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ORIGINAL PAPER

Cenozoic tectonic evolution in the Central Andes in northern Chile and west central Bolivia: implications for paleogeographic, magmatic and mountain building evolution

Reynaldo Charrier · Gérard Hérail · Luisa Pinto · Marcelo García · Rodrigo Riquelme · Marcelo Farías · Nelson Muñoz

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Abstract A review of available stratigraphic, structural, and magmatic evolution in northernmost Chile, and adjacent Peru and Bolivia shows that in this region: (1) compression on the Paleogene intra-arc during the middle Eocene Incaic phase formed the NNE-SSW-oriented Incaic range along the present-day Precordillera and Western Cordillera, and (2) post-Incaic tectonic conditions remained compressive until present, contrasting with other regions of the Andes, where extensional episodes occurred during part of this time lapse. A late Oligocene–early Miocene peak of deformation caused further uplift. The Incaic range formed a pop-up structure bounded by two thrusts systems of diverging vergencies; it represented a major paleogeographic feature that separated two domains with different tectonic and paleogeographic evolutions, and probably formed the Andean water divide. This range has been affected by intense erosion and was symmetrically flanked by two major basins, the Pampa del Tamarugal and the Altiplano. Magmatic activity remained located along the previous Late Cretaceous–early Eocene arc with slight eastward shift. Further compression caused westvergent thrusting and uplift along the western Eastern Cordillera bounding the Altiplano basin to the east by another pop-up shaped ridge. Eastward progression of deformation caused eastvergent thrusting of the Eastern Cordillera and Subandean zone.

R. Charrier · L. Pinto · M. García · M. Farías
Departamento de Geología, Facultad de Ciencias Físicas y Matemáticas, Universidad de Chile, Casilla 13518, Correo 21, Santiago, Chile
e-mail: lpinto@ing.uchile.cl

M. García
e-mail: mgarciagodoy@ing.uchile.cl

M. Farías
e-mail: mfarias@dgf.uchile.cl

R. Charrier (✉)
Escuela de Ciencias de la Tierra, Facultad de Ingeniería, Universidad Andres Bello, Santiago, Chile
e-mail: rcharrie@ing.uchile.cl; rcharrier@unab.cl

R. Charrier · L. Pinto · M. García · M. Farías
Advanced Mining Technology Center (AMTC), Facultad de Ciencias Físicas y Matemáticas, Universidad de Chile, Santiago, Chile

R. Charrier · G. Hérail · L. Pinto · M. García · R. Riquelme · M. Farías
Agreement IRD-Departamento de Geología, Universidad de Chile, Santiago, Chile
e-mail: gerard.herail@ird.fr

R. Riquelme
e-mail: riquelme@ucn.cl

G. Hérail
Institut Français pour la Recherche en Developement (IRD), LMTG, 14 Avenue Edouard Belin, 31400 Toulouse, France

Present Address:
G. Hérail
IRD Mission in Peru, Casilla 18-1209, Lima 18, Peru

R. Riquelme
Departamento de Geociencias, Universidad Católica del Norte, Avenida Angamos 0610, Antofagasta, Chile

N. Muñoz
Origin Energy, Isidora Goyenechea 3000, Floor 13, Santiago, Chile
e-mail: munoz.eyp@gmail.com

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Introduction and geologic setting

The Andean Cordillera, developed along the western continental margin of South America, is the typical example of a subduction-related mountain belt (cordilleran-type mountain belt sensu Dewey and Bird 1970), formed through crustal shortening and thickening, magmatic addition and intense erosion along the border of the overriding plate above the underthrusting Nazca Plate and the intervening mantle wedge (Isacks 1988). However, the shallow and deep-seated lithospheric processes or combination of processes that concur to the deformation, uplift and to maintain the altitude of the orogene are still poorly constrained. The enormous crustal thickening and uplift of the Altiplano-Puna region has particularly stimulated discussion on this subject and several processes and models, and combinations of them, have been proposed to explain orogenic development in this region, for example: (1) Large-scale monoclinal folding (Isacks 1988; Jordan et al. 2010), (2) Absolute or relative motion of the intervening plates (Soler and Bonhomme 1990; Silver et al. 1998; Ramos 2010), (3) High shear stress at the plate interface resulting from lack of sediment in the trench (Lamb and Davis 2003), (4). Erosion versus isostatic recovery (Montgomery et al. 2001), (5) Lithospheric removal (Platt and England 1993) or delamination (Kay and Kay 1993; Schurr et al. 2006; Garzione et al. 2006; Jiménez et al. 2009), (6) Isostatic response of crustal shortening and thickening (Gubbels et al. 1993; Muñoz and Charrier 1996; Victor et al. 2004; Elger et al. 2005; Farías et al. 2005). Convergence between these plates seems to have been rather continuous since Early Jurassic (Coira et al. 1982; Jordan et al. 1983, 1997; Ramos 1988; Mpodozis and Ramos 1989; Kay and Abbruzzi 1996; Kay et al. 1999; Charrier et al. 2007). However, major deformation and associated episodes of uplift occurred in late Early Cretaceous (Peruvian Phase), Middle Eocene (Incaic phase), and Late Oligocene-Early Miocene times indicating the existence of peaks or episodes of increased deformation. The last of these phases is associated with a major change in the relative movement and a considerable increase of the convergence rate between the oceanic and continental plates that occurred after breakup of the Farallon into the Nazca Plate (Pilger 1983; Pardo-Casas and Molnar 1987; Somoza 1998; Reutter 2001). On the other hand, some authors consider that a generalized uplift event occured somewhat later at 10–7 Ma (Gregory-Wodzicki 2000; Garzione et al. 2006; Hoke et al. 2007). Although the geodynamic conditions along the continental margin seem to have been continuous since Late Cretaceous until present, recent studies indicate that the tectonic regime affecting the southern Central Andes varied along strike (Charrier et al. 2009). The present article is part of a broader project attempting to determine along strike variations in the tectonic evolution in the southern central Andes.

Intensive studies in the last decades in the northern Chilean Andes provided abundant information allowing detailed reconstructions of the stratigraphic, structural, magmatic, and paleogeographic evolution in this region, which has been summarized in Charrier et al. (2007). However, there is no integration of this information with that of the eastern flank of the Andes hindering a coherent view of the evolution of the entire mountain range. This is due to the sparse information that existed on the Altiplano region, a situation that has recently changed with the work of different multidisciplinary groups working in that region.

In this article, we synthesize available stratigraphic, sedimentologic, structural, and chronologic evidence for the late Paleogene to present evolution on both sides of the Andes in northern Chile, southern Peru, western Bolivia and northwestern Argentina, in order to determine the prevailing tectonic conditions that controlled evolution of the mountain range. This information is essential to understand the formation and evolution of the major paleogeographic features (ridges and basins) associated with the development of subduction-related orogens. This information and the interpretations deduced from it in this article will be compared with that of other regions in the central Andes, where alternating compressive and extensional tectonic regimes have been detected, in order to deduce the causes for such differences and their bearing on surface uplift, dimensions of the orogen, and on the magmatic and metallogenic evolution. This attempt resulted in an integrated view of both sides of the Andes in this region.

The region analyzed in this study is part of the southern, N–S-oriented portion of the Bolivian Orocline (Figs. 1, 2). This region corresponds to one of the highest regions of the mountain range with an average altitude of $\sim$3,800 m in the Altiplano plateau, and summits reaching over 6,000 m in the Western and Eastern cordilleras. Here, the crustal thickness attains >65 km below the Altiplano and diminishes westward to $\sim$35 km in the forearc region and eastward to $\sim$30 km in the foreland region (Yuan et al. 2002).

The first, major, late Early Cretaceous Peruvian orogenic phase in this Andean region caused strong deformation, uplift and a major paleogeographic reorganization in this Andean region (Mpodozis and Ramos 1989; Charrier et al. 2007). After this phase, the arc, which was previously located along the Coastal Cordillera, in Jurassic and Early Cretaceous times, shifted to the east and formed a NNE–SSW-oriented band of magmatic rocks that extends

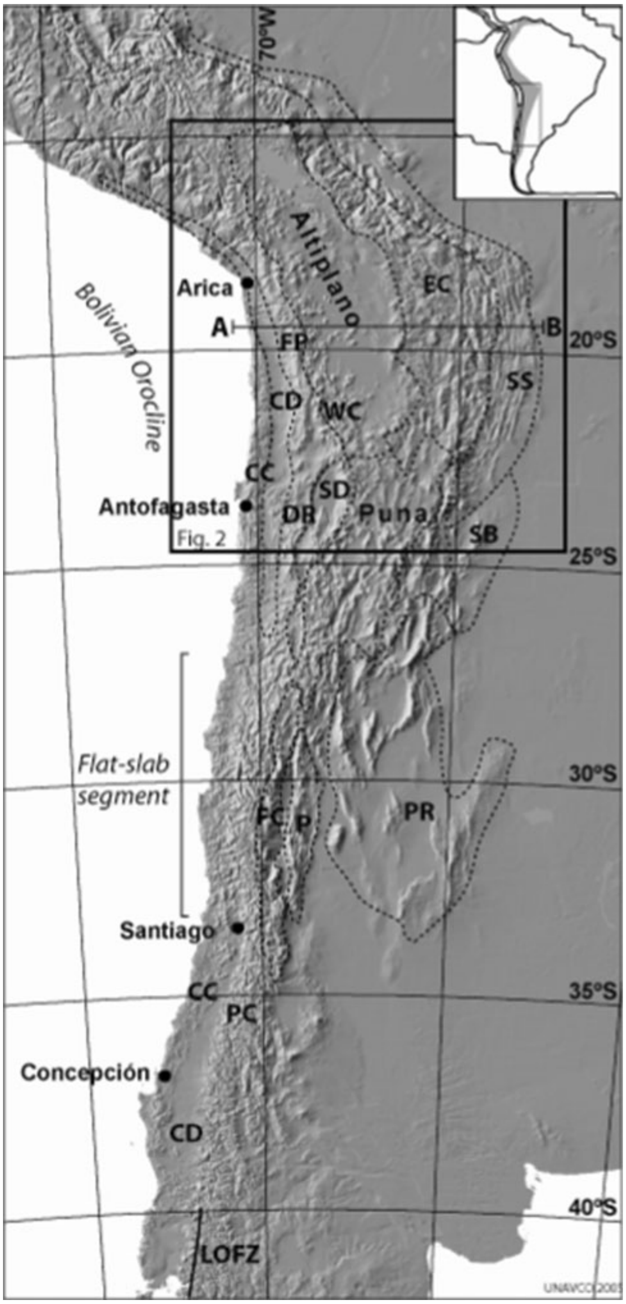

Fig. 1 Digital elevation model of the Andes, between 15° and 40°S, with indication of the Bolivian Orocline, the flat-slab segment in central Chile and Argentina, the morphostructural units and the region analyzed in this article in southern Perú, northern Chile, central western Bolivia and northwestern Argentina (*rectangle*). *Line AB* indicates cross-section in Fig. 11. *Abbreviations CC* Coastal Cordillera, *CD* Central Depression, *DR* Domeyko Ridge, *EC* Eastern Cordillera, *FC* Frontal Cordillera, *FP* Forearc Precordillera (western flank of the Altiplano) and, further south, the Sierra de Moreno and the Domeyko Ridge (DR), *LOFZ* Liquiñe-Ofqui Fault Zone, *P* Precordillera in Argentina, *PC* Principal Cordillera, *PR* Pampean Ranges, *SB* Santa Barbara System, *SD* Salars Depression, *SS* Subandean Sierras, *WC* Western Cordillera

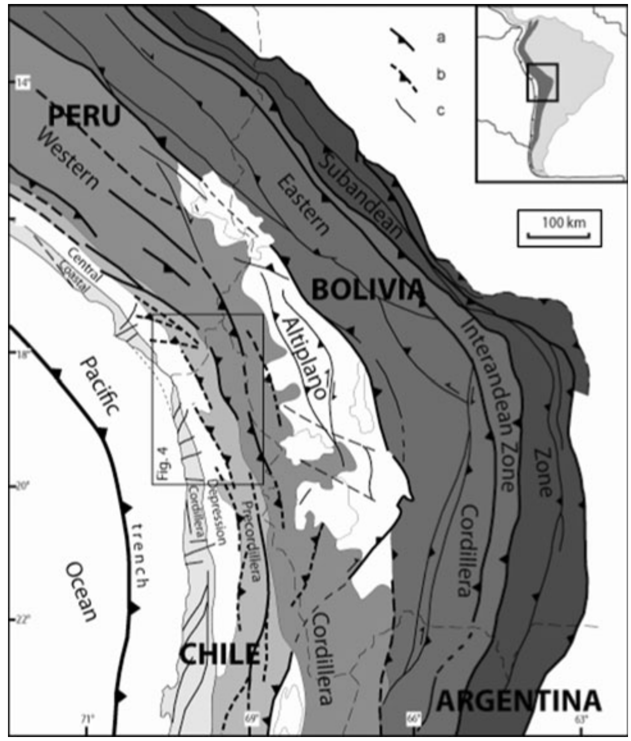

Fig. 2 Structural outline of the Central Andes, in northern Chile, southern Peru, western Bolivia and north-western Argentina, showing the major faults that control the morphostructural units. (*a*) Major Neogene reverse fault; (*b*) covered or partially-reactivated (during the Neogene) reverse fault; (*c*) pre-Neogene fault or minor Neogene fault; (*d*) international boundaries. *Rectangle* indicates region in Fig. 4

along most of northern and central Chile, and northward into southern Perú (Fig. 3). The magmatic activity, which developed in an extensional intra-arc setting, lasted until mid-Eocene times, with only one short interruption close to the Cretaceous-Paleogene boundary ('K–T' tectonic event). At this moment a new, major compressive tectonic event, the Incaic phase (Steinmann 1929; Charrier and Vicente 1972; Maksaev 1978; Noble et al. 1979; Mégard 1984; Mpodozis and Ramos 1989; Reutter 2001; Cornejo et al. 2003; Charrier et al. 2007), caused tectonic inversion of the previous (Late Cretaceous–Early Paleogene) magmatic arc. The resulting uplifted region formed an elongated NNE–SSW-oriented ridge, the Incaic range (Charrier et al. 2007, 2009), which developed along the present-day Domeyko range and its northward prolongation, the Sierra de Moreno (Fig. 1). In northernmost Chile, although the exact width of the Incaic range is somewhat obscured by the young and widely extended volcanic and sedimentary cover, its axis was located approximately along the present-day eastern Forearc Precordillera and Western Cordillera (Figs. 1, 3). This range can be followed northward into southern Perú, where it forms the "substructure of the Precordillera" (Tosdal et al. 1984).

Fig. 3 Distribution of the over 3,000-km-long Incaic range in southern Peru and Chile, north of 39°S, after Charrier et al. (2007). Extent of the range is based on the distribution of magmatic outcrops of the Late Cretaceous to early Eocene dissected arc, according to SERNAGEOMIN (2003)

Uplift of the Incaic range caused intense erosion on the exposed Paleozoic and Mesozoic units that formed its core and supplied great volumes of detritic sediments, which accumulated in the depressed regions on both sides of the range (Charrier et al. 2007, 2009). The Incaic range played an essential paleogeographic role during most of the Cenozoic evolution in the southern Central Andes and represented a sort of "divide axis" between two domains with different tectonic and paleogeographic evolutions. For this reason, we will analyze separately the tectonic and paleogeographic evolutions west and east of this axis, or Incaic axis.

In the study region, the western part of the Andean range consists of the following morphostructural units, from West to East: Coastal Cordillera, Central Depression, Forearc Precordillera (from here on, Precordillera), Volcanic or Western Cordillera, and Altiplano. In the eastern part of the range, the morphostructural units are the Eastern Cordillera, the Interandean Zone, and the Subandean Zone, in Bolivia and northern Argentina (Figs. 1, 2). The Coastal Cordillera is slightly oblique to the coast-line and its width diminishes gradually northwards until it disappears in the Arica region (18°15′S); further north it reappears in southern Peru. Its altitude is less than 1,200 m and consists essentially of Mesozoic rocks (Salas et al. 1966; García et al. 2004) and forms smooth hills and shallow valleys (Mortimer and Saric 1972). The Central Depression is approximately 40–55 km wide and filled with Oligo-Miocene flat-lying sediments and ignimbrites, which are dissected by deep valleys, such as the Lluta, Azapa, Vítor Camarones and Tiliviche-Camiña, along which the melt waters from the Western Cordillera flow westward toward the ocean (Fig. 4). The flat interfluves between these gorges, the "pampas" (Börgel 1983), correspond to the surface of the Tarapacá pediment (not to be confused with the Tarapacá Coastal pediplain, defined by Mortimer and Saric 1972) (Mortimer et al. 1974; Isacks 1988; Lamb et al. 1997; García et al. 2004; Farías et al. 2005; García and Hérail 2005) that extends eastwards to the Precordillera. This surface increases its altitude from 500 to 1,000 m in the Central Depression, to 1,900–2,300 in the eastern Precordillera. The Altiplano, comprised between the Western and the Eastern cordilleras (Figs. 1, 2), consists of a high plateau with altitudes between 3,500 and 4,500 m that forms an up to 200-km-wide internally drained basin in a backarc position. In this region of the Andes, the Altiplano is mostly located in Bolivia, and only its westernmost part is located in Chile. The Eastern Cordillera is an uplifted block consisting mainly of Early Paleozoic sedimentary rocks bounded on both sides by thick-skinned divergent thrust systems, whereas the Subandean Zone consists of a thin-skinned, east-vergent fold-thrust belt facing the Andean foreland. The Interandean Zone corresponds to the transition zone between the thick and thin-skinned thrust systems developed in the Eastern Cordillera and the Subandean Zone, respectively (Kley 1996).

Stratigraphy in the Forearc Precordillera and Western Cordillera

Pre-Oligocene substratum

The basement rocks correspond to: the Belén Metamorphic Complex (BMC), remnants of Carboniferous-Permian and

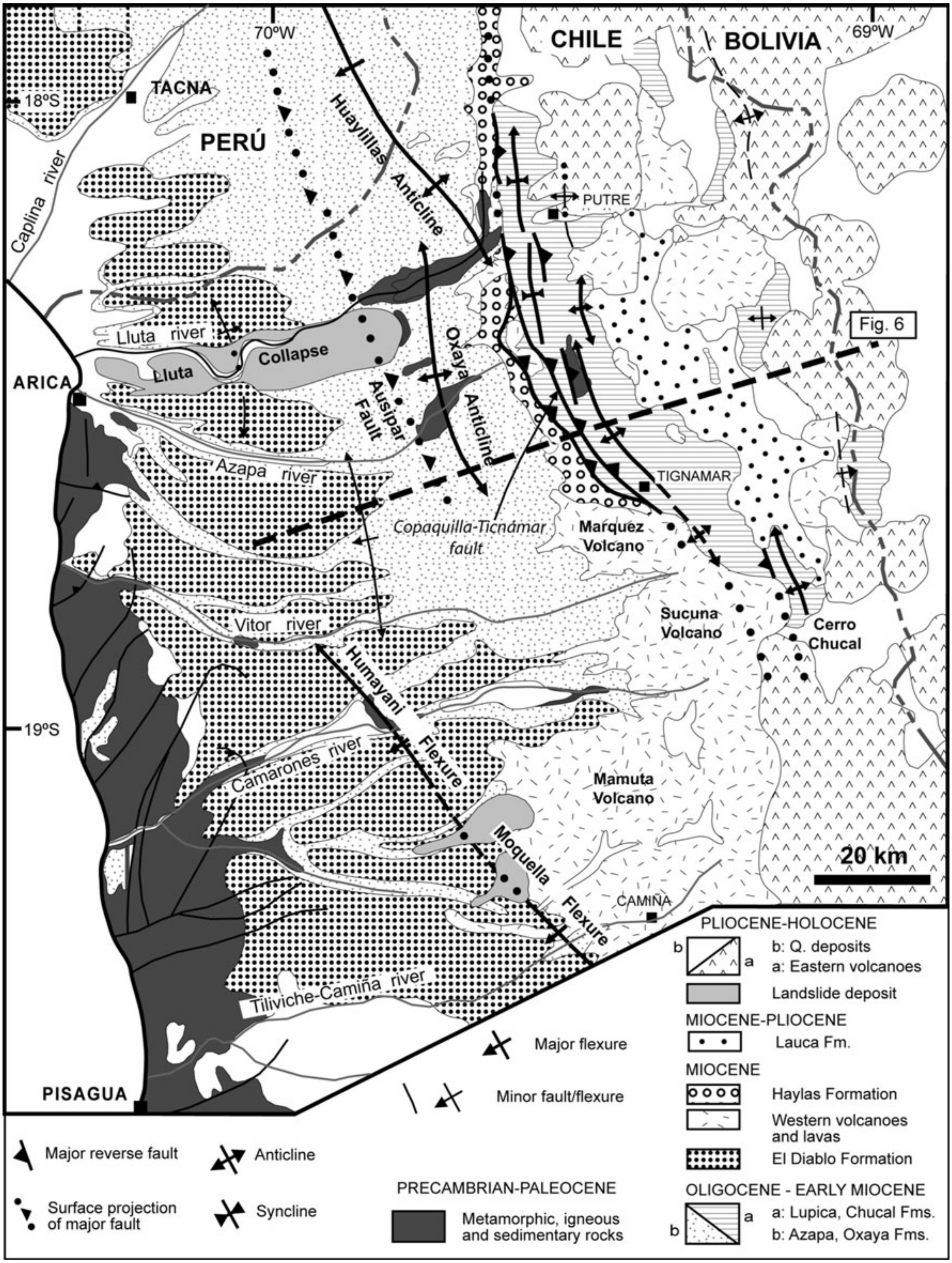

Fig. 4 Simplified geologic and structural outline of northernmost Chile, southernmost Perú and northwestern Bolivia in the southern part of the Bolivian Orocline, showing the main lito-stratigraphic units and the main structural features (based on Galli and Dingman 1962; García et al. 2004; Pinto et al. 2004a). *Thick dashed line* indicates location of balanced cross-section of the Precordillera in Fig. 6

Jurassic-Early Cretaceous sedimentary rocks, and Late Cretaceous-Paleocene intrusive rocks (Fig. 4). The BMC forms a narrow strip in the western border of the Western Cordillera and consists of amphibolites, quartz mica schists, gneissic schists and orthogneisses (Salas et al. 1966; Pacci et al. 1980; Lucassen et al. 1994, 1996; García et al. 2004; Wörner et al. 2000). The BMC has been dated as pre-Cambrian (1,000 Ma, Pacci et al. 1980), but more recent datings indicate a rather complicated thermal history that apparently has begun only in “middle” Cambrian (Basei et al. 1996) or Ordovician (Loewy et al. 2004) times. The BMC is unconformably covered by thin discontinuous remains of conglomerates, marine sandstones and dark siltstones in which Lezaun et al. (1996) reported the existence of brachiopods assigned to the Carboniferous-Permian? (Quichoco Beds; García et al. 2004). The BMC is also covered by Late Jurassic marine limestones and sandstones correlated with the western Livilcar Formation (Covacevich 1987; García et al. 2004). In the bottom of the Precordillera valleys, the substratum corresponds to the marine and continental sedimentary Jurassic to Early Cretaceous Livilcar Formation (Muñoz et al. 1988) intruded by Late Cretaceous-Paleocene granodiorites, monzodiorites, and monzonites (Muñoz and Charrier 1996; García et al. 2004).

Oligocene—present deposits

The Cenozoic successions vary considerably their thickness depending on their geographic location. In the Coastal Cordillera, they consist of gravel deposits onlapping Mesozoic rocks. In the Central Depression, the oldest deposits exposed correspond to the coarse detritic Azapa Formation, which is overlain by the ignimbrites of the Oxaya Formation and the sedimentary El Diablo Formation. These deposits accumulated in the Pampa del Tamarugal basin (Nester and Jordan 2012). In the eastern Precordillera, the Oxaya Formation is locally overlain by the detritic Huaylas Formation. In the Western Cordillera, the volcanic Lupica Formation is the age equivalent of the Oxaya Formation.

Azapa formation

In the Central Depression, only the upper 450–500 m of this formation is exposed and there is no evidence for its total thickness. These consist of thick, subhorizontal fluvial to alluvial conglomerates and sandstones (Salas et al. 1966; Vogel and Vila 1980; Tobar et al. 1968; Parraguez 1998; García 2002; García et al. 2004), which are visible only in the deeper parts of the valleys (Fig. 4). Grain size diminishes westwards. In the eastern part of the Central Depression, the Azapa Formation consists of clast-supported boulder to cobble conglomerates. To the west, it consists of red to brown pebbly conglomerates, sandstones and siltstones. Clasts consist of volcanic rocks (andesites and rhyolites), granodiorites, sandstones and metamorphic rocks (schists, gneisses, and amphybolites) indicating a source from all stratigraphic units that underlie the Cenozoic successions in the region. Provenance from the metamorphic basement has been also determined by the heavy mineral content (garnet-staurolite and actinolite-epidote assemblages) of finer-grained sediments; other, minor sources are limestones, felsic to intermediate volcanites, intrusive bodies (probably granodiorites), and contact metamorphic and hydrothermally altered rocks (Pinto et al. 2004a). Clast imbrication indicates a sedimentary transport from northeast and east. In the eastern part, more proximal facies are associated with coalescent fluvial fans, braided rivers, and torrential alluvial flows; further west, facies correspond to flood plains (Parraguez 1998; García et al. 2004).

Based on age determinations of the conformably overlying Oxaya Formation (late Oligocene-early Miocene; see below) and of the underlying intrusive rocks (Cretaceous-Paleocene), the Azapa Formation has been broadly assigned an Oligocene age (García et al. 2004). Cosmogenic ^{21}Ne analyses on quartz from the surface of similar deposits onlapping Mesozoic rocks in the Coastal Cordillera at $\sim$19°30′S, yield late Oligocene to Miocene exposure ages (Dunai et al. 2005). This formation can be followed to southern Peru, where it has been named Moquegua Superior Formation (Tosdal et al. 1984; Marocco et al. 1985). South of the study region these deposits are continuously exposed and have been thoroughly described by Pinto et al. (2004b), Victor et al. (2004), Farías et al. (2005). In this region, they are covered by the mainly coarse clastic and ignimbritic Altos de Pica Formation (Galli and Dingman 1962).

The upper portion of the Azapa Formation in the study region has been dated at 24.7 ± 0.3 Ma (García 2002; García et al. 2004) and there is no age constrain for its lower portion. Age determinations on its southern prolongation, between 19°30′S and 21°S, indicate that the formation is older than 26.0 ± 0.4 Ma (Farías et al. 2005) and that its lower levels have an estimated age of $\sim$27 and 29 Ma (Victor et al. 2004), suggesting that the Azapa Formation reaches, at least, the early Oligocene. Further evidence for this assumption is that thicknesses of 1,800 m have been reported for the southern part of the basin, between 20°S and 22°S, and that thicknesses of 1,200 m have been extrapolated to the study region based on seismic profiles (Nester and Jordan 2012). Such thicknesses suggest that in the study region the Azapa Formation could be thick enough to reach an early Oligocene age, as was calculated by García (2002), and even a late Eocene age.

Oxaya formation

It consists of an approximately 1,000-m-thick succession of dacitic to rhyolithic welded ignimbritic flows with minor detritic intercalations (Salas et al. 1966; Vogel and Vila

1980; Parraguez 1998; Wórner et al. 2000; García et al. 2004). Lithic components are of volcanic nature. The lower part consists of an alternation of tuffs, conglomerates and sandstones, and its upper part, of ignimbrites. To the west, in the Central Depression, this formation can be traced almost up to the Coastal Range, where its thickness reaches only 80–100 m. The age is constrained by about forty-five ^{39}K–^{40}Ar and ^{40}Ar/^{39}Ar radiometric determinations (Naranjo and Paskoff 1985; Parraguez 1998; Wörner et al. 2000; García et al. 2000, 2004) that yield values between 25.5 ± 0.6 and 16.9 ± 0.5 Ma. These ages are in good agreement with magnetostratigraphic data obtained by Parraguez (1998) and Tapia et al. (2000), which permit individualization of several flows in the Oxaya Formation emplaced in a time lapse of ~5 myr. The ignimbritic deposits derive from vents in the Western Cordillera from where they moved downslope to the west (Vogel and Vila 1980; García et al. 2000; Wörner et al. 2000). Similar deposits are known in southern Peru (Huaylillas Formation, Marocco et al. 1985). To the south, thinner ignimbritic intercalations of the same age form the Lataguala (Pinto et al. 2004b) and the Altos de Pica formations (Galli 1968; Galli and Dingman 1962; Lahsen 1982; Farías et al. 2005; Victor et al. 2004; Muñoz-Tolorza 2007).

Lupica formation

This unit is exposed in the Western Cordillera (Salas et al. 1966; García et al. 2004; Wörner et al. 2000; Charrier et al. 2005) (Fig. 4). Its thickness is ~1,500 m and has been subdivided into three facies associations (García et al. 2004). The lower association consists of a 300–600 m thick, well stratified and rather altered lava, volcaniclastic and sedimentary succession. The volcanic deposits consist of porphyric and andesitic breccias and tuff intercalations. The sedimentary deposits consist of sandstone, conglomeratic sandstones, tuffitic layers, sedimentary breccias with volcanic fragments, and ignimbritic tuffs. The 400–2,000-m-thick middle part of the formation consists of a succession of felsic ignimbrites and tuffs interbedded with andesitic lavas and fluvial-alluvial sandstones, conglomerates and breccias. Sandstone and tuffitic layers are well defined. The sandstones are polimictic and fragments are dominantly volcanic. The upper facies association consists of 400–600 m of siltstones, shales, marls, limestones and chert, with intercalations of conglomerate and volcanic deposits. Several ^{39}K–^{40}Ar, Ar–Ar, and U–Pb age determinations constrain the age of the Lupica Formation, with values between 25.7 ± 1.4 and 18.6 ± 0.6 Ma (Aguirre 1990; Muñoz and Charrier 1996; García et al. 2004; Wörner et al. 2000). According to these results, the Lupica Formation in this region can be assigned a late Oligocene to early Miocene age.

Joracane formation

It consists of a 500–600-m-thick, 25°–30° east-dipping conglomeratic succession with rare volcanic and silstone intercalations (García 1996; García et al. 2004). The conglomerates and sandstones consist essentially of andesitic and basalt-andesitic clasts. Calcareous clasts from the upper member of the Lupica Formation are also present; however, no clasts of the BMC have been found, which suggests that the basement was not exposed at that time. The conglomerates were deposited in a weakly erosive fluvial environment and are organized in layers of one to several meters thick. According to the imbrication as well as to the westward decrease in grain size these sediments were supplied from areas located to the east. A ^{39}K–^{40}Ar age determination on biotite from tuffs intercalated 150 m above the base of this unit yield a 18.2 ± 0.8 Ma age. According to this, it is assigned an early Miocene age.

Chucal formation

It is exposed on the east side of the Chapiquiña-Belén Ridge and is deformed by the Chucal anticline (Muñoz 1991; Riquelme 1998; García et al. 2004; Charrier et al. 2005) (Fig. 4, 5). The formation is separated from the Lupica Formation by an erosional and weathering surface. The sedimentary deposits of this unit are organized in several members, some of which cannot be traced from one limb of the Chucal anticline to the other. On the west side of the anticline, the formation is 600 m thick, and has been subdivided into four informal members. These deposits correspond to a continuous environmental evolution reflecting a fluctuating transition from fluvial to lacustrine conditions in the lower portion of the succession, and a return to fluvial conditions, accompanied by abundant supply of ash material in its upper portion. The deposits on the east-flank form a 365-m-thick, steeply east-dipping to vertical succession overlying the thinly stratified white–gray fluvial deposits of the upper part of the Lupica Formation, and unconformably overlain by the Macusa Formation (see bellow) (Fig. 5). This comparatively thin succession has been subdivided into three informal members, which are separated from each other by three progressive unconformities. The three stage subdivision on the east-flank of the Chucal anticline corresponds to a discontinuous environmental evolution reflecting a transition from lacustrine to high energy flood-plain conditions, however, with persistent lacustrine influence indicating continuous proximity of a lake within the depositional system, passing through lake filling, and flood-plain deposits with lacustrine influence.

Beginning of deposition of the Chucal Formation occurred in early Miocene times, after 21.7 ± 0.8 Ma,

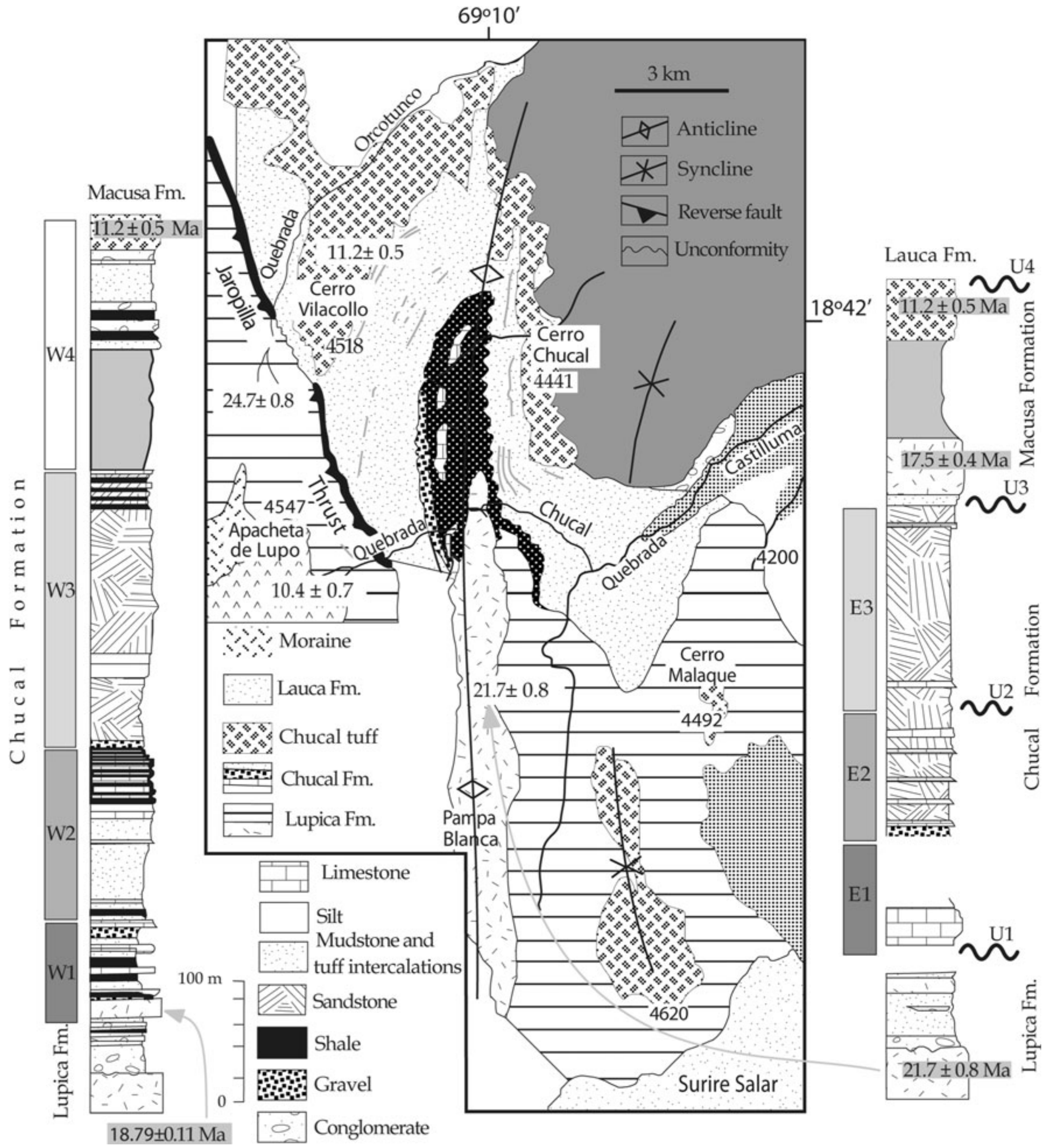

Fig. 5 Geologic map of the Chucal region in the Western Cordillera showing the east-vergent structural features in this region, and the stratigraphic columns of the west and east flanks of the ~15° northward plunging Chucal anticline, comprising the uppermost Lupica, the Chucal and Quebrada Macusa formations, and the basal Lauca Formation, indicating the stratigraphic position of progressive unconformities, and radioisotopically dated horizons. Unconformities indicated in the eastern stratigraphic column (U1–U4) correspond to progressive unconformities in the east-flank of the fold. Unconformities that separate the Chucal Formation from the underlying Lupica (21.7 ± 0.8 Ma) and from the overlying Quebrada Macusa (17.5 ± 0.4 ma) formations are separated by a 4.2 Ma time span, during which syntectonic deposition of the Chucal Formation took place

which is the age obtained in the rhyolitic tuff at the top of the Lupica Formation, and shortly before 18.79 ± 0.11 Ma, which is the age of an ignimbrite dated by Wörner et al. (2000) "at the base of the sedimentary Chucal Formation". Deposition lasted until shortly before 16–17 Ma (García et al. 2004; Charrier et al. 2005), which is the age of the lowest tuff of the Macusa Formation. These ages constrain the Chucal Formation to the early Miocene-middle Miocene. An abundant fossil mammal fauna has been recovered from several levels of the Chucal Formation (Charrier et al. 1994b; Bond and García 2002) consisting of several species of Notoungulates, Litopterns, Rodents and Glyptodonts (Flynn et al. 2002). According to the biochronologic ranges of the mammalian taxa, this fauna is restricted to the interval between ~20 and ~9 Ma (late early Miocene to middle Miocene) (Flynn and Swisher 1995), coinciding with the radioisotopic datings. Fossil leaves and pollen from calcareous levels in the lower portion of the west-side succession form an association that indicates the coexistence of forests and steppes developed in dry, temperate to cold climatic conditions (Charrier et al. 1994a), which is compatible with the presence of the abundant herbivore mammal fauna.

El Diablo formation

It is exposed in the Precordillera and Central Depression and can be traced up to the eastern border of the Coastal Cordillera (Tobar et al. 1968; Vogel and Vila 1980; Parraguez 1998; García et al. 2004) (Fig. 4). It overlies with erosional unconformity the Oxaya Formation, and consists of an almost undeformed, upward coarsening and thickening continental succession composed of conglomerates and sandstones, with intercalations of limestone, fine-grained sandstones and reworked tuffs. The 350 m maximum thickness in the study region is reached in the middle part of the Central Depression. These deposits can be subdivided into two members: (1) A lower member, consisting of quartz-rich sandstones, decimeter to several-meter thick layers of silicified white limestones, and brown and gray biotite rich calcareous sandstones, which locally contain imbricated pumice fragments and sporadic intercalations of reworked quartz-rich cineritic tuffs and siltstones. Silicified limestones contain rests of plants in life position. The components of the basal sandstones derive mainly from the ignimbritic Oxaya Formation. (2) An upper member consisting at its lower portion of gray to dark gray, medium to coarse sandstones with cross-bedding and dessication cracks, and at its upper portion of brown and black, poorly stratified, medium to badly selected conglomerates with pebbles that reach 2 m in diameter. The transition from the fine-grained to the coarse-grained deposits is gradual and occurs within 10–20 m. The thickness of the conglomeratic succession varies between 30 and <100 m.

The abundant andesitic and basaltic-andesite clasts give these deposits their characteristic dark color. The matrix of the conglomerates consists of coarse, poorly sorted sandstone. Imbrication indicates an eastern provenance. The 11.9 and 18.8 Ma age obtained in andesitic clasts (García et al. 2004) and the heavy mineral suite (clinopyroxenes, orthopyroxenes and olivine) in the fine-grained matrix indicates an important contribution of the middle Miocene volcanics at the time of deposition of the El Diablo Formation (Pinto et al. 2004a, 2007). These deposits accumulated in large coalescent fluvial-alluvial fans formed by debris and sporadic gravity flows in the proximal zones, and turbulent channelized flows and braided rivers in the medial to distal zones.

A ^{39}K–^{40}Ar age determination from a reworked fine-grained tuff at the uppermost part of the lower member of the El Diablo Formation yielded a 15.7 ± 0.7 Ma age (García et al. 2004). This age plus the abundance of clasts derived from the middle Miocene volcanics in the upper portion of the El Diablo Formation indicates that: (1) Deposition of the El Diablo Formation is coeval with the Zapahuira Formation (see below), as has been observed further south in the Precordillera by Farías et al. (2005), and (2) The middle Miocene volcanoes were undergoing intense erosion during deposition of the upper part of the El Diablo Formation.

Between the Suca and Camiña river valleys, the El Diablo Formation is covered by a basaltic-andesite lava flow (Mortimer et al. 1974; Pinto et al. 2004b). A sample of this flow yield a K–Ar age on whole rock of 8.2 ± 0.5 Ma (Muñoz and Sepúlveda 1992). Similar ages have been obtained by Mortimer et al. (1974) recalculated by Naranjo and Paskoff 1985, and García et al. 2004). In the Suca river valley, the El Diablo Formation overlies the 16.2 ± 0.7 Ma old Tarapacá Ignimbrite (Nama vitrophyre in Pinto et al. 2004b). According to the radiometric dates, the El Diablo Formation can be assigned a middle to late Miocene age. This formation can be followed further north into southern Perú, where it has been named Magollo Unit (Flores 2004).

Middle Miocene volcanics and Zapahuira formation

These rocks correspond to remains of andesitic and basaltic lava flows (García 1996; García et al. 2004) (Fig. 4). On the eastern Precordillera, these lavas conformably overlie the Oxaya Formation. Several Ar–Ar and ^{39}K–^{40}Ar ages from different stratigraphic levels constrain its age between 17.0 ± 0.6 and 11.0 ± 0.4 Ma (García et al. 2004; Wörner et al. 2000).

Macusa formation

The Macusa Formation forms an approximately 200-m-thick series, of light-colored, massive ignimbritic deposits, tuffaceous sandstones, and conglomeratic horizons (Riquelme 1998; García et al. 2004). The contact with the underlying Chucal Formation is conformable on the western flank of the Chucal anticline, whereas it is unconformable on the eastern flank. The abundance of ash deposits in the Macusa Formation indicates an increase in explosive volcanism after deposition of the Chucal Formation. Riquelme (1998) observed that on the eastern side the strata of the Macusa Formation gradually decrease their steepness with decreasing age suggesting syntectonic deposition. Three ^{40}Ar/^{39}Ar age determinations from the basal, white tuff of the Macusa Formation yield the following ages: 16 ± 3, 17.5 ± 0.4 Ma and 15.3 ± 0.2 Ma, of which the first is considered analytically more reliable (García et al. 2004). Four K–Ar dates on whole rock constrain the age of its upper part: 11.2 ± 0.5 and 11.4 ± 0.7 Ma for the brown Chucal Ignimbrite, and 10.4 and 10.3 Ma for the upper andesitic lavas (García et al. 2004) (Fig. 5).

Huaylas formation

This formation consists of an up to 350 m thick, almost flat-lying conglomeratic succession, with sandstone, limestone and tuffaceous intercalations. It is exposed in the Precordillera in a N–S-oriented depression formed by anticline growth on the east side of the Oxaya anticline (Salas et al. 1966; García et al. 2004) (Figs. 4, 6). The fine-grained basal deposits conformably overlie the east-dipping Zapahuira Formation (east limb of the Oxaya anticline), whereas the coarser upper levels lie almost horizontal and onlap westwards on the older Oxaya Formation.

The clasts of the upper cobble and pebbly conglomerates consist of fragments eroded from all pre-existing units exposed in the Western Cordillera, including fragments from the BMC. The heavy mineral suite in the sandy matrix of the Huaylas Formation is similar to that of the older Azapa Formation in the Central Depression indicating that during the time interval between deposition of these two formations the same substratum was exposed (Pinto 2003; Pinto et al. 2007) and that during all this time erosion in the eastern Precordillera was deep enough to reach the metamorphic basement. Stratification is irregular and paleochannels are not deeply cut in the underlying deposits. Coarseness of conglomerates increases upward both stratigraphically and to the east, and the uppermost levels of the eastern outcrops contain blocks of up to 1.5 m in size, indicating the eastern provenance for these deposits. Considering that: (1) The tuff levels intercalated in the upper Huaylas Formation yield several K–Ar and Ar–Ar radioisotopic ages of 9.9–11.4 Ma (García et al. 2004; García and Hérail 2005), and (2) The finer-grained lower deposits of the Huaylas Formation contain mammal

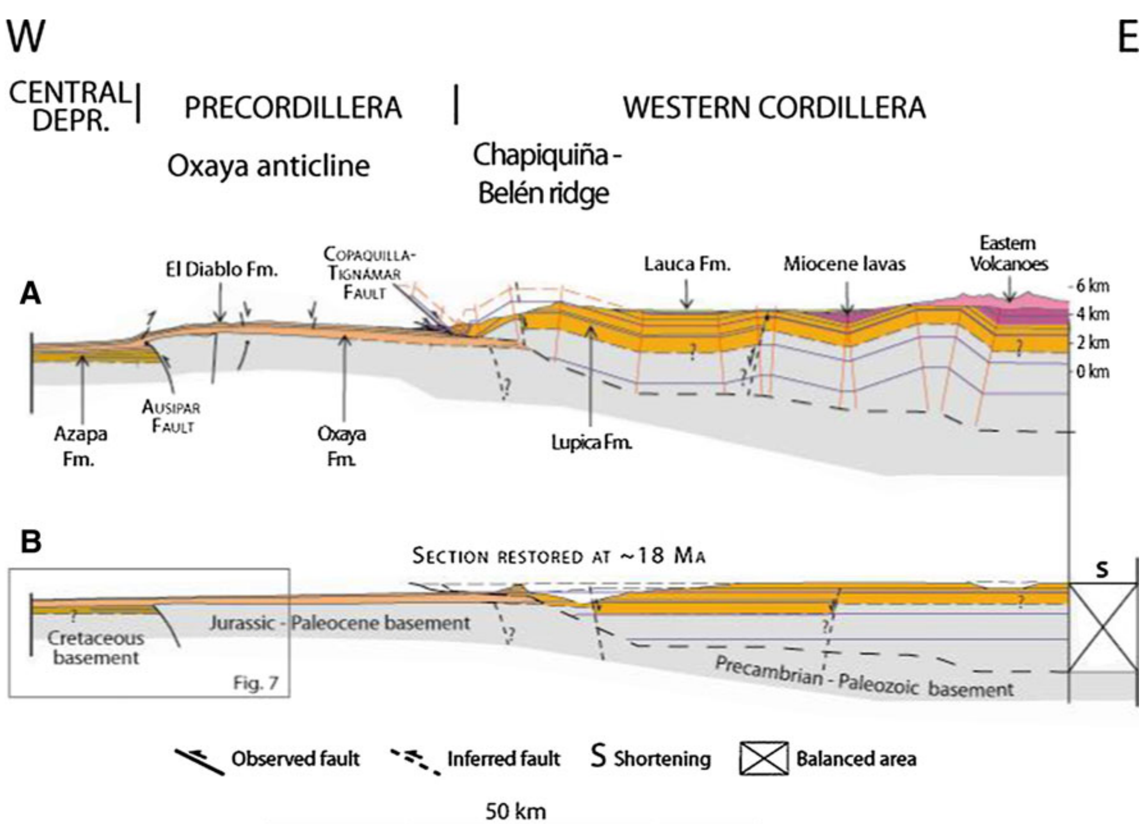

Fig. 6 **a** Present-day cross-section of the Precordillera and Western Cordillera, at 18°30′S, with 30 % vertical exaggeration, and **b** restoration at ~18 Ma (pre-second movement on the Ausipar fault), modified from García (2002). Figure shows the steep west-vergent Ausipar fault contrasting with the flatter Copaquilla—Tignámar detachment. The Chapiquiña—Belén ridge in this section is underlain by the fault-bend fold developed by the Copaquilla—Tignámar fault (Belén anticline). *Note* the small shortening (less >7 km) in this region

fossils (Nothoungulates) (Bargo and Reguero 1989; Salinas et al. 1991; Flynn et al. 2005) of a post-Friasian-pre-Huayquerian SALMA age (Middle to early Late Miocene) (Flynn et al. 2005), deposition of the Huaylas Formation is constrained to the middle to late Miocene.

Late Miocene–Holocene volcanics

A series of andesitic and dacitic volcanoes and relicts of eroded volcanos are exposed in the Western Cordillera (García et al. 2004). Abundant K–Ar and Ar–Ar determinations constrain their ages to this time span.

Lauca formation

This formation, informally defined by Muñoz (1988), fills an elongated, NNW–SSE-oriented lacustrine basin ($\sim$3,000 km^2) on the western Altiplano (Kött et al. 1995; Muñoz and Charrier 1996; García et al. 2004). On the east side of the Chucal anticline, the slightly east-dipping basal deposits of the Lauca Formation overlay with slight angular unconformity the steeper, upper tuff of the Macusa Formation, while further east, the Lauca deposits are flat-lying (Kött et al. 1995) suggesting syntectonic deposition on the east side of the anticline. The basal deposits of this formation consist of a coarse conglomerate, with boulders of ignimbrite, andesite, and basaltic-andesite. The Lauca Formation consists of an up to 300-m-thick sedimentary series (Muñoz and Charrier 1996), of which 120 m are fully lacustrine (Kött et al. 1995). The lacustrine deposits consist of fine-grained sandstones, and mudstones, frequently tuffaceous, and sporadically calcareous, with lesser amounts of evaporitic deposits, containing a few conglomeratic sandstone intercalations. These deposits contain ostracodes and diatoms (Kött et al. 1995). Lavas of the Guallatire and Nevados de Payachatas volcanoes (0.5–0.3 Ma) overlie young deposits of the Lauca Formation (Wörner et al. 1988; Kött et al. 1995). Geochronologic dates in the deposits underlying and overlying the Lauca Formation constrain its age between $\sim$10 and $\sim$0.5 Ma (late Miocene to Pleistocene).

Lauca ignimbrite

This is a conspicuous, rhyolitic, up to 150-m-thick ignimbrite intercalated in the middle part of the Lauca Formation in the Western Cordillera that extends westward to the Precordillera and Central Depression. The late Pliocene age for this deposit has been well constrained with several ^{39}K–^{40}Ar and ^{40}Ar/^{39}Ar age determinations with a mean age between 2.7 and 2.9 Ma (Kött et al. 1995; Muñoz and Charrier 1996; Wörner et al. 2000; García et al. 2004). The ignimbrite has been correlated with the Pérez ignimbrite (Evernden et al. 1977; Lavenu et al. 1989; Marshall et al. 1992) well exposed in the Bolivian Altiplano. Based on its age and paleomagnetic characteristics (Roperch et al. 2006), this ignimbritic flow is now considered to be the same flow as the Pérez ignimbrite in the Bolivian side. The Lauca (or Lauca-Pérez) ignimbrite is correlative with the Pachía ignimbrite in southern Peru (Flores 2004).

Tectonic evolution on the Precordillera and Western Cordillera

Direct evidence for the middle Eocene (Incaic) deformation has been reported for areas south of the study region, mostly in the Domeyko range (Charrier et al. 2007, 2009). Eocene compression in this region was apparently responsible for exhumation of the metamorphic basement prior to deposition of the Lupica and Oxaya formations, as indicated by the heavy mineral content in the Azapa Formation. In this region, two, thrust systems with opposite vergencies concentrated the Miocene to Holocene compressive activity (Muñoz and Charrier 1996; García et al. 2004; García and Hérail 2005). The tectonic style in northern Chile, west of the Incaic axis, is thick-skinned.

Precordillera

The Precordillera, east of Arica, consists of a major asymmetric west-vergent fold, the Oxaya anticline, which is bounded to the west by the surface projection of the Ausipar fault (Muñoz and Charrier 1996; García et al. 2004; García and Hérail 2005) (Figs. 4, 6). This is mostly a blind fault, exposed only in the deeper parts of the Lluta and Azapa valleys, with an eastward dip of 60°E, and the Oxaya anticline formed during the fault activity. Shortening along the Oxaya anticline is somewhat less than 100 m and its maximum amplitude is 850 m. The 850 m amplitude is considered as the structural relief formed during folding. The small shortening together with the rather high uplift of the crest of the anticline suggests that the dip of the Ausipar fault is subvertical at depth.

The hangingwall of the Ausipar fault consists of the Jurassic-Early Cretaceous Livilcar Formation unconformably overlain by the upper Azapa Formation and the Oxaya Formation. The footwall consists of the lower and middle members of the Azapa Formation conformably overlain by distal portions of the Oxaya Formation. The absence of the lower to middle Azapa Formation in the hangingwall and the presence of the Oxaya Formation on both sides of the fault indicate the existence of two tectonic episodes associated with this structure (Muñoz and Charrier 1996; Parraguez 1998; García et al. 2004; García and Hérail 2005) (Figs. 6, 7). Movement along the fault can, therefore, be separated into a first pre- or syn-Azapa episode (in any case pre-Oxaya), and a second, post-Oxaya episode. Because of the broad Oligocene age assigned to the Azapa Formation and the possibility that sedimentation of this formation began in late Eocene times, the first tectonic activity along the Ausipar thrust fault can be related to: (1) The late Oligocene–early Miocene tectonic event, and/or (2) The mid-Eocene Incaic phase. Development of the Oxaya anticline is related to the second episode. Based on the ages constraining the growth strata (Huaylas Formation) on the east limb of the Oxaya anticline, this second episode on the Ausipar fault occurred between 11.7 ± 0.7 and 10.7 ± 03 Ma, in middle to late Miocene (García et al. 2004; García and Hérail 2005). The Oxaya anticline is continued southward by the Sucuma homocline (Fig. 4).

Western cordillera

Early Miocene explosive volcanic event

In the western border of the Western Cordillera, near the headwaters of the Belén River, lavas and breccias of the lower member of the Lupica Formation are in contact with the BMC on the vertical Nacientes de Belén fault (García et al. 2004; Lezaun et al. 1996). The Lupica deposits contain fragments of the BMC, some of which are 1 m long. The basal andesitic layers of the Lupica Formation immediately west of the fault dip ~60°E and are overlain by gradually less steep andesitic layers, which cover the trace of the fault to the east. This geometry suggests the existence of growth strata associated with a normal fault close to the Oligocene–Miocene boundary. Such feature is probably related to the development of a caldera associated with the explosive volcanism that originated the Oxaya and Lupica formations (García et al. 2000).

Miocene–Holocene compression

In the western part of the Western Cordillera, at the latitude of Arica, the main topographic feature is the Chapiquiña-Belén ridge (Fig. 6), which exceeds in about 4,000 m the average altitude of the Central Depression and in 800 m the altitude of the Lauca basin, in the western Altiplano. This ridge forms the present-day water divide. The N–S trending Chapiquiña-Belén ridge has a pop-up geometry resulting from the activity of the two thrust systems with opposite vergencies. Deformation and uplift on these thrust systems caused development of relief and controlled sediment supply to the adjacent Pampa del Tamarugal and the Altiplano basins.

West side of the Chapiquiña-Belén ridge Deformation in this side of the ridge is west-vergent. Asymmetric anticlines and synclines in the Lupica Formation are relatively tight and thrusted to the west along an east-dipping thrust fault system: Socoroma, Copaquilla-Tignámar, and Belén-Tignámar faults (Fig. 6). The Chapiquiña-Belén ridge consists of a major fault-bend fold anticline (Belén anticline) of 15–20-km half-wavelength that exposes basement blocks in its core; its western flank is partially eroded and cut by thrusts, and its eastern flank is covered by 16–15 Ma old lavas showing a progressive unconformity, and the Pliocene–Pleistocene Lauca Formation (Fig. 6). The western thrusts are associated with the development of syntectonic coarse conglomerates in the Joracane and Huaylas formations. According to balanced cross-sections, the estimated post-18 Ma shortening is less than 7 km (García 2002), which is associated with a surface uplift of ~1,000 m in the Chapiquiña-Belén ridge (Fig. 6). According to the age of the syntectonic Joracane Formation, activity in the eastern Precordillera began shortly

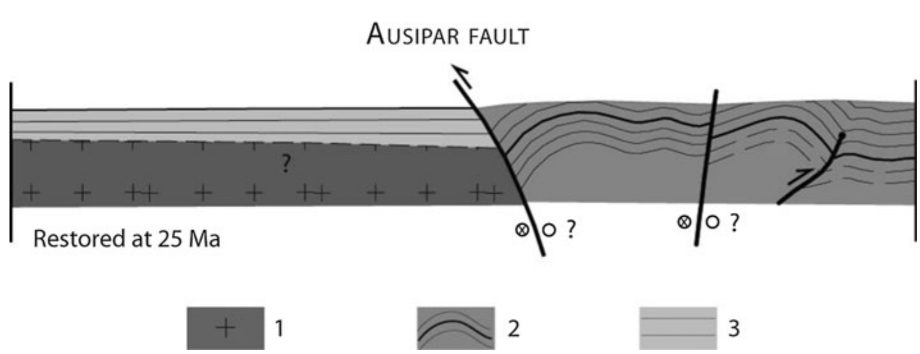

Fig. 7 Restoration of the pre-Oxaya situation on the Ausipar fault at 25 Ma. For location see rectangle in Fig. 6b. *1* Cretaceous plutonic basement, *2* Folded Jurassic—Late Cretaceous Livilcar Formation, *3* Oligocene Azapa Formation. The first activity on the Ausipar fault caused westward thrusting of the east-side block (hanging-wall) on the Azapa Formation and the Cretaceous plutonic basement in the west-side block (foot-wall). In early Miocene, the Oxaya Formation covered the fault and the two blocks, east and west of the fault. In late Miocene, renewed activity on the Ausipar fault (second movement) partially cut through the deposits of the Oxaya Formation (see Fig. 6a)

before 18.2 ± 0.8 Ma and ceased before deposition of the middle-late Miocene Huaylas Formation, which is related to the activity of the youngest fault in this region (Copaquilla-Tignámar fault). A somewhat later activity on the Copaquilla-Tignámar fault caused a vertical displacement of 100–150 m of the Lauca ignimbrite, indicating tectonic activity after 2.7–2.9 Ma (Fig. 4).

East side of the Chapiquiña-Belén ridge Deformation on this side of the ridge forms an east-vergent thrust system (ETS) and consists in the Chucal region of the steep-dipping Jaropilla thrust fault, to the west, and several folds including the major northward plunging Chucal anticline, to the east (Fig. 5). Tectonic activity in this region apparently concentrated first on the Jaropilla fault and later on the blind fault that caused development of the Chucal anticline and accommodation spaces on both sides of the anticline axis. The eastern flank of the Chucal anticline dips 40–80°E and the western flank dips only 10–30°W. Deformation in this region caused a total shortening of ~6 km (García 2002). The deposits of the Chucal Formation are exposed on both flanks of the anticline, but only the western succession shows no interruption. The succession on the east-flank is interrupted by several progressive unconformities indicating a rather continuous east-vergent deformation event. This fold affects the Chucal Formation, the somewhat less deformed Macusa Formation, and on the eastern flank of the anticline, the Lauca Formation that unconformably overlies the Macusa Formation (Fig. 5). Compressional deformation began between 21.7 ± 0.8 (Lupica Fm. white tuff) and 18.79 ± 0.11 Ma (ignimbrite "at the base of the sedimentary Chucal Formation"; Wörner et al. 2000). The east-vergent thrust system has been followed southward up to ~20°S.

Cenozoic stratigraphy, and tectonic and paleogeographic evolution in Bolivia

Based on available literature on the Altiplano, Eastern Cordillera, and Subandean Zones, we describe next the evolution east of the Incaic axis, in order to obtain a general picture of the evolution across the Andean range in this region (Fig. 8).

Mid-Paleocene to early Eocene

Reduced sedimentation has been reported for the Altiplano during mid-Paleocene to mid-Eocene times (Horton et al. 2001; McQuarrie et al. 2005). Sedimentation in late Paleocene (upper Puca Group; Sempere et al. 1997) resulted in the mainly fine grained, distal fluvial, locally lacustrine, with calcareous paleosol intercalations Santa Lucía Formation (Rochat et al. 1998; Horton et al. 2001; Rochat 2002). Paleocurrent measurements in the upper, sandstone-dominated portion of the Santa Lucía Formation or Cayara Formation (Marocco et al. 1987; Sempere et al. 1997; Rochat 2002) indicate sediment supply from the east (Horton et al. 2001) and the northeast (Rochat 2002) (Fig. 8). The lower part of the upper Santa Lucía Formation has been assigned by magnetostratigraphy a late Paleocene age of ~58 Ma (Sempere et al. 1997).

No major interruption of sedimentation has been recorded between the upper Santa Lucía deposits and the overlying 3,000 a 6,500-m-thick succession of clastic deposits of the Potoco Formation (Horton et al. 2001; Jiménez et al. 2009) and equivalent stratified units in the Altiplano. However, the sharp contact separating the two formations marks an abrupt change characterized by fine to very fine-grained sandstones and siltstones, and a rather continuous development of paleosols in the lower portion of the Potoco Formation (Horton et al. 2001; Rochat 2002). Despite the lack of age determinations along most of the Potoco Formation, it is possible to assign a late Paleocene to early-mid-Eocene age to its lower portion. The described features indicate extreme low rates of subsidence at this moment (Horton et al. 2001; McQuarrie et al. 2005).

The NNE–SSW-oriented and 90–100-km-wide Late Cretaceous to early Eocene magmatic arc was located mainly along the Precordillera (Charrier et al. 2007) (Fig. 3). The region east of the arc corresponded to the foreland of the mountain range, and consequently the depocenter located immediately east of the arc, in the present-day Altiplano, corresponded to the retroarc foreland basin at that time. Development of east-vergent compressional structures, like those in the Chucal region in Chile, on the rear side of the arc was probably still reduced and subsidence in the foreland basin was caused mainly by the load of the arc. Additionally, subsidence was probably enhanced by superimposed thermal subsidence associated with the late stages of the Salta Rift.

Middle Eocene

The fine grained and paleosol-dominated lower portion of the Potoco Formation is followed by a thick succession of sandstones and intercalated mudstones that form the bulk of the formation (main Potoco interval). The deposits in the Altiplano show at this moment a clear west to east transition from proximal to distal facies and contain paleocurrent data indicating a west to east sediment transport (Sempere et al. 1990; Rochat et al. 1998; Horton et al. 2001; Rochat 2002) (Fig. 8). The reorganization of the sedimentation pattern and the western provenance of the sediments indicate a major paleogeographic modification that we

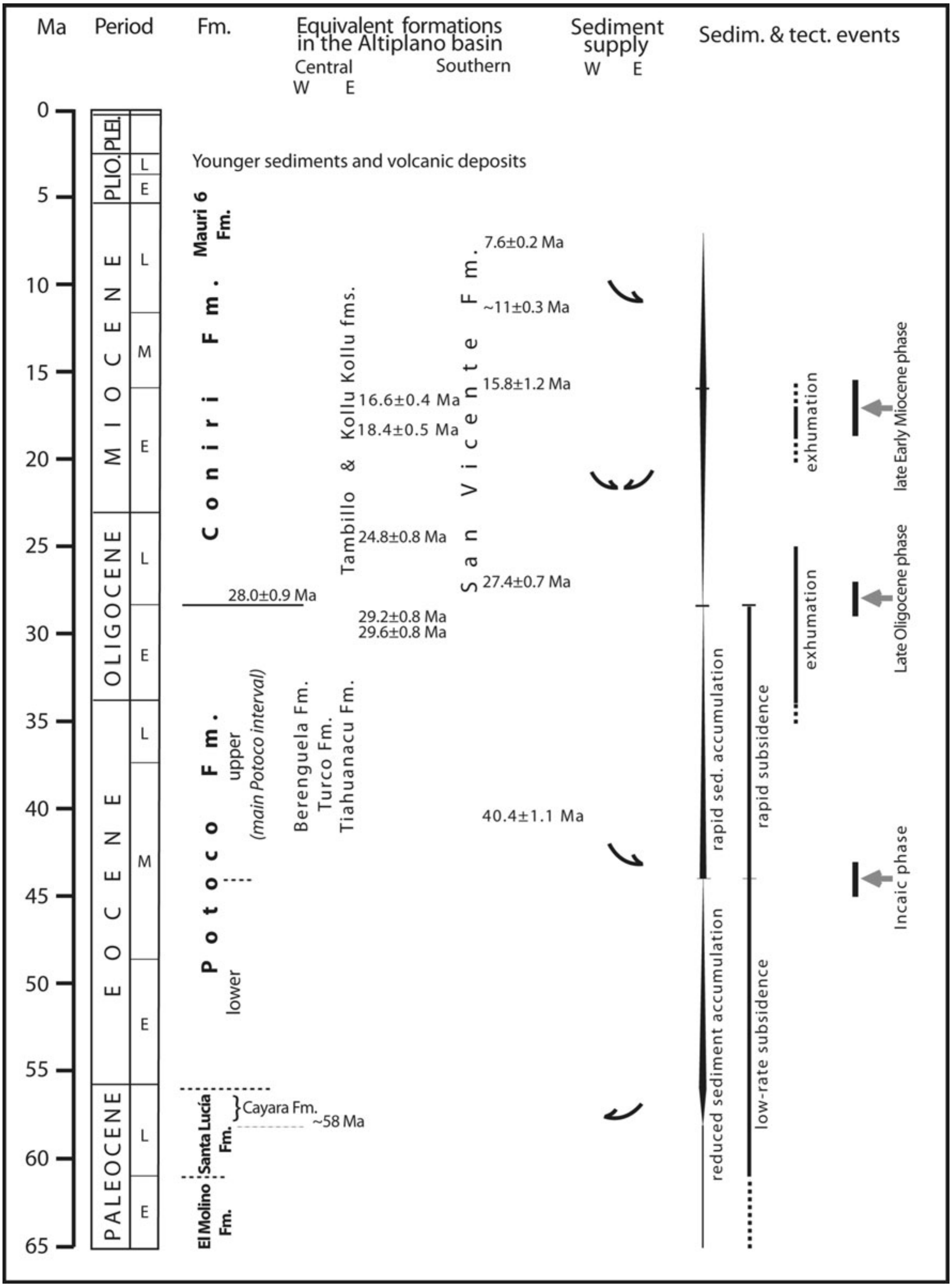

Fig. 8 Sketch for the Late Paleogene to Neogene chrono-stratigraphic evolution in the Altiplano Basin in western central Bolivia described in text. Major tectonic events deduced from the sedimentary and paleogeographic evolution are indicated on the *right side*: Incaic Phase, late Oligocene and late Early Miocene peaks of increased de formation

associate with uplift of the Incaic range (Maksaev 1978; Cornejo et al. 2003; Reutter 2001; Charrier et al. 2007, 2009), which would correspond to the "western orogenic source" suggested by Horton et al. (2001) for the deposits of the main Potoco interval.

Late Eocene to early Oligocene

Most of the main Potoco interval was deposited during this time (Horton et al. 2001) (Fig. 8). Rapid sediment accumulation characterizes these deposits (Horton et al. 2001). Sedimentary units equivalent to the main Potoco interval correspond, among others, to the following formations: Berenguela, to the west, Turco, further east, and Tiahuanacu, to the east (Suárez and Díaz 1996; Hérail et al. 1997; Rochat et al. 1998; Rochat 2002). Two K–Ar age determinations (29.2 ± 0.8 Ma and 29.6 ± 0.8 Ma) from tuffs intercalated in the upper Tiahuanacu Formation indicate an early Oligocene age for this part of the unit (Swanson et al. 1987). Palynomorph assemblages from the upper Potoco Formation also indicate an early Oligocene age (Horton et al. 2001). In the western southern Altiplano, a somewhat older age of 40.4 ± 1.1 Ma has been obtained on a lower tuff layer (K–Ar on biotite) interbedded in the Potoco Formation (Silva 2004). According to these ages, the main Potoco interval was deposited in late Eocene and early Oligocene. Sediment supply in all of them came from the west (Sempere et al. 1990; Rochat et al. 1998; Horton et al. 2001; Rochat 2002; Pinto et al. 2004a; Silva 2004) suggesting existence of an uplifted region to the west of the Altiplano basin.

Evidence for generalized uplift in the Bolivian Andes associated with the Incaic phase is provided by low temperature thermochronological analyses. Between ~15°S and 19.5°S, analyses yield mid-Eocene–Oligocene initial exhumation ages for the eastern Altiplano margin, and the Eastern Cordillera back and fore-thrust belts (~40–25 Ma and 36–27 Ma, respectively) (Barnes et al. 2006, 2008). In the southern Bolivian Altiplano, apatite fission-track cooling ages support an important exhumation event in late Eocene to Oligocene (Ege 2005): (1) Ages of 32–25 Ma have been obtained on basement clasts in the basal conglomerate in the late Oligocene San Vicente Formation, (2) a 33.1 ± 5.2 Ma age has been obtained on a resistant population of apatite crystals from a sample collected in the lower portion of the Potoco Formation, and (3) 34–30 Ma ages have been obtained in samples of Paleozoic rocks exposed in the hangingwall of two east-vergent thrust associated with the Uyuni-Khenayani fault zone. The oldest ages mentioned above coincide with cooling ages of 43.9 ± 4.4 Ma and 34.6 ± 4.2 Ma obtained by Carrapa et al. (2006) further south on the western flank of the southern Puna in the Fiambalá region, in Argentina, between 27° and 28°S. These ages indicate that the late Eocene to Oligocene exhumation event occurred coevally in a wide region comprising the Altiplano-Puna and Eastern Cordillera regions.

Uplift of the Incaic range greatly increased sediment supply from the west, which resulted in the thick upper Potoco Formation (main Potoco interval). Accordingly, rapid subsidence would have occurred in the retroarc foreland basin immediately east of the Incaic range (Hérail et al. 1997; Rochat et al. 1998; Horton et al. 2001; Rochat 2002; McQuarrie et al. 2005). Relative to its pre-middle Eocene location (Santa Lucía and lower Potoco formations), the basin axis and the forebulge were now located further east (Horton and DeCelles 2001; McQuarrie et al. 2005).

Although uplift of the Incaic range and eastward shift of the adjacent basin are evident, the structures associated with these processes have not yet been clearly identified in the Altiplano region and may be concealed underneath the regionally extended volcanic and sedimentary Cenozoic cover.

Late Oligocene

Approximately at this time, tectonic activity began in the west-vergent Huarina and SanVicente thrust belts (central Andean back-thrust belt) on the western margin of the present-day Eastern Cordillera. This activity caused deposition of coarse-grained deposits, development of growth strata, and progressive unconformities next to the contact between the Potoco and the overlying syntectonic deposits ("Oligocene conglomerate units" in the northwestern Altiplano, the basal conglomerates of the Coniri Formation, in the central Altiplano, and the San Vicente Formation, in the southern Altiplano), and the unconformable contact of these deposits over older units (Hérail et al. 1997; Horton et al. 2001; Mertmann et al. 2001; Rochat 2002; Elger 2003; Silva 2004; Ege 2005; Leier et al. 2010). Several K–Ar age determinations and fission-track dates have been obtained from tuff layers at the base of the Coniri Formation and equivalent deposits that yield ages between 28.0 ± 0.9 and 23.0 ± 0.8 Ma (Swanson et al. 1987; MacFadden et al. 1985; Sempere et al. 1990; Kennan et al. 1995; Horton et al. 2001). Recent U–Pb provenance analyses on detrital-zircon yield for the "Oligocene conglomerate units" in the Salla region in the northwestern Altiplano a 28.3 ± 0.6 Ma maximum depositional age that confirms the K–Ar ages referred above (Leier et al. 2010).

Similarly, in the southern Altiplano a K–Ar age of 27.4 ± 0.7 Ma (K–Ar in biotite; Mertmann et al. 2001; Silva 2004) has been obtained in a tuff layer interbedded in the lower conglomeratic deposits of the San Vicente Formation (Fig. 8). In the northern Altiplano, along the western margin of the Eastern Cordillera, where a similar tectonic and sedimentary situation has been recognized, $^{40}Ar/^{39}Ar$ age determinations in the syntectonic Aranjuez and Peñas formations indicate a coincident depositional age range from ~27.5 to ~25 Ma (Murray et al. 2010).

The westernmost equivalents of the Coniri and San Vicente formations, the Azurita and Tambillo formations (Hérail et al. 1997; Rochat 2002) contain eastern and western derived components, indicating existence of reliefs on both sides of the basin. Among the latter, abundant clasts of red granites and garnet-gneiss of pre-Cambrian age (Tosdal et al. 1984; Lamb and Hoke 1997) derive from rocks presently exposed only in the western Altiplano (Troeng et al. 1994; Hérail et al. 1997; Rochat et al. 1998; Rochat 2002; Pinto 2003).

According to Sempere et al. (1990), this late Oligocene tectonic activity took place at ~26 Ma. The 29.2 ± 0.8 and 29.6 ± 0.8 Ma old dates reported by Swanson et al. (1987) from tuffs intercalated in the upper Tiahuanacu Formation, an equivalent of the upper Potoco Formation (Fig. 8), and the dates mentioned above for the base of the Coniri Formation and the detrital-zircons in the Salla area indicate that this event is slightly older, and that it occurred by 28 Ma, in the late Oligocene. Thermochronological dates in both sides of the Eastern Cordillera indicate that the exhumation initiated in this region in mid-late Eocene–early Oligocene times continued until late Oligocene to early Miocene (25–19 Ma) (Barnes et al. 2008).

In the southern Altiplano, the San Vicente Formation contains ignimbritic, basaltic and andesitic intercalations (Fornari et al. 1993; Martínez et al. 1994; Mertmann et al. 2001) that decrease their thickness eastward (Elger et al. 2005; Silva 2004) indicating a provenance from an arc located to the west. These deposits have been considered equivalent to the Tambillos and Rondal lavas, the latter exposed in the Lípez region, in Bolivia, with K–Ar ages between ~23 and 18 Ma (Kussmaul et al. 1975; Fornari et al. 1989, 1993).

Although it has been presumed that normal faults have been active during deposition of the Potoco Formation and the lower San Vicente Formation (Silva 2004; Ege 2005), our view is that upon this moment the Altiplano basin was under compression and bounded by two thrust systems facing each other with uplifted ridges on their hanging-walls: (1) An east-vergent system, to the West, in the present-day Western Cordillera and western Altiplano, and (2) A west-vergent thrust system, to the East, in the present-day western Eastern Cordillera. Activity on the western thrust system caused additional uplift of the Incaic range, which would have been considerably eroded since mid-Eocene times. Activity on the two thrust systems determined that the Altiplano region became an internally drained basin (Sempere et al. 1990; Hérail et al. 1997; Rochat 2002).

The west-vergent thrust system on the west side of the Eastern Cordillera and the east-vergent fold-thrust system on the east side of the present-day central Eastern Cordillera that formed the Andean tectonic front at that time caused the development of an incipient, pop-up shaped Eastern Cordillera and triggered development of backbulge deposits further east. Erosion on the incipient Eastern Cordillera shaped a low relief, erosional surface on top of the Eastern Cordillera, the Challanta Surface, lying today at more than 4,000 m a.s.l. (Servant et al. 1989) (Fig. 8). Although, according to McQuarrie and DeCelles (2001) and McQuarrie et al. (2005), the west-vergent thrust system in the western Eastern Cordillera might have begun to develop in late Eocene, it is only in late Oligocene times that the fault activity had a noticeable effect on the sedimentation pattern. This would be so, unless older deposits have been eliminated (cannibalized) by erosion during early uplift of the western Eastern Cordillera (see Leier et al. 2010).

According to the mentioned authors, this west-vergent thrust system is the result of a basement megathrust of 10–12-km slip propagating over a bend (ramp to flat transition) in a crustal-scale, east-vergent detachment. It is interesting to point out, although their origin are probably not directly related, that development in late Oligocene times of the Challanta Surface on top of the incipient Eastern Cordillera roughly coincides with development of the Choja pediplane on the western flank of the Incaic range (Galli 1957; Farías et al. 2005), which is the southern prolongation of the Peruvian Altos de Camilaca Surface of Tosdal et al. (1984).

In northernmost Argentina, the late Oligocene compressive event characterizing this period caused eastward shift of the thrust front and the forebulge. At this moment began deposition of the Angastaco Formation that unconformably overlies older units and contains clasts from the Puncoviscana Formation also indicating sediment supply from the west (Galli and Hernández 1999).

Early Miocene

In Early Miocene times, sedimentation in the Altiplano continued without interruption. In fact, the deposits at this moment correspond to the upper portions of the Coniri and San Vicente formations and stratified equivalents (Fig. 8).

On the eastern side of the central Altiplano, the upper portion of the Coniri Formation and its western equivalent, the Kollu Kollu Formation, form an upward coarsening and westward fining conglomeratic and sandstone succession with tuff intercalations, some of which have regional distribution (Sempere et al. 1990; Horton et al. 2001; Rochat 2002; McQuarrie et al. 2005). K–Ar age determinations in the Kollu Kollu Formation yield early Miocene ages of

18.4 ± 0.5 and 16.6 ± 0.4 Ma (Swanson et al. 1987). Upward coarsening of the deposits indicates a continuous and increased deformation in the west-vergent thrust system in the western Eastern Cordillera and associated uplift of the ridge bordering the Altiplano to the east. Further west, no regional unconformities and an almost complete absence of growth strata has been detected by this time (Rochat 2002; McQuarrie et al. 2005).

Tectonic activity in the west-vergent thrust system ceased in the early Miocene (~19 Ma, Sempere et al. 1990; ~20 Ma, Horton 1998, and McQuarrie et al. 2005). However, east-vergent activity east of the Eastern Cordillera continued between ~18 and ~13 Ma (Horton 1998; McQuarrie et al. 2005) shifting further east the orogenic wedge. The associated foredeep was located by this time in the present-day Subandean Zone (McQuarrie et al. 2005). The tectonic activity on both sides of the proto-Eastern Cordillera and the continuation of activity on the east-vergent thrust system between ~18 and ~13 Ma, once the activity on the west-vergent thrust system ceased, indicates that eastward compression on the orogen did not cease after the late Oligocene event.

Thermochronologic datings indicate that after the late Oligocene–early Miocene (25–19 Ma) exhumation event in the Eastern Cordillera (~19.5°S), uplift shifted eastward to the Interandean zone in early Miocene times (22–19 Ma) and was followed by a second exhumation pulse in middle Miocene (16–11 Ma) (Barnes et al. 2008) causing abundant sedimentation further east (Strub et al. 2005). In the western Interandean zone, flat-lying, coarse-grained deposits (Cangalli Formation) fill a deeply incised paleorelief cut in folded Ordovician rocks on the western side of the Tipuani-Mapiri basin (Mosolf et al. 2011). ^{40}Ar/^{39}Ar analyses with laser fusion on sanidine crystals from a tuff 90 m above the base of this undeformed succession yield a weighted mean age 9.16 ± 0.07 Ma. This data gives a minimum age for initiation of deposition and cessation of upper crustal shortening in this region. These authors concluded that these undeformed deposits post-date the 16–11 Ma exhumation pulse referred to above and indicate that at the moment of sedimentation the locus of deformation had shifted further east, into the Subandean zone.

Ages from the western part of the Subandean zone indicate initial exhumation in Miocene times (20–8 Ma) and an eastward progression of exhumation during late Miocene to Pliocene. The above mentioned exhumation pulses that followed the mid-late Eocene to Oligocene exhumation event in the eastern Altiplano margin and Eastern Cordillera (Barnes et al. 2006, 2008) occured close one after the other, indicating that the compression after the Incaic phase remained continuous in the retroarc region affecting progressively more eastern regions, and causing eastward progression of the Andean deformation front.

Late early to late Miocene

In the eastern Altiplano and western Eastern Cordillera the middle Miocene deposits overlie unconformably pre-16 Ma old deposits (Hérail et al. 1997; Rochat et al. 1998; Rochat 2002). However, in the westernmost Altiplano basin (Corque Basin), middle Miocene deposits form a continuous succession with the older deposits and reach a thickness of about 6,000 m that diminishes toward the sides of the basin. These deposits form an upward fining succession that begins with sandstones and mudstones with conglomeratic intercalations.

In the southern Altiplano, south of the Uyuni Salar, at 21°S, conglomeratic clasts in the lower and upper portions of the westernmost outcrops of the ~3,000-m-thick San Vicente Formation have compositions that indicate a western provenance (Silva 2004). The lower conglomerates being of late Oligocene age are supportive of the late Oligocene tectonic event. Renewal of tectonic activity in middle to late Miocene is indicated by progressive unconformities in the upper conglomeratic horizons interbedded with tuff layers dated (K–Ar) at 15.8 ± 1.2, ~11 ± 0.3 and 7.6 ± 0.2 Ma (Silva 2004). The less resistant apatite population in the sample collected by Ege et al. (2007) in the westernmost outcrops of the lower Potoco Formation in this same region yield a cooling age of 17.9 ± 1.7 Ma. These early to late Miocene ages obtained for exhumation, deformation and coarse clastic sedimentation indicates that compression was continuous during most of the Miocene.

The arc-related volcanic activity that resumed in late Oligocene times continued during Miocene times in the present-day Western Cordillera and western Altiplano pouring dark, porphyric predominantly basaltic-andesitic lavas, and locally ignimbritic flows interbedded with epiclastic deposits (Jiménez et al. 2008). In late Miocene and Pliocene, arc-related volcanism reached a climax (Jiménez et al. 2008, 2009). Extended volcanic fields and complexes mostly associated with calderas developed along the arc.

In late Miocene, sedimentation continued in the Altiplano basin and tectonic activity was mainly concentrated further east at the deformation front. According to McQuarrie et al. (2005), the foreland basin was now located far to the east, almost at its present-day position. Further uplift of the Eastern Cordillera controlled by the east-vergent thrust fold belt caused uplift of a low relief geomorphic surface, the 13–9 Ma old San Juan del Oro Surface, presently located at altitudes between 3,000 and 4,000 m in south-western Bolivia and north-western Argentina (Servant et al. 1989; Sempere et al. 1990; Gubbels et al. 1993; Kennan et al. 1995). The San Juan del Oro deposits seal the Cenozoic structures formed in the central-to-eastern Eastern Cordillera indicating that deformation wanned shortly after 13 Ma in this region (Gubbels et al. 1993). This conclusion is consistent with the ages obtained in the uppermost deformed levels of the western San Vicente Formation in the southern Altiplano, in Bolivia (Silva 2004), after which only little deformation affected these deposits.

The early to late Miocene tectonic activity described above for the retroarc most certainly continued causing uplift in the arc region, which enhanced sediment supply and subsidence in the basins to the east. Late Miocene sedimentation in the Altiplano resulted principally in volcanogenic units and low-gradient fluvial deposits (Rochat 2002; McQuarrie et al. 2005). These deposits are not deformed.

In the western Corque Basin in the central Altiplano, late Miocene sedimentation is characterized by the presence of conglomerates of western origin. These deposits contain at their base a 9.03 ± 0.07 Ma old tuff layer (Callapa Tuff) (Marshall et al. 1992). The late Miocene deposits in the west and east sides of the central Altiplano basin exhibit growth strata that indicate that tectonic activity continued to be present in the region (Rochat 2002).

Pliocene to present

In this period, volcanic activity continued uninterruptedly along the arc, as well as, the sedimentary filling of the internally drained Altiplano basin.

Correlation between the tectonic and paleogeographic evolutions on both sides of the Incaic axis

Based on the presented information, we next correlate the post-mid-Eocene tectonic events on both sides of the Incaic axis, and determine how these events controlled further evolution in the study region (Figs. 9, 10). Being the Incaic range the main morphologic feature since the mid-Eocene, evolution in this Andean region has been strongly influenced by its presence. The particularly good preservation of rather old morphologic features in the western side of the Andes in this region has been favored by the extreme aridity prevailing since approximately late Oligocene times (Alpers and Brimhall 1988; Dunai et al. 2005).

Middle Eocene

West of the Incaic axis, in the study region, the upper Azapa Formation is the only known Cenozoic unit older than late Oligocene. Therefore, a paleogeographic reconstruction for the Precordillera in middle Eocene time can only be made on the basis of the evidence provided by: (1) The Azapa deposits further south, between 19°30′ and 21° S (Pinto et al. 2004a; Victor et al. 2004; Farías et al. 2005), and (2) The middle Eocene to Oligocene detrital Sichal Formation exposed still further south, between 21°30′S and 22°S, in the Domeyko range (Sierra de Moreno) (Maksaev 1978; Skarmeta and Marinovic 1981; Blanco and Tomlinson 2006). In the first region, the great thickness of the deposits, the early Oligocene age estimated by Victor et al. (2004) for their lower portion, and their eastern provenance suggest existence of an uplifted relief supplying detrital material to its western flank. In the second region, the mid-Eocene to Oligocene Sichal Formation consisting predominantly of coarse, alluvial sediments resulting from erosion on Paleozoic units indicates also existence of intense erosion on an uplifted relief. This general picture can be extrapolated to the study region, where the paleogeographic setting for the late Oligocene was similar. According to the oldest age in the Sichal Formation (~43 Ma, Blanco and Tomlinson 2006), we associate this relief with the Incaic phase (Fig. 10b).

East of the Incaic axis, the major reorganization of the sedimentation pattern described between the lower and upper Potoco Formation indicating a new stage of greater energy of sediment transport and a clear west to east transport direction occurred at this moment. This major paleogeographic modification is considered as an evidence for development of a relief to the west of the present-day Altiplano, the Incaic range (Figs. 8, 10a).

Late Eocene to mid-Oligocene

In the Precordillera, east–west-oriented and westward flowing waters formed a parallel drainage pattern that transported sediments to the present-day Central Depression, and eastern Coastal Cordillera regions (Pampa del Tamarugal Basin: Mortimer 1980; Jordan et al. 2010; Nester and Jordan 2012) (Fig. 10c). The resulting deposits correspond to the thick, coarse-grained, alluvial, Oligocene Azapa Formation. Westward fining, clast composition, heavy mineral content and sediment geochemistry indicate a sediment source located in the present-day eastern Precordillera.

In the Altiplano basin, the thick deposits of western provenance of the upper Potoco Formation indicate that the western uplifted source of sediments mentioned previously for mid-Eocene times still existed (Fig. 9). This uplifted region apparently was the same that supplied sediments to the Pampa del Tamarugal Basin, and would have corresponded to the Incaic range. The mid-late Eocene to Oligocene ages for exhumation in eastern regions of the eastern Altiplano and Eastern Cordillera (Barnes et al. 2008), are interpreted as evidence for continuous and eastward shift of the Incaic compressive event (Fig. 8).

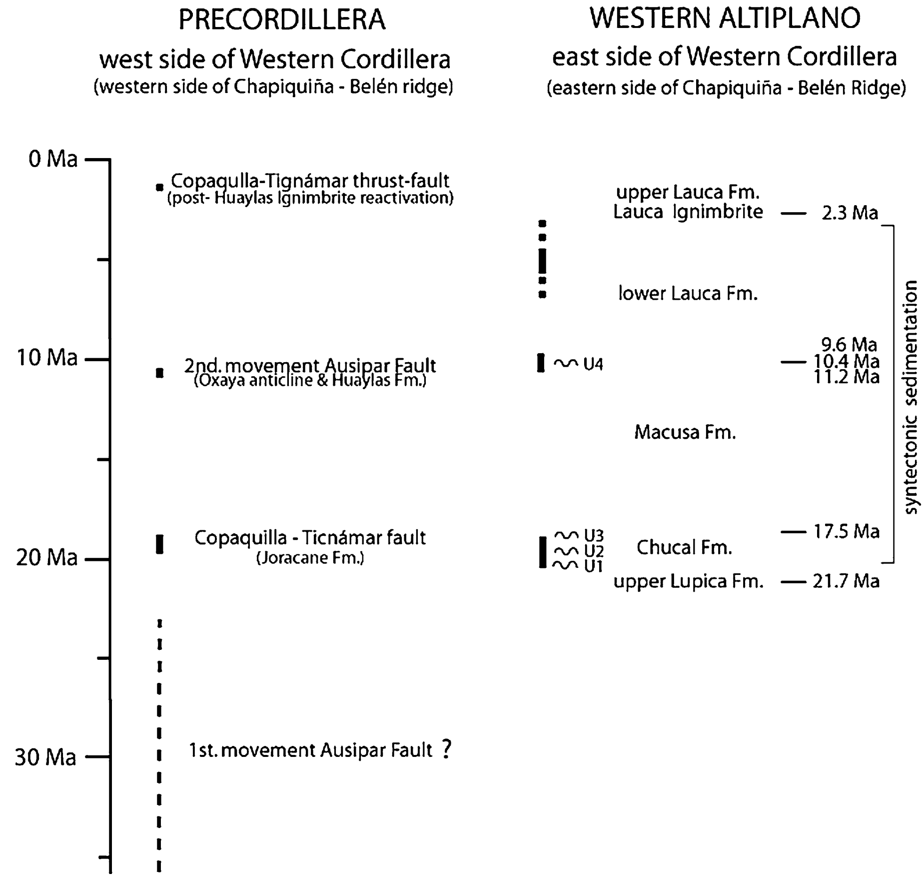

Fig. 9 Correlation of stratigraphic units and main tectonic events between both sides of the Incaic axis in Chile: the Precordillera and Western Cordillera

Late Oligocene

West of the Incaic axis, erosion on the west-flank of the Incaic range was still active at this time and continued its contribution of detritus to the upper Azapa Formation. In the western Altiplano, east of the Incaic axis, the western provenance of some components in the Azurita and Tambillo formations is considered evidence for further uplift of the Incaic range or, at least, further existence of an uplifted region.

Paleogeographic conditions in the Precordillera remained as described for late Eocene to Oligocene times. Shortly before 24.7 ± 0.3 Ma, intense felsic, explosive volcanic activity began that formed the Oxaya and Lupica formations, in the Precordillera and Western Cordillera, respectively. Volcaniclastic deposits in the Chucal and Macusa formations, in the Western Cordillera, belong to this felsic volcanic event. This volcanism is coeval with the continuous arc-related volcanic activity reported in western central Bolivia from late Oligocene (San Vicente Formation, in the southern Altiplano) to Miocene (Jiménez et al. 2008, 2009). Beginning of this volcanic activity coincides with the late Oligocene–early Miocene tectonic event that developed the westvergent Huarina and San Vicente thrust belts on the western margin of the Eastern Cordillera, in Bolivia (late Oligocene phase) (Fig. 8). With uplift of the western Eastern Cordillera, endorehic conditions began in the Altiplano.

Early to late Miocene

Early Miocene tectonic activity occurred on both sides of the Incaic axis: (1) West of the axis, in the eastern Precordillera, at 18.2 ± 0.8 Ma, activity in the WTS, along the Belén-Tignámar and Copaquilla-Tignámar faults, caused deposition of the syntectonic Joracane Formation, and between 11.7 ± 0.7 and 10.7 ± 0.3 Ma activity along the Ausipar Fault, caused development of the Oxaya anticline, (2) East of the axis, in the Western Cordillera, in the ETS, between 21.7 ± 0.8 Ma (Lupica Fm.) and sometime between 11.2 ± 0.5 and 10.4 ± 0.7 Ma (Macusa Fm.), tectonic activity caused growth deposits and progressive unconformities in the Chucal region, and, further east, in the eastern Altiplano basin, the upward coarsening of the Coniri and San Vicente formations and equivalents (Fig. 8).

The described activity in the thrust systems of opposed vergencies (WTS and ETS) caused the ~1,000 m uplift of the Chapiquiña-Belén ridge and rejuvenation of the reliefs previously formed with the Incaic phase, enhancing erosion

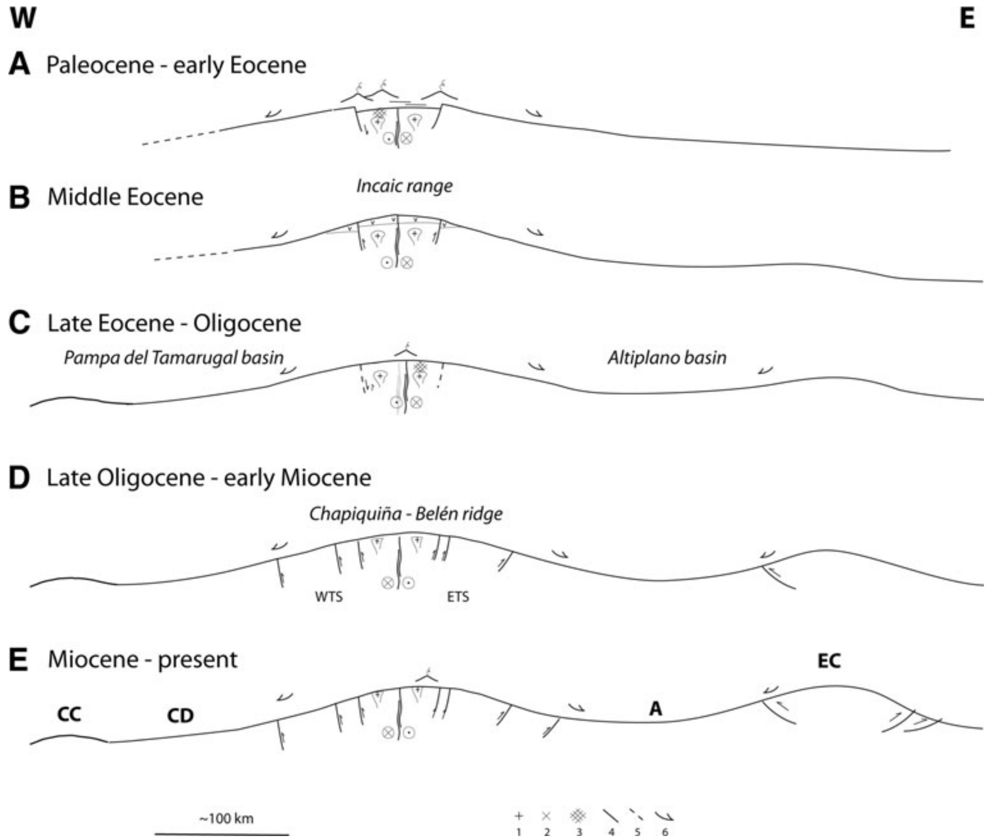

Fig. 10 Schematic paleogeographic development since Paleocene in the study region in northern Chile and western Bolivia, modified from Charrier et al. (2009). Figure shows development of the Incaic Range and adjacent sedimentary basins: Pampa del Tamarugal forearc basin and Altiplano foreland retro-arc basin. **a** Pre-Incaic intra arc, **b** tectonic inversion of the intra arc and development of the Incaic Range separating domains with different evolutions from this moment onward, **c** erosion of the Incaic Range and development of the foreland retroarc Altiplano Basin and incipient forebulge, **d** Late Oligocene-Early Miocene compression, relief rejuvenation, and development of the Chapiquiña-Belén Ridge (pop-up); tectonic compression in the Incaic foreland caused west-vergent thrusting in the Huarina Belt and uplift on the western side of the Eastern Cordillera. **e** Miocene to present paleogeographic situation. In sequence eastward thrusting continued to the east of the Eastern Cordillera in the Subandean zone

and sediment supply to both sides of the ridge (Fig. 10d). The Chapiquiña-Belén ridge separated two different paleogeographic environments on its western and eastern sides, as previously did the Incaic range.

On the western side of the ridge a large slope slightly inclined to the west formed the general landscape. This slope connecting the upper ridge with the western depocenter in the present-day Central Depression was continuous and not dissected at this time. Sediments onlapped eastwards the western flank of the ridge forming the Tarapacá pediment. Here, extremely dry climatic conditions prevailed and only sporadic water flows came from the high regions.

Development of the Oxaya anticline is roughly coeval with the development of the Miocene flexure-folds formed by propagation of subvertical blind faults (involving little horizontal shortening and great uplift) described in the Precordillera of southern Peru (Tosdal et al. 1984; David 2007) and further south in Chile (Galli and Dingman 1962; Pinto 1999; Victor and Oncken 1999; Pinto et al. 2004a; Victor et al. 2004; Farías et al. 2005; Muñoz-Tolorza 2007) (Fig. 4). Development of the Oxaya anticline interrupted the smooth westward-dipping slope modifying the parallel drainage pattern on the Tarapacá pediment in an orthogonal pattern (treillis pattern of García and Hérail 2005).

During this time, the exhumation event that began in middle Eocene continued practically uninterruptedly until late Oligocene–early Miocene (25–19 Ma) in the Eastern Cordillera (Barnes et al. 2008). Further east, in the Interandean zone, exhumation occurred in early Miocene (22–19 Ma) and was followed by a second exhumation pulse in mid Miocene (16–11 Ma) (Barnes et al. 2008) (Fig. 8). Deformation and continuous exhumation at this time east of the Incaic axis suggests that the mid-Eocene compressional event that uplifted the Incaic range propagated gradually and essentially uninterruptedly eastward toward the retro-arc region until, at least, late Miocene times.

Incision in the Tarapacá pediment, in northern Chile and southern Perú, began in late Miocene and represents initiation of the last stage of morphological evolution in northern Chile and southern Perú, and probably also in southern Bolivia and northwestern Argentina with incision on the San Juan del Oro Surface (Gubbels et al. 1993; Hérail et al. 1996; Barke and Lamb 2006).

Late Miocene to present

Giant landslides in the study region resulted from instabilities on the western flank of the Oxaya anticline, like the Lluta avalanche or Lluta collapse (Naranjo 1997; Uhlig 1999; Pinto 1999; García 2002; Pinto et al. 2004b, 2008; García and Hérail 2005; Strasser and Schluneger 2005) (Fig. 4). The Lluta collapse was controlled by N–S-oriented extensional cracks developed along the hinge of the Oxaya anticline (crestal grabens, Fig. 6) and the western steep slope on the western side of the Oxaya anticline. In the same tectonic context, mega-landslides associated with the folding and flexuring of the western flank of the Precordillera have also been observed in southern Perú (David 2007), and further south in Chile, in the Miñimiñi and Camiña regions (Pinto et al. 2004b, 2008) (Fig. 4). Apart from these, other local landslides occur on the slopes of the valleys, the origin of which is directly related to gravitational destabilization of the slopes during incision rather than to tectonic activity, like the General Alcérreca, in the upper Lluta river valley (Mortimer 1980).

In the present-day Coastal Cordillera and Central Depression, canyon-forming incision (~600–1,000 m deep) through the Tarapacá pediment began after 11.9–11.2 Ma in the region of the Lluta and Azapa river valleys (García et al. 2011) and after ~6 Ma in the Tiliviche and Camiña region (Naranjo and Paskoff 1985; Hoke et al. 2007). Further north, in southern Perú, the main incision developed between 13–9 and 3.8 Ma (Thouret et al. 2007). Incision through the Tarapacá pediment represents a major change in the drainage system and certainly also in the amount of water supply. In this extremely arid region, such a change can only have occurred once the mountain range had reached a certain altitude and, thus, being triggered by locally increased rain and snow-fall (García et al. 2011).

A late evidence of tectonic activity is the Pliocene reactivation of the Copaquilla-Tignámar fault that uplifted the eastern side of the Lauca ignimbrite relative to its western part. This activity correlates well with the eastward thrusting of the lower deposits of the Lauca Formation on the east side of the Chucal anticline. Compressive activity ended before deposition of the undeformed Lauca-Pérez ignimbrite, at ~2.7–2.9 Ma (Fig. 9).

Thermochronologic dates for the Subandean zone indicate initial exhumation in Miocene times (20–8 Ma) for the western part and an eastward progression of exhumation during late Miocene to Pliocene.

The tectonic and morphologic evolution described for the western side of the Incaic axis correlates with the tectonic activity that caused deposition of the upper conglomerates in the San Vicente Formation and the existence of progressive unconformities in these conglomeratic horizons.

Synthesis and discussion

We next discuss relevant aspects raised in this article relative to the paleogeographic and tectonic evolutions of the Andean range at this latitude, and compare this evolution with that of the Andes south of 27°S in central Chile and Argentina.

Relative to tectonic regime and chronology

Since the major Incaic phase in middle Eocene the tectonic regime in the study region remained compressive until present, with a peak of deformation in late Oligocene–early Miocene that apparently coincides with the Quechua I phase of Mégard (1984), in Peru. As shown in this article, protracted contraction in middle to late Miocene along the thrust systems in the retroarc region caused continuous surface uplift to considerable altitude, and probably westward tilting of the orogene. After the late Oligocene–early Miocene event, compression remained uninterruptedly active through the rest of the Cenozoic causing deformation in different regions at different moments, although mainly concentrated on the eastward shifting orogenic front. Accumulated contraction in the brittle upper crust and westward tilting needs some compensation that we propose was achieved by westward mass transfer in the lower lithosphere as Farías et al. (2010) proposed for the Andeas in the central region of Chile and Argentina (Fig. 11). Continuous compression since middle Eocene suggests that the altitude of the orogen in this region has been and still is mechanically supported by such stresses.

The compressional peak that gave rise to the Incaic range is coincident with the period of high convergence rate, between approximately 50 and 42 Ma, reported by Pardo-Casas and Molnar (1987). The compressional event that caused activity in the WTS and ETS, in northern Chile, as well as, in the Huarina and San Vicente thrust belts, in the Altiplano basin, in western central and southern Bolivia, coincides with the period of high convergence rate detected in late Oligocene–Miocene by the mentioned authors and Somoza (1998).

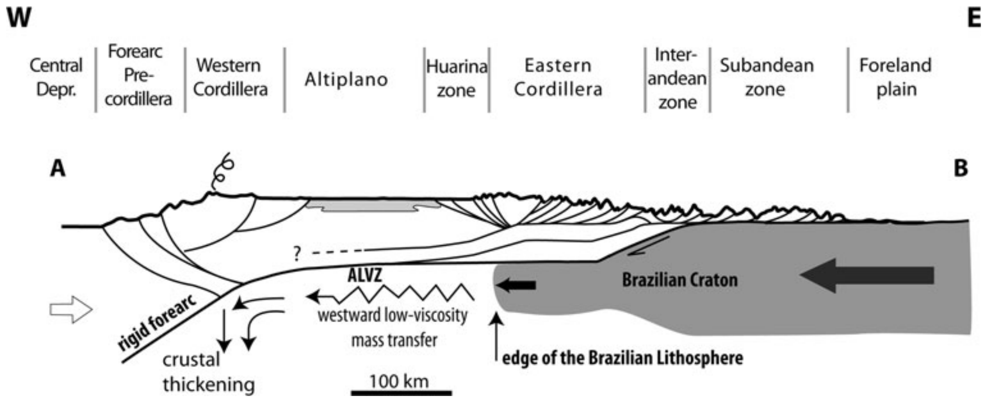

Fig. 11 Skematic architecture for the Andes in northernmost Chile and central western Bolivia, based on McQuarrie et al. (2005) and Farías et al. (2005). Pop-up like geometries in the Precordillera and Western Cordillera, and underneath the Eastern Cordillera developed on top of ramp to flat transitions. *Shaded zone* beneath the Subandean zone and Eastern Cordillera is the rigid upper crust of the Guaporé craton underthrusting westward the eastern side of the Andes (A subduction). Location of the downgoing craton is based on Beck and Zandt (2002). The westward penentration of the craton underneath the lithospheric décollement causes ductile flow in the weekened lower crust below the Altiplano and Western Cordillera (Altiplano low-velocity zone—ALVZ; Victor et al. 2004). Ductile flow further west is restrained by the rigid forearc, causing an increase of crustal thickness in this region

Relative to paleogeography

The middle Eocene Incaic phase developed a NNE–SSW-oriented relief, the Incaic range that formed the water divide at that time and separated two paleogeographic domains in the Andean region. The late Oligocene–early Miocene tectonic event rejuvenated this relief and formed the Chapiquiña-Belén ridge, which was uplifted on two divergent thrust systems, the WTS, in the Precordillera, and the ETS, in the Western Cordillera and Altiplano. The relief formed by the Incaic range and later the Chapiquiña-Belén ridge separated two paleogeographic domains with different climatic conditions and has been affected by intense erosion, thus, supplying with abundant sedimentary material the basins developed on both sides: The Pampa del Tamarugal basin (Mortimer 1980; Jordan et al. 2010; Nester and Jordan 2012), to the west, and the Altiplano Basin, to the east.

The Pampa del Tamarugal basin extending northward into southern Perú was bounded to the west by another uplifted ridge in the present-day location of the Coastal Cordillera (Mortimer et al. 1974; Parraguez 1998; García et al. 2004), the Coastal Cordillera precursor ridge (Fig. 10c, d). In the study region, the basin remained apparently isolated from the ocean until late Miocene times (∼9–8 Ma), when fluvial incision through the basin infill and the Coastal Cordillera precursor ridge created local connections to the Pacific Ocean (Mortimer et al. 1974; Pinto et al. 2004b; Farías et al. 2005; García and Hérail 2005). This basin hosted thick successions of generally coarse, westward fining and thinning, detrital and pyroclastic deposits of the Azapa, Oxaya and El Diablo formations that onlapped eastward onto a pre-Oligocene erosion surface (Coastal Tarapacá Pediplain of Mortimer and Saric 1972, and/or Choja pediplain of Galli 1968). Continuous sediment accumulation in the basin formed a regional gently west-sloping surface that corresponds to the top of the El Diablo Formation that forms the surface of the Tarapacá pediment. In the Coastal Cordillera precursor ridge, a surface similar to the Choja pediplain was covered with late Oligocene deposits of the Azapa Formation (Parraguez 1998; García et al. 2004; Dunai et al. 2005). This surface has been named "Coastal Tarapacá pediplain" by Mortimer et al. (1974).

Additional evidence for the existence of a relief bounding the basin to the east is: (1) Supergene alteration ages of 34–35 Ma in the Cerro Colorado mining district (Sillitoe and McKee 1996; Bouzari and Clark 2002), in the Precordillera, at 19°30′S, indicating existence of weathering surfaces in late Eocene times, and (2) A 37 Ma old exposure age on a clast extracted from a ∼23-Ma-old depositional surface in the Coastal Cordillera at the same latitude, indicating clast exposure since middle Eocene times before being incorporated to the deposit (Dunai et al. 2005).

To the east, in the present-day Altiplano, the middle-late Eocene to late Oligocene Altiplano basin can be characterized as a retro-arc fore-land basin. A similar view had been advanced by Sempere et al. (1990). Here, as a consequence of the Incaic phase, rapid subsidence and major drainage reorganization occurred already in middle Eocene times with abundant sediment supply from the west. In late Oligocene times, with activation of the Huarina and San Vicente thrust belts, the Altiplano basin became an internally drained depression bounded on both sides by inward verging thrust belts and uplifted ridges. At the same time, eastward progression of the deformation front (fold-thrust belt) continued east of the incipient Eastern Cordillera and at this moment the Altiplano became a hinterland basin (Horton 2012). The Altiplano and the Pampa del Tamarugal basins are symmetric features on both sides of the Incaic range, resulting from tectonic contraction against the rigid block formed by the uplifted Incaic range (Fig. 10c). Furthermore, the Coastal Range precursor ridge, on the western side, and the incipient Eastern Cordillera, on the eastern side, emphasizes this symmetric paleogeographic organization: two internally drained basins separated from each other by an uplifted ridge. Cenozoic deposits in the Altiplano basin form a thick succession consisting of the main Potoco Formation interval (deposited after the reorganization of the drainage pattern in middle Eocene), and the Coniri Formation and stratigraphic equivalents. This succession is coeval with the one exposed on the western side (Pampa del Tamarugal basin), consisting of the Azapa, Oxaya and El Diablo formations.

Relative to the magmatic arc

The Late Cretaceous to early Eocene magmatic arc, in northern Chile and southern Perú was located in the present-day Precordillera and Western Cordillera. Subsequent arc development was located at about the same position until present, with only slight eastward displacement relative to the position of the arc in early-middle Eocene time. In this Andean region, although Paleocene to middle Eocene arc deposits are almost inexistent, there is good evidence for coeval plutonic activity and emplacement of associated porphyry copper deposits (Cornejo et al. 1997; Camus 2003; Cornejo 2005). Similarly, late Eocene to early Oligocene arc-related volcanic deposits are practically absent (Lahsen 1982). This situation has been interpreted as a major interruption of the arc magmatic activity after the Incaic phase ("Oligocene magmatic lull") that lasted 8 m.y., between 37 and 29 Ma (Reutter 2001). Since then, arc activity remained at about the same position until present, only slightly shifted to the east of the previous magmatic centers, indicating that in this region the arc has been almost static ("quasi-static arc"). Arc volcanism in late Oligocene and early Miocene times produced the voluminous explosive deposits and lavas on both sides of Incaic axis.

We argue that the absence of most of the Paleocene to late Oligocene volcanic deposits in this region can be explained by the uplift and long lasting (over 15 m.y.) erosion on the Incaic range that coincided with the location of the previous magmatic arcs. The long exposure to weathering and erosion on this uplifted block is also considered to be the cause for intense late Eocene to early Oligocene, and late Oligocene to early Miocene supergene enrichment of the porphyry and allied copper deposits developed along the previous arc, to the south of the study region, at 19°30′S (Bouzari and Clark 2002). The two mentioned stages of supergene enrichment coincide with periods during which erosion has been particularly intense due to the episodes of uplift, first of the Incaic range, and then of the Chapiquiña-Belén ridge.

In this region, the Western Cordillera (present-day volcanic arc) consists of recent volcanic edifices (>6,000 m altitude) built on top of the Belén–Chapiquiña ridge, whereas on the eastern side of the Andes their basement consist mainly of tectonically uplifted early Paleozoic intrusive rocks. The pop-up structure of the Chapiquiña-Belén ridge in the Western Cordillera, bounded by two thrust systems of opposite vergency (the WTS and the ETS), can be compared with the Eastern Cordillera, which is also developed between two thrust systems with opposite vergencies, east and west of the Eastern Cordillera (Fig. 10e). These two major pop-up ridges bound the Altiplano basin and determine its endorehic condition.

Relative to crustal structure and thickening

It has been suggested that uplift of the Incaic range was controlled by two thrust systems with opposite vergencies (Charrier et al. 2007, 2009), which were reactivated in Miocene times favoring the uplift of the Chapiquiña-Belén ridge. Tectonic shortening in the Precordillera-Western Cordillera during Neogene have been rather small (only 8 km) and consequently crustal thickening has been small (García 2002; García et al. 2004; Rochat 2002). However, crustal thickening in the Andean forearc has been gradually increasing from west to east since Mesozoic times with alternate magmatic accretion and crustal deformation, in: late Early Cretaceous (Peruvian Phase), late Cretaceous–early Cenozoic (K–T Phase), middle Eocene (Inacaic phase), and Cenozoic times during which continuous compression occurred as indicated in the preceding item. With the superposed effect of this tectonic activity, the forearc became a rigid crustal block (Roperch et al. 2000; Arriagada et al. 2000, 2003; Farías et al. 2005; David 2007). If we consider that there has been no major shift of the magmatic arc since Late Cretaceous times, both magmatic accretion and protracted contraction that caused uplift of the Incaic range and later the Chapiquiña-Belén ridge are the main factors that contributed to crustal thickening in the arc region. Thus, thick-skinned deformation apparently prevailed in the region west of the Eastern Cordillera, that is, in the forearc, arc, and inner retro-arc. This situation strongly constrasts with the one developed on the eastern side of the mountain range (east of the Inerandean zone), where shortening and crustal thickening took place along a thin-skinned fold-thrust belt based in thick early Paleozoic ductile deposits (Roeder 1988; Baby et al. 1989, 1997; Sheffels 1990; McQuarrie et al. 2005). The enormous (over 300 km) east–west shortening calculated for this region (Roeder 1988; Sheffels 1990; Baby et al. 1997; McQuarrie et al. 2005, among others) corresponds to the upper crustal accommodation of strain accumulated by the eastward movement of a crustal-scale ramp-flat structure, which connects the Andean front in the eastern Subandean Zone and the subduction zone (Farías et al. 2005). Therefore, deformation in the brittle upper crust depends, apart from the stresses transmitted from the subduction zone, on the geometry of the crustal-scale ramp-flat structure.

Considering that subduction was continuous during the Cenozoic, stresses were most certainly transmitted without interruption all over this period from the subduction zone to the upper crust. Stress transmission must have been considerably increased in the Eocene, when frontal convergence began and continued through the Neogene (Pardo-Casas and Molnar 1987). Development at several stages during late Oligocene to late Miocene of: growth strata on both sides of the Altiplano basin, the Oxaya anticline on the hangingwall of the Ausipar fault, in the Precordillera, and flexures related to blind west-vergent thrust faults, further south in the Precordillera (Victor et al. 2004; Pinto et al. 2004b; Farías et al. 2005) (Fig. 4), plus present-day dextral transpressional shallow seismicity along the west-vergent thrust system in the Precordillera (Farías et al. 2005), and in the Eastern Cordillera (Vega and Buforn 1991; Dorbath et al. 1993; Funning et al. 2005) support the idea that contractional deformation has been constant throughout late Cenozoic times, although activity has been mainly concentrated on the Andean deformation front. Therefore, we suggest that the Incaic phase and the late Oligocene event represent episodes of more intense stress transmission and widespread strain, and can be related to periods of increasing convergence rate and more intense coupling between the intervening plates (Charrier et al. 2002, 2007, 2009), or to accelerated westward shift of the South American Plate (Ramos 2010).

Uplifted crustal blocks, parallel to the continental margin, like the Incaic range and its successor, the Chapiquiña-Belén ridge, and the Eastern Cordillera bounded by thrust systems with opposite vergencies, are apparently located on top of bends (transition from ramp to flat segments) along the ramp-flat structure (Fig. 11). There is little evidence for the structural control of the Coastal Range precursor ridge in this region. However, in the Atacama region, between 23°45′S and 29°S, the Atacama basin, a southern forearc basin of the same nature as the Pampa del Tamarugal basin, has been dammed to the west by activity in Middle Miocene times along the El Salado–Vallenar segment of the Atacama fault system (Riquelme et al. 2003). According to these authors, the western part of the present-day Coastal Cordillera has been uplifted, through a vertical fault, relative to the eastern part since the middle Miocene causing isolation of the basin from the ocean, and creating favorable conditions for accumulation of detrital material transported from the eastern uplifted reliefs and for development of the Atacama pediment. Recent uplift in the Coastal Cordillera in the Antofagasta region, south of the study region, has been related to underplatting underneath the continental wedge (González et al. 2003). Such process might also be responsible for movements along the Atacama fault system as claimed by Riquelme et al. (2003).

Protracted tectonic shortening and likely magmatic accretion in this region resulted in considerable crustal thickening underneath the eastern forearc, arc, inner retroarc, and Eastern Cordillera. Upper crustal shortening developed above the crustal-scale ramp-flat detachment is not enough to explain the present-day Moho depth and the altitude of the Western Cordillera, Altiplano and Eastern Cordillera (Housson and Sempere 2003). However, thick-skinned underthrusting of the Brazilian craton below the Altiplano since the late Miocene can be considered as a simple-shear thickening process that has accommodated more than 200 km of shortening according to geophysical images of the Altiplano lithosphere (Beck and Zandt 2002). The westward advance of the rigid craton beneath the major detachment should have produced westward likely low viscosity mass (Fig. 11). However, rigidity of the forearc region would have impeded this advance thus producing crustal thickening below the Western Cordillera several hundreds of kilometers west of the underthrusting zone, resulting in general surface uplift of the entire width of the orogen (Farías et al. 2005; Tassara 2005) (Fig. 11).

Relative to Andean uplift and origin of the Altiplano-Puna plateau

Several studies coincide with the existence of an uplift event in northern Chile, western central Bolivia and northwestern Argentina between ~11 and ~7 Ma (Gubbels et al. 1993; Garzione et al. 2006; Hoke et al. 2007). Paleofloristic observations from the Altiplano in the study region indicate that a considerable altitude has been acquired in late Miocene (Charrier et al. 1994a; Gregory-Wodzicki et al. 1998; Gregory-Wodzicki 2000). Similar views have been expressed on the basis of the late Miocene age of incision in northern Chile (Riquelme et al. 2003, 2007; Pinto et al. 2004a, 2007; Farías et al. 2005). According to the evidence presented in this paper, uplift of the Andes in this region can be regarded as the result of a continuous and long lived compressive tectonic regime that began in middle Eocene with the Incaic phase (and even before with the Peruvian phase in late Early Cretaceous) and continued until present. A similar view has been expressed for the Neogene on the western side of the range (Jordan et al. 2010). In our opinion, the uplift event detected between ~11 and ~7 Ma would represent a peak of increased uplift rate during a continuous, long lasting uplift process and not an individual, isolated orogenic phase. The pre-10 Ma period of continuous compression and associated uplift would correspond to the slow and steady rise model for the Andes described by Barnes and Ehlers (2009), whereas the post-10 Ma stage of more rapid uplift would correspond to the rapid and recent deformation stage of these authors.

Hyperaridity has been invoked to have triggered Andean uplift (Lamb and Davis 2003). Considering that in the Andean region in northern Chile and adjacent Perú, Bolivia and Argentina crustal shortening and thickening are processes that have been occurring uninterruptedly since mid-Eocene time until present (this paper; Haschke and Günther 2003; Mamani et al. 2010), we think that uplift has also been gradual and uninterrupted during that time. If this were so, uplift would have been occurring since long before onset of hyperarid climatic conditions, even accepting the oldest dates obtained for this climatic event at 23–25 Ma (Dunai et al. 2005). Considering that, according to Rech et al. (2006), onset of hyperaridity most probably occurred between 19 and 13 Ma, we argue that hyperaridity might have enhanced and accelerated, but not necessarily triggered, surface uplift between ~11 and ~7 Ma.

Conclusions

1. Protracted contraction since the major Incaic tectonic phase in middle Eocene until present, punctuated by two peaks of increased deformation in late Oligocene–early Miocene, and late Miocene, controlled morphologic evolution and sedimentation in the study region.
2. The Incaic phase caused development of a NNE-SSW-oriented relief or Incaic range. This relief formed in late Eocene and Oligocene times the Andean water divide in this region and separated two paleogeographic domains. The late Oligocene–early Miocene peak of deformation caused uplift and rejuvenated this relief forming the Chapiquiña-Belén ridge bounded by two divergent thrust systems, a westvergent (WTS) and an eastvergent (ETS).
3. The Incaic phase coincides with the episode of high convergence rate, between approximately 50 and 42 Ma reported by Pardo-Casas and Molnar (1987) and Somoza and Ghidella (2005). The late Oligocene–early Miocene peak of increased deformation that caused activity in the WTS and ETS, in northern Chile, and the Huarina and San Vicente back-thrust belts, in the Altiplano, coincides with the episode of increased convergence rate detected by Pardo-Casas and Molnar (1987) and Somoza (1998) at this time.
4. The major morphologic feature formed by the Incaic range and later by the Chapiquiña-Belén ridge formed an axis of symmetry in this Andean region (Incaic axis). On both sides of the axis a basin and a ridge were formed: (1) West of the Incaic axis, the Pampa del Tamarugal basin and further west the precursor ridge of the Coastal Cordillera, and (2) East of the axis, the Altiplano basin and the Eastern Cordillera (Fig. 10e).
5. During late Eocene and Oligocene, the Altiplano basin corresponded to a retroarc foreland basin. Development of the Huarina and San Vicente back-thrust belts on the western side of the Eastern Cordillera transformed the Altiplano basin into an internally drained basin. Further eastward deformation modified this condition into a hinterland basin.
6. Two main processes seem to be responsible for surface uplift, relief, and altitude of the Andean Cordillera in northern Chile and western central Bolivia, namely: (1) Accumulated contraction of the brittle upper crust transported eastward over a crustal-scale ramp-flat thrust, and (2) westward underthrusting of the Guaporé craton underneath the Subandean zone and the Eastern Cordillera that triggered westward flow in the ductile lower crust enhancing crustal thickening. Additionally, these processes caused westward tilting on the western flank of the mountain range.
7. Cenozoic deformation after the Incaic phase shows a clear and uninterrupted eastward progression. West-vergent structures in the forearc Precordillera are backthrust systems related to transitions from ramps to flats in the crustal-scale ramp-flat structure.

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