Influence of subduction history on South American topography

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Abstract

The Cenozoic evolution of South American topography is marked by episodes of large-scale uplift and subsidence not readily explained by lithospheric deformation. The drying up of the inland Pebas system, the drainage reversal of the Amazon river, the uplift of the Sierras Pampeanas and the uplift of Patagonia have all been linked to the evolution of mantle flow since the Miocene in separate studies. Here we investigate the evolution of long-wavelength South American topography as a function of subduction history in a time-dependent global geodynamic model. This model is shown to be consistent with these inferred changes, as well as with the

migration of the Chaco foreland basin depocentre, that we partly attribute to the inboard migration of subduction resulting from Andean mountain building. We suggest that the history of subduction along South America has had an important influence on the evolution of the topography of the continent because time-dependent mantle flow models are consistent with the history of vertical motions as constrained by the geological record at four distant areas over a whole continent. Testing alternative subduction scenarios reveals flat slab segments are necessary to reconcile inferred Miocene shorelines with a simple model paleogeography. As recently suggested, we find that the flattening of a subduction zone results in dynamic uplift between the leading edge of the flat slab segment and the trench, and in a wave of dynamic subsidence associated with the inboard migration of the leading edge of flat subduction. For example, the flattening of the Peruvian subduction contributed to the demise of Pebas shallow-water sedimentation, while continental-scale tilting also contributed to the drainage reversal of the Amazon River. The best correlation to P-wave tomography models for the Peruvian flat slab segment is obtained for a case when the asthenosphere, here considered to be 150 km thick and 10 times less viscous than the upper mantle, is restricted to the oceanic domain.

1. Introduction

Mantle flow influences the long-wavelength topography of continents (Gurnis, 1993), but progress has been relatively slow in quantifying the time-dependence of dynamic topography. South America is an ideal continent to investigate the evolution of subduction-driven long-wavelength dynamic topography (e.g. Lithgow-Bertelloni and Gurnis, 1997; Flament et al., 2013) because subduction has been continuous along its

western margin at least since the Jurassic (Seton et al., 2012) and the last plume-related large igneous province to be emplaced regionally was the Parana-Etendeka at ~ 132 Ma (Bryan and Ferrari, 2013). Several studies have investigated the influence of changes in subduction characteristic on the evolution of the topography of South America since the Miocene (Guillaume et al., 2009; Dávila et al., 2010; Shephard et al., 2010; Dávila and Lithgow-Bertelloni, 2013; Eakin et al., 2014; Dávila and Lithgow-Bertelloni, 2015). Changes in the geometry of subduction have been considered in series of instantaneous mantle flow models. The uplift of Patagonia (Guillaume et al., 2009; Pedoja et al., 2011; Fig. 1a) has been linked to the opening of a slab window under Patagonia, where an active spreading center has been intersecting the Chilean trench since ~ 14 Ma (Fig. 1b). Changes in slab dip angle, particularly the flattening of the South American slab under Chile and Peru since the late Miocene (Fig. 1b; Kay and Mpodozis, 2001; Ramos and Folguera, 2009), are proposed to explain the uplift of the Sierras Pampeanas (Dávila et al., 2010; Dávila and Lithgow-Bertelloni, 2013; Dávila and Lithgow-Bertelloni, 2015; Fig. 1a), and the uplift of northwestern South America that ended Pebas shallow-water sedimentation (Wesselingh et al., 2010; Eakin et al., 2014; Fig. 1a). In addition, a time-dependent inverse model of mantle flow predicted a dynamic tilting to the east of northern South America during the Miocene, which is proposed to have resulted in the drainage reversal of the Amazon River (Shephard et al., 2010). Other features of South American paleogeography may be associated with mantle flow. For instance, the eastward migration of the depocentre of the Chaco foreland basin (Fig. 1a) and associated change in drainage direction (Uba et al., 2006) is reminiscent on a smaller scale of the migration of the depocentre of the Western Interior Basin of the United States, attributed to subduction-driven dynamic subsidence (Liu et al., 2011). Here we

investigate the influence of subduction on the evolution of the large-scale topography of the South American continent in a series of time-dependent global forward mantle flow models that progressively assimilate tectonic reconstructions. We investigate the influence of flat slab subduction (Fig. 1b) and of the asthenosphere across four model cases.

2. Modelling the evolution of South American dynamic topography in global time-dependent mantle flow models

2.1 Mantle flow models assimilating tectonic reconstructions with flat slab subduction

We solve the convection equations with the finite-element method in a spherical domain using *CitcomS* (Zhong et al., 2008), modified to progressively assimilate surface plate velocities, the thermal structure of the lithosphere and the shallow thermal structure of subducting slabs (Bower et al., 2015), including flat slab segments (Figs 3 and 8 of Bower et al., 2015). The method of progressive data assimilation is fully described in Bower et al. (2015), and similar models are fully described in Flament et al. (2014), where governing equations are also given. Here we only briefly describe the assimilation method and models.

Surface plate velocities, the age of the ocean floor and the location and polarity of subduction zones are determined in one million year intervals from global plate tectonic reconstructions (e.g., Seton et al., 2012) with continuously closing plates (Gurnis et al., 2012) and deforming plates along the margins of South America (as in Arriagada et al., 2008; Flament et al., 2014). The tectonothermal ages of the continental lithosphere, assimilated in the models, were modified from Artemieva (2006) so that Proterozoic lithosphere is at least 100 km away from the margin of

98 Andean deformation (Arriagada et al., 2008; grey outline in Fig. 1b) and includes the
99 Guyana Shield (Fig. 1b), which displays a small elastic thickness (Tassara et al.,
100 2007).

101 Tectonic reconstructions including the position of present-day and past flat slab
102 segments are implemented using *GPlates* (Boyden et al., 2011), using global present-
103 day subduction zone geometries (Hayes et al., 2012) and geological constraints
104 (Ramos and Folguera, 2009) to define the outline of the ongoing Peruvian and
105 Pampean flat slab subduction. The evolution of the extent of flat slab segments is
106 based on (i) previous work (Ramos and Folguera, 2009), (ii) requiring that the leading
107 edge of the flat slab segment never falls landward of arc-related volcanics (Trumbull
108 et al., 2006) while accounting for deformation (Arriagada et al., 2008) and (iii) the
109 velocity of the subducting plate. With these constraints, we implement flat slab
110 subduction in the Payenia (11-5 Ma), Pampean (11-0 Ma) and Peruvian (9-0 Ma)
111 segments (Fig. 1b) forming slightly (1-2 Myr) later, but within uncertainty of the ages
112 proposed by Ramos and Folguera (2009).

113 In order to isolate the effect of flat slab subduction on dynamic topography, flat slab
114 segments are considered in all cases except for case 2 (Table 1) in which the
115 subduction dip angle is maintained to 45° until the present day. To investigate the
116 uncertain past configuration of flat slab segments, we consider cases in which
117 subduction progressively flattens over a few million years (cases 1 and 3), and a case
118 in which subduction becomes flat as soon as the flat slab segment appears (case 4).

119 The initial condition, derived from the tectonic reconstruction, is defined at 230 Ma to
120 ensure that the models have reached a dynamic equilibrium (which takes ~50 Myr;
121 Flament et al., 2014) for the period of interest (50-0 Ma). The initial condition

includes a basal thermochemical layer 113 km thick just above the core-mantle boundary (CMB) that consists of material 3.6% denser than ambient mantle, which is enough to prevent the formation of mantle plumes in the model and allows subduction-driven topography to be isolated. The temperature and thickness of the lithosphere is derived using a half-space cooling model and the synthetic age of the ocean floor (Bower et al., 2015). The thickness of the thermochemical continental lithosphere is derived from simplified tectonothermal ages: Archean lithosphere is assumed to be 250 km thick, Proterozoic lithosphere 165 km thick, and Phanerozoic lithosphere 135 km thick (as in Flament et al., 2014). The global thermal structure of slabs is built based on the location of subduction zones and on the age of the ocean floor (Bower et al., 2015). In the initial condition, slabs are inserted down to 1,400 km depth, with a dip of 45° down to 425 km depth and a dip of 90° below 425 km. Subduction zones thought to have initiated too little time prior 230 Ma to be 1,400 km deep are inserted to a depth based on subduction duration and assuming a descent rate of 3 cm/yr in the upper mantle and 1.2 cm/yr in the lower mantle (van der Meer et al., 2010). Slabs are initially twice as thick in the lower mantle than in the upper mantle to account for the advective thickening observed in tests in which subduction zones are only inserted to the upper mantle in the initial condition. Subduction zones that appear during the model run are progressively inserted in the upper mantle. The global thermal structure of the lithosphere and of subducting slabs is assimilated in the dynamic models in 1 Myr increments, to a depth of up to 350 km depth at subduction zone (Bower et al., 2015).

The Rayleigh number that determines the vigour of convection is defined by

$$Ra = \frac{\alpha_0 \rho_0 g_0 \Delta T h_M^3}{\kappa_0 \eta_0},$$

146 where α is the coefficient of thermal expansivity, ρ the density, g the acceleration of
 147 the gravity field, ΔT the temperature change across the mantle, κ the thermal
 148 diffusivity, η the viscosity, and the subscript “0” indicates reference values. Values
 149 are listed in Table 2.

150 As in Flament et al. (2014), viscosity varies by 1,000 due to temperature-dependence
 151 following

$$152 \quad \eta = \eta_0(r) (1 + \eta_C C) \exp\left(\frac{E_\eta}{T + T_\eta} - \frac{E_\eta}{T_b + T_\eta}\right),$$

153 where $\eta_0(r)$ is a pre-factor defined with respect to the reference viscosity η_0 for four
 154 layers: it is equal to 1 above 160 km, either to 1 (for case 4 without asthenosphere, see
 155 Table 1) or to 0.1 (for cases 1-3 with asthenosphere, see Table 1) between 160–
 156 310 km depth, to 1 between 310–660 km depth and to 100 in the lower mantle, below
 157 660 km depth. C is the composition field, and η_C is the compositional viscosity pre-
 158 factor equal to 100 in the continental lithosphere and to 10 in the continental
 159 asthenosphere (for cases 1 and 2 in which the asthenosphere is restricted to the
 160 oceanic domain). E_η is the activation energy (E_{UM} in the upper mantle and E_{LM} in the
 161 lower mantle), T is the temperature, T_η is a temperature offset, and T_b is the ambient
 162 mantle temperature (values are listed in Table 2).

163 In cases 1 and 2, in addition to the buoyant and viscous tracers that define the
 164 continental lithosphere (see Flament et al., 2014), we use neutrally buoyant, viscous
 165 tracers between 160 and 310 km depth under the continents to ensure that the
 166 asthenosphere is only present under the oceans. The asthenosphere extends globally in
 167 case 3, and it is absent in case 4 (Table 2). The average model resolution, obtained
 168 with $\sim 13 \times 10^6$ nodes and radial mesh refinement, is $\sim 50 \times 50 \times 15$ km at the surface,
 169 $\sim 28 \times 28 \times 27$ km at the core–mantle boundary (CMB), and $\sim 40 \times 40 \times 100$ km in

the mid-mantle. Model parameters that are common to all cases are listed in Table 2,
and further model details are available in Flament et al. (2014).

2.2 Computing dynamic topography

There has been significant recent progress in measuring the amplitude of residual topography in the oceanic domain (Winterbourne et al., 2014), but constraining the amplitude of residual topography in the continental domain remains a challenge (Flament et al., 2013). The amplitude of dynamic topography predicted by different flow models also vary widely depending on model set up and boundary conditions (Thoraval and Richards, 1997; Flament et al., 2013). These challenges in constraining the amplitude of residual and dynamic topography make it important to compare time-dependent model predictions to time-dependent geological constraints. Here, we aim to obtain dynamic topography amplitudes of ± 1 km, in line with large-scale compilations of estimates of residual topography carried out in the oceans (Winterbourne et al., 2014). This is achieved by computing dynamic topography using free-slip boundary conditions and ignoring buoyancy and lateral viscosity variations above 350 km depth, which is the depth to which tectonic reconstructions are assimilated in the models. The dynamic topography h is obtained by scaling the radial stress σ'_{rr} on the top model surface following

$$h = \frac{1}{\Delta\rho g_0} \frac{\eta_0 \kappa_0}{R_0^2} \sigma'_{rr},$$

where $\Delta\rho$ is the density difference between the shallow mantle $\rho_{\text{UM}} = 3340 \text{ kg m}^{-3}$ and either air ($\rho_a = 0 \text{ kg m}^{-3}$) or sediment ($\rho_s = 1950 \text{ kg m}^{-3}$), R_0 is the radius of the Earth and other parameters are defined in Section 2.1. The resulting dynamic topography

globally varies with time between -1,300 m and +1,030 m for case 4, and between -1,250 m and +910 m for case 3, in which the low-viscosity asthenosphere reduces the peak-to-peak amplitude of predicted air-loaded dynamic topography by ~ 200 m. The predicted dynamic topography is air-loaded in subaerial areas such as the Central Patagonian basin (Figs 5d, 6d and 7d), and sediment loaded in areas in which the sedimentary record indicates significant shallow-water environments or detrital deposition over time, such as the Pebas system (Figs 5a, 6a and 7a), the broken foreland basin of the Sierras Pampeanas (Figs 5b, 6b and 7b) and the Chaco foreland basin (Figs 5c, 6c and 7c). With the assumed upper mantle and sediment densities (see above), sediment-loaded dynamic topography has amplitude $(\rho_{UM} - \rho_a)/(\rho_{UM} - \rho_s) = 2.4$ times greater than air-loaded dynamic topography.

3. Effect of the viscosity of the asthenosphere on mantle structure across the Peruvian flat slab segment

We compare the predicted mantle structure along a cross-section across the Peruvian flat slab segment (dashed black line in Fig. 2d) to two P-wave mantle tomography models (Montelli et al., 2006; Li et al., 2008) in Figure 3e-f, because mantle tomography is not used in the mantle flow models. In the mantle flow models, we consider mantle 10% colder than ambient to be representative of subducted slabs. Taking flat slab subduction into account clearly improves the agreement between model mantle temperature field and tomography, as seen by the poor match between case 2 (no flat slab segments; green outlines in Figs 3e-f) and tomography 72-78°W and 200-550 km depth. In comparison, cases 1, 3 and 4 (with flat slab segments) successfully predict the subducting slab to be offset to the east and centred on $\sim 73^\circ\text{W}$

at ~250 km depth and on ~72°W at 350 km depth. The position of the subducting slab below 500 km depth depends on the viscosity of the asthenosphere. Without asthenosphere (case 4, cyan contour in Figs 3e-f), the western edge of the predicted subducted slab is shifted ~2° (~220 km) west of the fast seismic velocity anomalies imaged by both tomography models between 500 and 1,300 km depth (Figs 3e-f). Considering a global low-viscosity asthenosphere (case 3, yellow contours in Figs 3e-f) reduces this offset by ~0.8° (~90 km), and limiting the asthenosphere to the oceanic domain (case 1, magenta contours in Figs 3e-f), essentially aligns the western edge of the model slab with fast seismic velocity anomalies (Figs 3e-f). This eastward shift of the predicted slab when the asthenosphere is restricted to the oceanic domain is explained by the reduction of the net rotation induced to the lower mantle with this viscosity structure (Rudolph and Zhong, 2014). Such a viscosity structure is consistent with the findings of Ricard et al. (1991) that the viscosity of the asthenosphere must be an order of magnitude lower under the oceans than under the continents to match the present day net-rotation of the lithosphere in instantaneous global mantle flow models with lateral viscosity variations. It is also consistent with the findings of Zahirovic et al. (2015) that tectonic plates consisting of more than 50% continental lithosphere have moved slower than their oceanic counterparts over the last 200 Ma.

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4. Influence of subduction on South American topography

4.1 Contribution of subduction-driven mantle flow to the Miocene geography of South America

Over the last 50 Myr, the westward motion of South America over the sinking slab produced by continuous subduction along its western convergent margin has resulted in a broadening of negative dynamic topography towards the east (Flament et al., 2014; Fig. 2). Space-time variations in low topography along the convergent margin (Fig. 2) primarily reflect variations in the age of the subducting lithosphere (e.g. Seton et al., 2012) that is progressively assimilated in the model (Bower et al., 2015). In addition, flat slab subduction results in dynamic subsidence at the leading edge of the flat slab segment, where dynamic topography becomes more negative, and in dynamic uplift between the trench and the leading edge of the flat slab segment, where dynamic topography becomes less negative (Figs 2d, 3a-d, 5a-b, 6a-b and 7a-b; Dávila and Lithgow-Bertelloni, 2013; Eakin et al., 2014; Dávila and Lithgow-Bertelloni, 2015).

To investigate the implication of mantle flow for South American paleogeography, we remove for each case the predicted present-day dynamic topography from observed topography (Amante and Eakins, 2009) and add the contributions of dynamic topography predicted for each case and long-term sea level change with respect to present-day (Haq et al., 1987) back in time (Fig. 4). This model paleogeography is intended to illustrate long-wavelength shallow-water environments associated with mantle flow rather than to constitute an accurate and complete evolution of paleotopography. It does not account for lithospheric deformation and flexure, erosion and sedimentation, and it is at best determined within the relatively

large uncertainty of long-term sea level change (e.g. Spasojevic and Gurnis, 2012). We therefore limit its interpretation to a visual comparison in Figure 4 with two global sets of paleoshorelines (Smith et al., 2004; Golonka, 2009) and a more detailed set of paleoshorelines for northern South America (Wesselingh et al., 2010). The left-hand-side panels of Figure 4, in which only long-term sea level changes are considered, illustrate that the shallow-water deposits inland of the Rio de la Plata and inland of the Colorado Basin, present in both sets of global paleoshorelines between 21 Ma and 10 Ma, occur primarily as a consequence of long-term sea level change.

In contrast, in northern South America, considering variations in long-term sea level change alone results in shallow-water environments centred on the equator at 60°W (Fig. 4a-b-c), whereas also considering variations in dynamic topography results in shallow-water environments centred approximately on 75°W/5°S (Fig. 4d-e-f), a location that is compatible with paleo-environments (Wesselingh et al., 2010). This is because most of the Amazon river was dynamically more elevated than at present whereas the Pebas system, which falls above the Peruvian flat slab, was dynamically less elevated than at present (Fig. 4g-h-i). The predicted shallow-water environments are also broadly consistent with paleoshorelines along the Amazon River ~10 Ma, and with the extent of the Pebas system at ~10 Ma and ~16 Ma (Fig. 4; Wesselingh et al., 2010). The predicted extent of shallow-water environments in the Pebas system back in time is due to the dynamic uplift associated with the flattening of the subduction zone forward in time (Figs 1c-d, 4g-h-i, and 5a-d), as confirmed by the difference in shallow-water environment patterns obtained for case 2 (no flat slab segments) and case 3 (with flat slab segments; Supplementary Fig. 1). The similarities in the extent of Pebas shallow-water environments at 21 Ma (Fig. 4d) and 16 Ma (Fig. 4e) reflect that dynamic topography did not change much between these times for this area (Fig.

4g-h), and points to the limit of this simple model paleogeography based on the difference between present-day and past dynamic topography. In addition, the model does not fully capture the narrow strip of shallow-water sedimentation along the Andes that extends north of the equator between ~21-16 Ma (Fig. 4 and Supplementary Fig. 1), suggesting that the combined effects of mantle flow and flexure are required to explain the Miocene shallow-water depositional environments of north western South America (Sacek, 2014).

In contrast to flat slab subduction (Eakin et al., 2014), the demise of the Pebas system had previously been associated to the reversal of the drainage direction of the Amazon River and attributed to dynamic tilting of northern South America towards the east since 45 Ma (Shephard et al., 2010). Indeed, our model predicts a dynamic tilt to the east of the Amazon River since 50 Ma (Figs 2, 5a, 6a and 7a). Case 1 predicts tilting by $\leq 1,140$ m since 30 Ma, with up to 780 m of cumulative uplift at 77.3°W and up to 360 m of cumulative subsidence at 51.4°W (Fig. 7a). Without flat slab, case 2 predicts tilting by ≤ 745 m since 30 Ma, with up to 340 m of cumulative uplift at 78.7°W and up to 405 m of cumulative subsidence at 52.5°W (Supplementary Fig. 2e). This continental tilting down to the east is consistent with the findings of Shephard et al. (2010), whereas flat slab subduction results in a wave of dynamic topography at shorter wavelength, characterised by dynamic uplift between the trench and the leading edge of the flat slab segment, and dynamic subsidence at the leading edge of the flat slab (Figs 5a, 6a, 7a and Eakin et al., 2014). Case 1 predicts up to ~ 80 m/Myr of dynamic uplift during slab flattening at 7 Ma (Fig. 5a), whereas uplift rates would not exceed 15 m/Myr in the absence of slab flattening (case 2, Supplementary Fig. 2c). Together, these results suggest that a combination of flat slab subduction and continental-scale mantle-plate interactions played a role in shaping the large-scale

topography of northern South America since the Miocene, in addition to tectonic deformation (e.g. Arriagada et al., 2008; Flament et al., 2014) and flexural loading (e.g. Dávila and Lithgow-Bertelloni, 2013; Eakin et al., 2014; Sacek, 2014) at shorter wavelengths (< 500 km).

4.2 Dynamic uplift of the Sierras Pampeanas due to Pampean flat slab subduction

Similarly to the Peruvian flat slab segment, the formation of the Pampean flat slab segment (Fig. 1b) is expected to have had an influence on topography (Dávila et al., 2010; Dávila and Lithgow-Bertelloni, 2013; Dávila and Lithgow-Bertelloni, 2015). Dynamic uplift between the trench and the leading edge of the flat slab is suggested according to one estimate of large positive residual topography for the elevated Sierras Pampeanas (Dávila and Lithgow-Bertelloni, 2013; Dávila and Lithgow-Bertelloni, 2015). This positive residual topography at 32°S has been proposed to be the result of the flattening of the Pampean slab, based on the difference in the dynamic topography given by two instantaneous mantle flow models at 30 Ma and present-day, respectively without and with flat slab subduction (Dávila and Lithgow-Bertelloni, 2013; Dávila and Lithgow-Bertelloni, 2015). Our time-dependent mantle flow models are consistent with these findings, as they predict up to ~ 580 m (at 69.5°W , Fig. 7b) of cumulative uplift since 30 Ma associated with slab flattening at 32°S, with uplift rates up to 53 m/Myr at 7 Ma (Fig 6b). In the absence of subduction flattening, this uplift would be limited to ~ 365 m (at 70.6°W , Supplementary Fig. 2f), and uplift rates would not exceed 18 m/Myr at 7 Ma (Supplementary Fig. 2d). With the implemented scenario of flat subduction (Fig. 1b), most of the uplift in the Sierras Pampeanas is predicted to have occurred since ~ 11 Ma (Figs 6b and 7b). The

predicted amplitude of cumulative uplift since 30 Ma at 32°S (~ 580 m, Fig. 7b) although approximately 30% smaller than that predicted by a previous model (900 m; Dávila and Lithgow-Bertelloni, 2013), falls within the range of residual topography estimates for the area (between ~500 m and ~2.2km; Dávila and Lithgow-Bertelloni, 2013; Dávila and Lithgow-Bertelloni, 2015).

4.3 Migration of the depocentre of the Chaco foreland basin due to trench migration

In addition to slab flattening, lithospheric shortening can also modify the location of the subduction zone with respect to the over-riding plate, and have a significant effect on the evolution of dynamic topography. In South America, this process is best illustrated in the Bolivian orocline that has undergone the most deformation since the Eocene (Arriagada et al., 2008). This deformation is explicitly considered in the models (see Flament et al., 2014), which is reflected by the changing shape of the subduction zone along the Andes in Figures 1b, 2 and 4. In the Central Andes, the depocentre of the Chaco foreland basin (Fig. 1a) has migrated eastward by ~ 220 km (~2.2° of longitude) since ~ 25 Ma (Uba et al., 2006; Figs 5c, 6c and 7c), and river drainage from the east prior ~ 25 Ma was replaced by river drainage from the west by 10 Ma (Uba et al., 2006). Indeed, our results for instantaneous dynamic topography show an eastward migration of the dynamic topography low associated with the subduction zone of ~ 320 km since 30 Ma (Fig. 5c). The increasingly negative dynamic topography between 50 Ma and ~ 20 Ma (Fig. 5c) reflects the increasing age and thickness of the subducting oceanic lithosphere (Seton et al., 2012). The most negative dynamic topography is between ~300 km and ~ 400 km to the west of the depocentre through time (Fig. 5c). The evolution of the dynamic subsidence imposed

by the inboard migration of the subduction zone can be seen in the evolution of the rate of change of dynamic topography (Fig. 6c), with dynamic subsidence being consistent in time and space with the observed migration of the depocentre of the Chaco foreland basin (Fig. 6c). The model depocentre migrates from ~64.8°W at 25 Ma to ~60.7°W at 2 Ma (Fig. 6c), while the depocentre reconstructed from the rock record migrates from ~65.3°W at 25 Ma to ~63°W at present (Uba et al., 2006; Fig. 6c). In addition, the predicted pattern of large-scale dynamic subsidence is consistent with river drainage from the east (coloured arrows in Fig. 6c) from 50 Ma that progressively wanes between 18 Ma and 10 Ma. This predicted waning of river drainage from the east is consistent with the observations Uba et al. (2006). Collectively, these results suggest that in addition to crustal shortening and flexural processes likely at play in the crust and lithosphere (Uba et al., 2006), subduction-driven dynamic subsidence influenced the post-Oligocene history of sedimentation in the Chaco basin. The predicted cumulative dynamic subsidence is ≤ 460 m since 30 Ma. Since we have sought to minimise the amplitude of dynamic topography in this study (see Section 2.2), we infer that subduction-driven dynamic subsidence accounts for at least 8.5% of the 5.4 km thick Neogene strata deposited in the Chaco foreland basin (Uba et al., 2006).

The eastward migration of the slab relative to the South American continent is not only accompanied by that of the Chaco depocentre but also by a decreasing subsidence underneath the Andean belt. Indeed, the model predicts an eastward migrating uplift between 65°W and 70°W between 20 Ma and 0 Ma (Figs 6c and 7c) that could explain part of the Late Miocene uplift of the Eastern Cordillera and Bolivian Altiplano (e.g. Garzzone et al., 2006).

4.4 *Dynamic uplift of Patagonia basin due to the opening of a slab window*

The geomorphology of Patagonia revealed that (i) Pleistocene marine terraces are tilted down to the north (Pedoja et al., 2011) and (ii) river terraces north of the present-day latitude (46.5°S) of the Chile Triple Junction (CTJ) have been tilted down to the north since the Miocene, whereas river terraces south of 46.5°S were tilted down to the north until the late Miocene, and down to the south since then (Guillaume et al., 2009; pictograms in Figs 6d and 7d). These vertical motions were attributed to the northward migration of the intersection of the Chile Ridge and the South American subduction zone based on a piecewise series of instantaneous Stokes flow models (Guillaume et al., 2009). Our time-dependent model confirms these results with dynamic topography tilted down to the north (northward) at 15 Ma then southward south of 47.3°S at 11 Ma. This reversal progressively migrated to the north to reach 46.6°S at present (Fig. 5d). Uplift rates show a maximum of ~18 m/Myr at 46.4°S, with southward tilting south of 46.4°S and northward tilting to the north of 46.4°S (Fig. 6d). The maximum uplift rate is predicted to have remained between 47.1°S and 45.4°S since 13 Ma, and tilting is northward south of 46.5°S before 13 Ma (Fig. 6d). The cumulative dynamic topography since 30 Ma shows tilting mainly down to the north between 30 and 15 Ma, followed by northward migration of a reversal of tilting down to the south southward of 47.9°W at 9 Ma and southward of 46.7°W at present (Fig. 7d). Consistency between these model results and the tilt recorded by river terraces (Guillaume et al., 2009) gives us confidence in the history of vertical motions they describe for the area.

5. Implications for the geodynamic interpretation of paleogeographic reconstructions and paleotopography

Our results for the paleogeography of South America since the Miocene (Fig. 4) confirm that large-scale (wavelength $> 400\text{-}500$ km) episodes of uplift and subsidence observed in the geological record over a few million years may result from the interplay between long-term sea level change and large-scale mantle flow (Gurnis, 1993; Spasojevic and Gurnis, 2012), in addition to shorter-wavelength (< 500 km) processes including tectonic deformation (e.g. Arriagada et al., 2008; Flament et al., 2014) and flexural loading (e.g. Dávila and Lithgow-Bertelloni, 2013; Eakin et al., 2014; Sacek, 2014).

The formation of a slab window, where an active spreading centre intersects a subduction zone, is expected to result in several hundred meters of dynamic uplift that is of greatest amplitude in a direction parallel to the ridge axis and propagates up to several hundred kilometers inland from the trench (Guillaume et al., 2009 and Figs 2, 5d, 6d and 7d). Episodes of dynamic uplift should have occurred as a result of proposed Cenozoic slab windows along western North America (Madsen et al., 2006) and East Asia (Seton et al., 2015). The geometry and duration of these transient uplift episodes is expected to depend on the convergence velocity between the subducting and overriding plate, and on the intersection angle between the mid-oceanic ridge and the subduction zone. In addition to the evolution of magmatic history (Madsen et al., 2006), geological indicators of evolving large-scale topography and drainage system may help reconstructing the evolution of ancient slab windows (Madsen et al., 2006; Seton et al., 2015).

433 Here, we have focused on the evolution of mantle-driven topography associated with
434 subduction, ignoring sources of buoyancy above 350 km depth (Fig. 4). This choice
435 results in smaller amplitudes of predicted dynamic topography, since shallower
436 sources of buoyancy have a larger contribution to surface topography (e.g. Flament et
437 al., 2013 and references therein). The buoyancy associated with slabs lying flat at
438 ~100 km depth is not considered in the calculation of dynamic topography (Fig. 3a-d),
439 which is consistent with previous studies in which the buoyancy of flat slabs segment
440 was assumed to be zero (Dávila and Lithgow-Bertelloni, 2013; Eakin et al., 2014;
441 Dávila and Lithgow-Bertelloni, 2015). Including shallower sources of buoyancy in
442 our calculation of dynamic topography introduces higher frequency variations
443 associated with density variations in the thermal lithosphere, which is assimilated in
444 the model down to 250 km depth for Archean provinces. This shorter wavelength
445 topography is not easily reconciled with the geological record and interferes with the
446 subduction-driven long-wavelength topography. On the other hand, taking shallower
447 sources of buoyancy into account would increase the contribution of dynamic
448 subsidence to the total subsidence of the Chaco depocentre, which is ~10% when only
449 considering buoyancy sources deeper than 350 km. Accounting for the uppermost
450 mantle in the computation of dynamic topography might also reconcile the timing and
451 location of the model uplift of the Eastern Cordillera and Bolivian Altiplano with
452 observations (e.g. Garzzone et al., 2006). As for flat slab segments, at lithospheric
453 depths, the flattening of a subduction zone is also expected to result in lithospheric
454 shortening and associated isostatic topography because of increased contact and
455 coupling between the subducting and overriding plate (Bird, 1988; Martinod et al.,
456 2010). Although a definitive link between the subduction of oceanic plateaus and the
457 flattening of subduction zones remains to be established (Gutscher et al., 2000; van

Hunen et al., 2002; Skinner and Clayton, 2013), flat slab subduction under the Central Andes (Kay and Mpodozis, 2001) is proposed to have occurred following the southward migration of the subduction of the Juan Fernandez Ridge.

Our results confirm the importance of the history of subduction in shaping the foreland of mountain belts (Husson et al., 2014) > 200 km away from the tectonic load, where the effect of elastic flexure becomes negligible for the Andes (e.g. Dávila and Lithgow-Bertelloni, 2013; Eakin et al., 2014). Furthermore, they underline that in addition to the topography associated with isostatic compensation (Bird, 1988) and to the dynamic subsidence associated with the inboard migration of the leading edge of the flat slab segment (Liu et al., 2011; Eakin et al., 2014), the flattening of a subduction zone is expected to result in dynamic uplift between the leading edge of the flat slab segment and the trench (Dávila and Lithgow-Bertelloni, 2013; Eakin et al., 2014; Dávila and Lithgow-Bertelloni, 2015).

6. Conclusions

We have presented a series of time-dependent global mantle flow models assimilating tectonic reconstructions that are compatible with mantle tomography and with the long-wavelength paleogeography of South America as constrained by the geological record. The best match between model temperature field and P-wave tomography for the Peruvian flat slab is obtained when considering a low-viscosity asthenosphere restricted to the oceanic domain, which is consistent with previous studies (Ricard et al., 1991; Zahirovic et al., 2015).

The agreement between the evolution of topography predicted by our time-dependent mantle flow models and that constrained by the geological record in several distant areas over a whole continent (this study and Flament et al., 2014), highlights the important role of subduction history in shaping Earth's surface. We find that long-wavelength surface topography depends on subduction history in several ways. The flattening of a subduction zone results in dynamic uplift above the flat slab (Dávila and Lithgow-Bertelloni, 2015), and in dynamic subsidence at the leading edge of the flat slab. As a consequence, flat slab subduction must be taken into account, as well as long-term sea level change, to match the Miocene shallow-water deposition history of the continent. Flat subduction and continental-scale tilting have both shaped northern South America since the Miocene, characterized by the demise of Pebas shallow-water deposition and drainage reversal of the Amazon River. The relative motion of the trench with respect to the over-riding plate contributes to the migration of foreland basins in cordilleras, as shown to be the case for the Chaco foreland basin. The formation of a slab window results in several hundred meters of uplift of greatest amplitude parallel to the ridge axis and propagating up to several hundred kilometers inland from the trench, as shown here for Patagonia.

Acknowledgements N.F. and R.D.M. were supported by Statoil ASA through ARC IH130200012, M.G. and D.J.B. were supported by Statoil ASA and by the NSF (through EAR-1161046 and EAR-1247022). This research was undertaken with the assistance of resources from the National Computational Infrastructure (NCI), which is supported by the Australian Government. We thank Simon Williams for comments on an early version of the manuscript, and F. Davila and an anonymous reviewer for constructive comments that improved the manuscript. The evolution of the dynamic

topography predicted by all four cases for South America since 50 Ma is available at
<http://www.earthbyte.org/resources.html>.

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Figure captions

Figure 1: a/ Topography of South America, geological and geographical features referred to in this study, plate boundaries (white lines) and flat slab segments (triangles indicate the overriding plate for subduction zones). The extent of the Pebas system is shown at 16 Ma for the paleogeography of Wesselingh et al. (2010), the extent of Neogene strata in the Chaco foreland basin is taken from Uba et al. (2006), that of the Sierras Pampeanas from topography and Dávila and Lithgow-Bertelloni (2013), and that of the Central Patagonian basin from Guillaume et al. (2009). b/ Scenario of flat slab subduction along South America. Subduction zones and mid-oceanic ridges in the reference frame of South America, colour-coded by age, and with triangles on the overriding plate. Flat slab segments from Ramos and Folguera (2009) are plotted in 1 Myr increment, and other subduction zones and mid-oceanic ridges are plotted in 10 Myr increment. Present-day coastlines are shown in black, and the outline of Andean deformation (Arriagada et al., 2008) is shown in grey. The extent of Proterozoic (Archean) continental lithosphere is represented as a light orange (light magenta) shading.

Figure 2: Evolution of the deformation of air-loaded dynamic topography from sources deeper than 350 km and for free-slip boundary conditions, in the South-America fixed reference frame for case 1 at a/ 50 Ma, b/ 29 Ma, c/ 15 Ma and d/ present day. Mid-ocean ridges and transform faults are shown in red, subduction zones (including flat slab segments) in black with triangles on the overriding plate, and present-day coastlines in grey. The locations of cross-sections used in Figures 5, 6 and 7 are shown: along the Amazon river (dark red), at 21°S (dark violet), at 32°S (magenta), and at 69.5°W (light orange). Continental geological features labelled in Figure 1 are shown as white outlines in d.

Figure 3: Temperature and dynamic topography for case 1 along a section down to 750 km depth between -82°W/-12.6°S and -57°W/-7.9°S (dashed black in Fig. 2d) reconstructed in the mantle frame of reference at a/ 11 Ma, b/ 9 Ma, c/ 5 Ma and d/ present-day. Grey contours denote mantle 10% colder than ambient. The amplitude of air-loaded dynamic topography (magenta lines, with reference to black lines) is exaggerated 200 times. P-wave tomography along that section down to 1,300 km depth from e/ (Li et al., 2008) and f/ (Montelli et al., 2006), on which the magenta (case 1), green (case 2), yellow (case 3) and cyan (case 4) contours denote mantle 10% colder than ambient. In a-f, the dashed black lines are the upper-lower mantle boundary (660 km depth), and buoyancy is ignored above the red-dashed lines (350 km depth) in the calculation of dynamic topography.

Figure 4: Effect of long-term fluctuations in sea level ("SL only"; Haq et al., 1987) on continental shallow-water environments at a/ 21 Ma, b/ 16 Ma and c/ 10 Ma, model paleogeography considering fluctuations in sea level and dynamic topography ("SL and DT", see text for more details) at d/ 21 Ma, e/ 16 Ma and f/ 10 Ma, and difference in South American dynamic topography ("DT difference") between present-day and g/ 21 Ma, h/ 16 Ma and i/ 10 Ma. Results on panels d-i are for case 4. Inferred paleoshorelines are shown in blue (Wesselingh et al., 2010), magenta (Smith et al., 2004), and dark green (Golonka, 2009). Other lines are as in Figure 2.

Figure 5: Evolution of dynamic topography, colour-coded by age, sampled on the cross-sections shown in Figure 2, a/ for case 1 (sediment-loaded) along the Amazon River (dark grey shading denotes the approximate extent of Miocene Pebas flooding; Shephard et al., 2010; Wesselingh et al., 2010), b/ for case 1 (sediment-loaded) at 32°S (light grey shading denotes the approximate extent of positive residual topography; Dávila and Lithgow-Bertelloni, 2013; Dávila and Lithgow-Bertelloni,

2015), c/ for case 2 (sediment-loaded) at 21°S (vertical dashed lines, colour-coded by age, denote inferred evolution of the location of the Chaco basin depocentre; Uba et al., 2006) and d/ for case 2 (air-loaded) at 69.5°W (vertical dashed black line at 46.3°S denotes intersections with inferred tilt axis; Guillaume et al., 2009).

Figure 6: Evolution of the rate of change of dynamic topography, colour-coded by age, sampled on the cross-sections shown in Figure 2, and ordered as in Figure 5. a-b/ case 1, c-d/ case 2. Results are for sediment-loaded dynamic topography in a-c (air-loaded in d). The arrows in c, colour-coded by age, represent the paleo-direction of drainage. Pictograms in d, also colour-coded by age, represent the tilt direction of river terraces either side of 46.5°S (Guillaume et al., 2009). Other lines are as in Figure 5.

Figure 7: Evolution of cumulative dynamic topography since 30 Ma, colour-coded by age, sampled on the cross-sections shown in Figure 3, and ordered as in Figures 5 and 6. a-b/ case 1, c-d/ case 2. Results are for sediment-loaded dynamic topography in a-c (air-loaded in d). Pictograms in d, colour-coded by age, represent the tilt direction of river terraces either side of 46.5°S (Guillaume et al., 2009). Other lines are as in Figure 5.

Table 1: Case-specific model parameters.

Case	Flat slabs	Progressive flattening of subduction	Asthenosphere
1	Y	Y	Oceans only
2	N	-	Oceans only
3	Y	Y	Y
4	Y	N	N

743 **Table 2:** Parameters common to all model cases. Subscript “0” denotes reference
744 values.

Parameter	Symbol	Value	Units
Rayleigh Number	Ra	7.8×10^7	
Thermal expansion coefficient	α_0	3×10^{-5}	K^{-1}
Density	ρ_0	4000	$kg\ m^{-3}$
Gravity acceleration	g_0	9.81	$m\ s^{-2}$
Temperature change	ΔT	2825	K
Mantle thickness	h_M	2867	km
Thermal diffusivity	κ_0	1×10^{-6}	$m^2\ s^{-1}$
Viscosity	η_0	1×10^{21}	Pa s
Compositional viscosity pre-factor (continents)	η_{Cc}	100	
Compositional viscosity pre-factor (asthenosphere)	η_{Ca}	10	
Activation energy (upper mantle)	E_{UM}	100	$kJ\ mol^{-1}$
Activation energy (lower mantle)	E_{LM}	33	$kJ\ mol^{-1}$
Temperature offset	T_η	452	K
Background mantle temperature	T_b	1685	K
Radius of the Earth	R_0	6371	km

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Figure