Simulation of Late Cenozoic South American Flat-Slab Subduction Using Geodynamic Models with Data Assimilation

Jiashun Hu*, Lijun Liu, Armando Hermosillo, Quan Zhou

Department of Geology, University of Illinois at Urbana-Champaign, 605 E.

Springfield Ave, Champaign, IL, 61820, USA

Email: jhu16@illinois.edu

Abstract

The formation mechanisms of flat slabs in South America remain unclear. To quantitatively evaluate the earlier proposed mechanisms, we simulate the post-100 Ma subduction history below South America using 4-D geodynamic models by progressively incorporating plate kinematics, seafloor ages and key tectonic features including the buoyant oceanic crust, continental cratons, oceanic plateaus (i.e. the inferred Inca plateau, subducting Nazca Ridge and Juan Fernandez Ridge), as well as deformable trench profiles according to recent geological reconstructions. We find that, in the absence of an overriding plate and subducting buoyancy features, the seafloor age affects slab dip angle by controlling the slab's mechanical strength (i.e., the resistance to bending) and negative buoyancy (integrated positive density anomaly that enhances bending). Our models show that slab strength dominates its buoyancy at age > 30 Ma and the opposite for younger ages. The existence of a thick overriding plate reduces the slab dip by increasing dynamic suction, and individual cratonic roots further lead to along-trench variations of dip angle reduction. While dynamic suction from the overriding plate generates a permanent reduction of the long-wavelength slab dip angle, it is the final addition of

subducting oceanic plateau and aseismic ridges that produces the transient and localized flat-slabs as observed. These results suggest that all mechanisms except the buoyancy features affect the slab dip only at large spatial scales. Our best-fit model with all the above tectonic features included provides a good match to both the upper mantle Benioff zones and the temporal evolution of volcanic arcs since the mid-Miocene. The imperfect match of the Peruvian flat-slab is likely associated with the uncertain 3-D configuration of the Amazonian craton.

Key words: South America, flat slab, aseismic ridge, oceanic plateau, suction force, seafloor age

1. Introduction

South America has the longest continuous subduction zone in the world that stretches 7500 km from $\sim 10^{\circ}\text{N}$ to $\sim 60^{\circ}\text{S}$ (Klotz et al., 2001). Subduction along this ocean-continent convergent boundary started no later than the Cretaceous during the final separation of supercontinent Pangaea (Müller et al., 2008). South America, therefore, provides an ideal test yard for understanding subduction dynamics and the associated mantle processes. The present South American subduction zone is characterized by along-strike variation of slab dipping angles. There are several segments of flat slab including Bucaramanga, Peruvian and central Chile flat slabs (Gutscher et al., 2000), which are intervened by normal to steep subduction (Fig. 1).

In theory, the dip angle of a subducting slab is controlled by two competing forces, the negative buoyancy (i.e., integrated density anomaly) of the slab and resistance from the surrounding mantle and the overriding lithosphere (Stevenson & Turner, 1977). Over the past several decades, a number of possible mechanisms have been proposed to explain the formation of flat-slab subduction. These mechanisms include: (a) the subduction of buoyancy anomalies such as oceanic plateaus, aseismic ridges, and seamount chains (Cloos 1993; Gutscher et al., 1999a; van Hunen et al., 2002; Mason et al., 2010); (b) the fast active overriding of a continent over young oceanic lithosphere (van Hunen et al., 2000); (c) the hydrodynamic suction force due to thick continental roots that couples the subducting slab with the overriding plate (Jones et al., 2011; Roda et al., 2011; Rodríguez-González et al., 2012; O'Driscoll et al., 2009; 2012).

According to recent plate reconstructions (Müller et al., 2008; Seton et al., 2012), South America has not experienced exceptionally fast westward motion since the Cretaceous, implying that the fast overriding continent mechanism may not apply to South America, but the young oceanic lithosphere may play a role. Among existing studies, Gutscher et al. (2000) and Anderson et al. (2007) proposed that the central Chile flat slab, one of the most prominent flat slabs in South America, may be flattened by the Juan Fernandez Ridge. However, the same flat slab was also attributed to the suction force of the Rio de la plata craton (Manea et al., 2012). For the Peruvian flat slab, Gutscher et al. (1999a) argued that it was due to subduction of the “lost Inca plateau” and the Nazca Ridge, although in theory it could also be influenced by the thick Amazonian craton (O'Driscoll et al., 2009; 2012).

In contrast, some studies proposed that moderate-sized buoyancy features could not result in significant slab flattening (Martinod et al., 2005; Gerya et al., 2009). Furthermore, if the basalt-to-eclogite transformation finishes at shallow depths (<70 km), the subduction of oceanic plateau would prohibit the formation of flat slabs (Arrial and Billen, 2013). It is also suggested that the correlation between South American flat slabs and subducting bathymetric highs is rather poor (Skinner and Clayton, 2013). These existing debates raise the need for a systematic evaluation of the different mechanisms for the formation of flat-slab subduction in South America.

Here we report the first 4D quantitative geodynamic modeling on the subduction history of South America. We test both the respective and accumulative effects of seafloor age variation, hydrodynamic suction, and buoyancy feature subduction on the formation of flat slabs. These mechanisms influence the slab dip angle by controlling the negative buoyancy and mechanical strength (i.e., resistance to bending) of the slab, reducing the dynamic pressure in the mantle wedge, and decreasing slab density locally, respectively. By incorporating all these physical processes in the subduction models, we attempt to predict the present-day slab geometry and the temporal evolution of flat-slab formation, which are evaluated with the distribution of deep seismicity (Hayes et al, 2012) and volcanic arc migration from the online catalog of Peruvian Mining and Metallurgical Geological Institute – INGEMMET (<http://www.ingemmet.gob.pe>).

Among the three present-day flat-slab segments in South America (Barazangi and Isacks, 1976, 1979; Pennington, 1981; Ramos et al, 2002; Ramos and Folguera,

2009), the central Chile flat-slab (i.e. Pampean flat slab) that has been in existence since ~12 Ma currently extends from 27° to 33°30'S, the Peruvian flat-slab existing since 11 Ma extends from 5° to 14°S, and the Bucaramanga flat-slab since ~13 Ma covers 5° to 9°N. Here, we mainly focus on the present-day Central Chile and Peruvian flat slabs, since the Bucaramanga segment is too close to the edge of our model domain.

2. Model Setup

We use the 3D spherical finite element code CitcomS (Zhong et al., 2000; Tan et al., 2006) to simulate subduction and mantle convection. Our model domain extends from 10° W to 100° W in longitude, from 70° S to 20° N in latitude, and covers the whole mantle in depth. It is discretized with $257 \times 257 \times 65$ nodes in latitude $\times$ longitude $\times$ depth using a spatially variable finite-element mesh, with the finest resolution of $27 \times 27 \times 8$ (km) occurring in the central 2/3 of the mesh.

2.1. Governing equations

We solve the thermal-chemical mantle convection following the conservation of mass, momentum and energy. We assume the mantle flow is incompressible, with a variable viscosity structure, and satisfies the Boussinesq approximation. The governing equations are:

$$\nabla \cdot \vec{u} = 0 \quad (1)$$

$$-\nabla P + \nabla \cdot [\eta(\nabla \vec{u} + \nabla^T \vec{u})] + (\rho_m \alpha \Delta T + \Delta \rho_c) \vec{g} = 0 \quad (2)$$

$$\frac{\partial T}{\partial t} + \vec{u} \cdot \nabla T = \kappa \nabla^2 T \quad (3)$$

$$\frac{\partial C}{\partial t} + \vec{u} \cdot \nabla C = 0 \quad (4)$$

Where $\vec{u}$ is velocity, P is dynamic pressure, T temperature, ρ_m ambient mantle density, $\Delta\rho_c$ compositional density anomaly, ΔT temperature anomaly, η dynamic viscosity, α thermal expansion coefficient, κ thermal diffusivity, $\vec{g}$ gravitational acceleration, and C composition, respectively.

In Equation (2), both thermal and compositional buoyancy are taken into account. Equation (4) describes the advection of chemical particles, and this is solved using a ratio method for Lagrangian particles (Tackley & King, 2003).

120

121 **2.2. Boundary conditions**

The model utilizes the plate kinematics and seafloor age evolution from the plate reconstruction of Müller et al (2008) to update the surface velocity boundary condition and the upper thermal boundary layer profile, respectively (Figs. 2a-c). All other boundaries are free slip. The core-mantle boundary has a fixed temperature that is set to be the background mantle temperature. The surface velocity and seafloor age are reconstructed over an interval of 1 Ma, using the open source paleogeographic software GPlates (www.gplates.org/; Gurnis et al., 2012). At any given time within the interval, we linearly interpolate the surface velocity and seafloor age. We also take into account the history of central Andean deformation (Fig. 2d) by changing the trench profile through time using the most recent Andean shortening reconstruction (Arriagada et al., 2008).

The model assimilates the seafloor age at each time step to update the top thermal boundary layer within oceanic plates, where we define an error function distribution of thermal profile. In order to overcome the numerical diffusion which exists within the top thermal boundary layer as the finite grid resolution would smear out the sharp thermal gradient in the boundary layer (Davies et al., 2007; Liu and Stegman, 2011), we reduce the thermal gradient across the boundary layer from ~ 1200 °C to ~ 700 °C while maintaining the total buoyancy of the boundary layer unchanged by applying an age-dependent correction to the error function (similar to Liu and Stegman, 2011). The continent is assumed at the ambient mantle temperature, in order to maintain its overall neutral buoyancy.

2.3. Rheology

Three-dimensional temperature- and composition-dependent viscosity is applied in the model. The background mantle viscosity follows a 4-layer profile (Fig. 2e): lithosphere (0-100 km), asthenosphere (100-300 km), transition zone (300-660 km), and lower mantle (660 – 2890 km). Their respective viscosity values are 10^{23} , 3×10^{19} , 10^{21} , and 3×10^{22} Pa s. In the lower mantle, the viscosity increases from 3×10^{22} Pa s right below 660 km depth to 10^{23} Pa s at >1200 km, and then gradually decreases below 2000 km toward the core-mantle boundary to 3×10^{22} Pa s. Such a viscosity profile is obtained by matching the present-day mantle seismic structure with predicted slab geometry. This profile is also consistent with recent petrological experiment by Marquardt and Miyagi (2015), suggesting a significantly increased strength of ferropericlasite in the upper part of lower mantle. This is further

supported by the observation of global slab stagnancy above 1000 km depth from recent tomography images (Fukao & Obayashi, 2013).

In order to facilitate subduction, the model has also incorporated sharp lateral viscosity variations, including a weak zone (with a viscosity of 10^{19} Pa s) along the subduction zone and a weak mantle wedge above the down-going slab, similar to earlier models by Liu & Stegman (2011). To simulate the dynamic suction force from cratons, we prescribe these continental units as high-viscosity bodies. We first define a 75-100-km thick continental lithosphere, and then prescribe individual cratonic roots, which extends to about 200 km depth. The geometry of cratonic roots at 75 km depth is defined according to Loewy et al. (2004) (Fig. 1). The area of the roots shrinks as it goes from the surface to 200 km depth, in order to avoid a step-like craton viscosity profile. These high-viscosity features regulate the flow pattern in the upper mantle and generate suction force to flatten the slab.

2.4. Implementation of oceanic crust, plateau and mid-ocean ridges

We use compositional tracers to define the oceanic crust and the oceanic plateaus. The geometry of Inca Plateau, Nazca Ridge and Juan Fernandez Ridge are prescribed according to Gutscher et al. (2000) (Figure 1). These features are embedded in the subducting oceanic plate. We assume that the oceanic crust is 8 km thick with a density of 2.9 g/cm^3 , and that the oceanic plateau and aseismic ridges are 15 km thick with the same density. These features reduce the density of the down-going slab, and thus could potentially flatten the slab. We simulate the density change due

to basalt-to-eclogite phase transformation by turning off tracer buoyancy at ~100 km depth for both the crust and oceanic plateaus.

The oceanic crust is implemented into the model since 100 Ma. As seafloor spreads, new oceanic crust forms at the mid-ocean ridge where the seafloor is young. We update the composition of young seafloors (<10 Ma) at each time step, in order to mimic the oceanic crust generation along mid-ocean ridges. We introduce the Inca Plateau, Nazca Ridge and Juan Fernandez Ridge into the model at 15 Ma, with their initial configuration reconstructed starting from their present-day positions (Gutscher et al., 2000) using past plate motions (Müller et al., 2008) with GPlates (Boyden et al., 2011). In our model, the subduction of Inca Plateau and Nazca Ridge began at ~12 Ma and that of Juan Fernandez Ridge began at ~9 Ma. We realize the uncertainty in the earlier subducted length of Juan Fernandez Ridge (Yáñez et al., 2001; Kay and Mpodozis, 2002), but the simulation of a longer ridge subduction history results in a similar present-day slab structure to that of a shorter history. This may be related to the change of the ridge's strike orientation in the past and its southward migration relative to South America (Huene et al., 1997). We conclude that only the recent subduction history (post-10 Ma) of the ridge has contributed to the formation of the Central Chile flat slab.

2.5. Assimilation of a deformable South American trench

There are multiple palinspastic reconstructions of the Cenozoic shortening of the central Andes (Isacks, 1998; Kley and Monaldi, 1998; McQuarrie, 2002; Oncken et al., 2006; Arriagada et al., 2008). Among these reconstructions, we choose the one from

Arriagada et al. (2008) to represent the trench profile evolution during the geological past. This is because their model incorporates the most extensive tectonic shortening events and paleomagnetically determined tectonic rotations.

In practice, we first collect the trench profiles at 45, 15, and 0 Ma from the reconstruction by Arriagada et al. (2008) (there was no significant shortening before 45 Ma). We then linearly interpolate these trench profiles and calculate the shortening rate at any given time and space, assuming shortening occurs only along the east-west direction. To the first order, this is a valid approximation, since the convergence rate between the Farallon/Nazca plate and South American plate was mainly along the east-west direction during the past 100 Ma (e.g., Müller et al., 2008). After we obtain these trench profiles, we rotate them back in time, in order to get the absolute locations of the South American trench history (Fig. 2d).

Subsequently, we update both the velocity and thermal boundary conditions according to these deformable trench locations. The shortening of the overriding plate was accommodated in a zone of ~ 700 km width behind the trench, where the shortening rate decreases linearly inland-ward. Effectively, our approach may be similar to the one by Gurnis et al. (2013) and Flament et al. (2014), and their approach also requires substantial work in code-development and data-compilation.

3. Results

In this section, we present model results testing various physical mechanisms for flat-slab formation, and the predicted modern slab configuration in South America.

3.1 The physical mechanisms for flat-slab formation

As discussed in section 1, among earlier proposed mechanisms for flat-slab formation, those relevant to South American subduction include dynamic suction caused by strong cratonic roots inside the overriding plate (Manea et al., 2012), subduction of intra-slab buoyancy anomalies associated with aseismic ridges and oceanic plateaus (Gutscher et al., 1999a), and subduction of young seafloors (van Hunen et al., 2000). In this paper, we attempt to evaluate the relative contributions of these different mechanisms for the resulting present-day slab geometry.

3.1.1 Seafloor age

To explore the seafloor age effect, we first design two different models: one has a uniform seafloor age of 30 Myr during the entire subduction history (model I), and the other has temporal-spatial variations of seafloor age (model II) derived from the plate reconstruction of Müller et al., (2008) (Fig. 2). These two models do not include a high-viscosity overriding plate and intra-slab buoyancy features in order to isolate the effect of seafloor age. In model I, the present-day slab subducts steeply along the entire trench (Figs. 3a, 4). In contrast, the slab in model II is shallower almost everywhere along the trench (Figs. 3b, 4). The slab dip angle (defined by the upper surface of the slab between the trench and at 200 km-depth, Fig. 5b) varies along the trench, with two shallow-dipping slab segments centered at $\sim 10^{\circ}\text{S}$ and $\sim 29^{\circ}\text{S}$, respectively (Figs. 3a, 4). Since the only difference between model I and II is the seafloor age distribution (Fig. 3a, b), their different dip angles reflect the competition between the slab negative buoyancy and mechanical strength. However, the dependence of this competition, therefore slab dip, on seafloor age is elusive. For example, compared to model I, both younger (<30 Ma, north of 10°S and

south of 30°S) and older (>30 Ma, between 10°S and 30°S) seafloors lead to shallower subduction in model II, although we do observe a larger dip angle reduction at the central part of the trench where seafloor is older.

In order to better discern the effect of seafloor age on slab dip, we run an additional model with a uniform seafloor age of 50 Ma (model S), an age similar to the oldest part (between ~15°S and ~30°S) of that in model II at the present day (Fig. 4). The predicted present-day slab in model S is systematically shallower than that in model I. It is also shallower than the slab in model II, except for the central portion (Fig. 4), where the two models have similar seafloor ages (Fig. 2a-c). Comparing model I (uniform age at 30 Ma) and model S (uniform age at 50 Ma), we can see that older seafloor age tends to induce shallower subduction, which suggests that slab mechanical strength dominates the effect of negative buoyancy on slab bending. This conclusion can also be drawn from the comparison of model II and model S, where between ~15°S and ~30°S, the slab dip angles in the two models are similar, consistent with their similar seafloor ages in this region; while in the other regions, the slab of model S is clearly shallower, corresponding to its older seafloor ages. Together with the comparison between model I and II, we suggest that for seafloor ages >30 Ma, the slab strength dominates its negative buoyancy in determining the slab dip, and that this relation reverses for ages <30 Ma. It is worth noting that, between ~25°S and ~30°S, all three models have close slab dip angles, which is largely related to the anchoring effect of slabs at greater depth due to early subduction, a topic outside the scope of this paper. Furthermore, in model I, the

sharp kinks in slab dip distribution between $\sim 15^{\circ}\text{S}$ and $\sim 20^{\circ}\text{S}$ (Fig. 4) may be related to the variation of trench curvature (Fig. 3b).

Overall, with the tectonically reconstructed seafloor age history (i.e. model II), the predicted present-day slab dip angle distribution along trench is opposite to that outlined with Benioff zones, where shallow-to-flat slabs occur both north and south of central Andes (Figs. 1, 3b). Furthermore, the predicted subduction dip angle is consistently steeper than that observed (Fig. 3b). We note that, although an increase in the background mantle viscosity or slab strength could lead to shallower slabs, it would not explain the N-S variation of slab dip angle. This suggests that additional physical mechanisms are needed in order to explain the observed slab geometry.

3.1.2 Hydrodynamic suction

In order to investigate the effect of hydrodynamic suction, we perform two additional models, which have a uniform 100 km-thick overriding plate (model III) and a uniform 75 km-thick overriding plate with three 200 km-thick embedded cratons (model IV), respectively. Compared with model II, inclusion of a mechanically strong overriding plate (model III) leads to a reduction of slab dip along most of the South American trench (Figs. 3c, 5, 6). Addition of thick cratons (Amazonia, Sao Francisco, and Rio de la plata cratons; Fig. 1) inside the continent (model IV) reduces the dip angle further and generates long-wavelength (>1000 km) low-amplitude perturbations to the slab dip distribution (Figs. 3d, 5). This is due to enhanced hydrodynamic suction force from the overriding continent, which is reflected as a reduction of dynamic pressure inside the mantle wedge (Figs. 7, 8)

(Stevenson & Turner, 1977; Manea et al., 2012). Among models II, III and IV, the magnitude of dynamic pressure reduction in the mantle wedge increases progressively (Fig. 7b-d), resulting in shallower and shallower slabs (Figs. 3b-d, 6). Physically, the reduction of dynamic pressure in the mantle wedge, and thus increased suction force, results from the existence of a high-viscosity overriding plate which confines the mantle flow to occur within a narrow region (Fig. 7; Stevenson & Turner, 1977). In a vertical profile, when a thick overriding plate exists, the lowered dynamic pressure occurs inside a very broad upper mantle region above the slab (Fig. 8), with the lowest pressure present in the asthenosphere whose viscosity is the lowest.

In addition to confirming earlier studies on the origin of hydrodynamic suction force (Manea & Gurnis, 2007; Manea et al., 2012), our results further quantify the magnitude and spatial pattern of this force associated with the 3D geometry of the overriding plate including cratonic roots. The observation that the predicted slab dips at the long-wavelength in models III and IV are close to that observed suggests that their predicted magnitude of suction force is reasonable, and a further increase of suction force would over-predict the flattening of slabs along most of the trench. Relative to model II, the reduction of slab dip angle in model IV is larger than that in model III, especially along the segments between $\sim 10^\circ$ and 17°S and between $\sim 25^\circ$ and 33°S , which correspond to the locations of the Amazonia and Rio de la plata cratons (Fig. 5a). This is because these cratons further confine the nearby mantle flow, resulting in greater suction (Fig. 7d).

In summary, model IV predicts the observed slab dip angles slightly better than model III, but both models still fail to predict the localized flat slabs in Southern Peru and Central Chile (Fig. 3c, d). In fact, the central South American trench flattening is already over predicted at depths above 150 km, but the northern and southern parts, especially the two flat-slab segments, are still under predicted. We do not attempt to improve this fit by tuning the thickness and viscosity profile of these cratons, because, first, geometry of these craton roots is not well constrained in observation; second, suction-induced flat slabs are likely permanent in time (Manea et al., 2012; Taramón et al., 2015), inconsistent with the recent history of these flat slabs. Finally, the fact that the width of the central Chile flat slab (Fig. 1) is much smaller than that of the Rio de la plata craton (thus the dimension of suction-induced permanent flat slab) dismisses a direct causal relationship between the two.

3.1.3 Subduction of buoyancy features

We then further incorporate (model V) the three pieces of over-thickened oceanic crusts associated with the Inca Plateau, Nazca Ridge, and Juan Fernandez Ridge (Fig. 1) whose subduction started since late Miocene (Gutscher et al., 1999a; Ramos, 2009) (Fig. 2a). The extra positive buoyancy from these features relative to the ambient mantle causes additional flattening of the slabs in Peru and Central Chile, greatly improving the fit to the *Slab1.0* contours (Fig. 3d, e). The predictions for the Central Chile and Southern Peru flat slabs are almost exact, given the observation that it is the hydrodynamic suction from the continent that slightly over-predicts the slab dip at 100 km depth between 20°S and 30°S (Fig. 3c, d). The amount of flattening in Northern Peru at 250 km depth is still slightly under predicted (Fig. 3e),

and this is likely related to the under-estimation of suction force due to the uncertain geometry of Amazonia Craton. Overall, compared with model IV where only seafloor age variation and dynamic suction are considered (Fig. 3d), the buoyancy features contribute to slab flattening at much smaller spatial scales (<500 km), comparable to the widths of these buoyancy features (Fig. 5a). We emphasize that the magnitudes of the background mantle viscosity and slab strength affect the overall slab dip along the entire trench but contribute little to the local variations of slab dip angle. So we utilize the same viscosity structure for all five models.

Besides the spatial scales, the temporal effect of subducting buoyancy features also differs from that of continental suction. Since the suction force resulting from the overriding plate exists throughout the subduction history, it exerts a prolonged influence on the subducting slab, which is reflected as a continuous reduction of slab dip angle over depth (thus time) in models III and IV relative to that in model II (Figs. 3, 8). The addition of aseismic ridges, however, affects the slab dip angle only since their initial subduction during the late Miocene, which corresponds to ~500 km depth at the present. That also explains why there is almost no difference between model IV and model V in slab geometry at greater depths (Figs. 3, 6, 8).

One of the most significant structural differences between our results and earlier studies is the development of a tear inside the flat slab associated with the subduction of buoyancy features (Fig. 9). Due to the progressive accumulation of strain at the down-dip side and the two flanks of the aseismic ridge during subduction, the flat slab is highly deformed internally and eventually a slab hole forms. The two pairs of spines in the outline of slab dip angle distribution at ~13°S

and $\sim 31^\circ\text{S}$ in Figure 5a are caused by these slab holes (Figs. 3e, 5b and 9). In order to be comparable to the Slab 1.0 contours that are spatially interpolated from the seismicity distribution, we also interpolated these parts of the slab dip angle using dashed lines (Fig. 5a). This way, we could see more clearly the small spatial scales ($< 500\text{ km}$) of slab flattening due to the extra buoyancy of these over-thickened oceanic crusts (Fig. 5a, b).

Analysis of the temporal evolution of flat slabs in model V suggests that the formation of flat slabs was usually delayed by several million years relative to the start of buoyancy-feature subduction. This reflects the 3D structure of a flat slab that consumes the earlier subducted portion of the buoyancy feature to ‘raise’ the slab relative to its surroundings, consistent with the results of analog models (Espurt et al., 2008). For example, the Nazca Ridge entered the trench at $\sim 12\text{ Ma}$, but the slab reached its flattest state at $\sim 4\text{ Ma}$ (Fig. 9), which is also consistent with the cessation of arc volcanism at this time (Fig. 9). A similar situation is seen during the subduction of the Inca Plateau (Fig. 9).

3.2 The predicted present-day flat slabs

Figure 10 shows six different vertical cross sections of our best-fit model (i.e. model V). We compare this model prediction with upper mantle Benioff zone geometry using *slab 1.0* (Hayes et al., 2012) and seismicity catalog from IRIS (<http://service.iris.edu/fdsnws/event/docs/1/builder/>). Three cross sections through the flat-slab segments (Fig. 10a, b, f) and three through the normal subduction parts (Fig. 10c-e) are shown. As can be seen, the predicted slab

377 geometry fits the Benioff zone generally well both in the normal and flat slab
378 segments (Fig. 3e, 10). We do not fine-tune the slab strength and overriding plate
379 thickness, in order to better match these profiles, since we mainly focus on the first
380 order effect of the different physical mechanisms for slab flattening, and these
381 models (I-V) demonstrate a clear improvement toward the final prediction of the
382 observed slab geometry.

383 In central Chile (Fig. 10f), the Juan Fernandez Ridge has effectively flattened the
384 slab locally. This predicted slab geometry fits both the *slab 1.0* contour and the
385 seismicity pattern above 150 km depth very well, but not below this depth (Fig. 10f).
386 We suggest that this misfit is due to the interpolated nature of *slab 1.0* below 150
387 km depth, where there is no seismicity between 250 and 500 km along the entire
388 South American trench (Hayes et al., 2012). The same situation and reasoning apply
389 to the Peruvian flat slab, where the Benioff zone is interpolated with depth (Fig. 10a,
390 b). However, a confirmed mismatch in the predicted Peruvian flat slab is that its
391 overall depth is slightly larger compared to the Benioff zone at <200 km depth. This
392 is likely due to two reasons: 1) The buoyancy of Inca Plateau and Nazca Ridge is
393 insufficient to cause a wide flat slab as observed; 2) A larger continental suction is
394 needed in order to generate a shallower slab.

395 Significantly, our study reveals some important differences from earlier 2-D
396 simulations on flat-slab formation. According to the traditional view, subduction of a
397 buoyancy feature (e.g., van Huneun et al., 2002) or suction from a thick craton
398 (Manea et al., 2012) would result in a local decrease of dip angle inside a continuous
399 slab. Our results, however, suggest that the down-dip part of the flat-slab steepens

abruptly with increasing depth (Fig. 10a, b and f). This increase of slab dip below the flat slab is also consistent with analog model for buoyancy feature subduction (Martinod et al., 2005) and confirmed by tomographic images (Gutscher et al., 2000). This geometry variation reflects the reduced sinking speed of the flattening slab segment relative to surrounding slabs, and this different motion ultimately results in the slab holes as discussed earlier (e.g., Fig. 9). These apparent 3D effects are missing in earlier 2D models due to the assumption of infinite length in the third dimension.

4. Discussion & Conclusions

According to our study, the modern flat slab segments in South America are caused by a combination of seafloor age variation, hydrodynamic suction from the overriding plate, and subduction of buoyancy features. This may explain why there is a lack of clear correlation between seafloor age and slab dip global-wide (Jarrard 1986; Lallemand et al., 2005; Cruciani et al., 2005). We find that with seafloor age > 30 Ma, older seafloor age (< 60 Ma in South America) tends to cause flatter subduction (model I vs. model S), which is consistent with earlier modeling work (Billen and Hirth, 2007). This suggests that slab strength dominates negative buoyancy in controlling slab dip in this age range. However, with seafloor age < 30 Ma, we find that younger seafloor age leads to shallower slab dip (model I vs. model II), which likely results from the overall small negative buoyancy of the young slabs and the stronger effect of the buoyant oceanic crust, relative to older slabs. This conclusion on how seafloor age influences slab dip, however, should be subject to future verification.

The flexibility of data assimilation models also allows us to quantify the respective roles of the overriding plate and subducting buoyancy features on influencing the slab dip angle. According to our results, hydrodynamic suction only controls the long-wavelength variation of slab dip angle along trench (Figs. 3, 5). In particular, a uniformly thick overriding plate causes a systematic reduction of the slab dip angle (Fig. 5); the incorporation of cratons induces additional perturbations to the slab dip over ~ 1000 km spatial scale, which is considerably larger than that induced by the subduction of buoyancy features (Fig. 5). As a result, the sharp kinks of the slab contour at 100 km depth in Chile and Peru (Fig. 1) are most likely caused by the subduction of two narrow aseismic ridges, since the suction from nearby cratons would have elevated a much wider segment of the slab.

It is worth noting that, in our model, flattening of the Peruvian slab is slightly under-predicted (Fig. 3e). This requires either further reduced slab density or a stronger interpolate coupling across this region. The latter case could result from either a thicker overriding plate or a higher mantle-wedge viscosity (Manea and Gurnis, 2007; Rodríguez-González et al., 2012; 2014) than those assumed in our models. In reality, the subduction of buoyancy features could have flattened the slab to a critical angle, which then initiated a positive feedback between slab flattening and increased suction (Stevenson & Turner, 1977; Taramón et al., 2015), further reducing the slab dip angle (Fig. 1). As shown in a recent modeling study with non-linear rheology (Taramón et al., 2015), this positive feedback could have formed a permanent flat slab in Peru where the seafloor is relatively young (Fig. 2).

There is still an ongoing debate on whether subducting buoyancy features can cause flat slabs. While earlier 2D numerical models (e.g., van Hunen et al., 2000) support this hypothesis, a recent study (Gerya et al., 2009) suggests that aseismic ridges with moderate sizes could not generate flat slabs. Our 3D models with data assimilation demonstrate that the subduction of buoyancy features clearly contributes to the formation of the localized flat slabs in Peru and Chile (Fig. 5a). By simultaneously considering other physical mechanisms, we show that although down-going buoyancy features can effectively reduce the slab dip angle, the presence of other mechanisms such as dynamic suction is also necessary in order to predict the exact amount of flattening. The inadequate flattening of the Peruvian slab in our model represents a potential example for the need of additional lifting forces besides slab buoyancy.

The same reasoning may help to explain the poor spatial and temporal correlation between subducting buoyancy features and observed flat slabs (Skinner and Clayton, 2013), since buoyancy perturbation within the down-going slab is not the only driving force for flat slabs to form. Another possible reason is related to the temporal delay between slab formation and the initiation of buoyancy feature subduction. For example, the Juan Fernandez Ridge started to subduct at ~12 Ma (Yáñez et al., 2001), but the cessation of volcanism began at ~5 Ma (Espurt et al., 2008). A similar delay is also observed in Peru as discussed earlier (Fig. 9). This may further explain why the subduction of Carnegie Ridge and Iquique Ridge do not correlate with on-going flat slabs, given the short history of the Iquique Ridge subduction (<2 Ma, Rosenbaum et al., 2005) and the nascent slab flattening process

associated with the Carnegie Ridge (Gutscher et al. 1999b). In addition, the poor correlation between flat slabs and buoyancy feature subduction can also be due to the tendency of the slab to stay flat long after the subduction of the buoyancy feature, as observed in both analog (Martinod et al., 2005) and numerical (Taramón et al., 2015) models. In this case, an extant flat slab could be due to an ancient buoyancy feature subduction.

The volcanic history in Peru provides useful constraints on the development of flat slabs during the subduction of buoyancy features. Prior to 4 Ma, the arc volcanism is largely continuous in Southern Peru. After this time, a clearly defined volcanic gap emerged (Fig. 9), which coincides with the slab flattening due to the Nazca Ridge subduction. While this confirms the general belief on the relation between volcanic gap and flat slab (Nur and Ben-Avraham, 1981; McGeary et al., 1985; Ramos and Folguera, 2009), our model further shows a significant delay of slab flattening and the cessation of volcanism, relative to the initial subduction of the buoyancy feature. Furthermore, a possible delay by ~2-4 Myr of the volcanic gap formation from initial slab flattening is also observed (Fig. 9). If correct, this may reflect the secular cooling of the mantle wedge above a flat slab (van Hunen et al., 2002).

On the other hand, there seems to be long-lived magmatic quiescence between the equator and 6°S (Fig. 9). Subduction of the Inca Plateau further cleared out some sporadic volcanisms along these latitudes that were erupted prior to 8 Ma (Fig. 9). Temporally and spatially, this lack of volcanism correlates with an overall shallow to flat slab segment that has been in existence for more than 14 Myr (Fig. 9). This

abnormal slab segment is not due to the subduction of buoyancy features. Possible reasons for its formation may include the small trench curvature, past plate motions, and the earlier subduction history, given the presence of this slab feature in most models presented (Fig. 3). Since all our models use the same plate kinematics, we could not separately evaluate the dependence of slab dip on the overriding plate motion (Jarrard, 1986; van Hunen et al., 2000; Lallemand et al., 2005), or trench curvature (Tovish and Gerald, 1978; Schellart et al., 2007). A better understanding on this phenomenon will require future modeling efforts.

Finally, we emphasize that future research is necessary to further verify the dynamics of flat slab subduction. An important aspect is mantle rheology. In our models, we search for an effective viscosity structure that best matches the present-day mantle seismic image. A more physically self-consistent mantle rheology is nonlinear viscosity including pseudo-plasticity (Bello et al., 2015) and strain rate-weakening (Taramón et al., 2015). However, to what extent the nonlinear rheology can approximate subduction dynamics is still debated (Stegman et al., 2006; Stadler et al., 2010; Liu and Stegman, 2011; Arrial and Billen, 2013), which is likely due to the incapability of any single rheology to capture the natural complexity of the Earth. Since we are modeling the long-term subduction history of South America by assimilating many different data types, it is impractical to include such a complex rheology into our model. Future research, therefore, should focus on improved data assimilation techniques and more efficient computation methods.

Acknowledgement: We thank Xiaodong Song, Jay Bass, Craig Lundstrom and William Guenther for helpful discussions. The numerical models are performed using CitcomS (www.geodynamics.org) and GPlates (www.gplates.org). Figures are prepared using the GMT software package (<https://www.soest.hawaii.edu/gmt/>). This work is supported by NSF grant EAR-1345135.

Reference:

Anderson, M., Alvarado, P., Zandt, G., & Beck, S. (2007). Geometry and brittle deformation of the subducting Nazca Plate, Central Chile and Argentina. *Geophysical Journal International*, 171(1), 419-434.

Antonijevic, S. K., Wagner, L. S., Kumar, A., Beck, S. L., Long, M. D., Zandt, G., ... & Condori, C. (2015). The role of ridges in the formation and longevity of flat slabs. *Nature*, 524(7564), 212-215.

Arriagada, C., Roperch, P., Mpodozis, C., & Cobbold, P. R. (2008). Paleogene building of the Bolivian Orocline: Tectonic restoration of the central Andes in 2-D map view. *Tectonics*, 27(6).

535 Arrial, P. A., & Billen, M. I. (2013). Influence of geometry and eclogitization on
 536 oceanic plateau subduction. *Earth and Planetary Science Letters*, 363, 34-43.
 537
 538 Barazangi, M., & Isacks, B. L. (1976). Spatial distribution of earthquakes and
 539 subduction of the Nazca plate beneath South America. *Geology*, 4(11), 686-692.
 540
 541 Barazangi, M., & Isacks, B. L. (1979). Subduction of the Nazca plate beneath Peru:
 542 evidence from spatial distribution of earthquakes. *Geophysical Journal*
 543 *International*, 57(3), 537-555.
 544
 545 Bello, L., Coltice, N., Tackley, P. J., Müller, R. D., & Cannon, J. (2015). Assessing the
 546 role of slab rheology in coupled plate-mantle convection models. *Earth and*
 547 *Planetary Science Letters*, 430, 191-201.
 548
 549 Billen, M. I., & Hirth, G. (2007). Rheologic controls on slab dynamics. *Geochemistry,*
 550 *Geophysics, Geosystems*, 8(8).
 551
 552 Boyden, J. A., Müller, R. D., Gurnis, M., Torsvik, T. H., Clark, J. A., Turner, M., ... &
 553 Cannon, J. S. (2011). Next-generation plate-tectonic reconstructions using GPlates.
 554 *Geoinformatics: cyberinfrastructure for the solid earth sciences*, 95-114.
 555

556 Cloos, M. (1993). Lithospheric buoyancy and collisional orogenesis: Subduction of
 557 oceanic plateaus, continental margins, island arcs, spreading ridges, and seamounts.
 558 Geological Society of America Bulletin, 105(6), 715-737.
 559
 560 Cruciani, C., Carminati, E., & Doglioni, C. (2005). Slab dip vs. lithosphere age: no
 561 direct function. Earth and Planetary Science Letters, 238(3), 298-310.
 562
 563 Davies, D. R., Davies, J. H., Hassan, O., Morgan, K., & Nithiarasu, P. (2007).
 564 Investigations into the applicability of adaptive finite element methods to two -
 565 dimensional infinite Prandtl number thermal and thermochemical convection.
 566 Geochemistry, Geophysics, Geosystems, 8(5).
 567
 567 Espurt, N., Funiciello, F., Martinod, J., Guillaume, B., Regard, V., Faccenna, C., &
 568 Brusset, S. (2008). Flat subduction dynamics and deformation of the South American
 569 plate: Insights from analog modeling. Tectonics, 27(3).
 570
 570 Flament, N., Gurnis, M., Williams, S., Seton, M., Skogseid, J., Heine, C., & Müller, R. D.
 571 (2014). Topographic asymmetry of the South Atlantic from global models of mantle
 572 flow and lithospheric stretching. Earth and Planetary Science Letters, 387, 107-119.
 573
 574 Fukao, Y., & Obayashi, M. (2013). Subducted slabs stagnant above, penetrating
 575 through, and trapped below the 660 km discontinuity. Journal of Geophysical
 576 Research: Solid Earth, 118(11), 5920-5938.
 577

578 Gerya, T. V., Fossati, D., Cantieni, C., & Seward, D. (2009). Dynamic effects of aseismic
579 ridge subduction: numerical modelling. *European Journal of Mineralogy*, 21(3), 649-
580 661.

581

582 Gurnis, M., Turner, M., Zahirovic, S., DiCaprio, L., Spasojevic, S., Müller, R. D., ... &
583 Bower, D. J. (2012). Plate tectonic reconstructions with continuously closing plates.
584 *Computers & Geosciences*, 38(1), 35-42.

585 Gurnis, M., TURNER, M., BOWER, D., FLAMENT, N. E., & MÜLLER, R. D. (2013,
586 October). Advancing cyberinfrastructure for plate reconstructions: Present day
587 mantle structure and the stratigraphic architecture of continents as signals of a
588 dynamic Earth. In *Geol. Soc. Am. Abstr. Programs* (Vol. 45, No. 7).

589

590 Gutscher, M. A., Olivet, J. L., Aslanian, D., Eissen, J. P., & Maury, R. (1999a). The “lost
591 Inca Plateau”: cause of flat subduction beneath Peru?. *Earth and Planetary Science*
592 *Letters*, 171(3), 335-341.

593

594 Gutscher, M. A., Malavieille, J., Lallemand, S., & Collot, J. Y. (1999b). Tectonic
595 segmentation of the North Andean margin: impact of the Carnegie Ridge collision.
596 *Earth and Planetary Science Letters*, 168(3), 255-270.

597

598 Gutscher, M.A., Spakman, W., Bijwaard, H., Engdahl, E. R. (2000). Geodynamics of flat
599 subduction: seismicity and tomographic constraints from the Andean
600 margin. *Tectonics*, 19(5), 814-833.

601

602 Hayes, G. P., Wald, D. J., & Johnson, R. L. (2012). Slab1. 0: A three-dimensional model
603 of global subduction zone geometries. *Journal of Geophysical Research: Solid Earth*
604 (1978–2012), 117(B1).

605

606 Huene, R. V., Corvalán, J., Flueh, E. R., Hinz, K., Korstgard, J., Ranero, C. R., &
607 Weinrebe, W. (1997). Tectonic control of the subducting Juan Fernández Ridge on
608 the Andean margin near Valparaíso, Chile. *Tectonics*, 16(3), 474-488.

609

610 Isacks, B. L. (1988). Uplift of the central Andean plateau and bending of the Bolivian
611 orocline. *Journal of Geophysical Research: Solid Earth* (1978–2012), 93(B4), 3211-
612 3231.

613

614 Jarrard, R. D. (1986). Relations among subduction parameters. *Rev. Geophys*, 24(2),
615 217-284.

616

617 Jones, C. H., Farmer, G. L., Sageman, B., & Zhong, S. (2011). Hydrodynamic
618 mechanism for the Laramide orogeny. *Geosphere*, 7(1), 183-201.

619

620 Kay, S. M., & Mpodozis, C. (2002). Magmatism as a probe to the Neogene shallowing
621 of the Nazca plate beneath the modern Chilean flat-slab. *Journal of South American*
622 *Earth Sciences*, 15(1), 39-57.

623

624 Kley, J., & Monaldi, C. R. (1998). Tectonic shortening and crustal thickness in the
625 Central Andes: How good is the correlation?. *Geology*, 26(8), 723-726.
626

627 Klotz, J., Khazaradze, G., Angermann, D., Reigber, C., Perdomo, R., & Cifuentes, O.
628 (2001). Earthquake cycle dominates contemporary crustal deformation in Central
629 and Southern Andes. *Earth and Planetary Science Letters*, 193(3), 437-446.
630

631 Lallemand, S., Heuret, A., & Boutelier, D. (2005). On the relationships between slab
632 dip, back - arc stress, upper plate absolute motion, and crustal nature in subduction
633 zones. *Geochemistry, Geophysics, Geosystems*, 6(9).
634

635 Liu, L., & Stegman, D. R. (2011). Segmentation of the Farallon slab. *Earth and*
636 *Planetary Science Letters*, 311(1), 1-10.
637

638 Loewy, S. L., Connelly, J. N., & Dalziel, I. W. (2004). An orphaned basement block: The
639 Arequipa-Antofalla Basement of the central Andean margin of South
640 America. *Geological Society of America Bulletin*, 116(1-2), 171-187.

641 Marquardt, H., & Miyagi, L. (2015). Slab stagnation in the shallow lower mantle
642 linked to an increase in mantle viscosity. *Nature Geoscience*.

643 Mason, W. G., Moresi, L., Betts, P. G., & Miller, M. S. (2010). Three-dimensional
644 numerical models of the influence of a buoyant oceanic plateau on subduction
645 zones. *Tectonophysics*, 483(1), 71-79.

646

647 Manea, V., & Gurnis, M. (2007). Subduction zone evolution and low viscosity wedges
648 and channels. *Earth and Planetary Science Letters*, 264(1), 22-45.

649

650 Manea, V. C., Pérez-Gussinyé, M., & Manea, M. (2012). Chilean flat slab subduction
651 controlled by overriding plate thickness and trench rollback. *Geology*, 40(1), 35-38.

652 Martinod, J., Funiciello, F., Faccenna, C., Labanieh, S., & Regard, V. (2005). Dynamical
653 effects of subducting ridges: insights from 3-D laboratory models. *Geophysical*
654 *Journal International*, 163(3), 1137-1150.

655 McGeary, S., Nur, A., & Ben-Avraham, Z. (1985). Spatial gaps in arc volcanism: the
656 effect of collision or subduction of oceanic plateaus. *Tectonophysics*, 119(1), 195-
657 221.

658 McQuarrie, N. (2002). Initial plate geometry, shortening variations, and evolution of
659 the Bolivian orocline. *Geology*, 30(10), 867-870.

660 Müller, R. D., Sdrolias, M., Gaina, C., & Roest, W. R. (2008). Age, spreading rates, and
661 spreading asymmetry of the world's ocean crust. *Geochemistry, Geophysics,*
662 *Geosystems*, 9(4).

663 Nur, A., & Ben-Avraham, Z. (1981). Volcanic gaps and the consumption of aseismic
664 ridges in South America. *Geological Society of America Memoirs*, 154, 729-740.

665 Oncken, O., Hindle, D., Kley, J., Elger, K., Victor, P., & Schemmann, K. (2006).

666 Deformation of the central Andean upper plate system—Facts, fiction, and
667 constraints for plateau models. In *The Andes* (pp. 3-27). Springer Berlin Heidelberg.

668 O'Driscoll, L. J., Humphreys, E. D., & Saucier, F. (2009). Subduction adjacent to deep
669 continental roots: Enhanced negative pressure in the mantle wedge, mountain
670 building and continental motion. *Earth and Planetary Science Letters*, 280(1), 61-70.

671 O'Driscoll, L. J., Richards, M. A., & Humphreys, E. D. (2012). Nazca–South America
672 interactions and the late Eocene–late Oligocene flat-slab episode in the central
673 Andes. *Tectonics*, 31(2).

674

675 Pennington, W. D. (1981). Subduction of the eastern Panama Basin and
676 seismotectonics of northwestern South America. *Journal of Geophysical Research:*
677 *Solid Earth* (1978–2012), 86(B11), 10753-10770.

678

679 Ramos, V. A., Cristallini, E. O., & Pérez, D. J. (2002). The Pampean flat-slab of the
680 Central Andes. *Journal of South American Earth Sciences*, 15(1), 59-78.

681 Ramos, V. A., & Folguera, A. (2009). Andean flat-slab subduction through
682 time. *Geological Society, London, Special Publications*, 327(1), 31-54.

683 Roda, M., Marotta, A. M., & Spalla, M. I. (2011). The effects of the overriding plate
684 thermal state on the slab dip in an ocean-continent subduction system. *Comptes*
685 *Rendus Geoscience*, 343(5), 323-330.

686 Rodríguez-González, J., Negredo, A. M., & Billen, M. I. (2012). The role of the
687 overriding plate thermal state on slab dip variability and on the occurrence of flat
688 subduction. *Geochemistry, Geophysics, Geosystems*, 13(1).
689

690 Rodríguez-González, J., Billen, M. I., & Negredo, A. M. (2014). Non-steady-state
691 subduction and trench-parallel flow induced by overriding plate structure. *Earth
692 and Planetary Science Letters*, 401, 227-235.
693

694 Rosenbaum, G., Giles, D., Saxon, M., Betts, P. G., Weinberg, R. F., & Duboz, C. (2005).
695 Subduction of the Nazca Ridge and the Inca Plateau: Insights into the formation of
696 ore deposits in Peru. *Earth and Planetary Science Letters*, 239(1), 18-32.
697

698 Schellart, W. P., Freeman, J., Stegman, D. R., Moresi, L., & May, D. (2007). Evolution
699 and diversity of subduction zones controlled by slab width. *Nature*, 446(7133), 308-
700 311.

701 Seton, M., Müller, R. D., Zahirovic, S., Gaina, C., Torsvik, T., Shephard, G., ... & Chandler,
702 M. (2012). Global continental and ocean basin reconstructions since 200Ma. *Earth-
703 Science Reviews*, 113(3), 212-270.

704 Skinner, S. M., & Clayton, R. W. (2013). The lack of correlation between flat slabs and
705 bathymetric impactors in South America. *Earth and Planetary Science Letters*, 371,
706 1-5.
707

708 Stadler, G., Gurnis, M., Burstedde, C., Wilcox, L. C., Alisic, L., & Ghattas, O. (2010). The
709 dynamics of plate tectonics and mantle flow: From local to global scales. *science*,
710 329(5995), 1033-1038.

711

712 Stegman, D. R., Freeman, J., Schellart, W. P., Moresi, L., & May, D. (2006). Influence of
713 trench width on subduction hinge retreat rates in 3 - D models of slab rollback.
714 *Geochemistry, Geophysics, Geosystems*, 7(3).

715

716 Stevenson, D. J., & Turner, J. S. (1977). Angle of subduction. *Nature*, 270, 334–336

717 Tackley, P. J., & King, S. D. (2003). Testing the tracer ratio method for modeling
718 active compositional fields in mantle convection simulations. *Geochemistry,*
719 *Geophysics, Geosystems*, 4(4).

720 Tan, E., Choi, E., Thoutireddy, P., Gurnis, M., & Aivazis, M. (2006). GeoFramework:
721 Coupling multiple models of mantle convection within a computational
722 framework. *Geochemistry, Geophysics, Geosystems*, 7(6).

723 Taramón, J. M., Rodríguez - González, J., Negredo, A. M., & Billen, M. I. (2015).
724 Influence of cratonic lithosphere on the formation and evolution of flat slabs:
725 Insights from 3 - D time - dependent modeling. *Geochemistry, Geophysics,*
726 *Geosystems*, in press.

727 Tovish, A., & Gerald, S. (1978). Island arc curvature, velocity of convergence and
728 angle of subduction. *Geophysical Research Letters*, 5(5), 329-332.

van Hunen, J., van den Berg, A. P., & Vlaar, N. J. (2000). A thermo-mechanical model of horizontal subduction below an overriding plate. *Earth and Planetary Science Letters*, 182(2), 157-169.

van Hunen, J., Van Den BERG, A. P., & Vlaar, N. J. (2002). On the role of subducting oceanic plateaus in the development of shallow flat subduction. *Tectonophysics*, 352(3), 317-333.

Yáñez, G., Cembrano, J., Pardo, M., Ranero, C., & Selles, D. (2002). The Challenger–Juan Fernández–Maipo major tectonic transition of the Nazca–Andean subduction system at 33–34 S: geodynamic evidence and implications. *Journal of South American Earth Sciences*, 15(1), 23-38.

Zhong, S., Zuber, M. T., Moresi, L., & Gurnis, M. (2000). Role of temperature-dependent viscosity and surface plates in spherical shell models of mantle convection. *Journal of Geophysical Research: Solid Earth* (1978–2012), 105(B5), 11063-11082.

Figure 1. Geological setting of South America with depth contours of slab 1.0 (Hayes et al., 2012) indicated by thin black lines, subducting oceanic plateaus translucent gray and continental cratons translucent white. The major flat slabs in South America are outlined with thick black lines. The locations of oceanic plateaus, cratons and flat slabs are modified from Gutscher et al. (2000), Loewy et al. (2004) and Ramos and Folguera (2009), respectively. The present-day plate motion is shown as black arrows. Tooth-shaped line represents the South American trench. Seafloor ages to the west of South America are shown with colorful lines with numbers indicating the age in Ma.

753

754 **Figure 2.** Surface velocity boundary conditions and seafloor age distribution at (a) 20 Ma, (b) 10 Ma
755 and (c) present. White arrows show the direction and magnitude of surface velocities. Background
756 color represents the seafloor age. Black regions represent the oceanic plateau and aseismic ridges
757 with dashed contours indicating the already-subducted parts. The South American continent is
758 shown with dark gray with cratons indicated by light gray. (d) Trench profiles at 45 Ma, 15 Ma and
759 present. Red lines indicate the temporally deformable trench profiles (Arriagada et al., 2008), and
760 black lines represent the present-day trench geometry. (e) Background 1D mantle viscosity structure.
761 Key features include a low-viscosity asthenosphere and a viscosity maximum in the upper part of the
762 lower mantle.

763

764 **Figure 3.** Comparison of present-day slab geometry from five different subduction models. (a-e)
765 Predicted slab geometry (blue color) from models I, II, III, IV and V (see main text), respectively.
766 Different rows show different depths with the first row at 100 km, the second at 150 km, the third at
767 250 km, and the fourth at 500 km depth. Green lines represent the interpolated Benioff zones from
768 *slab 1.0* (Hayes et al., 2012). Present-day configuration of the cratons and buoyancy features are
769 shown with brown and blue, respectively, where we assume no lateral distortion in the geometry of
770 the buoyancy features after subduction. Dotted lines outline the overriding plate. Note the
771 progressive flattening of the slab from model I to model V.

772

773 **Figure 4.** Along-trench slab dip angle distribution (solid lines) from models I, II, and S and the
774 present-day seafloor age distribution at trench (dashed line) in model II. Note the similar seafloor
775 age distributions from 10 Ma to the present, the period relevant to the modern flat slab formation.
776 Slab dip angle is measured according to the horizontal distance between the trench and the slab
777 surface at 200 km depth along the east-west direction. Slab surface is defined as the upper interface
778 between the slab and the ambient mantle, where the interface is the isotherm of 300 °C cooler than

the ambient mantle.

Figure 5. (a) Present-day slab dip angle distribution along South American trench in models I, II, III, IV and V. The locations of subducting buoyancy features are marked with light blue, while the location of continental cratons are shown with dashed brackets. The dotted lines mark the location of the cross sections shown in Fig. 6. (b) Slab upper surface (300 °C cooler than the ambient mantle) at 200 km depth for the five models. Dashed lines outline the cratons, and light blue represents the buoyancy features.

Figure 6. Comparison of the slab upper surface of models I, II, III, IV and V at (a) 6°S, (b) 11°S, (c) 19°S and (d) 29°S, respectively, corresponding to the dashed lines in Fig. 5a. Notice the progressive flattening of the slabs across different models.

Figure 7. Maps showing the dynamic pressure distribution at 140 km depth for models (a) I, (b) II, (c) III, (d) IV and (e) V. The mantle flow field is over-plotted as arrows. Thin black contours indicate slab isotherms with an increment of 100 °C. The dynamic pressure in mantle wedge is progressively lowered from model I to model IV.

Figure 8. Cross sections showing the dynamic pressure distribution at 15°S for model (a) I, (b) II, (c) III, (d) IV and (e) V. Arrows represent mantle velocities. Yellow thick lines outline the overriding plate. Solid green contours indicate slab isotherms with an increment of 100 °C. The dynamic pressure within the mantle wedge is progressively lowered from model I to model IV, indicating an increasing hydrodynamic suction. Notice that the mantle flow above the slab is mainly concentrated in the asthenosphere when a high-viscosity overriding plate is present.

Figure 9. The evolution of slab geometry beneath Peru from 14 Ma to present day. Black lines represent the slab upper surface (300 °C cooler than the ambient mantle) at different depths. Dark gray patterns show the reconstructed locations of the Inca Plateau and the Nazca Ridge. Light gray

area represents now-subducted seafloors. Red triangles indicate the location of extrusive volcanism, while orange triangles represent that of intrusive volcanism. A slab hole forms since 6 Ma during the Nazca Ridge subduction. Note the delay between initial buoyancy feature subduction and flat slab formation.

Figure 10. Cross sections of the best-fit model from 5° to 30°S at an interval of 5°. Orange arrows mark the location of these cross sections. In each cross section, background color represents the temperature field with the yellow lines indicating the interpolated Benioff zone from *slab 1.0* (Hayes et al., 2012). Gray circles represent the locations of earthquakes with magnitude > 4.0 from IRIS earthquake catalog for years from 1970 to 2015. Black lines above each cross section delineate the topography, with the vertical scale amplified by 20 times. Note the overall match of the slab geometry to both individual seismicity and *slab 1.0 contour*.

817

818

819

820

821

822

823

824

825

826

827

828

829

849

850

851

852

853

854

855

856

857

858

859

860

861

862