

# Continental Margin Deformation along the Andean Subduction zone: Thermomechanical Models

Muriel Gerbault<sup>1\*</sup>, Cembrano<sup>2</sup> J., Mpodozis<sup>3</sup> C., Farias<sup>4</sup> M., Pardo<sup>5</sup> M.

1 IRD; LMTG; UPS (OMP), F-31400 Toulouse – France.

\* NOW at IRD, Geosciences Azur CNRS UMR5526 Bât 4, 250 rue Albert Einstein, Sophia Antipolis 06560 Valbonne – France .  
gerbault@geoazur.unice.fr.

2 Departamento de Ciencias Geológicas, Universidad Católica del Norte, Avda. Angamos 0610, Antofagasta, Chile

3 Antofagasta Minerals, Apoquindo 4001, Piso 18, Santiago, Chile.

4 Departamento de Geología, Universidad de Chile, Santiago, Chile.

5 Departamento de Geofísica, Universidad de Chile, Santiago, Chile.

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## Abstract

The Chilean Andes extend north-south for about 3000 km over the subducting Nazca plate, and show evidence of local rheological controls on first-order tectonic features. Here, rheological parameters are tested with numerical models of a subduction zone incorporating slab-pull and upper plate convergence, which calculate the development of stress and strain over a typical period of 4 Myr. The models test the effects of subduction interface strength, arc and fore-arc crust rheology, and arc temperature, on the development of superficial near-surface faulting as well as viscous shear-zones in the mantle. Deformation geometries are controlled by the intersection of the subduction interface with continental rheological heterogeneities. Upper plate shortening and trench advance are both correlated, and favored, to a first order by upper plate weakness, and to a second order by interface strength. In cases of a strong interface, a weak forearc crust is dragged downward by “tectonic erosion”, a scenario for which indications are found along the northern Chilean margin. In contrast for a resistant forearc, the slab-pull force transmits to the surface and produces topographic subsidence. This process may explain present-day subsidence of the Salar de Atacama basin and/or the persistence of a Central Depression. Specific conditions for northern Chile produce a shear zone that propagates from the subduction zone in the mantle, through the Altiplano lower crust into the Subandean crust, as proposed by previous studies. Models with a weak interface in turn, allow buoyant subducted material to rise into the continental arc. In case of cessation of the slab-pull, this buoyant material may rise enough to change the stress state in the continental crust, and lead to back-arc opening. In a case of young and hydrated oceanic plate forced by the slab-pull to subduct under a resistant continent, this plate is deviated and indented by the continental mantle, and stretches horizontally at ~100 km depth. This situation might explain the flat Wadati-Benioff zone of Central Chile.

**Keywords:** *Andes subduction, thermo-mechanical models, interface strength, trench motion, fore-arc deformation, flat slab.*

## 1. Introduction

Although the N-S trending Chilean subduction margin is nearly linear over a distance of about 3000 km, the width, height, and geological characteristics of the Andes vary significantly along and across strike. Several studies have related the differences in tectonic style of the Central and Southern Andes to changes in slab age, dip and convergence obliquity (e.g. Jordan et al., 1983; Lavenu & Cembrano, 1999; Gutscher et al., 2000; Yañez & Cembrano, 2004). Yet, other authors have shown that rheological heterogeneities within the continental plate also play a fundamental role on Andean tectonics, as western South American continental margin records a long and complex geological history of terrane accretion and amalgamation of diverse lithospheric blocks of variable composition and strength during the Late Proterozoic and Paleozoic (e.g. Isacks, 1988; Lamb & Hoke, 1997; Tassara, 2006; Cembrano et al., 2007, Ramos, 2009). Our aim here is to present and discuss thermo-mechanical models that give insight into the balance between thermal, compositional and frictional parameters that control first and second-order tectonic structures, such as conditions for the development of a Central Depression, magnitudes of crustal thickening, the vergence of major fault systems, and the connection of upper crustal structures with processes of ductile flow at depth. We also address why, in the Southern Andes, upper plate deformation has been mostly accommodated at the magmatic arc, whereas north of 36°S and even farther north of 23°S, it has produced much larger amounts of crustal shortening and has affected a much greater crustal volume. In some ways similar to previous studies (e.g. Yañez & Cembrano, 2004; Sobolev & Babeyko, 2005; Medvedev et al., 2007), this approach attempts to consider key aspects of the large-scale force balance between plates, together with rheological variations associated with local geology. However, our final objective is not to reproduce the history of the Chilean continental margin, but rather, to assess how variations in inner and interplate strengths may generate key orogen-wide tectonic features.

This contribution starts with a review of existing models of subduction zones (section 2). The large number of mechanical parameters involved in a subduction zone obviously result in a wide variety of modeling approaches, and we cite only some of them in order to present the context of our own approach. We then present the main tectonic features that characterize the Chilean subduction zone, and especially specific areas that are discussed in comparison to our models (section 3). The experienced reader may directly jump to section 4, in which we explain our modeling assumptions, based on a 2D Lagrangian approach that accounts for elastic-viscous-brittle behavior. With initial conditions close to present day conditions in the Chilean Andes, the models calculate stress build-up and propagation of deformation when applying upper plate motion and basal slab-pull over an intermediate time scale of 4 Ma. We have chosen this time span because (1) it represents well the order of magnitude of typical discrete deformation events affecting active continental margins (which however, superimpose on regional features resulting from first-order kinematic and rheological parameters that remain essentially the same for several tens of millions of years), and (2) it represents a reasonable time-scale from the thermo-mechanical point of view, for which dynamical processes linked to stress and strain propagation have time to develop, but more complex processes, such as phase change and gravitational instabilities, can still be neglected. In section 5 we present the dynamics of deformation, with series of models in which the effects of differences in the rheology of the crust, arc temperature anomaly, and subduction channel strength are displayed. Then in section 6 these results are discussed in the framework of first- and second-order features of the Chilean margin, fore-arc tilting and subsidence, trench motion, location of crustal shortening and thickening. We illustrate a specific model developed for the Northern Chile subduction zone in a previous study that focused on its double seismic zone (Dorbath et al., 2008). We show a final model that proposes a cause for slab flattening as observed in Central Chile (section 6.5), as an alternative to the already well known slab buoyancy argument (e.g. Van Hunen et al., 2002).

## 2. Modeling subduction zones: a short review

The subduction interface in plate convergence zones accommodates thousands of kilometers of displacement during tens of millions of years, along a high strain contact zone only a few kilometers thick. Its rheological behavior is intrinsically related to that of adjacent crustal and lithospheric rocks with which it also combines. Yet, although these high-strain zones control the dynamics of our planet (e.g. Burlini & Brun, 2005; Regenauer-Lieb & Yuen, 2003), it has proven difficult for models based on continuum mechanics to reproduce the large amounts of shear deformation involved. In addition, the precise nature of Benioff zones at depths greater than 50 km depth remains unclear, as well as their relation to subducted oceanic crust. Whether there still is a subduction channel (see for example recent seismic imaging by Rondenay et al., 2008), or whether it has been completely scratched off, and only the top of the subducting oceanic

mantle remains in direct contact with the asthenosphere, is beyond the scope of this paper. In a broader view than the subduction channel alone, all thermal, mechanical, and chemical processes involved in subduction are nearly impossible to take into account collectively in a single model: numerous studies have been developed up to now, and still, they have been able to account for only part of these processes. Here we attempt to briefly review and discuss key studies in relation to our present approach, rather than make a thorough review. We recall first the role of large-scale asthenosphere-lithosphere interactions, and then the role of more superficial rheological factors, on arc deformation styles. We also cite previously published models specifically applied to the Chilean Andes.

## **2.1 Surface deformation resulting from deep lithosphere-asthenosphere interactions**

Forsyth and Uyeda (1975) demonstrated the dominant importance of the negative buoyancy of subducting slabs (slab-pull) over other plate-scale forces (e.g. ridge push or basal drag). Deformation of the overriding plate and arc topography result from a dynamic equilibrium between the slab-pull that drags the overriding plate downward (and deep interactions with the mantle at greater depths) and the interplate buoyant force that acts in the opposite sense (e.g. Jyschke 1975; Chemenda, 1994). The Andes has long been regarded as “the” typical example of a mountain belt associated to oceanic subduction under a continental plate, where the dynamics of orogenic processes are controlled by factors such as convergence rate (high) and age of the subducting plate (young) (e.g. Molnar and Atwater, 1978; Somoza, 1998; Uyeda and Kanamori, 1979). Chile's subduction zone satisfies Jarrard's (1986) findings, that young and buoyant slabs produce shallow dips, and strong seismic coupling between both plates. Jarrard (1986) also looked for a correlation of the upper plate strain regime with slab age and upper plate absolute motion, but despite the Nazca's plate young age, the Chilean Andes belong to the most compressional regime. Until now, the debate remains in determining what controls the extensional or compressional regime of the upper plate (e.g. Heuret & Lallemand, 2005), and besides subducting plate age and velocity, several studies indicate that upper plate motion should be a prime factor. Intraplate deformation in the central Andes is likely to be driven by the absolute motion of South America, based on estimated amounts of crustal contraction, plate rollback and tectonic erosion of the margin (Russo & Silver, 1996; Marret & Strecker, 2000; Yañez & Cembrano, 2004; Sobolev & Babeyko, 2005).

A large family of models demonstrate the effects of temperature on the dynamics of subduction zones. Initially, simple thermal models assumed kinematical conditions and resolved the heat equation through time: with such models Gutscher et al. (2000) for example, explored the thermal effects of buoyant flat slabs in inhibiting the occurrence of arc magmatism in parts of the Andes. More advanced thermal Eulerian methods incorporate some mechanics: they have the advantage of accounting for thermally dependent viscous rheology, but the inconvenient of fixing or predefining the velocity of the top of the subducting and of the overriding plates. In these approaches, input parameters such as plate convergence rate, temperature of the subducting slab, and shear heating all affect the vigor of mantle wedge corner flow (e.g. Molnar & England 1990; Peacock, 2000). Based on observations of low bathymetry, gravity anomalies and seismic attenuation like in the Tonga-Kermadec subduction zone, Billen & Gurnis (2001) show that corner wedge viscosity should remain low, a property that is best satisfied in warm and shallow mantle wedges, and is associated with a warm overriding plate (e.g. van Keken et al., 2002; Currie et al., 2004; Conder, 2005; Abers et al. 2006). With similar numerical assumptions, Gerya et al. (2003, 2006) developed coupled petrological–thermo-mechanical viscous models to investigate the dynamical effects of metamorphic reactions and melting on ascending plumes of slab fluids. In these models, fluid flow develops beneath volcanic arcs, causing hydration and partial melting of the mantle wedge, and recycling of crustal rocks back into the fore-arc. Arcay et al. (2006), in turn, model rheological weakening of the arc-mantle lithosphere when accounting for water fugacity in the rheological creep behavior of mantle rocks. However, some subduction zones may instead have a cold mantle wedge, as indicated by surface heat flow data (e.g. Honda, 1985) or tomographic investigations compared with laboratory experiments on lattice preferred orientation of olivine (e.g. Kneller et al., 2007, see also review by Wiens et al., 2008). In extreme cases, models show that the fore-arc mantle wedge may be entrained to sink together with the subducting slab (e.g. Boutelier et al., 2003). We will illustrate the effect of cold versus warm mantle wedge in the approach presented here.

With analog experiments and numerical models, the stiffness, viscosity and density contrasts of the slab with the surrounding mantle were shown to determine its mode of deformation as it meets the 670 km deep boundary (bending, unbending, buckling, e.g. Christensen, 1996, 2001; Becker et al., 1999; Ribe et al., 2007; Funiciello et al., 2003a,b; Capitanio et al., 2007; Schellart et al., 2008a). As the slab crosses, bends or folds along this boundary, trench motion, slab dip and fore-arc regime become dynamically variable at the time-scale of millions of years (e.g. Van Hunen et al., 2004; Bellahsen et al., 2005; Morra et al., 2006; Clark et al., 2008). More recently, 3D mantle flow processes around a freely subducting slab have been modeled and show effects on the evolution of trench motion and curvature (e.g. Funiciello et

al., 2003; Morra et al., 2006; Schellart et al., 2007). In our study we will not consider these causes of fore-arc deformation, as we restrict our modelling approach to 2D, as well as to the upper 300 km shell of the Earth. On the other hand, these latter studies do not focus on the detailed elasto-plastic rheology of the overriding lithosphere, which however will be investigated here. Bonnardot et al. (2008) developed a new finite-element method that accounts for solid-fluid coupling of elasto-plastic lithosphere with a viscous asthenosphere: these models not only show the role of elasto-plasticity of the subducting plate on topography and stress regimes of the overriding plate, but also the fundamental role of a low asthenospheric viscosity in producing extensional stresses at the lithospheric surface. Contrarily to Billen & Gurnis (2001), Bonnardot et al. (2008) obtain less forearc subsidence for high corner wedge viscosities. Such contrasting results illustrate the importance and yet controversial hypotheses of subduction zones models. Recent semi-analytical studies show the effect of dynamic topography on top of subduction zones (dynamic topography is the vertical response of the Earth surface to viscous flow of the underlying mantle). The presence of dense slabs may decrease topography by as much as 2 km according to Husson (2006). According to Royden & Husson (2006), “*viscous stresses surrounding a subducting slab act as a driving engine in the deeper parts of the subduction system while shallower parts (<100-150km) act as a ‘brake’ on trench motion; trench migration reflects a competition between these two effects.*” These authors also conclude that the rheology and density of the frontal prism (overriding the subduction) exerts “unexpected” effects on trench migration rates and slab dynamics, effects that will be at least partly illustrated in the following models.

## 2.2 Deformation resulting from superficial lithospheric rheology

For more than a century diverse analog and numerical models have explored the mechanisms of deformation in layered crust and lithosphere. Structural styles of deformation are shown to depend on elastic, viscous and brittle yielding properties of rocks, producing, from the simplest to most complex layering, either periodical buckling and faulting of rather homogeneous layers (e.g. Biot, 1961; Ramberg 1969; Gerbault et al., 1999), asymmetric wedges and flower structures in heterogeneous layers (e.g. Davis et al., 1983; Lallemand et al., 1994), or double-vergent wedges and lateral lower crustal flow in contexts of continental collision (e.g. Willet et al., 1993; Beaumont et al., 2000; Ellis et al., 1995; Willingshofer et al., 2005). Layers deform and thicken when indented or pinched by stronger layers, and may connect at depth with weak and thin decollement levels along which huge geological terranes can slide horizontally over tens of kilometers. Similarly, deformation of the fore-arc, arc and back-arc areas of subduction-related plate margins obviously results from the geometry of crustal and mantle rheologies (e.g. jelly-sandwich structure, Watts & Burov, 2003), and develops both flexural (Audet et al., 2008) and viscous flow features (e.g. Gorczyk et al., 2007).

Structural characteristics of a fore-arc compares to orogenic-scale accretionary wedges, in that they are controlled by some kind of basal friction exerted by the subduction interface. The analytical studies of Chapple (1978) and Davis et al. (1983) first related the angles of internal friction and basal friction ( $\phi_b$ ) to tapered angle and topographic slope, using the theory of plasticity. This concept of basal friction was then extrapolated to subduction interplate friction, which was shown to control upper-plate state of stress and deformation, with elasto-plastic plates (e.g. Hassani et al., 1997). Buiter et al. (2001) explored vertical surface displacements and trench motion in models with elastic lithospheres, evidencing the elastic contribution of first order features of subduction zones. Hampel & Pfiffner (2006) also illustrated the state of forearc stress with visco-elastic plates, producing either uplift or subsidence patterns depending on this interplate friction. Interplate friction in these contexts is considered as an effective value that includes the “pore pressure” effect of lubricating sediments entering the trench on lowering the “basal” shear stress (e.g. Davis et al. 1983; Lamb, 2006).

At the scale of the seismic cycle, the concept of elastic crustal wedge is used to estimate seismogenic strength along subduction interfaces: Wang & He (1999) studied low-stress fore-arcs typical of Nankai or Cascadia, and found a basal friction close to  $\mu = \tan \phi_b = 0.05$  ( $\phi_b = 2.86$ ), emphasizing the importance of the distance along the slab interface over which this friction acts (seismogenic depth). With a similar approach, Lamb (2006) evaluated shear stresses in subduction zones worldwide, and identified the Northern Chile trench as a high friction and low pore-fluid pressure case with  $\mu = 0.095$  ( $\phi_b = 5.4$ ). Elastic-viscous-plastic thermo-mechanical models of the Central Andes over 25 Myr used a friction  $\mu = 0.015-0.05$  (Sobolev & Babeyko, 2005; Sobolev et al., 2007). In the present study we will test the role of overriding plate rheology and that of the interface channel strength.

Geodetic displacements in the Chilean Andes are also used to model post-seismic motion (e.g. Khazaradze & Klotz, 2003; Liu et al. (2000) with 3D visco-elastic overriding plate), and domains of either low or high viscosity lower crust appear, at the time-scale of the seismic cycle. These models are numerically not expensive, because deformation remains small, but they do not intend to explain fault zones and topography over a million years time-scale.

The synthetic thin viscous sheet approximation, which considers subduction-related forces applied on the borders of a plate of homogeneous viscosity (England & Mc Kenzie, 1982) was also applied to the Central Andes. Assuming

depth-dependent shear stress, Wdowinski & Bock (1994) linked trench topography to convergence velocity, overriding plate viscosity, interplate viscosity and thickness, and dip angle of the Benioff plane. Yañez & Cembrano (2004) used a similar approach to evaluate mean viscosities of the overriding plate and subducting channel, which they found to be in direct correlation with the age of the subducting slab. High viscosities were found at latitudes 18 to 28°S where the Altiplano and Puna plateaus extend, and where the slab is the oldest (~50 Myr). Finally, Medvedev et al. (2007) developed a three-dimensional thin-sheet model that incorporates changes in upper plate strength, lower crust channel flow, and anisotropic interplate coupling. They correlated the observed decrease in amounts of shortening between Northern and Southern Chile, first with contrasts in effective viscosity between a weaker Andean margin with respect to the foreland Brazilian shield, and second with a decrease in interplate coupling. We will look at similar effects here.

The Chilean subduction zone has also been modelled by Sobolev & Babeyko (2005) and Sobolev et al. (2007) by using an elasto-visco-plastic approach similar to ours, but on a time-scale of 30 Myr. Sobolev et al. (2007) argue that the major factor that controls tectonic shortening in the Andes is the fast and accelerating, westward drift of the South American Plate. They infer that the peculiar orogeny in the Central Andes (see section 3) took place because three major conditions were met: a high overriding velocity of the South American Plate, a back-arc pre-thickened crust, and a high friction coefficient of the subduction channel due to an arid climate. Sobolev et al. (2007) show that a model with such parameters generates more than 300 km of tectonic shortening within 35 million years, whereas a model with initially thinner continental crust and a lower interface friction results in less than 40 km of shortening, close to the situation in the Southern Andes. Sobolev et al. (2007) also argue that delaminating lithosphere interacts with subduction-zone corner flow, influencing both the rate of tectonic shortening and magmatic-arc productivity. In contrast to this study, our aim here will be to look in more detail at surface structures and their connection with rheological structures at depth.

The latter results, together with several other studies exploiting data from the Andean subduction zone (e.g. gravity anomalies, Iaffaldano & Bunge, 2008) contradict Schellart et al. (2008b) arguments that trench motion and overriding plate regime depend mostly on 3D slab-mantle deep interactions rather than on overriding plate properties and kinematics. Yamato et al. (2009) recently showed that the incorporation of a fixed upper plate, significantly reduces these 3D effects. These discrepancies between results in the community of subduction modellers, show once more the problems rising by not being able to reconcile yet, realistic elasto-viscous-plastic plate behaviour and deep slab-mantle processes.

### 3. Andean tectonics and rheological characteristics

The Andean orogen is a tectonically segmented mountain range whose changes along strike have been related to first-order variations in the geometry and age of the subducting Pacific plates and/or to the nature and rheology of the overriding South American Plate (e.g. Jordan et al. 1983; Mpodozis and Ramos, 1989; Kley et al, 1999; Ramos, 2009). The modern Andes of Chile and Argentina include three major segments that coincide with areas of changing steep to shallow subduction. The main tectonic, geologic and geophysical features of these segments are summarized below, including nature and timing of internal boundaries and regional fault systems, and total amounts of shortening.

#### 3.1 The Central Andes from 18 to 28°S

In the Central Andes of northern Chile between 18° and 28°S, the Nazca plate dips at ca 25°-35° below the South American continent (Baranzangi and Isacks, 1976). From 18° to 22°S the orogen is up to 700 km-wide, and is underlain by a very thick crust (ca. 70-75 km, ANCORP Group 1999, 2003), the Altiplano region (**Fig. 1**). This segment is characterized by a mainly undeformed forearc including a 1-3 km-high Coastal Cordillera, a longitudinal Central Depression and a 2-4 km-high Precordillera. Farther east, the more than 6 km high stratovolcanoes of the active Central Andean Volcanic Zone form the Western Cordillera, which along with the 4-6 km-high Eastern Cordillera flank the relatively flat, average 3.7 km-high, Altiplano region. Inland, east of the Eastern Cordillera are the 150-200 km wide Sub Andean ranges, an active fold-and-thrust belt that form the tectonic front of the Andes, and the low elevation (< 500m) Beni-Chaco foreland plains (Baby et al., 1997).

The most prominent structural feature of the Coastal Cordillera is the Atacama Fault Zone (AFZ, **Fig. 1**), a trench-parallel structural system extending over more than 1000 km between 20° and 30° S (e.g. Gonzalez et. al.2003; Riquelme et al., 2003; Cembrano et al., 2005). It was formed in the Early Jurassic as a left-lateral, intra-arc fault (Scheuber & González, 1999), and was partly reactivated, not only in the Cretaceous, but also in the Neogene and the Quaternary. Today, the AFZ separates the uplifting Coastal Cordillera from the inner forearc (e.g., Adam & Reuther, 2000), showing neotectonic kinematics dominated by normal faulting and minor dextral strike-slip movements (González et al., 2003).

North of 22 °S, the boundary between the forearc and the Precordillera includes a broad zone of west-vergent, monoclinical folds and high-angle reverse faults formed as a consequence of uplift of the western Cordillera and Altiplano between the Oligocene-Late Miocene (e.g. Muñoz and Charrier, 1996; García et al., 2002; Victor et al., 2004; Farías et al., 2005). To the east of the Altiplano, the Eastern Cordillera is a thin-skinned, essentially east verging, fold-and-thrust belt that has accommodated more than 200 km of horizontal shortening (Muller et al., 2002; McQuarrie et al., 2008). Most of the shortening leading to the main phase of uplift and exhumation of the Eastern Cordillera seems to have occurred during the Eocene-Oligocene, between 45-25 Ma, (see Roperch et al., 2000; Arriagada et al., 2008 and references therein). Deformation ceased in the Eastern Cordillera during the Miocene as the deformation front migrated eastwards into the Sub-Andean zone where the rigid lithosphere of the Amazonian craton is now being underthrust below the Altiplano Plateau (Isacks, 1988; Dorbath et al., 1996). Total amount of horizontal shortening in the Central Andes at the latitude of northernmost Chile since the Eocene (~40 Ma) may have reached between 300 and 400 km (e.g. Kley and Monaldi, 1998; McQuarrie et al. 2008). Crustal shortening decreases southward between 22° and 28°S in association with systematic changes in the deformation style from thin-skinned to thick skinned both in the Eastern Cordillera and the Sub Andean belt, (e.g., Kley et al., 1999, Hogn et al, 2007) as the Altiplano becomes narrower and higher giving way to the Puna region of NW Argentina (see Allmendinger et al., 1997).

Geophysical studies have shown that the fore-arc region in northern Chile is cold and rigid, From 2D flexural analysis, Tassara (2005) showed that the fore-arc region show high rigidities between 15° and 23°S, that decrease gradually southward, an observation consistent with magnetotelluric measurements and Qp tomography that do not show indications of fluids in the fore-arc region, but rather cold material containing only minor free fluid phases. In contrast, rigidity sharply decreases eastwards in the Altiplano-Puna where the crust is thick and warm, a situation leading Tassara (2005) to suggest that the rigid fore-arc acts as a crustal indenter abutting against the weak Altiplano lithosphere.. The Altiplano and Western Cordillera present first order geophysical anomalies at depth (see review in Schilling et al, 2007, and references therein), including a very high conductivity zone at 21°S and at depths greater than 20 km (e.g. Schwarz et al. 1994; Brasse et al., 2002) and high seismic attenuation (Qp, Rietbrock, 1999), located below a mid-crustal low seismic velocity zone (between 20 and 25°S, e.g. Yuan et al., 2002). Despite the proposition of the occurrence of a sill-like magma body at 20 km depth (e.g. Zandt et al., 2003), large amounts of melts remain controversial (see review by Schilling et al., 2007). Yet, the presence in the plateau region of extensive fields of Neogene ignimbrites (Altiplano-Puna Volcanic Complex), is consistent with the occurrence of crustal melts (Zandt et al., 2003; De Silva & Gosnold, 2007) The thick crust below the Eastern Cordillera is commonly related to Paleogene tectonic shortening (Schmitz, 1994; Isacks, 1988; McQuarrie, 2002), in contrast with the arc and forearc regions, for which crustal thickness would also include a component of magmatic addition derived from partial melting of hydrated mantle (Giese et al. 1999; Sempere et al., 2008). Deep thickening related to lateral crustal flow has also been proposed (Lamb et al., 1997; Farías et al., 2005, Husson & Sempere, 2003; Gerbault et al., 2002,; Medvedev et al., 2007).

### **3.2 The Andes of Central Chile, 28 to 33°S**

At 28° the central Depression and the Altiplano-Puna plateau disappear and, from 28 to 33°S the Andes overlies the “Pampean” flat slab region of central Chile and Argentina (Ramos et al., 2002) also characterized by a lack of active arc magmatism (**Fig. 1**). In this segment, the absence of volcanic rocks younger than 6 Ma (Kay et al., 1987) is attributed to the disappearance of the asthenospheric wedge below the continent, and a consequence of Miocene slab shallowing (Kay et al, 1987; Kay & Mpodozis, 2002). Here, the Main Andean Range (including the Principal and Frontal Cordilleras), corresponds to a narrower, ca 100 to 150 km-wide mountain belt. However, active deformation reaches up to 800 km inland from the trench axis across the thick-skinned Precordillera fold-and-thrust-belt in the San Juan-Mendoza region in Argentina and the Sierras Pampeanas even further to the east (i.e. Jordan et al., 1983; Siame et al., 2005). The Frontal Cordillera forms a double-vergent uplifted block, thrust to the west over the Coastal Cordillera, and to the east over the Argentinian foreland. Total shortening amounts for this segment range from 130 to 150 kms (e.g. Allmendinger et al., 1997). The Andean basement is here formed by a mosaic of tectono-stratigraphic terranes accreted against the Gondwana margin, during the early Paleozoic beginning with the accretion of the Pampia terrane (Eastern Sierras Pampeanas) at 530 Ma, the Cuyania terrane (Precordillera and westernmost Sierras Pampeanas terranes) in the Ordovician-Silurian, and finally the Chilenia terrane during the Devonian (Astini et al., 1995; Thomas & Astini, 2003; Ramos, 2009). This terrane collage forms an heterogeneous basement rheology that certainly influenced the tectonic development of the region during the Mesozoic and Tertiary. The spatial correlation of the southern termination of the Pampean flat-slab segment with the Juan Fernandez ridge at 33°S suggests a causative relationship, linked either to buoyancy or to indentation mechanics (e.g., Gutscher et al., 2000; Yañez et al., 2002). However, it remains difficult to tell whether the onset of

compression is associated with the arrival of this ridge into the subduction zone (e.g. Yañez et al., 2002), since tectonic events associated to the beginning of slab-shallowing also affect areas far away from the subducting ridge (Kay and Mpodozis, 2002).

Various workers have explored the deep structure of the Pampean flat slab region, providing data on the seismotectonics Moho depths and crustal rheology (Pardo et al., 2002, Alvarado et al., 2005, 2007; Fromm et al., 2004, Gilbert et al., 2006). Crustal thickness reaches up to 45-50 km, with altitudes locally exceeding 6.5 km (e.g. Tassara et al., 2006; **Fig. 1**). While the seismic flat segment is about 250 km long (Pardo et al., 2002; Anderson et al., 2007), the crust is up to 65 km deep below the High Cordillera, and associated with seismic properties similar to dry and cold partially eclogitized upper mantle (Alvarado et al., 2007). Wagner et al. (2005) also studied the seismic velocity structure of the mantle above the horizontal portion of the flat slab region, and suggest that it is consistent with highly depleted, cold, Mg rich compositions found in cratonic lithosphere.

### 3.3 The Southern Andes from 33.5°S to 47°S

South of 33.5°S, a major tectonic transition occurs at the southern end of the Pampean flat slab zone, as the active, Quaternary, volcanism and the Central Valley reappear, and the foreland basement uplifts disappear as the subducting slab became again steeper (Cahill and Isacks, 1992, Ramos, 2009; **Fig. 1**). The Main Andean Range in the northern part of this third segment is narrow (c.a. 100 km wide) and average elevations decrease continuously from more than 4500 at 33°S to less than 2000 m at 38°S, in correlation with a decreasing crustal thickness from 45 km at 33-34°S to 35 km to the south (Tassara et al., 2006; Alvarado et al., 2007, **Fig. 1**). The Main Cordillera of this segment is essentially formed by a thick pile of Neogene (Oligocene-Miocene) volcanics in the western flank of the range (Charrier et al., 2002, 2005; Kay et al., 2005), and a thin skinned (Aconcagua) and mixed (Malargue) east-verging fold-and-thrust belts involving Mesozoic sedimentary strata of the Neuquén basin along its eastern side (Ramos & Kay 2006; Giambiagi et al., 2003). The estimated tectonic shortening since the Late Oligocene is rather modest (<100 km, Kley & Monaldi, 1998; Giambiagi & Ramos, 2002).

Towards the south, the width of the Central Depression decreases from 38°S (Lavenu and Cembrano, 1999), coinciding with the appearance of the 1200 km long, “intra-arc” trench-parallel, dextral strike-slip Liquiñe-Ofqui Fault Zone (LOFZ, Lavenu and Cembrano, 1999; Lange et al., 2008; **Fig. 1**), which dominates the Neogene tectonic evolution of the Main Andean range. Tectonic shortening in the eastern foreland and the Neuquén basin diminish (Kley et al., 1999; Vietor & Etchler, 2006). It becomes even smaller further south between 42°S and 47°S where dextral transpression in the Mio-Pliocene (ca. 6-4 Ma) led to differential exhumation and uplift of the Patagonian Batholith, which defines a 60 km wide crustal scale pop-up structure, that forms the topographic axis of the north Patagonian Main Cordillera (Cembrano et al., 2002, Adriasola et al., 2005, Thomson et al., 2001). This regional-scale pop-up structure compares with three-dimensional numerical models (Braun & Beaumont, 1995) of orogenic blocks bounded by double-verging thrust zones. Geochronological studies (Cembrano et al., 2002, Adriasola et al., 2005) indicate that long-term dextral transpressive regime would be driven by oblique subduction during the Cenozoic (e.g. Dewey & Lamb, 1992; Hervé, 1994; Cembrano et al., 1996), whereas the short-term deformation from 6 Ma to present would rather be controlled by the collision of the northwards-migrating Southern Chile Triple Junction now located at ~47°S (Forshyte & Nelson, 1985). Southward decrease in crustal thickness and orogen width (**Fig. 1**) suggest that shortening also decreases, consistent with the increase in obliquity of subduction system.

## 4. Numerical Assumptions

### 4.1 Numerical method

Here, we use a 2D Lagrangian approach that accounts for elastic-viscous-brittle behavior, over an intermediate time scale of about 4 Ma: the idea is to setup a model with reasonable initial thermal and rheological structures, close to the generic present-day Chilean margin conditions, and to record stress build-up and propagation of deformation when applying upper plate velocity and basal slab-pull. Sobolev & Babeyko (2005) and Sobolev et al. (2007), used a similar numerical code to ours (Flac, described below), and very similar initial conditions (see next section), to model the Andean orogeny over a time-span of 30 Myr. Instead, our aim here is to look in more detail at a short-scale control on variation of topography, and the connection of localised deformation at the surface with rheological structures at depth.

We use a version of the numerical code Parovoz (Poliakov & Podladchikov, 1992), based on the FLAC method (Fast Lagrangian Analysis of Continuum, Cundall & Board, 1988). This two-dimensional plane-strain method is

recognized in engineering mechanics, for its capability to model the initiation and propagation of faults or shear zones, because it handles non-associated elasto-plasticity and a failure criterion expressed in the three-dimensions of the stress field (e.g. Hobbs and Ord, 1989; White et al. 2004). It has been adapted to many geodynamical applications, at crustal and lithospheric scales, in collision and extensional settings, in oceanic and in continental environments (e.g. Poliakov et al., 1993; Lavier et al. 1999; Burov et al., 2007; Yamato et al., 2007; Gerbault et al., 1998, 2005).

The equations of motion are solved in a large-strain Lagrangian formulation (the mesh moves with material) and with an explicit time-marching scheme:

$$\rho \partial \frac{\partial u / \partial t}{\partial t} - \text{div } \sigma - \rho g = 0,$$

where  $u$ ,  $\sigma$ ,  $\rho$ ,  $g$ ,  $t$  stand for the displacement, stress, density, acceleration due to gravity, and time respectively.

Visco-elasto-plastic constitutive equations relate stress to deformation, expressed with a general functional  $F$ :

$$D\sigma/Dt = F[\sigma, u, \partial u / \partial t, T, \dots].$$

The terms  $\partial / \partial t$ ,  $D / D t$ , and  $T$  denote a time derivative, an objective time derivative, and temperature, respectively. More specifically, elastic-viscous-brittle behavior is accounted for with the total strain increment in each numerical element defined by the sum of elastic, viscous and brittle deformation. At any given element of the model and at any time-step, the stress behavior is self-consistently chosen from the minimum of the elastic-plastic  $\sigma_{ep}^{II,dev}$  and visco-elastic deviatoric stress  $\sigma_{ve}^{II,dev}$ , with the second invariant of deviatoric stress being defined by

$$\sigma^{II,dev} = 1/2 [\sigma_{ij} \sigma_{ij}]^{1/2}.$$

Elasto-plasticity follows the linear elastic Hooke law ( $\sigma_{ep} = \sigma_e$ ) until a failure yield stress is reached ( $\sigma_{ep} = \sigma_p$ ).

Hook's law linearly relates stress and strain (components  $\epsilon_{ij}$ ), with Lamé's parameters  $\lambda = \mu = 3.10^{10}$  Pa,  $I$  the identity matrix and  $tr[]$  the trace function:

$$\epsilon_{ij} = 1/2 \left[ \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right], \quad \sigma_e = \lambda tr[\epsilon] I + 2 \mu \epsilon$$

The failure yield is prescribed with a Mohr-Coulomb non-associative criterion (dilatancy is 0), and occurs when the tangential stress along any given plane reaches the yield stress  $\tau_p$ , and is a function of the normal stress  $\sigma_n$ , friction angle  $\varphi$  and cohesion  $S_0$ :  $\tau_p = S_0 - \tan \varphi \cdot \sigma_n$ . The normal stress expresses as  $\sigma_n = P + \sigma^{II,dev} \cdot \sin \varphi$ . In terms of principal stresses, the yield criterion can also be written as  $\sigma_1 - \sigma_3 = \sin \varphi [\sigma_1 + \sigma_3 - 2 S_0 \cdot \tan^{-1} \varphi]$ .

Maxwell visco-elasticity depends on the deviatoric strain-rate  $\dot{\epsilon}^{dev}$  and the effective viscosity  $\nu^{eff}$ , that incorporates the power-law stress-strain relationship and exponential temperature dependency obtained by laboratory rock experiments on creep:

$$\sigma_{ve}^{dev} = \left[ \mu \dot{\epsilon}^{dev} dt + \sigma^0 \left[ 1 - \frac{\mu dt}{2 \nu_{eff}} \right] \right] \cdot \left[ 1 + \frac{\mu dt}{2 \nu_{eff}} \right], \quad \nu^{eff} = 1/4 [4/3A]^{1/n} \cdot [\dot{\epsilon}^{II,dev}]^{1/n-1} \cdot \exp^{Q/nRT}.$$

where  $\sigma^0$  are stress components from the previous time-step,  $A$  is the material parameter,  $n$  is its effective stress exponent,  $Q$  its creep activation energy, and  $R$  is the universal gas constant. Specific values of these creep properties are given for each material in Table 1 (e.g. Kirby & Kronenberg, 1987; Chen & Morgan, 1990; Kohlstedt et al., 1995; Ranalli, 1995, and references therein).

Temperature is initially calculated from oceanic and continental thermal ages (Burov & Diament, 1995), and by inserting an arc anomaly of arbitrary geometry (see section 4.3). At each time-step, the heat transfer equation is solved:

$$\rho C_p \frac{\partial T}{\partial t} + \dot{u} \nabla T - k \Delta T - H_r - H_s = 0.$$

$T$ ,  $\rho$ ,  $C_p$ ,  $k$ ,  $H_r$ ,  $H_s$  denote temperature, density, specific heat, thermal conductivity, internal heat production, and shear heating, respectively. The advective term  $\dot{u} \nabla T$  is implicitly resolved by the displacements of the Lagrangian grid. Radioactive heating is depth-dependent and is defined as  $H_r = H_0 \cdot \exp(-y/h_r)$ , with  $H_0 = 9.e-10$  W/kg,  $h_r = 10$  km. Shear heating is produced by dissipative nonrecoverable processes and depends on viscous and plastic deviatoric stress and strain-rate, and is written as  $H_s = \xi \cdot \sigma^{II,dev} \cdot \dot{\epsilon}^{II,dev}$ , with efficiency  $\xi = 0.9$  (e.g. Kaus & Podladchikov, 2006). Shear heating in the subduction channel is also included in the initial conditions (with 10 MPa, based on evaluations of surface heat flow and observed metamorphic reactions, e.g. Molnar & England, 1990; Peacock, 1996; Lamb, 2006).



Note that in our models, we assume that the initial thermal state is close to present-day steady state. Temperature is thus basically useful for the evaluation of temperature-dependent rheologies, e.g. brittle-ductile transitions and viscosities. This version of Parovoz does not account for phase reactions and subsequent density evolution.

The remeshing technique is based on the use of passive markers (e.g. Yamato et al., 2007; Dorbath et al., 2008): initially, nine regularly spaced markers are inserted in each mesh element, which passively move with the Lagrangian grid during normal computation, carrying elementary stresses and linearly interpolated nodal temperatures. All 9 markers of an element are assigned elementary material properties, which include density, conductivity, and rheological parameters, and which are represented by a colour “phase” (see setup **Fig. 2**). When the angle of any triangular element becomes smaller than  $13^\circ$ , remeshing occurs and a new regular mesh replaces the old one. Then, markers are counted in each new element, in order to identify the dominant material properties (“phase”) that composes it. During remeshing, markers are also used to evaluate new stress components, averaged over all markers located inside the new sub-element.

## 4.2 Initial conditions

We now “translate” geological and geophysical properties of the Central and South Andean margin, described in section 3, into thermo-mechanical parameters for the models. We first assign initial geometries of rheological layers, that define the oceanic and continental crust, mantle lithospheres, asthenosphere and subduction channel (**Fig. 2**). The domain is 1800 km wide and 300 km deep, which is a trade-off between a reasonable depth where mantle viscosity increases (e.g. Husson & Royden, 2006), and computational cost of the domain size. It comprises a flexuring oceanic plate dipping at  $35^\circ$ , an average for Chilean conditions. This domain is meshed with 600 x 100 elements, so that horizontal resolution is 3 km, and vertical resolution decreases downward from 2 to 4 km.

The continental mantle is composed above 100 km depth by a lithospheric layer and below by an asthenospheric layer (e.g. Yuan et al., 2002, for the Central Andes; Tassara et al., 2006), and both have a density of  $3350 \text{ kg/m}^3$ , corresponding to a fresh harzburgite or a fertile lherzolite. Power-law creep parameters of wet olivine are assigned to the upper lithospheric part and to the first 200 km east of the subduction zone below the arc, and dry olivine parameters are assigned elsewhere. In addition, the depth to the  $1300^\circ\text{C}$  isotherm is chosen equal to either 120 km or 160 km, following the Lithosphere-Asthenosphere Boundary range obtained by Tassara et al. (2006) for South America. An effective friction angle is also assigned to these mantle layers as a cutoff value (e.g. Watts & Burov, 2003): we set friction to  $10^\circ$  and  $20^\circ$ , respectively for the continental and oceanic mantle lithospheres.

The oceanic lithospheric mantle is assigned a mean thermal age of 40 Myr and a constant density of  $3360 \text{ kg/m}^3$ . This value corresponds to that determined by Tassara et al. (2006) down to about 100 km depth along segments of the Nazca plate facing Chile, based on age distribution (<50 Myr) and observed long-wavelength Bouguer anomaly. For comparison, Sobolev et al. (2007) chose  $3280 \text{ kg/m}^3$  and  $3300 \text{ kg/m}^3$  respectively for the oceanic mantle and asthenosphere. The oceanic crust is assigned power-law creep parameters of a mafic granulite, and a density of  $3050 \text{ kg/m}^3$ , a high value according to classical worldwide averages, but a lower value than that determined by Tassara et al. (2006) for a mixture of 85% fresh basalts and gabbro and 15% fully metamorphosed basalts. This crust together with the oceanic mantle composes a lithosphere that remains buoyant with respect to the underlying asthenosphere. This buoyancy may appear wrong compared to classical estimates such as by Cloos (1993). However, new calculations by Afonso et al. (2007) indicate that although the density structure of oceanic lithosphere makes it statically unstable after 35 Ma, depth-averaged density of a mature oceanic plate including a 7 km thick crust remains close to  $\sim 3315 \text{ kg/m}^3$ , never exceeding that of the underlying mantle. The magnitude of the slab-pull force is controlled by the excess weight of the Nazca plate at greater depths where its age increases, all the more that it continues way beyond the 660 km discontinuity (e.g. Li et al., 2008).

We now define the subduction interface channel. It is set 12 km thick, chosen first as a proxy for the  $\sim 10$  km resolution of seismic tomography at the 100 km scale of subduction zones, and second for the “numerical” reason that at least 4 elements are necessary to form a numerical shear zone (Sobolev & Babeyko, 2005; Sobolev et al., 2007, also chose 12 km). Densification with depth results from metamorphic reactions that occur with increasing temperature and pressure during subduction, and involves hydration and dehydration processes (e.g., Peacock, 2000; Hacker et al., 2003a,b; Rondenay et al., 2008). We do not go into the complexity of simulating these reactions, also because the specific structure of the Chilean subduction interface is not well known. This interface channel is thus here defined by 3 layers of different densities. The top 10 km of unconsolidated material are simulated with a plagioclase rheology of density  $2800 \text{ kg/m}^3$  and a friction angle of  $5^\circ$ . The density of this channel increases with depth, it is  $3200 \text{ kg/m}^3$  from 10 to 100 km depth, and  $3400 \text{ kg/m}^3$  from 100 km depth downwards, rendering the oceanic plate negatively buoyant with respect to the underlying asthenosphere.

The continental crust is composed of upper crust, of density  $2750 \text{ kg/m}^3$ , and lower crust, of density  $3050 \text{ kg/m}^3$ ,

an average between the high values evaluated by Tassara et al. (2006) for the Andes, and worldwide observations (e.g. Christensen & Mooney, 1995). The crust can be very complex in its superposition of layers of different composition and strength, and our aim here is not to reproduce all behaviors. Therefore, we consider only two extremes, a weak upper and weak lower crust (WR), as opposed to a strong upper and a strong lower crust (SR). Specific creep parameters are given in the next section (also Table 1). Effective friction is set to  $20^\circ$  for the upper crust, and  $10^\circ$  for the lower crust.

At the top-free surface, initial topography is deduced from the assumption of Airy isostasy at 300 km depth (implying that deeper, lateral density variations are negligible). This simplified assumption of the 300 km limit of the model is consistent with studies indicating that below, mantle viscosity increases by about an order of magnitude (e.g. Royden & Husson, 2006). No significant differences occurred when comparing with a test model extending down to 410 km depth. A diffusion law simulates erosion and sedimentary processes (Table 1), which are known to participate in the long-term mass balance that controls orogenic growth (e.g. Avouac & Burov, 1996; Montgomery et al., 2001).

Plate reconstructions between South America and the Pacific oceanic plates (Somoza & Ghidella, 2005) reveal a 26-0 Ma stage corresponding to the subduction of the Nazca plate beneath most of the South American trench, marked by a convergence rate reaching 15 cm/yr which gradually decreased during the last 10 My, and a rather constant direction of convergence. A convergence velocity of 7.5 mm/yr is obtained for the last 4.9 Myr at  $22^\circ\text{S}$  (Somoza & Ghidella, 2005), which the authors find consistent with a linear decrease down to present-day velocities measured by GPS (Norabuena et al., 1998). Somoza & Ghidella (2005) also argue that the correlation of South America's motion with respect to various hotspots (including the Indo-Atlantic reference frame from O'Neill et al., 2005) and the evolution of the Central Andes remains ambiguous to the order of at least 1 cm/yr, whatever references frames of mantle plumes are chosen. Consequently, we choose O'Neill's et al. (2005) prediction of South America's westward velocity close to 2.5 cm/yr. The slab-pull velocity is then equal to 5 cm/yr, oriented parallel to the slab and where it intersects the bottom of the model domain at 300 km depth. Lithostatic compensation is applied on the other boundaries of the model (**Fig. 2**).

The choice of a traction applied to the slab is justified by the occurrence of “classical” downdip-tension from ~100 to about 300 km depth, measured along the “normal” sections of the Nazca slab (Fig. 1, Chen et al., 2001), which is also obtained in modeling studies (e.g. Funicello et al., 2003b; Goes, pers. Comm.). Second, the fixed direction of this traction, although disputable, is justified by the fact that the Nazca slab is likely to be anchored in the lower-mantle (e.g. Conrad & Lighthow-Bertelloni, 2002; Li et al., 2008), so that any modification of its motion at ~300km depth and on the time-scale of 4 Myr of our models, should not be significant. In order to quantify the artifacts induced by the initially bent geometry of the slab as well as by initial temperatures, a test-model is ran in which these initial velocities are set to zero, together with zero density contrast of the subducting channel. This model first shows induced velocities of 3 mm/yr that reduce to 1.5 mm/yr within 2 My. Second, it shows that within 4 Myr, mean temperatures within the subducting channel (where highest gradients occur) evolve from initially about  $4^\circ\text{C}/\text{km}$  initially to  $6^\circ\text{C}/\text{km}$  down to 100 km depth.

### 4.3 Defining the parameters to test

Deformation patterns in the upper plate typically result from the combination of horizontal compression induced by upper plate advance, and counterclockwise shear drag induced by the descending slab. Out of more than 200 tests, the parametric space capable of generating significant differences in deformation patterns was reduced to three significant parameters :

- A first parameter is the rheology of the continental crust, which controls the location of shortening and the occurrence of decoupling stress levels. The 35 km thick crust is thus set, either weak (WR) with creep parameters of wet granite and plagioclase for the upper and lower crust, respectively, or, in contrast, strong (SR), with plagioclase and mafic granulite creep parameters for the upper and lower crust, respectively (Table 3, 4). Section 5.2 and Fig. 3 show how variations in the position of a weak crust control the location of deformation zones..
- A second parameter is the strength of the subduction interface, which tends to drag upper plate material downwards along the subduction zone. While in reality this channel strength results from a complex combination of metamorphic reactions with fluids and material belonging either to the upper plate or to the downgoing plate, in our models instead this channel strength is approximated, either strong (SF, models in Fig. 3) with plagioclase creep parameters and a friction angle of  $2.86^\circ$  (corresponding to the highest friction coefficient of 0.05 tested by Sobolev & Babeyko, 2005), or weak (WF, models in Fig. 4) with wet granite creep parameters and a friction angle of  $1.43^\circ$  (consistent with the coefficient of 0.025 proposed by Lamb, 2006). By accounting for a self-consistent brittle-ductile behavior, we take a step further in modeling the behavior of the subduction shear zone, in comparison to models that assume either a constant friction (e.g. Hampel & Pfiffner, 2006) or a constant viscosity (e.g. Medvedev et al., 2007, section 2.2). Results are discussed in section

5.3. Note that Sobolev & Babeyko (2005) and Sobolev et al. (2007) use a similar self-consistent approach but only vary friction angles.

– The third parameter is the temperature “anomaly” of the fore-arc mantle below the active arc, which controls its strength and its capacity to localise deformation. Whereas a cold mantle fore-arc sustains stresses and transmits slab drag into the overriding lithosphere, a warm fore-arc mantle has the opposite decoupling effect (section 5.4, Fig. 4). Here too, we oversimplify real thermal processes at the fore-arc, resulting from complex solid and fluid mass transfers and metamorphic reactions occurring over millions of years. We define a Gaussian perturbation added to the background geotherm, with a geometry that is as close as possible to the temperature profiles evaluated by Springer (1999) and Tassara et al. (2006), along the Chilean margin. This Gaussian is centered at  $X=300$  km and varies both in vertical and lateral extents (WT defines a wide and strong anomaly, CT defines a narrow extent). Results are discussed in section 5.4. An additional model will illustrate the effects of releasing the slab-pull force, section 5.5.

## 5. Parametric study

### 5.1 Generic results

We first describe the main stress and strain characteristics that develop in all models, for which specific results are shown in the next sections, and sketched **Fig. 5** :

- As the oceanic lithosphere is dragged downwards by the applied basal velocity, flexural stresses develop at its pre-bended geometry, and tensional stresses develop parallel to the dipping direction from about 100 km depth. Tensile brittle shear zones develop at the oceanic fore-bulge, down to about 20 km depth. This is consistent with observed outer-rise seismicity along the Chilean margin (e.g. Clouard et al., 2007).
- The subduction channel displays a brittle behavior with increasing shear stress until the brittle-ductile transition depth, which is self-consistently determined between 30-60 km, according to different models (**Fig. 6**). This depth range and the associated shear stress are discussed in detail in section 5.3, and compared to estimates from previous studies.
- The mantle in intraplate domains is the strongest layer of the continental lithosphere, as is shown here in the undisturbed domain to the east where “normal” temperatures prevail (**Fig. 7c**, right column, commented in section 5.2). Approaching the subduction zone, this continental mantle is weakened by the predefined arc thermal anomaly, and intense shear deformation and corner flow develop at the contact with the descending plate. This flowing mantle wedge connects with shear zones that develop through the lithosphere up to the surface.
- The crustal domain, from the trench to the main mountain range (the “arc” domain), deforms into one or more pop-up blocks, whose extent are controlled by the location of rheological discontinuities at depth, and by their intersection with the subduction interface (the Moho and brittle-ductile transitions). These intersection points at depth are also combined with horizontal decollements in weak ductile levels. The horizontal length-scale of a pop-up block at the surface relates to the depth at which it roots by the classical orientation of conjugate faulting at angle  $\pi/4 \pm \phi/2 = 30^\circ$  from the mainly horizontal compressional stress. Crustal buckling also develops in some models.
- Note that in the models, the trench-to-coast distance, initially equal to about 150 km, shortens to about 50 km, whereas along the Chilean margin, this distance is of about 100 km. This discrepancy can be attributed to the initial loading stage in which the highly deforming trench zone shortens exceedingly in comparison to the present-day real domain that has “equilibrated dynamically” since tens of million of years of subduction.

### 5.2 The effect of crustal rheology

This first series of 4 models include a “strong” subduction interface (SF). Results are displayed **Figs. 3**, and sketched on the top line of **Fig. 5**. Each one of these 4 models has a different geometry of crustal strength.

The first model (M1, **Fig. 3** top-left) has a strong continental crust (SR): crustal deformation develops in the arc domain where the initial thermal anomaly was set. Naturally the location, width and intensity of this localised deformation depend directly on the location, width and intensity of the initial thermal anomaly. A crustal-scale pop-up forms, about 60 km wide, bounded by symmetric conjugate shear zones in a “Y” geometry, and producing a topographic maximum. Diffuse faulting also affects the forearc and foreland domains (top-left **Fig. 3b**). At the tip of the forearc, a smaller scale east-verging shear-zone accommodates sliding along the subduction channel, and is linked to a second topographic bulge. The entire crust and upper mantle sustain relatively high shear stress and are under compression, except in the corner mantle wedge where tensional stresses develop, induced by the slab's downward pull (top-left, **Fig. 3c**). This model is consistent with localised deformation in the magmatic arc domain in the Southern Andes (e.g. Hervé,

1994; Cembrano et al., 2002).

The second model (M2, top-right **Fig. 3**) has a wide weak continental crust extending initially 300 km eastward (WR) from the trench. Shortening concentrates in this weak domain, forming a widespread thickened crust, still enhanced at the location of the thermal anomaly. Crustal conjugate shear zones are now multiple and root in the ductile lower crust decollement. The entire domain of weak crust displays negligible shear stress (top-right **Fig. 3c**). This model explains that the increasing eastward width of the Andes, from South to North, can be attributed to the presence of weak crust over an increasing width (e.g. Allmendinger et al., 1997). Differences in amounts of shortening between models are discussed in section 6.2.

The third model (M3, **Fig. 3** bottom-left) has a narrow weak crust, extending from the trench over only 150 km inland (WRfore). Consequently, shortening is accommodated over a shorter distance with respect to model M2, and induces greater thickening in the fore-arc domain. Eventually the fore-arc lower crust subducts and disappears into the mantle along the subduction channel (Cartoon **Fig. 5**). This process could explain the loss of fore-arc crust by episodic subduction erosion as has been suggested for the margin of Central and Northern Chile (e.g. Kay et al., 2005), and/or controversial proposals of about 2 km of fore-arc subsidence in the Late-Miocene, deduced from sedimentological and paleontological marine sequences in south and central Chile (Encinas et al., 2008).

The fourth model (M4, **Fig. 3** bottom-right) has a weak crust located inland 200 km away from the trench (WRback), over 200 km. The strong fore-arc then buckles and indents the weak crust to the east. This indentation produces a crocodile geometry: the weak upper crust over-thrusts the fore-arc with a west-verging fault, whereas the weak lower crust under-thrusts the fore-arc along a westward descending shear zone (Cartoon **Fig. 5**). This typical fault structure is observed in collision zones such as the Southern Alps of New Zealand, the Svecofennian orogen, or the Eastern Alps (references in Gerbault & Willingshoffer, 2004). It may be compared to the Northern Chile fore-arc indenter of Tassara (2005), and to observations of fault vergence and geophysical images in the western border of the Precordillera of northern Chile (e.g. Farías et al. 2005; Victor et al., 2004).

### 5.3 Test of a weak subduction interface

The next series of 4 models have a weak subduction interface (WF, friction angle reduced to  $1.43^\circ$  and wet granite creep parameters, among the weakest of rocks tested in laboratories). Crustal rheology is then varied in 4 different ways similar to the previous series of models (see Table 4). In order to limit the number figures, only models M7 and M8 are shown (top-left and top-right of **Figs. 4**), but all models M5 to M8 are sketched in the middle-row of **Fig. 5**.

Now, relative plate convergence is accommodated by easier shearing along the subduction interface, and less shortening of the upper plate in comparison to previous models. Also, because interplate shear-stress is lower, weak fore-arc crust is not dragged anymore along the subduction zone. Instead, light material from the subduction channel migrates upwards, suggesting magma and/or hydrothermal fluid ascent (yellow layers of **Figs. 4d**). It is important to recall that magmas at convergent margins are largely the result of mantle wedge melting as triggered by dehydration of the subducting slab, a process that is only very approximately represented in our models by these ascending “light material plumes”. As these “plumes” rise, they occupy the mantle domain into which the crust would thicken otherwise (first series of models). The proximity of this buoyant material with thickened crustal root invokes the occurrence of mixing processes, and the possible formation of crustal pathways taken by fluids towards the surface, using already active mechanical shear zones. In the models, the depth at which this material ascent occurs and the geometry with which it rises, are controlled by the depth of the 600-700°C isotherms in the fore-arc mantle, which determines the olivine brittle-ductile transition.

**Figure 6** displays the shear stress, viscosity and temperature along the subduction channel, which can be compared to previous estimations by Yañez & Cembrano (2004) or Medvedev et al. (2007). These estimations give a (depth-constant) viscosity of the subduction channel in the range of  $4.5 \cdot 10^{20}$  Pa.s to  $3 \cdot 10^{20}$  Pa.s, associated with a shear stress of 20-95 MPa (Yañez & Cembrano, 2004). Our models provide depth-variable viscosities that bracket this range. Model M1 with a strong interface provides viscosities close to  $10^{21}$  Pa.s and shear stress greater than 60 MPa, at 30-60 km depth. Models MH8 and MH6 with a weak interface provide viscosities below  $10^{20}$  Pa.s and a shear stress below 20 MPa, at similar depths.

**Figure 6c** shows minimum and maximum temperatures inside the subduction channel. Above 60 km depth, minimum temperatures in the subduction channel are below 300°C. Maximum temperatures however, can be 200°C higher (dashed lines in **Fig. 6c**). Rising buoyant material in weak interface models can carry temperatures >700°C to 50 km depth

(model MH8). Modeled temperatures at ~100 km depth along the interface, show values as low as 400°C, but maximum values reach 800°C at lower depths, on the outer contact with the continental mantle. These temperatures are similar to those estimated by Van Keken et al (2002) along the subduction interface.

In our models, the brittle-ductile transition occurs at 35-55 km depth, which is close to evaluations of seismogenic depths along the Chilean margin. Along the Chilean subduction interplate, published estimates provide a rather consistent seismogenic depth of approximately 50 km along the entire margin (e.g., Tichelaar & Ruff 1991; Pardo et al. 2002), but estimates of currently locked segments using GPS surface velocities vary from about 33 km depth in the north to about 52 km depth in the south (e.g. Khazaradze & Klotz, 2003). In contrast, estimates of the effective coefficient of friction indicate significantly greater values in the north than in the south (e.g. Wang & He, 1999; Oleskevich et al., 1999; Lamb & Davies, 2003; Lamb, 2006). Hoffmann-Rothe et al. (2007) observe that these inversely evolving properties along the Chilean margin may compensate each other, and that coupling forces remain more or less constant on a regional scale. Sobolev et al. (2007) attribute these inversely evolving properties to the deeper brittle-ductile transition that logically results from a brittle yield slope of lower friction angle.

By accounting for a self-consistent brittle-ductile behavior, the present models allow one to question the behavior and composition of the subduction interface channel in reality. Our “strong interface” models tend to overestimate seismogenic values, whereas our “weak interface” models tend to underestimate them. From these models, we propose that whereas quantification of strength at shallow depth remains appropriately described by the traditional friction concept, seismogenic depths could be better described by ductile creep properties. Then, plagioclase-like compositions would overestimate high viscosities at deep depths, despite an a priori dominantly mafic composition of real subduction channels. However, the brittle-ductile transition in real subduction channels is also very probably controlled by the amount of fluids produced by hydration and dehydration reactions in the subducting slab (e.g. Hacker et al., 2003, Rondenay et al., 2008). Therefore, alternatively to the proposition of Sobolev et al. (2007), we propose that the dominant rheological composition of the subduction interface along the Chilean margin, while probably becoming drier to the north with respect to the south (related to climate-driven sediments to the trench, e.g., Von Huene et al., 1999), also becomes less mafic, an observation that is consistent with differences in upper plate crustal composition between north and south Chile (Tassara et al., 2006). Indeed, the composition of the subduction interface should not only be determined by lubricating processes, but also by the composition of the overriding plate. Further high resolution seismic imaging is needed to test this hypothesis.

#### 5.4 Test of Temperature

Models with higher temperatures are then tested (both the background geotherm and the arc anomaly are modified). They produce generally wider crustal deformation, because of temperature-dependant enhancement of creep versus brittle behaviour. Table 4 displays the characteristics of four models (models MH\*). We show the results for only two extreme models, one with a strong crust (MH5, SR- WF-WT, bottom-left **Fig. 4**) and one with a weak back-arc crust (MH8, WRback-WF-WT, bottom-right **Fig. 4**).

Higher temperatures weaken the subduction channel and fore-arc domain, and facilitate the rise of buoyant material. A somehow paradoxical result occurs for models with a weak forearc crust: the forearc crust thickens so much that temperatures are advected downwards, and the fore-arc cools down (e.g. model M7).

**Figure 7** displays shear stress, viscosity and temperature, along three vertical sections through the forearc, arc, and undisturbed continental domains. For sake of clarity we plot only two extreme models, M1 which is cold, and MH8, which is warm (see undisturbed initial isotherms on bottom-right plot, and Table 4 for parameters details). We see that Moho temperatures at ~35km are only about 400°C in the forearc domain (column **Fig.7a**). The subduction channel at 45-70 km depth, shows viscosities greater than  $10^{20}$  Pa.s, and temperatures of 300-500°C. Below the arc (column **Fig.7b**), Moho temperatures exceed 800°C at 65 km depth for model M1, and at 40 km depth for model MH8 (the thicker weak crust, however, lies deeper further east, see **Fig. 4**). Buoyant material rising below the arc in model MH8 has a viscosity below  $10^{20}$  Pa.s., and temperatures locally close to 1000°C. In the undisturbed continent (column **Fig.7c**), the Moho remains at 35 km depth with temperatures of 400°C and 500°C, and the lithospheric mantle has a viscosity greater than  $10^{22}$  Pa.s. down to 60-70 km depth, respectively for models M1 and MH8.

Models in which temperatures are cold, provide interesting results when the overriding crust is set strong (not shown), a situation that better corresponds to ocean-ocean subduction systems. In this case, the entire crustal forearc can be underthrust by the main continental crust, with significant surface subsidence. A crustal scale west-verging shear zone bounds the main cordillera parallel to the subduction zone. The entire fore-arc slice may eventually be entrained downwards coupled to the subducting slab, and trigger a “subduction jump” inland, as modeled by Boutelier et al. (2003).

## 5.5 The effect of cessation of subduction

Numerous models have explored the modes and consequences of slab break-off. Andrews & Billen (2009) distinguish 2 types of deep or shallow slab detachment, produced either by heating from the surrounding mantle in case the slab is stiff, or by internal yielding in case the slab is relatively weak. Furthermore, numerical and analog experiments show slab break-offs around 250 km depth (e.g. Sobolev et al., 2007; Regard et al., 2008). Although it is not our aim to study slab-detachment specifically in this paper, we aim at showing what happens with our typical model setup when slab-pull velocities are removed, which approximates the consequences of a “deep” slab break-off (**Fig. 8**, model M<sub>10</sub>). This model has thermo-mechanical properties identical to model M1. As the slab-pull related stresses vanish, a huge amount of buoyant material (yellow layer) can rise below the arc, enough to generate significant vertical stress on the upper crust that changes the regional convergence regime. The initially horizontal most compressive principal stress ( $\sigma_1$ ) becomes vertical, the crust thins, and back-arc opening develops over a distance of about 100 km within 2 Myr.

This model shows that back-arc opening may occur when relative plate convergence rate diminishes, even under constant upper-plate motion (only the slab-pull force reduces). This scenario could have occurred sometimes in the past, perhaps explaining the formation of extensional intra-arc basins both in northern Chile (i.e. Calama and Atacama basins, Pananont et al., 2004; Blanco, 2008), and the volcano-sedimentary Abanico basin in Central Chile (e.g. Charrier et al., 2002), extending southward probably down to 45°S along the forearc (Jordan et al., 2001), and which developed when the rates of plate convergence decreased dramatically in the Oligocene.

## 6. Comparative discussion

### 6.1 Forearc tilting and subsidence

Fore-arc deformation shows decoupling from the underlying lithosphere deformation, depending on the existence of crustal ductile levels. **Figure 9** displays a detail of fore-arc deformation for models M1, M4 and MH6 (see Table 4 for parameter details). We plot instantaneous rotation values, with the reference color scale given for example by the top of the subducting slab showing clock-wise rotation values (orange) due to bending, and by the subducting channel showing counter-clockwise rotations (blue) due to its sense of shear. While these instantaneous values change from one snapshot to the other (like strain-rate), general features are detectable. The extreme edges of the forearc show counter-clockwise rotation and tension at the western most tip in contact with the plate interface, and clockwise rotation under compression on the eastern flank of arc topography. In models M1 and M4 which have a strong fore-arc crust (**Fig. 9a,b**), vertical deformation within the fore-arc shows clockwise motion in the western-coastal side (like entrained and coupled with the subducting slab), combined with clockwise “up-to-the-east” uplift on the eastern flank of arc topography. In models displaying ductile levels in the forearc like MH6 (**Fig. 9c**), vertical deformation decomposes into several smaller blocks which alternate in up-to-east and up-to-the-west tilting.

In model M4 (**Fig. 9a,b**), relative subsidence occurs in the central fore-arc domain. Whereas the forearc is under a general uplift and compressional regime, a subsiding basin develops approximately 250 km east of the trench, dividing an outer-forearc coastal high, and the inner-forearc that borders the main arc relief. The shear-zone that accommodates these two differential movements should display complex relative senses of shear, not only in reverse mode, but also in apparent normal modes.

These processes (tilting and subsidence) suggest an analogy with the Central Depression and Central Valleys of the northern and central Chile forearc, and possibly also with the present-day compressional subsidence of the Salar de Atacama basin (e.g. Arriagada et al., 2006). However, the precise tectonic nature of the boundary between the Coastal Cordillera, the Central Depression, and the Main Cordillera remains poorly constrained. The eastern limit of the Central Valley from 33°S to 35°S, is currently regarded as an eroded fault system made up of Miocene to Pliocene strike-slip and west-verging reverse faults (e.g. Cembrano et al., 2002; Fock et al., 2006; Rauld, 2002). According to Farías et al. (2008), the Coastal Cordillera in Central Chile (2 km high at the latitude of Santiago) uplifted about 1 km less than the Main Cordillera since the late Pliocene, with a maximum of 600 m of relative subsidence of the Central Depression, and ~300 m of coastal uplift.

Farther North from 18.5°S to ~27°S, the Atacama fault system bounds two longitudinal structural domains (Cembrano et al., 2007) : the outer forearc domain comprises the Coastal Cordillera which exhibits east–west directed extension and an uplifted landscape, while the inner forearc and magmatic arc domain (comprising the Central Depression, the Precordillera and the Western Cordillera) is characterized by uplift in a compressive regime exhibiting, at

least in the Arica-Antofagasta region, high-angle reverse faulting and minor reactivation of older structures. While east–west extension is attributed to processes of tectonic erosion and upper plate bending, uplift would result from underplating of crustal material and progressive indentation of the South American margin by the Nazca plate (e.g. Cembrano et al., 2007 and references therein). Compression in the inner forearc and magmatic arc domain (Precordillera and Western Cordillera), in turn, would correspond to shortening due to convergence. These interpretations are consistent with our models, which show how all these processes combine.

## 6.2 Crustal shortening

Upper plate shortening in the models can be estimated in two ways. The simplest way is to take the difference in length of the continent, from the eastern edge of the model to the trench between 0 and 4 Myr, which we call “total upper plate” shortening (this value is given for each model at the bottom of Figs 3 and 4). Another interesting method, because it may better relate to field estimates, is to sum up the amount of horizontal strain ( $\epsilon_{xx}$ ) in each element of the top of the emerged the continent in the models (which we call “top continent”). It may be interesting to compare these two amounts of upper plate shortening, shown **Fig.10**.

Models with a strong interface (M1 to M4, **Fig. 10a**) show “top continent” shortenings of 100 to 150 km, and “total upper plate” shortenings of 150 to 200 km. This ~30% difference in amounts of shortening shows that total shortening is not only horizontal, but has a strong shearing component (exerted from the subduction interface).

Shortening increases with increasing width of weak crust, with M1 and M5 displaying less than 100 km of emerged continent shortening, while M2 and MH6 display about 50 km more. Trench motion is directly correlated with total upper plate shortening, and is thus also sensitive to the rheology of the upper plate, with about 50 km more landward advance for models with a weak fore-arc crust (M2 and M3) than for models with a strong crust (M1). We verify that total upper plate shortening (160 to 200 km for models M1 to M4) closely equals the sum of overriding plate motion (100 km within 4 Myr) and trench motion (60 to 200 km for models M1 to M4).

To a second order, models with a weak interface display smaller amounts of shortening (**Fig. 10b**), simply because relative plate convergence is more easily accommodated by shearing along the plate interface. Logically, differences between “top continent” shortening and “total upper plate” shortening are smaller, because of less shear stress at the subduction interface inducing less shear deformation in the upper plate. Trench landward motion is equivalently reduced with a weak plate interface.

In the models, the first Myr of total shortening (and trench advance) is due to elastic loading prior to failure, and should thus be subtracted to “steady state” values. This initial shortening is roughly estimated by evaluating the horizontal strain ( $\epsilon_{xx}$ ) necessary to reach the failure yield stress in the crust. First, we relate principal stress and strain with Hooke's law  $\sigma_i = 2G \cdot \epsilon_i$  (in plane strain with  $\nu = 0.25$ ,  $G = 5 \cdot 10^{10} Pa$  is bulk modulus, and  $\epsilon_{xx}$  is approximated to  $\epsilon_1$ ). Then, the failure yield state allows one to relate the principal stresses with the friction angle ( $\phi = 20^\circ$ ), neglecting cohesion and approximating the vertical principal stress  $\sigma_3$  with the lithostatic stress  $\rho gh$ ,  $\sigma_1 = (1 + \sin\phi) / (1 - \sin\phi) \sigma_3 \sim 2 \cdot \rho gh$ . The brittle crustal thickness  $h$  reaches 10–30 km depth. All numbers accounted for, we obtain an estimate of the required amount of shortening of about  $h$ , thus 10–30 km (see Gerbault, 2000, for example, for more details). This effect of elastic loading needs to be subtracted when making any comparison with the real Andes, for which this stage of elastic loading has occurred a long time ago, when compression started.

In agreement with previous modelling studies, our models allow one to simply attribute different amounts of measured shortening and different styles of deformation between Northern (north of  $33^\circ S$ ) and Southern (south of  $33^\circ S$ ) Chile, to differences in overriding crustal strength and in strength of the subduction interface. Little total shortening such as in models M5 and MH5, which have a strong fore-arc and a weak subduction interface, is consistent with the Southern Andes being set in a context of a resistant (plagioclase-rich) fore-arc block, aside a weaker quartz and fluid-rich arc domain, and a “weak” subduction interface lubricated by high-erosional sediment input (e.g. Lamb & Davies, 2003). In contrast, high amounts of shortening like in model M2 or M4 with a strong fore-arc and a strong subduction interface, better correspond to the Central Andes, which are characterised by a sediment-starved deep trench generating large earthquakes, a rigid fore-arc, and highly deformed arc and back-arc domains (e.g. Von Huene & Ranero, 2003; Lamb, 2006; Tassara et al., 2006).

## 6.3 Modelled structures at depth

The present models show that upper plate shortening is not uniformly distributed at depth in the continental lithosphere. In other words lithospheric deformation is generally in a “simple shear” mode rather than a “pure shear” mode (according to the terminology used by e.g. Allmendinger & Gubbels, 1996). The following remarks compare the

conditions in which the models develop specific deep structures (e.g. the crustal root), with what is known of the Andes:

- Only models with a uniform overriding crust, either strong or weak, actually produce a “pure shear” mode of plate shortening, centered at the thermally weakened forearc. In other models, crustal “delamination” sensu lato occurs: subduction erosion occurs in cases with a weak the fore-arc crust and a resistant interface. In other cases, the thickened crust below the arc is dragged downward by mantle corner flow. The short time-scale of our models (4 Myr) impedes the development of Rayleigh-Taylor gravitational instabilities, which require typical time-scales of about 10 Myr. Such delamination events seem to have occurred under the Andes, especially below the Altiplano-Puna plateau (e.g. Kay & Kay, 1993; Sobolev et al., 2007; De Celles et al., 2009).
- In models with a weak interface, the rise of weak and buoyant material “plumes” from the subducting channel is in many ways an oversimplification of real fluid-related processes developing at the ~50-100 km depth range. Their location and “volume” in the model depend on initially chosen density contrasts and temperatures of the subduction channel and asthenospheric wedge. These modeled plumes only give a qualitative geometrical picture of their interaction with the thickened crust, and the reader is directed to more precise patterns reproduced by other numerical studies that especially focus on this subject, such as Gerya et al. (2003, 2006) and Gorczyk et al. (2007).
- In most of our models, the forearc mantle wedge generally remains cold, often less than 1000°C at 100 km depth (**Fig. 7**), which is cold for the initiation of partial melting. While cold conditions are likely to be valid in Northern Chile where numerous geophysical data indicate cold fore-arc mantle conditions (see sections 2.2 and 6.4), they may not apply in all other sections of the Chilean margin. Other thermo-mechanical models of mantle wedge corner flow usually obtain higher temperatures (e.g. Conder, 2005; van Keken et al., 2002; Arcay et al., 2006), but they apply to high heat flow subduction zones. Warm conditions were also proposed by Tassara et al. (2006) along the Chilean forearc, with the 1300°C isotherm lying as shallow as 60 km depth. Nevertheless, Stern (2002) argues that a shallow dipping subduction zone, as found below Chile, restricts flow of asthenosphere in the mantle wedge, and diverse combined data all indicate that the arc domain should remain cooler than the surrounding mantle. More advanced models that account for long-term thermal and fluids flow processes will better determine these thermal conditions.
- Crustal thickening in our models localises at the arc thermal anomaly and in domains of weak crust. A weak crust located in the forearc produces west-verging lower crustal shear zones on the western flank of the Main Cordillera, whereas a weak crust located in the backarc produces east-verging lower crustal shear zones on the eastern flank of the Main Cordillera (cf. Cartoons **Fig. 5**). In addition, this crustal thickening and the axis of the orogeny may migrate westward, if the forearc domain is weak enough to shorten. If instead, the forearc is strong, the axis of orogenic growth remains approximately at the same place, with respect to the trench.
- The magmatic arc in Northern Chile has systematically migrated to the east from ~160 Ma to the last 20 Ma, and has been attributed to subduction erosion. Letting aside “external causes” such as thermal heating of the mantle wedge or changes in slab dip, fore-arc shortening (and subduction erosion) is a scenario consistent with our models containing a weak forearc crust and a strong channel interface. If this scenario has indeed occurred along portions of the Chilean forearc where missing crust is observed (e.g. Kay et al., 2005), the causes for relative forearc crustal weakening with respect to strengthening of the subduction interface remains to be explained. To the South instead, the magmatic arc has remained relatively static since the Jurassic, only changing its width. A static arc associated with little shortening means that the subduction interface remains weak relative to the upper plate strength, and accommodates most of the plate convergence on the long-term.

## 6.4 A specific model for the Altiplano

The Altiplano area is now modeled by incorporating rheological and thermal parameters that best fit known geophysical data (see section 2). These geophysical data are discussed in detail in Dorbath et al. (2008), and may perhaps summarize in the statement of exceptionally low heat flow values measured both in the oceanic lithosphere and in the fore-arc (Springer & Forster, 1998; Springer, 1999), as opposed to high values in the Altiplano domain. The initial conditions, similar to those presented in Dorbath et al. (2008) resembles our generic case M4, but the oceanic and continental geotherms are both set colder, and the Altiplano domain is defined by a relatively weak pre-thickened lower-crust and a wide thermal anomaly (Model  $M_{ap}$ , **Fig. 11**, Table 4). An additional difference with previous models is the inclusion of asymmetric surface processes, which favor the development of deformation on the side of the orogen that is subjected to a more active climate, a process that has long been recognized to control the development of the Central Andes (e.g., Hoke et al., 2007; Garcia-Castellanos, 2007, and references therein). Arid vs. tropical climatic conditions are thus modeled by choosing a coefficient of diffusion 100 times lower west of the Eastern Cordillera, than to the east.

In the model after 4 Myr, deformation bypasses the forearc, forms an east-verging structure departing from the



subduction interface at ~100 km depth, propagates as a horizontal shear zone in the warm and pre-thickened lower crust of the Altiplano (corresponding to TRAC1 in Yuan et al., 2002), and reaches the surface in the eastern SubAndean superficial crust (**Fig. 11**). This pattern of deformation has been previously proposed by Tassara (2005) or Farias et al. (2005). Seismicity shown in **Fig. 11** is from relocated data by Dorbath et al. (2008) at 18-19°S. The east-verging straight seismic pattern that links the subduction interface to the Western Cordillera at the surface (also shown by David et al., 2005) is located within the 400-600°C isotherm range in the crust, which corresponds to brittle-ductile transition conditions for mafic crust. This geometry corresponds to a first-order mechanical feature, which constitutes a lithospheric-scale shear-zone conjugate to the subduction channel shear-zone, and which delimits the fore-arc lithospheric wedge. The similar geometry and depth of this seismic structure in other subduction zones such as Peru, central America, and Japan, suggests that it may also be linked with the blueschist-eclogite pressure-dependent transition.

Stresses in the fore-arc are high because it is cold and resistant, which is consistent with geophysical observations of low  $Q_p$  attenuation and specific SKS splitting directions (e.g. Schurr & Rietbrock, 2004; Kneller et al., 2007). In our models, the fore-arc is submitted to both horizontal compression and slab-pull downward tension. The compressional state of the upper plate transmits into the subducting plate and explains the occurrence of double-plane seismicity along the subduction zone, despite the relatively young age of the Nazca plate (see discussion in Dorbath et al., 2008).

### 6.5 Rheological controls on slab flattening

A final model now illustrates the effect of basically switching the rheological properties of the oceanic and continental mantle. Now, the oceanic lithosphere is warmer and weaker, assuming it is hydrated and behaves with wet olivine creep, whereas the continental mantle is simulated colder, with dry olivine creep parameters. The brittle yield of both oceanic and continental mantle are now set equal, with friction angle at 8° and cohesion at 20 MPa (thus the brittle yield is lower than in previous models for the oceanic plate, and higher than previous models for the continental mantle). The oceanic geotherm is set warmer by assuming a younger plate age, arc thermal anomaly is also reduced ( $M_{fl}$ , Table 4).

Justification for such drastic rheological changes may in fact be found at the latitudes of the “Pampean” flat-slab. On the one hand, both geological and seismological data give indications of a dry and cold continental mantle above the flat-slab segment in Argentina (e.g. Wagner et al., 2005; Ramos, 2002, see section 3.3). With time, the subduction zone approaches the more resistant core the South America, and may eventually face a relatively strong overriding plate. On the other hand, the incoming Nazca mantle at this latitude may be hydrated. Previous studies vainly focused on the crustal properties of the Juan Fernandez ridge in order to explain slab-flattening by additional crustal buoyancy (Yañez et al., 2002; Van Hunen et al., 2004). Seismic profiles at latitudes 32-33.3°S (Kopp et al., 2004), indicate abnormal crustal thickness of only 1 km, but also indicate reduced crustal and upper mantle velocities. Kopp et al. (2004) interpret these widespread low velocities as hydrated lithosphere due to hot-spot magmatism in the pervasively faulted region of the Challenger Fracture zone. Independently, Stern (2002) argued that hydration of oceanic mantle lithosphere can occur when formed either at a relatively slow spreading ridge, or coeval with the development of deep fracture zones. Here, the Challenger Fracture zone extending over latitudes 30-35°S, accommodated rifting during the Farallon-Nazca plate reorganisation at ~25 Ma, and it separates now the northern part of the Nazca plate originating at the East-Pacific Rise, from its southern part originating at the -two-times slower- Antarctic-Nazca rise (Cande & Haxby, 1991; Yañez et al., 2002). Based on their seismic velocities, Kopp et al. (2004) argued that the mantle could be ~15% serpentinized, which

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should lead to a decrease in bulk density of less than 15 kg/m<sup>3</sup>. Thus for sake of consistency with these values in our model, and recording that 35 Myr is the critical age for neutral buoyancy of oceanic plates (Afonso et al., 2007), we also

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removed the 10kg/m<sup>3</sup> density excess of the modeled oceanic mantle.

As a result of the model, this new rheological distribution prevents the oceanic plate to slide rigidly below a deformable continental mantle. Instead, it slides by stretching itself horizontally along the base of the resistant continent, at the characteristic ~100 km depth where olivine lowers its viscosity. Within about 3 Myr, a flat slab forms about 200 km long ( $M_{fl}$ , **Fig. 12**). Eventually, this stretching separates the upper and lower parts of the slab, into a horizontal type of slab break-off, with opening of an east-west pathway for asthenospheric flow across the subduction zone, and progressive heating from below (which may be described as a type of delamination).

Surface deformation is decoupled from the slab at depth and therefore is insensitive to whether the slab is flattening or not, because of the rigidity of the entire the continental “lid” (mantle). Hence, any active deformation or surface relief is more likely a continuation of inherited previous dynamics. The upper part of the slab above 100 km depth becomes disconnected from the slab-pull driving force, and it becomes passively attached to the westward moving overriding plate,

and trench retreat occurs. This final scenario, of westward drifting of the remnant top of the slab, is inconsistent with the Chilean subduction zone, because there, the three-dimensional continuity of the Nazca plate with its northern and southern parts that are still submitted to slab-pull, makes it still driven into the subduction zone (it may also still be submitted to the East-Pacific-Rise and Chile Rise ridge-pushes). However, the fact that the Nazca plate is submitted to differential slab-pull forces between the flat-slab area and its surroundings implies that shear forces act on the horizontal portion of the plate prior to subduction. It would therefore be useful in future studies to identify and quantify active shear deformation on the sea-floor north and south of the flat-slab area.

Seismicity at 31.5°S, when superimposed on the model, coincides with zones of high deviatoric strain above and within the stretched slab (**Fig. 12**). The model shows high temperature gradients there, from ~700 to 1100°C, which is high in comparison to the brittle-ductile transition range of olivine. However, temperatures must be hot enough at these depths in order for the slab to be able to stretch ductily. Cooler conditions generate brittle-like slab break-off and sinking. And if the oceanic lithosphere is made denser as in previous models, then flattening by stretching occurs over only about 100 km before sinking, unless the domain is taken even warmer. Further modeling investigations are needed to resolve this temperature discrepancy with observed seismicity.

Pardo et al. (2002) show extensional mechanisms within the horizontal flat portion of the slab, which are consistent with the occurrence of brittle-ductile stretching in this area of the modeled slab. The overriding mantle lithosphere displays a tensional state of stress. Whereas the flat-slab of Costa-Rica is nearly aseismic, the overriding plate is also reported to be under a tensional state of stress (Perez-Campos et al., 2008). Finally, a double seismic zone about 20 km thick is also reported from 50 to 100 km depth in Central Chile between latitudes 31°S and 33°S (Pardo et al., 2002). This observation is consistent with the presence of a hydrated oceanic mantle (e.g. Peacock, 2001). However, if one simply follows the rule of thumb proposed by Brudzinsky et al. (2007), of a double-seismic zone thickness equal to 0.14 km per million year age, then it should be only 5 km thick. We suggest that here, together with the hydrated state of the oceanic plate, compressional stresses transmitted into the oceanic plate favor the occurrence of this characteristic seismicity, despite the relatively young age of the slab (as also proposed for Northern Chile by Dorbath et al., 2008).

This model addresses a new cause for slab flattening in Central Chile. Whereas the role of the Juan Fernandez ridge in adding buoyancy to the slab and producing its flattening is questioned (e.g. van Hunen et al., 2002; 2004), we suggest that instead, it is the broader Challenger Fracture Zone that plays a role in having weakened and hydrated the entire oceanic plate thickness. The added effects of relative weakening of the subducting plate, and relative strengthening of the overriding plate, could have triggered the observed Central Chile's flat subduction.

In fact, van Hunen et al. (2004) modeled and discussed three main mechanisms for flat subduction: overthrusting of the subducting plate, subduction of a buoyant oceanic plateau, and slab suction forces. When comparing their models with the Peruvian flat slab, these authors conclude that the buoyancy of an oceanic plateau is insufficient to cause flat subduction, while overthrusting would better generate it (also suggested by Cross and Pilger, 1978, who had correlated continental motion and slab dips). While these authors addressed relative buoyancy and motion between subducting and overriding plates, our model here shows directly, the importance of their relative strength.

## 6.6 Proposed causes for along strike changes in the tectonic style of the Chilean margin

**Fig.13** displays a series of 4 schematic cross sections, corresponding to variations in either channel strength, crustal rheology and/or arc thermal state, which could be roughly compared to the known tectonic style of the Chilean Andes arranged from north (top) to south (bottom). This is only a working hypothesis of possible spatial rheological variations that might explain some of the Chilean arc first-order structural heterogeneities. Obviously, each latitudinal domain is characterized by more complex conditions. However, let us assess the following scenarios for Andean arc deformation:

- The top section results when the overriding plate has a warm and wide thermal state bounded by cold and resistant material. This condition produces various decoupled levels in the continental crust and multiple sub-crustal blocks. The pink thick line represents the crustal-scale shear zones proposed by Tassara (2005) and others, linking deformation at slab-mantle depths to the Sub-Andean belt to the east (e.g. our model **Fig. 11**).

- If the arc crust becomes more competent or cools down, a smaller number of crustal blocks form, with a larger characteristic length size. Towards the south, a strengthening crust combined with a narrowing arc-geotherm reduces the extent of continental deformation, which remains localised along the thermal anomaly. In this case, resistance of the continent competes with the strength of the subduction channel: if the latter remains sufficiently high, the continental fore-arc remains coupled to the slab and can be locally dragged downwards, forming a central depression, and west-verging faults bounding the flanks of the Main Cordillera. Increasing evidence show that the Central Depression is bounded by reverse faults in the area around Santiago (33°S, Rauld, 2002), and in Southern Chile (where the Principal

Cordillera Quaternary arc overthrusts the Central Depression, Cembrano et al., 2002).

– An extreme case may occur when the relative strength of the subduction interface dominates, and the forearc is dragged downwards as a whole, with significant fore-arc erosion. This situation may have occurred at various reprises in northern Chile, during the Mesozoic and Paleogene, where the geological history is punctuated by episodic eastward “jumps” and migration of the magmatic arc axis. The same process, to a less extent, may have occurred in south-central Chile during the Neogene, where Encinas et al. (2008), based on disputable paleontological evidence, have suggested that a period of rapid forearc tectonic subsidence (as much as 2 km) occurred in the Late Miocene. The rheology of the subduction interface may be changing in south-central Chile, subjected to a peculiar set of changing tectonic conditions, which associate a southward decreasing age of the slab (possibly also more hydrated due to formation at the relatively slower Antarctic-Nazca rise), and an overriding plate becoming more mafic (which may strengthen it from below, see section 5.3, and Tassara et al., 2006). In fact the subduction interface in southern Chile displays paradoxical properties, at the same time being able to generate sufficient shear stresses to produce, in the short term, very strong earthquakes, but also being sufficiently lubricated by sediment transport in the trench to produce, in the long term, relatively low orogenic topography (e.g. Lamb & Davis, 2003).

## 7. Conclusion

The presented numerical models reproduce the development of strain and stress patterns in the fore-arc and back-arc regions of an Andean-type subduction zone. By resolving the thermo-mechanical equations of motion and incorporating elastic-brittle-ductile behaviour all together, these models offer a physical robustness to purely conceptual published models (see most recent Ramos, 2009; deCelles et al., 2009). We show the relative control of crustal rheology, arc thermal conditions and subduction interface strength, on the geometry of crustal-scale fault zones, the location of crustal shortening and thickening, forearc tilting and subsidence, and rise of buoyant material from the subduction interface. If one takes a snapshot at a time-interval of several millions of years, the subduction process modifies continental rheologies and reaches domains that were located initially hundreds of kilometres inland. From one snapshot to the other, the relative strength of continental rocks with respect to that of the channel interface may evolve sufficiently to the different situations that were modelled here, over the relatively short time-scale of 4 Myr.

Although some authors have suggested a direct link between convergence rates and the amount of shortening along the Andes (e.g. Pardo-Casas & Molnar, 1987; Charrier et al., 2002), other studies show that this correlation may also be negative (e.g. Jordan et al., 2001; Yañez & Cembrano, 2004; Somoza & Ghidella, 2005). We have shown here that the amount of shortening depends significantly on the rheology of both the channel interface and the upper plate (e.g. Yañez & Cembrano, 2004; Medvedev et al., 2007; Sobolev & Babeyko, 2005; Sobolev et al., 2007). We show that under constant boundary conditions, plate convergence is differentially accommodated by the competing processes of overriding plate shortening and sliding along the subduction interface, in correlation with their respective strength. The weaker the continental crust and the stronger the subduction channel, lead to more widespread and greater superficial shortening of the overriding plate. Significant amounts of crustal shortening are accommodated at depth by lower crust and lithospheric mantle “delamination” *sensu lato*. The thickened continental crust can be dragged downwards, at the forearc (subduction erosion), but more generally below the arc into the mantle corner flow. We show that total upper plate shortening includes a component of shear exerted from the subduction interface, but also verify that it roughly equals the sum of upper plate trenchward motion and trench landward advance (all values taken positive, to compare with Jarrard, 1986; Schellart, 2008a,b).

Along the Andes, the double-vergence of regional-scale shear zones is geometrically obvious, but on the long-term most of the shortening is accommodated asymmetrically in the eastern flank of the orogen. Our models show that dominant fault vergence also depends on the rheology of the crust. Eastward vergence is favoured when the forearc crust is resistant, so that deformation propagates inside the continent. We also show that at the surface, the outer-forearc is uplifted under tensile conditions whereas the inner-forearc and arc domain are uplifted under compressive conditions, consistent with observations along the Chilean margin. When the forearc crust is resistant and mechanically coupled to the subduction interface, both zones are separated by a subsiding basin. Then, the fore-arc is tilted up-to-the-east (as indicated in Northern Chile, e.g. Riquelme et al., 2003; Farias et al., 2005). At depth, crustal thickening can also occur asymmetrically, depending on crustal strength heterogeneity, and produce typical “crocodile” structures seen in collisional contexts.

Specific initial conditions need to be accounted for in order to make these models fit geological and geophysical

data along a specific section, as done to some extent, with the same approach, for the Altiplano area in Dorbath et al. (2008). There, the combination of cold and resistant forearc and back-arc domains, a wide warm arc domain and asymmetric surface processes, produces a lithospheric-scale fault-bend-fault that propagates from the subduction interface at ~80 km depth towards the surface in the Sub-Andean zone, as proposed previously (e.g. Tassara, 2005; Farias et al., 2005).

We also show that the vanishing of the large-scale slab-pull force, mimicking deep slab detachment, generates crustal tension and thinning in the arc domain, triggered by the rise of voluminous amounts of buoyant material from the subduction interface. This scenario may explain episodes of crustal thinning and intra-arc volcanism along the Chilean margin, such as for example, the development of the intra-arc volcano-sedimentary Abanico basin in Central Chile (e.g. Charrier et al., 2002), contemporaneous with a dramatic decrease of plate convergence in the Oligocene.

Finally, we propose an alternative cause for slab flattening that differs from the common hypothesis of relative buoyancy of the subducting plate (e.g. Gutscher et al., 2000; Yañez & Cembrano, 2004), or asthenospheric viscosity increase (Van Hunen et al., 2004; Bonnardot et al., 2008): the relative weakness of the oceanic plate with respect to the continent. While geological evidences suggest the occurrence of several discrete episodes of slab flattening (e.g. Ramos, 2009; de Celles et al., 2009), water content and composition of both plates in contact is very likely to have varied during similar periods of time along different segments of the Andes.

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## Figures Captions

**Figure 1:** The Nazca-Chile margin topography and tectonic features: oceanic plate ages (red lines and labels) and ridges names. Andean morphological structures (labelled in black), the Central Depression (in shaded gray). Volcanoes are plotted as red triangles, and depths contours of the Nazca plate are plotted as dotted lines (from Cahill & Isacks, 1992). Three north-south main sections are distinguished, according to descriptive section 3. Seismic lines shown in figures 11 and 12 are located as pink lines. On the left, average Moho depths (green), trench depths (blue), and maximum heights (red) are shown from north to south.

**Figure 2:** Initial models main setup: each colour layer has specific density, friction and power-law creep parameters (see Table 2). The subduction channel is 12 km thick and shown in light orange (top 10 km), yellow (down to 100 km), and light green (from 100-300 km depth). The continental upper and lower crust are at 17 and 35 km depth respectively, either both weak (dark brown and red) or both strong (light brown and orange), and at different locations depending on different models. The oceanic crust is 10 km thick (orange), the oceanic and continental mantle go down to 100 km depth (ingreen and light blue respectively). The asthenosphere and the “wet” corner wedge region extending 300 km away from the slab (in blue). Isotherms spaced by 300°C are shown in dark red. The top free surface has initial topography deduced from initial density differences with Airy isostasy assumed at 300 km depth. Boundary conditions are lithostatic compensation elsewhere (thin black arrows) except for slab-pull velocity (5 cm/yr) applied at the base of the subducting plate and convergence velocity (2.5 cm/yr) applied on the continental lithosphere of the right-hand border (thick black arrows). See text for details on plate ages, compositions and densities.

**Figure 3:** Models 1-4, deformation and stress induced by specific models testing rheology, all with a relatively high channel strength, and after 4 Myr. For each model and from top to bottom: **a)** topography and erosion rate, **b)** shear strain rate (second invariant, most intense shear zones are shown in red, and velocity vectors, which can both change pattern from one time-step to the other according to numerical treatment of dynamical deformation. **c)** Shear stress (second invariant) with highest values in dark red, and nonlithostatic pressure lines tensile in red and compressive in black. **d)** Rheological layers shown with same colour convention as setup Fig. 2. The continental crust is either strong (model M1, top-left, SR), weak over 300 km (model M2 top-right, WR), weak over only the first 150km (model M3, bottom-left, WRfore), or weak from 200-400 km inland (model M4, bottom-right, WRback). See text and tables for explanations of different models.

**Figure 4:** Deformation and stress induced by a weak channel, and then also by warm arc-temperatures. For each model the same features as in Fig. 4 a,b,c,d) are displayed from top to bottom. Top-left, model M7 has a weak forearc crust and temperatures similar Model M3, but a weak channel (WR-WF-CT). Top-right, Model M8 has similar weak back-arc crust and temperatures to Model M4, and a weak channel (WR-WF-CT). Bottom-left, model MH5 has a wide thermal arc anomaly and a warm continent, a strong crust and weak channel (SR-WF-WT). Bottom-right, Model MH8 is similar weak crust and weak channel as M8, but has a wide thermal arc anomaly and a warm continent (WR-WF-WT). All these models let buoyant (yellow) material rise from the subduction interface beneath the arc.

**Figure 5:** Sketch of deformation induced by specific models according to rheology (varied from right to left, weak and strong crust in light and dark blue respectively), and channel strength (strong on top-line figures and weak in middle-line figures). Buoyant material is shown in yellow. Shear zones are shown in red. Specific cases are shown in the bottom-line, from left to right: model  $M_{10}$  in which slab-pull is removed (Fig.8), model  $M_{ap}$  applied to the Altiplano (Fig. 11), and model  $M_{fl}$  that produces a flat slab (Fig. 12). See text and Table 4 for explanations.

**Figure 6:** Subduction channel properties for several models: Model M1 (resistant interface, resistant crust) in blue, model MH8 (weak interface, weak back-arc crust and warm conditions), and model MH6 (weak interface, weak crust and warm conditions). **a)** Deviatoric shear stress, High friction and plagioclase rheology of model M1 leads to stresses > 60 MPa at 40-60 km depth, whereas both weak interface models M8 and MH6 display shear stresses <10 MPa from 40 km depth. **b)** Logarithm of viscosity, with decreasing values with depths close to 1021Pa.s for M1, and more than an order of magnitude less for MH8 and MH6. **c)** Minimum (solid lines) and maximum (dashed lines) Temperatures. Note that maximum values also represent buoyant material rising into the continental mantle. The grey areas show the

seismogenic depth range estimated along the Chilean margin (shallow in Northern Chile and deep in Southern Chile).

**Figure 7:** Three vertical sections at horizontal locations **a)** in the forearc (left column), **b)** in the arc (central column), and **c)** in the undisturbed continent (right column). Mechanical properties are displayed from top to bottom, deviatoric shear stress, logarithm of viscosity, and temperature. Profiles in blue correspond to model M1 (resistant interface, resistant crust), and profiles in red correspond to model MH8 (weak interface, weak back-arc crust, and warm conditions). Depths of intersection with the Moho are shown in circles, and domains of buoyant material coming from the subduction channel are shown in thick lines.

**Figure 8:** Model  $M_{10}$ , similar to M1, but without applying slab-pull. As stresses in the slab and along the subduction relax, buoyant material rises up to the surface and produces crustal thinning in the arc domain. The continuing convergence applied from the continental edge to the right, pushes the entire forearc and trench westward.

**Figure 9:** Detail of fore-arc deformation and rotation in models, **a)** M1 to the left, **b)** M4 at the centre, and **c)** MH6 to the right. From top to bottom: rotation rate, instantaneous strain rate and velocity vectors as in figure 3b), and rheological layers. The outer-forearc is under clockwise rotation (yellow to red on figures), corresponding to a downward tilt towards the slab, while the inner-forearc domain is under counter-clockwise tilting (greenish on figures). Alternating tilting blocks in model MH6 are due to a weak and decoupling crust.

**Figure 10:** Surface deformation on the continent, **a)** for strong interface models M1 to M4, and **b)** for weak interface and warm models MH5, MH6 and MH8, and then also for M8 in dashed lines. From top to bottom: topography, local shortening deduced from horizontal deformation  $\epsilon_{xx}$ , function of distance from the trench, and then, as a function of time, “emerged continent” shortening deduced from the above local shortening, “total upper plate” shortening, and trench motion. Note that models with a strong interface produce highest amounts of shortening and trench advance, as opposed to models with a weak interface. Independently, models with a weak or warm crust produce higher amounts of shortening than models with a strong crust. Dotted lines show approximate amount of shortening and trench motion taken up by elastic loading during the first Myr.

**Figure 11:** Model  $M_{ap}$  of deformation and stress for the Altiplano area (from Dorbath et al., 2008). From top to bottom same a,b,c,d) plots as those shown for each model of figure 4. White and black dots are seismicity relocated by Dorbath et al. (2008). Topographic profile in black is that at 18.5°S (see location Fig. 1).

**Figure 12:** Model  $M_{fl}$  of slab flattening, in which mantle rheologies are basically inverted: the oceanic mantle is young and wet, whereas continental mantle is cold and dry. After 3.5 Myr, the slab stretches horizontally as it is forced by the slab-pull boundary condition, but deforms ductily more easily than the overriding rigid continental mantle. White and black dots are seismicity at 31.5°S. Topographic profile in black is that at 32°S (see location Fig. 1).

**Figure 13:** Idealised rheological spatial evolution of the Chilean subduction zone from north to south, as a function of crustal strength (blue lines and ductile level shown in orange), arc temperature (green dots), and interplate strength (red). The thick pink line on top displays the crustal-scale fault-bend-fold structure proposed by several authors.