The West Andean Thrust, the San Ramón Fault, and the seismic hazard for Santiago, Chile

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[1] The importance of west verging structures at the western flank of the Andes, parallel to the subduction zone, appears currently minimized. This hampers our understanding of the Andes-Altiplano, one of the most significant mountain belts on Earth. We analyze a key tectonic section of the Andes at latitude 33.5°S, where the belt is in an early stage of its evolution, with the aim of resolving the primary architecture of the orogen. We focus on the active fault propagation–fold system in the Andean cover behind the San Ramón Fault, which is critical for the seismic hazard in the city of Santiago and crucial to decipher the structure of the West Andean Thrust (WAT). The San Ramón Fault is a thrust ramp at the front of a basal detachment with average slip rate of ~0.4 mm/yr. Young scarps at various scales imply plausible seismic events up to M_w 7.4. The WAT steps down eastward from the San Ramón Fault, crossing 12 km of Andean cover to root beneath the Frontal Cordillera basement anticline, a range ~5 km high and >700 km long. We propose a first-order tectonic model of the Andes involving an embryonic intracontinental subduction consistent with geological and geophysical observations. The stage of primary westward vergence with dominance of the WAT at 33.5°S is evolving into a doubly vergent configuration. A growth model for the WAT-Altiplano similar to the Himalaya-Tibet is deduced. We suggest that the intracontinental subduction at the WAT is a mechanical substitute of a collision zone, rendering the Andean orogeny paradigm obsolete. **Citation:** Armijo, R., R. Rauld, R. Thiele, G. Vargas, J. Campos, R. Lacassin, and E. Kausel (2010), The West Andean Thrust, the San Ramón Fault, and the seismic hazard for Santiago, Chile, *Tectonics*, *29*, TC2007, doi:10.1029/2008TC002427.

1. Introduction

[2] The Andean orogeny is considered the paradigm for mountain belts associated with subduction plate boundaries [e.g., *Dewey and Bird*, 1970; *James*, 1971]. Yet, no mechanical model can explain satisfactorily the Andean mountain building process as a result of forces applied at its nearby subduction margin, along the western flank of the South America continent [e.g., *Lamb*, 2006]. Part of the problem arises from a geometric ambiguity that is readily defined by the large-scale topography (Figure 1): the Andes mountain belt is a doubly vergent orogen that has developed a large back thrust margin at its eastern flank, with opposite (antithetic) vergence to the subduction margin. To avoid confusion, the tectonic concept of subduction margin used here is equivalent to the proflank (or prowedge) concept used for collisional belts [*Malavieille*, 1984; *Willett et al.*, 1993; *Adam and Reuther*, 2000; *Vietor and Oncken*, 2005] and is preferred to the magmatic concept of fore arc, which has nearly coincident horizontal extent (Figure 1). Similarly, the notion of back thrust margin is used as an equivalent to that of retroflank (or retrowedge) in collisional belts.

[3] The doubly vergent structure of the Andes mountain belt is defined by distinct orogenic thrust boundaries at the East and West Andean fronts (Figure 1). While the East Andean Front coincides with the basal thrust of the back thrust margin over the eastern foreland (the South America continent), the orogenic West Andean Front is located at significant distance from the basal megathrust of the subduction margin. There is a wide western foreland (~200 km wide horizontally) separating the orogenic West Andean Front from the subduction zone, which is designated here as the Marginal (or Coastal) Block. Consequently, a fundamental mechanical partitioning occurs across the subduction margin and the marginal block, between the subduction interface, a megathrust that is responsible of significant short-term strains and the occurrence of repeated large earthquakes, and the West Andean Front thrust that appears important in regard to the long-term cumulative deformation and other processes associated with the Andean orogeny. However, very few specific observations are available at present to describe and to model this fundamental partitioning.

[4] It is generally admitted that the high elevation of the Andes and of the Altiplano Plateau result from crustal thickening (up to ~70 km thickness), which is associated with significant tectonic shortening (up to ~150–300 km shortening) and large-scale thrusting of the Andes over the South America continent (the South America craton plus other terrane accreted to the western margin of Gondwana in the Late Paleozoic), at the back thrust margin [*Wigger et al.*, 1994; *Allmendinger et al.*, 1997; *Kley and Monaldi*, 1998; *Kley*, 1999; *Kley et al.*, 1999; *Coutand et al.*, 2001; *ANCORP Working Group*, 2003; *Oncken et al.*, 2006]. On the other hand, the role of the subduction margin and of the West Andean Front in the thickening processes is often considered

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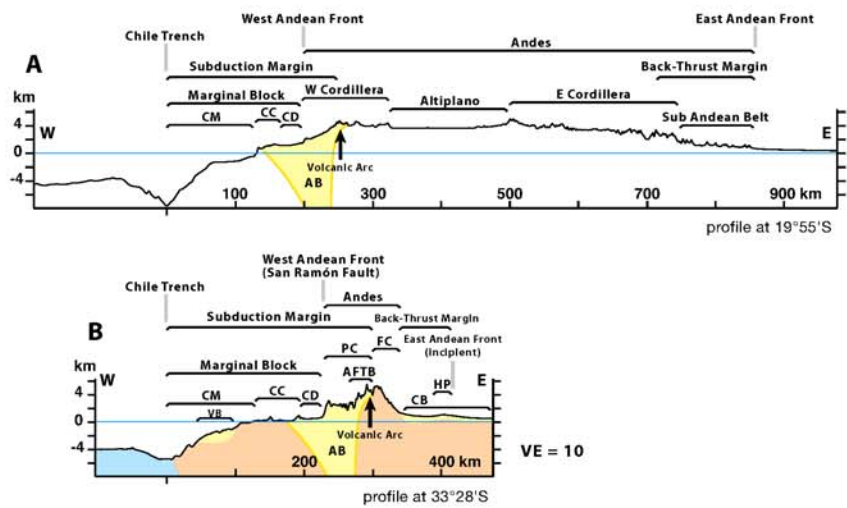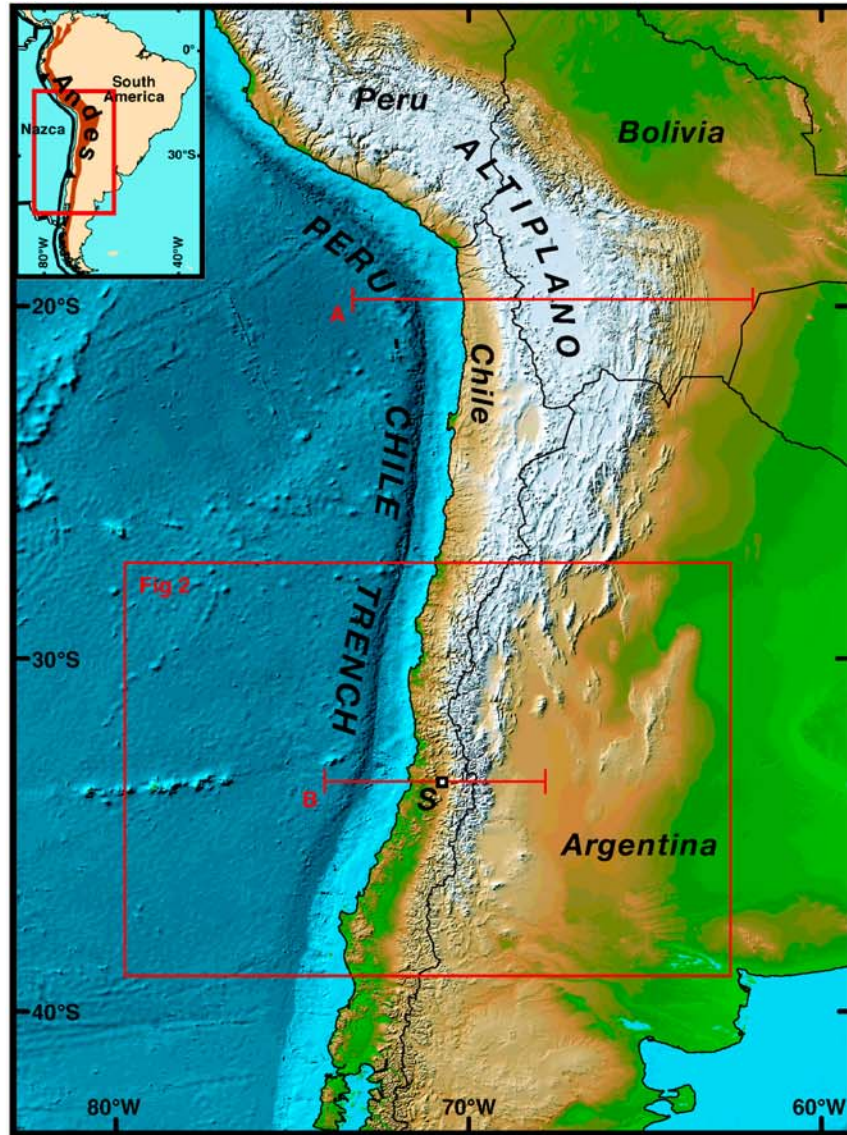[3]Departamento de Geofísica, Universidad de Chile, Santiago, Chile.

Figure 1

negligible [e.g., *Isacks*, 1988]. Yet the Andean Subduction Margin stands as one of the largest topographic contrasts on Earth (up to ~12 km), substantially larger than its back thrust counterpart (Figure 1). The present study is aimed at revising our knowledge of the large-scale tectonics of the Andes and its interaction with subduction processes. So we specifically deal with the overlooked West Andean Front associated with the subduction margin and we attempt to reassess its relative importance during the Andean orogeny. Purposefully, we choose the region of the Andes facing Santiago, because it includes a key section of the Andes crossing a key structure: the San Ramón Fault.

[5] We analyzed and revised critically the Geomorphology and the Geology of the Andes covering the region near Santiago between ~33°S and ~34°S (Figures 1 and 2a) and focusing on morphologically active tectonic features to assess the seismic hazard associated with the West Andean Front. Santiago nestles in the Central Depression, which for long has been described as an extensional graben, bounded to the east and west by normal faults [*Brüggen*, 1950; *Carter and Aguirre*, 1965; *Thiele*, 1980]. In our work, we show that the San Ramón Fault, crossing the eastern outskirts of Santiago, is a major active fault with many kilometers of thrust slip [*Rauld*, 2002; *Rauld et al.*, 2006; *Armijo et al.*, 2006]. The West Andean Front as defined by the San Ramón Fault is precisely where the Quaternary and older sediments of the Central Depression are overthrusted by the deformed rocks of the Andes Principal Cordillera.

[6] Our study of the San Ramón Fault aims at describing fault scarps at a range of scales (meters to kilometers) along with uplift of datable morphological surfaces to determine slip rates over a range of ages (10^3 to 10^7 years). We combine high-resolution air photographs and digital topographic data with a detailed field survey to describe the morphology of the piedmont and fault scarps across it. Large cumulative scarps and single event scarps can be identified and mapped with good accuracy. Fault parameters (length of segments, fault dip, and possible fault slip rate) can be discussed with a view to assess seismic hazard. The multikilometric-scale folding of the San Ramón structure during the past tens of million years can be used to constrain the thrust geometry to depths down to ~10 km and more.

[7] At the large scale, key tectonic observations were gathered and analyzed critically throughout the study region, to incorporate our observations of the San Ramón Fault into a complete tectonic section across the Andes, from the Chile Trench to the stable basement of South America (see location in Figures 1, 2a, and 2b). This unifying approach allows us to set together, strictly to scale, the most prominent Andean tectonic features, specifically the West Andean Front, which as we show, appears associated with the large-scale West Andean Thrust (WAT). We discuss the main results emerging from this study, particularly the true geometry and possible tectonic evolution of this segment of the Andes, which allow us to reassess the role of the subduction margin and to suggest a broad range of implications that challenge the Andean orogeny paradigm.

2. Tectonic Framework

[8] The deformation styles generally described along the Andes are based almost exclusively on the structure of its back thrust margin [*Kley et al.*, 1999; *Ramos et al.*, 2004]. Two different large-scale sections can be used to characterize the doubly vergent margins of the Andes; one at 20°S latitude crossing where the belt is largest and its structure fully developed, the other at 33.5°S where the belt is relatively narrow and less developed (Figure 1). The first section (profile A) crosses the largest Andean back thrust, namely, the sub-Andean Belt, which is a thin-skinned thrust belt detached over the basement of stable South America [e.g., *Mingramm et al.*, 1979; *Allmendinger et al.*, 1983; *Kley*, 1996; *Schmitz and Kley*, 1997]. Clearly, the basal detachment of the sub-Andean Belt (reaching the surface at the East Andean Front, Figure 1) is very distant from the trench (850 km), thus also from forces applied across the subduction plate boundary. The second section (profile B, Figure 1; see also Figures 2a and 2b for location of tectonic elements) includes another classical example of Andean back thrust, which is the thin-skinned Aconcagua Fold-Thrust Belt (AFTB) [*Ramos*, 1988; *Ramos et al.*, 1996b; *Giambiagi et al.*, 2003; *Ramos et al.*, 2004]. However, this belt is not located along the eastern flank of the Andes Mountains, but right in the middle of them (see profile B in Figures 1, 2a, and 2b). The basal detachment of the AFTB is shallow (~2–3 km depth) and its front found at high elevation (~4000 m), atop a huge basement high of the Andes (the Frontal Cordillera). So at present there is no clear flexural foreland basin directly in front of the basal detachment [*Polanski*, 1964; *Ramos*, 1988; *Ramos et al.*, 1996b; *Giambiagi*

Figure 1. Topography and very rough geology of the central Andes. Red box locates Figure 2. Square marked with S locates Santiago. The main tectonic features are identified on two selected profiles (A and B, traces marked in red at 20°S and 33.5°S). Vertical black arrows indicate the present-day volcanic arc. The subduction margin (synthetic to subduction) coincides with fore-arc extent. The sub-Andean Belt in profile B is part of the back thrust margin, (antithetic to subduction). The Principal Cordillera (PC) includes the Aconcagua Fold-Thrust Belt (AFTB), both made of volcanic/sedimentary rocks of the Andean Basin (AB) overlying basement of the Frontal Cordillera (FC). The relatively shallow Cuyo Basin (CB) overlies the basement structure of the Hidden Precordillera (HP). The marginal block is formed of Central Depression (CD), Coastal Cordillera (CC), and continental margin (CM). Profile B depicts major crustal features deduced from the geology in light colors: Triassic and pre-Triassic continental basement (brown), post-Triassic basins (yellow) and oceanic crust (blue). The deep basin represented in the two profiles (A and B) is the Andean Basin (AB) that is crossed by the trace of the West Andean Front. VP is Valparaíso Basin. Vertical exaggeration (VE) in profiles is 10. Map and profiles are based on topographic data from the NASA Shuttle Radar Topography mission (SRTM) and the global grid bathymetry of *Smith and Sandwell* [1997] (available at http://topex.ucsd.edu/WWW_html/srtm30_plus.html).
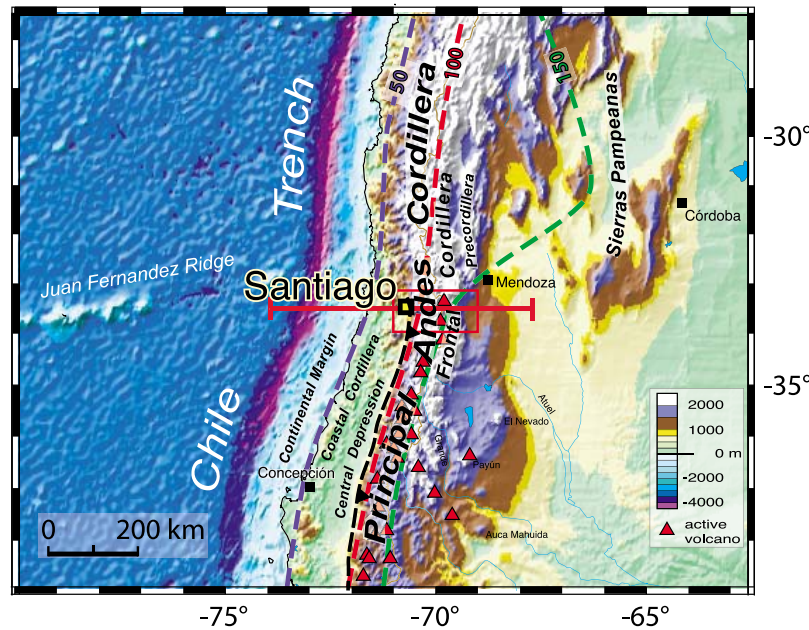
Figure 2a. Tectonic framework and physiography of the Andes and the Nazca–South America plate boundary and the study region. The capital of Chile (Santiago) is located at the northern end of the Central Depression where the West Andean Front is well defined (black dashed line with triangles, compare with profile B in Figure 1). The Andes Principal Cordillera overthrusts the 230–250 km wide marginal block, formed south of Santiago of Continental Margin, Coastal Cordillera, and Central Depression, which over-thrusts the Nazca Plate at the subduction zone (trace marked by Chile Trench). At 33°S latitude, north–south changes in Andean tectonics, topography, and volcanism appear associated with changes in the shape of the trench and in the dip of the subducting slab (Nazca Plate under South American Plate), which in turn may be associated with subduction of the Juan Fernandez Ridge. Active volcanism (red triangles) is limited to the region south of 33°S. Thick dashed lines (blue, red, and green) are the horizontal projection of lines of equal depth at the top of the Benioff zone (at 50, 100, and 150 km, respectively), as deduced by seismicity studies [*Cahill and Isacks*, 1992]. At 33°S the strike of the Benioff zone bends ~15° as defined from the trench down to 100 km depth (from N20°E south of 33°S to N5°E north of 33°S). South of 33°S the lower part of the subducting slab (between 100 and 150 km depth) appears to dip steeply (~35°), whereas for the same depth range immediately north of that latitude a flat slab geometry is suggested [*Cahill and Isacks*, 1992]. Our study region (displayed in Figure 2b) is outlined by red box. Red line at 33.5°S corresponds to topographic profile B in Figure 1 and to sections in Figures 3 and 8.

et al., 2003]. Therefore, identifying the AFTB with the main back thrust margin of the Andes is problematic. Mitigating this problem, a late thick-skinned basement thrusting at the eastern flank of the Frontal Cordillera is proposed [*Giambiagi et al.*, 2003; *Ramos et al.*, 2004].

[9] In contrast with back thrusts, synthetic thrusts along the western flank of the Andes are poorly known. However, the two sections used for comparison (profiles A and B, Figure 1) reveal a clear, continuous West Andean Front that is expressed in the topography at ~200 km distance from the trench and which appears larger and sharper than features at the same latitude along the back thrust margin. The paucity of seismic activity associated with this major synthetic contact may be a real feature, but may also result, at least in the region south of 33°, from lack of an appropriate local network [*Pardo et al.*, 2002; *Barrientos et al.*, 2004]. The most studied part of the West Andean Front is in northern Chile (along profile A, Figure 1), where large volumes of

Neogene volcanic rocks blanket its structure and obscure its tectonic significance, remains under debate [*Isacks*, 1988; *Muñoz and Charrier*, 1996; *Victor et al.*, 2004; *Farías et al.*, 2005; *García and Hérail*, 2005; *Hoke et al.*, 2007]. By contrast, the West Andean Front is particularly well exposed by the tectonic section at the latitude of Santiago, capital of Chile (profile B in Figure 1), which is the region retained for this study. The physiographic map in Figure 2a displays the main tectonic belts in that region and the structural map of the Andes in Figure 2b focus on its fundamental elements, as defined at 33°S–34°S latitude.

[10] Some characteristic elements of the central Andean physiography have been defined in the region around Santiago. South of 33°S and for more than 1000 km, the western 230–250 km wide fraction of the subduction margin between the trench and the western flank of the Principal Cordillera is characterized by three parallel zones (Figures 1 and 2a): an offshore continental margin ~100–130 km wide; a
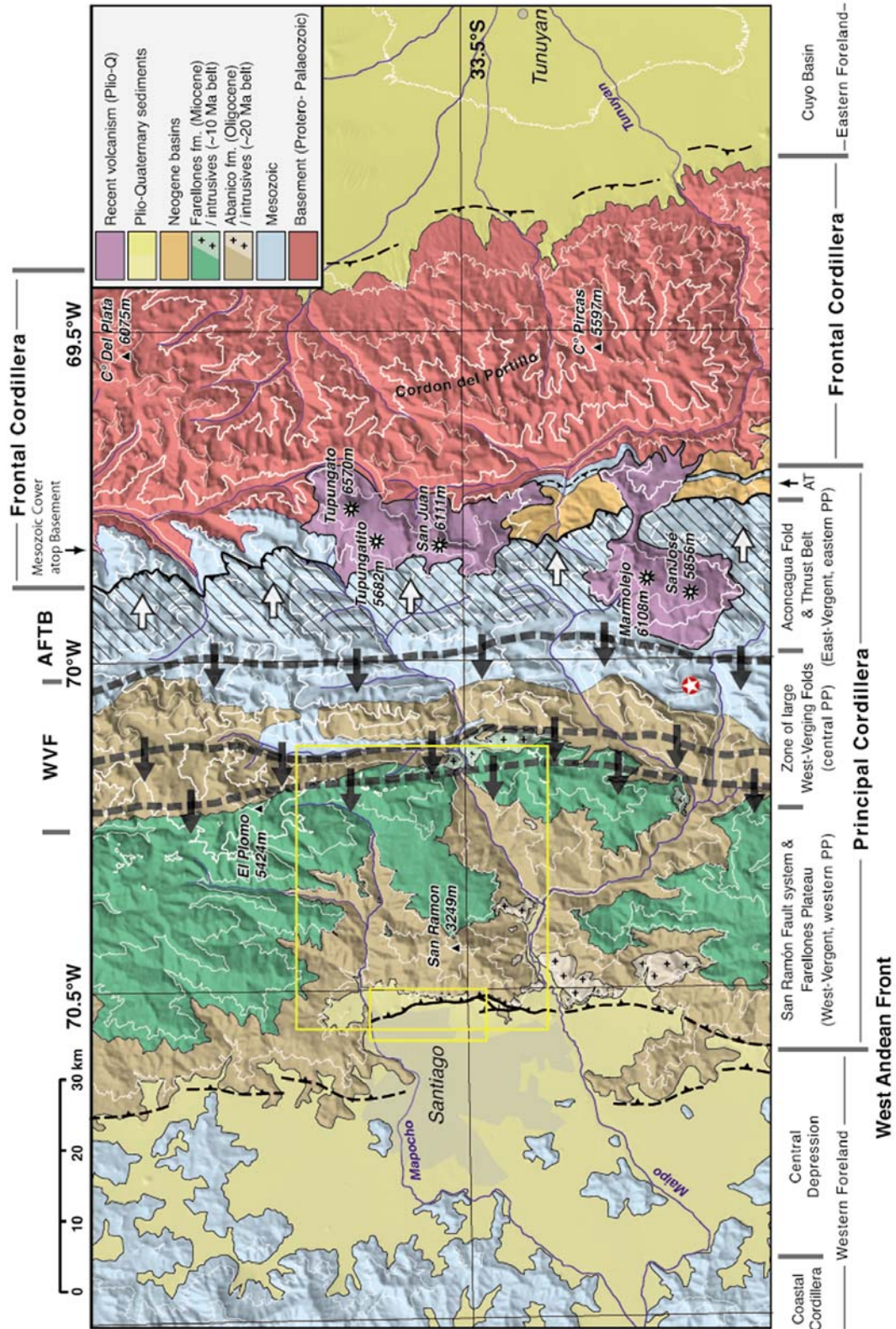
Figure 2b

coastal cordillera ~30–60 km wide, made of mature landforms peaking at more than 2000 m elevation; and a flat central depression ~30–60 km wide and averaging ~500 m in elevation near Santiago, which is filled with less than 1 km of quaternary sediments. Structurally those three zones together represent the western foreland of the Andes (the marginal or coastal block). East of the marginal block, a large region with elevation over 1 km associated with the Andes extends at this latitude over a total width of ~200 km. However, the Andes mountain belt strictly, with elevation >2 km, is restricted to a narrower western belt, which is only about 100 km wide, including the Principal Cordillera (which in turn includes the AFTB) and the Frontal Cordillera (Figures 1, 2a, and 2b). Santiago is located in the Central Depression facing the West Andean Front of the Principal Cordillera where it is particularly well defined by the topography (profile B, Figure 1). Despite its accepted designation, the "Frontal" Cordillera is flanked to the west by the Principal Cordillera and to the east (north of 33°S) by the Precordillera (Figure 2a), so it appears located far from any major structural front (i.e., the East and West Andean fronts; profile B, Figure 1). The tectonic significance of the Frontal Cordillera appears capital and thus is addressed along with discussions in this section and in sections 4 and 5.

[11] The study region is immediately south of 33°S latitude (Figures 2a and 2b) where significant lateral (along-strike) changes in the Andean tectonics, topography and volcanism appear associated with changes in the shape of the trench and in the dip of the subducting slab, i.e., the Nazca Plate under the South American Plate [Isacks, 1988, Cahill and Isacks, 1992]. Those changes may in turn be associated with subduction of the Juan Fernandez Ridge [von Huene et al., 1997; Gutscher et al., 2000; Yañez et al., 2001]. At 33°S, the average strike of the Benioff zone bends ~15° as defined from the trench down to 100 km depth (from N20°E south of 33°S to N5°E north of 33°S). South of 33°S, the lower part of the subducting slab (between 100 and 150 km depth) appears to dip steeply (~35°) whereas for the same depth range a flat slab geometry is suggested immediately north of that latitude [Cahill and Isacks, 1992]. This change in the subduction geometry would explain the presence of active volcanism to the south of 33°S and its absence north of 33°S, where volcanism has ceased at ~10 Ma [Kay et al., 1987; Isacks, 1988].

[12] The physiography of the west Andean flank also changes to some extent in front of the "flat slab" segment (between 33°S and 27°S). The western boundary of the Principal Cordillera shifts westward, the Central Depression disappears and the West Andean Front is expressed by a wide, gradual topographic contrast. North of 27°S the West Andean Front is again very sharp in the topography (profile A, Figure 1). Changes of the physiography in front of the "flat slab" segment are more significant along the eastern flank of the Andes, where the Precordillera and the Sierras Pampeanas are defined (Figure 2a). The Precordillera is a belt located to the east of the Frontal Cordillera and separated from it by a narrow depression (Uspallata-Iglesia basin). It is described as a thin-skinned back thrust belt involving Paleozoic rocks (Paleozoic cover of Cuyania terrane [Ramos, 1988]) and absorbing significant tectonic shortening, similar to the sub-Andean Belt [Allmendinger et al., 1990]. East from the Precordillera are found the thick-skinned Sierras Pampeanas, which correspond to several thrust blocks within the basement of Gondwanan South America [Allmendinger et al., 1990; Ramos et al., 2002; Ramos, 1988] (see Figure 2a for location of these belts).

[13] Geologically, the front of the Principal Cordillera east of Santiago crosses the deep, very long (several 10^3 km long) and relatively narrow (of the order of 10^2 km wide) Andean Basin that can be followed nearly parallel to the Andes between the equator and latitude 48°S [Mpodozis and Ramos, 1989; Vicente, 2005]. This huge feature (in yellow and labeled AB in profiles A and B, Figure 1) has earlier been called the Andean "Geosyncline" [e.g., Aubouin et al., 1973] and appears closely associated with subduction and Andean cycle orogenic processes operating continuously along the western margin of the South America continent since the Jurassic. The Andean Basin is formed of Early Jurassic to Miocene volcanics, volcanic-derived rocks, clastics and some marine rocks with an overall thickness of ~12–15 km or more [Mpodozis and Ramos, 1989; Robinson et al., 2004] (for a thorough stratigraphical review, see Charrier et al. [2007]). Those rocks have been deposited over the pre-Andean basement assemblage consisting of

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Figure 2b. Structural map of the Andes in the study region compiled from geological information at diverse scale, in Chile [from Thiele, 1980; Gana et al., 1999; Sellés and Gana, 2001; Fock, 2005], in Argentina [from Polanski, 1964, 1972; Giambiagi et al., 2001; Giambiagi and Ramos, 2002], and our own observations. The Principal Cordillera is subdivided in three units (western PP, central PP, and eastern PP) according to vergence of observed structures (discussed in the text). The thin-skinned Aconcagua Fold-Thrust Belt (AFTB) is hatched with frontal fault trace in black, adorned with triangles. Its main eastward vergence is indicated by white arrows. The zone of large west verging folds (WVF) is illustrated by dashed black stripes adorned with westward directed arrows, indicating the approximate position of basal layers of main nearly vertical limbs (with top-to-the-west geometry). The red-circled white star indicates location of Neocomian beds in Figure 9. The shallow westward dipping basal contact of the Mesozoic cover (eastern side of Andean Basin) is clearly seen atop the Frontal Cordillera basement anticline (towering at more than 6000 m) in the northern half of the map. Two outcrops of the relatively small intermontane Alto Tunuyán basin (AT) are seen in the southern half of the map (consistent with maps by Giambiagi et al. [2001] and Giambiagi and Ramos [2002]). The San Ramón Fault marks the synthetic West Andean Front, at the contact between the Principal Cordillera and the Central Depression (western foreland). At this latitude, the east verging, thick-skinned back thrust that disrupts the contact of the Frontal Cordillera with the Cuyo basin (eastern foreland) is mostly hidden. Elevation contours (white, each 1000 m; thicker contour for 4000 m) derived from SRTM data. The yellow rectangles locate maps in Figures 3b (larger rectangle) and 4 (smaller rectangle).

magmatic and metamorphic rocks amalgamated, during the Early to Middle Paleozoic, with the western margin of Gondwana (south of 15°S under the Andean Basin are found specifically the Arequipa-Antofalla, the Mejillonia and the Chilenia terranes [e.g., *Hervé et al.*, 2007; *Charrier et al.*, 2007; *Lucassen et al.*, 2000; *Ramos*, 2008; *Vaughan and Pankhurst*, 2008]). At the latitude of Santiago, the western side of the Andean Basin along the Coastal Cordillera overlays the extreme west margin of the pre-Andean Gondwana basement (west margin of the Chilenia terrane), which is made of a metamorphic accretionary prism system (involving rocks of Paleozoic and possibly Late Proterozoic age [see *Hervé et al.*, 2007, and references therein]), intruded by a granitoid batholith of Late Paleozoic age [e.g., *Mpodozis and Ramos*, 1989; *Hervé et al.*, 2007]. The thick rock pile that has accumulated in the Andean Basin has undergone extensive burial metamorphism with low-grade, subgreenschist facies [e.g., *Levi et al.*, 1989; *Aguirre et al.*, 1999; *Robinson et al.*, 2004, and references therein]. The Andean Basin started to form in a back-arc environment, which lasted stable until the end of the Mesozoic [*Mpodozis and Ramos*, 1989; *Charrier et al.*, 2007]. Then progressive eastward migration of the magmatic arc during the Late Cretaceous and Cenozoic, which may be associated with subduction erosion processes [*Coira et al.*, 1982; *Scheuber et al.*, 1994; *Kay et al.*, 2005; *Charrier et al.*, 2007], has ultimately put most of the basin in its present fore-arc position (see profiles A and B, Figure 1).

[14] The first-order structure of the Andean Basin in our study region is simple, but with a clear asymmetry. In its western (external or coastal) flank, the Jurassic rocks at the base of it rest unconformably, with relatively shallow eastward dip (~25° on the average, but increasing eastward to maximum values of ~40°), on top of the pre-Andean basement rocks of the marginal block, which crop out at relatively low elevation (no more than ~300 m) in the Coastal Cordillera. In its eastern (internal or Andean) flank, the equivalent Jurassic rocks at the base of the Andean Basin rest unconformably on top of basement rocks of similar pre-Jurassic age, but situated in a structural high cropping out at more than 5 km elevation in the Frontal Cordillera (Figures 1 and 2b). The Frontal Cordillera high is a large arched ridge elongated in N–S direction, of which the top surface is made of the Permian-Triassic Choiyoi Group. These rocks overly the magmatic and metamorphic, Protero-Paleozoic, Gondwana basement (considered as Chilenia terrane in the Frontal Cordillera [see *Mpodozis and Ramos*, 1989; *Heredia et al.*, 2002; *Llambías et al.*, 2003]). The basal sedimentary contact of the Andean Basin rocks over Triassic and basement rocks of the Frontal Cordillera has shallow westward dip (Figure 2b). However, west of the Frontal Cordillera, the sediments of the Andean basin in the high Principal Cordillera (which includes the AFTB) are strongly deformed with dominant very steep westward dip across most of the AFTB. That steep westward dip appears the main characteristic of the eastern flank of the Andean Basin (Figure 2b) and a result of strong pervasive shortening across the whole Principal Cordillera. Hence, overall the Andean Basin emerges as a crustal-scale asymmetric syncline inclined westward, located west of the main basement high of the Andes (the Frontal Cordillera) and thus structurally constituting the original basin formed in the western foreland, west of the Andean belt. However, the present-day mountain front of the Principal Cordillera intersects that large syncline in the middle, so reducing the actual width of the foreland (profile B, Figure 1). The same tectonic configuration is observed 800 km northward along strike for the subduction margin of the Andean belt (profile A, Figure 1).

[15] Summarizing the foregoing, the West Andean Front near Santiago appears as a major tectonic contact between the Principal Cordillera, which corresponds to the pervasively shortened eastern side of the Andean Basin, with significant westward dip, and the marginal block, which constitutes the relatively shallow dipping western side of the Andean Basin and is devoid of significant Andean deformation [*Thiele*, 1980] (Figure 1). It has been shown recently that this fundamental contact is not a normal fault as stated by the general belief for more than half a century [*Brüggen*, 1950; *Carter and Aguirre*, 1965; *Thiele*, 1980; *Nyström et al.*, 2003], but an important thrust system [*Rauld*, 2002; *Rauld et al.*, 2006; *Armijo et al.*, 2006]. The observations of the San Ramón Fault along the West Andean Front in section 3 define constraints on the tectonic mechanisms by which the significantly shortened Principal Cordillera overthrusts the relatively rigid western foreland represented by the marginal block, which in turn overthrusts the even stiffer Nazca Plate at the subduction zone.

3. San Ramón Fault

3.1. Basic Observations

[16] Cerro San Ramón is the 3249 m high peak on the mountain front that overlooks the city of Santiago from the east and gives the name to the fault at its base (Figure 3a). The corresponding west vergent San Ramón frontal structure, which is interpreted and discussed hereafter, is determined by structural elements mapped accurately in Figure 3b, which are displayed in an E–W section constructed below the structural map in Figure 3b and extended for tectonic interpretation in Figure 3c. The trace of the San Ramón Fault bears a morphological scarp that is readily visible across the piedmont in the outskirts of the city [*Tricart et al.*, 1965; *Borde*, 1966]. However, the detailed mapping of the San Ramón Fault has been prompted only once it was recently identified as an active thrust representing seismic hazard for Santiago [*Rauld*, 2002; *Rauld et al.*, 2006; *Armijo et al.*, 2006] (Figure 4) (see the auxiliary material).[1] Techniques used to describe the San Ramón Fault include satellite imagery, high-resolution air photographs and digital elevation models (DEMs) at three different scales, which were combined systematically with published geological maps [*Thiele*, 1980; *Gana et al.*, 1999; *Sellés and Gana*, 2001; *Fock*, 2005] and detailed field observations (e.g., Figure 5). Various sources of accurate digital topography (particularly SRTM with 90 m horizontal resolution and a DEM with ~30 m resolution derived from 1:25,000 scale local maps) and imagery (Landsat 7, ortho-rectified georeferenced SPOT images with resolution of 5 m, and aerial photographs) have

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[1]Auxiliary materials are available in the HTML. doi:10.1029/2008TC002427.
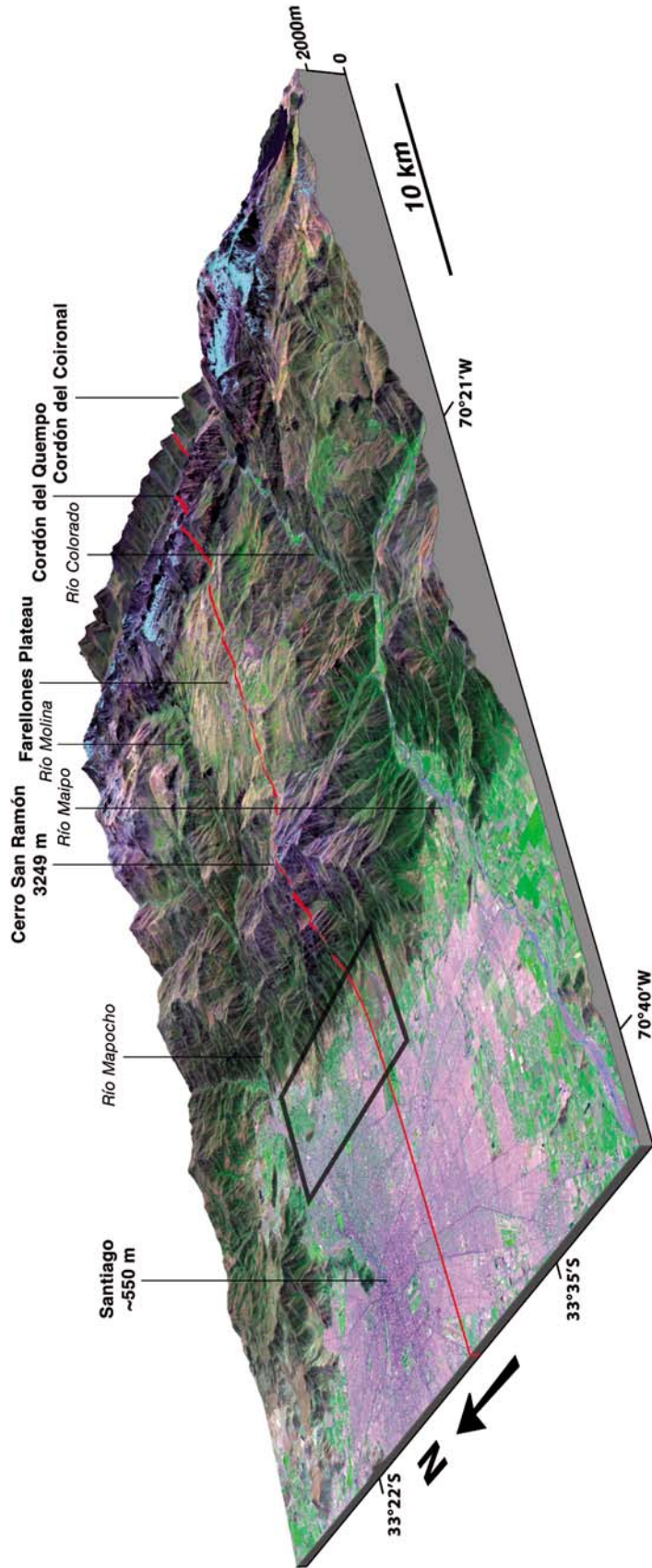
Figure 3a. Morphology and structure of the West Andean Front and the western Principal Cordillera facing Santiago. Three-dimensional view of DEM with Landsat 7 imagery overlaid. Oblique view to NE. The most frontal San Ramón Fault reaches the surface at the foot of Cerro San Ramón, across the eastern districts of Santiago. Rectangle shows approximate area mapped in Figure 4, which gives details of the fault trace. To the east of Cerro San Ramón, Farellones Plateau is incised ~2 km by rios Molina-Mapocho and Colorado-Maipo, which grade to the Central Depression (Santiago basin). Red line marks approximate trace of section shown in Figures 3b and 3c.
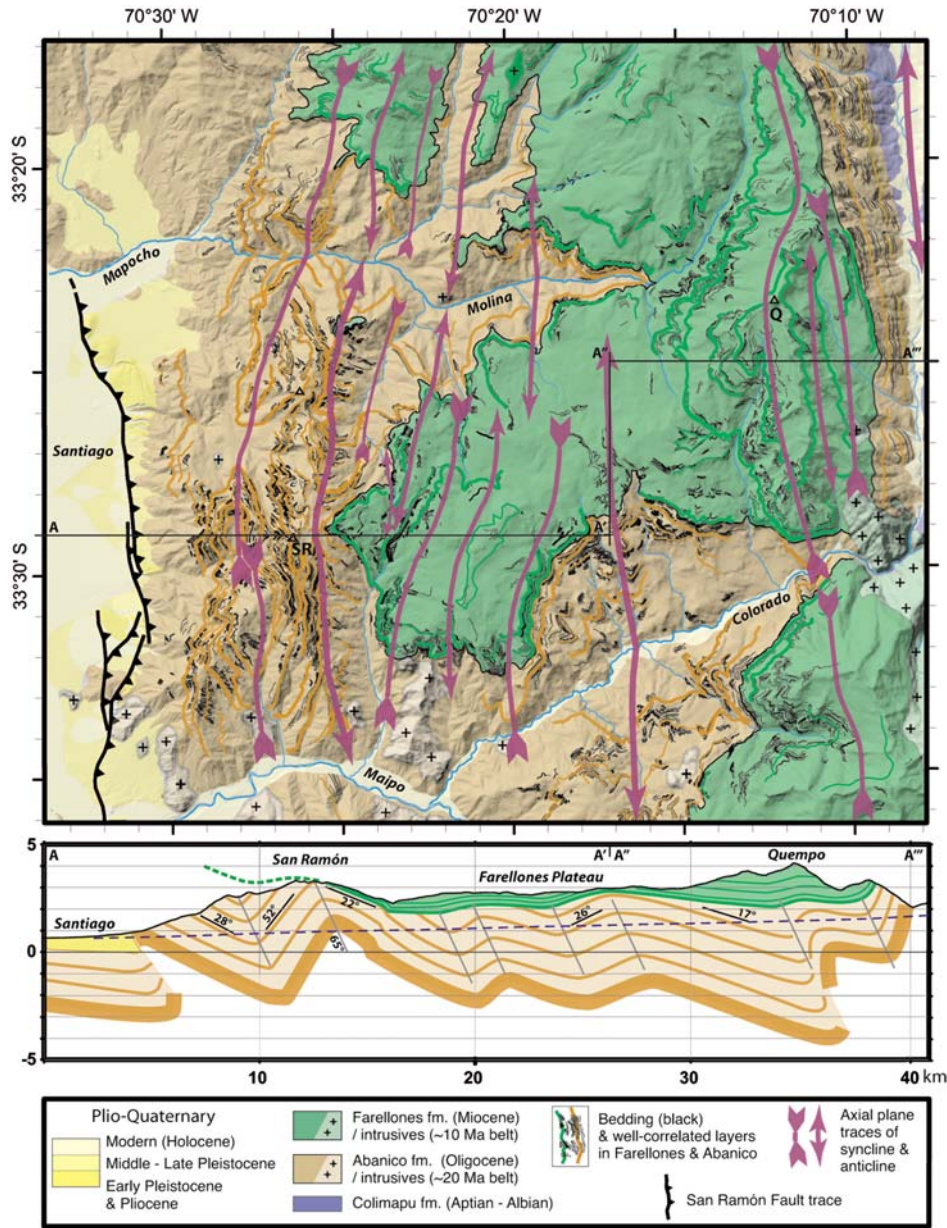
Figure 3b.   Structural map of the San Ramón–Farellones Plateau region and corresponding E–W section (see location in Figure 2b). Bedding attitudes are determined by mapping systematically over a DEM the most visible layers (thin black lines) with overall horizontal resolution of 30 m, using SPOT satellite imagery, aerial photographs, and field observations. Well-correlated layers (correlated over distances of kilometers) are indicated by thicker brown and green lines (in Abanico and Farellones formations, respectively). The basal contact of Farellones Formation has been modified from *Thiele* [1980], to be consistent with details of the mapped layered structure. The trace of axial planes of main folds is indicated. SR locates summit of Cerro San Ramón (3249 m); Q locates summit of Cordón del Quempo (4156 m). Section shift (from AA′ to A″A‴) is chosen to better represent the fold structure of San Ramón and Quempo. Bedding and axial plane attitudes in the section are obtained by projection of structural elements mapped over the DEM, directly constrained by the surface geology between the high mountainous topography and the longitudinal profile of Colorado River (dashed blue), and extrapolated below that level. The geometry of the base of the Abanico Formation is consistent with the structure observed in the upper part of the section, and it has been tentatively interpolated between where it pinches out with shallow eastward dip beneath sediment of the Santiago Basin (see Figure 3c) and its outcrop with steep westward dip in the Olivares river valley, east of Quempo.
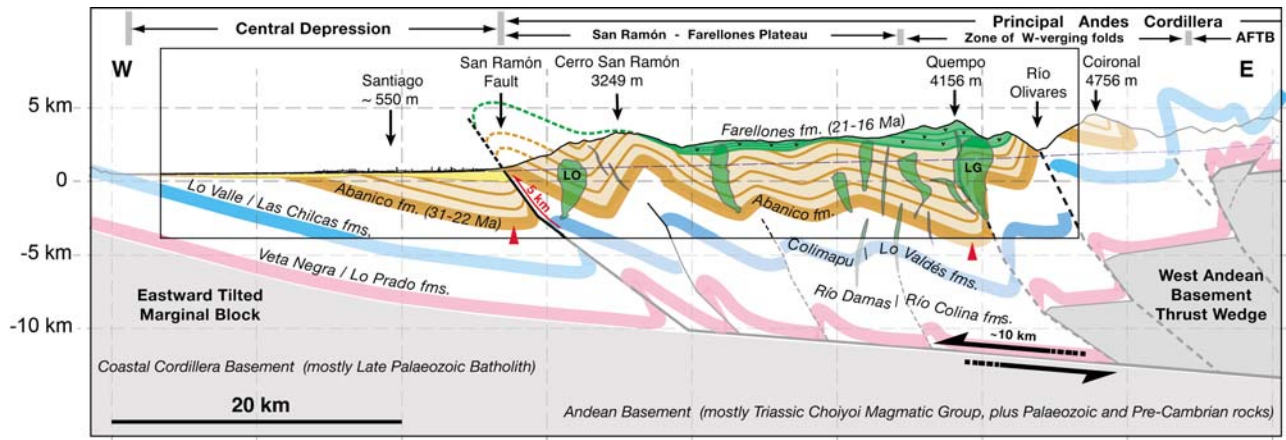
Figure 3c. Interpreted structure of the western Principal Cordillera associated with the east dipping ramp flat geometry of the San Ramón Fault. Box with plain colors shows surface geology, exactly as in the structural section constructed to −4 km in Figure 3b. Geology inferred farther downward and observed sideward is shown in opaque colors. Our interpretation implies that the western Andean front is characterized by the west vergent fold thrust structure of the >12 km thick volcanic-sedimentary cover of the Andean Basin (Jurassic-Cenozoic), which is pushed over the eastward tilted Marginal Block (or coastal block, formed of Central Depression warp atop Coastal Cordillera basement) by the (backstop or "bulldozer-like") the West Andean Basement Thrust System (involving magmatic rocks of Triassic and older age). A low-angle basal decollement would be localized at the ductile evaporitic layers of Late Jurassic age (Río Colina Formation; "Yeso principal del Malm"). The San Ramón Fault alone has net thrust slip of ~5 km (red half arrow), corresponding to the offset of the base of the Oligocene–early Miocene Abanico Formation. The total horizontal shortening by folding and thrusting across the San Ramón–Farellones Plateau structure (between the two red triangles) results from ~10 km westward transport of the volcanic-sedimentary pile relative to the underlying basement (bold double arrows). The larger overall shortening associated with the West Andean Thrust (>30 km) must be rooted in a major east dipping ramp crossing the basement and deep crust underneath the high Andes (to the east, outside of the shown section, see Figure 8). Main Miocene plutons are indicated: La Obra (LO, ~20 Ma) and La Gloria (LG, ~10 Ma).

been used to constrain the morphology and the structure at the scale of kilometers (Figures 3a and 3b). For the piedmont scarp, a DEM based on photogrammetry (horizontal resolution of 10 m; vertical precision of 2.5 m) (Figure 6) and various sets of aerial photographs were used (Servicio Aerofotogramétrico (SAF) Fuerza Aérea de Chile, scales 1:70,000 (1995) and 1:50,000 (1997)). Because large parts of the piedmont scarp are now obliterated by human settlement, old air photographs (taken in 1955 by Instituto Geográfico Militar, scale 1:50,000) were used to determine the exact position of the fault trace. So the morphological and tectonic features could be mapped at 1:5000 scale, then the overall information compiled in a map at 1:25,000 scale (Figure 4). To characterize a smaller-scale fault scarp across young alluvium, a higher-resolution DEM (horizontal resolution of 2 m; vertical precision of 10 cm) was created over an area of ~400 × 300 m^2 using a DGPS survey (Figure 7).

3.2. Multikilometric Frontal Thrust: Shallow Structure, Morphology, and Stratigraphy

[17] The shortening structures affecting the Cenozoic sequences of the western Principal Cordillera are very well

defined and well exposed in the San Ramón massif and the Farellones Plateau, which are located immediately eastward of the San Ramón Fault (Figure 3b). That frontal shortening appears associated with significant structural uplift of the Principal Cordillera relative to the Central Depression. A first-order measure of that uplift is given by the recent morphological evolution, in particular, the incision of deep canyons across the Farellones Plateau by ríos Mapocho and Maipo, and their tributaries Molina and Colorado, respectively (Figure 3a). The reasoning behind the foregoing is as follows. The total incision observed at the present is of the order of ~2 km and it has necessarily occurred since the time the Farellones Plateau formed as a continuous and relatively flat surface (now at elevation of ~2200–2500 m) atop a pile of Miocene volcanic lava flows (Figure 5a). Mostly coevally with the incision of the canyons, the ríos Mapocho and Maipo appear to have discharged sediment in the Santiago basin (Central Depression), where a maximum of ~500 m of alluvium and colluvium of Quaternary and possibly late Neogene age has accumulated [*Araneda et al.*, 2000]. The Central Depression therefore represents a relative base level where at least part of the sediment supply
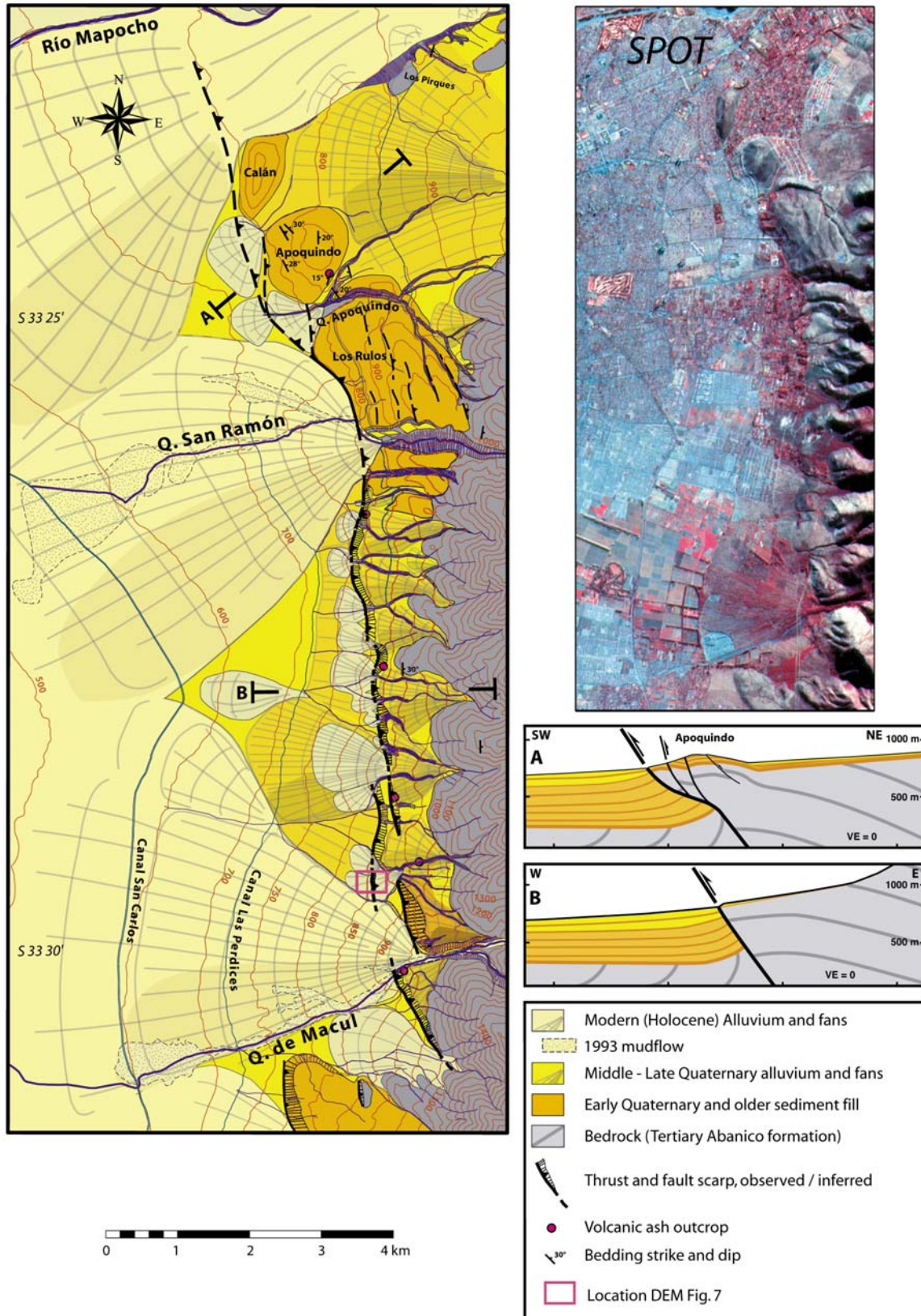
Figure 4

provided by erosion of the Andes Principal Cordillera is trapped. However, most of that sediment supply transits through the Central Depression and Coastal Cordillera to be deposited on the continental margin and ultimately in the trench [e.g., *von Huene et al.*, 1997]. Conversely, there is very little recent incision in the Central Depression, which behaves mostly as a sedimentary basin. A mature relief with relatively minor young incision by rivers characterizes the rest of the eastward tilted marginal block, formed by the eastward dipping edge of the Andean Basin on top of Coastal Cordillera basement rocks (Figure 3c). This structure suggests that the whole marginal block has experienced moderate, non-uniform, uplift associated with the eastward tilt, and gradual erosion, possibly throughout a significant part of the Cenozoic. Hence, the cause of the vigorous young incision of the Farellones Plateau and more generally of the whole Principal Cordillera must be a relative base level drop localized at its boundary with the Central Depression, so associated with thrusting and relative uplift at the West Andean Front, deduced only from the morphology, of at least ~2 km.

[18] The young uplift across the West Andean Front can be explained by some fault geometry and kinematics, which evolve during a certain time span. The relatively simple fold thrust structure of the San Ramón massif–Farellones Plateau, representative of the western Principal Cordillera (Figure 2b), is particularly fitted to the purpose, because it can be readily interpreted as a growing west vergent fault propagation fold system (Figures 3b and 3c). Kinematic models describing faults propagating through layered rocks and generating folds ahead of their tip lines [e.g., *Suppe and Medwedeff*, 1990] can be applied to the San Ramón structure. So if the geometry and the timing of the deformation are sufficiently constrained, then deformation rates can be derived.

[19] The simplified structure depicted to 4 km depth in the San Ramón section (in Figure 3b, below, and in the box with plain colors in Figure 3c) is well reconstructed from the direct observations of the surface geology reported in our structural map (Figure 3b). Uncertainties remain because not all the layers in the volcanic sequences can be followed continuously, but we are confident that the overall geometry constrained by the elements of our map is accurate, and in any case, correct enough for the first-order estimates that we make in section 3.3. This superficial part of the section spans mostly continental deposits of Paleogene-Neogene age, with the Farellones Formation on top of the Abanico Formation [*Thiele*, 1980; *Vergara et al.*, 1988, *Nyström et al.*, 2003]. The Abanico and Farellones formations are regionally mapped one over the other for more than ~300 km along the strike of the Andes (for thorough descriptions and discussions of these two

formations, see *Charrier et al.* [2002, 2005]). They represent the uppermost units deposited in the Andean Basin syncline. Folding in the Abanico Formation is significant and it decreases gradually upward into the Farellones Formation, so the contact between these two units is described as progressive, with no clear time hiatus and no development of a regional unconformity [*Godoy et al.*, 1999; *Charrier et al.*, 2002, 2005]. The elusive definition of this contact has created confusion and inconsistencies between the different published sections and maps [e.g., *Kay et al.*, 2005]. However, pronounced angular unconformities with the overlying Farellones Formation are visible where localized deformation and erosion in the Abanico Formation is more intense, as it can be appreciated in some spots along the San Ramón section (Figure 3b, bottom). Folding appears to have continued throughout and after deposition of the Farellones Formation, so this unit is syntectonic. As a result the two formations (Abanico and Farellones) form a progressively folded asymmetric synclinorium ~30 km wide at the center of the Andean Basin (Figure 3c).

[20] The Abanico Formation consists of volcaniclastic rocks, tuffs, basic lavas, ignimbrites and interbedded alluvial, fluvial and lacustrine sediments [*Charrier et al.*, 2002, 2005, and references therein], with a minimum exposed thickness of ~3 km in the western flank of Cerro San Ramón (Figure 3b). The maximum age range compiled regionally for the Abanico Formation is from 36 Ma to 16 Ma, indicating a late Eocene–late early Miocene age [*Charrier et al.*, 2002]. More precisely, the K/Ar and ^{40}Ar/^{39}Ar dates in the huge stratified pile of volcanic rocks of the Abanico Formation close to Santiago range from 30.9 to 20.3 Ma, and are intruded by stocks, porphyry dikes and volcanic necks as young as 16.7 Ma [*Gana et al.*, 1999; *Nyström et al.*, 2003; *Vergara et al.*, 2004]. The more prominent pluton named La Obra leucogranodiorite intruding the San Ramón massif (Figure 3c) has a ^{40}Ar/^{39}Ar biotite age of 19.6 ± 0.5 Ma [*Kurtz et al.*, 1997].

[21] The Farellones Formation as defined at its type section east of Santiago consists of a thick series of intermediate and basic lava flows with volcaniclastic rocks and minor ignimbritic flows [*Beccar et al.*, 1986; *Vergara et al.*, 1988]. Its thickness there is variable, of 1–2 km (Figure 3b). The published K/Ar and ^{40}Ar/^{39}Ar dates as well as U-Pb zircon analyses in the Farellones Formation east of Santiago range from 21.6 to 16.6 Ma [*Beccar et al.*, 1986; *Nyström et al.*, 2003; *Deckart et al.*, 2005]. Elsewhere the Farellones Formation may exceed thicknesses of 2 km and span ages from middle to late Miocene [*Charrier et al.*, 2002]. The age of the Farellones Formation is also variable at the regional scale. Unconformable volcanic rocks as old as 25.2 Ma are

Figure 4. Map, satellite SPOT image and sections describing the San Ramón Fault and its piedmont scarp in the eastern districts of Santiago. Map and SPOT images covering the same area (shown in Figures 2b and 3a). Sections tentatively interpreted across the fault (labeled A and B) are located in the map. The San Ramón Fault trace is at the foot of a continuous scarp east of which the piedmont is uplifted and incised by streams. The more incised Cerros Calán, Apoquindo, and Los Rulos (to the north) expose an anticline made of early Quaternary sediments, cored by bedrock of the Abanico Formation and possibly cut by subsidiary thrusts, as those better exposed and mapped in Cerro Los Rulos (illustrated in section A). The gently sloping piedmont that is uplifted in the central part of the segment (section B) is covered with middle-late Pleistocene alluvium containing lenses of volcanic ash correlated with Pudahuel ignimbrites (see text). The map has been compiled and georeferenced at 1:25,000 scale, from original mapping on a DEM at 1:5,000 scale (shown in Figure 6).
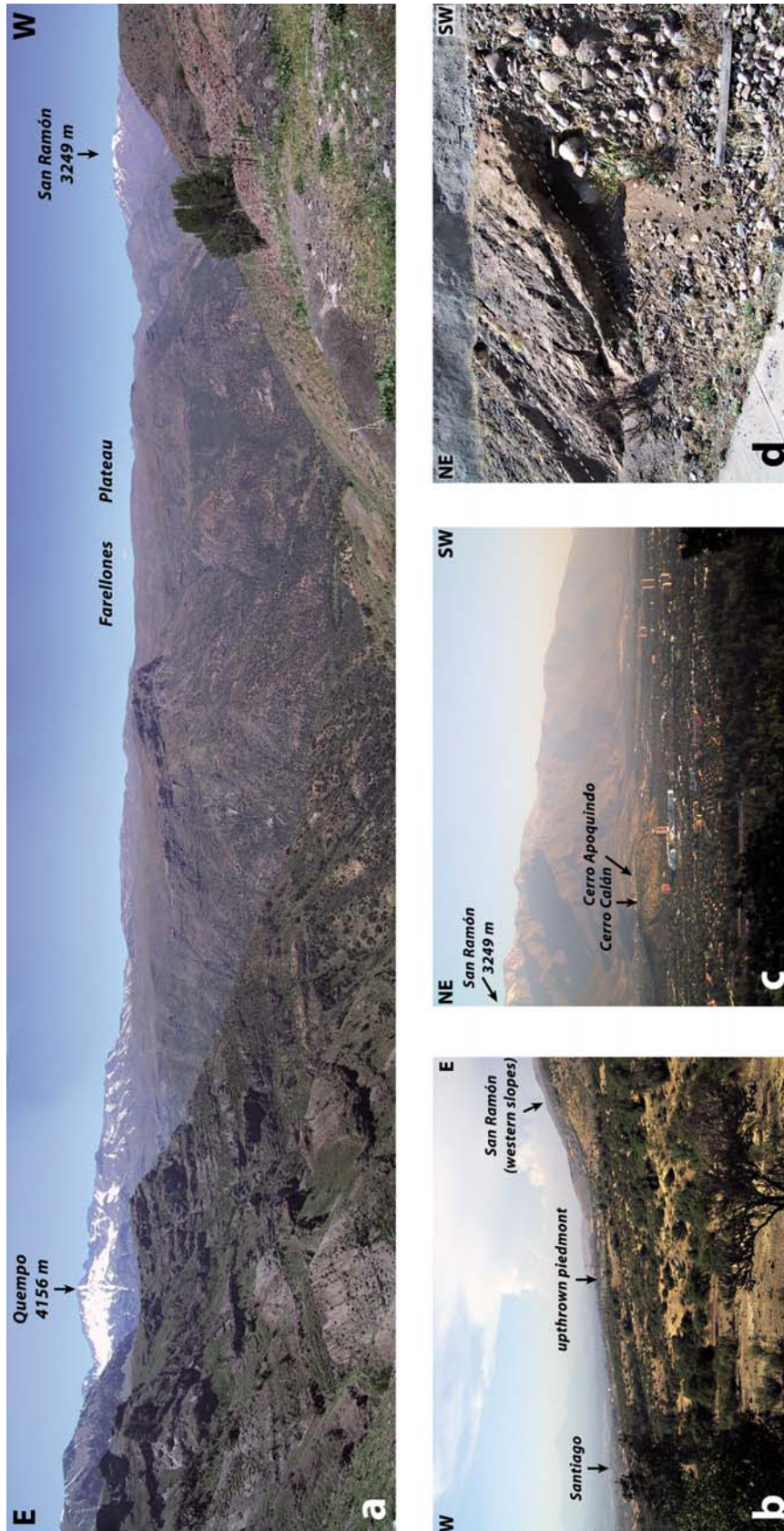
Figure 5. Field photographs (area of San Ramón and Farellones). (a) Panorama of Farellones Plateau, view to south from the north side of Río Molina canyon (see Figure 3a). (b) San Ramón piedmont scarp, view to north. Piedmont step with Pleistocene alluvium is upthrown by San Ramón fault (to its left) and dominates Santiago (see Figures 4 and 6 for location of piedmont scarp). (c) San Ramón piedmont scarp, view to southeast. Cerro Calán and Cerro Apoquindo are part of an anticline made of folded alluvial sediments of Río Mapocho (of early Quaternary or older age). (d) Outcrop of NE tilted alluvial sands and gravels on the NE limb of the anticline at Cerro Apoquindo (see section A in Figure 4), view to southeast.
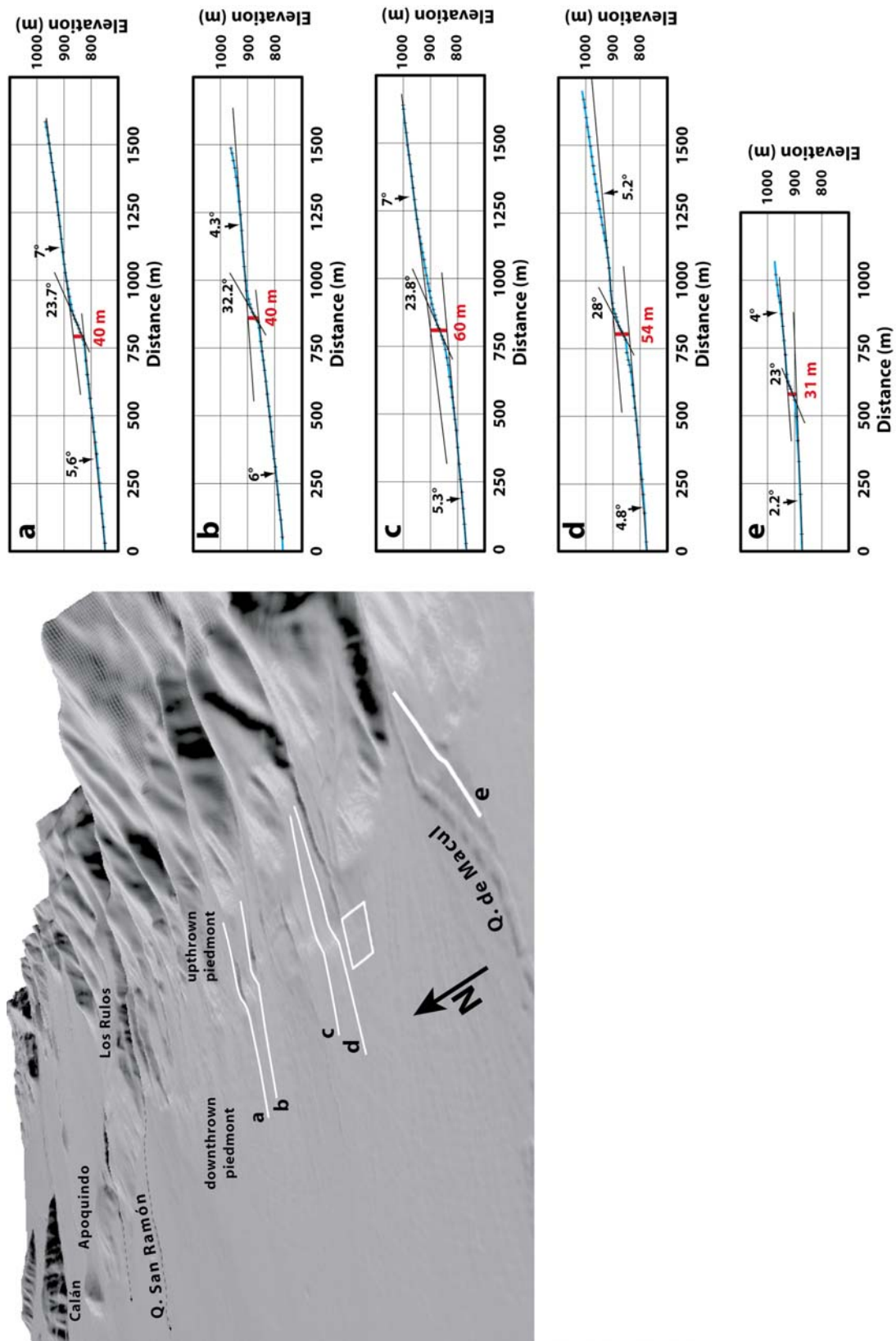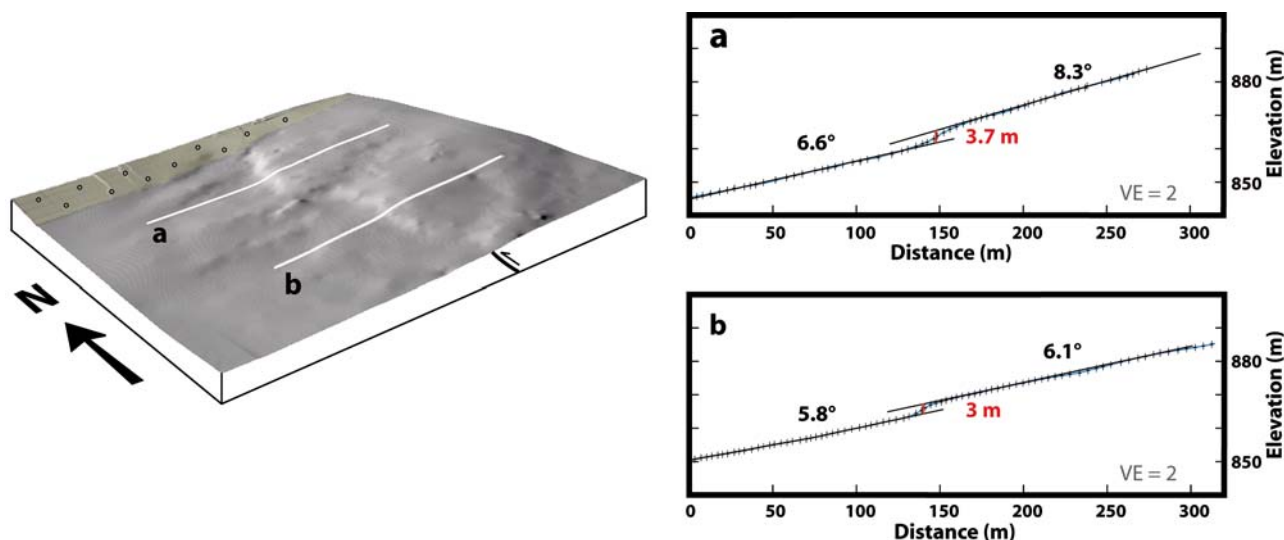
Figure 6

Figure 7. Morphology of most recent scarp across young (Holocene?) alluvial fan. (left) A DEM with 2 m horizontal resolution; 0.1 m vertical precision, based on a differential DGPS survey covering an area about 400 × 300 m^2 (location shown in Figures 4 and 6). The younger alluvium to the north (yellow) conceals the scarp. (right) Profiles a and b show a scarp 3–4 m high that may have resulted from a single event. Vertical exaggeration is 2. Profile symbols are as in Figure 6. The GPS data are not referenced, so the elevation base is within error of about ±10 m.

attributed to the Farellones Formation to the north of Santiago, at 32°S–33°S [*Munizaga and Vicente*, 1982]. The Farellones Formation is also correlated southward with rocks of the Teniente Volcanic Complex (at ~34°S latitude), which have K/Ar ages ranging from 14.4 to 6.5 Ma [*Kay et al.*, 2005]. The large uncertainties on the age of the Farellones Formation may stem from diachronism concomitant with progression of unconformities, suggesting a north-to-south propagation of the onset of shortening deformation [*Charrier et al.*, 2005].

[22] The section in Figures 3b and 3c suggests that the shortening deformation in the western Principal Cordillera has occurred after deposition of most, but perhaps not all, of the Abanico Formation. However, it could not have started after the deposition of the basal layers of the Farellones Formation. So conservatively the onset of the shortening deformation is probably in the late Oligocene to the early Miocene (~25–22 Ma) and strictly not later than 21.6 Ma, consistent with the age inferred by *Charrier et al.* [2002] for the onset of the regional shortening (interpreted by these authors as tectonic inversion) in the Abanico Formation. Besides, most of the incision of the Farellones Plateau by the Mapocho-Molina and Maipo-Colorado rivers has occurred after the Farellones Formation has been entirely deposited, so significantly later than the initiation of the shortening.

However, assigning a precise age for the inception of this incision from the published data is difficult, because of the stratigraphic uncertainty associated with the top of the Farellones Formation. Besides, apatite fission track ages documented the western Principal Cordillera provide no further constraint, because they fit well with depositional ages of the Abanico and Farellones formations [*Farías et al.*, 2008]. However, the young ages of lavas and plutons in the high-elevation mining districts of Los Bronces and El Teniente (porphyry copper deposits intruding the Farellones Formation) suggest that significant river incision has occurred after 5 Ma [*Farías et al.*, 2008]. So, conservatively, the incision of the Farellones Plateau could not have started earlier than ~16 Ma and it is still in progress, so occurring at a minimum long-term average rate of 0.125 mm/yr. Regardless of its rate, the young spectacular incision of the Farellones Plateau suggests that the shortening process associated with the West Andean Front has continued until the present.

3.3. Multikilometric Frontal Thrust: Deeper Structure, Kinematics, and Evolution

[23] The section in Figure 3c suggests that the San Ramón Fault has reached the surface with steep eastward dip and

Figure 6. Morphology of San Ramón piedmont scarp. Oblique NE view (3-D) of DEM shows upthrown and downthrown piedmont surfaces. White lines with lowercase letters mark location of profiles (labeled a to e, from north to south). Los Rulos-Apoquindo-Calán anticline is seen on the NW extension of upthrown piedmont. Profiles (in blue) across piedmont scarp indicate a minimum throw of 30–60 m (vertical red bars). Crosses in profiles correspond to pixels in the DEM; black lines with angles approximate average piedmont and scarp slopes. White rectangle in 3-D view locates area covered by a higher-resolution DEM where the most recent scarp is observed (Figure 7). DEM has 10 m horizontal resolution; 2.5 m vertical precision, based on aerial photogrammetric map at a scale of 1:5000 with elevation contours each 5 m.

producing a minimum throw of ~3.5 km (vertically measured from the highest peak in the mountain front to the lowest point in the bedrock bottom of the Santiago Basin). Besides, according to the reconstruction of the Abanico and Farellones formations in Figure 3c, the total structural thrust separation across the San Ramón Fault (measured on the ~55° dipping fault plane, red half arrow) would be of about 5 km (or ~4 km throw). Correspondingly, the minimum average slip rate would be of the order of a few tenths of mm/yr (0.2 mm/yr taking 5 km in 25 Myr). However, the growing fault propagation fold structure of the San Ramón massif implies that the fault has probably reached the surface much more recently than 25 Ma. To derive deformation rates in such a structure, the deeper geometry must be reasonably determined. The larger box with opaque colors in Figure 3c shows one possible simplified structure at depth that we deduce from regional maps [*Thiele*, 1980; *Gana et al.*, 1999; *Sellés and Gana*, 2001; *Fock*, 2005], the known stratigraphy of the Andean Basin [*Charrier et al.*, 2002; *Charrier et al.*, 2005; *Robinson et al.*, 2004, and references therein] and our own field observations. Our interpretation fulfils the most important geological constraints, within uncertainties, and is intended to draw up a set of first-order quantitative estimates, which are discussed below. It is clear that our observations and interpretation of the structure evolution as a fault propagation fold system imply a continuing shortening process across the West Andean Front, from ~25 Ma to the present, thus modifying drastically the concept of a regional shortening pulse (or a tectonic inversion) ending not later than ~16 Ma [e.g., *Charrier et al.*, 2002; *Farías et al.*, 2008]. However, the purpose here is to discuss the main first-order results, not the details of our structural reconstruction, or the formal modeling of the fault propagation folding and associated uncertainties (e.g., using trishear formalism [*Erslev*, 1991]), which will be presented elsewhere (R. Rauld et al., manuscript in preparation, 2010).

[24] The San Ramón–Farellones Plateau structure, which appears representative of the western Principal Cordillera (Figure 2b), is seen at the surface as a series of leading and trailing anticline-syncline pairs (for the fold and thrust vocabulary see *McClay* [1992]) growing bigger progressively westward, with a folding wavelength of ~8 km (Figures 3b and 3c). That structure requires a fault propagation fold mechanism with appropriate footwall flat-and-ramp geometry, to scale with the folding wavelength. The basal detachment underneath must have dip gently eastward to explain the steady uplift of the Farellones Plateau. A likely location for that detachment is close to the base of the Andean Basin, at ~12 km or more of stratigraphical depth, where the well-known, regionally widespread thick layers of ductile gypsum of Late Jurassic age should occur (Yeso Principal del Malm [*Thiele*, 1980]). These layers are elsewhere associated with significant diapirism and décollement of the Mesozoic-Cenozoic cover from the pre-Jurassic basement [e.g., *Thiele*, 1980; *Ramos et al.*, 1996b]. The anticline-syncline pairs developing bigger westward require a sequence of 3 or 4 steeply dipping ramps branching off upward from the basal detachment. Hence the tip line of each ramp appears to have propagated westward progressively closer to the surface, in agreement with the westward deformation gradient and with

the frontal San Ramón Fault ultimately reaching the surface. As a result of the deformation gradient, the Farellones Formation appears to have been deposited mostly in a piggyback basin configuration, contained between large thrust structures at its eastern and western sides (the Quempo and San Ramón thrust structures, respectively, Figure 3c). This configuration has probably preserved the Farellones Plateau located at the center of the piggyback basin from being deformed as much as rocks of same age on its sides. Under such circumstances, the Farellones piggyback basin may have been syndepositionally uplifted by hundreds of meters without being much eroded. It may have also been syndepositionally transported westward by motion on the basal detachment, probably by kilometers.

[25] The westward fault-propagating fold structure of the San Ramón–Farellones Plateau can be restored to deduce amounts of shortening and uplift across the western Principal Cordillera during the past ~26–22 Myr. The total horizontal E–W shortening across the Abanico Formation according to our schematic section (measured by restoring the idealized thick layer at the base of the Abanico Formation, between the two red triangles, see Figure 3c) is ~10 km, representing ~25% of the initial length. This includes 7–8 km of shortening due to folding and nearly 3 km of discontinuous shortening across the San Ramón Fault. The basal detachment goes from about 12 km to 10 km depth with 4.5° eastward dip beneath the Farellones Plateau. With a dip that shallow, the slip on the basal detachment associated with the total shortening of the Abanico Formation is also roughly 10 km.

[26] Assessing the net uplift of the Farellones Plateau from the structure is difficult, however, because several effects may contribute to the total apparent uplift of 2 km deduced from the incision by rivers. A basal slip of 10 km over the detachment ramp with 4.5° dip would contribute about 800 m uplift of rocks under the Farellones Plateau. The second contribution is penetrative shortening of the cover. Taking a reduced horizontal shortening of 5% (1/6 of the total average) in a rectangle 10 km thick and 20 km wide under the Farellones Plateau would produce 500 m of thickening of the cover above the décollement, which would contribute by a similar amount to the uplift. A negative contribution should be taken into account if some erosion of the top Abanico layers has occurred before deposition of the basal Farellones layers (~300 m?). The last contribution would be the whole thickness of the lavas that have piled up in the center of the Farellones piggyback basin (~1 km?). So despite the large uncertainties in any of the foregoing structural effects, the amount of uplift that can be deduced from summing them up does not appear inconsistent with the uplift deduced directly from river incision. However, the uplift of rocks under the Farellones Plateau due to those structural effects must have commenced much earlier than river incision of its top surface (after 16 Ma). So the minimum incision rate of 0.125 mm/yr (2 km of incision in 16 Myr) reveals a very weak constraint on uplift rates in the western Principal Cordillera.

[27] Clearly the shortening, uplift and erosion rates are inhomogeneously distributed across the Andes, because their respective intensity is causally connected [e.g., *Charrier et al.*, 2002]. Uplift and subsequent erosion are stronger where

shortening has been more intense, creating a structural high. For example, since deposition of the Farellones Formation the incision rate of the ríos Maipo and Mapocho across the intensely folded and faulted San Ramón frontal range would be of 0.25 mm/yr (4 km of incision in 16 Myr), twice as much as across the piggyback Farellones Plateau, where folding is less intense (Figure 3c). So the variation in the degree of erosion in the Principal Cordillera appears intimately correlated with the structure wavelength, which is relatively short. Under such conditions, no regional erosion surface (or peneplain) can develop. This observation casts a serious doubt on the validity of the approach used by *Farías et al.* [2008], who have identified a high elevated peneplain in this region, used to describe quantitatively the Andean uplift and morphologic evolution.

[28]   The minimum average shortening rate across the San Ramón–Farellones Plateau structure and the slip rate on its basal detachment since the inception of shortening are both of the order of 0.4 mm/yr (10 km in 25 Myr). In principle, the slip rate on the San Ramón Fault can be estimated for the time elapsed since its propagating tip line has crossed the base of the Abanico Formation, but no direct observation is available to constrain that time. Assuming the geometry in our section is valid (Figure 3c), the westernmost ramp of the propagating thrust system has formed after deposition of the Farellones Formation. Then the minimum long-term average slip rate on the San Ramón Fault would be of about 0.3 mm/yr (5 km in 16 Myr), or a throw rate of ~0.25 mm/yr (4 km in 16 Myr). This inference is consistent with most of the slip on the basal detachment being transferred since 16 Ma to the San Ramón Fault, making of it the frontal ramp of the Principal Cordillera.

3.4. Piedmont Scarp

[29]   The San Ramón mountain front was first interpreted as the expression of a normal fault and the sediments on its piedmont identified as mostly glacial in origin [*Brüggen*, 1950]. Recently compiled geological maps still miss the piedmont scarp and interpret the piedmont sediments as mostly derived from massive gravitational sliding [e.g., *Gana et al.*, 1999]. However, the geomorphology of the piedmont scarp was recognized long since [*Tricart et al.*, 1965; *Borde*, 1966], but as yet ignored by geologists. Here we make a synthetic quantitative description of the San Ramón piedmont scarp, building on previous work that to our knowledge is the first attempt to elucidate the faulting processes behind that scarp [*Rauld*, 2002; *Rauld et al.*, 2006; *Armijo et al.*, 2006]. A more detailed analysis will be presented elsewhere (R. Rauld et al., manuscript in preparation, 2010).

[30]   The best surface expression of the San Ramón Fault is found along the 14–15 km long segment with a sharp fault trace at elevation between 800 and 900 m, approximately between Cerro Calán and Quebrada Macul, so covering a large part of the 25 km separating ríos Mapocho and Maipo along the San Ramón mountain front (Figures 3a, 4, and 6). The quality of the exposure stems from the occurrence of a ~3 km wide piedmont (roughly between 700 and 1000 m elevation) formed by rough stratified alluvium and colluvium, which is clearly cut by the trace of the fault. The exposure of

structural features in that rough and little dissected material is not only poor and scarce, but also heavily hampered by urbanization. So most of the information described here derives from the morphology, which is well constrained by the accurate digital topography and imagery. The piedmont has generally regular slopes, reaching a maximum of ~7° by the mountain front and decaying gradually downward, away from it (Figure 6). Those relatively gentle slopes are sharply cut and offset by the scarp, creating an upthrown piedmont balcony that overlooks Santiago (Figure 5b; the center of Santiago is at ~550 m elevation).

[31]   The southern part of the scarp has a N5°W strike on the average, but approximately from Quebrada San Ramón northward, toward the Río Mapocho, it turns into a N25°W strike (Figure 4). That northern part of the fault scarp is characterized by a string of three arch-shaped hills (Los Rulos, Apoquindo and Calán hills; Figures 4, 5c, and 6) overhanging by 100–300 m the rest of the upthrown piedmont. So sediments cropping out in those hills appear stratigraphically older than those in the gently sloping piedmont. The hills correspond to eroded remnants of a gentle NW striking anticline structure deforming the Quaternary sediments, particularly the alluvium deposited by Río Mapocho. Layers of fluvial sands and gravels are tilted northeastward (so toward the mountain side, opposite to the drainage direction of the Río Mapocho) reaching dips up to 30° along the northeast limb of the anticline (Figure 5d). Development of stepped terraces in the valleys that cross the fold structure (like in Quebrada Apoquindo; see map in Figure 4), suggests that those terraces have formed during alternating periods of erosion and aggradation, which have occurred syntectonically across the forming anticline. Subsidiary reverse faulting is observed in the anticline (Figure 4, map and section A). However, the folding of the 5 km long, 1.5 km wide anticline appears to involve folding at the same scale of the underlying bedrock (Abanico Formation), which forms the core of Los Rulos and Apoquindo hills (Figure 4, section A). So the presence of the Los Rulos-Apoquindo-Calán bedrock anticline may be indicative of some near-surface complexity in the process of fault propagation in this area, as is tentatively illustrated in section A (Figure 4).

[32]   Between the prominent Quebradas San Ramón and Macul is the younger, most regular part of the piedmont, where only minor streams traverse its surface (Figure 4). Unlike the arched northern part of the piedmont fault scarp, the gently inclined piedmont here expresses no surface folding, suggesting that no significant near-surface complexity of the fault plane occurs in the bedrock behind (Figure 4, map and section B). However, the piedmont sediments cover an erosion surface at the foot of the mountain front, so if an earlier bedrock structural complexity had occurred, it has been erased by that erosion. The piedmont surface is made of a series of contiguous alluvial fans forming a bajada. Modern streams have caused fan head entrenchment across the bajada on the upthrown block. The lower end of entrenchment reveals clearly the trace of the piedmont fault, because the modern streams incising the bajada grade to the top surface of the downthrown block, west of the fault scarp, where the modern alluvial fans are being deposited (Figure 4). Between the streams on the upthrown block, the top surface of the

bajada is abandoned and well preserved from surface erosion. However, the continuation of that top surface on the down-thrown block is partly covered by the modern alluvium. The modern deposition of alluvium by recent fans on top of that older piedmont surface appears modest, within mapping uncertainties of the modern fans on top of the older piedmont surface. May be not more than ~20 m sediment thickness have been accumulated by the small alluvial fans fed by the small streams in this part of the piedmont, while ~100 m thickness of recent alluvium may have been accumulated by the fault scarp by the larger fans in front of Quebradas San Ramón and Macul.

[33] The morphology of the San Ramón piedmont bajada is well determined by the DEM (Figure 6). Profiles across this topography provide a precise measure of the piedmont fault scarp (profiles labeled a to e). Then, the apparent component of vertical slip (throw) derived from the topography varies between a minimum of 30 m and a maximum of 60 m within an uncertainty of ~10%. Strictly, these are minimum estimates for the piedmont offset. However, because erosion of the piedmont surface on the upthrown block is negligible and the thickness of modern deposition of alluvium in this part of the downthrown block appears modest, then retaining a minimum average throw of 60 m as an estimate of the piedmont offset appears reasonable.

[34] To date the rough heterogeneous material of the offset piedmont is difficult. The alluvium of the piedmont is formed of a poorly layered and poorly sorted sequence of boulders and angular pebbles embedded in a matrix composed of silts and clays, locally including layered gravels, sand lenses of fluvial origin and conspicuous lenses of volcanic ash. Hence a significant part of the piedmont sediment appears to have been deposited by debris flows and mudflows. A modern example is the catastrophic mudflow of 1993, triggered by a flash flood rain in the nearby slopes of the San Ramón massif, which contributed with up to ~5 m thickness of new sediment over an area of ~3–4 km^2 on the large fans at the exit of Quebradas San Ramón and Macul (Figure 4, map).

[35] However, the frequent occurrence of ash lenses exposed in the upthrown block of the San Ramón piedmont may provide us with an accurate stratigraphical mark [Brüggen, 1950; Tricart et al., 1965; Rauld, 2002] (Figure 4, map). The ash lenses of the San Ramón piedmont can be correlated with the pumice deposits called Pudahuel ignimbrites, found extensively in the Santiago valley [Gana et al., 1999] and with lithologically similar rhyolitic pyroclastic flow deposits that occur more discretely on terraces of several rivers and are distributed at a broad regional scale, on both the east and the west flanks of the Andes [Stern et al., 1984]. Stern et al. [1984] dated those pyroclastic flows at 450 ka ± 60 ka (with zircon fission tracks) and suggested that their deposition may have followed large eruptions (volume erupted estimated as ~450 km^3) associated with the collapse of the noticeable Maipo volcano caldera. If the correlation of ash lenses embedded in the San Ramón piedmont with the Pudahuel ignimbrites and their inferred ages are correct, then the top surface of the San Ramón bajada is younger than 450 ka and a minimum throw rate of ≥0.13 mm/yr (≥60 m in ≤450 kyr) can be deduced for the San Ramón Fault.

This represents about half the minimum throw rate deduced over the long term. Conversely, taking the long-term estimate of average slip rate on the San Ramón basal detachment (0.4 mm/yr), the abandoned bajada surface of the uplifted piedmont would have an age of 150 ka, consistent with the inferred age of the ashes. Alluvial sediments and fluvial terraces in the Maipo and Mapocho river valleys can be unambiguously correlated with the uplifted San Ramón piedmont. Unpublished age determinations of these deposits (Ar-Ar ages from pumice rhyolitic pyroclastic deposits and optically stimulated luminescence (OSL) ages of alluvial sediments (G. Vargas et al., manuscript in preparation, 2010)) suggest that the younger age estimate (~150 ka) consistent with the long-term slip rate of the San Ramón detachment is close to the age of the bajada abandonment.

3.5. Scarp Corresponding to the Last Event (s) and the Seismic Hazard

[36] To find well-preserved small-scale scarps is now difficult in the densely urbanized outskirts of Santiago. Hereafter we describe what appears to be the last testimony to late scarp increments left for study along the San Ramón Fault.

[37] The 15 km long piedmont fault segment discussed in section 3.4 has a simple trace and is well preserved over most of its length, so it crosses the stream drainage close to the apexes of the modern alluvial fans (Figure 4). In those places the stream power is high and no scarp increment has apparently survived to persisting erosion in the stream channel and to rapid knickpoint headward retreat. The trace of the piedmont fault is more complex near its two ends, near Cerro Apoquindo and near Quebrada de Macul. There significant fault branches cross the modern alluvial fan surfaces, offering opportunities for preservation of young scarp increments. Unfortunately, the young scarps crossing the small modern alluvial fans to the west and south of Cerro Apoquindo (which are readily detected in the high-resolution DEM and in old aerial photographs; see location of those scarps in Figure 4) are now out of reach for study because of the rapid urbanization of the city.

[38] To the south of the piedmont scarp one scarp is still preserved, which we have been able to describe in the field. Two branches about 300 m apart make echelons in the morphology over a length of about 3–4 km near Quebrada de Macul (see Figure 4). The westernmost fault branch is only 1 km long and it splays northward and southward, entering into the downthrown piedmont. There it crosses the small modern alluvial fans being deposited in front of the main piedmont scarp, which at those places follows the fault branch located eastward. A clear fault scarp can be followed across one of those small fans, for no more than ~300 m, where the preservation conditions appear to have been exceptionally favorable (Figures 4 and 7). South of this fan along the fault branch no scarp is visible across the large fan at the exit of Quebrada de Macul. This absence may be due to the relatively rapid modern accumulation of alluvium on this fan during repeated catastrophic mudflow events like that in 1993, which can conceal any young fault scarp [Naranjo and Varela, 1996; Sepúlveda et al., 2006]. The

preservation of a small scarp across the smaller alluvial fan in Figure 7 can be explained by a recent abandonment of this fan associated with the northward diversion of the stream feeding deposition of alluvium to the small fan just north of it, where the continuation of the same scarp appears to have been concealed (Figures 4 and 7).

[39] To determine the offset produced across the topography of the small fan by the fault, a DGPS survey has been conducted over a limited area (Figure 7). The fan has steep slopes decaying westward from 8° to 6° and the scarp across it suggests an apparent throw that decays southward along strike from ~3.7 m to ~3 m (Figure 7). The sharp simple morphology of that scarp suggests it may have resulted from a single seismic event, although the possibility of multiple events cannot be excluded. Considering that the fault near the surface may have steep 50°–60° eastward dip (Figure 3c), the net thrust slip corresponding to the total measured throw would be of about 4 m, to be accounted for by a single event or by several events with thrust slip of the order of ~1 m or less. Taking an average thrust slip ranging between 1 and 4 m over the 15 km long piedmont fault segment with rupture width of 15 km (corresponding to the frontal ramp in Figure 3c, breaking from 10 km depth to the surface) would yield seismic moments of Mo ~0.75 to 3×10^{19} N m, corresponding to events of magnitude Mw 6.6 to Mw 7.0. This range of magnitudes is higher than that of the sequence of three consecutive shocks, all together known as the 1958, Las Melosas earthquake, which correspond to the largest events recorded instrumentally in the upper plate near Santiago [see *Sepúlveda et al.*, 2008; *Alvarado et al.*, 2009, and references therein]. The 1958 sequence occurred within an interval of 6 min with hypocentral depth of 10 km in the center of the Principal Cordillera ~60 km SE of Santiago, with intensity values reaching IX in the epicentral area. The larger first shock has been assigned a revised magnitude Mw 6.3 [*Alvarado et al.*, 2009].

[40] However, the estimate above is not conservative because the seismicity recorded under the Principal Cordillera shows well-constrained hypocenters down to 15 km and more (Figure 8c). This suggests that the basal detachment of the San Ramón–Farellones Plateau structure could contribute to the seismic release, thus increasing significantly the width of a potential fault rupture of the San Ramón Fault. Given the shallow dip of the detachment, a rupture confined to depths of less than 15 km could reach widths in the range of 30–40 km. Taking as before the same range of average slip of 1–4 m but considering a rupture extending over a width of 30 km and over the entire length of ~30 km of the San Ramón mountain front facing the Santiago valley would yield seismic moments of Mo ~0.3 to 1.2×10^{20} N m, corresponding to events of magnitude Mw 6.9 to Mw 7.4.

[41] Earthquakes with large magnitudes would not be frequent because the loading rate of the San Ramón Fault appears low. If the present-day slip rate is assumed to be as slow as the average estimate for the basal detachment over the long-term (0.4 mm/yr) then the time to recharge events with slip of 1–4 m is 2500–10,000 years. The probability of having recorded historically such an event is low, given the

Figure 8. Simplified section across the Nazca–South America plate boundary and the Andes at the latitude of Santiago (33.5°S, see Figures 1 and 2a for location; main Andean geological features correspond to those mapped in Figure 2b). (a) The hypothesized west Andean megathrust (depicted as an embryonic intracontinental subduction under the Andes) reaches the surface at the San Ramón Fault (West Andean Front), parallel and synthetic to the subduction interface with the Nazca Plate. The two synthetic systems are separated by the rigid Marginal Block, down-flexed eastward as is underthrust beneath the Andes. The eastern side of the Andean Basin (>12 km thick Mesozoic-Cenozoic "back-arc" basin filled with sedimentary and volcanic rocks, also called "Andean Geosyncline" [*Aubouin et al.*, 1973]) is inverted and deformed in the Principal Cordillera as a prowedge, pushed westward by the Frontal Cordillera basement backstop (leading edge of South America). Once reduced to scale, the Aconcagua Fold-Thrust Belt (AFTB) appears as a shallow minor back thrust. The Marginal Block appears a rigid board balancing coastal uplift (associated with subduction processes) and orogenic load by the Andes (circled yellow arrows), between the two megathrusts. The Cuyo Basin in the eastern foreland is mildly deformed as an incipient retrowedge. Total Andean shortening across the section is not less than about 35 km, not more than about 50 km (30–40 km across the Principal Cordillera prowedge and 5–10 km across the Cuyo Basin retrowedge). Shortening at lithospheric scale (measured by Moho offset) is drawn consistent with shortening in the upper crust. Continental lithosphere (brown, crust; green, mantle) and oceanic lithosphere (light blue, crust; dark blue, mantle) are schematized. Base of lithosphere is highly simplified and does not take into account (although possibly important) lithosphere thickness variations under the Andes, which are not relevant for the primary purpose retained for this paper. Faults are in red, dashed where very uncertain. Coupled circled cross and dot (in black) indicate likely locations of strike slip. Basins with sedimentary/volcanic fill (adorned with layers) are shown in yellow (Cenozoic) and green (Mesozoic). Positions of present-day volcanic arc and possible feeding across lithosphere are indicated. Gray rectangle locates details shown in Figure 3c. No vertical exaggeration. AW, accretionary wedge; FP, Farellones Plateau; WVF, west verging folds; T, Tupungato; AT, Alto Tunuyán basin. (b) Outline section with bulldozer representing appropriate boundary conditions. The West Andean Thrust (WAT) is in bold red. (c) Outline section testing for consistency of model with seismicity and rough crustal thickness. The red circles represent the complete record of the 2000–2005 projected seismicity, including the best located hypocenters (events with Ml > 4.0 and hypocentral location RMS < 0.3) by Servicio Sismológico Nacional of Universidad de Chile (seismicity section is half degree wide, centered at 33.5°S). Projected smooth Moho profile, roughly interpolated from broadband seismological data (bold dashed white line from *Gilbert et al.* [2006]), is represented for comparison. The better resolved images obtained with receiver functions by *Gilbert et al.* [2006] (to the north and south of our section, so not represented but discussed in the text) show structural complexity consistent with the stepped Moho structure proposed in our model.
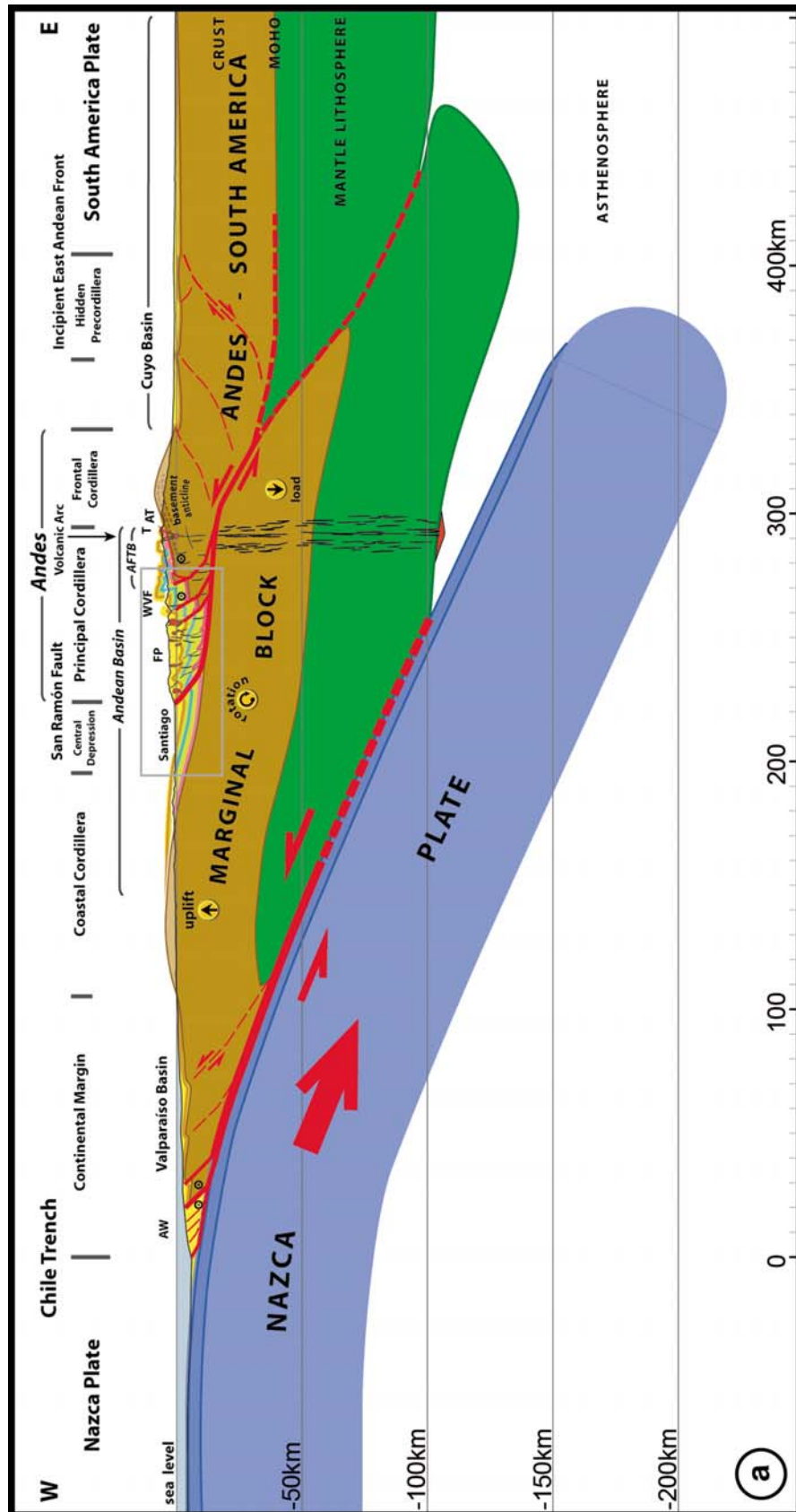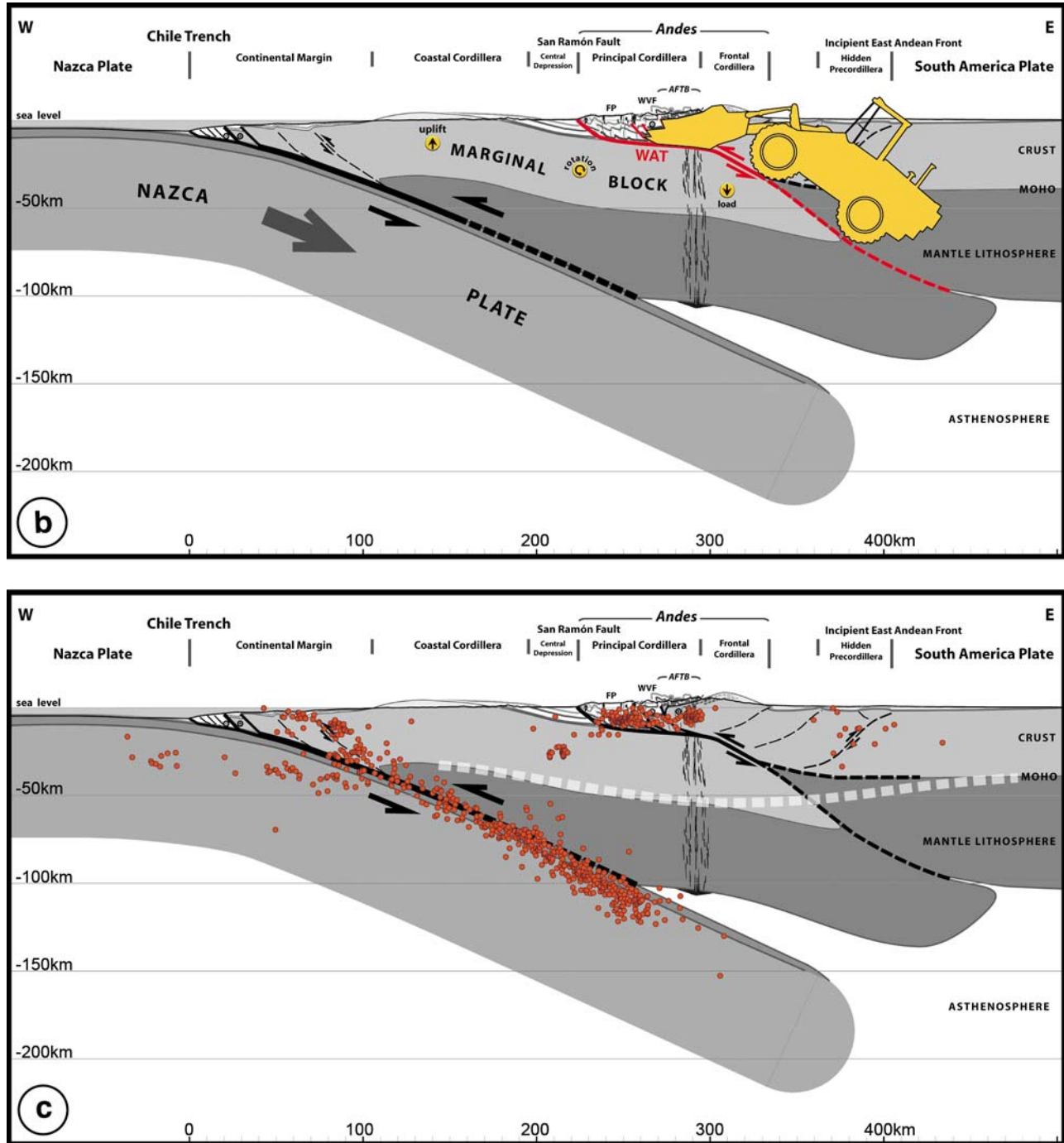
Figure 8

Figure 8. (continued)

short period of settlement in the Santiago valley (the city was founded by the Spanish in 1541). However, the earthquake that destroyed most of the city in May 13, 1647 was probably not a subduction interface event [*Barrientos*, 2007]. The historical accounts suggest that the destructive source could have been either a slab-pull event in the subducting plate at intermediate depth (~100 km depth, similar to the 2005 Tarapacá earthquake [*Peyrat et al.*, 2006]) or a shallow

intraplate event close to Santiago in the nearby Andes [*Lomnitz*, 2004]. Only a paleoseismological study of the San Ramón Fault could prove or disprove the second hypothesis.

4. Discussion

[42] The foregoing description of the San Ramón thrust front has significant consequences on our understanding of

the large-scale Andean architecture, thus implying substantial changes from the currently accepted interpretations. It is generally accepted that shortening in the Principal Cordillera is due to a process of inversion of the Andean Basin associated with dominant large-scale east vergent thrusts extending downward to the west, specifically beneath the Andean Basin [*Ramos*, 1988; *Mpodozis and Ramos*, 1989; *Ramos et al.*, 1996b; *Godoy et al.*, 1999; *Cristallini and Ramos*, 2000; *Charrier et al.*, 2002, 2005; *Fock*, 2005; *Farías et al.*, 2008; *Giambiagi et al.*, 2003; *Ramos et al.*, 2004; *Kay et al.*, 2005]. The ultimate consequence of those interpretations is that the east vergent thrust system would step deeper and deeper down westward, beneath the Central Depression and the Coastal Cordillera, to meet the subduction interface [*Farías*, 2007] (for a similar interpretation for northern Chile, see also *Farías et al.* [2005]). Our results suggest that those large-scale geometries are mechanically inconsistent with the observed present-day architecture of the Andes at the latitude of Santiago, which clearly indicates a primary westward vergence.

[43] The following discussion is intended to explain briefly the first-order arguments, constraints and uncertainties underlying our new interpretation of the Andean tectonics at this latitude. The most relevant structural elements that we use for this discussion are represented in map view (Figures 2b and 3b) and in section (Figures 3c and 8a). A more complete development discussing details of the Andean evolution in space (comparison with other tectonic sections located northward and southward from the one discussed here) and time (propagation of deformation) is out of the scope of this paper and will be provided elsewhere (R. Lacassin et al., manuscript in preparation, 2010).

4.1. Crustal Structure Asymmetry and the Scale of the Mechanical Problem

[44] The observations of wholesale uplift, shortening and décollement of the thick Andean Basin deposits throughout the Principal Cordillera imply an overall Andean fold-thrust belt ~80 km wide, which requires appropriate structure geometry at crustal-scale depth, associated with reasonable kinematics and boundary conditions. The source of mechanical energy supplied to that large Andean fold-thrust belt needs to be explained. In other words, a rigid buttress (or bulldozer) must be found at the appropriate scale, in the inner thicker part of the orogen [*Davis et al.*, 1983; *Dahlen*, 1990]. Such boundary conditions appear sound and generally accepted for orogenic wedges like the Alps and the Himalayas [e.g., *Bonnet et al.*, 2007; *Bollinger et al.*, 2006]. Clearly, there is no such a basement buttress west of the Principal Cordillera in the region of the Central Depression and the Santiago Valley (or more westward, in the Coastal Cordillera or Continental Margin), capable of pushing eastward a Coulomb wedge made of Andean Basin deposits, as suggested earlier [e.g., *Giambiagi*, 2003; *Giambiagi et al.*, 2003; *Farías*, 2007]. Similar mechanisms, which we find improbable, are required by all tectonic models of the Principal Cordillera based on large-scale eastward vergence [*Ramos*, 1988; *Mpodozis and Ramos*, 1989; *Ramos et al.*, 1996b; *Godoy et al.*, 1999; *Cristallini and Ramos*, 2000;

Charrier et al., 2002, 2005; *Fock*, 2005; *Farías*, 2007; *Farías et al.*, 2008; *Giambiagi et al.*, 2003; *Ramos et al.*, 2004; *Kay et al.*, 2005].

[45] Besides, the propagating West Andean Front as inferred from the observed structure of the San Ramón–Farellones Plateau must be rooted in downward to the east, beneath the higher Principal Cordillera. Thus, the basal detachment under the Farellones Plateau and the thick Mesozoic-Cenozoic cover (shown in Figure 3c) must ultimately step down into the Andean hinterland basement, i.e., penetrating deeply into the basement of the Frontal Cordillera and into the Andean crust. That large-scale west vergent thrust system (designated hereafter as the West Andean Thrust, or WAT), which would be the main deep-seated feature responsible of the present-day architecture of the Andean fold-thrust belt, may have a complicated staircase trajectory consisting of multiple flats and ramps. As first-order approximation, however, the simple geometry illustrated in Figure 8a is consistent with the main structural features of the surface geology (Figures 2b and 3b), as discussed gradually from one stage to the next below. We also discuss the apparent consistency of that geometry with the available geophysical data. Clearly, however, the details of the present architecture of the WAT need to be refined and constrained further with the rapidly growing set of geophysical data (e.g., gravity, GPS and seismological data). It has been shown that the well-established chronostratigraphical constraints available for the region [e.g., *Charrier et al.*, 2002; *Giambiagi et al.*, 2003 and references therein] fit well tectonic interpretations very different from the one presented here [e.g., *Giambiagi et al.*, 2003; *Farías et al.*, 2008]. Alternatively, however, these data can also be used to constrain the possible evolution of main structures taking as final stage the present-day architecture proposed in Figure 8a. We discuss (in section 4.8) some of the alternative interpretations, along with the chronology and shortening estimates.

4.2. Frontal Cordillera Bulldozing Westward the Whole Andean Fold-Thrust Belt

[46] The Frontal Cordillera is the only Andean basement high to scale with the mechanical function of a rigid buttress that backstops the shortening and the high elevation of the Principal Cordillera. This function is represented in Figure 8b by a bulldozer. The Frontal Cordillera forms a huge basement anticline (30–50 km wide) that elongates probably more than ~700 km along the grain of the Andes (at least between 28°S and 34.5°S [e.g., *Mpodozis and Ramos*, 1989]) and is located side by side, east of the similarly long Andean fold-thrust belt represented by the Principal Cordillera (Figure 2b). The anticline shape of the Frontal Cordillera is outlined in section (Figure 8a) by the unconformable basal contact of the Choiyoi Group rocks (Permian-Triassic) over the Paleozoic Gondwanan basement, which includes Proterozoic metamorphic rocks [*Polanski*, 1964, 1972; *Ragona et al.*, 1995; *Heredia et al.*, 2002; *Giambiagi et al.*, 2003]. The Frontal Cordillera appears thus as a crustal-scale ramp anticline providing the necessary boundary conditions to maintain the high elevation in the Principal Cordillera and to produce the

westward propagation the San Ramón thrust system. Similar fold-thrust structures reaching the surface northward or southward from the San Ramón system along the West Andean Front have probably the same origin and relation with the Frontal Cordillera. Therefore, the main crustal-scale ramp under the Frontal Cordillera anticline must dip necessarily eastward and the overall Andean structure at the latitude of Santiago has a decided westward vergence. Albeit much less pronounced, the localized westward dipping thrust features described to the east of the Frontal Cordillera (Figures 2b and 8a) must also be of crustal scale [Ramos et al., 1996b] and are discussed in section section 4.5.

4.3. Aconcagua Fold-Thrust Belt: A Secondary Feature

[47] The Aconcagua Fold-Thrust Belt (which is conventionally limited to the eastern ~30 km wide part of the Principal Cordillera, where Mesozoic sediments crop out) appears to be a shallow secondary feature, structurally overlying the Frontal Cordillera basement, thus located well above the main ramp of the WAT beneath the Frontal Cordillera anticline. As depicted in Figures 2b and 8a, the AFTB is a shallow back thrust décollement (at ~2–3 km depth below the high Cordillera surface) that detaches most of the Mesozoic platform sediments deposited at the eastern margin of the Andean Basin from its basement (specifically the Frontal Cordillera; e.g., see the classical section at 33°S by V. A. Ramos and collaborators, reproduced by Ramos et al. [2004, Figure 8]). The décollement appears to be localized at the Jurassic gypsum layers [e.g., Ramos et al., 1996b] and is associated with significant diapirism and kilometer-scale disharmonic folding [Thiele, 1980]. In fact the AFTB displays both, an eastward vergence on its eastern side and a westward vergence on its western side. To the east, the shallow AFTB detachment ramps up to the surface, where a back thrust front associated with relatively modest displacement (generally less than about 3 km) generally separates the detached layers of the Mesozoic cover from similar layers, stratigraphically at the very base of the same Andean platform cover, which have remained undetached from their basement [Polanski, 1964, 1972] (see Figure 2b). Those basal layers, rest with shallow westward dip (~20°–25°), on the western limb of the Frontal Cordillera anticline [Polanski, 1964, 1972; Ramos et al., 1996a]. That basal contact appears nearly conformable over the Choiyoi Group rocks (Permian and Triassic age), but is regionally unconformable over older rocks of the Gondwana basement. In some places along the eastern front of the AFTB, small intermontane basins (<10 km wide and <2 km thick) that are filled with continental conglomerates of late early Miocene (~18 Ma) to late Miocene age (so mostly coeval with the Farellones Formation) are spectacularly involved in the back thrust deformation (Santa María and Alto Tunuyán basins; see location of Alto Tunuyan basin (AT) in Figures 2b and 8a) [Ramos, 1988; Ramos et al., 1996b; Giambiagi et al., 2001, 2003]. However, it is clear that for most of its length the back thrust front of the AFTB is a shallow detachment mechanically supported almost directly by the basement structure, not by any sig-

nificant foreland basin of flexural origin. Thus the AFTB appears a relatively minor feature of the Andean tectonics, which is passively transported westward, atop the basement ramp anticline forming the Frontal Cordillera. Another second-order feature of the regional tectonics is the present-day volcanic arc, which appears to cross both, the basement and the AFTB, without any significant structural modification (Figures 2b and 8a).

4.4. Large West Verging Folds Ahead of the West Andean Basement Thrust Wedge

[48] The central Principal Cordillera, at the western side of the AFTB (between the AFTB and the Farellones Plateau) is characterized by a 20 km wide zone of strong deformation, consisting of a cascading sequence of two or three very large asymmetric west verging folds (Figures 2b and 3c and WVF in Figure 8a). Those features include an impressive ~5 km wide vertical limb (dip varying from steep westward to vertical, or even overturned locally to steep eastward dip) exposing a complete section of the Mesozoic sequence (top-to-the-west geometry). That large limb is outlined in the topography by the massif continental conglomerates, andesitic lavas and breccias constituting the ~3000 m thick Río Damas Formation (Kimmeridgian: Late Jurassic), which arises as an almost continuous structural ridge with the same geometry over more than ~250 km along strike (from 33°S to 35°S). Westward alongside of that continuous wall, the vertical beds of the calcareous Lo Valdés Formation (Neocomian: Late Jurassic–Early Cretaceous) are characterized by a weak penetrative cleavage that dips steeply eastward (~70°E, Figure 9), consistent with the dominant westward vergence. Farther west alongside of the Mesozoic sequence, there is a prominent syncline system with an overturned eastern limb, topped by rocks of the Abanico Formation (Coironal in Figure 3c), which can be easily followed for about 100 km along strike (roughly between 33°S and 34°S [González-Ferrán, 1963; Thiele, 1980; Baeza, 1999; Fock, 2005]). The steeply eastward dipping (or vertical) Chacayes-Yesillo fault disrupts the steep eastern limb of the syncline near 33.8°S, and similar faults are described extending tens of kilometers southward [Baeza, 1999; Fock, 2005; Charrier et al., 2005]. To the west of the Cerro Coironal syncline are found the Río Olivares anticline, then the Quempo ridge at the eastern edge of the Farellones Plateau, where the top of the Mesozoic sequence continues the cascade of west verging folds, plunging westward under the Cenozoic sequence (Abanico and Farellones formations, Figures 2b and 3c). Taken as a whole, that 20 km wide zone of west verging folds with large vertical limbs must be associated with a vertical structural separation comparable with its width. In our sections, the vertical separation of the base of the Abanico Formation across this zone (similarly as that of the other Mesozoic formations below), as deduced from the published stratigraphy and maps [Thiele, 1980; Fock, 2005], is at least of ~15 km (Figures 3c and 8a). The importance of vertical limbs in this zone of the Principal Cordillera is precisely what determines the crustal-scale westward inclination and overall wedged asymmetry of the
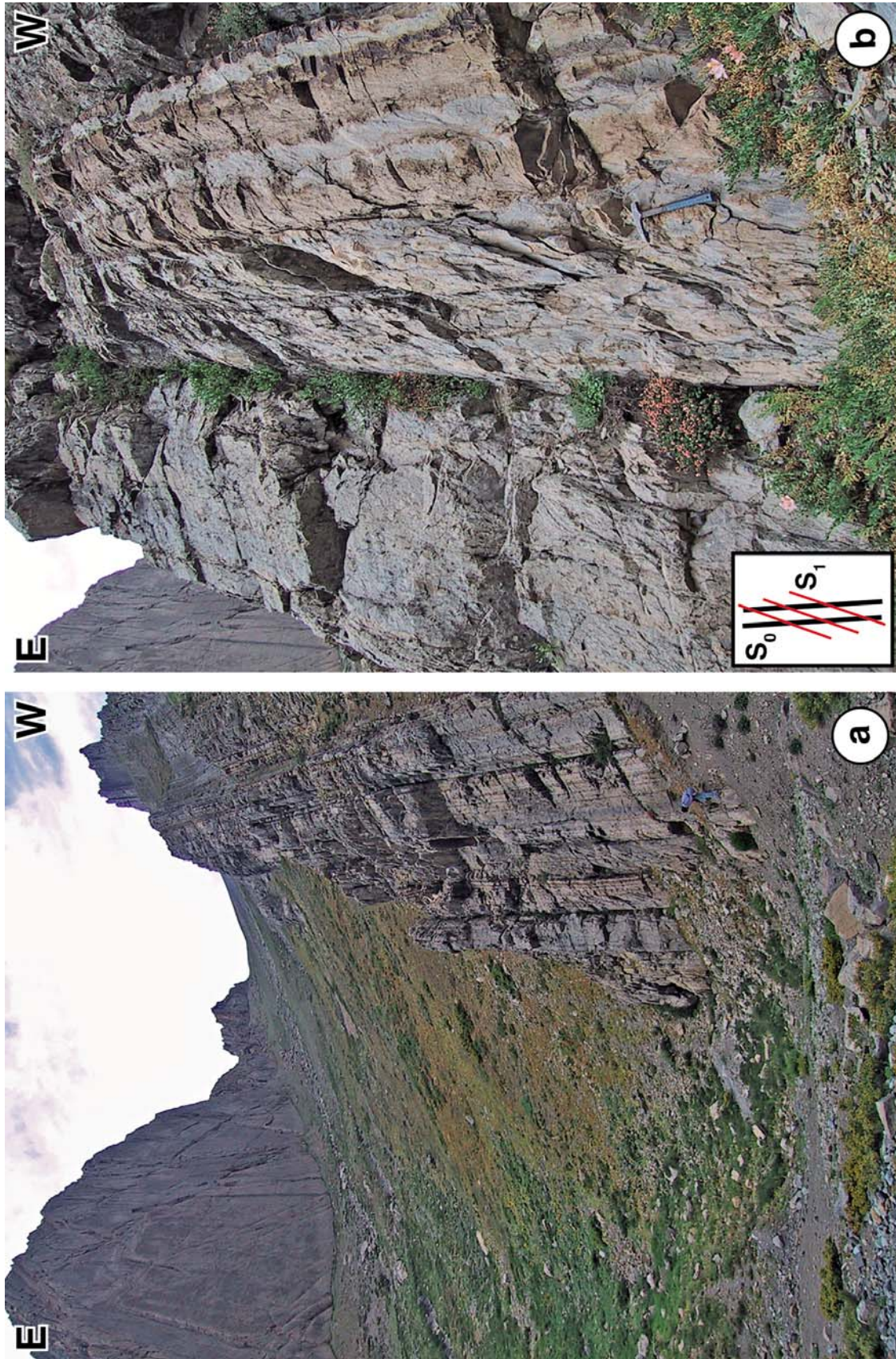
Figure 9

Andean Basin, as discussed in section 2 (profiles A and B in Figure 1).

[49] However, the main consequences of the foregoing concern the deeper structure. The prominent zone of west verging folds of the Andean cover in the middle of the Principal Cordillera, as well as its apparent continuity over hundreds of kilometers along strike strongly suggest a crustal-scale fault propagation fold structure involving the Andean basement to significant depth. Consequently, the section in Figure 8a suggests the zone of west verging folds is ahead of the tip line of the propagating main thrust ramp of the WAT, extending from beneath the Frontal Cordillera anticline across the cover to the surface. Similar steep limbs in the cover above basement-involved structures have been described elsewhere and their relatively complex kinematics formalized [*Narr and Suppe*, 1994]. It is thus reasonable to interpret the western leading edge of the Frontal Cordillera as a hidden west vergent thrust wedge, characterized by kinematic complexity (West Andean Basement Thrust Wedge in Figure 3c). Then the main cascading west verging folds seen in the thick Andean Basin sediment pile may be the result of a series of (2–3) westward propagating ramps, thrusting westward, at the western side of the basement wedge. Conversely, the detachment of the cover at the base of the AFTB appears the result of a shallow back thrust on top of the gentle western slope of the rigid wedge, in a similar situation as that of the excess of material pushed by a bulldozer, which can overtop the bulldozer's hoe and flow backward on top of it (Figure 8b). As a consequence, all the shortening observed across Andean Basin cover in the Principal Cordillera (i.e., the overall ~80 km wide Andean fold-thrust belt, specifically including the San Ramón–Farellones Plateau frontal system, the west verging folds and the AFTB; see Figure 2b) has to be accounted for by basement shortening across the crustal ramp of the WAT beneath the Frontal Cordillera. The details of those mechanisms deserve an extended discussion that is beyond the scope of the present paper and will be provided elsewhere (R. Lacassin et al., manuscript in preparation, 2010).

4.5. Eastern Foreland: Hidden Back Thrust Margin and the Incipient East Andean Front Beneath the Cuyo Basin

[50] The changes proposed here for the tectonic interpretation of the Andes at 33.5°S latitude need to be confronted with our knowledge of structures in the eastern back thrust margin. East of the Frontal Cordillera is the eastern foreland of the Andes, represented by the Cuyo Basin (Figures 2b and 8a), which is a wide, relatively shallow Cenozoic basin, founded on the basement of the San Rafael block (Permian-Triassic Choiyoi Group over Cuyania terrane deformed during the Early Permian San Rafael orogenic phase [e.g., *Llambías et al.*, 2003; *Mpodozis and Ramos*,

1989]). The Cuyo Basin is very well studied because beneath the overlapping units of Andean synorogenic deposits it also contains some valuable Late Triassic oil-rich rocks which are interpreted to have been deposited in a Pre-Andean continental rift system (which alone is also called "Cuyo Basin" by petroleum geologists [e.g., *Irigoyen et al.*, 2000]). The structures formed during that Triassic Pre-Andean extensional phase are described as being widespread over the Late Paleozoic Gondwanan margin of South America [e.g., *Charrier et al.*, 2007, and references therein]. Indeed, those structures are interpreted to have played a very significant role in the subsequent basin inversion processes during the Andean cycle [e.g., *Ramos et al.*, 1996b; *Giambiagi et al.*, 2003; *Charrier et al.*, 2002, 2007]. The Andean eastern foreland deposits in the Cuyo Basin have been dated with magnetostratigraphy calibrated with ^{40}Ar-^{39}Ar dates from interbedded tephra [*Irigoyen et al.*, 2000]. According to these results, the deposition of the main synorogenic units, reaching a total thickness of ~2–4 km, has occurred since ~16 Ma, the early middle Miocene [*Irigoyen et al.*, 2000].

[51] The structures represented in Figure 8a under the synorogenic Cuyo Basin are inspired from published sections in the area [*Ramos et al.*, 1996b; *Brooks et al.*, 2003]. Our section at latitude 33.5°S (Figure 8a) cuts the hidden southward extension of the Precordillera structures, which attenuate rapidly southward [*Ramos et al.*, 1996b]. Thus, according to *Ramos et al.* [1996b], at this latitude little finite shortening has occurred across the Cuyo Basin and the bulk of the shortening appears restricted to the Principal Cordillera (across the AFTB), as the Frontal Cordillera has been uplifted as a (single) rigid block. Admittedly, the finite shortening is generally poorly constrained in the eastern foothills of the Frontal Cordillera [*Ramos et al.*, 2004]. In our cross section (Figure 8a), all the shortening across the Principal Cordillera is accounted for by the West Andean Thrust under the Frontal Cordillera. However, it seems improbable that the hidden back thrust at the eastern flank of the Frontal Cordillera and the hidden, attenuated Precordillera structures at this latitude (southward extension of Chacheuta–La Pilona–Tupungato and Barrancas anticlines [e.g., *Irigoyen et al.*, 2000]), which possibly correspond to inversion of structures in the Cuyo Basin [*Giambiagi and Ramos*, 2002], may represent an amount of finite shortening comparable with that in the Principal Cordillera. It is probably much less for two reasons.

[52] 1. The overall geometry and the relative shallowness of the synorogenic Cuyo Basin over its relatively flat rigid basement (Figure 8a) preclude any significant foreland flexure at the east flank of the Frontal Cordillera. Thus, a large hidden back thrust at this boundary appears unlikely.

[53] 2. The top of the basement flooring the Cuyo Basin appears only mildly affected by the Andean deformation,

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Figure 9. Photographs of the structure in the zone of large west verging folds (WVF, represented in Figures 2b and 8a; see precise location of photograph in Figure 2b). (a) In the largest vertical limb (~5 km wide, top-to-the-west geometry), ridges of limestone beds of Neocomian Lo Valdés Formation (locally dipping ~80°W) bearing weak cleavage dipping steeply eastward (~70°E). (b) Detail showing S_1/S_0 relation.

and the topography of the surface landscape is modest, so the actual finite shortening across the foreland east of the Frontal Cordillera must be relatively small.

[54]   Therefore, in contrast with earlier interpretations suggesting a large east vergent shallow dipping detachment in the basement under the Frontal Cordillera (which significantly maximize the shortening estimates [e.g., *Ramos et al.*, 1996b; *Brooks et al.*, 2003; *Giambiagi and Ramos*, 2002; *Ramos et al.*, 2004]), our interpretation in Figure 8a suggests that a series of steep crustal-scale ramps (which moderate the shortening estimates) may have developed on the back of the Frontal Cordillera anticline. Such an incipient eastern Andean Back Thrust Margin (see also profile B in Figure 1) cannot compare with the WAT at this latitude.

4.6. Western Foreland: Marginal Block as a Balance Between the Andes and the Subduction Zone

[55]   The large-scale monocline architecture of the marginal block in front of the Principal Cordillera is illustrated in the section at 33.5°S (Figure 8a) and is defined, at least over the more than 1500 km separating sections A and B in Figure 1, by the western, eastward dipping contact of the Andean Basin on top of the Coastal Cordillera basement. That architecture strongly suggests overall crustal-scale flexure of the western Andean foreland, partly explained by its eastward underthrusting beneath the Andes and its loading by the advancing thrusting under the Frontal Cordillera (Figure 8a). Furthermore, the eastward tilt of the marginal block appears also associated with uplift and substantial erosion of the Coastal Cordillera. The overall upward bulging of the Coastal Cordillera is documented by the observation of erosion surfaces beveling the east dipping Andean Basin sequence as well as Cretaceous granite intrusions. Relicts of such surfaces are preserved at high elevation (up to 2200 m) in the eastern Coastal Cordillera [*Brüggen*, 1950; *Borde*, 1966; *Farías et al.*, 2008]. Therefore, a long-lasting erosion process appears to have gradually reduced the relief created at the western edge of the Andean Basin by the bulging of the Coastal Cordillera, as illustrated in Figure 8a. It follows that the uplift of the Coastal Cordillera could be interpreted simply as elastic fore-bulging ahead of the foreland flexure. However, the proximity of the Coastal Cordillera with the subduction zone suggests that mechanical coupling across the subduction interface is the leading boundary condition. The following is an attempt to settle some basic features of that boundary condition, in view of the published observations.

[56]   The current mechanical processes described along the Chilean subduction zone appear to change significantly north and south of 33°S [*von Huene et al.*, 1997; *Yañez et al.*, 2001; *Laursen et al.*, 2002; *Ranero et al.*, 2006]. North of 33°S, a limited amount of sediment (<1 km thickness) accumulated in a narrow trench is associated with a dominantly erosive margin, which has substantially receded landward over the long term. South of 33°S, a trench 40 km wide flooded by a 2.5 km thick pile of turbidites is associated with a margin where active accretion of recent sediment dominates, but where episodes of accretion and erosion may have alternated over the long term [*Bangs and Cande*, 1997; *Kukowski and Oncken*, 2006]. The present-day point of intersection of the

Juan Fernandez Ridge (a hot spot seamount chain) with the front of the subduction zone is at 33°S. That collision point appears to have migrated southward along 1400 km of the margin since 20 Ma [*Yañez et al.*, 2001]. South of the Juan Fernandez Ridge, the south central Chile margin (between 33°S and 45°S) is characterized by a string of discrete, shelf to subshelf, margin-parallel basins of differing size, tapering toward the slope and the coast, separated from each other by subtle basement knolls [*González*, 1989; *Melnick and Echtler*, 2006]. Those basins are filled with Cretaceous and mostly Cenozoic sequences with thicknesses of up to ~3 km and appear to be "fore-bulge" basins, rather than "fore-arc" basins, because their structural development has no direct relation with the present-day volcanic arc and appears clearly confined to the western margin of the Coastal Cordillera bulge (Figure 8a).

[57]   The accretionary nature of basins south of the Juan Fernandez Ridge is substantiated by the occurrence of clear accretionary wedge structures in the frontal ~25 km of the margin, which may also be the site of some right-lateral decoupling [*González*, 1989; *Laursen et al.*, 2002; *Ranero et al.*, 2006]. The Valparaíso Basin (illustrated schematically in Figure 8a) can be considered as the northernmost basin of the south central Chile margin. It has been suggested that the sediment in the Valparaíso Basin might have been deposited and shortened at the back of a growing frontal accretionary wedge, against a continental backstop [*Ranero et al.*, 2006]. However, the Valparaíso Basin appears to be modified by young extensional faulting associated with the recent arrival of the Juan Fernandez Ridge, which marks the end of the dominant accretionary regime and the onset of dominant tectonic erosion [*Ranero et al.*, 2006]. Those incipient extensional features are not represented in Figure 8a. The removal of material from the upper plate by basal erosion may have also contributed to thinning and overall subsidence of the margin, putting the Valparaíso Basin into somewhat deeper water than other basins located southward (its present-day average depth of ~2400 m [*Laursen et al.*, 2002; *Laursen and Normark*, 2003; *Ranero et al.*, 2006]).

[58]   The late Miocene–early Pliocene sediments in the Navidad Basin, which is located immediately south of the Valparaíso Basin, contain benthic foraminifers and ostracodes that indicate deposition at lower bathyal depths (>2000 m [*Encinas et al.*, 2006, 2008; *Finger et al.*, 2007]). The landward side of the Navidad Basin sediments are now found forming cliffs along the coast at elevations reaching 200 m. Based on the stratigraphical and sedimentological evidence collected in the Navidad Basin and in similar basins located southward, so overall between 34°S and 45°S, *Encinas et al.* [2006, 2008], *Finger et al.* [2007], and *Melnick and Echtler* [2006] suggest rapid margin subsidence of >1.5 km in the late Miocene (starting at ~11 Ma), followed by rapid uplift by nearly the same amount since the early Pliocene (~3.6 Ma). Clearly, those features may be neither uniform, nor strictly synchronous along strike, but they indicate that the margin has probably undergone both, substantial subduction erosion [*Encinas et al.*, 2008] and tectonic uplift. Thus, the long-term uplift process of the Coastal Cordillera bulge may be punctuated by significant episodes of subsidence, suggesting the underlying dynamics is governed by alternating cycles of

accretion and erosion as deduced independently earlier [*Bangs and Cande*, 1997; *Kukowski and Oncken*, 2006]. We suspect, again as previously suggested [*Adam and Reuther*, 2000], that large-scale underplating of crustal rocks associated with basal erosion may have also occurred under the Coastal Cordillera, thus contributing to sustain the uplift.

[59] The foregoing suggests that the evolution of the margin adjacent to the Coastal Cordillera has involved over the long term a southward decreasing degree of subduction erosion for nearly 3000 km between ~18°S and ~45°S, apparently paralleling the southward decreasing degree of development of the Andes. Despite local changes associated with collision and migration of oceanic ridges, the erosive character of subduction appears to have been particularly intense and a permanently dominant feature during the whole Andean cycle (since the Jurassic) between ~18°S and ~33°S, parallel to where the Andes orogen is fully developed. On the average, subduction erosion of the margin appears to have been less intense and episodic between ~33°S and ~45°S, thus paralleling the region where the development of the Andes orogen tapers gradually southward.

[60] Summarizing, the marginal block in our section at 33.5°S (including the eroding/accreting margin, the basement of Coastal Cordillera and the undeformed part of the Andean Basin, see Figure 8a) may be seen as a large rigid board that is progressively inclined landward by the increasing Andean orogenic load over the long term, coevally with shortening (roughly ~2.5 × 10^7 yr), while it may also swing gently as a seesaw over shorter periods of time (some 10^6 yr), as a response to alternating cycles of subduction erosion and accretion at the continental margin. Within variations (which are not discussed here), this view of the marginal block appears to be valid for thousands of kilometers along the Andes.

4.7. Primary Vergence of the Andean Orogen and Its Possible Evolution

[61] Our complete tectonic section across the Andes (Figure 8a) emphasizes the west vergent structure and the primary asymmetry of the orogen at this latitude. That is expressed in the Principal Cordillera by the asymmetric deformation of the Andean Basin, west of the Frontal Cordillera backstop, and by the west propagating Andean fold-thrust system, with clear prowedge geometry (synthetic with the subduction zone, compare Figure 8a with sections in Figure 1). The prowedge geometry of the Andean Basin is comparable to the prowedge geometry of the continental margin, suggesting processes of similar scale may occur on the two leading edges of the marginal block. In our tectonic section (Figure 8a), we suggest that the West Andean Thrust may involve the lithospheric mantle and interpreted as an embryonic intracontinental subduction.

[62] On the back of the Frontal Cordillera there is a significant, albeit incipient, back thrust margin, which is known to be progressively more developed northward of 33.5°S (in the Precordillera, the Eastern Cordillera and the sub-Andean Belt [e.g., *Ramos et al.*, 2004]). This suggests that the Andean system at 33.5°S latitude may be evolving into a wider and more symmetric, doubly vergent orogen, as deformation propagates both eastward into the eastern foreland, and southward alongside the more developed back thrust system. The growth of an efficient back thrust margin should cause the relaxation of stresses across the WAT (the embryonic west vergent subduction), which could then decay or abort. As a consequence, progression of deformation across the back thrust margin should be further increased, in positive feedback. This very simple evolution scenario of the Andean orogen, derived from our section, predicts progressive thickening and widening eastward, and appears consistent with the occurrence of the Altiplano. A double-vergent growth process of the Altiplano, widening eastward by progressive back thrusting behind the West Andean Thrust would thus appear comparable to that proposed for the Tibet Plateau, widening north and northeastward by north verging progressive thrusting behind the Himalaya thrust system [e.g., *Tapponnier et al.*, 2001].

[63] Therefore, the architecture depicted in our tectonic section appears a fundamental stage of the Andean evolution (west vergent stage with dominance of the WAT). If this inference is correct, then a similar west vergent stage may have occurred in the past in the regions where the Andean orogen is more developed and where the rigid marginal block is identified, as in northern Chile (see profile A, Figure 1). So, building further on our evolutionary model, we anticipate that some features of a WAT stage of probable "Incaic" age (Paleogene age; after the name of a tectonic "phase" of the classical Andean geology, so older than the west vergent stage at 33.5°S) are preserved along the West Andean Front of northern Chile (studied among others by *Muñoz and Charrier* [1996], *Victor et al.* [2004], *Farías et al.* [2005], and *García and Hérail* [2005]), specifically in the Cordillera Domeyko, beneath the blanket of Neogene volcanic rocks. We also note that the occurrence of giant porphyry deposits in Chile (in a specific association with magmatism [e.g., *Maksaev et al.*, 2007]) are correlated with tectonic thickening during the Paleogene "Incaic" and the Neogene "Quecha" "phases" in northern and central Chile, respectively. So the hypothesis of propagation of both, the emplacement of porphyry deposits and the shortening seems attractive, and challenges the commonly used concept of synchronous tectonic "phase," particularly in the Andean geology. Altogether the foregoing arguments suggest that deformation would have propagated diachronically southward along the West Andean Thrust, and eastward to the Eastern Cordillera and the sub-Andean Belt.

[64] Finally, if our evolutionary model is correct, then the origin of the Andes is intrinsically associated with initiation, then propagation of the crustal-scale West Andean Thrust, so with the ripping apart by tectonic shear of the rigid marginal block from the main Gondwanan basement of South America. However, concerning the cover, the Andean orogenic cycle is closely associated with formation of the back-arc Andean Basin since the Jurassic, then during the Cenozoic with its deformation in the Principal Cordillera as a prowedge thrust over the down-flexed marginal block. Taken together, the two arguments suggest that the WAT has formed during the Cenozoic in a preexistent zone of weakness of the Mesozoic back arc. That zone of weakness may correspond to mechanical damage at crustal or litho-

spheric scale, by intense fracturing, stretching and thinning. In other words, the Andean orogenic cycle, characterized by long-lasting subduction processes, has first prepared by damaging the region where subsequent localized rupture of the West Andean Thrust has produced the orogenic shortening and thickening.

4.8. Impact of Measurable Constraints

4.8.1. Chronology

[65] The best chronostratigraphical constraints come from absolute dates of volcanic rocks in the Andean Basin and magnetostratigraphy calibrated with ^{40}Ar-^{39}Ar dates of tephra in the Cuyo Basin, as described by *Charrier et al.* [2002] and *Irigoyen et al.* [2000], respectively. According to our interpretation, on the one hand, the thick infill of the Andean Basin in the Principal Cordillera has started to deform by westward fault propagation folding in the late Oligocene to the early Miocene (~25–22 Ma), strictly not later than 21.6 Ma [*Charrier et al.*, 2002], and the shortening process has continued throughout to the present time by westward propagation of the WAT up to the surface at the West Andean Front. On the other hand, synorogenic deposition on the back of the Frontal Cordillera ramp anticline (in the eastern foreland Cuyo Basin) appears to have commenced since the early middle Miocene (~16 Ma) [*Irigoyen et al.*, 2000]. Thus, the orogenic uplift of the Principal Cordillera (pushed by the Frontal Cordillera anticline, above the WAT) would have been followed by a sedimentary response in the eastern foreland with a delay of about 8–11 Myr. Results in the small intermontane Alto Tunuyán basin are less well constrained by direct geochronology than in the larger Andean and Cuyo basins [*Giambiagi et al.*, 2001]: Deposition of synorogenic units there appear bracketed between a maximum age provided by a sample of volcanic rock dated at 18.3 Ma (base) and the minimum age of an andesitic flow dated by K/Ar at 5.8 Ma (top). So according to these observations, the shallow back thrust deformation associated with the Aconcagua Fold-Thrust Belt would have started in the late early Miocene (~18 Ma). The tectonic model in Figure 8a is basically consistent with all these information.

[66] The differences between available ages are relatively small, reducing the resolution of detailed scenarios for the progression of deformation, but it is not unreasonable to envision the following one.

[67] 1. By ~25 Ma, onset of the Andean deformation with the growth process and the upward propagation of the West Andean Thrust beneath the Frontal Cordillera.

[68] 2. By ~18 Ma, the basement-involved ramp would have already produced ahead of it a significant amount of deformation in the Andean Basin cover. As a result, the zone of west verging folds and large vertical limbs would have generated there the first upheaval of kilometric-scale topographic relief of the Andes at this latitude (in the central eastern Principal Cordillera), the erosion of which having provided the first important source of new sediment (see Figure 8a). This is consistent with the clast composition (dominance of Cenozoic volcanic clasts) and paleocurrent observations (dominantly eastward transport) in the oldest unit (Tunuyán Conglomerate) of the Alto Tunuyán basin,

immediately east of the AFTB (see location in Figure 2b) [*Giambiagi et al.*, 2001].

[69] 3. By ~16 Ma, the upbulging of the Frontal Cordillera ramp anticline would have been enough to create topographic relief and to start providing sediment to the Cuyo Basin (eastern foreland). This issue deserves more discussion. According to sedimentary evidence, the Alto Tunuyán basin (located to the west of the Frontal Cordillera, see Figure 2b) would have started to record the uplift of the Frontal Cordillera with the deposition of the Palomares Formation [*Giambiagi et al.*, 2001]. A maximum age of ~12 Ma has been attributed to those deposits, based on a correlation with dated distant sediments deposited on the other flank of the Frontal Cordillera (the eastern foreland properly), in the Cuyo Basin [*Giambiagi et al.*, 2001]. This combined argument is at the origin of the generally accepted inference that the Frontal Cordillera is a very late feature of the Andes [e.g., *Giambiagi et al.*, 2003; *Ramos et al.*, 2004; *Farías et al.*, 2008]. In turn, that inference is difficult, although not impossible, to reconcile with the architecture of the Andes depicted in Figure 8a. Alternatively, in the absence of more precise age constraints, the Palomares Formation could be as old as ~16 Ma (consistent with the ages obtained directly in the Alto Tunuyán basin) and the gradual upbulging of the Frontal Cordillera would have been recorded on its two flanks roughly the same time.

[70] 4. The occurrence of discrete "out-of-sequence" thrust faults within the Principal Cordillera has been proposed [e.g., *Ramos et al.*, 2004; *Farías et al.*, 2008]. Our model describes a large prowedge associated with the WAT, producing shortening throughout the Principal Cordillera, from the frontal San Ramón Fault to the AFTB, so explaining any discrete thrust within that region. The evolution of that system may include the proposed "out-of-sequence" thrusts. Our model is also consistent with a late propagation of deformation into the eastern foreland, on the back of the Frontal Cordillera, so present-day shortening may occur as well throughout the hidden back thrust margin beneath the Cuyo Basin (Figure 8a).

4.8.2. Cumulative Shortening, Shortening Rates, and GPS Velocities

[71] Our section at 33.5°S implies shortening of about 30–40 km throughout the Principal Cordillera prowedge that deforms the Andean Basin, west of the Frontal Cordillera backstop (so associated with the WAT, Figure 8a). Specifically, those rough minimum and maximum shortening estimates result from adding the shortening of 10 km observed across the San Ramón–Farellones Plateau (western Principal Cordillera), to a minimum-maximum of 15–20 km shortening across the zone of large west verging folds (central Principal Cordillera), and a minimum-maximum of 5–10 km shortening across the AFTB (eastern Principal Cordillera). By contrast, deformation across the Cuyo Basin, corresponding to the retrowedge east of the Frontal Cordillera implies probably no more than ~10 km shortening. Thus our model suggests overall cumulative shortening of 35–50 km throughout the Andes at this latitude. The corresponding average shortening rate across the Andes over the past 25 Ma would be quite slow, of ~1.4–2 mm/yr (consistent with the 0.4 mm/yr slip rate on the San Ramón basal detachment).

[72] For comparison, *Giambiagi and Ramos* [2002] estimated for the same transect of the Andes a total shortening of 70 km (47 km across the Principal Cordillera, 16 km in the Frontal Cordillera and 7 km across the Cuyo Basin), which over their suggested maximum age of 17 Ma yields an average shortening rate of ~4 mm/yr. Their estimate of shortening across the Principal Cordillera is not very much higher than ours, so the two estimates could be considered consistent with each other. However, the structural interpretation of the AFTB and the Frontal Cordillera is radically different in the two models. We do not believe that the large-scale, east verging, very complex, system of thrusts (and duplexes) involving deeply the basement (thick-skinned thrusting), which *Giambiagi and Ramos* [2002] and *Giambiagi et al.* [2003] propose under the AFTB and the Frontal Cordillera are directly justified by any compelling (geological or geophysical) evidence. Therefore, even if that hypothetical, nonunique structure (among many other possible complex structures) may appear plausible (and we cannot rule it out), we prefer to keep our more conservative estimate based on the structural evidence and arguments presented in sections 2, 3, and 4.

[73] Regardless of the foregoing discussion, significantly higher values of shortening (~300 km) and shortening rates (6–7 mm/yr) have been found in the Andes northward of the transect at 33.5°S, specifically localized at the eastern Andean back thrust margin [e.g., *Kley and Monaldi*, 1998; *Ramos et al.*, 2004].

[74] The interpretation of the velocity field deduced from available GPS measurements across the Andes appears controversial [e.g., *Khazaradze and Klotz*, 2003; *Brooks et al.*, 2003; *Kendrick et al.*, 2006; *Norabuena et al.*, 1998; *Vigny et al.*, 2009]. There are two main reasons for discrepancies. On the one hand, the sparse coverage of stations within the Andes generally does not allow for resolving local deformation. On the other hand, on the regional scale, the strong interseismic signal produced by the coupling at the subduction interface appears dominant over distances of several hundreds kilometers from the trench. Therefore, the important subduction signal, which is transient, has to be removed from the total velocity field. Then a residual velocity field (supposed to be less transient and more permanent) is recovered, from which the "permanent" shortening rates in the Andes are inferred. Given the uncertainties concerning the degree of coupling across the subduction interface, it is clear that any inference on shortening rates in the Andes is model-dependent. Between 30°S and 34°S latitude, under the maximizing hypothesis of a fully locked subduction interface and a three-plate model (a third microplate is introduced between Nazca and South America), shortening rates of 4.5 mm/yr and less than ~3 mm/yr are obtained by *Brooks et al.* [2003] and *Kendrick et al.* [2006], respectively. Conversely, as shown by *Vigny et al.* [2009], the subduction is only partially locked between 30°S and 32°S, implying that the "permanent" shortening in the Andes at these latitudes would be within the error of the interseismic models and therefore unresolved by the present GPS network. This suggests that the transient character of the interseismic loading along the subduction zone may seriously alter the image of the subtler permanent Andean strain, as deduced from the GPS data on hand.

[75] We conclude that the shortening rates throughout the Andes at 33.5°S latitude, as determined with the available geological and geodetic observations, are very slow on the average (no more than ~2–4 mm/yr), compared to the much faster convergence rates at the subduction zone (in the range of 63–68 mm/yr [*Norabuena et al.*, 1998; *Brooks et al.*, 2003; *Vigny et al.*, 2009]). Shortening rates in the Andes are also possibly nonuniform in space and time. It is possible that episodes of relatively accelerated rates might be associated with particular stages of the faulting evolution. An efficient and well-localized back thrust margin, such as that observed north of 33°S in the Precordillera and the sub-Andean Belt [e.g., *Allmendinger et al.*, 1997; *Kley and Monaldi*, 1998; *Ramos et al.*, 2004], could represent a favorable boundary condition for such an acceleration.

4.8.3. Seismicity

[76] The shallow seismicity associated with the western flank of the Andes and the WAT is poorly known, mainly because of its poor record by local networks [*Barrientos et al.*, 2004]. The recent deployment of instruments in the region of the present study by the Chilean Servicio Sismológico Nacional (with good resolution near Santiago, between 33°S and 35°S) provides the image of the 2000–2005 seismicity with local magnitude Ml > 4.0 illustrated in Figure 8c. Beside the seismicity associated with the subduction zone, there is significant shallow seismicity (depth of <20 km) under the Principal Cordillera, which is consistent with our west vergent thrust model. However, the data are not accurate enough to discriminate between a variety of models including the alternative east vergent thrust model. In our model, the shallow seismicity is mostly concentrated ahead of the Frontal Cordillera ramp anticline and apparently above the basal detachment. It may thus be associated with deformation of the Andean Basin cover, and more precisely, with the westward fault-propagating fold structure behind the San Ramón Fault.

[77] The Principal Cordillera region has been the site of significant shallow earthquakes with M > 6.0, like the 1958 Las Melosas sequence discussed in section 3.5 [*Barrientos*, 2007; *Alvarado et al.*, 2009]. All these events occurred beneath the zone of west verging folds, thus probably associated with the ramp system beneath the Frontal Cordillera anticline. Fault plane solutions obtained for six M ≥ 5.0 events are consistent with P axes oriented NE–SW to NW–SE [*Barrientos et al.*, 2004; *Barrientos*, 2007; *Alvarado et al.*, 2009]. The three (or four) larger events discussed by *Barrientos* [2007] and by *Alvarado et al.* [2009] are strike-slip events. All but one of these events are consistent with E–W compression, thus with the Andean shortening. The remaining one is the Las Melosas main shock, indicating NW–SE compression, which is inconsistent with the right-lateral component of the Nazca–South America plate motion. The complex kinematics of events occurring in this region may be due to the complexity expected for the basement-involved structures that have propagated beneath [*Narr and Suppe*, 1994]. In particular, the steep front of the basement wedge appears a likely place for lateral decoupling (Figure 8a).

4.8.4. Crustal Structure

[78] Crustal thickness estimates in the region come from studies of gravity data [*Introcaso et al.*, 1992; *Tassara et al.*, 2006] and from studies using various seismological techniques [*Fromm et al.*, 2004; *Gilbert et al.*, 2006; *Alvarado et al.*, 2007]. Along the section at 33.5°S, the crustal thickness reaches a maximum of ~50–60 km beneath the high Andes and diminishes both eastward and westward. This is roughly consistent, within uncertainties, with the structure we suggest, as shown in Figure 8c. To test further the consistency of the west vergent thrust model requires a better resolved image of the deep Andean structure. Broadband data and receiver functions have been used to that purpose [*Gilbert et al.*, 2006]. Cross sections at 30.5°S and 35.5°S established by *Gilbert et al.* [2006], so to the north and south of our section in Figure 8, show a clear Moho recorded by a strong simple signal at ~40 km depth, by stations located far into the eastern foreland of the Andes. As noted by the authors, that signal cannot be followed westward when looking at stations close to the main Andes. The clear Moho arrivals disappear, or appear interrupted, possibly by crustal-scale faults. *Gilbert et al.* [2006] show that stations in the Andes do not record Moho arrivals that appear coherent on the multiple traces. Interestingly, the thickest crust of 64 km is found in the stacked receiver function at one of those stations with incoherent arrivals (USPA, located precisely at the eastern flank of the Frontal Cordillera). Thus this station records also significant arrivals above the inferred Moho, which are therefore interpreted as midcrustal arrivals [*Gilbert et al.*, 2006]. We suspect that the image of structural complexity given by the receiver functions beneath the eastern flank of the Andes (giving the impression of more than one Moho) probably reveals the superposition, by thrusting, of two crustal-scale units. This inference is consistent with the ~30–40 km shortening of the Andean prowedge, and, as suggested in Figure 8a, with the West Andean Thrust involving the lithospheric mantle and interpreted as intracontinental subduction.

5. Conclusions

[79] Our purpose was to show that the San Ramón thrust system is an active fault that is critical for the seismic hazard in the city of Santiago and also a key structure to describe the primary architecture of the Andes and its possible evolution. So, our conclusions are twofold.

5.1. Concerning the San Ramón Fault

[80] The San Ramón Fault is a multikilometric frontal thrust at the western front of the Principal Cordillera and interpreted as a growing west vergent fault propagation fold system. Its basal detachment is close to the base of the Andean Basin at ~12 km or more of stratigraphical depth and probably localized at ductile layers of gypsum of Late Jurassic age.

[81] The fault-propagating fold structure associated with the San Ramón Fault is well constrained by the mapped surface geology and can be restored to deduce amounts of shortening and uplift. Total shortening across the frontal system is ~10 km. It has occurred since ~25 Ma, according

to precise dates in the Abanico and Farellones formations. The shortening, uplift and erosion rates across the Principal Cordillera are inhomogeneously distributed and the degree of erosion appears correlated with structure wavelength. So, no extensive erosion surface has developed.

[82] The San Ramón Fault reaches the surface with steep eastward dip, producing a probable total throw of ~4 km (net thrust slip of ~5 km), according to the apparent structural thrust separation across the Abanico and Farellones formations. The minimum average slip rate on the San Ramón Fault would be of the order of 0.2 mm/yr (5 km in 25 Myr) and the slip rate on the basal detachment of the frontal system of the order of 0.4 mm/yr (10 km in 25 Myr). However, the growth process of the thrust front suggests that the most of the slip on the basal detachment has localized since 16 Ma in the San Ramón Fault, making of it the frontal ramp of the Principal Cordillera. Thus, the actual long-term average slip rate on the San Ramón Fault would be of ~0.3 mm/yr (throw rate of ~0.25 mm/yr).

[83] The San Ramón piedmont scarp of Pleistocene age has been mapped in detail along a 15 km long fault segment facing Santiago, despite structural complexities in the northern sector (Cerros Calán, Apoquindo and Los Rulos) and rapid urbanization of the eastern districts of the city, obliterating the fault trace. The younger, most regular part of the piedmont scarp reveals minimum average throw of ~60 m. The occurrence in the piedmont of ash lenses correlated with the Pudahuel ignimbrites yields a strictly minimum throw rate of 0.13 mm/yr (≥60 m in ≤450 kyr).

[84] Throw of about 4 m was measured across a well-preserved scarp that appears to be the last testimony to late scarp increments left for study along the San Ramón Fault. This feature is to be accounted for by a single event or by several events with thrust slip of the order of ~1 m or less. A conservative estimate using a range of average slip of 1 to 4 m, consistent with rupture of the entire length of the San Ramón mountain front facing the Santiago valley (~30 km) and with the well-constrained hypocenters down to 15 km depth under the Principal Cordillera, yields seismic moments of M_o ~0.3 to 1.2×10^{20} N m, corresponding to events of magnitude M_w 6.9 to M_w 7.4. Events that large could not be disregarded for seismic hazard assessment in the Santiago region. Recurrence time for such events would be very long, of the order of 2500–10,000 years.

5.2. Concerning the Primary Large-Scale Tectonics of the Andes

[85] The present study of the San Ramón Fault uncovers the primary importance of the propagating West Andean Front, interpreted as the tip of the West Andean Thrust (WAT), so implying substantial changes from the currently accepted interpretations. Our tectonic section at the latitude of Santiago synthesizes the main results (Figure 8), which are summarized hereafter step by step:

[86] We show that the West Andean Front must be rooted in downward to the east, beneath the high Principal Cordillera and probably beneath the basement of the Frontal Cordillera. The Frontal Cordillera is a huge basement anticline ~5 km high and more than ~700 km long, located side by side with

the Principal Cordillera. The thick Andean Basin (12 km thick or more), which constitutes the bulk of the Andean fold-thrust belt in the Principal Cordillera, appears clearly deformed as a west verging prowedge, ahead of the Frontal Cordillera. We infer that the Frontal Cordillera is the crustal-scale ramp anticline that, as a bulldozer, provides the necessary boundary conditions to maintain the high elevation in the Principal Cordillera and to cause the westward propagation of the San Ramón thrust system. So, the primary Andean structure at the latitude of Santiago has a decided westward vergence.

[87] A prominent zone of west verging folds of the thick Andean cover in the middle of the Principal Cordillera appears to mark at the Earth's surface the tip of the propagating main west vergent thrust ramp system associated with the Frontal Cordillera anticline. Huge vertical limbs (implying an overall ~15 km vertical separation of the Andean Basin infill) and a complex kinematics are observed at these basement-involved structures. The Aconcagua Fold-Thrust Belt (eastern part of the Principal Cordillera) appears to be a shallow subsidiary back thrust on top of the Frontal Cordillera anticline. On the back of the Frontal Cordillera is the eastern foreland of the Andes, represented by the relatively modest Cuyo Basin (no more than ~2–4 km thickness), which cannot be interpreted as a large flexural basin. An incipient back thrust margin probably including a series of steep crustal-scale ramps on the back of the Frontal Cordillera anticline appears hidden beneath the Cenozoic sediments of the Cuyo Basin. So, the structure of the Andes at this latitude is strongly asymmetric and its doubly vergent character very incipient.

[88] At the subduction margin, the rigid marginal block appears to act as a balance between forces applied by the Andes across the WAT and the subduction zone. The extensively eastward dipping Andean Basin on top of the Coastal Cordillera basement indicates crustal-scale flexure of the western foreland associated with eastward underthrusting of the marginal block beneath the WAT, and its consequent loading by the weight of the Andes. Alternating cycles of subduction erosion and accretion at the continental margin punctuate the long-term uplift process of the Coastal Cordillera. The marginal block has similar characteristics for thousands of kilometers alongside the Andes, suggesting it is a fundamental feature of the mechanical partitioning between orogenic and subduction processes.

[89] The chronostratigraphic constraints suggest slow deformation processes across the Andes. Orogenic uplift of the Principal Cordillera would have been followed by a sedimentary response in the eastern foreland with a relatively long delay of about 8–11 Myr. Cumulative shortening of

35–50 km throughout the Andes at this latitude implies a modest average shortening rate of the order of ~2 mm/yr, consistent with GPS results. Shallow seismicity under the Principal Cordillera apparently ahead of the WAT is significant, but its record hampered by insufficient instrumental coverage. Maximum crustal thickness of ~50–60 km beneath the high Andes is consistent with our suggested structure. The complex image of the deep Andean structure given by receiver functions reveals interruption of Moho arrivals, suggesting to us superposition by the West Andean Thrust of crustal-scale units and involvement of the lithospheric mantle in an embryonic intracontinental subduction.

[90] We note that the stage of primary westward vergence with dominance of the WAT at 33.5°S is evolving into a doubly vergent configuration, consistent with the overall eastward and southward propagation of deformation in the central Andes and the Altiplano (south of 18°S). A growth model for the WAT-Altiplano similar to the Himalaya-Tibet is suggested. We anticipate that the west vergent stage is ubiquitous in the central Andes and that it should have occurred earlier in the regions where the Andean orogen is more developed (specifically in northern Chile between 18°S and 26°S). It is deduced that the shear on the WAT has localized during the Cenozoic in a preexistent zone of weakness of the Mesozoic back arc, characterized by damage at crustal or lithospheric scale. The thrusting of the marginal block by Gondwanan South America has given way to the inception of partitioning between subduction and orogenic processes. So, the origin of the Andes appears intrinsically associated to the occurrence and propagation of the West Andean Thrust, improving our mechanical understanding of the Andean orogenic cycle and its specific association with a long-lasting subduction. The occurrence of the WAT reduces the differences between the Andean orogen and other doubly vergent orogens associated with continental collision, like the Himalayas: The intracontinental subduction at the West Andean Thrust may act as a mechanical substitute of the collision zone. In any case, the Andean orogeny paradigm may be considered obsolete.

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