Along-strike variation in subducting plate velocity induced by along-strike variation in overriding plate structure: insights from 3D numerical models

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Abstract

Subduction dynamics can be understood as the result of the balance between driving and resisting forces. Previous work has traditionally regarded gravitational slab pull and viscous mantle drag as the main driving and resistive forces for plate motion respectively. However, this paradigm fails to explain many of the observations in subduction zones. For example, subducting plate velocity varies significantly along-strike in many subduction zones and this variation is not correlated to the age of subducting lithosphere. Here we present three-dimensional and time-dependent numerical models of subduction. We show that along-strike variations of the overriding plate thermal structure can lead to along-strike variations in subducting plate velocity. In turn, velocity variations lead to significant migration of the Euler pole over time. Our results show that the subducting plate is slower beneath the colder portion of the overriding plate due to two related mechanisms. First, the mantle wedge beneath the colder portion of the overriding plate is more viscous, which increases mantle drag. Second, where the mantle wedge is more viscous, hydrodynamic suction increases, leading to a lower slab dip. Both factors contribute to decreasing subducting plate velocity in the region; therefore, if the overriding plate is not uniform, the resulting velocity varies significantly along-strike, which causes the Euler pole to migrate closer to the subducting plate. We present a new mechanism to explain observations of subducting plate velocity in the Cocos and Nazca plates. These results shed new light on the balance of forces that control subduction dynamics and prove that future studies should take into consideration the three-dimensional structure of the overriding plate.

Highlights

Observed correlation between thick overriding plate and slower subduction

We present three-dimensional thermo-mechanical models of subduction

Non-uniform overriding plate leads to along-strike variations of subducting plate velocity

The Euler pole migrates with time if the overriding plate is non-uniform

1. Introduction

Subduction processes can often be understood in terms of the balance between driving and resisting forces [Billen, 2008]. Many studies have approached the study of subduction dynamics through statistical analyses of observations and numerical and analog models, but some aspects are still poorly understood. Early statistical studies showed that the main force driving subduction is gravitational pull and the main force resisting subduction is the viscous drag of the mantle [e.g., Forsyth and Uyeda, 1974; Vlaar and Wortel, 1976], possibly modulated by slab bending forces [Buffett and Rowley, 2006; Conrad and Hager, 1999]. If gravitational pull and viscous resistance were the only forces controlling subduction processes, a good correlation between subducting plate age and subducting plate velocity would be expected. However, such a correlation does not exist or is very weak [Di Giuseppe et al., 2009; Lallemand et al., 2005]. Similarly, no correlation is found between subducting plate age and slab dip [Cruciani et al., 2005]. This lack of correlation indicates that there are other factors controlling subduction dynamics.

1.1. First-order Expectations. The motion of a rigid plate on a sphere can be represented by a rotation about an axis through the centre of the sphere and intersecting its surface at the so-called Euler Pole. This kinematic model implies that plate motion is tangential to a small circle centred at the Euler pole and the plate velocity varies progressively with distance to the pole. When the Euler pole is located at 90° from the centre of the subducting plate, the plate velocity is maximum

and symmetrical to the middle of the subducting plate, with minimal variability. If the position of the pole is closer to the plate, the portion of the plate closest to the pole rotates at a slower linear velocity relative to the portion of the plate farther away. Therefore, the location of the Euler pole is a clear way of quantifying the variation in subducting plate velocity along a trench. Due to the many factors affecting subducting plates, the Euler pole for real tectonic plates is rarely located at 90° from the centre of the plate [DeMets *et al.*, 2010]. These factors include, but are not limited to, the irregular geometry of plates or the presence of multiple subduction zones on different sides of a plate. The location of the Euler pole of a given subducting plate (with respect to the overriding plate) can also vary significantly in time [e.g., Pardo-Casas and Molnar, 1987; Somoza, 1998; Somoza and Ghidella, 2005; 2012], indicating that the spatial balance between processes driving and resisting the plate along a subduction zone is also time-dependent. Neither why the Euler pole is rarely located at 90° nor why it migrates with time has been fully explained. We suggest that subducting plate velocity can be affected by diverse factors, such as lateral variations in the overriding plate thermal state, which will cause the Euler pole to migrate closer toward the end of the subduction with colder overriding plate.

1.2. Factors Controlling Subducting Plate Velocity. Previous two-dimensional numerical models have investigated many of the factors proposed to affect subducting plate velocity. By comparing observations and numerical models, previous studies demonstrated the importance of slab pull as a driving force [Conrad and Lithgow-Bertelloni, 2002; Forsyth and Uyeda, 1974]. These studies showed that plate motions were best predicted if slabs were included in numerical models, and that gravitational pull accounts for at least half of the total driving force. Other studies have investigated the importance of viscous dissipation and have shown that a change in mantle viscosity by an order of magnitude, changes subduction velocity by a factor of four [Conrad and

69 *Hager, 2001*] to ten [*Capitanio, 2013*]. Analytical studies also show that approximately 75-80%
70 of the energy from slab pull is dissipated through viscous resistance in the ambient mantle
71 [*Capitanio et al., 2007; Stegman et al., 2006*]. Similarly, an increase in viscous resistance at the
72 upper-lower mantle boundary (ULMB) causes subduction velocity to decrease dramatically
73 [*Lithgow-Bertelloni and Richards, 1998; Quinteros and Sobolev, 2013; Čížková et al., 2012*].
74 Thus, the balance between slab pull and viscous resistance plays a primary role in controlling
75 subducting plate velocity.

76 Several studies also demonstrate the importance of energy dissipation within the subducting
77 plate due to internal deformation or bending [*Buffett and Rowley, 2006; Capitanio et al., 2007;*
78 *Conrad and Hager, 1999; Di Giuseppe et al., 2009; Doglioni, 2009; Irvine and Schellart, 2012;*
79 *Stegman et al., 2006*]. Previous two-dimensional numerical models of a bending plate showed that
80 around 60% of the energy is dissipated during plate bending [*Conrad and Hager, 1999*]. Similarly,
81 location of the Euler pole for different subducting plates was used to determine that at least 40%
82 of the slab pull energy is dissipated during plate bending [*Buffett and Rowley, 2006*]. However,
83 other studies show that dissipation during plate bending might be less important, around 2-20%,
84 when the bending radius can evolve freely [*Capitanio et al., 2007; Doglioni, 2009; Irvine and*
85 *Schellart, 2012; Stegman et al., 2006*]. Numerical models have also shown that even though old
86 slabs subduct faster, strong slabs subduct slower because of the increased energy dissipation within
87 the slab [*Di Giuseppe et al., 2009*]. These studies highlight the importance of the subducting plate
88 strength, and show how stronger slabs, which are more difficult to bend at the plate boundary,
89 subduct at lower velocities than predicted based on their negative buoyancy. This result partially
90 explains the lack of correlation between plate age and subducting plate velocity. However, other
91 studies show opposing results [*Capitanio et al., 2007; Doglioni, 2009; Irvine and Schellart, 2012*].

These studies show that during an initial transient stage, dissipation due to plate bending increases with increasing slab viscosity or thickness. However, once subduction reaches steady-state, the trend is inverted and less energy is dissipated to bend more viscous or thicker slabs. This effect appears because these models are allowed to evolve freely and the more rigid slabs accommodate the deformation with a larger bending radius, which requires less energy.

Interplate coupling has also been studied as a candidate to control subducting plate velocity. Numerical models have shown that increasing the frictional coefficient at the plate boundary significantly decreases subduction velocity [Di Giuseppe *et al.*, 2008; van Dinther *et al.*, 2010]. Analogue models have validated this result, and show that high interplate coupling decreases subduction velocity [Duarte *et al.*, 2013].

1.3. Along-strike Variations in Subducting Plate Velocity. While previous studies identify many of the factors that can affect subducting plate velocity and the magnitude of expected variations, they do not explicitly consider why subducting plate velocity often varies systematically along an individual subduction zone. One possibility is that the rheology intrinsic to the subduction zone varies systematically along-strike. For example, the viscosity of the mantle wedge could vary along-strike if there were variations in fluid input through sediment subduction, or mantle wedge temperature (e.g., due to cooling by an older plate). However, mantle viscosity is only weakly dependent on water content and melt fraction [Hirth, 2003]. Conversely, significant variations in mantle wedge viscosity are expected if there is a significant difference in the slab sinking rate due to the non-linear dependence on strain-rate (or stress). If this were a primary control of along-strike variation in subducting plate velocity, then the slab thermal parameter, which determines the available negative buoyancy of the slab, should be correlated with subducting plate velocity. However, this correlation is not observed [Lallemand *et al.*, 2005].

Finally, a systematic variation in intrinsic plate boundary viscosity, due to for example, differences in sediment input, could result in large variation in plate velocity. However, large variations in plate velocity along individual subduction zones occur even where there is uniform sediment input (e.g., compare plate velocities from Lallemand et al., [2005] and sediment thickness from Heuret et al., [2012]). Therefore, while along-strike variations in rheology likely do exist along individual subduction zones, these do not appear to be the primary cause of along-strike variations in subducting plate velocity.

While variations in intrinsic plate boundary and mantle wedge rheology may not explain observed variations in plate velocity, other extrinsic processes that could modify plate coupling have been proposed. For example, the excess weight of an orogen can increase the frictional stress on the plate boundary locally, which can cause the subduction velocity to decrease [*Iaffaldano et al.*, 2006; *Meade and Conrad*, 2008]. Similarly, the existence of flat slab segments below South America in the past would have increased coupling (higher shear stress) between the slab and the overriding plate [*Martinod et al.*, 2010]. This increased coupling could potentially decrease subducting plate velocity, explaining the past and present history of the subducting velocity of the Nazca Plate.

Previous two-dimensional numerical models have also shown that subduction is slower for older overriding plates [*Garel et al.*, 2014]. The reduction of subduction velocity is interpreted as consequence of slower ocean-ward trench migration rates. However, we also observe that the structure of the overriding plate is correlated with the along-strike subducting plate velocity distribution in some subduction zones where the trench is hardly migrating. Specifically, in Middle America and South America, a lower subducting plate velocity is observed below regions with a thicker, colder overriding plate (Fig. 1).

1.4. Observations.

Because multiple factors can affect the location of the Euler pole for a given plate, only a few plates with simple geometry are suitable for examining the possible role of the overriding plate. Ideally, this would be a plate with a simple geometry, a single subduction zone on one side of the plate, and a single ridge on the opposite side. The four plates that are closest to meeting this ideal scenario are the Cocos, Nazca, Juan de Fuca and Philippine plates. However, part of the Philippine plate is involved in a collision in Taiwan and motion at its northern end may be significantly affected by the triple junction with the Japan and Izu-Bonin trenches.

The subducting plate velocity of the Cocos plate along the Middle America Trench varies from 4.6 cm/yr in the north to 9.2 cm/yr in the south (Fig. 1; Euler pole located at 26.91 N, 124.31 W). Similarly, the subducting plate velocity of the Nazca plate varies from 5.3 cm/yr in the north to 7.7 cm/yr in the south (Euler pole located at 46.23 N, 101.06 W). A similar along-strike velocity distribution is observed for the Juan de Fuca Plate. Note that in order to make a consistent comparison with the results of the models we show the present day observations of subducting plate velocity from MORVEL NNR, using a no-net rotation reference frame [DeMets *et al.*, 2010; Fig. 1]. This point is explained further in section 2.

Plate age varies along strike in both Middle and South America. However, age is not correlated with the subducting plate velocity: the subducting lithosphere is oldest at the centre of the plate boundary for both oceanic plates [~ 25 for Cocos and ~ 50 Myr for Nazca; Müller *et al.*, 1997] but velocity is maximum at one end of each subduction zone.

However, a correlation between plate velocity and thickness of the overriding plate is observed in both Middle and South America (Fig. 1). In both regions a thick overriding plate is located above the region where a slower subducting plate velocity is found. In the case of the Philippine Sea plate, faster subduction occurs in the south, beneath younger oceanic lithosphere

[Müller *et al.*, 1997]. The only counter example of this relationship is the Juan de Fuca plate, where the subducting plate velocity is slightly smaller in the south, even though the overriding plate is thicker in the north [Yuan and Romanowicz, 2010]

Here we present results from numerical models that offer a new mechanism to explain along-strike variability of subducting plate velocity. In this study we show that non-uniform overriding plate thickness can cause along-strike variations in subducting plate velocity through two related mechanisms. First, a non-uniform overriding plate can deform the slab, leading to lower dips beneath regions of colder overriding plate [Manea *et al.*, 2012; Rodríguez-González *et al.*, 2012; Rodríguez-González *et al.*, 2014; Taramón *et al.*, 2015]. These lower slab dips are less able to transmit the slab pull to the subducting plate, leading to lower subducting plate velocity. Second, the mantle wedge beneath regions of cold overriding plate will be colder and therefore more viscous; the higher viscosity will increase the resistance exerted on the slab thus further decreasing subducting plate velocity in the region, especially during the first stage of subduction.

2. Methods

We compute the solid-state flow of the lithosphere and sub-lithospheric mantle using the finite-element code CitcomS [Gurnis *et al.*, 2000; Tan *et al.*, 2006], in which the time-dependent equations of conservation of mass, momentum and energy are solved for a three-dimensional incompressible fluid. We adopt the Boussinesq approximation, which neglects shear and adiabatic heating and ignores density differences except for the gravitational force. All the boundaries have a free-slip condition, the top boundary temperature is set at 0°C, the bottom boundary is set at 1400°C, and the lateral boundaries are thermally insulated. The modelled domain represents a

three-dimensional tesseroïd that spans 5000 by 3900 by 1500 km in latitude, longitude and depth, respectively (Fig. 2). The model domain is discretized as a finite element grid with 480 by 488 by 160 nodes (37,478,400 elements total) with finer mesh resolution near the plate boundaries and slab. Previous studies have shown that including surrounding plates can significantly affect subduction dynamics [Yamato *et al.*, 2009]. Therefore, the models include not only the subducting and overriding plates, but also two lateral plates, one on each side of the subducting plate. The dimensions of the plates are chosen to reduce boundary effects and the subducting plate is wide enough that most of the slab is not affected by the return flow around the edges. More information regarding the solved equations, model setup, initial temperature distribution and rheology can be found in the supplementary material and in Rodríguez-González *et al.* [2014]. Note that the models implemented for this work are not intended to mimic any of the subduction zones discussed above in particular, even though the observations in all four subduction zones motivate this study.

In the models presented here, subduction is dynamically driven, that is, the models start with a partially subducted proto-slab that is negatively buoyant and provides a pull force that drives subduction. The proto-slab (Fig. S1) has a uniform dip of $\sim 75^\circ$ measured along the slab surface from 120 to 220 km depth. The subducting plate is surrounded by an imposed weak zone, which effectively decouples it from the lateral and overriding plates and from the west boundary. The age of the subducting plate varies initially from 0 Myr at the western boundary to 50 Myr at the trench. In the reference model (Model R), a uniform age of 50 Myr is used for the overriding and lateral plates. In models with a non-uniform overriding plate (Models A and B) we replace the northern portion of the overriding plate with colder lithosphere (150 Myr).

We use a temperature, pressure and strain-rate-dependent composite viscosity (dislocation and diffusion creep mechanisms) for the upper mantle, whereas we assume a linear viscosity

(diffusion) for the lower mantle [*Jadamec and Billen, 2010*]. We also impose a viscosity increase of an order of magnitude between the upper and lower mantle, in agreement with constraints from the geoid [*Hager, 1984*]. The parameters used for the viscosity law [*Hirth, 2003*] are listed in Supplementary Table 2. In order to account for plastic deformation occurring at low temperatures and pressures, the effective viscosity is limited by a depth-dependent yield stress.

Finally, we note that because of the fixity of the trench and overriding plate, our model can only consider the effect of the overriding plate structure on the motion of the subducting plate, not on the dynamics of the upper plate. For this reason we use the observed present-day plate velocities expressed in the no-net-rotation reference frame to compare with our model results. This limitation makes it more difficult to compare the time-dependence of the Euler poles because the studies of past plate motions tend to focus on the relative velocity of the subducting and overriding plate. Therefore, we will not directly evaluate our model against past pole locations but only against the variability in relative pole locations and magnitude of overall plate velocity.

3. Results

Here we present results from three different models. Model R is the reference model in which the overriding plate is uniform along strike. Model A has a non-uniform overriding plate with a colder portion extending 600 km to the north. Model B (setup represented in Fig. 2) also has a non-uniform overriding plate, but the cold portion is 1400 km wide (half of the trench length).

In order to study the evolution of the three models we compute different quantities. First, the shallow slab dip is determined by linearly fitting the 1000°C isotherm between 120 and 220 km. Second, we compute the subducting plate velocity, v_s , as the magnitude of the velocity in the nodes

of the top surface of the subducting plate. The average subducting plate velocity $\bar{v}_s$ is the average of the velocity magnitude on the top surface of the entire subducting plate. In addition, we also compute the minimum and maximum subducting plate velocities at the trench ($v_s^{\min}$ and $v_s^{\max}$ respectively). Finally, we compute the location of the Euler pole that best fits the surface velocity of the subducting plate using a modified version of the code Euler Pole Calculator [Goudarzi *et al.*, 2014]. We compute the Euler pole latitude λ_e and longitude φ_e and the angular velocity ω_e .

During the evolution of all the three models, three clear stages can be distinguished: 1) from the beginning of the simulation until the slab reaches the ULMB (around 4-5 Myr), 2) while the slab is crossing the ULMB (until around 6-7 Myr), and 3) once the slab has entirely penetrated into the lower mantle. For additional information regarding the evolution of the slab geometry, see videos of time evolutions of models in the supplemental material.

3.1. Reference model. Model R (Fig. 3a and Videos 1 and 4) is designed for comparison purposes and has a uniform overriding plate of 50 Myr (the depth of the 1000°C isotherm is $Z_{1000} = 58$ km). The initial proto-slab is uniform and has a dip of $\sim 75^\circ$ (Fig. 4.). During the evolution of this model the slab dip varies with time, but remains uniform along-strike, except for some curvature at the edges (Figs. 4a and S2a). The subducting plate velocity also varies with time, but is always symmetrical and almost uniform along strike (Figs. 4d, 5a and 6a). The Euler pole is located close to $\lambda_e = 90^\circ$ (from the centre of the plate) and remains there throughout the simulation (Fig 6d).

The temporal evolution can be divided in three stages (Figs. 5a and S2a). First, as the slab sinks into the mantle the slab dip remains approximately constant, varying between 70° and 75° . During this period the gravitational pull increases, which causes the average subducting plate

velocity $\bar{v}_s$ to increase from 4.4 cm/yr to a maximum of 19.0 cm/yr at ~4 Myr (Fig. 5a), leading to angular velocities of $\omega_e = 0.3\text{-}1.6^\circ/\text{Myr}$. The second stage begins when the slab reaches the ULMB and the viscous resistance increases due to the one order of magnitude viscosity increase. This causes the slab to flatten, decreasing its dip to 65° (Fig. S2a) and the average subducting plate velocity decreases to 6.6 cm/yr ($\omega_e = 0.5^\circ/\text{Myr}$) in just 2 Myr (Fig. 5a). The final stage occurs after the slab has crossed the ULMB (~6 Myr) and the models evolve more slowly with only minor variations of the slab dip (~ 65°). Subducting plate velocity continues decreasing reaching values of 2.5 cm/yr ($\omega_e = 0.3^\circ/\text{Myr}$) because, as the portion of the slab in the lower mantle lengthens, the viscous mantle drag continues increasing [Capitanio *et al.*, 2007].

3.2. Lateral variations of the overriding plate. In Models A (Videos. 2 and 5) and B (Videos. 3 and 6), the northern portion of the overriding plate has been replaced by older lithosphere (150 Myr, $Z_{1000} = 99$ km), which significantly affects model evolution. Even though both models have a uniform proto-slab, the slab deforms with time leading to a contorted geometry (Figs. 4b, c). In addition, subducting plate velocity is not uniform along strike and the Euler pole migrates closer to the subducting plate (Figs. 4e, f). As in the reference model, the time evolution is characterized by three stages (Figs. 5b, c).

At the beginning of the simulation there are already slight along-strike variations of subducting plate velocity and it is not symmetrical (Figs. 4e, f). The minimum value of the initial subducting plate velocity at the trench is reached to the north ($v_s^{\min} = 3.4$ cm/yr for Model A and $v_s^{\min} = 2.8$ cm/yr in Model B) and the maximum initial value is reached close to the southern edge of the subducting plate ($v_s^{\max} = 4.0$ cm/yr for Model A and $v_s^{\max} = 3.7$ cm/yr for Model B). This variation in subducting plate velocity causes the initial position of the Euler pole to shift southward closer to the plate (Figs. 6e, f): $\lambda_e = 71^\circ$ N in Model A and $\lambda_e = 60^\circ$ N in Model B.

During the first 5 Myr of evolution, the northern portion of the slab, beneath the cold overriding plate, flattens, which locally decreases slab dip by $\sim 20^\circ$ ($\sim 55^\circ$ in the north and $\sim 75^\circ$ in the south, Figs. 3b, c and 4b, c). With time, as the slab sinks into the upper mantle, the average subducting plate velocity increases (Fig. 5). Simultaneously, as the northern portion of the slab flattens (Figs. 4b, c and S2b, c), the gradient in the subducting plate velocity also increases (Figs. 4e, f). For Model A, the maximum velocity variation is reached at 3.4 Myr (Fig. 4e and 6b), approximately the moment when the slab reaches the ULMB, and $v_s^{\min} = 9.2$ cm/yr whereas $v_s^{\max} = 15.6$ cm/yr (Fig. 4e). The evolution of Model B is similar, but the variations of subducting plate velocity are even greater. The maximum difference is reached at 4.6 Myr (Figs. 4.f and 6.c), again around the time when the slab reaches the ULMB, with the minimum in the north ($v_s^{\min} = 3.8$ cm/yr) and the maximum in the south ($v_s^{\max} = 13.4$ cm/yr) (Fig. 4f). During this stage, as the variations of subducting plate velocity increase, the Euler pole migrates southward, getting closer to the subducting plate (Figs. 6.e, f). For Model A, the minimum latitude of the Euler pole is $\lambda_e = 39^\circ$ N at 2.7 Myr and for Model B it is $\lambda_e = 20^\circ$ N at 3.9 Myr.

Once the slab reaches the ULMB (stage 2) the southern portion of the slab flattens due to the increased viscous resistance in the lower mantle, and the variation in slab dip decreases significantly (Figs. 4b, c and S2b, c). The average velocity also decreases to around 5 cm/yr for both models due to the increased viscous resistance of the lower mantle (Fig. 5b, c). Variations in subducting plate velocity also decrease in both Model A and B (Figs. 5 e, f), which causes the Euler pole to migrate back to the north (Figs. 6e, f).

Finally, once the slab has completely crossed the ULMB (stage 3) the system evolves more slowly. In both models, some slab dip variations remain ($\sim 10^\circ$), the average plate velocity remains at around 4 cm/yr ($\omega_e = 0.3^\circ/\text{Myr}$), and the latitude of the Euler pole is at $\sim 85^\circ$ N for both models.

6. Discussion

The models presented here provide a new mechanism to explain the observed occurrence of slower subduction beneath cold/thick overriding lithosphere. Along-strike variations of the thermal state of the overriding plate can cause along-strike variations in the subducting plate velocity, with a slower velocity beneath colder lithosphere. Subducting plate velocity can vary by as much as 10 cm/yr in these models. Due to the rigidity of the plate, the effect of regional variations of the overriding plate is transmitted to the whole plate, leading to a smooth along-strike velocity distribution and variations in plate divergence at the spreading centre. These variations are strongly time-dependent, leading to long migration paths of the Euler pole from 90° to as close as 20° from the centre of the subducting plate. The along-strike variation in velocity is caused by two related mechanisms: stronger viscous coupling between the slab and the mantle wedge and a decrease in slab pull transmitted to the plate.

First, the cold lithosphere extends deeper into mantle wedge beneath the cold overriding plate. Because the viscosity is temperature dependent, this colder part of the mantle wedge also has a higher viscosity. Therefore, the slab is subject to a stronger viscous coupling to the shallow part of the mantle wedge. This increased viscous coupling between the slab and the mantle imparts a higher viscous drag on the slab, which acts to slow the plate. This effect is most noticeable at the beginning of the evolution of Models A and B, when the slab is still uniform, but some along-strike variation in the subducting plate velocity can already be observed. The effect can also be noticed in the differences in the initial location of the Euler pole between different models. Because this effect is only present in the mantle wedge, its importance diminishes as the slab sinks deeper into the mantle.

Second, along strike variations of the slab dip have an even more important effect in subducting plate velocity. Because slab dip decreases beneath the colder region the magnitude of gravitational slab pull transmitted to the plate is reduced. This can be understood from a purely geometric analysis in which the component of negative slab buoyancy aligned in the slab dip direction decreases as slab dip decreases. The lower slab dip in these regions is also a consequence of the colder overriding plate structure, which causes the slab to shallow in the cold region by about 20° relative to the warmer region. As demonstrated with numerical models, assuming a non-linear rheology, both a decrease in subducting plate velocity and an increase in the mantle wedge viscosity lead to a large negative dynamic pressure that flattens the slab [Manea *et al.*, 2012; Rodríguez-González *et al.*, 2012; Rodríguez-González *et al.*, 2014; Taramón *et al.*, 2015].

The results of the numerical simulations agree with a number of observations for the Cocos and Nazca oceanic plate subduction. Both the Cocos and Nazca plates formed from the break-up of the larger Farallon plate at about 23 Ma [Lonsdale, 2005]. Tectonic reconstructions based on finite rotations analyses indicate significant time variations of the rotation pole as well as convergence rates between Nazca (Farallon) and South America [Pardo-Casas and Molnar, 1987; Somoza, 1998; Somoza and Ghidella, 2005; 2012] and between the northern Cocos plate and North America [Schilt and Karig, 1982]. For Nazca-South America, the Euler pole has migrated ~30° since 30 Ma (Somoza, 1998) while at the same time the convergence rate between these plates has decreased from ~15 cm/yr to ~7 cm/yr [Martinod *et al.*, 2010 and references therein]. For the Cocos plate, the pole was 10° to 20° south of its present location and the convergence velocity was 2–4 times faster prior to 10–14 Ma [Schilt and Karig, 1982]. The models predict larger migration of the Euler pole than observed but this is likely because the models use a simple two-plate system and do not include a prior tectonic setting. The magnitude of along-strike variation in velocity and

the time spans over which changes in plate motion and subducting plate velocity take place agree with our model results. In addition, our results show that along-strike variations of subducting plate velocity are transient and their rate of change is significantly damped when the slab reaches the lower mantle. A significant portion of the Nazca slab has already reached the lower mantle and, according to our results, we would expect to observe a progressively more homogeneous subducting plate velocity. Moreover, our models are neglecting other possible along-strike variations that might increase subducting plate velocity variations, such as variable trench velocities.

As with all models we have had to make certain simplification to make the numerical models tractable. The low viscosity channel implemented to decouple both plates constitutes a major limitation of our models. First, in the models presented here, the interplate coupling is not allowed to vary. The reduced coupling is simulated by means of a weak viscosity channel at the plate boundary with an imposed maximum viscosity $\eta_{sz} = 10^{21}$ Pa·s. Because we used the same value for this parameter in all the simulations, we know that the along-strike variations that develop in our model are not related to variations in intraplate coupling. However, in real subduction zones, it is likely that thermal variations in the overriding plate act in combination with other factors controlling subducting plate velocity such as variable intraplate friction [*Di Giuseppe et al.*, 2008; *Duarte et al.*, 2013; *van Dinther et al.*, 2010] and enhanced coupling associated with topographic loads [*Iaffaldano et al.*, 2006; *Meade and Conrad*, 2008].

Second, the low viscosity channel has a fixed location, orientation and width. This means that the models do not allow for the trench to migrate and for dip in the first ~100 km to change, which would otherwise lead to even greater variations of the slab dip, and, in turn, to greater

subducting plate velocity variations. Therefore, we suggest that using a free trench would lead to larger subduction velocity variations and these variations would have a more lasting effect.

In addition, when the slab has reached the lower mantle, all the models shown here present an overturned geometry. This slab shape, with the tip dipping at angles higher than 90° is rarely found in nature. In this case, the geometry is a consequence of the combination of the fixed trench and relatively stiff slabs. The slab needs to bend to subduct, but once it has gone through the plate boundary it is too stiff to unbend and keeps the curled shape. Trench migration is likely to prevent this feature to develop in nature.

Finally, because of the fixity of the trench and overriding plate, our model can only consider the effect of the overriding plate structure on the motion of the subducting plate. For this reason, we use the plate velocities predicted in the no-net-rotation reference frame, so that we are comparing the present-day velocity of the subducting plate to the behaviour of the subducting plate in our models. This limitation makes it more difficult to compare the time-dependence of the Euler poles, because the studies of past plate velocities focus on the relative motion of the subducting and overriding plate. Therefore, we cannot directly compare model predictions with past Euler pole locations but instead only compare the magnitude of the variability in the relative pole locations and magnitude of overall plate motions.

In addition to providing insight into the mechanical processes affecting subduction dynamics, subducting plate velocity has important implications for other geophysical processes. For example, several studies have shown a good correlation between fast subduction, high seismicity rate [*Ide*, 2013], high maximum earthquake magnitude [*Duarte et al.*, 2013; *Jarrard*, 1986; *Ruff and Kanamori*, 1980], and the intensity of tectonic activity in the overriding plate [*Pardo-Casas and Molnar*, 1987]. Subducting plate velocity also has an important influence on the stress regime of

the overriding plate, producing higher stress in regions of fast subduction [Jarrard, 1986]. Therefore, our models would predict higher stress in regions of thin (or warm) overriding lithosphere. However, our models also suggest that higher stress could be found in regions of thick (or cold) overriding plate. We interpret this as a consequence of the lower slab dips obtained in regions where the overriding plate is colder and the mantle wedge more viscous. Because coupling between the overriding plate and the slab is higher for lower slab dips, we expect to find high maximum earthquake magnitude and high stress in the overriding plate in regions of older overriding plate. Therefore, more complete models of the longer-term dynamics are required to understand the opposing processes affecting seismicity rate, maximum magnitude and stress regime in the overriding plate.

5. Conclusions

We suggest that the observed correlation of subducting plate velocity and overriding plate thickness is due to the changes in slab shape that occur in response to increased viscosity in cold regions of the shallow mantle wedge beneath colder overriding plate. Increased viscous coupling between the slab and cold mantle wedge causes an initial reduction in the rate of subduction. The lower subducting plate velocity and higher mantle wedge viscosity also act to increase the magnitude of negative dynamic pressure in the mantle wedge, which causes the slab dip to decrease. At lower slab dip the proportion of gravitational slab pull transmitted to the subducting plate decreases, which further decreases the subducting plate velocity. The decrease in the slab dip has a more important effect than along-strike variations of viscous mantle drag. Corresponding to along-strike variations of subducting plate velocity, the Euler pole changes and it is located closer to the subducting plate when the overriding plate is not uniform. This process is strongly time-

dependent, with significant variations of subducting plate velocity occurring prior to the penetration of the slab into the lower mantle. All of these effects also depend on the width of the cold portion of the overriding plate, and are more significant for wider areas of cold lithosphere (up to half the trench length). Comparisons to the present-day and past evolution of the Cocos and Nazca plates demonstrate that, for a relatively simple subducting plate, the observed variation in subducting plate velocity at the trench can be related to the structure in the overriding plate. This suggest that overriding plate structure is also a factor affecting plate motion in other subduction zones although it may be difficult to demonstrate a direct correlation due to other complicating factors.

Figure captions

Figure 1: Kinematics and thermal state of the Middle America and South America subduction systems. Patterns on the North and South America plates indicate regions with increased depth of the lithosphere-asthenosphere boundary (LAB; solid patterns) and surface heat flow (hashed patterns); orange lines are slab contours (Hayes et al., 2012) every 40 km; colours on the Cocos and Nazca plates represent subducting plate velocity relative to the overriding plate and crosses represent the location of the Euler pole for Cocos (red) and Nazca (blue) according to the MORVEL plate reconstruction in the no net rotation reference frame [DeMets et al., 2010]. Areas of thick lithosphere in South America correspond to the Amazonian (north) and Rio de la Plata (south) Cratons, and in Middle America correspond to the Paleozoic-age continental margin.

Figure 2: Schematic diagram of the three-dimensional model configuration, indicating the geometry and thermal state of the subducting, lateral and overriding plates (red for warm portion

and blue for cold portion); plate boundary shear zone (green line) and the location of the trench (solid triangles). In Model R the overriding plate is uniform. In Models A and B the northern portion of the overriding plate is colder than the reference value. The depth of the 1000 °C isotherm is $Z_{1000} = 99$ km under the cold portion and $Z_{1000} = 58$ km under the warm portion.

Figure 3: Slab cross sections. a) Model R with a uniform overriding plate. b) Model A with a cold overriding plate width of 600 km. c) Model B with a cold overriding plate width of 1400 km. Cross section for Models A and B are shown beneath the cold and warm portions of the overriding plate (left and right column, respectively). For Model R the time shown corresponds to the time of maximum average subducting plate velocity; for Models A and B the time shown corresponds to the time of maximum velocity gradient at the trench. The viscosity structure (colours), temperature (contours every 200 °C) and vertical projection of the velocity field (arrows) are shown.

Figure 4: Slab dip between 120 and 220 km (top) and subducting plate velocity (bottom) variability for model R (a and d), Model A (d and e), and Model B (c and f). For Model R, the three times shown are the initial time, the time of maximum average subducting plate velocity and the final time. For Models A and B, the four times shown are the initial time, the time at which the Euler pole is closest to the subducting plate, time at which the subducting plate velocity gradient at the trench is maximum and the final time.

Figure 5: Time evolution of subducting plate velocity for Model R (a), Model A (b), and Model B (c). Average $\bar{v}_s$ (black), minimum ($v_s^{\min}$, blue) and maximum ($v_s^{\max}$, red) subducting plate velocities are computed at the trench. The minimum subducting plate velocity is reached beneath the cold portion of the overriding plate. Horizontal lines at the bottom of each plot show the duration of each of the three stages of the evolution discussed in the text.

Figure 6: Map view of the subducting plate velocity (top) and time evolution of the Euler pole (bottom) for model R (a and d), Model A (d and e), and Model B (c and f). In the top panels, the magnitude of subducting plate velocity (colours) and velocity field (arrows) are shown; the thermal state of the overriding and lateral plates is indicated (red for warm portion and blue for cold portion). Results are shown at the time of maximum velocity gradient for each model. In the bottom panels, the locations of the Euler pole are shown and the initial time, the time at which the Euler pole is closest to the subducting plate, and the final time are labelled. The overriding and subducting plates are also indicated (red for warm plates, blue for cold overriding plate and green for the subducting plate).

Video captions

Video 1: Temporal evolution of Model R (uniform overriding plate). Top) cross-sections taken at two different latitudes, showing the viscosity structure (colours), temperature (contours every 200°C) and vertical projection of the velocity field (arrows). Bottom left) Time evolution of the slab dip computed between 120 and 220 km deep at two latitudes. Bottom right) Slab dip variations along the trench computed between 120 and 220 km deep.

Video 2: Temporal evolution of Model A (non-uniform overriding plate with a 600 km wide cold portion in the north). Panels show the data as in Video 1.

Video 3: Temporal evolution of Model B (non-uniform overriding plate with a 1400 km wide cold portion in the north). Panels show the same data as in Video 1.

Video 4: Evolution of the Euler pole parameters of Model R (uniform overriding plate). Magnitude of the subducting plate velocity (blue to red colours), location of the Euler pole (longitude and latitude) and the magnitude of the angular velocity (colours) are shown.

Video 5: Evolution of the Euler pole parameters of Model A (non-uniform overriding plate with a 600 km wide cold portion in the north). The data is shown as in Video 4.

Video 6: Evolution of the Euler pole parameters of Model B (non-uniform overriding plate with a 1400 km wide cold portion in the north). The data is shown as in Video 4.

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Figure 1

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