Rapid tectonic and paleogeographic evolution associated with the development of the Chucal anticline and the Chucal-Lauca Basin in the Altiplano of Arica, northern Chile

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Abstract

The east-vergent Chucal thrust system, on the east side of the Chapiquiña-Belén ridge in the Western Cordillera, was continuously or almost continuously active for ~ 18 m.y. (<21 to >2.7 Ma). Contractional activity deformed late Oligocene tuffaceous, fluvial, or distal alluvial deposits of the uppermost Lupica Formation; fluvial and lacustrine deposits of the Miocene Chucal Formation; tuffaceous and coarse fluvial deposits of the Quebrada Macusa Formation; and the lower part of the westernmost, latest Miocene?—Pliocene, essentially lacustrine Lauca Formation. It controlled the paleogeographic and paleoenvironmental conditions in which these units were deposited. More humid conditions on the east side of the Chapiquiña-Belén ridge favored the development of an abundant mammal fauna and flora. The deformation is characterized by the Jaropilla thrust fault and the Chucal anticline, which is east of the fault. Deformation on the Chucal anticline began before the deposition of the Chucal Formation and was controlled by a blind thrust fault. The west flank has a nearly constant dip (45–50°) to the west and nearly continuous stratigraphic units, whereas on the east flank, the dip angle is variable, diminishing away from the axis, and the stratigraphic units are discontinuous. The anticline growth on this flank caused the development of three observable progressive unconformities. Deformation was particularly rapid during the deposition of the ~ 600 m thick Chucal Formation (between the 21.7 ± 0.8 Ma old uppermost Lupica Formation and the 17.5 ± 0.4 Ma old base of the Quebrada Macusa Formation, a 4 m.y. period). The deformation rate decreased during the deposition of both (1) the ~ 200 m thick Quebrada Macusa Formation (between the 17.5 ± 0.4 Ma age of its basal deposits and the ~ 11 Ma age of its uppermost levels, a 7 m.y. period) and (2) the lower Lauca Formation (between the ~ 11 Ma age of the upper Quebrada Macusa Formation and the 2.3 ± 0.7 Ma old Lauca ignimbrite, which is intercalated within its middle part). We interpret the contractional deformation to be associated with tectonic activity that led to the uplift of the Altiplano; however, paleobotanical evidence does not indicate any major altitude changes during the time period considered here but rather suggests that rapid uplift took place after the deposition of the Quebrada Macusa Formation.

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Keywords: Altiplano; Andean uplift; Cenozoic paleoenvironment; Chilean Andes; Chucal-Lauca basin; Late Cenozoic; Syntectonic sedimentation

Resumen

La actividad del sistema estructural de Chucal, con vergencia al este, ubicado al este del Cordón Chapiquiña-Belén en la Cordillera Occidental, fue continua o casi continua durante $\sim$18 m.a. (entre $<$21 Ma y $>$2,7 Ma). La contracción deformó depósitos tobáceos y fluviales o aluviales distales del Oligoceno Superior pertenecientes a la parte superior de la Formación Lupica, depósitos fluviales y lacustres del Mioceno de la Formación Chucal, y depósitos esencialmente lacustres del Mioceno?—Plioceno de la parte inferior y más occidental de la Formación Lauca. Controló también las condiciones paleogeográficas y paleoambientales en las que estas unidades se depositaron. Condiciones más húmedas al lado oriental del Cordón Chapiquiña-Belén favorecieron el desarrollo de flora y de abundantes mamíferos. La deformación se asocia a la falla inversa de Jaropilla y al Anticlinal Chucal, al este de la falla. La deformación en el Anticlinal Chucal comenzó antes del inicio de la depositación de la Formación Chucal y estuvo controlada por una falla inversa ciega. Los estratos del flanco occidental tienen un manteo casi constante de 45–50° al oeste y la serie estratificada es contínua, en cambio, en el flanco oriental el manteo disminuye alejándose del eje del pliegue y la serie es discontínua; el crecimiento del anticlinal determinó en este flanco el desarrollo de tres discordancias progresivas observables. La deformación fue particularmente rápida durante la depositación de los $\sim$600 m de espesor de la Formación Chucal (entre los 21,7 $\pm$ 0,8 Ma de la parte superior de la Formación Lupica y los 17,5 $\pm$ 0,4 Ma de la base de la suprayacente Formación Quebrada Macusa, un periodo de 4 m.a.). La velocidad de deformación disminuyó durante (1) El periodo de 7 m.a. que tardó la depositación de los $\sim$200 m de espesor de la Formación Quebrada Macusa, entre los 17,5 $\pm$ 0,4 Ma obtenidos en su base y los $\sim$11 Ma obtenidos en su parte superior y (2) el periodo de 7 m.a. que tardó la depositación de la parte inferior de la Formación Lauca, entre los $\sim$11 Ma de la parte superior de la Formación Quebrada Macusa y los 2,3 $\pm$ 0,7 Ma obtenidos en la Ignimbrita Lauca, intercalada en su parte media. Asociamos esta deformación con la actividad tectónica que causó el alzamiento del Altiplano, sin embargo, evidencia paleobotánica previa no indica cambios altitudinales mayores en el periodo aquí considerado y sugiere, más bien, que el rápido alzamiento se habría producido despúes de la depositación de la Formación Quebrada Macusa.

1. Introduction

The Andean Cordillera is the typical example of a subduction-related mountain belt (cordilleran-type mountain belts sensu Dewey and Bird, 1970), formed through crustal shortening and thickening and magmatic additions along the border of the overriding plate. In this case, the mountain range formed along the western continental margin of South America above the subducting Nazca plate (Jordan et al., 1983; Allmendinger, 1986; Isacks, 1988; Ramos, 1988; Kono et al., 1989; Kay and Abbruzzi, 1996; Allmendinger et al., 1997; Muñoz and Charrier, 1996). The tectonic evolution and resulting paleogeographic features of the southern central Andes therefore directly reflect subduction activity (Charrier, 1973; Frutos, 1981; Jordan et al., 1983, 1997; Ramos, 1988; Mpodozis and Ramos, 1989).

Although the convergence of the Nazca and the South American plates suggests a continuous, long-term compressive strain regime along the active continental margin, the resulting tectonic style suggests that the stress regime underwent a series of major changes during Andean evolution. Along the northern and central Chilean Andes, there is growing evidence of a late Cenozoic extensional event followed by a long episode of contraction (for the region discussed herein, see García, 1996; García et al., 1996; for farther south at 27°S, see Mpodozis et al., 1995; for localities between 33 and 36°S, see Charrier et al., 1994a, 1999, 2002; Godoy and Lara, 1994; Godoy et al., 1999; Jordan et al., 2001). The tectonic style that results from a contractional episode generally is characterized by high-angle thrust faults with variable vergence, which strongly suggests that they are inverted normal faults (Muñoz and Charrier, 1996; García, 1996, 2001; Farías et al., 2002). In the altiplano of northern Chile, the contractional episode began in the early Miocene and seems to have continued until the Pliocene (Muñoz and Charrier, 1996; García, 1996, 2001; García et al., 1996; Riquelme, 1998). The plant and palynologic content of the early Miocene deposits described subsequently strongly indicate that they were deposited at a rather low altitude above sea level (Charrier et al., 1994b). A similar conclusion has been derived from somewhat younger deposits located only 100 km away in Bolivia (Gregory-Wodzicki et al., 1998).

We present evidence of rapid east-vergent contractional deformation during the Miocene in the Western, or Volcanic, Cordillera of the Chilean altiplano in the Arica region (18–19°S), following a probable extensional episode. This deformation episode is associated with important local topographic changes.

We describe the early Miocene–Pliocene syntectonic evolution of a series of fluvial and lacustrine deposits, associated with a well-exposed, east-vergent, anticlinal structure in the Cerro Chucal region, next to Salar de Surire (Fig. 1). The excellent exposures, numerous fossil-bearing horizons, and frequent occurrence of abundant datable volcanic material provide a detailed chronology of the evolution of the east-vergent structural system that developed on the east side of a contemporaneously uplifted block: the Chapiquiña-Belén ridge (Charrier et al., 2000) (Fig. 2). The uplift of this block probably caused major modifications in the drainage pattern of the Arica region, as well as rapid paleogeographic variations on the eastern side of the Chapiquiña-Belén ridge. These rapid paleogeographic

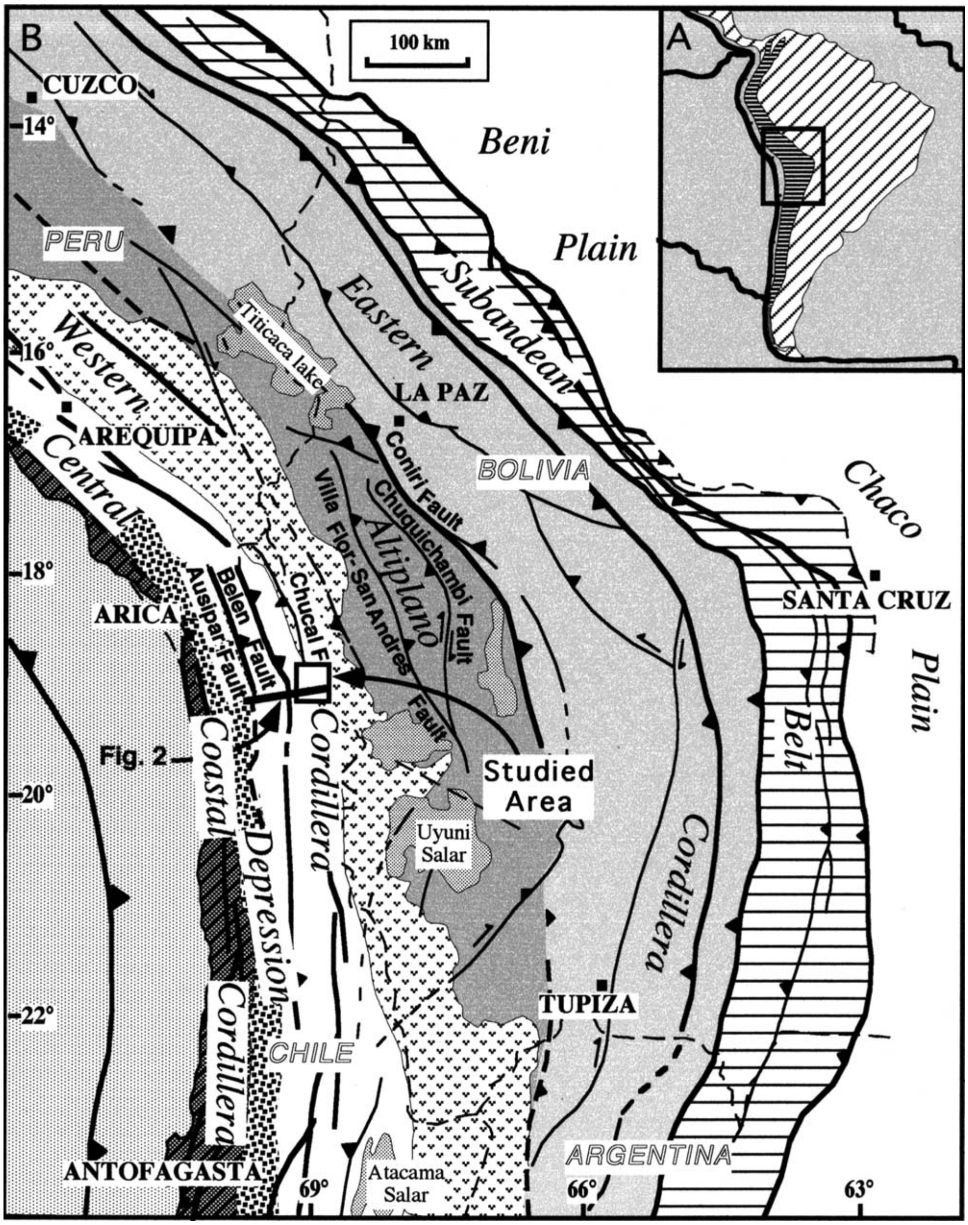

Fig. 1. Location map. (A) Location of the Bolivian orocline relative to South America and the Andes mountain range. Major oceanic features (trench and active ridges) are indicated. (B) Morphostructural units and major structural features of the Bolivian orocline, with the location of cross-section AB (Fig. 2) and map of the Chucal region (Fig. 3).

changes greatly influenced the sedimentation and environmental conditions and controlled the occurrence and preservation of a rich mammal fauna. The detailed chronology obtained in this region further enables us to infer the effects of this tectonic activity on the timing of Andean uplift, regional faunal and floral changes, and paleoclimate.

2. Geologic and tectonic setting of the study region

The study region is located at altitudes of 4000–5000 m in the Western Cordillera, a volcanic range developed on the western side of the Altiplano. The Altiplano is a morphostructural unit of the broader Andean Cordillera, located between approximately 15 and 27°S latitude. It consists of a high plateau with altitudes of generally 3500–4500 m above sea level. In this region, the majority of the Altiplano is located in Bolivia and Perú, and only a narrow, western section of it is located in Chile. Along the western side of the Altiplano in Chile and Bolivia, the Western or Volcanic Cordillera corresponds to the present-day volcanic arc. In this region, two morphostructural units occur: the central or longitudinal depression and, farther west, the Coastal Cordillera. Morphostructural units east of the Atiplano are the Eastern Cordillera and the sub-Andean belt (Figs. 1 and 2).

Along the western slope of the Altiplano, or the precordillera, a N–S- to NNW–SSE-oriented, high-angle,

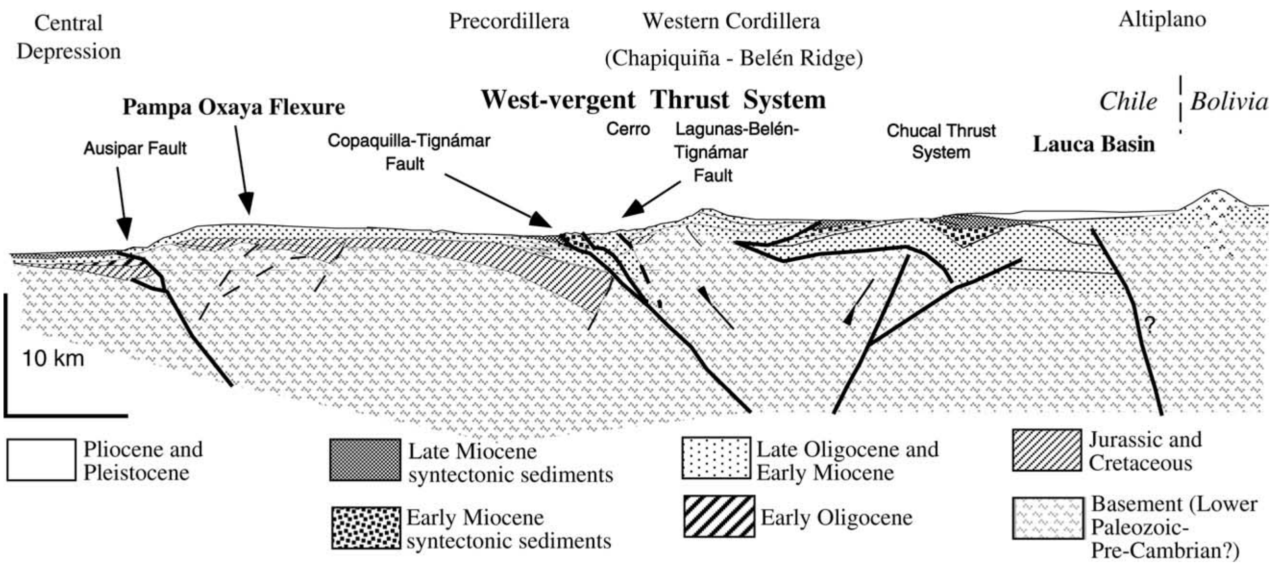

Fig. 2. East–west cross-section of the precordillera and Western Cordillera, or present-day volcanic arc, in northernmost Chile. West- and east-vergent thrust systems bounding the uplifted Chapiquiña-Belén ridge are indicated. The Ausipar fault to the west, part of west-vergent thrust system, separates the precordillera from the central depression. The Pampa Oxaya flexure is an anticlinal fold developed on the back side of the Ausipar fault.

westward propagating, west-vergent thrust fault system occurs (Muñoz and Charrier, 1996; García, 1996; García et al., 1996). This system comprises, from east to west, the Chapiquiña-Belén, Cerro Lagunas-Belén-Tignámar, and Copaquilla-Tignámar faults, as well as the Ausipar fault, which is located farther west (Muñoz and Charrier, 1996; García, 1996; García et al., 1996) (Figs. 1 and 2). The Chapiquiña-Belén fault thrusts pre-Cambrian?—Early Paleozoic metamorphic rocks of the Belén metamorphic complex over Cenozoic deposits. Activity along this thrust system is associated with syntectonic sedimentation. The resulting deposits are coeval with the described units from the Chucal-Lauca basin on the east side of the Western Cordillera.

West of both the precordillera and the extensive early Miocene ignimbritic cover that forms the Pampa Oxaya (Figs. 1 and 2), the major west-vergent Ausipar thrust fault separates the precordillera from the central depression (Figs. 1 and 2). A late movement in the late Miocene (~9–7.7 Ma) along this fault caused additional development of the extended Pampa Oxaya flexure (anticline) on its backside (García et al., 1999).

Tectonic activity along the west-vergent thrust system, which we associate with the uplift of the western side of the Altiplano, caused an approximate shortening of 14 km (García, 2001) and an almost, if not totally, continuous sequence of syntectonic deposits. The areal distribution and compositional features of these deposits are strongly variable and depend on the paleogeographic environment that exists at the moment of deposition, as well as the exposed preexistent units. On the basis of this and their stratigraphic position, the different syntectonic deposits recognized in this region can be separated into several units (García, 1996). Restricted to the eastern precordillera (Fig. 2), early and middle Miocene syntectonic deposits of the Joracane and Huaylas Formations are associated with activity along the Cerro Lagunas-Belén-Tignamar and Copaquilla-Tignámar faults, respectively (García, 1996, 2001; García et al., 1996) (Fig. 3).

The Joracane Formation (García, 1996) is an approximately 500–600 m thick, 25–30° east-dipping, conglomeratic series. It was deposited unconformably over the Lupica Formation. To the east, these deposits are thrust by the Lupica Formation, whereas to the west, they thrust over the Huaylas Formation (Fig. 3). Clasts are volcanic in composition, subordinately correspond to limestones, and derive from the Lupica Formation. No clasts of the Belén metamorphic complex are found, which suggests that this unit was not exposed at that time. The conglomerates were deposited in a weakly erosive fluvial environment and are organized in layers up to several meters thick. Sediment supply was from the east, probably from zones uplifted by the activity of the west-vergent Cerro Lagunas-Belén-Tignámar thrust fault (Fig. 3). Two K–Ar age determinations of biotite from tuffs intercalated in the conglomerates yield middle Miocene ages of 18.2 ± 0.8 and 16.8 ± 1.5 Ma.

The other syntectonic deposits associated with the tectonic activity in the eastern precordillera have been grouped in the Huaylas Formation (Salas et al., 1966; García, 1996, 2001; García et al., 1996). They consist of a thick, essentially horizontal, coarse- to medium-grained conglomeratic series with sandstone and limestone intercalations and few tuffaceous lenses. The thickness (500 m) rapidly diminishes and wedges out to the west on the east side of the Pampa Oxaya flexure (anticline). This unit is associated with activity on the west-vergent Copaquilla-Tignámar thrust fault (Figs. 2 and 3). East of the fault, the

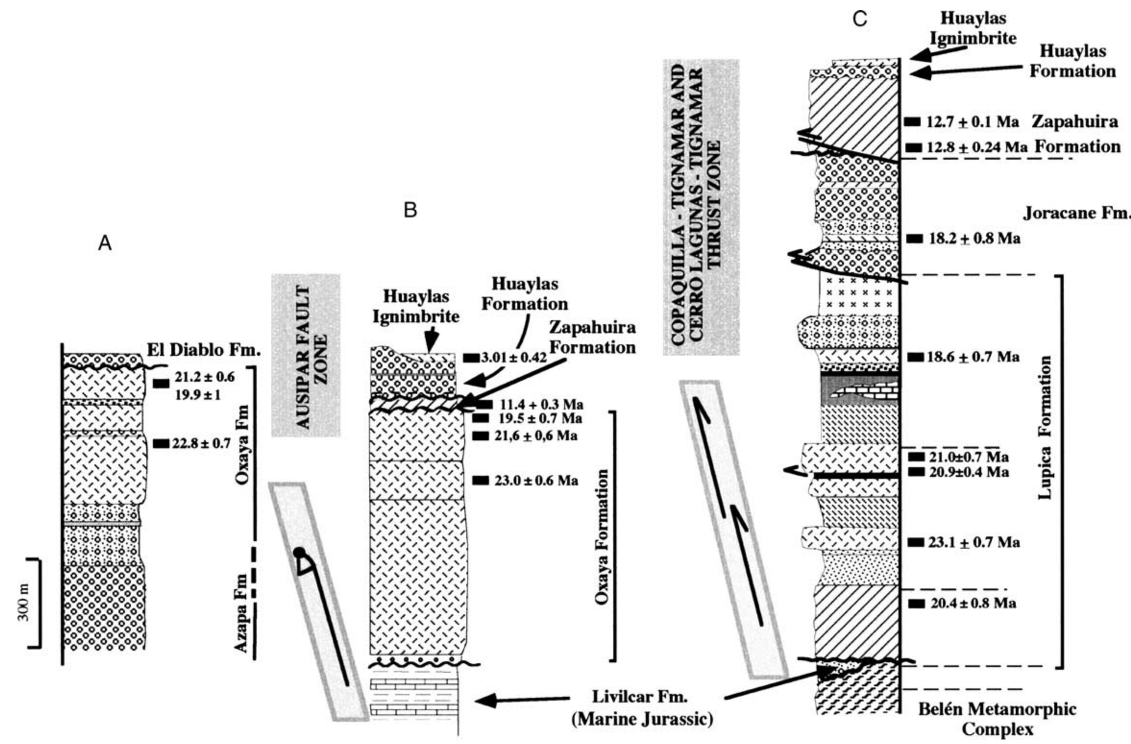

Fig. 3. Stratigraphic columns of Cenozoic deposits in the central depression and precordillera showing the east–west stratigraphic variations that permit the individualization of different paleogeographic environments during Late Cenozoic evolution. (A) The central depression west of the Ausipar fault; here, the essentially ignimbritic Oxaya Formation overlies the coarse clastic Oligocene Azapa Formation and is unconformably covered by the conglomeratic Miocene El Diablo Formation. (B) The easternmost flank of the Oxaya flexure (anticline) in the precordillera; here, the Oxaya Formation unconformably overlies Jurassic–early Cretaceous deposits (Livilcar Formation) accumulated in an extensional backarc basin and is covered by the volcanic Zapahuira Formation; the conglomeratic, syntectonic Huaylas Formation; and the Pliocene Huaylas Ignimbrite. (C) The eastern precordillera in the region of Belén (Fig. 1). The early Miocene Lupica Formation represents an eastern time equivalent of the Oxaya Formation unconformably covering Pre-Cambrian? or early Paleozoic metamorphic rocks and late Paleozoic? or Jurassic marine sediments. The early late Miocene Joracane Formation and the late Miocene Huaylas Formation correspond to syntectonic deposits associated with faults of the eastern west-vergent thrust system.

Huaylas Formation unconformably overlies the Lupica, Zapahuira, and Joracane Formations (Fig. 3). West of the fault, on the east flank of the Pampa Oxaya flexure (anticline), the lower levels of the Huaylas Formation conformably overlie either ignimbritic deposits of the Oxaya Formation (Salas et al., 1966) or the 11.4 ± 0.3 Ma old andesitic lavas of the Zapahuira Formation (García, 1996; García et al., 1996) (Fig. 3) and dip slightly to the east. The upper deposits of the Huaylas Formation are flat lying, prograde toward the west over the Oxaya Formation, and onlap this unit with a strong erosional unconformity (García, 1996; García et al., 1996). These upper deposits contain an ignimbritic intercalation that has been dated ($^{40}Ar/^{39}Ar$ on biotite) as 10.55 ± 0.07 Ma in the Tignámar area (Wörner et al., 2000) and as 10.8 ± 0.4 Ma in the Caragua area (García, 2001). The different relationships between the lower and upper parts of the Huaylas Formation and the underlying Oxaya and Zapahuira Formations indicate deformation on the east flank of the Oxaya anticline at approximately 11 Ma. These deposits contain fossil mammals (Nothoungulates) (Bargo and Reguero, 1989; Salinas et al., 1991; Flynn et al., 2002) that probably correspond to the Friasian/pre-Huayquerian South American land mammal ages (SALMAs) of the late Early Miocene–early Late Miocene (Flynn et al., 2005). The formation is unconformably covered by the 30 m thick Huaylas ignimbrite with K–Ar ages of 4.4 ± 0.3 to 4.8 ± 0.3 Ma (Naranjo and Paskoff, 1985). Recent authors have suggested that the Lauca ignimbrite (Lauca-Pérez ignimbrite of Wörner et al., 2000), which is exposed east of the Western Cordillera in the studied region, corresponds to the flow of the Huaylas ignimbrite (Wörner et al., 2000; García, 2001; García et al., 2002). On the basis of three $^{40}Ar/^{39}Ar$ dates on sanidine crystals (2.72 ± 0.01, 2.7 ± 0.2, and 2.4 ± 0.4 Ma) and one K–Ar date on biotite (3.0 ± 0.4 Ma) from the Lauca ignimbrite, these authors conclude that the Huaylas ignimbrite is late Pliocene in age. Therefore, deposition of the Huaylas Formation began before 10.55–10.8 Ma and ended before 2.7 Ma.

On the west flank of the Pampa Oxaya flexure (anticline) and in the central depression, other syntectonic series are extensively exposed, including the Azapa Formation (Salas et al., 1966) of Oligocene age (García, 2001) and the El Diablo Formation (Tobar et al., 1968). The El Diablo Formation contains at its base a fine-grained tuff deposit dated 15.7 ± 0.7 Ma (García, 2001) and is older than the

8.2 ± 0.7 and 9.0 ± 1.0 Ma old basalt flow that covers it (Naranjo and Paskoff, 1985; Muñoz and Sepúlveda, 1992; Muñoz and Charrier, 1996; Pinto, 1999; Pinto et al., 2004). The El Diablo Formation therefore is equivalent to the syntectonic deposits on the east side of the Western Cordillera.

Evidence from the precordillera and central depression (Fig. 2) indicates long-lasting tectonic activity that began in the Oligocene. The younger units (Joracane, Huaylas, and El Diablo) represent an almost continuous syntectonic depositional series beginning in the early Miocene (18.2 ± 0.8 Ma in the Joracane Formation) and continuing until the early Pliocene (Charrier et al., 1999, 2000; García, 2001). In the precordillera, along the southern prolongation of the Ausipar fault (Moquella flexure of Muñoz and Sepúlveda, 1992; Muñoz and Charrier, 1996; Pinto, 1999) 100 km south of the Quebrada Azapa, syntectonic conglomerates associated with progressive unconformities are covered by a 21.7 ± 0.6 Ma old ignimbrite (Suca ignimbrite in the Camiña member of the Latagualla Formation of Pinto, 1999; Pinto et al., 2004), which indicates a still older age for the beginning of the tectonic activity associated with the west-vergent thrust system along the precordillera.

At the same time in the Western Cordillera, deformation developed along a NNW–SSE-trending, east-vergent thrust system: the Chucal thrust system (Hérail and Riquelme, 1997; Riquelme and Hérail, 1997; Riquelme, 1998; García, 2001) (Fig. 2). This system is associated with progressive unconformities and syntectonic sedimentation. Because the two thrust systems have opposite vergencies, tectonic activity developed an uplifted block—the Chapiquiña-Belén ridge—between them (Charrier et al., 1999, 2000) (Fig. 2). The contractional episode and thrusting began in the early Miocene, at approximately 18 Ma, continued until the Pliocene, and probably remains active. The resulting late Cenozoic compressive structures control the present-day N–S-oriented relief of various structural/topographic highs.

The east-vergent structural system is located on the east side of the fault-controlled Chapiquiña-Belén ridge (Charrier et al., 1999, 2000) in the altiplano of Arica, adjacent to the Salar de Surire (Figs. 2 and 4). This system comprises the Jaropilla fault, which is covered (sealed) by the Quebrada Macusa Formation that constrains its most recent activity, as well as the fault-propagated Chucal anticline (Riquelme and Hérail, 1997; Riquelme, 1998) (Fig. 4). It deformed lower–upper Miocene rocks regionally and Pliocene rocks locally and caused the development of several progressive unconformities, mainly on the east flank of the anticline (Hérail and Riquelme, 1997; Riquelme and Hérail, 1997; Riquelme, 1998; Charrier et al., 1999, 2000; Chávez, 2001). The Lupica, Chucal, and Quebrada Macusa (Estratos de Quebrada Macusa of Riquelme, 1998) Formations, which form a superpositional stratigraphic series ranging in age from 21.7 ± 0.8 to 10.4 ± 0.7 Ma (dated horizons within the lowermost and uppermost units), are all deformed. The latter age corresponds to that of an undeformed lava of the Anocarire volcano that covers the upper brown tuff of the Quebrada Macusa Formation (Riquelme, 1998) (Fig. 5).

New detailed stratigraphic and structural studies on an E–W section across the east-vergent Chucal anticline (Chávez, 2001), radioisotopic dates of tuff layers (García, 2001; Bond and García, 2002), and recent discoveries of mammal fossils (Flynn et al., 2002) enable us to better constrain the chronology and deformation rate of the Chucal anticline, as well as of the development of the progressive intraformational unconformities, and better understand the effects of deformation on the stratigraphic sequence, sedimentation, and paleoenvironmental evolution of the Chucal-Lauca basin, as well as their possible significance for the timing of Andean uplift.

3. Geology of the Chucal region

In the study region, in contrast to the situation in the precordillera (Muñoz et al., 1988; García, 2001), no pre-Cambrian?—early Paleozoic metamorphic rocks (Montecinos, 1963; Pacci et al., 1980; Basei et al., 1996; García, 1996; Lezaun et al., 1996; Lucassen et al., 1996; Lezaun, 1997; Heber, 1997), Paleozoic sediments (García, 1996, 2001; Lezaun et al., 1996; Lezaun, 1997), or Mesozoic rocks have been found. The oldest exposed deposits in the study region correspond to the Lupica Formation (Riquelme, 1998), a sedimentary and volcano-clastic unit exclusively exposed south of the study region (Figs. 4 and 5). The core of the Chucal anticline is formed by a massive white tuff and a thin, white, tuffaceous, fluvial or distal alluvial series. Southeast of the anticline axis, younger levels of the Lupica Formation are exposed. Unconformably overlying the Lupica Formation on both flanks of the anticline are developed fluvial, lacustrine, and volcanoclastic deposits of the Chucal Formation. The sedimentary Chucal Formation is in turn covered by a less deformed, thick tuffaceous series assigned to the Quebrada Macusa Formation (Riquelme, 1998). On the east flank of the Chucal anticline, the slightly deformed Lauca Formation (Muñoz, 1988) unconformably overlies the Quebrada Macusa Formation.

The > 800 m thick deposits of the uppermost Lupica, Chucal, and Quebrada Macusa Formations have been affected by deformation of the NNW–SSE-oriented, 10–15° northward-plunging Chucal anticline, which is part of the east-vergent Chucal system.

3.1. The Lupica Formation: the core of the Chucal anticline (Fig. 5)

The core of the Chucal anticline is formed by a > 50 m thick, massive, white–gray tuff (Riquelme, 1998) covered by a stratified series of white sandstones. The massive white–gray, crystalline, and vitreous ash tuff comprises

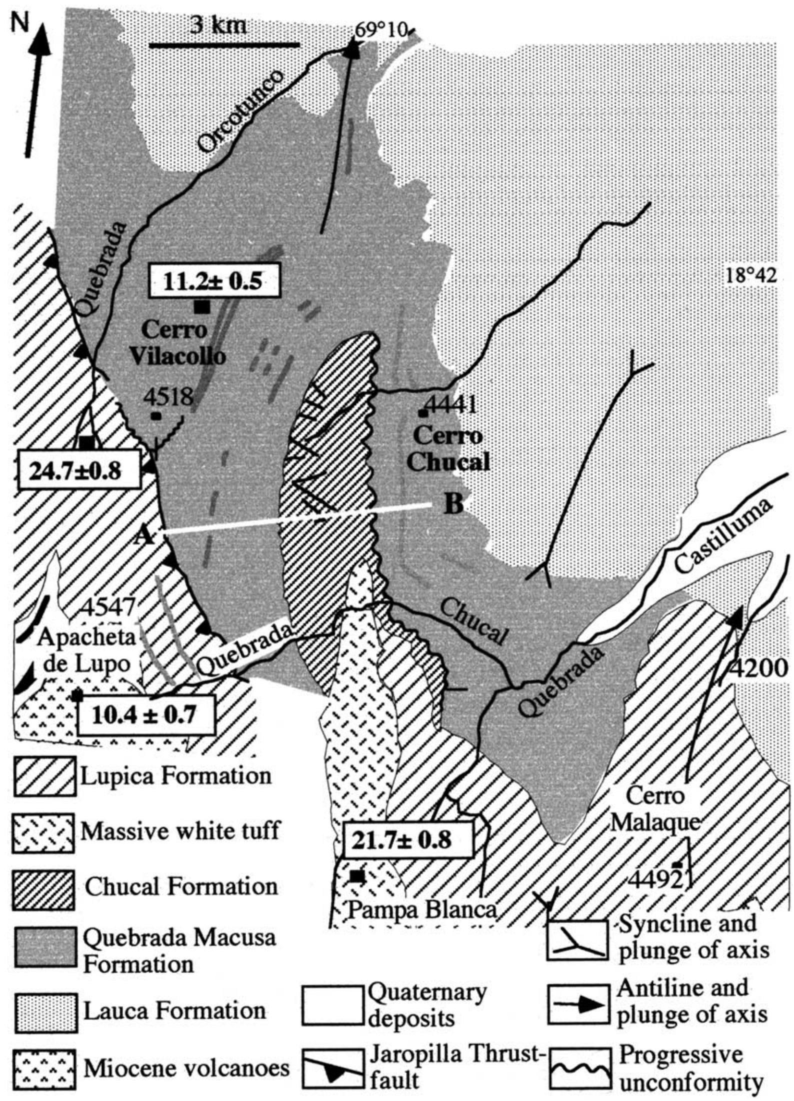

Fig. 4. Geologic map of the Chucal region in the Western Cordillera showing the major structural features associated with the east-vergent thrust system. The east flank of the $\sim 15°$N-plunging Chucal anticline registers progressive unconformities separating the different formations (Lupica, Chucal, Quebrada Macusa, and Lauca) and members inside the Chucal Formation. The location of structural cross-section of Fig. 7 is indicated.

subhedral and euhedral quartz, feldspar, and subordinated biotite crystals in a vitreous ash mass. It contains subspheric pumice fragments and shows slight chlorite alteration.

The stratified series that covers the massive white–gray tuff is formed of thinly stratified, white–gray, richly tuffaceous sandstone layers, most of which are tabular, though some are lenticular and finely conglomeratic at the base. Frequent thin, white–gray tuff intercalations are present. Near the top, this series contains a 15 m thick, gray–white ignimbritic tuff (CH-5). The thinly stratified white–gray series is 66 m thick on the west flank and approximately 50 m thick on the east flank. This series is paralleled on each flank, though with an erosional discontinuity, by the Chucal Formation (Muñoz, 1991), as we describe next (Fig. 5). Decimeter-deep paleochannels and cross-bedding are present. Paleochannels are E–W- and NE–SW-oriented, and cross-bedding indicates a sense of transport from E and NE. Grains in the sandstones are angular with low sphericity; they consist of plagioclase, quartz, and lithics (cineritic tuff).

We interpret the thinly stratified white–gray series as having been deposited in a fluvial or distal alluvial fan environment. After deposition, these deposits were exposed to climatic and biologic agents, weathered, and partially eroded. The upward-thinning tendency suggests a gradual diminution of the energy of the transport agent.

A whole-rock, ^{39}K–^{40}Ar analysis of the massive white–gray tuff underlying the fine fluvial or distal alluvial sediments yields an age of 21.7 ± 0.8 Ma (Riquelme, 1998), which is in agreement with the ages obtained for the Lupica Formation in the precordillera (Fig. 3). On the basis of the tuffaceous character of this series and the weathering and erosional episode that followed its deposition, we assign it to the upper part of the Lupica Formation.

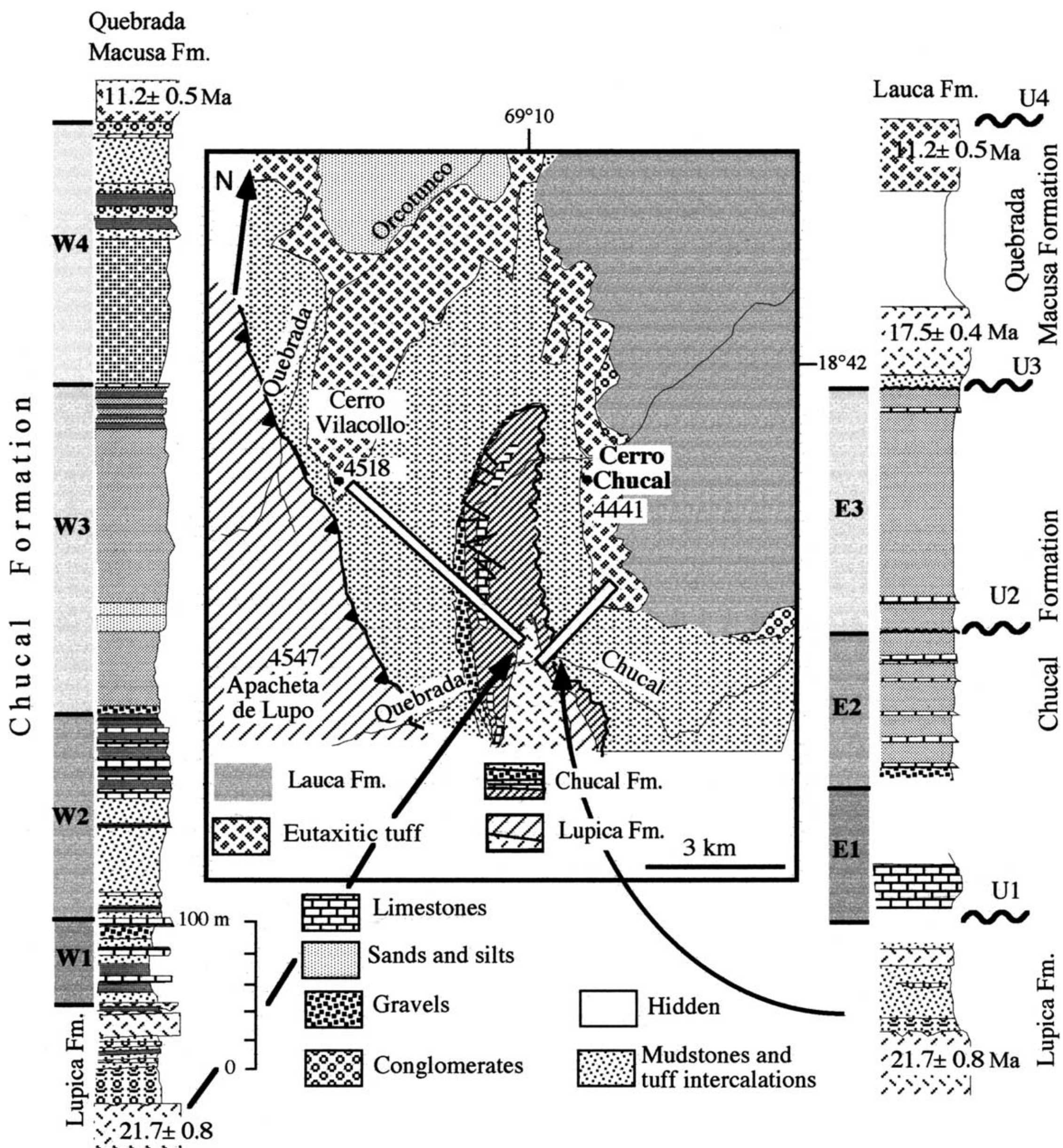

Fig. 5. Stratigraphic columns of the west and east flanks of the Chucal anticline, comprising the uppermost Lupica Formation, the Chucal and Quebrada Macusa Formations, and the basal Lauca Formation and indicating the stratigraphic position of progressive unconformities and radioisotopically dated horizons.

3.2. The Chucal Formation

The Chucal Formation parallels the tuffaceous, fluvial, or distal alluvial deposits assigned to the upper Lupica Formation on both flanks of the Chucal anticline. However, it is separated from this series by a weathering and erosional surface. The Chucal Formation is formed by several members, some of which are developed only on one of the flanks of the anticline.

3.2.1. The Chucal Formation on the west flank of the anticline (Fig. 5)

The Chucal Formation is 600 m thick on the west flank. It forms a west-dipping (35–40°), continuous series covered by the Quebrada Macusa Formation. We subdivide it, from bottom to top, into four informal members: W1, W2, W3, and W4.

Member W1 is 60 m thick and begins with a lenticular, orange, channelized paraconglomerate. The rest of the member is formed by three upward-fining sequences, approximately 20 m thick, that are composed of decimeter-thick, medium- to fine-grained sandstone layers with tuffaceous matrix, fine ash tuffs, and mudstones. The final series is 4–6 m thick brown and gray limestone. Some sandstone layers are laminated; others are calcareous. Some mudstone layers are calcareous and occasionally contain badly preserved plant remains. The limestone series are formed by centimeter- to decimeter-thick, tabular micritic levels that alternate with marls with a finely ondulated lamination. The orange color becomes more prevalent upward. Badly preserved plant remains and ostracodes have been found in this member. The orientation of one paleochannel is N20°E; no paleocurrent data are available for this member.

The fine-grained sediments indicate that deposition occurred in essentially low energy conditions. Facies associations indicate a transition from a probable distal fluvial plain to a lacustrine environment. This transition was repeated three times, as indicated by the sequences that each culminate with limestones. The paleochannel orientation may indicate a paleogeography that differs from the one that prevailed during the upper Lupica Formation. The tuffaceous intercalations indicate that volcanic activity continued throughout deposition of this member.

Member W2 is 144 m thick, white, and composed of several upward-fining sequences of variable thickness. It is formed by fine-grained sandstones, some with tuffaceous matrix, gray mudstones, and white limestones. Mudstones and limestones are more abundant toward the top. Limestones form decimeter- to several meters-thick intercalations throughout the series that are composed of alternations of laminated marls with black chert nodules, massive marls, fine calcareous sandstones, and thin micritic layers. Well-preserved plant remains occur in the laminated marly intercalations.

This mostly white-colored unit covers the third 4–6 m thick limestone series that culminates in the upward-fining sequence of Member W1. Environmentally, it corresponds to the same facies associations. However, its characteristics indicate that lacustrine conditions prevailed throughout it and more commonly than in Member W1.

Member W3 is 225 m thick. It is characteristically and distinctively green in color and predominantly composed of sandstones. Its base is the first green layer above member W2. It corresponds to an alternation of coarse- to fine-grained massive sandstones, coarse to conglomeratic sandstones, mudstones (frequently calcareous), and a few mostly thin limestone intercalations (marls and micrites). Some sandstone beds are lenticular; others contain abundant tuff material. Sporadic tuff intercalations were found, some of which contain accretionary lapilli. Some limestone layers at the top weather to a characteristic external red color. Poorly preserved plant remains are frequent in this member.

Facies associations indicate the renewal of a fluvial plain environment, representing the filling of the lake episode of Member W2. In the surrounding areas, explosive volcanism continued during this time.

Member W4, the uppermost member of the Chucal Formation on the west flank of the anticline, is 172 m thick. It is characteristically composed of brownish-gray mudstones in its lower part, similar to those observed in Member W2. Its uppermost part contains cross-bedded sandstones, some minor conglomerates, and thick intercalations of fluvially transported ashes.

The predominance of mudstones in its lower parts indicates a floodplain environment. The intercalation of coarser deposits in the upper portions indicate a transition to well-developed fluvial conditions. The thicker tuffaceous intercalations indicate an increase in explosive volcanism.

3.2.2. Facies evolution on the west flank

The west flank deposits form a continuous series that reflects the following four-stage environmental evolution: (1) fluvial or distal alluvial conditions in the upper Lupica Formation to fluctuating lacustrine conditions in Member W1 and well-developed lacustrine conditions in Member W2, (2) filling of the lake and expansion of the fluvial plain environment in Member W3, (3) renewed floodplain deposition in lower Member W4, and (4) fluvial (distal alluvial plain) conditions accompanied by an abundant supply of ash material derived from nearby explosive volcanic sources in the upper part of Member W4.

On the basis of this evolution on the west flank of the anticline, we deduce two stages of accommodation space creation for the Chucal Formation in the studied stratigraphic column: (1) during deposition of Members W1 and W2 when a lake developed and (2) principally during deposition of the lower part of Member W4. However, basin development is better represented in Member W2 than Member W4. These two stages of basin development are separated by a stage of increased transport energy and filling of the generated space during deposition of Member W3.

3.2.3. The Chucal Formation on the east flank of the anticline (Fig. 5)

On the east flank of the Chucal anticline, the Chucal Formation can be subdivided into three members, from bottom to top: E1, E2, and E3. The maximum thickness of the Formation in this flank is 365 m. The Chucal Formation unconformably overlies the thinly stratified white series of the Lupica Formation and is unconformably overlain by the Quebrada Macusa Formation (Riquelme, 1998). Next to the anticline axis, the dip is vertical and diminishes gradually to the east; immediately below the overlying Quebrada Macusa Formation, the dip is 35° east.

Member E1 is represented by a poorly exposed (no outcrops of its lower and upper contacts) white limestone series, formed by decimeter-thick micritic layers that contain black chert nodules and thin, gray marly intercalations. The exposed thickness of Member E1 is approximately 30 m. The gaps between the underlying white tuffaceous sandstones of the Lupica Formation and the overlying green sandstone Member E2 (Fig. 5) total 100 m, which may be its maximum possible thickness. No deposits that could be correlated to Member W1 have been observed on this flank between the thinly stratified white sandstones of the upper Lupica Formation and Member E1.

Member E2 is 100 m thick and characteristically green in color. It contains fine to coarse, often laminated, lenticular sandstones (calcareous and conglomeratic); sporadic, decimeter-thick limestone intercalations with an external red color; massive, several meters thick, brown and green mudstones; and green ash tuff intercalations with abundant chlorite. Layers in this member are vertical.

Member E3 is 165 m thick and consists of predominantly gray (less frequently brown), generally massive, medium- to coarse-grained sandstones with cross-bedding and fine conglomeratic breccias with hard, concretionary, calcareous horizons, possibly related to paleosoil development. Some thin, green sandstone intercalations occur just above its base. Lithic components consist of ash tuff, pumice and porphyritic andesite fragments, and plagioclase and pyroxene crystals. Although cross-bedding could not be measured, it indicates a general eastward transport direction. The contact with the underlying Member E2 is slightly unconformable, and E3 dips less steeply than Member E2. It has no equivalent on the west flank. Abundant mammal fossils have been found in Member E3.

3.2.4. Facies evolution on the east flank

The deposits on this side of the Chucal anticline reflect a three-stage evolution of paleoenvironmental conditions: (1) probable lacustrine conditions in Member E1, (2) lake filling and fluvial plain systems with possible lacustrine influence in Member E2, and (3) an abrupt intensification of the fluvial plain conditions in Member E3.

On the basis of this evolution, it is possible to deduce two stages of accommodation space development: (1) during deposition of Member E1 with the probable development of a lake and (2) during deposition of Members E2 and E3, when a considerable increase in transport energy took place.

3.2.5. Correlation of and comparison between the west and east flank series

The absence of exposures on the northern periclinal termination (because the plunge of anticlinal axis is to the north, there is no southern periclinal termination in the study region) precludes directly tracing the members of the east and west sides into one another. The criteria adopted to correlate the members of both sides of the anticline therefore are their relative stratigraphic position, the bracketing Lupica and Quebrada Macusa Formations, and their lithology (limestone for Members W2 and E1; green sandstone for Members W3 and E2). Considering the difficulty of correlating deposits from separated syntectonic subbasins, even when they are close to one another, the correlation that follows should be considered tentative.

The white, tuffaceous, fluvial or distal alluvial series of the Lupica Formation, which covers the thick, white–gray tuff, is developed on both flanks. The exposed thickness of this series is thicker on the west than on the east flank.

Member W1 of the Chucal Formation is apparently or practically not developed on the east flank. Alternatively, the white calcareous Member E1 may be compared with one of the brown and gray limestone series that culminates the three upward sequences of Member W1. However, the white color of the limestones in Member E1 and the presence of black chert nodules relates them to Member W2 rather than to Member W1.

The white calcareous Member W2 can be lithologically and stratigraphically correlated with the similarly white and calcareous, though thinner, Member E1, as well as the underlying green fluvial sandstones. E1 is developed above the white, fluvial or distal alluvial, tuffaceous series of the Lupica Formation.

The green sandstone Member W3 can be compared to Member E2. The nature of the deposits and their characteristic green color, together with their stratigraphic position above similar white limestone deposits of Members W2 and E1, respectively, supports this correlation. The green sandstones are considerably thicker, however, on the west (225 m thick) than on the east flank (100 m thick), which suggests either longer duration of the paleoenvironmental conditions responsible for its deposition or more rapid subsidence and greater sediment influx on the west side.

Mudstone Member W4 has no lithologic equivalent on the east flank. With its position between the green sandstone Member W3 and the lower white tuff at the base of the Quebrada Macusa Formation, it could be stratigraphically and chronologically correlated with all or part of Member E3 on the east flank. However, Member E3 represents an increase of transport energy not represented on the west flank, except at the uppermost portion of Member W4. Therefore, it could be an eastern representative of this phase of increased energy or simply correspond to a somewhat younger deposit not represented on the west flank.

3.2.6. Paleoenvironmental transformation during deposition of the Chucal Formation

The different paleogeographic histories observed in the stratified series developed on both flanks of the Chucal anticline may result from a contractional episode on the Chucal thrust system, probably with fluctuating deformation rates, associated with variable erosion and sedimentation rates. The existence of two distinct depositional series on either side of the anticline axis reveals different paleogeographic and paleoenvironmental histories in two minimally interconnected basins/subbasins. The west flank stratigraphic column apparently is continuous, whereas the east flank column is interrupted by at least three exposed unconformities.

A major change in the depositional regime occurred in the basin after the deposition of the Lupica Formation at the base of the Chucal Formation with the deposition of the basal conglomerate of the orange Member W1, which is exposed only on the west flank. According to the minimal evidence of the sense of transport (from the E and NE) of the white, tuffaceous, fluvial or distal alluvial deposits of the Lupica Formation, the slope of the basin floor—which is next to the present-day location of the anticline axis before the deposition of Member W1—was to the W and SW. This slope may indicate a previous paleogeographic feature or the development of a swell along the axis of the anticline before the start of deposition of the Chucal Formation.

On the east flank, the absence of deposits that are lithologically similar to Member W1 and the smaller thickness of the white calcareous Member E1, which contains limestone types similar to those in Member W2, may indicate that (1) erosion or sediment bypass prevailed on that side of the anticline at that moment and that no deposition occurred, (2) an equivalent to Member W1 was deposited to the east of the anticline axis but eroded later, or (3) Member E1 is the time equivalent of Member W1. The first explanation indicates the existence of a swell that controls the east margin of the western subbasin and the absence of sedimentation next to the east side of the swell. The second situation suggests that the uplift of the swell occurred after the deposition of the orange deposits and that this uplift caused erosion of these deposits on the east side of the anticline axis. The third possibility indicates the existence of a swell that separates two active depocenters with completely different facies associations at the same time.

On the west side of the anticline axis, there was a clear tendency toward the development of accommodation space that resulted in a lacustrine basin, as shown by the development of the 60 m thick Member W1. As Member W1 consists of three upward-fining cycles, each of which ends with a well-developed limestone series, it suggests episodic variation in sediment influx and/or the size of the depocenter. This sedimentation space may have been controlled by a topographic high that formed farther west and the probable development of a swell along the anticline axis to the east.

This tendency to form an accommodation space culminated on the west side of the anticline axis during the deposition of the white, calcareous Member W2. The thickness of the lacustrine-dominated deposits on the west flank, represented by members W1 and W2, totals 200 m. On the east flank, lacustrine environments may have been present at the time of deposition of members W1 and W2, as is revealed by the development of the lithologically similar, thinner (<100 m thick) Member E1. As indicated previously, it is impossible to determine from the available information whether the lacustrine deposition on the east flank (Member E1) corresponds to a gradual eastward transgression of the lacustrine facies over a west-dipping surface or if a second basin already developed on this side, separated from the western basin or subbasin by a low swell.

Slowing of basin development and rapid filling of the basin or subbasin occurred on both sides of the anticline during or initiated by the deposition of the green sandstones of Members W3 and E2. It is difficult to assess if a single depocenter was present across the region or if two coexisting depocenters were separated by a swell. The absence of exposures bracketing the contact between the white, calcareous Member E1 and the overlying green sandstone Member E2 at the northern (periclinal) end of the anticline precludes any conclusion about the nature of the contact or transitions in paleoenvironments.

The considerably different environments that developed on both sides of the present-day location of the anticline axis following the deposition of the green sandstones clearly indicate that two different depositional domains were present, separated by a swell. On the west side, a renewed floodplain deposition (lower part of Member W4) began to develop. On the east side, fluvial deposition continued but with higher transport energy.

3.2.7. Unconformities

As described previously, the west flank series of the Chucal Formation was deposited rather continuously, whereas the east flank series was interrupted by erosional or angular unconformities (Fig. 5). This difference indicates that the environmental evolution on the east flank was controlled by modifications caused by the tectonic activity. In addition to the erosional discontinuity separating the Lupica from the Chucal Formation, the observed and inferred unconformities on the east flank (see Chávez, 2001) are, from bottom to top (Fig. 6), as follows:

1. U1: This unconformity is inferred on the basis of the probable lack of deposition or total erosion of the orange member on the east flank and the deposition of the white, calcareous Member E1, which we consider an equivalent of Member W2 rather than of Member W1, directly over the upper Lupica Formation;
2. U2: The slight change in the dip angle between the almost vertical green sandstone Member E2 and

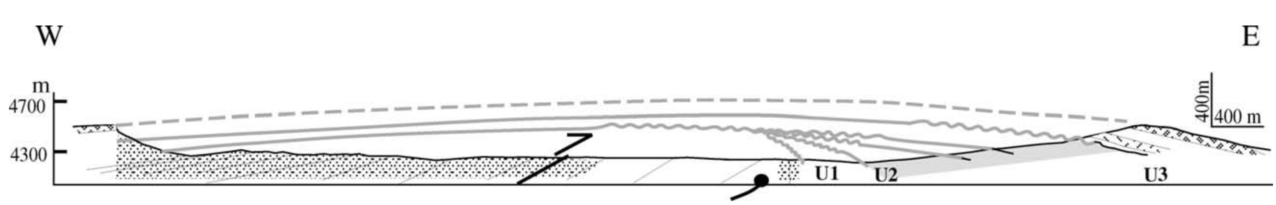

Fig. 6. East–west structural cross-section of the Chucal anticline (for location, see Fig. 4). The more steeply dipping east flank registers four observable progressive unconformities (U1–U4). Considering the general geometry of growth deposits and the low level attained by erosion, some progressive unconformities that developed during the growth of the Chucal anticline are not exposed. A blind propagation fault is indicated in the core of the anticline.

the overlying coarser Member E3 mark the presence of this unconformity; and

3. U3: It is indicated by the >20° difference in dip between the coarse sandstone Member E3, which forms the upper deposits of the Chucal Formation, and the overlying Quebrada Macusa Formation (Fig. 7).

3.2.8. Chronology of deformation

According to this evidence, the tectonic events that controlled the paleogeographic evolution of the study region probably began during the later depositional stages of the Lupica Formation (prior to 18–20 Ma in the early Miocene). The first tectonic movements, as revealed by the transport directions of the white, tuffaceous, fluvial or distal alluvial series that forms the upper Lupica Formation in this region, probably caused the development of the west-dipping slope over which these sediments were transported and deposited. This west-dipping slope may correspond to the west flank of a ridge or swell that formed along the present-day location (and likely paleoposition) of the anticlinal axis.

The second tectonic event is indicated by the deposition of the basal conglomerate of the Chucal Formation on the west flank, which likely reflects uplift in the basin borderlands, followed by a gradual tendency to develop a lacustrine basin, which reflects more rapid subsidence in the basin center and/or highland erosion or reduced sediment influx. During this tectonic event, a possible uplift movement farther west, possibly associated with the Jaropilla fault, appears to be a reasonable explanation for the development of the west margin of the basin. A tectonic event along or east of the anticlinal axis that might be responsible for U1 could explain the formation of the eastern subbasin. However, as we indicated previously, no information is available from the east flank to provide a definitive conclusion regarding the location of an eastern margin for this subbasin. Increased accommodation space

Fig. 7. Southward view of U3, which separates steeply dipping deposits of Member E3 of the Chucal Formation on the east flank of the Chucal anticline (A) from the moderately east-dipping, basal, white tuffaceous levels of the Quebrada Macusa Formation (B). In the back, the Miocene Anocarire volcano covers the Cenozoic stratified series.

developed during the deposition of the white calcareous Members W2 and E1, which suggests that tectonic movements controlling central basin subsidence continued throughout this time.

After the deposition of the green sandstone Members W3 and E2, significant tectonic movements along the axis of the anticline caused deformation and formation of an unconformity (U2) on the east flank. The green sandstones of Member E2 dip more steeply than the overlying, coarser sandstone series of Member E3. After these tectonic movements began on the west flank, the second stage of accommodation space creation occurred with the deposition of Member W4.

The last tectonic movement detected in the deposits of the Chucal Formation caused further steepening of both the green sandstone strata (Member E2) and the overlying sandstone Member E3, as well as the unconformable deposition (U3) over them of the basal deposits of the Quebrada Macusa Formation. Probably during this last tectonic movement, deposition of coarser intercalations in the upper part of the mudstone Member W4 began.

Episodes of deformation, associated with the development of a ridge or swell separating the west and east subbasins, and steepening of the east flank strata are related to anticline growth. Growth of the anticline presumably is associated with further movement of an east-vergent, blind thrust fault. Consequently, the east flank deposits of Member E3 and possibly part of Member E2 are interpreted as fault-controlled growth deposits.

Unconformities such as U2 and U3, which cause the differences in dip between Members E2 and E3 and between Member E3 and the basal deposits of the Quebrada Macusa Formation, dip steeper toward the anticline core. Together with their presence on the east or frontal flank of the anticline and their absence on the west or back flank of the anticline, this dip indicates that they are associated with anticlinal growth and therefore correspond to syndepositional, progressive unconformities (see Riquelme and Hérail, 1997; Riquelme, 1998; Charrier et al., 1999, 2000; Chávez, 2001).

3.3. The Quebrada Macusa Formation: the cap of the Chucal anticline

The Quebrada Macusa Formation in the Chucal region forms an approximately 200 m thick series comprised largely of light-colored, thick, massive ignimbritic deposits with minor lithic tuff, tuffaceous sandstones, and fine- to medium-grained, clast-supported conglomerate intercalations (Riquelme, 1998).

The strata of the Quebrada Macusa Formation are well exposed on both flanks of the Chucal anticline and form an ample anticline or draping blanket that envelops the deposits of the Chucal Formation. The contact with the Chucal Formation on the west flank is conformable, whereas on the east flank, it is unconformable. This difference reflects the stronger deformation of the eastern flank compared with the western flank, as well as the east-vergent deformation of the Chucal thrust system (Figs. 5 and 6).

Riquelme (1998) observes that on the west flank, the strata of the Quebrada Macusa Formation gradually decrease their steepness with decreasing age, which indicates that deformation occurred during deposition and probably stopped after deposition.

Two $^{40}Ar/^{39}Ar$ (biotite) dates of the basal white tuff of the Quebrada Macusa Formation on the east flank give ages of 16 ± 3 and 17.5 ± 0.4 Ma (García, 2001; Bond and García, 2002). Three ^{39}K–^{40}Ar dates constrain the age of its upper part to 11.2 ± 0.5 Ma (whole rock) on the upper eutaxitic, brown tuff (Chucal ignimbrite); 9.6 ± 0.7 Ma on amphibole crystals; and 10.4 ± 0.7 Ma on an andesitic lava from the Anocarire volcano that unconformably covers the Quebrada Macusa Formation south of Quebrada Chucal and Cerro Chucal. The age of the Chucal Formation is therefore late Early–Middle Miocene, consistent with the age information provided by its diverse mammalian fauna. The abundance of ash deposits in the Quebrada Macusa Formation indicates that after deposition of the Chucal Formation, explosive volcanism increased considerably.

3.4. The Lauca formation: Eastward shift of the syntectonic basin during the Late Miocene–Pliocene

These deposits originally were included in the Huaylas Formation by Salas et al. (1966) and Viteri (1979). Muñoz (1988; quoted first by Aguirre, 1990) informally separated these deposits from the Huaylas Formation and created the new stratigraphic Lauca Formation. Since then, subsequent authors studying the Arica Altiplano have referred to these deposits as the Lauca Formation (Aguirre, 1990; Bisso, 1991; Gröpper et al., 1991; Kött et al., 1995; Muñoz and Charrier, 1996; Riquelme, 1998; Gaupp et al., 1999; García, 2001).

The essentially flat-lying Lauca Formation fills an elongated, $\sim 3000\ km^2$, NNW–SSE-oriented basin. The basin developed between the Nevados de Putre or Taapacá volcano (probably farther north of Parinacota) and the Salar de Surire, on the westernmost side of the altiplano in Chile and western Boliva (Salas et al., 1966; Viteri, 1979; Muñoz, 1991; Muñoz and Charrier, 1996; Kött et al., 1995). The Lauca River runs along this basin in Chile, cutting its valley through the Lauca deposits, and then flows into Bolivia.

On the east flank of the Chucal anticline (Fig. 5), east-dipping basal deposits of the Lauca Formation cover, with a slight angular unconformity (U4), the steeper and east-dipping, approximately 10 Ma old upper tuff of the Quebrada Macusa Formation. However, the eastward continuation of the Lauca deposits is flat-lying and covers Miocene ignimbrites and lavas (Kött et al., 1995). On the eastern flank of the Chucal anticline, the Lauca Formation begins with a coarse conglomerate of boulders of ignimbrite, andesite, and basaltic-andesite. In the inner part of the basin, the Lauca Formation consists of an up to 300 m thick sedimentary series (Muñoz and Charrier, 1996), of which 120 m are lacustrine (Kött et al., 1995). The lacustrine deposits consist of predominantly light brown and yellowish, fine-grained sandstones and mudstones, frequently tuffaceous and sporadically calcareous, with lesser amounts of evaporitic lacustrine deposits that contain a few conglomeratic sandstone intercalations. These deposits contain ostracodes and diatoms. For a detailed description of these deposits and the evolution of the Lauca paleolake, see Kött et al. (1995).

It is difficult to determine a precise age for the base of the Lauca Formation. Kött et al. (1995), on the basis of the onlapping relationship of Lauca deposits over the 6.6 ± 0.2 and 7.0 ± 0.2 Ma old lavas of the Lauca volcano (Wörner et al., 1988), call this the age of its base. However, considering its superposition above older rocks of the Quebrada Macusa Formation immediately east of the Chucal anticline (Riquelme, 1998; García, 2001), the age of its basal deposits may be older. In this case, the observed onlapping at the Lauca volcano would represent later-stage basin infilling and onlap, and the 7.0–6.6 Ma old lavas of the Lauca volcano would represent the upper part of a series of unexposed lavas from this and other volcanoes that interfinger with the Lauca deposits. The basal Lauca Formation deposits observed immediately above the Quebrada Macusa Formation probably correspond to what Kött et al. (1995) consider part of the Huaylas Formation, which is the name originally given by Salas et al. (1966) and Viteri (1979) to all coarse-grained fluvial deposits of possible Plio-Pleistocene age in the precordillera and western Altiplano (Western Cordillera). We limit the name 'Huaylas Formation' to the predominantly coarse-grained, syntectonic deposits that accumulated in the precordillera, west of the Copaquilla-Tignámar thrust fault, to which they probably are genetically related.

A conspicuous ignimbritic intercalation located above the middle part of the Lauca Formation (Lauca ignimbrite of Kött et al., 1995; Muñoz and Charrier, 1996; Lauca-Pérez ignimbrite of Wörner et al., 2000), from its exposures in the Lauca Basin, provides three $^{40}Ar/^{39}Ar$ dates on feldspar crystals: 2.67 ± 0.25, 2.32 ± 0.18, and 2.88 ± 0.13 Ma (Kött et al., 1995, Table 2; Wörner et al., 2000) and a ^{39}K–^{40}Ar date on whole rock: 2.3 ± 0.7 Ma (Muñoz and Charrier, 1996). On the basis of this age of roughly 2.7 Ma, this intercalation has been correlated with the Pérez ignimbrite by Kött et al. (1995), Riquelme (1998), and Wörner et al. (2000). The Pérez ignimbrite is well known in the Bolivian part of the altiplano (Evernden et al., 1977; Lavenu et al., 1989; Marshall et al., 1992), and its occurrence in Chile suggests an even broader extent. Below this ignimbrite, a basaltic-andesite cinder cone (El Rojo Norte) gives a $^{39}K/^{40}Ar$ age of 3.1 ± 0.2 Ma (Kött et al., 1995).

The age of the Lauca Formation thus is constrained by the approximately 10 Ma age of the uppermost tuff of the underlying Quebrada Macusa Formation, though lacustrine sedimentation may have begun later, and the 0.5 and 0.3 Ma ages obtained by Wörner et al. (1988) for volcanic deposits of the Guallatire and Nevados de Payachatas volcanoes that cover the Lauca Formation deposits.

According to our observations, the accommodation space for at least the oldest and westernmost deposits of the Lauca Formation was created by the growth and eastward shift of deformation associated with development of the Chucal anticline. However, we have no evidence to disprove the existence of younger phases of sedimentation associated with extensional basin development, as proposed by Muñoz and Charrier (1996), Kött et al. (1995), and Gaupp et al. (1999).

4. Early Miocene–Middle Pliocene tectonic evolution

4.1. Tectonic activity of the Chucal thrust system and tectonic evolution on the east side of the Chapiquiña-Belén ridge

According to some authors, the Lupica Formation developed in extensional conditions (García, 1996; García et al., 1996). The contractional activity that followed the deposition and partial erosion of the Lupica Formation and controlled the paleogeographic evolution of the western Altiplano in the Chucal region probably was associated with the inversion of the previously existing normal faults.

The first tectonic activity recorded in the sedimentary record of the Chucal region is associated with paleogeographic changes related to the development of the first phase of accommodation space creation (basin) on the west flank of the Chucal anticline. In this case, maximum tectonic activity probably was concentrated to the west of the present-day anticlinal axis, possibly on the Jaropilla thrust fault. Along the anticline axis, some previous uplift may have caused the SW transport of the sediments, which formed the white, tuffaceous, fluvial series of the upper Lupica Formation. The absence or almost total absence of deposits equivalent to Member W1 on the east side of the anticline axis, that is, from the beginning of deposition of the Chucal Formation, makes it impossible to determine if the basin was there already and, if so, whether the sediments were removed by subsequent uplift and erosion. According to the available age determination of the massive, white tuff of the upper Lupica Formation, the first tectonic episode probably occurred shortly after 21.7 Ma.

Although the tectonic activity and consequent deformation were probably continuous, the unconformities provide evidence of moments or episodes during which deformation and erosion rates were probably higher and/or sedimentation rates were not rapid enough to fill the forming accommodation space.

Unconformities U1 and U2 provide evidence of two of such episodes during the syntectonic paleogeographic evolution. Unconformity U3, which represents a third such episode and separates the Chucal Formation from the Quebrada Macusa Formation, occurred before 17.5 ± 0.4 Ma, the age of the basal horizon of the Quebrada Macusa Formation. This last age and the 21.7 ± 0.8 Ma age obtained for the white–gray, massive tuff of the upper Lupica Formation bracket the Chucal Formation and its deformational history and indicate that deformation occurred in a short period of 4–5 m.y. Preliminary indications (Flynn et al., 2002; new fossils discovered in 2001) that the mammalian fauna does not span more than one South American land mammal 'age' (typically less than 2–3 m.y. during the Miocene, Flynn and Swischer, 1995) suggest the Chucal Formation may have accumulated, and its synsedimentary events occurred, during an even shorter time.

The activity of the Jaropilla thrust fault occurred during the accumulation of the Chucal Formation and ended before the deposition of the Quebrada Macusa Formation and therefore can be related closely to the development of the basinal accommodation space for syntectonic deposition of the Chucal Formation sediments.

Additional evidence of an episode of rapid deformation and erosion or relatively slow sedimentation includes the deformation of the Quebrada Macusa Formation. This episode occurred approximately 7.5 m.y. after U3, according to the 9.6 ± 0.7, 10.4 ± 0.7, and 11.2 ± 0.5 Ma dates obtained for the upper levels of the unit and the 17.5 Ma and older dates from horizons above U3.

Another episode of this kind may have occurred, as is indicated by the deformation of the westernmost exposures of the Lauca Formation that unconformably cover the upper Quebrada Macusa deposits. This last episode may be related to the east-vergent thrust system that affects the lower levels of the Lauca Formation (Muñoz and Charrier, 1996) and is well exposed 20 km north of the study region along the road from Putre to Surire, immediately south of the Guallatire village. This tectonic episode probably occurred during the Pliocene and certainly before the deposition of the 2.3 ± 0.7 Ma old Lauca ignimbrite intercalation, though a Late Miocene age for it cannot be rejected.

4.2. Correlation with the tectonic evolution on the precordillera (west of the Chapiquiña-Belén ridge)

Although deformation might have been rather continuous on the east side of the Chapiquiña-Belén ridge, the presence of unconformities and the existence of environmental changes suggest periods of more rapid activity, on which we base the chronology of the tectonic evolution of the region. The well-constrained data that we present for the chronology of the activity of the Chucal thrust system can be compared briefly with the available information from the precordillera on the west side of the Chapiquiña-Belén ridge.

The deformation associated with the Jaropilla thrust fault and the Chucal anticline, which occurred rather continuously between 21.7 ± 0.8 and 17.5 ± 0.4 Ma, during the development of the sedimentation accommodation space and the deposition of the syntectonic Chucal Formation, is approximately coeval with the activity of the Cerro Lagunas-Belén-Tignámar thrust fault. This fault produced the syntectonic, coarse-grained Joracane Formation on the west.

The probable continuous tectonic deformation that occurred during the deposition of the Quebrada Macusa Formation, between 17.5 ± 0.4 Ma (according to its basal deposits) and approximately 10 Ma (age of the upper tuff deposit), can be correlated with the beginning of deformation on the Oxaya anticline farther west. With this deformation, the dorsal flank of the Oxaya anticline, formed by deposits of the ignimbritic Oxaya and andesitic Zapahuira Formations, was tilted eastward before the deposition of the Huaylas Formation at ~11 Ma.

The folding of the Quebrada Macusa Formation, after approximately 11 Ma and before the unconformable deposition of the Lauca Formation, can be correlated with the deformation associated with the Copaquilla-Tignámar thrust fault on the precordillera. This thrust fault activity yielded the syntectonic, coarse-grained, >10?–9 Ma old deposits of the Huaylas Formation (Salas et al., 1966; Naranjo and Paskoff, 1985). However, new fossils and reanalysis of the sole specimen on which the 9–10 Ma age estimates for the Huaylas Formation was originally based (Flynn et al., 2005) indicate that deposition of this unit could have begun earlier, possibly contemporaneous with or slightly postdating the deposition of the lenticular Zapahuira Formation, which yields ages of 12.7 ± 0.1 and 12.8 ± 0.24 Ma.

Finally, the deformation on the eastern border of the Chucal anticline, which caused the deformation of the lower deposits of the Lauca Formation in the Late Miocene and/or early Pliocene, can be correlated with the two following events observed along the precordillera: (1) a further upwarping of the Pampa Oxaya anticline (Naranjo, 1997; Parraguez, 1997; García et al., 1999), probably caused by reactivation of the Ausipar thrust fault (Parraguez, 1997; García et al., 1999), and (2) the Pliocene reactivation of the Copaquilla-Tignamar thrust fault, which deformed the Huaylas ignimbrite and preexistent units at the end of the Pliocene.

5. Implications

5.1. Volcanism

The presence of abundant tuff and welded tuff deposits in the syntectonic depositional series indicates that explosive volcanism existed almost constantly during the early Miocene–Late Pliocene. Explosive volcanism appears to have been especially active during the deposition of the Quebrada Macusa Formation and the described tectonic activity.

As indicated by the presence of a 10.4 ± 0.7 Ma old andesitic lava from the Anocarire volcano, intercalated at the top of the Quebrada Macusa Formation, and the 7.0–6.6 Ma old andesitic lavas of the Lauca volcano, intercalated in the Lauca Formation, andesitic volcanism also was active at this time but was restricted to localized centers. Therefore, it is less evident in the stratified series.

5.2. Biota

5.2.1. Plant remains

Poorly preserved plant remains were found in several levels throughout these deposits. Pollen and some *Erytroxylum* sp. leaves collected in 1992 from Member W2 of the Chucal Formation indicate the coexistence of forests and grassland steppes developed in dry climatic conditions and temperate to warm paleotemperatures (Charrier et al., 1994b). A rather abundant collection of well-preserved leaves, apparently of poor to moderate diversity, were recently found in Member W2; these have not been studied yet.

Gregory-Wodzicki et al. (1998) and Gregory-Wodzicki (2002) have studied abundant, diverse, and well-preserved paleoflora from two stratigraphically close levels from Jakokkota, in the Bolivian Altiplano, approximately 100 km north of Cerro Chucal. This locality (Cerro Jancocata) also was studied by Berry (1922). A sample from an ashfall tuff located between the two fossiliferous horizons gives a weighted mean age of 10.66 ± 0.06 Ma ($^{40}Ar/^{39}Ar$ method with laser fusion of single and multiple crystals) in the early Late Miocene (Gregory-Wodzicki et al., 1998; Gregory-Wodzicki, 2002). Therefore, these deposits are coeval with the uppermost levels of the Quebrada Macusa Formation and more than 7 m.y. younger than the plant-bearing deposits of the Chucal Formation (Member W2).

The Jakokkota flora indicates that the climatic conditions of the early Late Miocene of at least part of the Altiplano were milder than the present conditions in this region and that precipitation was similar to that of the present day. Similar conditions were derived by Gregory-Wodzicki et al. (1998) for an older, but less well-constrained in age, fossil flora from Potosí in the Eastern Cordillera, bracketed between 20.7 and 13.8 Ma. The estimated paleoelevation for the Jakokkota region during the early Late Miocene coincides with the altitude suggested by Charrier et al. (1994b) for the Chucal region during the Early Miocene (21.7–17.5 Ma), which was estimated at 1000 m.

According to these results, the general climatic conditions and paleoelevation of these regions that lie relatively close to each other did not change dramatically between the Early Miocene and the early Late Miocene, over a time span of approximately 7 m.y.

5.2.2. Mammalian fossils

Following the first discovery of an isolated Toxodont humerus in 1992 (Charrier et al., 1994c), abundant mammalian remains have been found throughout much of the Chucal Formation (Flynn et al., 2002). In the anticline's west flank, they generally occur in floodplain facies and lacustrine-influenced intervals (i.e. in the lower parts of Members W3 and W4), whereas on the east flank, they predominate in fluvial facies (i.e. Member E3, but also E1). The Chucal fauna is the first mammalian assemblage known from the Chilean Altiplano and appears to represent a single, short temporal interval, most likely Santacrucian or Friasian (Flynn et al., 2002; 14–17.5 Ma, Flynn and Swischer, 1995). The new date from the basal levels of the Quebrada Macusa Formation overlying the mammal fossils constrains this fauna to be more likely Santacrucian than a youger SALMA, consistent with preliminary biochronologic data from new fossil mammal specimens recovered in 2001. Additional work on the fauna, paleomagnetic stratigraphy, and geochronology should determine this constraint with more precision.

Fossil mammals already identified from the Chucal Formation (Flynn et al., 2002) represent at least seven taxa, including four notoungulates (*Nesodon imbricatus*, toxodont; three mesotheriid species; hegetotheriine), the litoptern *Theosodon*, glyptodontid xenarthrans, and the oldest known chinchilline rodent. Bond and García (2002) also identify three teeth of a Toxodont from the upper Chucal Formation on the northwest slope of Cerro Chucal as *Palyeidon*, but they note that the material differs from, and is somewhat intermediate between, *Nesodon* and *Palyeidon* in various features. Abundant new material and additional localities discovered in 2001 from both the west and east flanks south of Cerro Chucal suggest that at least six to seven additional fossil mammal species occur in the formation.

The presence of an abundant flora and suitable paleoenvironments clearly fostered the occurrence and persistence of an abundant mammalian fauna in this region during the deposition of the Chucal Formation and its associated tectonic events. Although palynological evidence and plant macrofloral remains suggest the existence of rather dry climatic conditions (similar to but probably not as severe as modern conditions in the area), the development of large topographic lows filled with water (lakes) must have favored the development of grasslands (steppes) and some forests (Charrier et al., 1994b) and consequently fostered an abundant and diverse fauna. Many of the mammalian taxa recovered to date exhibit adaptations (particularly hypsodonty) to open, grassland environments, though others have dental or skeletal features (or living relatives) that suggest possible preferences for more forested or even riparian/wetland environments.

5.3. Paleoclimate

The development of the Chapiquiña-Belén ridge, between the Chucal region and the modern precordillera, probably generated considerable local environmental or climatic effects on both sides. A comparison of the paleoenvironmental conditions on both sides of the Chapiquiña-Belén ridge suggests that the west side was considerably drier than the east. In the precordillera, other than the well-developed lacustrine facies of the Lupica Formation, the only evidence for lacustrine deposits are the fine-grained basal deposits of the 10–9 Ma old or possibly older Huaylas Formation (Flynn et al., 2005).

On the east side of the ridge (the Chucal region), lacustrine facies existed within the basin almost continuously from the basal Chucal Formation to the Lauca Formation. Therefore, the conditions may have been more humid. This evidence also might indicate that the climatic regime was influenced by a humid westward atmospheric flux and that the mountain range was already high enough to isolate the west flank of the ridge from the humid air masses coming from the east, which concentrated precipitation on the eastern flank and led to more arid conditions on the leeward side of the ridge, at least during some periods during the year, as is the case today.

Deposition of the Chucal Formation, from <21.7 to 17.5 Ma, coincided with a global warm episode between 26 Ma and ~ 15 Ma (Zachos et al., 2001). This warm period, which nonetheless had some brief cold episodes, culminated with the Mid-Miocene climatic optimum of 17–15 Ma (Zachos et al., 2001). These mild climatic conditions certainly permitted the persistence of a diverse fauna and flora in this region but also may have regionally modified the tectonic influences and uplift in terms of their magnitude and effect and probably masked the real altitude/elevation reached by the mountain range. On the basis of the existence of similar climatic conditions in the Chucal Formation and the 7 m.y. younger early Late Miocene deposits at Cerro Jakokkota, we might conclude that no climatic changes occurred during the time period. However, the substantially younger Jakokkota flora, developed well into the gradual cooling period that followed the Mid-Miocene climatic optimum, reflects no climatic variation. Such apparent conflicts may be explained in various ways. Significant regional differences (environmental stability) may have occurred in response to global climatic changes/trends (cooling and drying) or through complex interactions between climate and tectonically influenced processes (e.g., uplift, paleodrainage, rainfall patterns, rainshadows). The various sequences may represent sufficiently distinct local habitats that their inferred climate and paleoenvironments cannot be directly compared. The floral and faunal data may reflect the paleoenvironment differently; for example, pollen and macroflora may sample the region differentially, and mammalian communities may sample a broader spectrum of local habitats but be less sensitive to particular climatic variables. In addition, our paleobotanical analytical methods, though powerful, are influenced by many variables, and they may not adequately distinguish the relative influences of complex series of changes in temperature (and subcomponents such as extremes, ranges, and seasonality), rainfall or humidity, soil moisture or drainage, elevation, and so on. Additional floral and faunal samples, well constrained in age and from different parts of the region, and an application of additional methods for paleoenvironmental reconstruction should refine our understanding of biotic responses to changing climate and tectonic regimes.

Evidence provided by Kött et al. (1995) for the younger Lauca Formation indicates that the climate between 7 Ma and the Late Pleistocene was arid to semiarid. According to them, the absence of fossil plants and animals documents the existence of an environment hostile to life. Gaupp et al. (1999), on the basis of the pollen assemblage in the lower part of the Lauca Formation, conclude that the climate was generally cold and semi-humid to semiarid.

5.4. Andean uplift

Available paleofloristic information indicates that the Andean range in the Early–early Late Miocene had a moderately low elevation. Although contractional processes were active throughout the deposition of the Chucal and Quebrada Macusa Formations, they do not seem to have been especially effective in uplifting the mountain range to great elevations. If we consider the similar climatic and altitudinal conditions inferred from the study of the palynologic content of the Early Miocene Member W2 of the Chucal Formation and the foliar physiognomy of the early Late Miocene Jakokkota flora in the Bolivian Altiplano, we can conclude that (1) the tectonic events detected during the deposition of the Chucal Formation did not increase substantially the altitude of this region and (2) the uplift of the altiplano to its current elevation (3800–4000 m) occurred after deposition of both the Chucal Formation and the Jakokkota flora, that is, after 10 Ma (and probably before 2.7 Ma, the age of the Lauca ignimbrite, above which no deformation has been observed in the Lauca Formation), in the Late Miocene or Pliocene (also see Gregory-Wodzicki, 2000, 2002). This conclusion would mean that all uplift occurred between 10 and 2.7 Ma (Lauca ignimbrite). However, though we cannot assess the exact contribution of this deformation to the uplift of the west side of the Altiplano, we believe that the uplift process was active during Oligocene times and continued until the middle Pliocene and possibly to the Present.

5.5. Continental margin–Nazca plate interactions

The continuous or almost continuous late Early Miocene–early Pliocene contractional episode described here coincides with a broad episode of increasing and high convergence between the Nazca and South American plates (Pardo-Casas and Molnar, 1987). According to these authors, this period occurred after approximately 26 Ma, the moment of the lowest convergence rate since the Oligocene. This situation is similar to that observed in the Cordillera Principal in central Chile for essentially volcanic deposits of the Abanico (=Coya-Machalí) and Farellones Formations (Charrier et al., 2002). Contraction began at approximately 25–21 Ma after an extensional episode and lasted until at least 16 Ma. In that contraction occurred in these two separate regions during a longer episode of increasing convergence, and apparently since the beginning of the episode of increasing convergence, rather than during the peak of convergence seems to indicate that deformation (i.e. folding, faulting, and accommodation of deformation along preexistent structures or inversion) is the immediate response to an increased compressive tectonic regime.

6. Conclusions

The syntectonic sedimentation of the 600 m thick west flank of the Chucal Formation, the development of the Chucal anticline, the progressive unconformities U1, U2, and U3, and the development of separate paleogeographies with abundant associated fauna and flora on both sides of the anticline occurred probably continuously between 21.7 ± 0.8 and 17.5 ± 0.4 Ma in a rather short time span of 4 m.y.

Additional contractional deformation was considerably lesser. Deformation that affected the Quebrada Macusa Formation and caused the development of U4, as well as further deformation of the underlying units, occurred after 11.2 ± 0.5 Ma and before 10.4 ± 0.7 Ma, whereas deformation of the lower part of the Lauca Formation occurred in an apparently longer time span between 10.4 ± 0.7 and 2.3 ± 0.7 Ma ($\sim$7 m.y.).

The total time elapsed between the syntectonically influenced sedimentary series of the massive, white–gray tuff of the upper Lupica Formation and the deformed lower part of the Lauca Formation is 18 m.y. Tectonic activity during those 18 m.y. was apparently continuous, though deformation rates apparently became slower with time.

The resulting east-vergent contractional deformation controlled the development of the paleogeographic features (i.e. Chapiquiña-Belén ridge), the different paleoenvironment, and the paleoclimatic conditions on either side of the ridge. More humid conditions on the east side fostered development of abundant fauna and flora. A similar situation seems to be the case for the precordillera on the west flank of the Chapiquiña-Belén ridge.

The development of the Chapiquiña-Belén ridge contributed to development of the Altiplano as an endorheic basin between the Eastern Cordillera to the east and the Chapiquiña-Belén ridge to the west.

Acknowledgements

This study is part of a collaborative, interdisciplinary effort designed to understand the evolution of Andean (altiplano) uplift in northern Chile and western Bolivia and the associated tectonic, geomorphologic, sedimentary, paleogeographic, faunistic, floristic, and climatic effects. We thank the National Geographic Society (USA) for Grant 5371-94, US NSF Grant DEB-9317943, Chile FONDECYT Grant 1224-91, the Field Museum Department of Geology, the John S. Guggenheim Foundation, and the IRD for fieldwork and other support, as well as for supporting A. Chávez's thesis. We thank G. Carrasco, A. Charrier, C. Hérail, the Flynn family, and S. Villagrán for collaboration in the field and the Field Museum preparator staff for their excellent preparation of the mammal fossils. Paleontological investigations have benefited greatly from continuing support from the Museo Nacional de Historia Natural (Santiago) and the Consejo Nacional de Monumentos Nacionales de Chile. We thank Adrian Hartley, Onno Oncken, and Víctor Ramos for reviews that significantly improved the manuscript. This is a contribution of the Centro de Estudios Atiplánicos (INCAS) of the Universidad de Chile.

References

Aguirre, E., 1990. Geología del Complejo Volcánico Choquelimpie-Ajoya, Altiplano de Arica, I Región. Thesis, Departamento de Geología, Universidad de Chile, 136pp.

Allmendinger, R.W., 1986. Tectonic development of the southeastern border of the Puna plateau, northwestern Argentina Andes. Geological Society of America Bulletin 97, 1072–1082.

Allmendinger, R.W., Jordan, T.E., Kay, S.M., Isacks, B., 1997. The evolution of the Altiplano-Puna plateau of the central Andes. Annual Review Earth and Planetary Sciences 25, 139–174.

Bargo, M.S., Reguero, M.A., 1989. El primer registro de un mamífero fósil en el extremo septentrional de Chile. Ameghiniana 26, 239.

Basei, M.A., Charrier, R., Hervé, F., 1996. New ages (U–Pb, Rb–Sr, K–Ar) from supposed Precambrian units in Chile: Some geotectonic implications, Proceedings Third International Symposium on Andean Géodynamics (Saint Malo), Ed. ORSTOM, Paris, France pp. 763-766.

Berry, E.W., 1922. Late Tertiary plants from Jancocata, Bolivia. John Hopkins Studies in Geology 4, 205–221.

Bisso, C.R., 1991. Geología y geoquímica del yacimiento de Oro y plata Choquelimpie, 1ª Región, Chile. Thesis, Departamento de Geología, Universidad de Chile, 142pp.

Bond, M., García, M., 2002. Nuevos restos de toxodonte (Mamalia, Notoungulata) en estratos de la Formación Chucal, Mioceno, Altiplano de Arica, norte de Chile. Revista Geológica de Chile 29 (1), 81–91.

Charrier, R., 1973. Interruptions of spreading and the compressive tectonic phases of the Meridional Andes. Earth and Planetary Science Letters 20, 242–249.

Charrier, R., Wyss, A.R., Flynn, J.J., Swisher III., C.C., Spichiger, S., Zapatta, F., 1994a. Nuevos antecedentes estratigráficos y estructurales para las Formaciones Coya-Machalí y Abanico, entre 33° 50′ y 35°S, Cordillera Principal Chilena, Proceedings 7th Congreso Geológico Chileno, Concepción 1994 pp. 1316–1319.

Charrier, R., Muñoz, N., Palma-Held, S., 1994b. Edad y contenido paleoflorístico de la Formación Chucal y condiciones paleoclimáticas para el Oligoceno tardío—Mioceno Inferior en el Altiplano de Arica, Chile, Proceedings 7th Congreso Geológico Chileno, Concepción, vol. 1 1994 pp. 434–437.

Charrier, R., Muñoz, N., Wyss, A.R., Flynn, J.J., Hérail, G., 1994c. Hallazgo de un húmero de Taxodonte (Mammalia) en la Formación Chucal (Oligoceno Tardío-Mioceno Inferior) en el Altiplano de Arica, Chile, Proceedings 7th Congreso Geológico Chileno, Concepción, vol. 1 1994 pp. 430–433.

Charrier, R., Hérail, G., Flynn, J.J., Riquelme, R., García, M., Croft, D., Wyss, A.R., 1999. Opposite thrust-vergencies in the Precordillera and Western Cordillera in northern Chile and structurally linked Cenozoic paleoenvironmental evolution, Proceedings 4th International Symposium on Andean Geodynamics (ISAG), Göttingen, Germany, Editions IRD, pp. 155–158.

Charrier, R., Hérail, G., Flynn, J.J., Riquelme, R., García, M., Croft, D., Wyss, A.R., 2000. El Cordón Chapiquiña-Belén en el borde occidental del Altiplano chileno: significado paleogeográfico y contexto tectónico regional, Proceedings 9th Congreso Geológico Chileno, Puerto Varas, vol. 1, pp. 763–767.

Charrier, R., Baeza, O., Elgueta, S., Flynn, J.J., Gans, P., Kay, S.M., Muñoz, N., Wyss, A.R., Zurita, E., 2002. Evidence for Cenozoic extensional basin development and tectonic inversion south of the flat-slab segment, southern Centralk Andes, Chile (33°.36°S.L.). Journal of South American Earth Sciences 15, 117–139.

Chávez, A.N., 2001. Evolución paleoambiental y sedimentación sintectónica durante el Mioceno en el sector del Cerro Chucal (Cordillera occidental, Región de tarapacá: 69°10′O–18°45′S), y su relación con la estructuración del Altiplano. Thesis, Departamento de Geología, Universidad de Chile, 73pp.

Dewey, J.F., Bird, J.M., 1970. Mountain belts and the new global tectonics. Journal of Geophysical Research 75 (14), 2625–2647.

Evernden, J., Kris, S., Cherroni, C., 1977. Potassium-Argon ages of some Bolivian rocks. Economic Geology 72, 1042–1061.

Farías, M., Charrier, R., Comte, D., Martinod, J., Pinto, L., Hérail, G., 2002. Active Late Cenozoic flexures in the Precordillera in Northern Chile: Correlations with the shallow seismic activity, and implications for the uplift of the Altiplano. Eos Trans. AGU 83 (47). Fall Meet. Suppl., Abstract T51A-1136.

Flynn, J., Swisher III., C.C., 1995. Chronology of the Cenozoic South American land mammal ages. In: Berggren, W.A., Kent, D.V., Hardenbol, J. (Eds.), Geochronology, Time-scale, and Global Stratigraphic Correlation, vol. 54. SEPM (Society for Sedimentary Geology) Spec. Publ., pp. 317–333.

Flynn, J.J., Croft, D., Charrier, R., Hérail, G., Wyss, A., 2002. The first Cenozoic mammal fauna from the Chilean Altiplano. Journal of Vertebrate Paleontology 22 (1), 200–206.

Flynn, J.J., Croft, D., Charrier, R., Wyss, A., Hérail, G., García, M., 2005. New Mesotheriidae (Mammalia, Notoungulata, Typotheria), Geochronology and tectonics of the Caragua Area, Northernmost Chile. Journal of South American Earth Sciences 17.

Frutos, J., 1981. Andean tectonic as a consequence of sea-floor spreading. Tectonophysics 72, T21–T32.

García, M., 1996. Geología y estructura del borde del Altiplano occidental, en el área de Belén (Chile). Thesis, Departamento de Geología, Universidad de Chile, Santiago, 111pp.

García, M., 2001. Evolution oligo-néogène de l'Altiplano Occidental (arc et avant-arc des Andes d'Arica, 18°–19°s). Téctonique, volcanisme, sédimentation, géomorphologie et bilan erosion-sédimentation. Thesis, Université Joseph Fourier (Grenoble 1), France, 178pp.

García, M., Hérail, G., Charrier, R., 1996. The Cenozoic forearc evolution in Northern Chile: the western border of the Altiplano of Belén (Chile), Proceedings Third International Symposium on Andean Geodynamics, (ISAG), Saint-Malo, Ed. ORSTOM, France, pp. 359–362.

García, M., Hérail, G., Charrier, R., 1999. Age and structure of the Oxaya Anticline: a major feature of the Miocene compressive structure of northern Chile, Proceedings 4th International Symposium on Andean Geodynamics, (ISAG), Göttingen, Ed. ORSTOM, France, pp. 249–252.

García, M., Hérail, G., Charrier, R., Mascle, G., Fornari, M., Pérez, de Arce, 2002. Oligocene-Neogene tectonic evolution of the altiplano of northern Chile (18°–19°S, Proceedings 5th International Symposium on Andean Geodynamics (IASG), Toulouse, IRD, France, pp. 235–238.

Gaupp, R., Kött, A., Wörner, G., 1999. Paleoclimatic implications of Mio-Pliocene sedimentation in the high-altitude intra-arc Lauca basin of northern Chile. Paleography, Paleoclimatology, Paleoecology 151, 79–100.

Godoy, E., Lara, L., 1994. Segmentación estructural andina a los 33°–34°: Nuevos datos en la Cordillera Principal, Proceedings 7th Congreso Geológico Chileno, Concepción, vol. 2. pp. 1344–1348.

Godoy, E., Yáñez, G., Vera, E., 1999. Inversion of an Oligocene volcano-tectonic basin and uplifting of its superimposed Miocene magmatic arc in the Central Chilean Andes: first seismic and gravity evidences. Tectonophysics 306 (2), 217–236.

Gregory-Wodzicki, K., 2000. Uplift history of central and Northern Andes: a review. Geological Society of America Bulletin 112 (7), 1091–1105.

Gregory-Wodzicki, K., 2002. A late Miocene subtropical-dry flora from the northern Altiplano, Bolivia. Palaeogeography, Palaeoclimatology, Palaeoecology 180, 331–348.

Gregory-Wodzicki, K., Macintosh, W.C., Velazquez, K., 1998. Climatic and tectonic implications of the late Miocene Jakokhota flora, Bolivian Altiplano. Journal of South American Earth Sciences 11 (6), 533–560.

Gröpper, H., Calvo, M., Crespo, H., Bisso, C.-R., Cuadra, W.A., Dunkerley, P.M., Aguirre, E., 1991. The epithermal gold-silver deposit of Choquelimpie, Northern Chile. Economic Geology 86, 1206–1221.

Heber, V., 1997. Das präandine Basement von Belén (Nordchile): Kartierung und Geochimie der metamorphen Gesteine. Thesis, Institut für Geochemie, Georg-August-Universität, Göttingen, Germany, 115pp.

Hérail, G., Riquelme, R., 1997. Estructuras de crecimiento sinsedimentario y de reactivación tectónica cenozoicas en la Región de Chucal (Altiplano de Chile), Implicaciones tectónicas. Proceedings 8th Congreso Geológico Chileno, Antofagasta, vol. 1. pp. 90–94.

Isacks, B.L., 1988. Uplift of the central Andean plateau and bending of the Bolivian orocline. Journal of Geophysical Research 93 (B4), 3211–3231.

Jordan, T.E., Isacks, B.L., Allmendinger, R.W., Brewer, J.A., Ramos, V.A., Ando, C.J., 1983. Andean tectonics related to geometry of subducted Nazca plate. Geological Society of America Bulletin 94, 341–361.

Jordan, T.E., Reynolds III., J.H., Erikson, J.P., 1997. Variability in age of initial shortening and uplift in the Central Andes, 16-33°30′ S. In: W. Ruddiman (Ed.), Tectonic Uplift and Climate Change. Plenum Press, New York, pp. 41–61.

Jordan, T.E., Burns, W.M., Veiga, R., Pángaro, F., Copeland, P., Kelley, S., Mpodozis, C., 2001. Extension and basin formation in the Southern Andes caused by increased convergence rate: a mid-Cenozoic trigger for the Andes. Tectonics 20, 308–324.

Kay, S.M., Abbruzzi, J.M., 1996. Magmatic evidence for Neogene lithospheric evolution of the central Andean "flat-slab" between 30° and 32°S. Tectonophysics 259, 15–28.

Kono, M., Fukao, Y., Yamamoto, A., 1989. Mountain building in the Central Andes. Journal of Geophysical Research 94 (B4), 3891–3905.

Kött, A., Gaupp, R., Wörner, G., 1995. Miocene to Recent history of the Western Altiplano in northern Chile revealed by lacustrine sediments of the Lauca Basin (18°15′/18°40′S–69°30′/69°05′W). Geologische Rundschau 84, 770–780.

Lavenu, A., Bonhomme, M., Vatin-Perignon, N., Pachtère, P., 1989. Neogene magmatism in the Bolivian Andes between 16° and 18°S: stratigraphy and K–Ar geochronology. Journal of South American Earth Sciences 2 (1), 35–47.

Lezaun, J., 1997. Geochronology of Andean basement and Tertiary volcanics. Thesis, Institut für Geochemie, Georg-August Universität, Göttingen, 92pp.

Lezaun, J., Heber, V., Beck, A., Hammerschmiedt, K., Wörner, G., 1996. Metamorphic basement units and the stratigraphy of overlying volcano-sedimentary rocks at 18°S: implications for style and timing of Andean deformations, Proceedings Third International Symposium on Andean Geodynamics (ISAG), Saint Malo, Ed. ORSTOM, Paris, France, pp. 423–426.

Lucassen, F., Wilke, H., Viramonte, J.G., Becchio, R., Franz, G., Laber, A., Wemmer, K., Vroon, P., 1996. The Paleozoic basement of the central Andes (18°–26°S): a metamorphic view, Proceeding of Third International Symposium on Andean Geodynamics (Saint Malo), ed. ORSTOM, Paris, pp. 779–782.

Marshall, L., Swisher, C., Lavenu, A., Hoffstetter, R., Curtiss, G., 1992. Geochronology of the mammal-bearing late Cenozoic on the northern Altiplano, Bolivia. Journal of South American Earth Sciences 2 (1), 1–19.

Montecinos, F., 1963. Observaciones de geología en el Cuadrángulo de Campanani, Departamento de Arica, Provincia de Tarapacá. Thesis, Departamento de Geología, Universidad de Chile, Santiago, 109pp.

Mpodozis, C., Ramos, V., 1989. The Andes of Chile and Argentina. In: Ericksen, G.E., Cañas, M.T., Reinemund, J.A. (Eds.), Geology of the Andes and its Relation to Hydrocarbon and Mineral Resources Earth Science Series, vol. 11. Circum-Pacific Council for Energy and Mineral Resources, Houston, TX, pp. 59–90.

Mpodozis, C., Cornejo, P., Kay, S.M., Titler, A., 1995. La Franja de Maricunga: Síntesis de la evolución del frente volcánico oligoceno-mioceno de la zona sur de los Andes centrales. Revista Geológica de Chile 22, 273–314.

Muñoz, N., 1988. Estratigrafía y paleontología de la Formación Blanco (Aptiano) y Formación Lauca (Mioceno), 1ª región. Unpublished report ENAP, Santiago, 9pp.

Muñoz, N., 1991. Marco geológico y estratigráfico de un sistema fluvio-lacustre Paléogeno, Altiplano de Arica, Norte de Chile, Proceedings 6th Congreso Geológico Chileno, Viña del Mar 1991, pp. 201–204.

Muñoz, N., Sepúlveda, P., 1992. Estructura compresiva con vergencia al oeste en el borde oriental de la Depresión Central, Norte de Chile (19°15′S). Revista Geológica de Chile 19 (2), 241–247.

Muñoz, N., Charrier, R., 1996. Uplift of the western border of the Altiplano on a west-vergent thrust system, northern Chile. Journal of South American Earth Sciences 9 (3–4), 171–181.

Naranjo, J.A., Paskoff, R., 1985. Evolución cenozoica del piedemonte andino en la Pampa del Tamarugal, norte de Chile (18°–21°S), Proceedings 4th Congreso Geológico Chileno, vol. 4 1985 pp. 5/149–5/165.

Pacci, D., Hervé, F., Munizaga, F., Kawashita, K., Cordani, U., 1980. Acerca de la edad Rb–Sr precámbrica de rocas de la Formación Esquistos de Belén, Departamento de Parinacota, Chile. Revista Geológica de Chile 11, 43–50.

Pardo-Casas, F., Molnar, P., 1987. Relative motion of the Nazca (Farallón) and South American plates since Late Cretaceous time. Tectonics 6 (3), 233–248.

Parraguez, G., 1997. Sedimentología y geomorfología productos de la tectónica cenozoica en la Depresión Central, Pampa de Chaca, I Región, Tarapacá, Chile. Thesis, Departamento de Geología, Universidad de Chile, Santiago, 104pp.

Pinto, L., 1999. Evolución tectónica y geomorfológica de la deformación cenozoica del borde occidental del Altiplano y su registro sedimentario entre los 19°08′–19°27′S (Región de Tarapacá, Chile). Thesis, Departamento de Geología, Universidad de Chile, 125pp.

Pinto, L., Hérail, G., Charrier, R., 2004. Sedimentación sintectónica asociada al flexuramiento neógeno en el borde oriental de la Depresión Central en el área de Moquella (19°15′S, Norte de Chile). Submitted to Revista Geológica de Chile 31 (1), 19–44.

Ramos, V.A., 1988. The tectonics of the Central Andes, 30° to 33°S latitude. Geological Society of America, Special Paper 218, 31–54.

Riquelme, R., 1998. Evolución tectosedimentaria post-oligocénica del borde occidental del Altiplano, entre Tignámar y Salar de Surire, I Región, Chile. Thesis, Departamento de Geología, Universidad de Chile, Santiago, 123pp.

Riquelme, R., Hérail, G., 1997. Puesta en evidencia de discordancias progresivas en el Cenozoico Superior del Altiplano de Arica: Implicancias en la interpretación tectónica de la Cordillera Occidental, Proceedings 8th Congreso Geológico Chileno, vol. 1 pp. 231–235.

Salas, R., Kast, R., Montecinos, F., Salas, I., 1966. Geología y recursos minerales del Departamento de Arica, Provincia de Tarapacá, vol. 21. Instituto de Investigaciones Geológicas, Santiago, 130pp.

Salinas, P., Villaroel, C., Marshall, L., Sepúlveda, P., Muñoz, N., 1991. Typotheriopsis sp. (Notongulata, Mesotheridae), Mamífero del Mioceno Superior en las cercanías de Belén, Arica, Norte de Chile, Proceedings 6th Congreso Geológico Chileno, Viña del Mar, vol. 1. pp. 314–317.

Tobar, A., Salas, R., Kast, R., 1968. Cuadrángulos Camaraca y Azapa, Provincia de Tarapacá. Instituto de Investigaciones Geológicas, Santiago. Carta Geológica de Chile, 19–20, 13pp.

Viteri, E., 1979. Geología y recursos minerales de la Formación Huaylas en el Altiplano de Arica, Proceedings of the 2nd Congreso Geológico Chileno, Arica, vol. 2 1979 pp. E33–E52.

Wörner, G., Harmon, R.S., Davidson, J., Moorbath, S., Turner, T.L., McMillan, N., Nye, C., López-Escobar, L., Moreno, H., 1988. The Nevados de Payachata volcanic region 18°S/69°W, N. Chile. I. Geological, geochemical and isotopic observations. Bulletin of Volcanology 30, 287–303.

Wörner, G., Hammerschmidt, K., Henjes-Kunst, F., Lezaun, J., Wilke, H., 2000. Geochronology (40 Ar/39Ar, K–Ar and He-exposure ages) of Cenozoic magmatic rocks from Northern Chile (18°–22°S): implications for magmatism and tectonic evolution of the central Andes. Revista Geológica de Chile 27 (2), 205–240.

Zachos, J., Pagani, M., Sloan, L., Thomas, E., Billups, K., 2001. Trends, rhythms, and aberrations in global climate 65 Ma to Present. Science 292, 686–692.