Tethyan closure, Andean orogeny, and westward drift of the Pacific Basin

Laurent Husson [a,*], Clinton P. Conrad [b], Claudio Faccenna [c]

[a] *Géosciences Rennes, Université Rennes 1, Université Rennes 1, Rennes, France*
[b] *Department of Earth and Planetary Science, Johns Hopkins University, Baltimore, Maryland, USA*
[c] *Dipartimento di Scienze Geologiche, Universita Roma Tre, Rome, Italy*

ABSTRACT

The deep-sea trenches that circumscribe the Pacific basin are moving westward relative to the deep mantle by a few cm/yr, together with the Pacific seafloor. While the westward drift of the Pacific basin dominates the observed net westward rotation of Earth's tectonic plates relative to the deep mantle source of the Pacific hotspots, its cause remains uncertain. We demonstrate that the source of this westward drift is tectonic in origin: the westward push of the high Andean topography along the eastern Pacific margin is large enough to shear the entire Pacific upper mantle with a surface velocity of ~30 mm/yr. In addition, because the Pacific upper mantle is confined between vertically coherent slabs of subducted lithosphere on both sides of the Pacific, slabs in the western Pacific are forced to advance toward upper plates because the eastern Pacific slabs are pushed westward. This cross-Pacific interaction provides an explanation for trench advance, which is observed in the western Pacific and is not easily explained by the local dynamics of subduction in which dense sinking slabs essentially tend to retreat from overriding plates. Thus, the basic asymmetry of the Pacific system, which has been noted for almost all tectonic features including slab dips, trench motions, and upper plate tectonics, is the result of one-sided Andean forcing on the volume of Pacific upper mantle. The net rotation of the lithosphere is then the result of surface tectonics: westward motion is an accident resulting from the current distribution of tectonic plates, and was triggered by the mid-Cenozoic closure of the Tethys Ocean, via a westward expansion of the Atlantic and Andean cordilleran building.

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1. Introduction

The tectonic plates that cover the surface of the Earth are, on average, drifting westward relative to the locations of hotspots that are tied to the deep mantle (Bostrom, 1971; Uyeda and Kanamori, 1979; Doglioni, 1990). This westward drift, a consistent feature of kinematic reconstructions (Minster and Jordan, 1978; Gordon and Jurdy, 1986), is reported to be occurring at rates as large as 48 mm/yr in the HS3 (Pacific hotspot) reference frame (Gripp and Gordon, 2002)] and is dominated by the rapid westward motion of the large Pacific plate (Fig. 1a). Although other hotspot reference frames exhibit smaller magnitudes of net lithosphere rotation, all choices require a net westward motion of the lithosphere relative to the deep mantle (see Becker, 2006 for a review).

The net rotation is a puzzling feature of Earth's surface tectonics because convective mantle flow, which drives plate motions, cannot induce a net surface rotation unless some plates are more tightly coupled to this flow than others. While net rotation of the lithosphere can be induced in numerical models that include thick roots beneath continental cratons (Ricard et al., 1991) and temperature and strain-rate dependent viscosities (Becker, 2006), these models predict

amplitudes of net rotation that are generally smaller than amplitudes constraints supplied by hotspot reference frames (Zhong, 2001; Becker, 2006). Furthermore, these models do not explain basic differences between the tectonics of the western Pacific, where subducting slabs are steeper and associated with extensional upper plate settings ("Marianas" type of Uyeda and Kanamori, 1979) compared to the eastern Pacific with shallower slabs ("Chilean" type of Uyeda and Kanamori, 1979) and compression that forms Andean-style mountain belts.

It has nevertheless been proposed that the westward motion of the Pacific system could be correlated to the cross-Pacific tectonic dichotomy (Ricard et al., 1991; Doglioni et al., 1999). The net rotation of the lithosphere and westward motion of the Pacific plate have also been related to the contrasted migration behavior of trenches, with predominantly advancing trenches along the western margin of the Pacific and retreating trenches along its eastern margin (Carlson and Melia, 1984; Heuret and Lallemand, 2005; Faccenna et al., 2007). The correlation between subduction kinematics and local intrinsic sub-duction properties such as slab age-buoyancy is not straightforward (Jarrard, 1986; Doglioni et al., 1999; Lallemand et al., 2005). However, a positive correlation between slab dip and the dynamics of the upper plate (Lallemand et al., 2005) suggests consideration of the possibility that upper plates may exert an external forcing not only on cross-Pacific subduction dynamics, but on the entire Pacific system.

* Corresponding author.
  E-mail address: laurent.husson@univ-rennes1.fr (L. Husson).
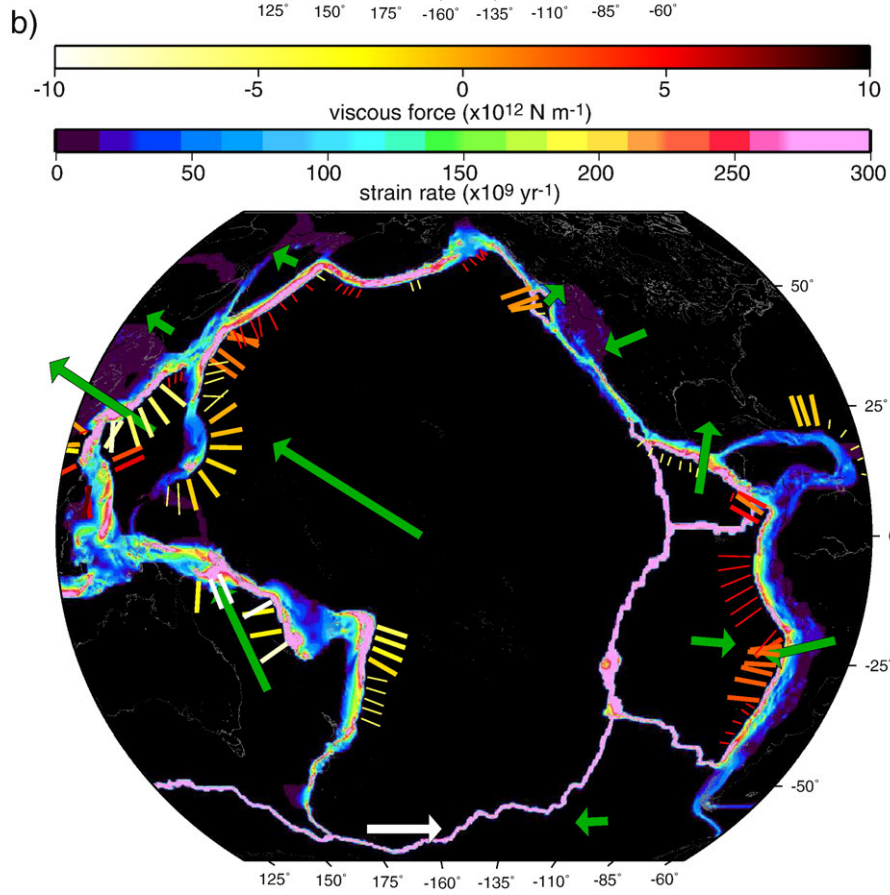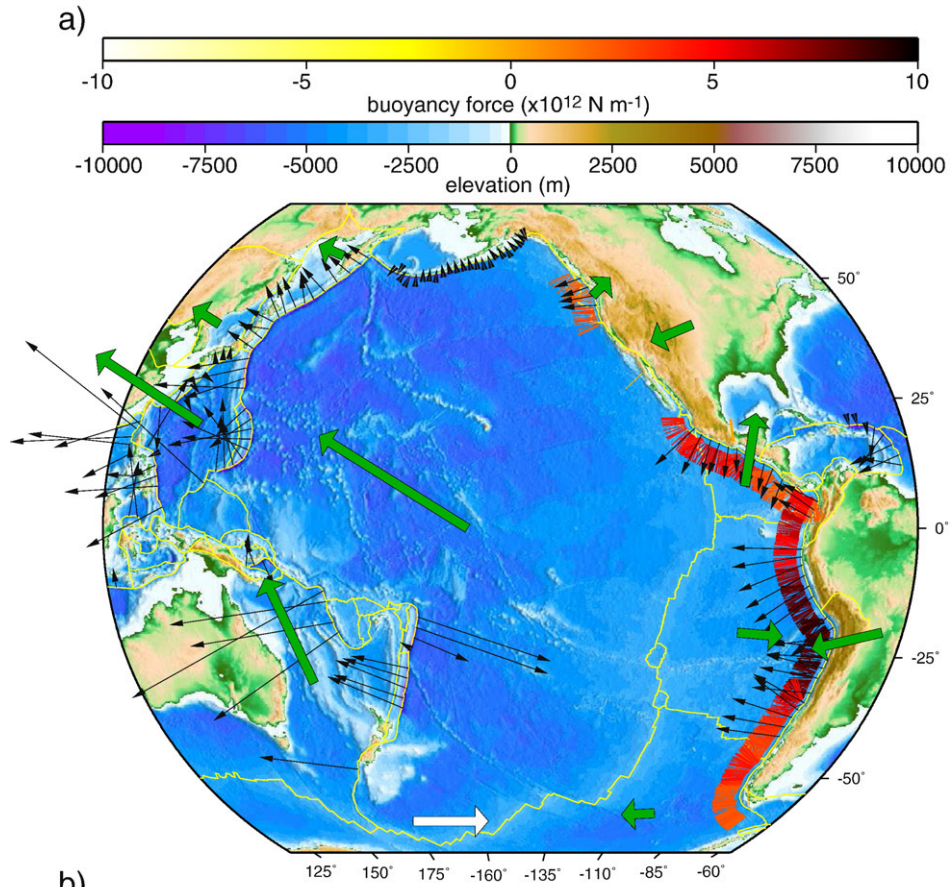
We therefore present an original view of tectonic forcing on the Pacific upper mantle that explains both the net rotation of the lithosphere and cross-Pacific differences in subduction kinematics. In this view, subduction zones are explored within a global framework that separates the upper mantle, slabs and lithosphere into two prominant and interacting, convection cells: the Pacific, or Pantha-lassan, and the Indo-Atlantic, or Pangean, domains (Collins, 2003; Davaille et al., 2005).

2. Net rotation of the Pacific system

A tectonic explanation for westward drift is suggested by the absolute motions of trenches, which are found where plates bend and enter the subduction system. Subduction in its most basic form involves a plate following itself into the mantle. Trenches can easily retreat toward the subducting plate; by contrast, trench advance (toward the overriding plate) is gravitationally unfavorable because it requires mantle to fill in the space beneath the slab, and requires specific boundary (Bellahsen et al., 2005) and rheologic (Billen and Hirth, 2007; Faccenna et al., 2007) conditions. A recent global survey of trench migration rates (Lallemand et al., 2005), however, shows that a plethora of trenches in the western Pacific advance at fast rates (Fig. 1a) in the Pacific hot spot reference frame (model HS3-NUVEL 1A, Gripp and Gordon, 2002). Along-strike averages of trench migration rates from Lallemand et al. (2005) indicate trench advance at ~ −23 mm/yr and retreat at ~37 mm/yr on the western and eastern Pacific sides, respectively. The fact that westward advance of the western Pacific trenches is accompanied by westward retreat of trenches in the eastern Pacific (at rates up to 40 mm/yr, Fig. 1a) suggests that the entire Pacific basin is migrating westward at ~30± 7 mm/yr, together with other known bimodal features of the Pacific (Table 1), (Jarrard, 1986; Doglioni et al., 1999; Lallemand et al., 2005). For example, west-subducting and east-subducting plates are moving at rates of 110 mm/yr and 50 mm/yr respectively (Gripp and Gordon, 2002), i.e 30 mm/yr faster and slower than the half-spreading rate of the intermediate East Pacific ridge. This indicates a ~30 mm/yr general westward drift of the entire Pacific system, including both overriding and subducting plates and the trenches between them. This pattern suggests that trench advance in the western Pacific and overall westward drift of the lithosphere are related issues.

Although global mantle flow models based on seismic tomography (Zhong, 2001; Becker, 2006) can reproduce the westward direction of the observed HS3 (Gripp and Gordon, 2002) net rotation, they generally do not reproduce the 48 mm/yr amplitude of this rotation. Instead, the maximum amplitude of the predicted net rotation matches the minimum net rotation estimated using Indo-Atlantic and Pacific hotspots (Gordon and Jurdy, 1986) and even remains lower than the amplitude estimated using the Indo-Pacific hot-spot tracks (Steinberger et al., 2004). Although the magnitude of the westward drift incorporated by the HS3 model is controversial and might be an end-member, we consider it to be more appropriate for our analysis of Pacific subduction zone kinematics because it is based on Pacific hot spots. In addition, note that the overall pattern of trench motion (advancing versus retreating) is a persistent feature because a large portion western Pacific trenches are advancing regardless of the reference frame (Funiciello et al., 2007).

The bimodal setting of the Pacific basin allows for a two-dimensional approximation; total trench lengths are comparable on both eastern and western sides (~12,000 km from Tierra del Fuego to northern British Columbia and from the south of New Zealand to Kamchatka, respectively, Table 1) and the subducting lithosphere

Table 1
Mean geometric/kinematic characteristics of W. Pacific and E. Pacific subduction zones (after Lallemand et al., 2005), in the hot spot reference frame (HS3-Nuvel1A Gripp and Gordon, 2002)

	Slab dip	Length	UP	Trench velocity	UP velocity	LP velocity
East	50° E	12000 km	Continental	37 mm/yr	42 mm/yr	30 mm/yr
West	70° W	11000 km	Oceanic	-23 mm/yr	-40 mm/yr	100 mm/yr

Positive velocities are trenchward. UP: upper plate, LP: lower plate. See details in text (neglecting the ~E–W trending Aleutian trench).

forms a N–S cylindrically folded sheet that wraps the Pacific upper mantle (Fig. 2). This volume is also bounded on its underside at 670 km by the high viscosity of the lower mantle, which is thought to be up to two orders of magnitude greater than that of the upper mantle (Lambeck and Chappell, 2001). The slabs that bound the Pacific upper mantle are slightly discontinuous: upper mantle "leaks", where the mantle can escape from the Pacific zone via toroidal flow, are mostly restricted to the ~6000 km wide southern boundary, which is orthogonal to the slabs, and the ~1700 km long transform boundary of North America, which together account for only ~20% of the total circum Pacific length (within the ~N–S oriented slabs bounding the eastern and western sides of the Pacific upper mantle, openings only amount to less than ~10% of the total trench length). Therefore the length and continuity of slabs not only moderate the rate of Pacific shrinkage but also prevent toroidal flow almost everywhere; hence if some slabs retreat, the trapped upper mantle must accompany the journey of the slabs. This in turn implies that any slab retreat on one side also has to be accompanied by slab advance on the other side. Thus, a westward-directed net force on the oceanic plates of the Pacific basin will cause the entire system of Pacific upper mantle, trenches, and plates to drift westward. This will tilt upper mantle slabs increasingly toward the east (Fig. 2).

3. Driving mechanism

What is the driving force that causes this westward drift of the Pacific? Forces that arise from upper plates may have a significant impact on subduction dynamics (Royden and Husson, 2006), and for the Pacific, significant differences between the eastern- and western-bounding upper plates are evident (Table 1). Most notable among these differences is the presence of a mountain range on the east (the Andes) and its absence on the west (Uyeda and Kanamori, 1979). Silver et al. (1998) demonstrated that the faster westward motion of South America triggered by the mid-Cenozoic closure of the Tethys sea (collision of Africa and India with Eurasia), is responsible for the Andean cordilleran orogeny. Here we consider the possibility that this forcing is also transferred through South America and on to the subducting plates of the Pacific, thus driving their westward motion. Thus, instead of regarding upper plate deformation as a by-product of subduction dynamics, we consider it as an external forcing on the Pacific system. In this respect, we develop a semi-analytical approach that avoids the technical difficulties and associated modeling uncertainties that are currently an inherent part of efforts to embed detailed subduction dynamics within numerical models of mantle flow (Conrad and Hager, 2001; Becker, 2006).

3.1. Force balance

Because it is challenging to model the dynamics of the upper plate, we alternatively use upper plate kinematics to infer its dynamics (England and Molnar, 1997). In this case, viscous stresses result from

Fig. 1. Kinematic and dynamic descriptions of Pacific tectonics. Black and green arrows indicate trench and plate velocities in the hot spot reference frame (Gripp and Gordon, 2002), respectively (white arrow is 50 mm/yr). Colours represent (a) topography and (b) strain rates (after Kreemer et al., 2000). Thick ticks show (a) the maximum buoyancy force per unit trench length in the upper plates (with ρ_c=2800 kg m^{-3} and ρ_m=3300 kg m^{-3}) and (b) the maximum viscous force per unit trench length in the upper plates. Thin ticks in (b) show compressive (red) and extensive (yellow) viscous forces (length is qualitative magnitude from Lallemand et al., 2005). Positive values denote trenchward forces.
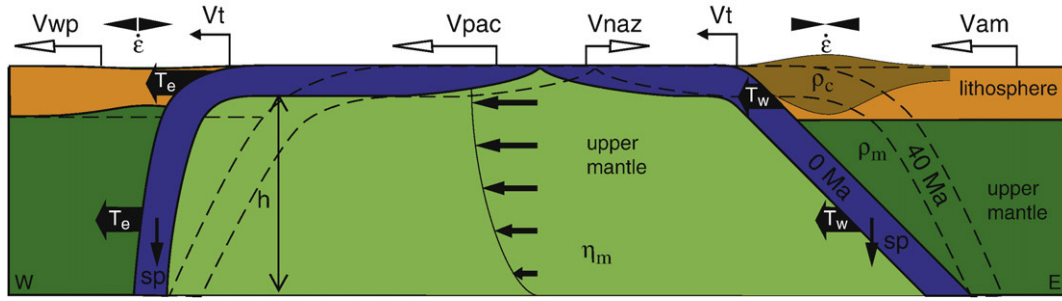
Fig. 2. Conceptual model for the Pacific upper mantle and plate circulation. Vt: trench velocity, Vnaz, Vpac, Vam, Vwp: Nazca, Pacific, American (North/South), West Pacific (Australia/Eurasia) velocities. $\dot{\varepsilon}$: viscous deformation; sp: slab pull. Thin black arrows indicate Poiseuille flow in the upper mantle.

the disequilibrium between boundary stresses and buoyancy stresses. In a continuous media, gradients of stress are balanced by the force of gravity per unit volume,

$$\partial \sigma_{ij}/\partial x_j = -\rho g_i, \tag{1}$$

where σ_{ij} is the ij^{th} component of the stress tensor ($\sigma_{ij} = \tau_{ij} - P\,\delta_{ij}$, where τ_{ij} is the deviatoric stress tensor and P is pressure), x_j is the jth coordinate direction, ρ is the density, g_i the ith component of the gravitational acceleration. In 2D, i and j only stand for the x (horizontal) and z (vertical) directions. The lithosphere can be conveniently approximated as a thin viscous sheet; this approximation has been extensively applied (England and McKenzie, 1982; Houseman and England, 1993) and here we only briefly develop the governing equations that are based on the vertical integration of the Navier–Stokes equation over the lithosphere thickness. Such models assume that the vertical variations of the horizontal velocity can be neglected within the lithosphere, and that deformation occurs at wavelengths larger than lithosphere thickness. The total vertical stress induced by a column of rock in the lithosphere is (neglecting 2nd order terms),

$$\sigma_{zz}(z) = g \int_0^z \rho(z')dz'. \tag{2}$$

Upper plate deformation is controlled by the distribution of basal shear stresses, either between the plates or at the asthenosphere/lithosphere boundary (Husson and Ricard, 2004). By vertically integrating the horizontal equilibrium over the lithosphere thickness L, we obtain,

$$\frac{\partial}{\partial x}\int_0^L \sigma_{xx}dz + \sigma_{xz} = 0, \tag{3}$$

which relates the horizontal gradient of the mean horizontal stress over the lithosphere to the basal shear stress σ_{xz}. Balancing horizontal and vertical stress terms via the pressure P, we write $\sigma_{xx} = \tau_{xx} - \tau_{zz} + \sigma_{zz}$. For a Newtonian fluid, $\tau_{ij} = \eta\left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i}\right)$, where η is the viscosity, and u_i and u_j are the components of the velocity vector. Plugging Eq. (2) into Eq. (3), the horizontal equilibrium becomes (again, neglecting 2nd order terms, Husson and Ricard, 2004)

$$4L\frac{\partial}{\partial x}\eta\frac{\partial u_x}{\partial x} + \sigma_{xz} = \frac{\partial M}{\partial x}, \tag{4}$$

where η is now the average viscosity of the lithosphere, and $M = \int_0^{+\infty} \delta\rho gzdz$ is the moment of the lithospheric mass anomalies (Artyushkov, 1973; Fleitout and Froidevaux, 1982). Restricting density variations in the lithosphere to crustal thickness variations (i.e. neglecting the density contrast between continental lithosphere and asthenosphere), we calculate the isostatic crustal thickness $S = h\rho_m/(\rho_m - \rho_c)$. Assuming isostasy, elevation h can therefore be used as a proxy for the moment because $M = \frac{1}{2}g\frac{\rho_c\rho_m}{\rho_m - \rho_c}h^2$, where ρ_c

and ρ_m are the crust and mantle densities. Eq. (4) can thus be conveniently written

$$\sigma_{xz} = \frac{1}{2}\frac{\rho_c\rho_m}{\rho_m - \rho_c}\frac{\partial h^2}{\partial x} - 4L\frac{\partial}{\partial x}\eta\frac{\partial u_x}{\partial x}, \tag{5}$$

where σ_{xz} is the traction at the base of the plate, and the first and second terms on the right hand side of the equation are the buoyancy and viscous components of stresses, respectively. When applied to plate margins, positive values of σ_{xz} indicate compression due to interplate traction and negative values quantify the resistance of the asthenosphere to the displacement of the lithosphere (Husson and Ricard, 2004).

3.2. Net westward force

The total force T transmitted from one plate to the other is the sum of buoyancy and viscous forces. It can be approximated by the horizontal integration of the shear stresses between the two plates given by Eq. (5), i.e. by the sum of the moment and the deviatoric stresses integrated over the lithosphere thickness. Along the eastern margin of the Pacific, buoyancy forces that build up as a result of compression (primarily the Andean mountain belt) are large and locally reach more than 10^{13} N m^{-1} (Fig. 1a), with an average of 4.8×10^{12} N m^{-1}. To obtain these estimates, we Fourier-filtered the topography in order to discard wavelengths smaller than 75 km (and thus the extreme peaks) and used the normal distance between the trench and the location of the maximum moment (i.e. the location of the maximum elevation) to approximate the integration distance Δ that corresponds to interplate traction (which in essence is given by the sum of positive values of σ_{xz}). From Eq. (5),

$$T = \int_\Delta \sigma_{xz}dx. \tag{6}$$

On the western margin of the Pacific, the buoyancy forces are negligible (beyond our approximation) and we discarded them. Viscous forces can be evaluated from deviatoric stresses inferred from strain rates in the upper plates (deduced from Lallemand et al.,

Table 2
Mean dynamic characteristics of W. Pacific and E. Pacific subduction zones, in the hot spot reference frame (HS3-Nuvel1A, Gripp and Gordon, 2002)

	Mean elevation, UP	Buoyancy force	Strain rate	Viscous force	Total force
East	2270 m	4.8×10^{12} N m^{-1}	-2×10^{-15} s^{-1}	1.2×10^{12} N m^{-1}	$T_e = 6.0\times10^{12}$ N m^{-1}
West	/	/	12×10^{-15} s^{-1}	-3.7×10^{12} N m^{-1}	$T_w = -3.7\times10^{12}$ N m^{-1}

Positive forces denote compression. UP: upper plate. See details in text.

2005). Of course they are highly dependent on the viscosity of the lithosphere which has an uncertainty of an order of magnitude. Choosing a low effective viscosity of the lithosphere of $3 \times 10^{21} \pm 2$ 10^{21} Pa s (England, 1986; Husson and Ricard, 2004), a characteristic width of deformation of 100 km and plate thicknesses of 100 km and 50 km for the continental and oceanic lithospheres, respectively (10 km for active back-arc systems), we find minimum mean viscous forces of 1.2×10^{12} N m^{-1} $\pm 2.5 \times 10^{12}$ N m^{-1} and -3.7×10^{12} N m^{-1} $\pm$ 2.5×10^{12} N m^{-1} along the eastern and western margins, respectively (Fig. 1b and Table 2). Along the eastern margin, they are significantly lower than buoyancy forces, indicating that buoyancy -or body-forces balance boundary -or tectonic-forces, and that tectonic equilibrium is almost achieved in mountain belts (Husson and Ricard, 2004). The mean total force transmitted from the upper plates to the lower ones are thus $T_e = 6.0 \times 10^{12}$ N m^{-1} and $T_w = -3.7 \times 10^{12}$ N m^{-1}, along the eastern and western margins, respectively.

3.3. Upper mantle shear

This result suggests that upper plates are pushing the lower plates westward along the eastern margin and pulling them westward along the western margin. A comparison between the kinematics (Table 1) and dynamics (Table 2) of the eastern and western margins of the Pacific makes this long recognized dichotomy (Uyeda and Kanamori, 1979; Jarrard, 1986) even more striking. In fact, when the Pacific domain is considered in 2D, it seems that the only difference between eastern and western sides of the Pacific arises from this additional forcing from upper plates. Slab pull is generally considered as the main driving force, and as such is presumed responsible for trench retreat. But because the Pacific upper mantle is bounded almost everywhere – laterally by slabs, beneath it by a lower mantle where the viscosity increases by two orders of magnitude – it essentially has to remain within the domain of the Pacific basin. There is no obvious contrast between eastern and western slab pull forces, which drive the relative motions of the Pacific and Nazca plates but are mutually annihilated in a net sense. This leaves the Pacific domain in a *status quo* of antagonistic forces that, by itself, precludes systematic trench retreat. In fact, Lallemand (1998) already suggested that "sub-Pacific mantle resistance to contraction may induce a delay in retrograde slab migration".

The forces that are transmitted from the upper plates are directed westward on both sides of the Pacific, i.e toward the subducting plate along the eastern margin, and toward the overriding plate along the western side. The sum of the westward components of the forces, $|T_e| + |T_w| \simeq 10 \times 10^{12}$ N m^{-1} (Table 2) is comparable in magnitude to the main plate-driving forces such as slab pull (Conrad and Lithgow-Bertelloni, 2002). This total traction is distributed over the Pacific width, shears the upper mantle, and induces westward drift of the entire Pacific system, including oceanic lithospheres, upper mantle, slabs, and trenches. The validity of this hypothesis is tested in Appendix A, in which the Pacific system is simplified into two vertical slabs of width W, facing each other and separated by a distance W. When one slab moves steadily toward the other, the Stokes flow between these two slabs shows that the open ends between the slabs do not permit enough mantle leakage to significantly diminish the transmission of compressive stresses between the converging slabs. This implies that only mantle that is close to Tonga and the San Andreas can escape the Pacific system, and that the presence of these openings does not significantly diminish the stress transmission required to shear the Pacific upper mantle.

The velocity at which the system drifts is controlled by the viscosity of the upper mantle, which deforms in response to surface plate motions. To estimate the viscosity, η_{um}, that produces the observed surface drift, we consider 2D Poiseuille flow of the upper mantle with a free surface slip, where slabs transmit surface stresses to the upper mantle because they are stiffer than the
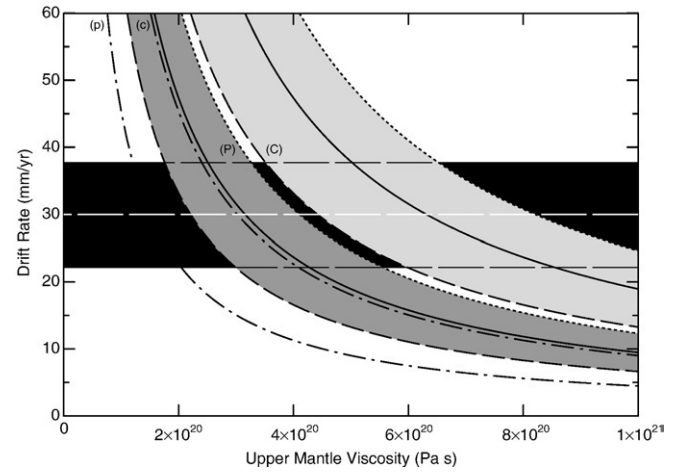
Fig. 3. Upper mantle flow in the Pacific described by Poiseuille (P, dark shaded) and Couette (C, light shaded) flows for a total force of 7×10^{12} N m^{-1} (dashed lines), 10×10^{12} N m^{-1} (solid lines) and 13×10^{12} N m^{-1} (dotted lines), and 4.8×10^{12} N m^{-1} (dotted-dashed line, (d) and (c), only considering buoyancy forces), $W = 10{,}000$ km, $h = 600$ km (removing a 70 km thick lithosphere thickness). Dotted line and grey box bounds the westward drifting rate of the Pacific domain (30 ± 7 mm/yr).

asthenosphere and anchored in the lower mantle. The velocity at the surface is:

$$u_0 = -\frac{h}{2\overline{\eta}_{um}} \frac{|T_e| + |T_w|}{W}, \qquad (7)$$

where W is the average width of the Pacific ocean and h is the thickness of the upper mantle. For the estimated total force, a surface velocity $u_0 = 30 \pm 7$ mm/yr requires a mean upper mantle viscosity $\overline{\eta}_{um} = 3.1 \times 10^{20} \pm 10^{20}$ Pa s (Fig. 3). The fact that this value is consistent with independent estimates (Lambeck and Chappell, 2001; Steinberger et al., 2004; Mitrovica and Forte, 2004) supports our model. If we neglect viscous forces in the overriding lithospheres, the required viscosity is smaller ($\overline{\eta}_{um} = 1.5 \, 10^{20} \pm 10^{20}$ Pa s). If slabs are not strong enough to transmit surface stresses down to the upper mantle, but the stresses are exerted directly on the subducting plates, then the drift of those plates will instead induce laminar (Couette) shear flow in the upper mantle. In this case, the force balance on the lithosphere yields a drift rate that is twice that of Eq. (7), which implies estimates for the upper mantle viscosity that are twice as large as those for Poiseuille flow for the same driving force and drift rate (Fig. 3), but still in the range of independent estimates (Lambeck and Chappell, 2001; Steinberger et al., 2004; Mitrovica and Forte, 2004). It may be argued that the convecting, heterogeneous mantle may prevent stresses from being transmitted across the system. In particular, in the strict case of Couette shear flow, the stresses on the system must be transmitted across the East Pacific ridge, which could be thought problematic because of the supposed weakness of ridge systems. However, there is no reason for ridges to be stress free. Because stresses are continuous and add up linearly, only deformation may vary rapidly as a result of spatial variations in viscosity, not stresses. Moreover, in a continuous media stresses can be transmitted laterally beneath the ridge via the upper mantle. The divergence that is observed at ridges only mirrors stresses arising from the underlying convecting mantle. Far field compressive stresses associated with the shrinking of the Pacific are exerted on top of the diverging convective stresses; it would be an extraordinary coincidence if these two sets of stresses cancel. We conclude that the drift rate of the Pacific is consistent with both the tectonic forces that drive drift and the viscous deformation that resists it.

For Poiseuille flow, the mean upper mantle velocity is $\overline{u}_{um} = \frac{2}{3} u_0$, which implies that plates and trenches drift at faster rates ($\overline{u}_0 \simeq 30$ mm/yr) than the upper mantle on average ($\overline{u} \simeq 20$ mm/yr).

Also notice that that lower estimates for the Pacific drift rate derived in different reference frames (Gordon and Jurdy, 1986; Steinberger et al., 2004) imply higher viscosities. Because deeper layers move more slowly, western slabs dip more steeply than eastern ones (Fig. 2), as noted by observations (Jarrard, 1986; Doglioni et al., 1999; Lallemand et al., 2005). A regional expression of the general westward drift of the upper mantle is (paradoxically) the rapid eastward retreat of the Tonga trench associated with the opening of the Lau basin (Fig. 1): the discontinuous northern edge of the Tonga slab is an uncommon setting in which toroidal flow around the edge that permits slab retreat. Significant unbounded edges are absent elsewhere in the western Pacific; as a result these trenches only advance. The presence of such "leaks" also explains the faster westward motion of eastern trenches than that of western ones: part of the upper mantle volume that is trapped under the Pacific manages to flow outside of the Pacific domain, which is shrinking in response to the growth of the Atlantic and Indian spreading systems.

4. Discussion

Implications are twofold. First, trench advance can be explained by an additional forcing from upper plates: heterogeneities in the vigour of subduction roll-back, boosted or not by stresses arising from upper plates at places that depend on the given distribution of tectonic plates, impose an overall shear of the upper mantle that causes trench advance at other locations. This mechanism is thus alternative to models where trench motion is driven purely by slab parameters (Bellahsen et al., 2005; Faccenna et al., 2007; Capitanio et al., 2007). In this regard, the change in trench migration behavior during Mid-Miocene (e.g. Sdrolias and Muller, 2006) toward more advancing and retreating regimes for western and eastern slabs, respectively, could be correlated to the onset of Andean building, which can be regarded as the triggering mechanism. It follows that subduction kinematics – in the Pacific in particular – can be considered as dependent on the dynamics of the broader Pacific system (the Panthalassan framework of Collins (2003), and not solely a result of the local subduction properties.

Second, the long-recognized net rotation of the lithosphere in general (Bostrom, 1971; Uyeda and Kanamori, 1979; Doglioni, 1990; Ricard et al., 1991), and of the Pacific in particular, is governed by the motion of the upper plates, in the sense that they form boundary layers that transmit stresses from one convecting unit to another. The current force exerted by the South American plate plays a major role; it accounts for about half of the total force that shears the Pacific upper mantle. This effect may be somewhat surprising, but it is in fact not unexpected since such large transmission of stresses across the interplate contact has been evidenced by other means: Andean growth may be responsible for the decrease in convergence rates between Nazca and South American plates (Iaffaldano et al., 2006). In Tibet, the very large gravitational potential energy compresses and buckles the Indian lithosphere (Chamot-Rooke et al., 1993; Martinod and Molnar, 1995): the long-wavelength undulations of the Indian plate are viscous folds resulting from the increased compression between the Indian plate and Tibet.

This situation connects the Pacific system to the current understanding of the Atlantic system and global plate circuit (Fig. 4): the mid-Cenozoic closure of the Tethys and subsequent aggregation of Africa to Eurasia led to the formation of a massive continent,
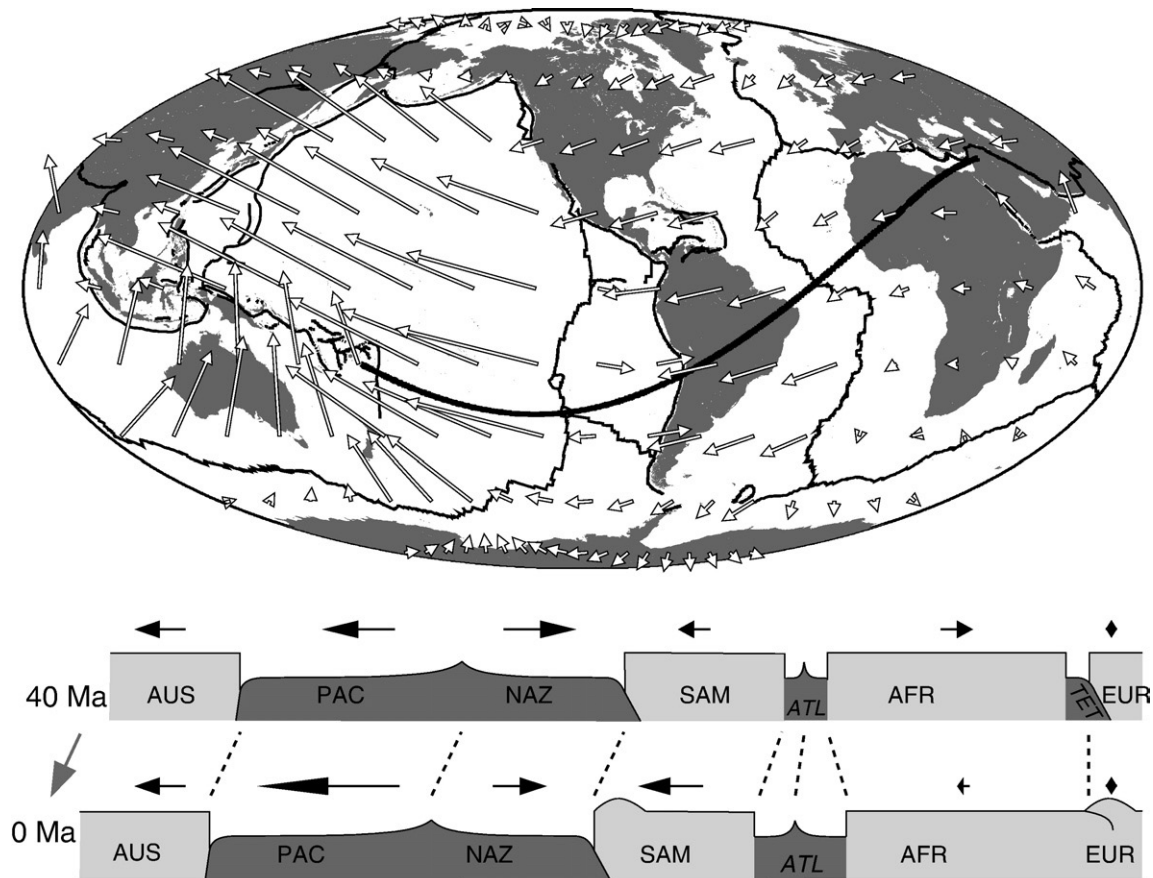
Fig. 4. Top) Map view of the present day plate motion in the hot spot reference frame (Gripp and Gordon, 2002; bottom) section along a great circle crosscutting the Pacific/Atlantic plate circuit (location on map) before (40 Ma) and after (0 Ma) the closure of the Tethys. Note that spreading of both Atlantic and Pacific oceans changed from symmetric to asymmetric. Dark (light) shaded areas are oceanic (continental) units. AUS: Australia; PAC: Pacific; NAZ: Nazca; SAM: South America; AFR: Africa; EUR: Eurasia; ATL: Atlantic; TET: Tethys. Black arrows indicate plate velocities in the hotspot reference frame. Dashed lines indicate the moving location (in the hotspot reference frame) of characteristic locations.

profoundly anchored into the mantle by its highly viscous lithospheric keel. The surrounding plates are thinner and easily move apart from it. This abrupt change disrupted the global plate circuit and in particular triggered a westward drift of the Atlantic ridge and a faster westward motion of the South American plate, subsequently increasing stresses along the South American margin sufficient to develop the Andean belt (Silver et al., 1998; Sobolev and Babeyko, 2005). Those large stresses are in turn transmitted to the subducting plates and the Pacific upper mantle. In the theoretical situation where the closure of the Tethys and subsequent western expansion of the Atlantic started abruptly, the total force was transmitted to the Pacific system instantaneously: it was transmitted via viscous stresses in the upper plate before the Andes formed and via buoyancy stresses associated with the presence of a mature Andean orogenic belt afterwards (Eq. (5)). The net transmitted force T would then be constant through time. This implies that the net rotation of the Pacific and Andean growth are coeval, unless interplate coupling evolved through time (Iaffaldano et al., 2006).

A small amount of upper mantle material leaks out of the Pacific domain, which explains why western subduction zones are moving westward more slowly than eastern ones by ~14 mm/yr, which causes the area of the Pacific basin to shrink by ~0.1% per Myr, while the Atlantic basin expands. We estimate that ~0.1×10^9 km^3/Myr of upper mantle material flows from the Pacific reservoir into the Atlantic reservoir (from the Panthalassan domain to the Pangean domain, (Collins, 2003; Davaille et al., 2005) through the 1700 km long North American transform margin, which can be considered the only significant outlet. This volume transfer is independent of the reference frame. Thus, even if the HS3 reference frame (Gripp and Gordon, 2002) overestimates the magnitude of the westward drift of trenches, other reference frames (see Becker, 2006 for a review) tend to accommodate Pacific shrinkage by holding western trenches steady while eastern trenches retreat. The cross-Pacific dichotomy is therefore persistent and we conclude in any case that the mid-Cenozoic plate circuit reorganization at least led to the onset of Pacific shrinking with the same mechanism of stress transmission.

Therefore, eastern Pacific mountain belts – the Andes in particular – transmit stresses from the Atlantic domain to the Pacific domain; note that these stresses will last as long as the spreading of the Atlantic prevails at such rate. Eventually, the entire Pacific upper mantle system is sheared, as multiple geometric and kinematic features suggest (Table 1). Similarly, the Laramide and Sevier orogenies, which formed during the Cretaceous when the Andean belt was rather small, may also have caused westward drift, this time imposed by the North American plate; the present-day location of the Farallon slab beneath eastern North America (Bunge and Grand, 2000) seems to testify for this drift. In spite of its early timing, the Hawaii–Emperor seamount bend at 50–42 Ma (Sharp and Clague, 2006) could potentially be interpreted as a record of the onset of this mechanism.

The westward drift of the lithosphere is thus the consequence of a particular setting of tectonic plates on their global course and is not an intrinsic property of the Earth, as has already been suggested by others (Ricard et al., 1991; Ranalli, 2000). In fact, net rotation, which requires degree one toroidal flow, may only exist if lateral viscosity variations exist in the upper mantle (Hager and O'Connell, 1979; Ricard et al., 1991). In this particular case, the African and Eurasian lithospheric keels form a very thick and stiff unit that is anchored into the mantle, allowing for a westward expansion of the Atlantic ocean and subsequently a net westward motion of the lithosphere. This result emphasizes that while plate motions are ultimately the result of forces that arise from mantle flow (Zhong, 2001; Becker, 2006), the dynamical deformation of these plates both at the surface (e.g., Andean tectonics) and at depth (e.g., the slabs as barriers to flow) modifies the interaction of mantle flow and plates in ways that cannot yet be predicted by numerical models of mantle convection.

Acknowledgments

Appendix A. A Stokes flow between 2 slabs

In order to validate the approximation to 2 dimensions, we evaluate how stresses can be transmitted between the slabs. We solve the biharmonic stream function Ψ for incompressible Stokes flow in a square domain of width $5W$ (Fig. A.1). A vertical slab of width W retreats at a prescribed velocity toward a steady parallel slab of width W. The biharmonic function is:

$$\nabla^4 \psi = 0, \tag{A.1}$$

with $u = -\frac{\partial \psi}{\partial y}$ and $v = -\frac{\partial \psi}{\partial x}$, u and v being the velocities along the x and y cartesian directions. Compressive stresses (positive) induced by the retreat of the eastern slab are transmitted across the entire width to the western slab, even close to the edges of the slabs (note the compressive singularities at the edges of the rigid slabs). This result suggests efficient stress transmission in the Pacific domain, where the situation is even more favorable for stress transmission because the width of the openings at the N and S ends of the domain are much narrower than at center. The uniformity of this stress transmission verifies that our 2D approximation is valid for this problem.
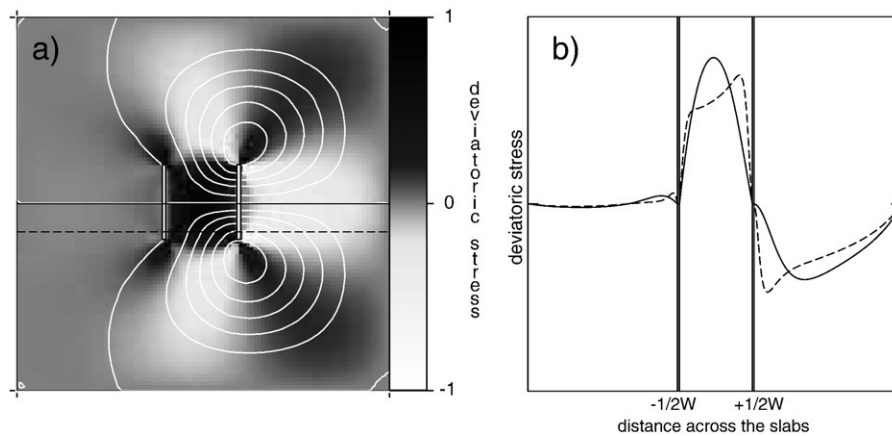

Fig. A.1 a) Map view of deviatoric stress for the Stokes flow between two vertical slabs of the W separated by a distance W; deviatoric stress (color coded) and stream lines (white curves). b) Profile of the deviatoric stresses on a section across the slabs, at the center (solid curve) and at the edges (dashed).

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