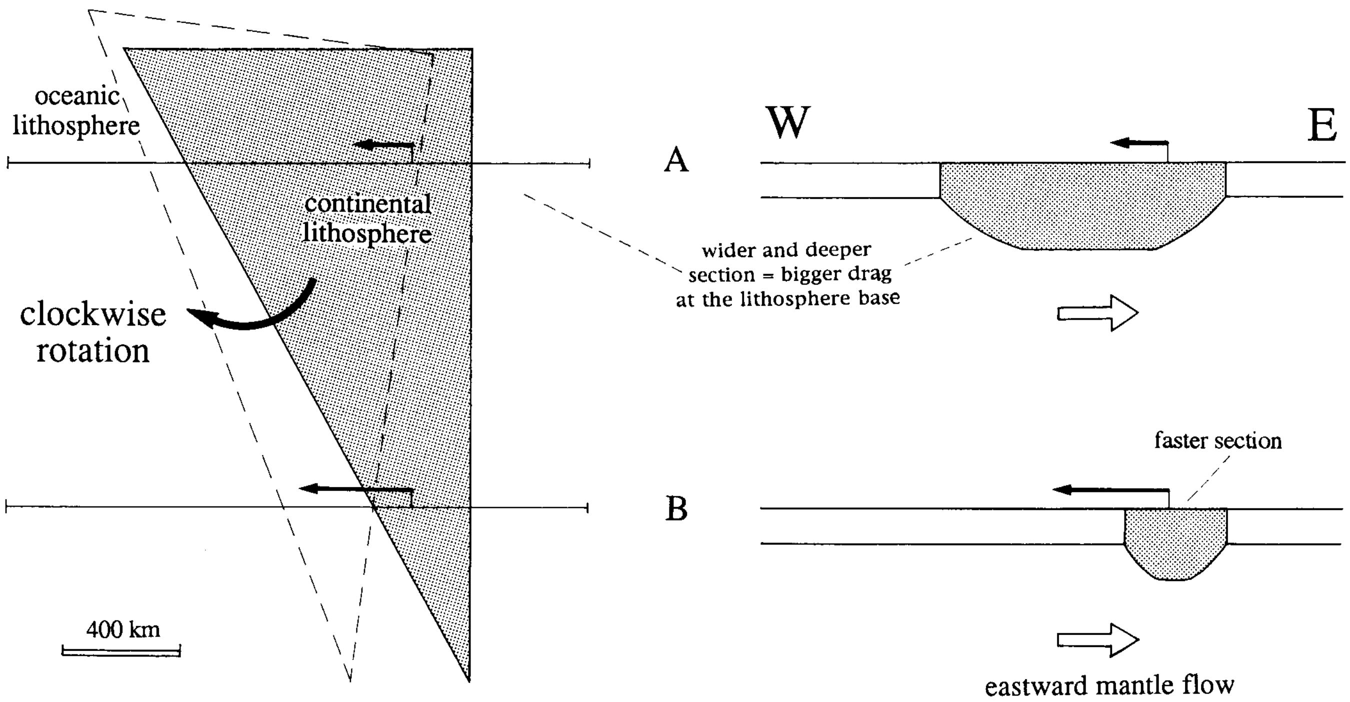

Fig. 7. The areal shape and thickness of the lithosphere controls the drag at its base. If we introduce longitudinal thickness and dimensional variations to the plate, we obtain different relative westward velocities with respect to the mantle. The figure shows a map view on the left side and cross-sections on the right. The upper part of the plate (A) has a bigger continental cross section and it is slower with respect to the lower part (B) due to the greater drag at its base. The 'triangle' effect is to generate rotation of the plate, in this case clockwise. A similar interpretation may account for several examples of plate rotation, such as the clockwise rotation of South America.

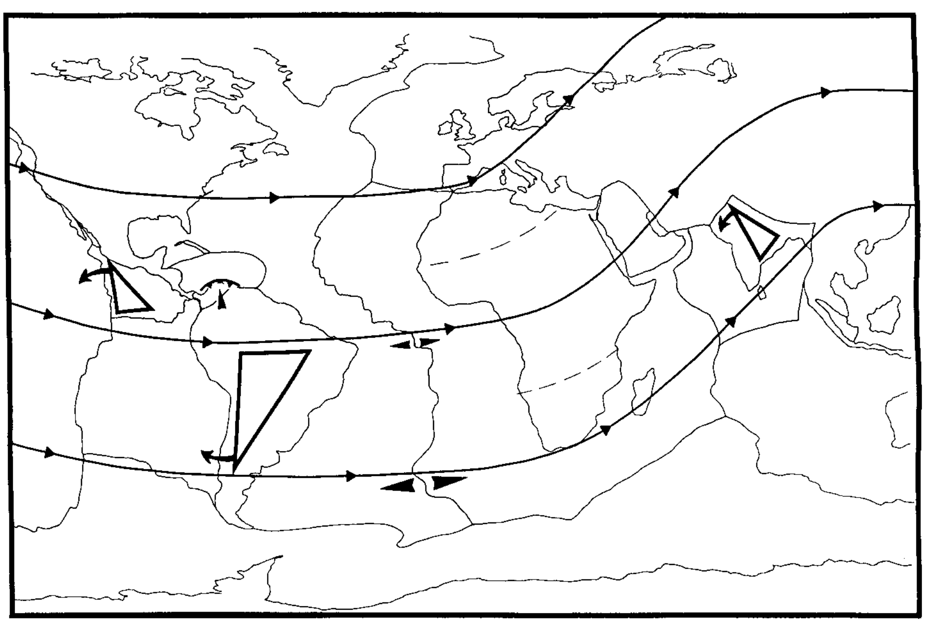

Fig. 8. Tentative application of the triangular shape of plates in controlling well documented plate rotations. The lines represent the inferred mantle flow over which plates are moving at different velocities, and moving relatively 'westward' due to variable decoupling. The triangular shape has to be seen with the long axis of the triangle perpendicular to the eastward or northeastward mantle flow. The clockwise rotation of South America is responsible for the Late Cretaceous–Tertiary faster opening of the southern Atlantic ocean and the coeval compression in the Panama–Southern Caribbean–Venezuela thrust belt. The Indian plate rotated counterclockwise during its general north-northeastward motion; with respect to the global flow lines such a rotation should be expected because the Indian continental lithosphere is shorter in its western part with respect to the mantle flow. A similar interpretation might be applied to the Cocos plate, even in a completely oceanic environment, the triangular shape controlling the counterclockwise rotation responsible for the eastward increase of the extension rates in the Galapagos ridge and the coeval compression in the Nicaragua–Costa Rica trench.

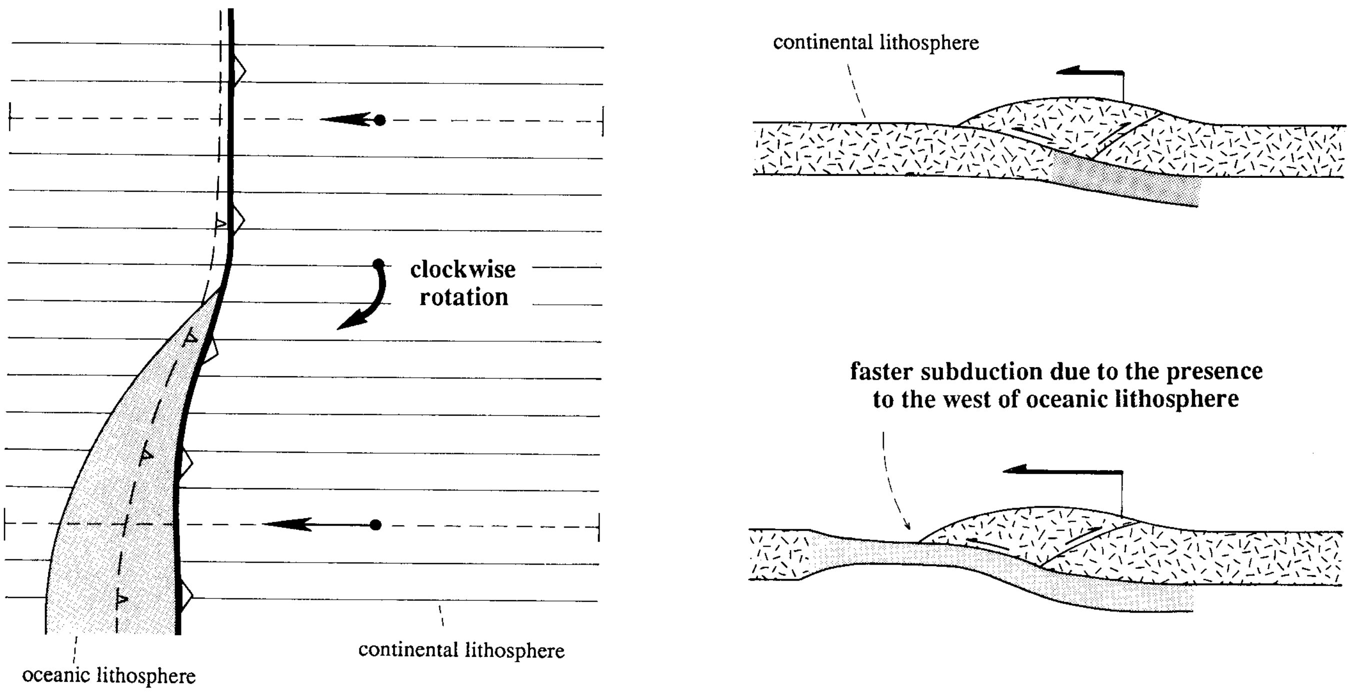

Fig. 9. Subduction zones are controlled by gradients in lithospheric thickness and composition. A subduction system which is characterized by along strike lithospheric variations will result in different rates of convergence. To the left of the figure, the northern arm of the subduction system is almost over because the plate to the west is very thick and continental in origin, while to the south the subduction is still very active because to the west there is still thinner oceanic lithosphere. This will control the clockwise rotation of the eastern plate and the undulation of the subduction zone and the related thrust belt with right lateral transpression along the transition between the oceanic and continental lithosphere of the western subducting plate. This model could be applied to the Dinaric–Hellenic subduction zone.

Therefore the analysis of the distribution of lithospheric anisotropies is important in order to unravel the rotation mechanisms of plates. The amount of subduction, or the relative velocity between plates, is controlled by the magnitude of the lithospheric gradients in the three-dimensional field.

We may interpret the Eulerian description of plate motion as resulting from a combination of the motion of one plate along the global flow lines (Fig. 1) and the rotation induced by lithospheric anisotropies, e.g. the triangular shape or the variations along strike of a subducting plate. Such rotations also control the generation of triple junctions over the Earth's surface. The global flow of plate motion needs further testing and the second step is to describe it more carefully, especially in the polar regions and at the Asiatic zone of deflection. Ultimately the amount of relative motion between lithosphere and asthenosphere should be quantified in centimetres per year for different parts of the Earth.

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