

Evolution of shallow and deep structures along the Maipo–Tunuyán transect ($33^{\circ}40'S$): from the Pacific coast to the Andean foreland

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Abstract: We propose an integrated kinematic model with mechanical constraints of the Maipo–Tunuyán transect ($33^{\circ}40'S$) across the Andes. The model describes the relation between horizontal shortening, uplift, crustal thickening and activity of the magmatic arc, while accounting for the main deep processes that have shaped the Andes since Early Miocene time. We construct a conceptual model of the mechanical interplay between deep and shallow deformational processes, which considers a locked subduction interface cyclically released during megathrust earthquakes. During the coupling phase, long-term deformation is confined to the thermally and mechanically weakened Andean strip, where plastic deformation is achieved by movement along a main décollement located at the base of the upper brittle crust. The model proposes a passive surface uplift in the Coastal Range as the master décollement decreases its slip eastwards, transferring shortening to a broad area above a theoretical point S where the master detachment touches the Moho horizon. When the crustal root achieves its actual thickness of 50 km between 12 and 10 Ma, it resists further thickening and gravity-driven forces and thrusting shifts eastwards into the lowlands achieving a total Miocene–Holocene shortening of 71 km.

The Andes, the world's largest non-collisional orogen, is considered the paradigm for geodynamic processes associated with the convergence of an oceanic plate (Nazca) below a continental plate margin (South American western margin). Although these mountains have been described as a consequence of crustal shortening that leads to crustal thickening and surface uplift (Isacks 1988; Sheffels 1990; Allmendinger *et al.* 1997), the mechanisms by which this crustal shortening is achieved remain

controversial (e.g. Garzione *et al.* 2008; DeCelles *et al.* 2009; Ehlers & Poulsen 2009; Armijo *et al.* 2010; Farías *et al.* 2010).

Tectonic plates near subduction zones are displaced towards the trench by buoyancy forces, such as slab pull and ridge push. These driving forces are resisted by forces associated with mantle viscosity, slab bending around the outer-rise and interplate friction (Forsyth & Uyeda 1975; Heuret & Lallemand 2005; Lamb 2006; Iaffaldano & Bunge

2008; Schellart 2008). Part of the work done by the driving forces is stored as elastic strain energy associated with the deformation of the upper plate above the seismogenic portion of the locked subduction interface. The elastic deformation accumulated during the decades to centuries of the interseismic period is cyclically recovered by the coseismic slip of both plates along the megathrust during earthquakes and their post-seismic relaxation phases (Savage 1983; Wang *et al.* 2012). If the Earth were purely elastic, then no permanent deformation would accumulate at active continental margins at geological timescales, all the convergence would be absorbed by coseismic interplate slip and no Andean-type mountains would exist. The elastic behaviour is appropriated to describe the rheology of the cold and rigid forearc and foreland regions, whereas the hot and weak arc-back-arc region is dominated by brittle (plastic) deformation of the upper crust and ductile (viscous) deformation of the lower crust and lithospheric mantle (Hyndman *et al.* 2005). In this context, the long-term structure of the arc-back-arc region is the result of the imbalance between compressive strain accumulated during the interseismic period and extensional strain activated by the co- and post-seismic phases, but summed over thousands or millions of seismic cycles.

Andean-type margins are geodynamically controlled in the long term by a strong coupling between the forearc and the down-going slab (Lamb 2006) and a comparatively rapid advance of the foreland (and the entire upper plate) toward the forearc with respect to the ocean-wards rollback velocity of the subducted slab (Heuret & Lallemand 2005; Schellart 2008). The permanently deforming arc-back-arc region located between the colliding forearc and foreland regions accumulates shortening and thickening to construct the Andes and its crustal roots. However, the mechanisms and structures by which this process actually occurs are not well understood. Our study explores a possible solution. We focus on this problem by studying the Maipo-Tunuyán transect across the Southern Central Andes ($33^{\circ}40'S$) at the latitudes of the cities of Santiago and Mendoza, which has been the subject of a number of tectono-structural studies including the pioneering observations of Darwin (Giambiagi *et al.* 2009). Because of the amount of pre-existing data, this transect constitutes a key area for understanding the role of deep-seated structural and tectonic processes on the constitution of the central Chilean-Argentinean Orogen.

Several conceptual models of deep crustal deformation have been proposed for the Andes between 18° and 38°S . These models can be divided into two types: east-vergent and west-vergent models. Among the first to be proposed are the wedge model

(Allmendinger *et al.* 1990), the duplex model (Schmitz 1994) and the east-vergent décollement model (Allmendinger & Gubbels 1996; Giambiagi *et al.* 2003b, 2012; Ramos *et al.* 2004; Farías *et al.* 2010). The east-vergent models assume a shallow subhorizontal to gently west-dipping detachment located at different depths in the upper crust or at the transition between the upper and lower crust. This crustal-scale detachment has been identified within a shallow brittle-ductile transition (e.g. Tassara 2005; Farías *et al.* 2010) and it concentrates nearly all the horizontal crustal shortening between the forearc and foreland. The eastwards propagation of deformation into the foreland generates predominantly east-verging upper crustal thrusts and folds. These models propose an underthrusting of the rigid, cold South American craton under the mechanically and thermally weakened Andean sector (Allmendinger & Gubbels 1996). The diametrically opposed west-vergent model proposed by Armijo *et al.* (2010) argues for the existence of a ramp-flat décollement dipping to the east and the growth of the Andes mountain belt as a bivergent orogen. In this model, it is the forearc (coastal crustal-scale rigid block) that underthrusts beneath the Principal Cordillera.

The aim of this paper is to propose an integrated Miocene to present kinematic model with mechanical constraints for the Maipo-Tunuyán transect ($33^{\circ}40'S$) that correlates with the large volume of geological and geophysical studies carried out by numerous authors. For that purpose, we construct a conceptual model of mechanical interplay between deep and shallow deformational processes based on thermomechanical and kinematic modeling.

The Andes between 33° and 34°S latitudes: geological framework

The $33^{\circ}40'S$ transect (Fig. 1) is located in a transition zone between a subhorizontal subduction segment north of 33°S and a zone of normal subduction south of 34°S (Stauder 1975; Barazangi & Isacks 1976; Cahill & Isacks 1992; Tassara *et al.* 2006). The abrupt southwards disappearance of the Precordillera and Sierras Pampeanas has been related to variation in the slab geometry (Jordan *et al.* 1983; Charrier *et al.*, this volume, in review). At this latitude, the Andes of Argentina and Chile are composed from West to East by the Coastal Range, the Central Depression, the Principal Cordillera, the Frontal Cordillera and the Cerrilladas Pedemontanas range, where the pre-existing extensional structures of the Triassic Cuyo basin are partially inverted (Fig. 2a).

The Coastal Range can be divided into two sectors. The western sector with low topographic

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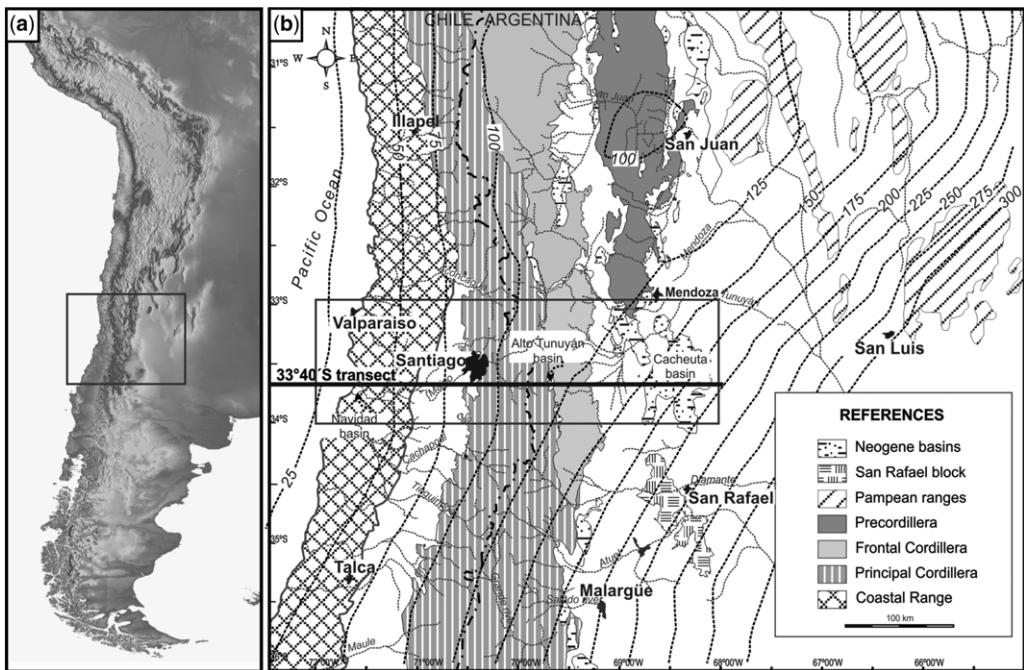


Fig. 1. (a) Shaded relief map of the Andes. Box indicates location of map in Figure 1b. (b) Regional map of the Andes at latitudet 31–36°S highlighting the present-day morphostructural units. The 33°40'S transect crosses the Coastal Range, the Central Depression, the Principal and Frontal cordilleras and the foreland. Dashed black lines represent depth of the subducted Nazca plate (from Tassara & Echaurren 2012). Notice the location of the transect above the transition segment between flat and normal subducted slab. Box indicates location of map in Figure 2.

altitude (<500 m) is composed of a series of Late Pliocene–Pleistocene marine ablation terraces (Wall *et al.* 1996; Rodríguez *et al.* 2012) carved on Late Palaeozoic–Middle Jurassic plutons (Sellés & Gana 2001). The eastern sector with altitudes up to 2000 m and relicts of high-elevated peneplains at different elevations (Brüggen 1950; Fariñas *et al.* 2008) is made up of Cretaceous plutons within a Late Jurassic–Early Cretaceous sedimentary and volcanic country rock.

The Central Depression at an elevation of 700–500 m.a.s.l separates the Coastal Range from the Principal Cordillera. It consists of a Quaternary sedimentary and ignimbritic cover of up to 500 m thickness beneath the Santiago valley (Araneda *et al.* 2000) and basement rocks cropping out in junction ridges and isolated hills reaching 1600 m a.s.l. (Rodríguez *et al.* 2012).

The Principal Cordillera is also subdivided into western and eastern sectors. The western sector consists of Eocene–Early Miocene volcaniclastic rocks of the Abanico extensional basin (Charrier *et al.* 2002), covered by the Miocene volcanic-arc rocks of the Farellones Formation (21.6–16.6 Ma; Aguirre *et al.* 2000) which young southwards (Godoy 2014). The eastern Principal Cordillera

consists of a thick sequence of Mesozoic sedimentary rocks, highly deformed into the Aconcagua fold-and-thrust belt (Giambiagi *et al.* 2003a).

The Frontal Cordillera at this latitude corresponds to the Cordon del Portillo range, where Proterozoic metamorphic rocks, Late Palaeozoic marine deposits, Carboniferous–Permian granitoids and Permo-Triassic volcanic rocks crop out (Polanski 1964). This range is uplifted by several east-vergent faults of the Portillo fault system.

The Cerrilladas Pedemontanas range is marked by the inversion of Triassic extensional faults of the Cuyo basin and moderately dipping basement faults (García & Casa 2014). Triassic continental rocks are covered by Neogene to Quaternary synorogenic deposits of the Cacheuta basin (Irigoyen *et al.* 2000). The Quaternary volcanic arc at this latitude is represented by the Marmolejo-Espíritu Santo-San José volcanic centre, located along the crest of the Andes 300 km east of the Chile trench.

Thermomechanical modelling

A necessary first-order constraint for the construction of a crustal-scale balanced cross-section is the

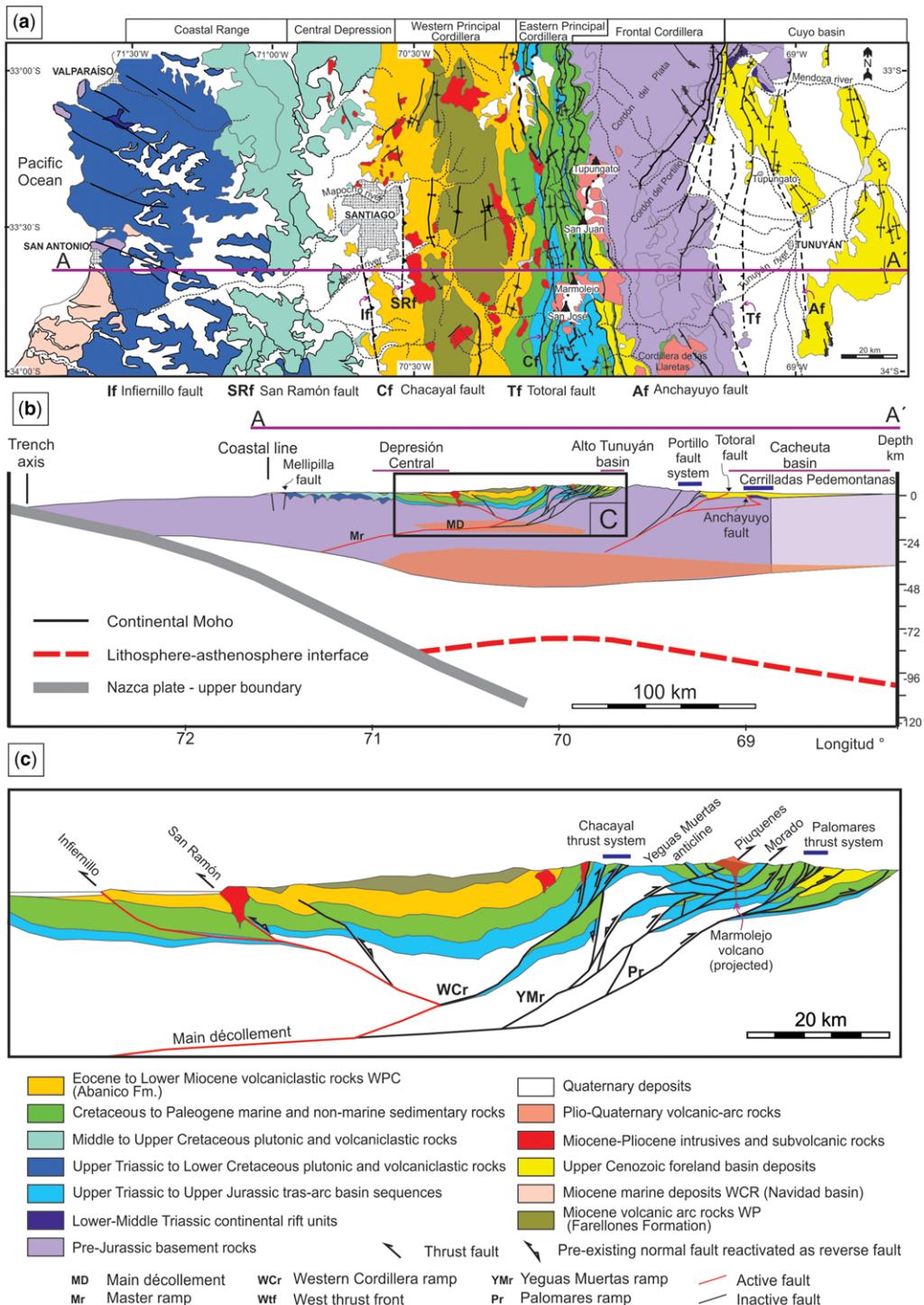


Fig. 2. (a) Simplified map of the Andean orogen between 33° and 34° S. Modified from Polanski (1964), Gana *et al.* (1996), Wall *et al.* (1996), Sellés & Gana (2001), Giambiagi & Ramos (2002) and Giambiagi *et al.* (2003a). (b) Balanced cross-section from coast to foreland, based on our own data and detailed compilation from Polanski (1964).

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expected mechanical structure of the lithosphere, particularly the identification of eventual ductile zones associated with brittle–ductile transitions that can serve as master detachments. The gravity-based and seismically constrained three-dimensional density model of the Andean margin developed by Tassara *et al.* (2006) and recently upgraded by Tassara & Echaurren (2012) contains the geometries for the subducted slab, the lithosphere–asthenosphere boundary of the South American plate (LAB), the continental Moho and the intra-crustal discontinuity, which separates the upper crust (density 2700 kg m^{-3}) from the lower crust (density 3100 kg m^{-3}). Based on these geometries and analytical formulations of the heat transfer equation (Turcotte & Schubert 2002) with appropriate boundary conditions, a 3D thermal model of the Andean margin has been developed (Morales & Tassara 2012; Tassara & Morales 2013). This model accounts for heat conduction from the LAB and subduction megathrust with radiogenic heat generation into the crust and thermal advection by the subducted plate, and predicts the distribution of temperature inside the Andean lithosphere.

The 3D thermal model and the original density model serve as the base to construct a 3D mechanical model based on the concept of the yield strength envelope (Goetze & Evans 1979; Burov & Diament 1995; Karato 2008). This envelope predicts the mechanical behaviour (brittle, elastic or thermally activated ductile creep) of a compositionally layered lithosphere with depth along a given 1D geotherm via the extrapolation of constitutive rheological laws for different Earth materials from experimental conditions to lithospheric space–temporal scales.

We use a preliminary version of the thermo-mechanical model (Tassara 2012) to extract an east–west cross-section along the Maipo–Tunuyán transect (Fig. 2b). This section shows that most of the upper-middle crust has a brittle-elastic behaviour, particularly for the cold and rigid forearc and foreland regions. However, a ductile behaviour is predicted by the model below the Principal Cordillera within a thin layer ($<5 \text{ km}$) at mid-crustal depths (15–20 km) and for the entire lower crust (i.e. deeper than 30 km).

Kinematic model of subduction orogeny

Our kinematic model with thermomechanical constraints (Fig. 3) accounts for the long-term

deformation of the Andean convergent margin between the cratonic interior of the South American plate and the forearc. It considers a coupled slab–forearc system with elastic loading and release associated with the megathrust seismic cycle. We kinematically model a velocity gradient ΔV_x between the continental lithospheric plate and the underlying asthenospheric mantle by fixing the slab–forearc interface and applying a westwards movement of the continental plate along an artificial line at the base of the Moho. This artificial line has no geological significance; it has been designed for the purpose of kinematical modelling and represents a broad area with dislocation creeping and a vertical velocity gradient ΔV_x constrained between the Moho and the lithosphere–asthenosphere boundary. Displacement is transmitted along this base using the *trishear* algorithm with p/s (propagation/displacement) ratio between 0 and 0.5 until the singularity point S below the Cordilleran axis (Fig. 3). At the singularity S, which is located above the downdip limit of the coupled seismogenic zone, shortening is transmitted to a ramp-flat master décollement, modelled with the fault parallel flow algorithm as a passive master fault. Above this décollement, crustal block B experiences brittle deformation. Below the décollement, the crust thickens by ductile deformation. Inside crustal block A, deformation diffuses westwards and upwards in a triangular zone of distributed shear. This deformation represents flexural permanent deformation above the S point.

The geometry of our proposed master décollement coincides with previous conceptual models for deep crustal deformation proposing an eastwards vergence of the orogen (Allmendinger *et al.* 1990; Allmendinger & Gubbels 1996; Ramos *et al.* 2004) and with the active ramp-flat structure dipping 10°W beneath the western Principal Cordillera and $25\text{--}30^\circ\text{W}$ beneath the Coastal Range proposed by Farías *et al.* (2010) based on seismological studies. In our model, the décollement ramps up beneath the easternmost sector of the Chilean slope of the Andes. The vertical component of slip in this ramp would be much greater than in the Argentinean slope, where low-angle thrusting develops. The abrupt rising of the international border sector may therefore be a consequence of localized rapid uplift on this segment of the transect.

The rationale follows Isacks' (1988) conceptual model in which brittle crustal horizontal shortening in the back-arc upper crust is compensated by

Fig. 2. (Continued) Gana & Tosdal (1996), Alvarez *et al.* (1999), Godoy *et al.* (1999), Giambiagi & Ramos (2002), Giambiagi *et al.* (2003a, b), Rauld *et al.* (2006), Godoy (2011). Red lines indicate faults with Quaternary activity. Geophysical data was extracted from the ACHISZS electronic database (www.achiszs.edu.cl/~achiszs/accessdb.html). (c) Detail of the cross-section in (b) for the sector of Principal Cordillera. Red lines indicate faults with Quaternary activity.

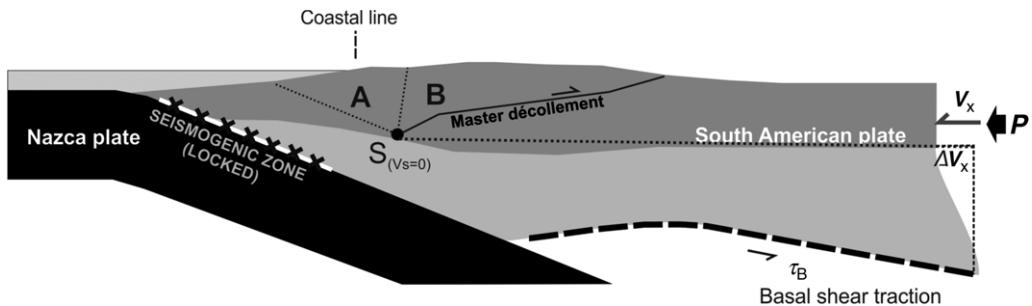


Fig. 3. Conceptual model for the 33°40'S transect. A velocity gradient ΔV_x is applied at the back of the foreland area during the 100% locked interseismic period. We kinematically model deformation by applying movement along an artificial basal line at the Moho, which transmits displacement westwards until the singularity point S. At this point S shortening is transmitted to a ramp-flat master décollement, modelled as a passive master fault. Above this décollement, crustal block B experiences brittle deformation. Below this line, the crust thickens by ductile deformation. Inside crustal block A, deformation diffuses westwards and upwards in a triangular zone of distributed shear.

ductile thickening in the arc and back-arc lower crust, and with the Andean microplate model of Brooks *et al.* (2003) in which a west-dipping décollement below the Andes mountain belt is viscously coupled to the South American craton and allows continuous creeping and stress transmission across the boundary.

We perform a forward model using the 2DMove academic license, taking into account the geological and geophysical constraints and the thermo-mechanical model. We assume plain strain along the transect, without magmatic additions or subduction erosion. Shortening estimates rely on upper crustal deformation measured all along the transect and the balanced cross-section, which provides a minimum estimate of horizontal shortening. First of all, we reconstruct the Late Cretaceous–Early Paleocene and Eocene–Early Miocene compressional and extensional events using the available geological and geochemical studies. For the Early Miocene–present shortening period, we carry out 36 steps each of 2 km shortening, constrained by the timing of movement along the main structures of previous studies (Godoy *et al.* 1999; Giambiagi & Ramos 2002; Giambiagi *et al.* 2003a), isotopic analysis (Ramos *et al.* 1996; Kay *et al.* 2005) and exhumation analysis (Kurtz *et al.* 1997; Maskaev *et al.* 2004, 2009; Hoke *et al.* 2014). With these data we calculate an average shortening rate for each period. Every 2–3 steps, we simulate erosion and sedimentation in accordance with sedimentological, palaeogeographic and provenance studies from the foreland and forearc basins (Irigoyen *et al.* 2000; Giambiagi *et al.* 2001; Rodríguez *et al.* 2012; Porras *et al.*, this volume, *in review*), and allow flexural-isostatic adjustments of the lithosphere due to local load changes assuming a default value for the Young's modulus $E = 7 \times 10^{10}$ Pa and the effective elastic thickness (T_e) calculated

in Tassara *et al.* (2007). Considering that we take into account the effects of erosion and sedimentation, topographic data presented here are purely qualitative.

Structure of the 33°40'S transect

The present-day structure of the 33°40'S transect is represented in Figure 2b, c. The crust below the Andes appears to reach its greatest thickness of 50 km at a longitude near 69°45'W. This longitude closely corresponds to the location of the highest peaks at this latitude that reach an altitude of 6000 m in the Marmolejo volcano. The Coastal Range (Fig. 4a) represents an east-dipping gentle homocline of Upper Palaeozoic–Cretaceous rocks (Wall *et al.* 1999). No major Andean thrust fault has been identified in this range, which is affected mainly by high-angle pre-Andean NNW–NW-trending faults such as the Melipilla fault (Yáñez *et al.* 2002), which may have been inherited from the Late Triassic continental-scale rifting.

The western Principal Cordillera comprises the Late Eocene–Miocene volcanic arc and is dominated by the inversion of the Abanico extensional basin (Fig. 4b). Fock *et al.* (2006) proposed that this inversion has a bivergent sense. According to Godoy *et al.* (1999) however, a double vergence for Abanico is developed only south of this latitude; at the transect latitude, the Abanico master faults are only inverted in its western edge.

The Front Range east of Santiago city represents the western thrust front of the Principal Cordillera. It was uplifted by inversion of the east-dipping Abanico master fault system, including the Infiernillo and San Ramón faults (Godoy *et al.* 1999; Fock *et al.* 2006; Rauld *et al.* 2006). The latter of which has been described by Rauld *et al.* (2006)

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and Armijo *et al.* (2010) as a feature active during the Holocene. The hanging wall of the San Ramón fault is folded into a tight syncline–anticline pair interpreted to be generated by an underlying basement ramp (Godoy *et al.* 1999). Toward the east, east-dipping faults with small throws are inferred to fold the Abanico and Farellones strata (Armijo *et al.* 2010).

The east-vergent Chacayal thrust system marks the border between western and eastern Principal Cordillera sectors (Giambiagi & Ramos 2002). This thrust system, which uplifts a thick sheet of Upper Jurassic sedimentary rocks (Fig. 4c), runs from the Las Cuevas river ($32^{\circ}50'S$) to the Maipo river ($34^{\circ}10'S$) for more than 150 km along-strike and corresponds to the most important structure in the western slope of the Principal Cordillera. The Aconcagua fold-and-thrust belt presents an overall geometry of a low-angle eastwards-tapering wedge, characterized by a dense array of east-vergent imbricate low-angle thrusts (Fig. 4d) with subordinate west-vergent out-of-sequence thrusts (Fig. 4d) (Giambiagi & Ramos 2002). The western sector of the belt is dominated by a broad anticline related to the inversion of the Triassic–Jurassic Yeguas Muertas extensional depocentre (Alvarez *et al.* 1999).

The Frontal Cordillera represents a rigid block uplifted by the Portillo fault system, located at its eastern margin (Fig. 4e). This system corresponds to a series of east-vergent deeply seated thrust faults (Fig. 4f), which uplift the pre-Jurassic basement rocks on top of the Middle Miocene–Quaternary sedimentary rocks deposited in the foreland basin (Polanski 1964; Giambiagi *et al.* 2003a). The foreland area comprises Neogene–Quaternary sedimentary deposits gently folded into open anticlines (Fig. 4g). The Triassic Cuyo basin is partially inverted in this sector by reactivation of pre-existing structures, such as the Anchayuyo fault, and generation of new thrusts (Fig. 4h).

Deformational periods

Even although the tectonic evolution of the transect can be regarded as a continuous deformational event from the Early Miocene to the present, with crustal thickening and widening and uplift of the Andean ranges, we can separate this evolution into several periods of deformation during which rock uplift and erosion shape the orogen.

The Late Cretaceous–Paleocene deformational event

The Andean orogeny started in several segments with contractional deformation during Late Cretaceous time when the back-arc basins began to be

tectonically inverted (Mpodozis & Ramos 1989). Deformation in the southern Central Andean thrust belts north and south of this transect started during the Upper Cretaceous–Palaeogene shortening event (Mpodozis & Ramos 1989; Tunik *et al.* 2010; Orts *et al.* 2012; Mescua *et al.* 2013). According to Tapia *et al.* (2012), in the studied transect this contractional event was localized in the Coastal Cordillera and the western sector of the Principal Cordillera. However, Godoy (2014) argues that the evidence for this event in the Main Range is dubious and interprets the area to represent a structural knot that resisted K–T (Cretaceous–Tertiary) deformation.

In the eastern Principal Cordillera on the other hand, continental sediments were deposited in the northern Neuquén foreland basin (Tunik *et al.* 2010) which culminated with an Atlantic marine ingress during Maastrichtian time (Tunik 2003; Aguirre-Urrreta *et al.* 2011). Further east, geothermocronological analysis indicates that the foreland area east of Frontal Cordillera was stable (not deformed nor uplifted) during the Jurassic–Palaeogene period (Ávila *et al.* 2005). Further structural studies are needed to understand the deformation during this stage, which is beyond the scope of this study.

Given the lack of precise constraints for the deformation during Cretaceous–Palaeogene time in the studied transect, we estimated the shortening based on the assumption of conservation of the crustal area along the section. We used the Tassara & Echaurren (2012) geophysical model to calculate the crustal area at present. Taking into account the shortening estimations based on structural studies (Godoy *et al.* 1999; Giambiagi & Ramos 2002; Giambiagi *et al.* 2003a; Fig. 2b), an initial crustal thickness of $T_0 = 40$ km is needed to accommodate Andean shortening along the transect. However, crustal thickness was not constant across-strike before the Neogene, since a Palaeogene extensional event led to a thin crust (of 30–35 km of thickness) in the western Principal Cordillera (see the following section). In order to preserve crustal area, this region of thin crust should be compensated with a thick region which we associate with the Cretaceous–Palaeogene contractional deformation developed to the west of the extended sector. A block of 42-km-thick crust in the present Coastal Cordillera and Central Depression can account for the missing crustal area. We modelled this block of thicker crust with 10 km of Late Cretaceous–Palaeogene shortening (Fig. 5a).

The Eocene–Early Miocene extensional event

During Eocene–Early Miocene time, a protracted extensional event took place with deformation concentrated in the western sector of the Principal

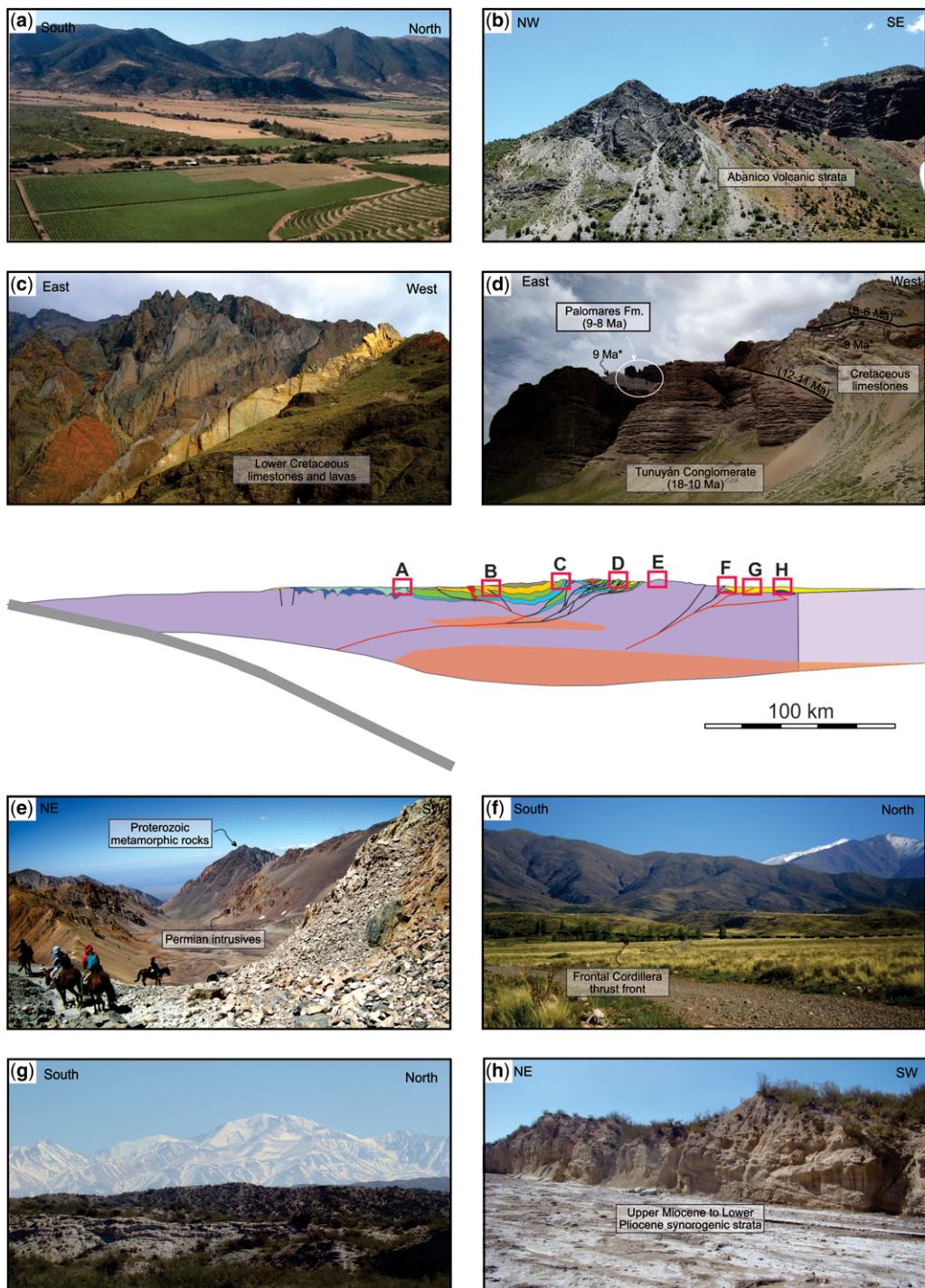


Fig. 4. Photographs of the 33°40'S transect. (a) Central Depression and Coastal Range (photograph P. Alvarez). (b) Abanico strata deformed into open west-vergent folds in the western Principal Cordillera (photograph J. Suriano). (c) Vertical mesozoic beds uplifted by the Chacayal fault and back-tilting by the Aconcagua FTB faults (photograph L. Pinto). (d) Palomares fault system in the Aconcagua FTB, uplifting Mesozoic strata over the Neogene synorogenic

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Cordillera (Charrier *et al.* 2002). This event has been linked to a segmented roll back subduction event (Mpodozis & Cornejo 2012) in contrast to the compressive deformation registered in the Andean margin north of 25°S (Carrapa *et al.* 2005; Hongn *et al.* 2007). Normal faulting was associated with crustal thinning and tholeiitic magmatism of the Abanico Formation (Nyström *et al.* 1993; Kay & Kurtz 1995; Zurita *et al.* 2000; Muñoz *et al.* 2006), whose $^{40}\text{Ar}/^{39}\text{Ar}$ ages at the latitude of Santiago (33°S) range from 35 to 21 Ma (Muñoz *et al.* 2006). The extensional basin was filled with up to 3000 m of volcaniclastic deposits, and acidic to intermediate lavas with sedimentary intercalations (Charrier *et al.* 2002). Even although geochemical analyses suggest it was formed upon a *c.* 30–35-km thick continental crust (Nyström *et al.* 2003; Kay *et al.* 2005; Muñoz *et al.* 2006), there is no evidence of marine sedimentation. Isotopic analysis and comparison between western and eastern outcrops of the Abanico Formation indicate higher crustal contamination, suggesting larger crustal thickness to the east (Fuentes 2004; Muñoz *et al.* 2006) and an asymmetric geometry of the extensional basin, due to the concentration of extensional deformation in the western master fault system.

During this period, the Coastal Range should have been elevated due to isostatic compensation of the Late Cretaceous–Paleocene crustal thickening event and locally by isostatic rebound close to the Abanico east-dipping master fault. Further east in the Neuquén basin, Paleocene continental distal sediments were deposited tapering towards the east.

The extensional structures of the Abanico basin have been obliterated by the later compressional faults. Inferred master faults have been suggested to be located beneath the Central Depression (Godoy *et al.* 1999; Fock *et al.* 2006). Much more speculative is the existence of eastern basin-border faults. Apatite and zircon fission track cooling ages from the eastern Coastal Cordillera constrain an exhumation period between 36 and 42 Ma (Fariñas *et al.* 2008), which could be related to uplift of the rift shoulder westwards of the master fault (Fock 2005; Charrier *et al.* 2007) at the beginning of the extensional period (since Late Eocene).

The Early Miocene (21–18 Ma)

The last major shortening event began during the Early Miocene with the inversion of the Abanico

basin (Godoy *et al.* 1999; Charrier *et al.* 2002; Fock *et al.* 2006). After that, chronological data suggest that deformation has accumulated during a single period of shortening between the Early Miocene and the present (Giambiagi *et al.* 2003a; Porras *et al.*, this volume, in review). The onset of deformation in the western slope is marked by the change from low-K Abanico Formation tholeiites to calc-alkaline dacites of the Teniente Volcanic Complex (Kay *et al.* 2005, 2006), between 21 and 19 Ma (Charrier *et al.* 2002, 2005). This inversion occurred coevally with the development of the Farellones volcanic arc of Early–Middle Miocene age (Vergara *et al.* 1999). A 22.5 Ma U–Pb zircon SHRIMP age of the base of the Farellones Formation along the Ramón–Damas Range anticline marks the beginning of the Farellones arc at this transect (Fock 2005).

In our model, the western Principal Cordillera is related to the generation of the Main décollement (MD) rooted in the singularly point S at the contact between the Moho and mantle sub-arc lithosphere, *c.* 40 km depth from the surface, and the Western Cordillera ramp (WCr) inferred to be located below the Western thrust front (WTF). We interpret this event as the generation of an important detachment connected to the Chilean ramp proposed by Fariñas *et al.* (2010) and located between 12 and 15 km depth beneath the Principal Cordillera. Eastwards movement of the upper crust relative to the stable and long-lived lower crust MASH (melting, assimilation, storage and homogenization) zone proposed by Muñoz *et al.* (2012) could explain the width of up to 40 km of the Farellones volcanic arc. The main uplifted sectors for this period are the eastern Coastal Range and the western Principal Cordillera, above the Western Cordillera ramp (the Front Range east of the city of Santiago). The passive uplift of both sectors is in agreement with provenance studies of heavy mineral assemblages for the lower Navidad Formation (Rodríguez *et al.* 2012), whose radiometric ages indicate deposition during 23–18 Ma (Gutiérrez *et al.* 2013) and apatite fission track (AFT) age (18.3 ± 2.6 Ma) for an Upper Cretaceous deposit (Fock 2005).

The passage of the pre-existing Abanico master normal faults over the WCr could have reactivated these structures as passive back-thrusts and favoured the intrusion of the La Obra pluton (19.6 Ma; Kurtz *et al.* 1997). In the foreland, retro-arc volcanism of the Contreras Formation (18.3 Ma;

Fig. 4. (Continued) units (photograph M. Tunik). (e) Frontal Cordillera from the Portillo pass (photograph M. Tunik). (f) Frontal Cordillera active thrust front (photograph L. Giambiagi). (g) Looking westwards from the Cacheuta basin. Strata on front correspond to uplifted Neogene synorogenic deposits (photograph L. Giambiagi). (h) Neogene deposits uplifted and folded at the easternmost sector of the transect (photograph L. Giambiagi).

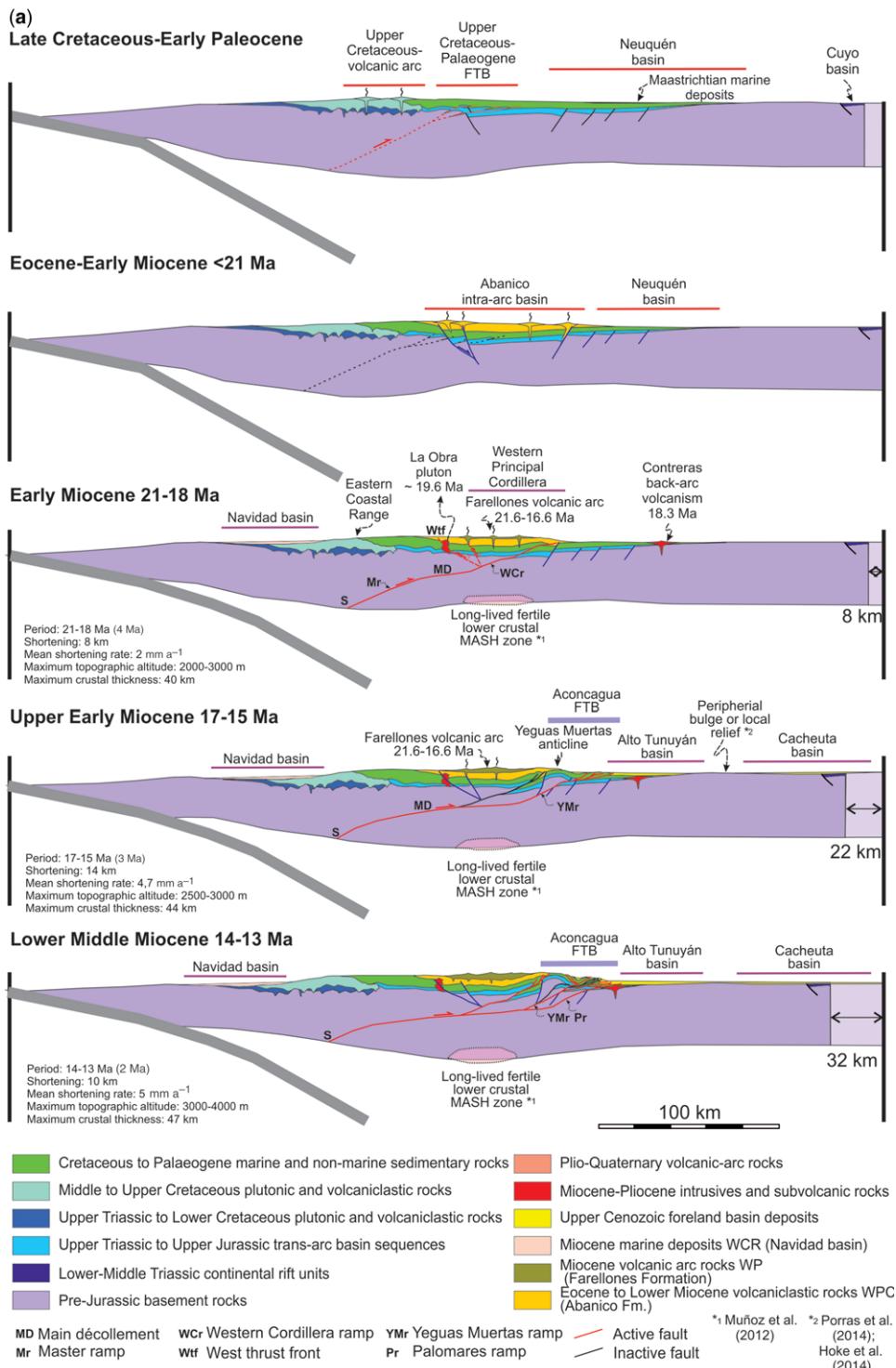
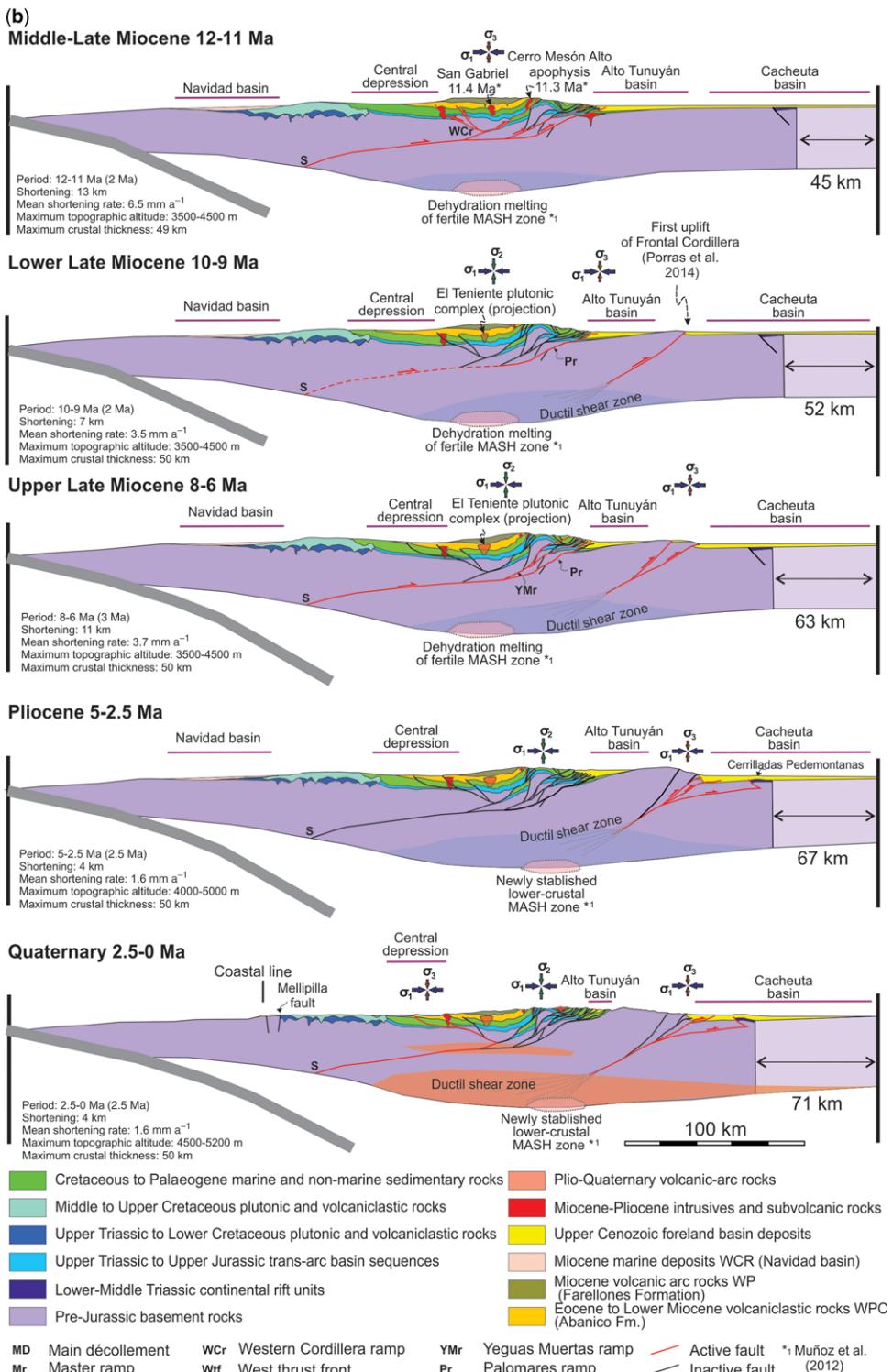


Fig. 5. Model for the evolution of the 33°40'S transect from the Late Cretaceous to the present. (a) Deformational periods from the Late Cretaceous–Early Paleocene compressional event to the early Middle Miocene phase.

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**Fig. 5. (Continued) (b)** Deformational periods from the Middle–Late Miocene to the present.

Giambiagi & Ramos 2002) with geochemical signatures of unthickened crust (Ramos *et al.* 1996) pre-dates the formation of the Alto Tunuyán foreland basin.

We calculated 8 km of shortening concentrated in the western Principal Cordillera (7 km) and the Coastal Range (1 km), with a mean shortening rate of 2 mm a^{-1} for this period. Beneath the western Principal Cordillera, the crust achieved its maximum thickness (40 km) and maximum topographic elevation ranges between 2000 and 3000 m a.s.l.

The Late Early Miocene (17–15 Ma)

This period marks the beginning of deformation along the Aconcagua FTB (fold-and-thrust belt), linked to the prolongation of the Main décollement eastwards and the formation of the Yeguas Muertas ramp beneath the homonymous anticline (Giambiagi *et al.* 2003a, b). By this time, driving forces cannot supply the energy needed to elevate the western Principal Cordillera by movement along the WCr (supercritical wedge stage), and the orogen begins to grow in width to lower the taper by propagating forwards towards the foreland. According to Muñoz *et al.* (2012), this shift in the locus of deformation should be related to the rapid ascent of subduction-related mantle-derived magmas that had little interaction with the upper lithosphere, such as those analysed by these authors in the western Principal Cordillera.

The earliest Neogene synorogenic strata appear in the Alto Tunuyán basin at *c.* 17–16 Ma, with source region restricted to the volcanic rocks of the Principal Cordillera and to the Frontal Cordillera (Porras *et al.*, this volume, *in review*). For this last source region, Porras *et al.* (*this volume, in review*) propose the existence of a peripheral bulge located in what is now the Frontal Cordillera, while Hoke *et al.* (2014) suggest a significant pre-Middle Miocene local relief. Further studies are required in order to decide which option is the best. During this period, the Cacheuta basin, apparently disconnected from the Alto Tunuyán basin, only received aeolian deposits (Irigoyen *et al.* 2000).

The uplift of the western Principal Cordillera evidenced in the provenance of the Navidad basin (Rodríguez *et al.* 2012) is due to overriding of this sector of the range across the Yeguas Muertas ramp. Movement along the Main décollement and this ramp achieves 14 km of shortening in the Aconcagua FTB, with an average shortening rate of 4.7 mm a^{-1} . Maximum topographic elevation is between 2500 and 3000 m in the western Principal Cordillera, and the crust thickens to 44 km below the westernmost Principal Cordillera as suggested by geochemical data (Kay *et al.* 2005). Flexural-isostatic compensation of this sector of the orogen

could explain the exhumation rates of the La Obra pluton ($0.5\text{--}0.6 \text{ mm a}^{-1}$) during this period (Kurtz *et al.* 1997).

The coeval uplift of the eastern Coastal Cordillera and the Principal Cordillera leads to the formation of a depocentre filled by up to 3000 m of arc-related volcanics of the Farellones Formation (Vergara *et al.* 1988; Elgueta *et al.* 1999; Godoy *et al.* 1999). Magma migration within the arc may have been enhanced by active movement along the Western Cordillera ramp and the associated back-thrusts.

The Middle Miocene (15–11 Ma)

During this phase, upper crust deformation is localized in the eastern Principal Cordillera with the development of the east-vergent in-sequence Aconcagua FTB faults and associated back-thrusts close to the international border. Cross-cutting relationships between thrust faults suggest the cyclic activation–deactivation of the Western, Yeguas Muertas and Palomares ramps with important shortening (17 km) in the Principal Cordillera. The period between 12 and 11 Ma corresponds to one of quiescence in deformation along the Aconcagua FTB, when the thrust front became inactive and erosion of the previously uplifted area was responsible for the generation of an important unconformity between the Cretaceous strata and the Middle Miocene synorogenic deposits (Giambiagi *et al.* 2001). Instead, deformation is concentrated in the Chilean slope of the Principal Cordillera with the reactivation of the Western Cordillera ramp and movement along the pre-existing San Ramón and Infiernillo faults (Fock *et al.* 2006). After this short quiescence period, the reactivation of the Yeguas Muertas and Palomares ramps leads to generation of out-of-sequence thrusts in the Aconcagua FTB.

This is a period of high crustal shortening (23 km) concentrated along the international border and the eastern Principal Cordillera, with the mean shortening rates of 5 mm a^{-1} and 6.5 mm a^{-1} for the 15–13 and 12–11 Ma periods, respectively. The crust almost achieves a thickness of 49 km, with an eastwards shift in the crustal keel. Important space creation due to flexural compensation of the tectonic load, immediately to the east of the Aconcagua FTB, is registered both in the Alto Tunuyán and Cacheuta basins (Irigoyen *et al.* 2000; Giambiagi *et al.* 2001).

At the end of this phase, the volcanic activity practically wanes and only very localized activity is recorded during late Middle Miocene–Late Pliocene time, in agreement with a highly compressive stress regime in the arc region. The present volcanic arc develops in the latest Pliocene along the watershed. Instead, barren and

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mineralization-hosting intrusives, such as La Gloria and San Gabriel plutons and Cerro Mesón Alto porphyry (Deckart *et al.* 2010), intrude the Miocene Farellones Formation or the Upper Cretaceous sedimentary rocks.

The Late Miocene (10–6 Ma)

At the beginning of this phase around 10–9 Ma, the crust achieves its present maximum thickness of 50 km in accordance with isotopic analysis (Kay *et al.* 2005). At this point, driving forces can no longer supply the energy needed to thicken the crust and uplift the range and the crustal root is likely to grow laterally in width instead of increasing its depth, with a reduction in shortening rates. This is in accordance with the proposition of Muñoz *et al.* (2012) for the existence of deep crustal hot zones. According to these authors, the repeated basalt intrusion into the lower crust induces a significant thermal perturbation, with the amphibolitic lower crust reaching temperatures up to 750–870°C and increased melt component. This promotes the ductile behaviour of the lowermost crust and the widening of the crustal root.

Overall, no volcanic activity is registered during this period. Instead, the Teniente Plutonic Complex intrudes the Abanico and Farellones formations between 12 and 7 Ma (Kay & Kurtz 1995; Kurtz *et al.* 1997; Kay *et al.* 2005).

During the early Late Miocene (10–9 Ma), reactivation of the Palomares ramp leads to out-of-sequence thrusting in the Aconcagua thrust front and the cannibalization of the Alto Tunuyán foreland basin deposits. Seven kilometres of shortening are achieved during this phase at the thrust front. Provenance analysis of the synorogenic fill of the Alto Tunuyán foreland basin indicates the beginning of uplift of the Frontal Cordillera (Porras *et al.*, this volume, in review), thrusts of which have been inferred to be deeply seated in a ductile shear zone of the lower crust (Giambiagi *et al.* 2012).

During the latest Late Miocene (8–6 Ma), the sedimentary record indicates important uplift of the Frontal Cordillera between 33°30' and 34°30'S (Irigoyen *et al.*, 2000; Giambiagi *et al.* 2003a; Porras *et al.*, this volume, in review). Thermochronological studies in Argentina both bordering and within the study area show that very little exhumation has occurred in this area during Cenozoic time. An apatite (U–Th)/He study in the Frontal Cordillera between 32°50' and 33°40' yields pre-Miocene exhumation rates of c. 12 m Ma⁻¹, with an increase to 40 m Ma⁻¹ at 25 Ma and an inferred onset of rapid river incision between 10 and 7 Ma, roughly within geological constraints (Hoke *et al.* 2014).

A couple of million years after the onset of uplift in the Frontal Cordillera, the Principal

Cordillera experiences important reactivation of the Yeguas Muertas and Palomares ramps with generation of out-of-sequence thrusts (e.g. Piuquenes and Morado) and the reactivation of the Palomares thrust system (Giambiagi & Ramos 2002).

At this time, movement along the Yeguas Muertas ramp would cause the Western Cordillera ramp to ramp up with a vertical component of slip much greater than the eastern Principal Cordillera and a localized rapid uplift. This is in agreement with fission track data from the Miocene plutons exposed in the western Principal Cordillera, compatible with <3 km of denudation of the El Teniente district since Late Miocene time (Maskaev *et al.* 2004), and the identification of a main stage of surface uplift in the Chilean slope during Late Miocene–Early Pliocene time with maximum vertical throw 0.7–1.1 km (Fariñas *et al.* 2008).

During Early Miocene–Early Pliocene time, the western sector of the Coastal Cordillera is submerged as evidenced by the marine deposits of the Navidad Formation and younger units (Encinas *et al.* 2008; Gutiérrez *et al.* 2013), and is subjected to local extensional stress field (Lavenu & Encinas 2005). Flexural elasticity of the rigid cold forearc region causes the western Coastal Range to subside while the eastern sector of the range is uplifted. This is in agreement with geologic evidence for subsidence between 10 and 4 Ma in this sector of the transect (Encinas *et al.* 2006).

The Pliocene (5–2.5 Ma)

During Pliocene time, the magmatic activity reassumes its current locus along the High Andean drainage divide. Around this time, the Main décollement becomes inactive and there is a lull in deformation in the Principal Cordillera, concomitant with the mineralization of the El Teniente porphyry whose magmatic-hydrothermal centre records 6 Ma of continuous activity during 9–3 Ma (Mpodozis & Cornejo 2012). Compressional deformation occurs only in the eastern Frontal Cordillera with generation of thrusts affecting the Miocene–Early Pliocene synorogenic deposits of the Cacheuta basin, and in the Cerrilladas Pedemontanas with the generation of an angular unconformity in the synorogenic units (Yrigoyen 1993; Irigoyen *et al.* 2000) and the inversion of the Anchayuyo fault (Giambiagi *et al.* submitted to Tectonics).

The compressional deformation in the Frontal Cordillera is coeval with movement along NE dextral and WNW sinistral strike-slip faults reported to affect the El Teniente porphyry copper deposits in western Principal Cordillera (Garrido *et al.* 1994). According to these authors, the strike-slip faults were active before, during and after the

formation of the giant porphyry Cu–Mo deposit (6.3–4.3 Ma, Maksaev *et al.* 2004), indicating a strike-slip regime during Early Pliocene time.

The widening of the crustal root during this stage favours the Coastal Cordillera uplifts by isostatic rebound, evidenced by the emergence of marine deposits between 4.4 and 2.7 Ma (Encinas *et al.* 2006) and the onset of knickpoint retreat in the western Coastal Range around 4.6 Ma (Fariás *et al.* 2008). This process would have partially blocked the drainage, inducing sedimentation in the Central depression (Fariás *et al.* 2008). The widening of the crustal root could in turn be responsible for the increase in topographic elevation by isostatic compensation. The enhanced erosion related to increased relief during 6–3 Ma suggested by Maksaev *et al.* (2009) could be attributed to this phenomenon.

The Quaternary (2.5 Ma–present)

The distribution of shallow earthquake epicentres in the Andes between 33° and 34°S gives important clues for the Quaternary deformation along the transect. In the Chilean slope of the Andes, shallow seismic activity is distributed mainly along the western flank of the Principal Cordillera at depths of 12 and 15 km (Barrientos *et al.* 2004), and beneath the Central depression at depths shallower than 20 km (Fariás *et al.* 2010). This suggests that at least the western portion of the Main décollement is active today. Beneath the Western Principal Cordillera, seismicity is concentrated on the steeper portion of the main décollement. We suspect that slip occurs aseismically on the gently dipping segments of this detachment zone. This is in agreement with minor Quaternary movement along the San Ramón fault, whose scarp has been linked to the vertical offsets (0.7–1.1 km) of the peneplains present in the western Principal Cordillera before 2.3 Ma (Fariás *et al.* 2008).

In the High Andes close to the Chilean–Argentine border, most shallow earthquakes of $M > 5$ show focal mechanisms related to strike-slip kinematics (Barrientos *et al.* 2004), such as the M_w 6.9 Las Melosas earthquake (Alvarado *et al.* 2005). Beneath the eastern part of the Frontal Cordillera and the Cuyo basin shallow focal mechanisms are related to compressional kinematics (Alvarado *et al.* 2007). The neotectonics of the Andean retrowedge at this latitude is characterized by movement along buried faults that fold the Quaternary deposits (García & Casa 2014).

During this phase, the backward tilting of the Principal and Frontal cordilleras by movement along the Portillo thrust system and the isostatic readjustment of the thickened crust could be responsible for the tilting to the west (1–3°) of several

flat erosional surfaces located between 2600 and 3200 m a.s.l. in the western Principal Cordillera (Fariás *et al.* 2008). We propose that the increase in gravitational potential energy due to crustal roots formation and mountain uplift prevents the main décollement from propagating eastwards. Instead, reactivation of the WPC back-thrusts should have implied less work against resisting stresses.

Discussion: implications of the model

The kinematic model presented in this work integrates structural, sedimentological, petrological, geochronological and geophysical data for the studied transect. The geological constraints available for the area allowed us to build a complete model that describes the relation between horizontal shortening, uplift, crustal thickening and activity of the magmatic arc, accounting for the main deep processes that have shaped the Andes since Early Miocene times. These geological and geochemical constraints reinforce previous hypotheses of a west-dipping detachment at the transition from brittle-elastic to ductile rheology in the crust below the Andean strip (Allmendinger & Gubbels 1996; Ramos *et al.* 2004; Fariás *et al.* 2010; Giambiagi *et al.* 2012), and are consistent with the east-vergent models. Furthermore, the model allows us to discuss some aspects of Andean history in light of our results.

Our model proposes passive surface uplift in the Coastal Range as the master décollement decreases its slip downwards transferring shortening to a broad area above the singularity point S, where the master detachment touches the Moho horizon (Figs 3 & 5). During the 18–5 Ma period, the main phase of deformation is located in the Aconcagua FTB. As the crust thickens from 40 to 50 km and the upper crust shortens to 52 km, the S point slowly migrates westwards (Fig. 5b). This migration generates a slowly propagating wave of surface uplift in the Coastal Range, from its eastern part at c. 18 Ma to its western part at the beginning of Pliocene time, in agreement with provenance studies from the Navidad basin (Rodríguez *et al.* 2012). Coeval uplift of the Coastal Range and Principal Cordillera creates the Central Depression, inducing thick sedimentary filling as suggested by Fariás *et al.* (2008). This explains why no important coastal-parallel faults have been observed in the Coastal Range, even although it experienced significant topographic uplift during Cenozoic time suggested by the exposure of the Miocene marine deposits (Encinas *et al.* 2008; Gutiérrez *et al.* 2013). Moreover, the presence of relict continental erosion surfaces with different elevations developed close to sea level and uplifted

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up to 2.1 km a.s.l. (Fariñas *et al.* 2008) indicates several pulses of Coastal Range uplift.

The eastward migration of the volcanic arc during Late Miocene–Quaternary time could be related to an eastwards migration of the trench due to tectonic erosion of the continental margin, as proposed by Stern (1989, 2011) and Kay *et al.* (2005); the delamination of an eclogitic root in the lower crust and mantle lithosphere (Kay & Kay 1993; DeCelles *et al.* 2009); and to shortening of the brittle crust above the major décollement. Migration of the volcanic arc can be explained by our model without invoking the decrease in the angle of subduction of the oceanic slab for this period of time, consistent with the proposal of Godoy (2005). Instead, the overall decrease in the volume of the asthenospheric wedge due to the construction of the crustal root could inhibit the influx of hot asthenosphere into the region and be responsible for the cooling of the subarc mantle and the eastward migration of the arc, as proposed by Stern (1989).

Underthrusting of the mechanically strong South American craton beneath the Andean strip leads to thickening of the crust. Kay *et al.* (1999) and Kay & Mpodozis (2001) argued that the transformation of hydrous lower-crustal amphibolite to garnet-bearing eclogite during crustal thickening can be responsible for the exsolution of fluids. These fluids can substantially decrease the strength of the lower crust. In this way, crustal thickening before Middle Miocene time may have favoured increased deformation during the Middle–Late Miocene (12–10 Ma) period, enhancing horizontal shortening in the Principal Cordillera and widening of the crustal roots. The forces that support the Andes provide an upper limit to the height of the mountain range and also to crustal thickness (Molnar & Lyon-Caen 1988). In our study area, the critical value seems to be of the order 50 km. Once this thickness is achieved, gravitational potential forces created by the buoyant crustal root are higher than unbalanced driving tectonic forces and therefore thrusting shifts eastwards into the lowlands. Even although convergence was steady, faults in the Principal Cordillera become inactive, the main décollement is deactivated and a new décollement is formed in the east to uplift the Frontal Cordillera.

Conclusions

In this paper we propose a kinematic model with thermomechanical constraints for the Miocene–present evolution of the Southern Central Andes at the latitude of the city of Santiago. Our model assumes a main décollement located at the base

of the upper crust (15–12 km in depth), which produces the underthrusting of the South American craton beneath the Andean strip. The total amount of horizontal shortening calculated with our master-detachment is 71 km, distributed between the western and eastern slopes of the Principal Cordillera (17 and 35 km, respectively), the Frontal Cordillera and foreland area (16 km). On the other hand, the Coastal Range undergoes passive surface uplift with only 3 km of shortening.

During the 18–5 Ma period, the main phase of deformation is located in the Aconcagua FTB. As the crust thickens from 40 to 50 km and the upper crust shortens by 52 km, the thrust front migrates from the Principal Cordillera to the Frontal Cordillera with a peak of deformation at c. 12–10 Ma. After the Andean crust achieves its present thickness of 50 km beneath the western Principal Cordillera, gravitational stresses drive the lateral expansion of the crustal root westwards and eastwards, driving surface uplift of the Central Depression and both slopes of the Principal Cordillera by isostatic response. Afterwards, during Pliocene time (5–2.5 Ma), there was a lull in deformation both in the eastern and western slope of the Andes with a deactivation of the master décollement. Uplift at this time is concentrated in the easternmost sector of the Frontal Cordillera and the Cachete basin. During Quaternary time, there is a reactivation of contractional deformation in the actual thrust front and in the Frontal range close to the city of Santiago as evidenced by seismological studies, suggesting the reactivation of the western sector of the main décollement.

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