Late Mesozoic to Paleogene stratigraphy of the Salar de Atacama Basin, Antofagasta, Northern Chile: Implications for the tectonic evolution of the Central Andes

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Abstract

The Salar de Atacama basin, the largest "pre-Andean" basin in Northern Chile, was formed in the early Late Cretaceous as a consequence of the tectonic closure and inversion of the Jurassic–Early Cretaceous Tarapacá back arc basin. Inversion led to uplift of the Cordillera de Domeyko (CD), a thick-skinned basement range bounded by a system of reverse faults and blind thrusts with alternating vergence along strike. The almost 6000-m-thick, upper Cretaceous to lower Paleocene sequences (Purilactis Group) infilling the Salar de Atacama basin reflects rapid local subsidence to the east of the CD. Its oldest outcropping unit (Tonel Formation) comprises more than 1000 m of continental red sandstones and evaporites, which began to accumulate as syntectonic growth strata during the initial stages of CD uplift. Tonel strata are capped by almost 3000 m of sandstones and conglomerates of western provenance, representing the sedimentary response to renewed pulses of tectonic shortening, which were deposited in alluvial fan, fluvial and eolian settings together with minor lacustrine mudstone (Purilactis Formation). These are covered by 500 m of coarse, proximal alluvial fan conglomerates (Barros Arana Formation). The top of the Purilactis Group consists of Maastrichtian-Danian alkaline lava and minor welded tuffs and red beds (Cerro Totola Formation: 70–64 Ma K/Ar) deposited during an interval of tectonic quiescence when the El Molino–Yacoraite Late Cretaceous sea covered large tracts of the nearby Altiplano-Puna domain. Limestones interbedded with the Totola volcanics indicate that this marine incursion advanced westwards to reach the eastern CD slope. CD shortening in the Late Cretaceous was accompanied by volcanism and continental sedimentation in fault bounded basins associated to strike slip along the north Chilean magmatic arc to the west of the CD domain, indicating that oblique plate convergence prevailed during the Late

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Cretaceous. Oblique convergence seems to have been resolved into a highly partitioned strain system where margin-parallel displacements along the thermally weakened arc coexisted with margin-orthogonal shortening associated with syntectonic sedimentation in the Salar de Atacama basin. A regionally important Early Paleocene compressional event is echoed, in the Salar de Atacama basin by a, distinctive, angular unconformity which separates Paleocene continental sediments from Purilactis Group strata. The basin also records the Eocene–Early Oligocene Incaic transpressional episode, which produced, renewed uplift in the Cordillera de Domeyko and triggered the accumulation of a thick blanket of syntectonic gravels (Loma Amarilla Formation).

1. Introduction

The Salar de Atacama basin is the largest negative topographic anomaly along the generally "smooth" western slope of the Central Andes in northern Chile (Isacks, 1988; Gephart, 1994, Figs. 1 and 2). The basin overlaps the prominent Central Andean Gravity High (CAGH), a first order oblique NW-trending feature of the residual gravity field (Götze and Kirchner, 1997; Götze and Krause, 2002). The basin has been a long-lived subsiding region since the Paleozoic (Ramírez and Gardeweg, 1982; Breitkreutz et al., 1992; Hartley et al., 1992a; Flint et al., 1993; Muñoz et al., 1997, 2002) and is now filled by more than 8 km of sedimentary rocks (4 s TWT, Macellari et al., 1991). One of the key elements of the basin fill (Fill Unit 2 of Flint et al., 1993) is a sedimentary, with minor volcanics, Cretaceous to Paleogene unit, almost 6 km thick, and generally described in the literature as the "Purilactis Group". The understand-

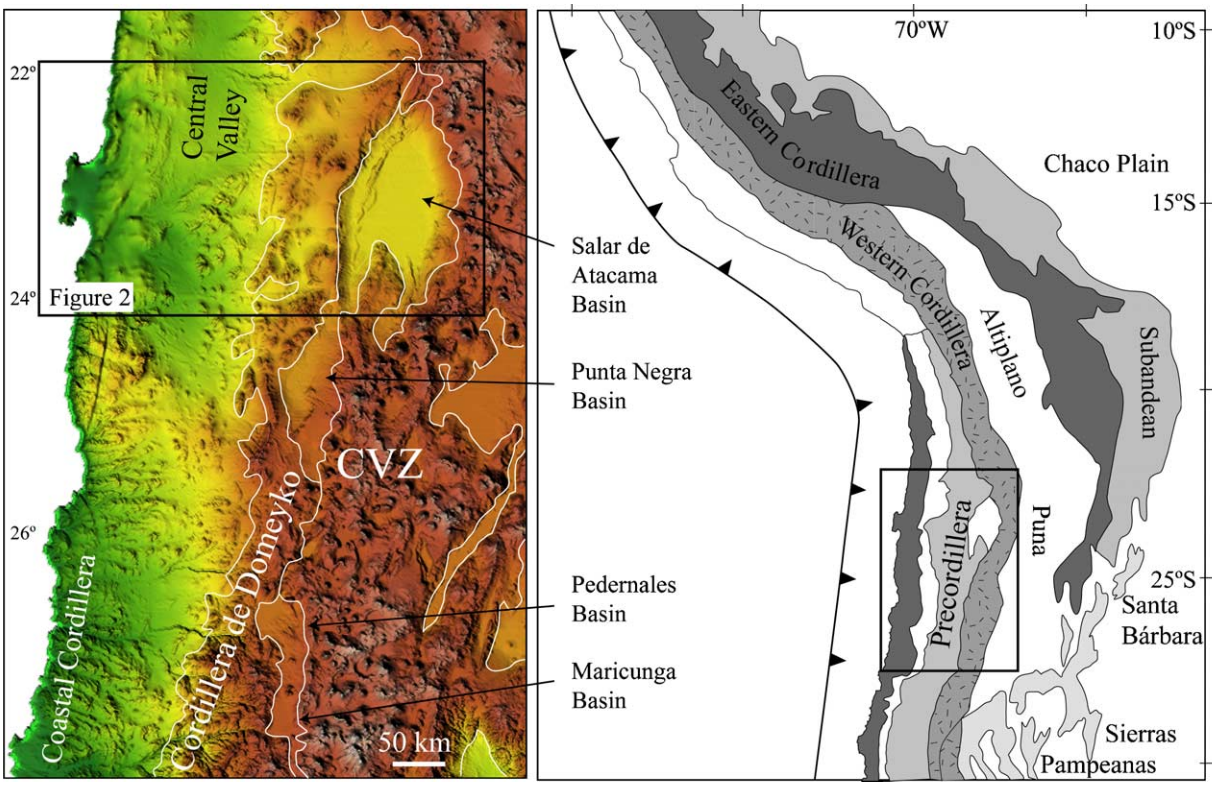

Fig. 1. Topography of the Central Andes between 10 and 30°S showing main morphological units and location of major pre-Andean basins (Salar de Atacama, Punta Negra, Pedernales and Maricunga) in the forearc region of northern Chile.

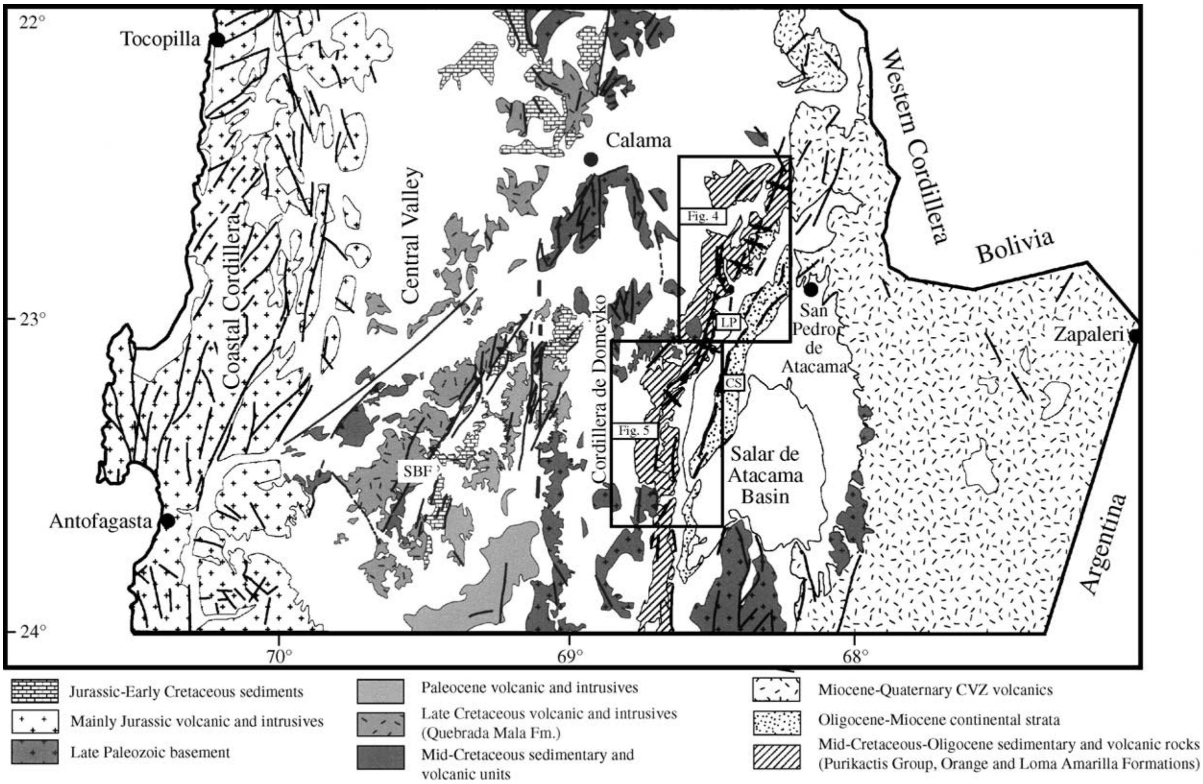

Fig. 2. Simplified geological map of the Antofagasta Region. CS, Cordillera de la Sal, LP, Llano de la Paciencia, SBF, Sierra El Buitre Fault. (Data from Marinovic and García, 1999; Cortés, 2000; 1:1,000,000 Geological Map of Chile, 2002 and Basso and Mpodozis, in preparation).

ing of the stratigraphy, age, facies and sedimentary history of the Purilactis Group is key to any model that tries to account for the Cretaceous to Paleogene tectonic evolution of the Central Andes, as the Purilactis Group has been frequently compared with the sediments of the Salta Rift System in Northwestern Argentina (Salfity et al., 1985; Salfity and Marquillas, 1994, 1999; Galliski and Viramonte, 1988; Coutand et al., 2001) and the sedimentary Cretaceous to Tertiary sequences of southwestern Bolivia (Welsink et al., 1995; Sempere et al., 1997; Horton et al., 2001; McQuarrie et al., 2005). Purilactis strata crop out along the western margin of the Salar de Atacama basin along the El Bordo Escarpment, an NNE-trending, 120 km long, 800 m high cliff, which bounds the basin to the west and defines the eastern edge of the Cordillera de Domeyko (Mpodozis et al., 1993a,b; Maksaev and Zentilli, 1999). Purilactis strata have been imaged by industry seismic lines in the modern Salar basin and tested at the Toconao 1 deep exploration oil well (Muñoz and Charrier, 1999; Muñoz et al., 1997, 2002).

Numerous authors have studied the stratigraphy of the Purilactis Group (Brüggen, 1934, 1942, 1950; Dingman, 1963; Ramírez and Gardeweg, 1982; Marinovic and Lahsen, 1984; Hartley et al., 1988, 1992a; Flint et al., 1989; Charrier and Reutter, 1990, 1994; Hammerschmidt et al., 1992; Arriagada, 1999; Mpodozis et al., 1999). In spite of this, important disagreements about the stratigraphy and age of the Purilactis Group persist, which is no surprise considering the non-fossiliferous nature of the sequence and the scarcity of interbedded volcanic horizons suitable for dating. Consequently, conflicting models about the tectonic evolution of the basin have been proposed. These discrepancies have made it difficult to validate proposed regional correlations with successions in southwestern Bolivian and northwestern Argentina and, in turn, has led to incomplete or faulty regional tectonic syntheses for this important period of the Central Andean history.

In this contribution we present a reassessment of the stratigraphy and age of the Purilactis Group and discuss its implications for regional tectonic models. Detailed mapping carried out during 1997–2002 the recognition of major unconformities, the study of lateral facies changes, the revision of all previously published age-significant data and 10 new K/Ar ages from previously undated volcanic horizons serve as the starting point to develop the new stratigraphic synthesis that we propose in this contribution.

2. Geological framework of the Salar de Atacama basin

The Salar de Atacama basin is relatively a low-lying region, 150 km (N–S) long by 80 km (E–W) wide, at a minimum altitude of about 2300 m, located in the modern Andean fore arc of northern Chile between 22°30′ and 24°30′S. Its center is occupied by the active Salar (salt pan) de Atacama (Figs. 1 and 2). The basin is located directly west of a sharp bend in the present Central Andean arc (Western Cordillera) which, in this particular area, retreats 60 km eastwards from its regional north–south trend (Fig. 1). The basin is bounded to the west by the Cordillera de Domeyko (or Precordillera), a basement range that was uplifted as a consequence of Cretaceous and Eocene tectonic events and reaches an average altitude of 3000 m.a.s.l. (Ramírez and Gardeweg, 1982; Flint et al., 1993; Charrier and Reutter, 1990, 1994; Mpodozis et al., 1993a,b, 1999).

The main morphological features of the western Salar de Atacama basin and eastern Cordillera de Domeyko region include, from west to east, the Cordillera de Domeyko, El Bordo Escarpment, Llano de La Paciencia, the Cordillera de la Sal and, finally, the active Salar de Atacama salt pan (Fig. 2). The Cordillera de Domeyko is formed by Late Carboniferous to Early Permian rhyolitic ignimbrites and domes, associated with volumetrically minor basaltic to andesitic lavas and intruded by granitoid plutons which yield K/Ar, Rb/Sr and U/Pb (zircon) ages ranging between 300 and 200 Ma (Davidson et al., 1985; Mpodozis et al., 1993a; Breitkreutz and van Schmus, 1996). The upper levels of the volcanic succession include, north of Cerro Quimal (Fig. 4), 1200 m of fossiliferous, plant-bearing, lacustrine strata, which recently have been attributed a Triassic age (fossil flora, U/Pb ages, Basso and Mpodozis, in preparation). The eastern border of the Cordillera de Domeyko coincides with the 'El Bordo' Escarpment where, unconformably overlying the Paleozoic and Triassic basement, a thick succession of Cretaceous to Miocene continental sediments, including the Purilactis Group, is exposed (see below). Further east the Llano de la Paciencia is a narrow 80 km long N–S, and 8 km wide sub-basin filled by coalescing Quaternary alluvial fans (Jolley et al., 1990) separated from the main Salar de Atacama basin by the Cordillera de la Sal, a still tectonically active north–south ridge of complexly deformed Oligocene–Pliocene, evaporite-rich, continental sediments and ignimbrites (Ramírez and Gardeweg, 1982; Flint, 1985; Flint et al., 1993; Wilkes and Görler, 1994; Kape, 1996; Mpodozis et al., 2000; Blanco et al., 2000). The older and best exposed sequences of the Mesozoic to Cenozoic Salar de Atacama basin fill occur in a series of major E–W to NW–SE trending canyons, cutting through the El Bordo Escarpment (Dingman, 1963; Ramírez and Gardeweg, 1982; Naranjo et al., 1994; Flint et al., 1993; Charrier and Reutter, 1994; Mpodozis et al., 2000). Further to the east, but unexposed, the Purilactis Group also forms the lower section of the Llano de la Paciencia and part of the modern Salar de Atacama basin sedimentary fill (Macellari et al., 1991; Muñoz et al., 1997).

3. Stratigraphy and structure of the El Bordo Escarpment area

Recent regional 1:50,000 and 1:100,000 mapping along the El Bordo Escarpment (Arriagada, 1999; Mpodozis et al., 1999; Arriagada et al., 2002) allowed the identification of seven major stratigraphic sequences, probably spanning the Late Cretaceous–Oligocene time interval. Mpodozis et al. (1999) included all seven of these units within the Purilactis Group and briefly described their main stratigraphic characteristics. However, after recognizing the importance of a regional unconformity between the Paleocene "Orange" Formation and the older strata (see Fig. 3) we prefer here to limit the Purilactis Group to the

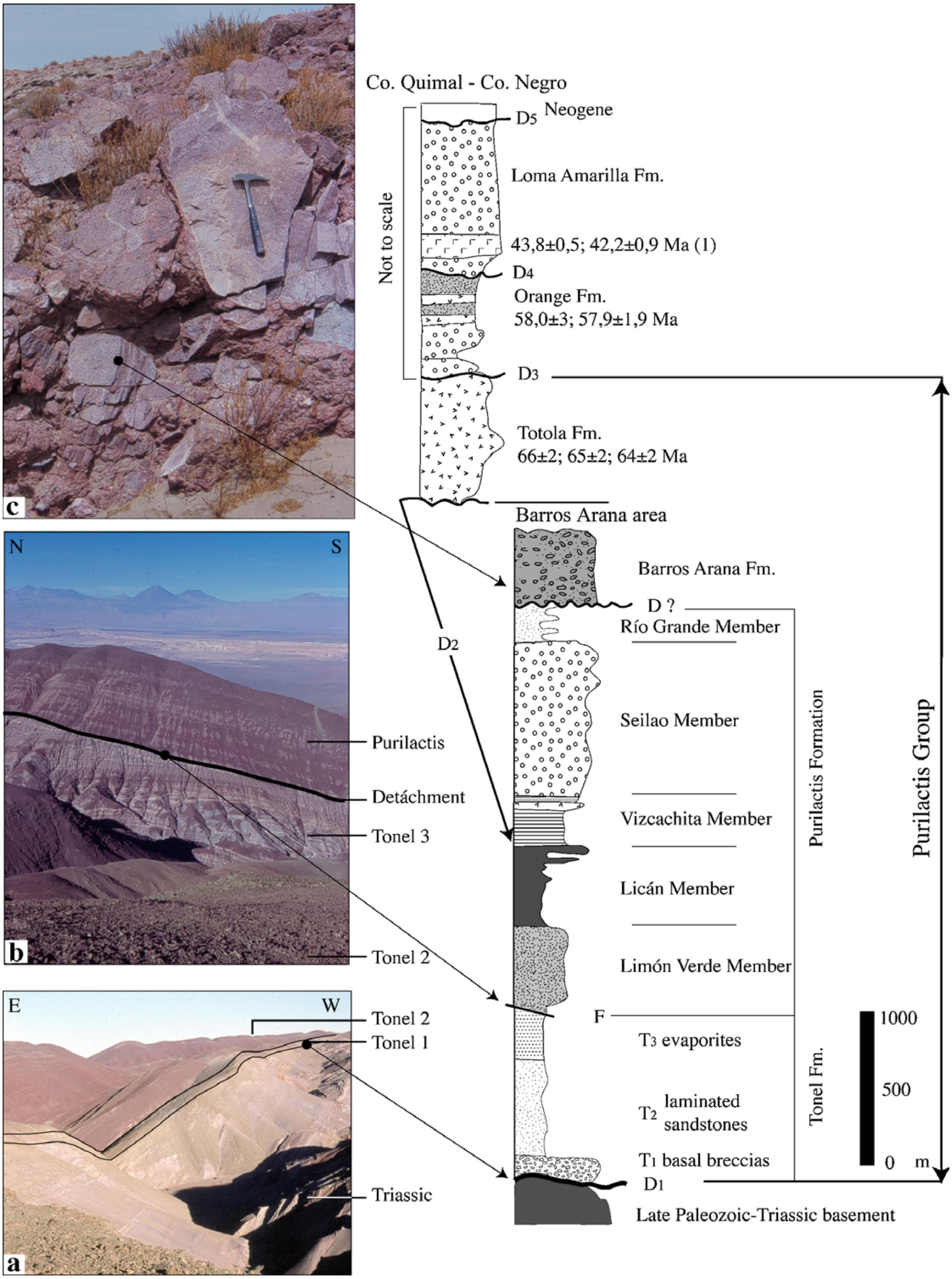

Fig. 3. Generalized stratigraphic section of the Purilactis Group and Paleogene strata cropping out along El Bordo Escarpment, showing main units and internal relations. (1) $^{39}Ar/^{40}Ar$ ages from Hammerschmidt et al. (1992); (D1–D5): major regional unconformities; (a–c) field photographs. More details in text.

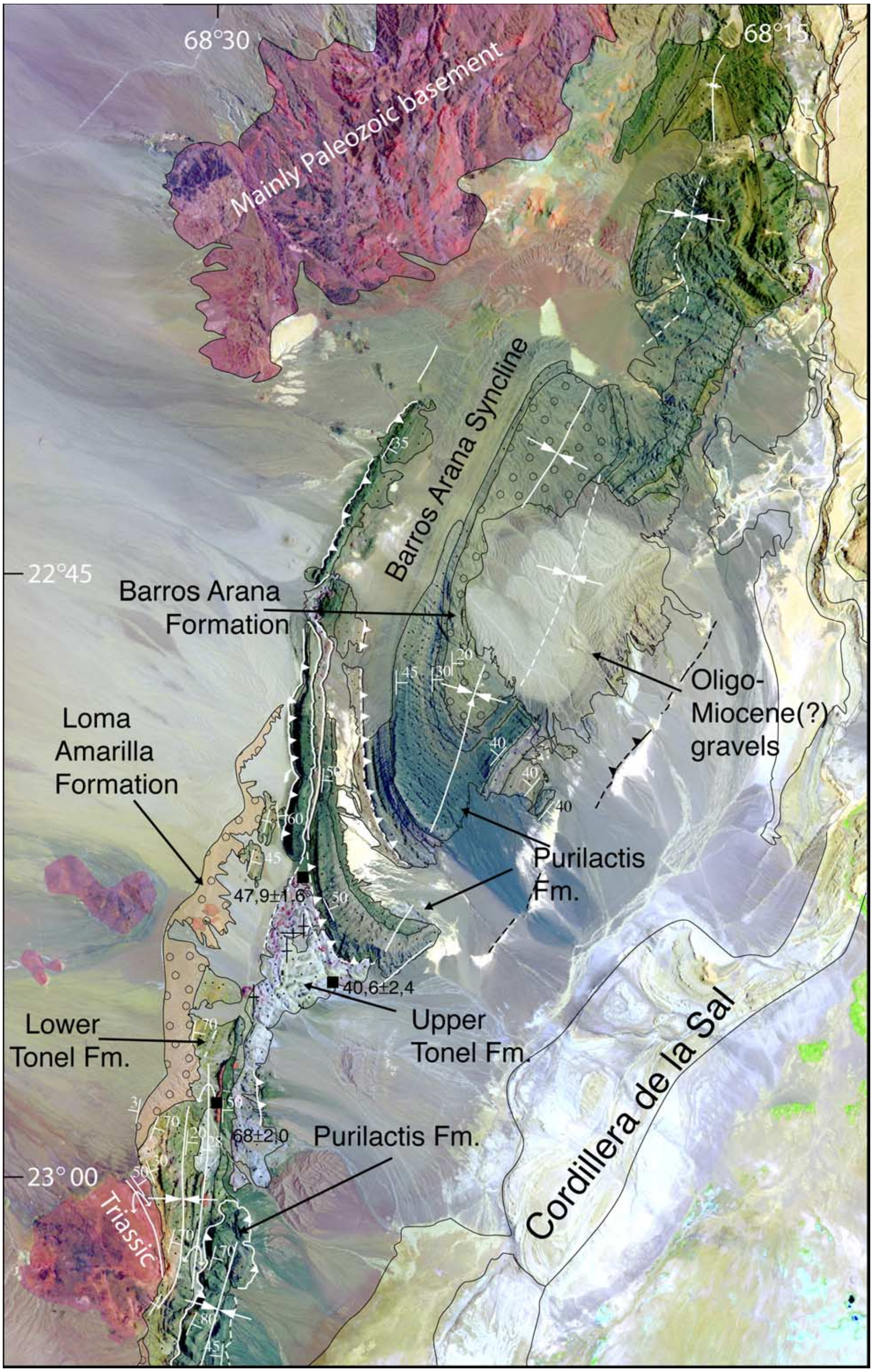

Fig. 4. Main geological elements of the northern El Bordo Escarpment domain (Cuesta de Barros Arana region). Location in Fig. 3.

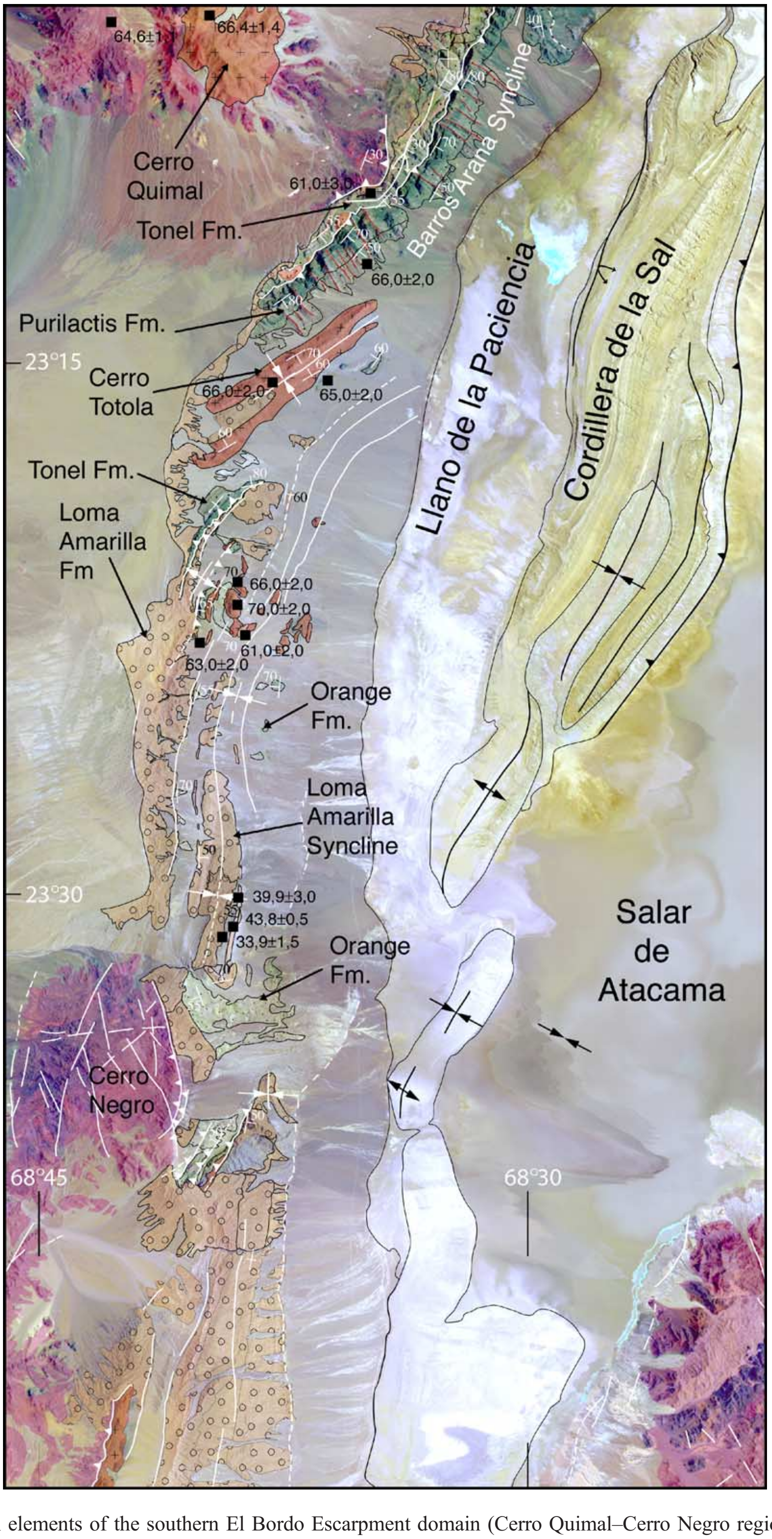

Fig. 5. Main geological elements of the southern El Bordo Escarpment domain (Cerro Quimal–Cerro Negro region). Location in Fig. 4.

lowermost four units, which share an interrelated evolution and history. The redefined Purilactis Group comprises (from bottom to top) the Tonel, Purilactis, Barros Arana and Totola Formations (Figs. 3–5). During the Neogene the basin continued to evolve showing a complex history of Neogene sedimentation, volcanism and deformation (see Wilkes and Görler, 1994; Blanco et al., 2000; Kape, 1996; Mpodozis et al., 2000), to be discussed elsewhere.

The main structure affecting the Cretaceous sedimentary sequences along the El Bordo Escarpment is a large (up to 15 km wavelength) syncline (Barros Arana Syncline), which extends for more than 80 km between Cerro Totola and Pampa Vizcachita (Figs. 4 and 5). Fold geometry changes from open-concentric to the north (Cuesta de Barros Arana) to tight-chevron in the south (east of Cerro Quimal) while the trend of its axial surface shifts from NNE (Barros Arana area, Fig. 4) to N60°E (Cerro Totola region, Fig. 5). Near Cuesta de Barros Arana, the syncline is bounded to the west by outcrops of east dipping Late Paleozoic volcanics which form the easternmost ranges of the Cordillera de Domeyko (Cerros de Tuina) although the contact between basement and Cretaceous strata is buried below Neogene gravels (Fig. 4). Only further south, around Cerro Quimal can a complete structural section be investigated. There, Tonel Formation strata forming the western limb of the Barros Arana Syncline unconformably overlie Triassic sediments and volcanics which are deformed in an east-verging, overturned anticline (Quimal Anticline, Fig. 4). The anticline appears to be a fault propagation fold associated with a blind thrust, rooted to the west in the Late Paleozoic basement core of the Cordillera de Domeyko.

All the Cretaceous units (Tonel, Purilactis, Barros Arana and Totola) are folded. A strong angular unconformity between the "Orange Formation" (see below) and the Purilactis Group south of Cerro Totola suggests that the Quimal-Barros Arana fold system was already formed by the early Paleocene although its original geometry was later modified during Eocene (Incaic) and Neogene deformation episodes (see Arriagada et al., 2000). Late stage reactivation can be seen along the western limb of the Barros Arana Syncline at Cerros de Purilactis where a detachment separating the Purilactis Formation from the upper evaporites of the Tonel Formation acts as the decollement level for a west verging back-thrust (Purilactis Fault) carrying Purilactis strata over Miocene gravels (Fig. 4). Secondary west-verging back-thrusts tectonically repeat Purilactis beds along the western limb of the syncline.

Folding of the axis of the Barros Arana Syncline around the Cerro Quimal basement block (Figs. 4 and 5) was considered by Arriagada et al. (2002, 2003) to be a consequence of the clockwise rotation of crustal blocks during the Eocene Incaic deformation. The effects of the Incaic episode seem to be especially important in the Cerro Negro area (Fig. 5) where discrete basement blocks are thrust eastwards over the Purilactis Group, post-Purilactis Paleocene–Eocene strata ("Orange" Formation) and syntectonic Eocene–Oligocene strata (Loma Amarilla Formation). Deformation is particularly important in the thrust footwalls where the Orange Formation show a complex array of tight, hectometric scale isoclinal to chevron folds with subvertical axial surfaces.

4. Lower Purilactis Group (Late Cretaceous?): Tonel, Purilactis and Barros Arana Formations

The lowermost sedimentary units of the redefined Purilactis Group are essentially the same originally recognized by Brüggen (1942). The Tonel Formation was first described by Brüggen as the Salina de Purilactis Formation and was renamed as the Tonel Formation by Dingman (1963). The Purilactis Formation was also described first by Brüggen ("Porphyritic Purilactis Formation"). Finally, the Barros Arana Formation (Barros Arana Strata in Arriagada, 1999 and Mpodozis et al., 1999), which crops out at the core of the Barros Arana Syncline (Fig. 4) is equivalent to the Conglomerados de Purilactis Formation of Brüggen. It was also recognized as a separate unit by Hartley et al. (1992a) and Flint et al. (1993), which they included within the Cinchado Formation, described further south by Ramírez and Gardeweg (1982).

5. Tonel Formation

This succession comprises red beds and evaporites deposited in gentle angular unconformity (D1, Fig. 3) over Late Triassic volcanic and sedimentary rocks northeast of Cerro Quimal (Fig. 4). Its outcrops extend for more than 60 km along the El Bordo Escarpment (Cordón de Barros Arana to Cerro Oscuro). The lower beds of the Tonel Formation (T1, Fig. 3) include 60 m of medium grained breccias and conglomerates with basement-derived andesitic, rhyolitic and sedimentary clasts (alluvial fan facies, deposited in a proximal alluvial fan/valley fill environment, Hartley et al., 1992a). The middle Tonel (T2), in contrast, is formed of 400 to 1000 m of finely bedded, laminated red-brown sandstones, some of them displaying planar cross bedding, and gypsum nodules. In places the sandstones alternate with thin (centimeter-thick) gypsum layers. At Cerros de Tonel, massive and complexly deformed anhydrite deposits of unknown total thickness form the upper Tonel Formation (T3, Figs. 3 and 4). According to Hartley et al. (1992a), the Tonel Formation represents a playa/sabkha facies association accumulated in playa mudflat and fringing sand flat environment. Paleocurrent measurements generally indicate a western source provenance. Numerous, disrupted lamproitic blocks, ("hornblendites" Dingman, 1963) several meters in diameter, clearly visible as "red specks" in the TM satellite images are embedded within the upper anhydrite sequence. These coarse, typically green, porphyritic rocks, present large (1–4 mm) amphibole, pyroxene and plagioclase phenocrysts. In many places they show clear intrusive contacts with the host rocks and possibly represent part of a dismembered, post Tonel, dyke swarm complex.

6. Purilactis Formation

A faulted (detached) contact over the Tonel Formation marks an abrupt upward transition to the Purilactis Formation in the Barros Arana Syncline (Fig. 4). The Purilactis Formation as a whole comprises almost 3000 m of sandstones, red mudstone and minor conglomerate with andesitic to rhyolitic clasts quartz and abundant potassium feldspar. Clasts of the mafic hornblende-rich dykes, that intrude the Tonel Formation, and fossil-bearing Jurassic limestone are also present (Brüggen, 1942; Dingman, 1963; Ramírez and Gardeweg, 1982). Paleocurrent data indicate a western source for the sediments (Hartley et al., 1988, 1992a).

Hartley et al. (1992a) divided the formation into five members of regional extent. The lowermost (Limón Verde) includes 400 m of fine to coarse, reddish, cross-bedded sandstones, conglomerates and thin red mudstones deposited as sheet floods in an alluvial fan setting (Hartley et al., 1992a). An overlying member of 700 m of red sandstones, mudstones, minor conglomerate and evaporites (Licán Member), which represent channel and sheet flood sediments accumulated in a medial/distal alluvial fan and playa environment. Continuing upwards (Fig. 3), the Vizcachita Member, is formed by almost 300 m of light-green, cross-bedded, eolian sandstones and structureless fluvial sands which, near the top in Quebrada Seilao, include a 40 m thick, strongly altered, porphyritic andesitic lava. The Vizcachita sandstones gradually merge upwards with more than 1000 m of coarse conglomerates and sandstones (Seilao Member, Fig. 3) with volcanic clasts up to 30 cm in diameter, which accumulated as sheet flood and channelized sandstones during repeated cycles of alluvial fan progradation from the west (Hartley et al., 1992a). Finally, the top of the Formation in the Barros Arana and Cerros de Purilactis areas corresponds to 250 m of "varved" black mudstones and sandstones (Río Grande Member) which Hartley et al. (1992a) considered to have been deposited in a shallow lake environment.

7. Barros Arana Formation

The Purilactis Formation is overlain by a sequence of 550 m of coarse-grained conglomerates, which crops out at the core of the Barros Arana Syncline (Figs. 3 and 4). The reddish, well-cemented conglomerates and associated sandstones form laterally extensive beds, up to 50 m thick, with subrounded and moderately sorted clasts up to 20 cm in diameter, composed mainly by Paleozoic granitoids and rhyolites and minor andesite and limestone. Imbrication indicates a western provenance. According to Hartley et al. (1992a), this unit corresponds to an association of facies accumulated as proximal sheet and channel conglomerates and high-density flood and debris flows deposited in a proximal alluvial fan setting.

Similar rocks are also exposed at Cerros de Ayquina, approximately 45 km to the north of Cerros de Purilactis (Marinovic and Lahsen, 1984). When compared to the underlying Purilactis Formation clasts, the Barros Arana Formation (and the Ayquina conglomerates) shows a noticeably larger volume of basement-derived granitoid clasts which may indicate deep exhumation of the Cordillera de Domeyko during an increasing period of erosion and uplift as is also reflected by the much more energetic depositional environment.

8. Age constraints

One of the main difficulties in relation to the Purilactis, Tonel and Barros Arana Formations concerns their poorly constrained ages. No hard data about the age of the Tonel Formation is available although the formation is undoubtedly post-Triassic as indicated by a new 240 Ma U/Pb age obtained on the underlying dacitic tuffs near Cerro Quimal (Basso and Mpodozis, in preparation). Two K/Ar whole rock dates of the altered, dismembered lamproitic dykes associated with the upper member of the Tonel Formation (PC-1 and PC-22, Table 1) collected by the authors south of Cordón de Barros Arana (Fig. 4) gave ages of 40.6±2.4 and 47.9±1.6 Ma. Macellari et al. (1991) obtained an older, whole rock, age of 63.6±2.8 Ma for the same hornblende-rich dykes at Quebrada Los Cóndores even though they admit that this may represent a minimum age considering the strongly altered state of the dykes. We concur and believe that these results represent only reset values as an andesitic dyke intruding the middle member of the Tonel Formation yielded a K/Ar age of ca. 68 Ma (see below). These data only allow the Tonel to be placed in the general Jurassic–Cretaceous interval.

Tighter constraints exist, however, for the Purilactis Formation. The occurrence of reworked mid-Jurassic marine fossils (*Vaugonia v. l. gottschei Moericke*, *Perisphinctes* sp.; Ramírez and Gardeweg, 1982) and the fact that part of the strata are unconformably covered by 66–64 Ma volcanic rocks of the Totola Formation (Figs. 4 and 5), as well as intruded by a late Cretaceous complex of dykes and stocks dated (K/Ar) between 70 and 61 Ma (see below) constrain the age of the Purilactis Formation to the Late Jurassic–Cretaceous interval. Flint et al. (1989) reported a $^{39}Ar/^{40}Ar$ date from a lava flow

Table 1
Radioisotopic ages

Sample	UTM (N, Lat)	Coordinates (E, Long)	Geological unit	Method	Material	%K	Ar rad. (nl/g)	%Ar atm.	Age (Ma)	Error (2 Sig)
PC-1	7465998	557760	Dismembered lamproitic dykes	K/Ar	Whole rock	0.565	0.901	59	40.6[a]	2.4
PC-22	7469772	555538	Dismembered lamproitic dykes	K/Ar	Whole rock	2658	5015	24	47.9[a]	1.6
KP3L (1)	7476621	565866	Lava in Vizcachita Member	Ar/Ar	Clinopyroxene				64.0[b]	10.0
PC-26	7428249	540074	Totola Formation	K/Ar	Whole rock	0.944	2429	20	65.0	2.0
PC-27	7428898	535909	Totola Formation	K/Ar	Whole rock	3704	9695	10	66.0	2.0
PC-28	7414484	535575	Totola Formation	K/Ar	Whole rock	1597	3863	26	61.0	2.0
									64	2.0
PC-29	7418040	536420	Dyke (intruding Totola Fm)	K/Ar	Whole rock	3.28	8604	26	66.0	2.0
PC-33	7434314	542265	Dyke (intruding Purilactis Fm)	K/Ar	Biotite	5712	14,872	13	66.0	2.0
PC-39	7460116	551591	Dyke (intruding Tonel Fm)	K/Ar	Biotite	7018	19,017	36	68.0	2.0
PC-40	7437536	542930	Dyke (intruding Tonel Fm)	K/Ar	Whole rock	2203	5315	58	61.0	3.0
PC-17	7415123	534538	Dyke intruding Purilactis Fm)	K/Ar	Amphibole	1372	3391	11	63.0	2.0
PC-52	7416397	536318	Porphyritic stock	K/Ar	Amphibole	1038	2.88	23	70.0	2.0
MAF-306 (2)	7446800	534900	Cerro Quimal stock	K/Ar	Biotite	7598	19,967	22	66.4	1.4
MAF-325-C (2)	7446500	529500	Cerro Quimal stock	K/Ar	Biotite	7071	18,055	15	64.6	1.1
Pu-1 (3)	7390991	537328	Loma Amarilla Fm	Ar/Ar	Biotite				43.8	0.5
Pu-13 (3)	7419224	536576	Loma Amarilla Fm	Ar/Ar	Biotite				42.2	0.9
To-432 (2)	7398000	535500	Loma Amarilla Fm	K/Ar	Plagioclase	0.362	0.567	76	39.9	3.0
PC-13	7398993	535322	Loma Amarilla Fm	K/Ar	Amphibole	0.355	0.473	54	33.9[a]	1.5

(1) Flint et al. (1989), (2) Ramírez and Gardeweg (1982), (3) Hammerschmidt et al. (1992).
[a] Minimum age.
[b] Value of unknown significance.

interbedded in the middle part of the Purilactis Formation (Vizcachita Member) at Quebrada Seilao, which yielded an "age" of 63.8±1.9 Ma (Fig. 3). However, the extremely disturbed Ar release spectra, the strongly altered nature of the sampled rocks and the very low K content of the analyzed clinopyroxene led us to consider this value as a "minimum age" with uncertain geological meaning.

Arriagada et al. (2000) carried out a detailed paleomagnetic survey on the Tonel and Purilactis Formations, sampling 10 sites in the middle member of the Tonel Formation and 20 other sites mainly in the lower part of the Purilactis Formation (Limón Verde and Licán). Results indicate that all these samples acquired their characteristic magnetization during a normal polarity interval. This observation and the Late Cretaceous age of the overlying volcanic rocks of Cerro Totola strata (which, in contrast, show both normal and reverse polarities) led Arriagada et al. (2000) to suggest that deposition and magnetization of the lowermost units of the Purilactis Group occurred during the Cretaceous normal polarity superchron, between 119 and 84 Ma (Cande et al., 1988; Besse and Courtillot, 1991). Hartley et al. (1991, 1992b) also report paleomagnetic data for the Purilactis Formation in the Barros Arana area (Fig. 4). Samples came, apparently, from higher stratigraphic levels and exhibit both normal and reverse polarities which is consistent with their younger age, although Arriagada et al. (2000) draw attention to the possibility that these may be secondary magnetizations and not primary values.

9. Tectonic controls on sedimentation

Flint et al. (1989), based on the study of the outcrops along the Barros Arana region, suggested that the Purilactis basin was formed in a contractional setting after a compressive deformation event, which led to the emergence of the Proto-Cordillera de Domeyko in the "mid-Cretaceous". However, the study of some of the ENAP (Chilean National Oil Company) seismic lines which seemingly show that the Purilactis Group strata thickens westwards later led Flint et al. (1993) to propose that the tectonic environment during deposition of the Purilactis was dominantly extensional instead. Macellari et al. (1991) suggested that the Purilactis Fault, which bound to the west the Cretaceous outcrop belt, may have formed in the Early Cretaceous as an east-dipping listric normal fault, which originally delineated the western boundary of the Purilactis basin and was inverted during the Tertiary.

However, even if structural relations in the Barros Arana area (Fig. 4) do not totally disagree with an extensional origin for a forerunner of the Neogene Purilactis fault, during the earliest stages of basin development, our observations near Cerro Quimal indicate that, at least there, the lower units of the Purilactis Group (Tonel Formation) were deposited under a compressional regime. The sandstones of the Tonel Middle Member (T2) exposed along the eastern limb of the east-verging Quimal Anticline, form part of a panel which exhibit progressive internal unconformities as shown by a systematic decrease in dip from very steep (80°E) to gentle (30°E) when moving away (eastwards) from the basement-cover contact. Individual bed thickness increase from west to east (see Fig. 6) while the strata are deformed by second-order upwards attenuated synclines, following a pattern described elsewhere as typical for syntectonic growth strata deposited in an active compressional environment (Riba, 1976; Barrier, 2002). Growth structures and compressional folding led us to infer that the lower and middle section (T1, T2, Fig. 3) of the Tonel Formation accumulated in a compressional context during the growth of the Quimal Anticline in the Late(?) Cretaceous (Arriagada et al., 2002).

The evaporites of the Upper Tonel indicate a period of reduced sediment supply (and relative tectonic quiescence?), which culminated with the emplacement of the Tonel mafic dyke swarm (hornblendites). The drastic facies change of the evaporites of Tonel Formation to the thick sequences of sandstones and conglomerates of alluvial fan and eolian-lacustrine facies of Purilactis Formation shows an important increase in the accommodation space and sediment supply, possibly triggered by renewed uplift and shortening in the Cordillera de Domeyko, although we are not been able to find, as in the Tonel case strata with clear "growth" geometrical arrays. Finally, the conglomeratic Barros Arana Formation can be considered an independent unit which prograde, probably, from the west into the Purilactis basin although its age is poorly constrained. The large volume of Paleozoic

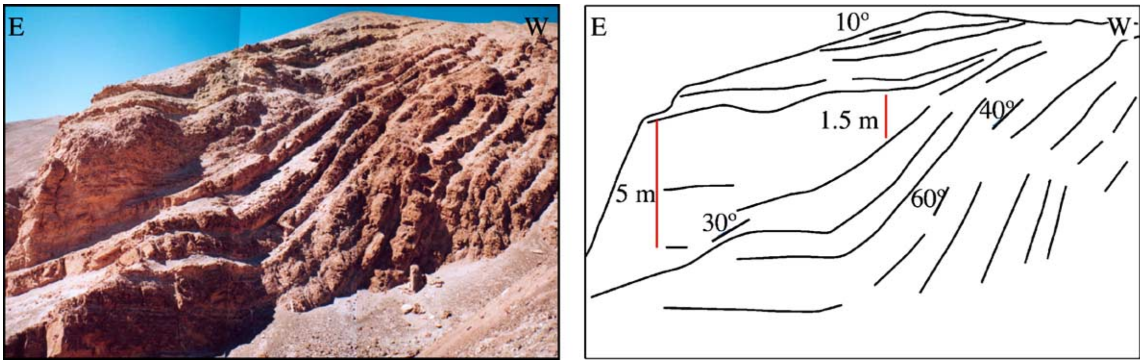

Fig. 6. Syntectonic growth strata in the middle member (T2) of the Tonel Formation, due east of Cerro Quimal.

granitoid clasts in the Barros Arana Formation (Hartley et al., 1992a), which are almost absent in the Purilactis Formation, indicate deep erosion and the unroofing of Paleozoic intrusive complexes, that today constitute most of the exposed rocks in the Cordillera de Domeyko to the west.

10. Upper Purilactis Group (Maastrichtian-Danian): the Totola Formation

To the southeast of Cerro Quimal, playa and alluvial playa fan facies of the Licán Member of the Purilactis Formation and red beds of the Tonel Formation are unconformably overlain (D2 unconformity, Fig. 3) by a volcanic-sedimentary sequence, described by Arriagada (1999) and Mpodozis et al. (1999). The sequence, at the core of the Barros Arana Syncline, includes 800 m of andesitic, andesitic-basaltic and minor dacitic lava flows, interbedded with welded rhyolitic ignimbrites, and, nears the base, coarse red, volcaniclastic conglomerates and sandstones. Some 60 km south of Cerro Totola, at Cerro Pintado, another small outcrop of the Totola Formation can be found. Its base is not exposed and to the west it is overthrust by mafic Late Paleozoic lavas. This outcrop is specially interesting because 100 m of andesitic to basaltic flows are capped by 150 m of red conglomerates and calcite-cemented sandstones which alternate with gray limestone beds up to 5 m thick composed mainly by sparry calcite but including some bioclastic material which point to an organic (marine) origin. The Cerro Totola basalts and basaltic andesites are porphyritic rocks with idiomorphic plagioclase and clinopyroxene phenocrysts. Groundmass is pilotaxitic in part and includes plagioclase microlites, pyroxene and opaque minerals. The intermediate portion of the lava flows is massive, while the flow tops exhibit quartz and zeolite-filled vesicles. The andesitic lavas include albitized plagioclase, amphibole and biotite phenocrysts embedded in a microgranular array of plagioclase and opaque minerals. Calcite occurs in fractures and veins throughout the flows. Dacites are also porphyritic with albitized plagioclase phenocrysts immersed in a devitrified glassy groundmass with plagioclase microlites and opaque minerals. Finally, the ignimbrites are rhyolitic, biotite-amphibole rich, welded tuff, with stretched pumice fragments, and andesitic lithic fragments in a devitrified groundmass.

11. Totola-related intrusives

At Cerro Totola the lavas are intruded by a swarm of subvertical basaltic to dacitic dykes, less than 10 m thick, which also cut the Tonel and Purilactis Formations southeast of Cerro Quimal where they seem to form a radial array around the Cerro Quimal stock. The latter is a Late Cretaceous granodiorite stock, 5 km in diameter, emplaced in Triassic volcanics of the Cordillera de Domeyko basement. North of Quebrada El Salto, dykes trend N60°W and N30–40°W. These dykes may represent feeders to the Totola volcanics as they share the same chemistry and age (Tables 1 and 2). Irregular, smaller stocks, hundred of meters in diameter, also intrude the Totola and Tonel Formations near Quebrada Tonel, Quebrada

Table 2
Representative major and trace elements whole-rock analyses from mafic lavas of the Totola Formation and related dykes

Sample	Lavas						Dykes		
	PC-49	PC-26	PC-28	PC-53	PC-45	PC-27	PC-40	PC-41	PC-33
SiO_2	66.8	63.7	66.0	66.9	66.3	64.2	67.2	72.2	62.7
SiO_2	46.42	46.49	49.53	50.30	53.13	56.46	44.81	47.99	51.69
Al_2O_3	17.50	17.66	17.32	17.26	19.84	20.49	15.04	16.16	17.92
TiO_2	1.21	1.25	0.93	0.92	0.68	0.48	1.05	0.79	0.77
Fe_2O_3	8.77	9.14	8.95	7.04	4.25	3.16	6.11	5.55	4.50
FeO	2.16	2.27	1.04	1.37	1.21	0.72	5.10	3.60	2.01
CaO	8.37	8.65	9.48	7.55	4.29	3.92	9.32	6.91	5.19
MgO	4.66	4.36	3.99	4.68	1.74	1.07	6.47	5.89	2.77
MnO	0.18	0.18	0.18	0.18	0.17	0.14	0.25	0.17	0.18
Na_2O	4.43	4.35	3.34	4.33	4.94	5.78	2.99	4.25	4.10
K_2O	1.63	1.41	2.57	2.39	5.35	5.02	3.37	3.03	5.79
P_2O_5	0.7	0.66	0.7	0.37	0.53	0.29	0.57	0.62	0.48
LOI	3.67	3.44	1.95	3.16	3.58	2.38	4.71	4.85	4.91
Total	99.69	99.86	99.98	99.55	99.71	99.92	99.8	99.82	99.91
Ba	1500	1500	615	645	2000	1400	852	1000	1800
Rb	27	13	61	57	60	91	73	59	118
Sr	1200	1400	944	718	1300	1600	534	699	407
Y	23	19	24	21	18	18	19	20	21
Ni	15	13	7	15	–	–	26	22	–
Cr	23	29	19	54	6	6	52	95	7
Nb	10	6	5	8	15	12	–	–	10
Zr	115	102	104	115	165	199	67	96	157
La	22	25	18.6	18.9	26	24	11	16.5	27
Ce	51	49	43	42	55	51	25	39	53
Pr	6	5.7	5.2	4.9	6.1	5.10	3.1	4.7	–
Nd	27	25	25	23	25	21	15.8	23	26
Sm	6	5.6	5.8	5	4.9	4.3	4.2	5.3	5.38
Eu	1.87	1.67	1.74	1.59	1.43	1.31	1.41	1.58	1.61
Gd	5.5	5.1	5.5	4.7	4.2	3.7	4.4	4.9	4.91
Tb	0.82	0.77	0.81	0.67	0.62	0.6	0.71	0.78	0.79
Dy	4.3	4	4.6	3.9	3.5	3.10	3.7	3.9	3.99
Ho	0.84	0.74	0.85	0.80	0.68	0.59	0.71	0.75	0.73
Er	2.2	2.1	2.5	2.2	1.7	1.77	2	2	2.12
Tm	0.29	0.26	0.32	0.29	0.26	0.23	0.24	0.25	0.28
Yb	1.98	1.89	2.3	2.1	1.94	1.85	1.65	1.79	2.01
Lu	0.29	0.28	0.32	0.31	0.29	0.27	0.24	0.26	0.29

El Salto and southeast of Cerro Pichungo (Fig. 5). The large (25 km^2) Cerro Quimal stock comprises medium-grained amphibole–biotite granodiorites but also small (1 km^2) pyroxene bearing diorite stocks occur west and south of Cerro Quimal. Dacitic stocks and dykes further east display a porphyritic texture, albitized plagioclase amphibole and a small percentage of clinopyroxene (<2%) while the basaltic andesitic to andesitic dykes, outcropping in the same general area are also porphyritic, with plagioclase and pyroxene phenocrysts including biotite in the less acidic facies.

12. Geochemistry and age

Nine whole-rock major and trace element analyses of mafic lavas from the Cerro Totola Formation and associated dykes are presented in Table 2. Silica content varies from 46% to 51% SiO_2. All the rocks show high Na_2O+K_2O values, which allow them to be classified as an alkaline suite according to the various criteria summarized in Rollinson (1993). Chemically the lavas are trachybasalts and basaltic according to the TAS classification of Lemaitre et al. (1989). The dykes (44% to 52% SiO_2) show a slightly higher alkali

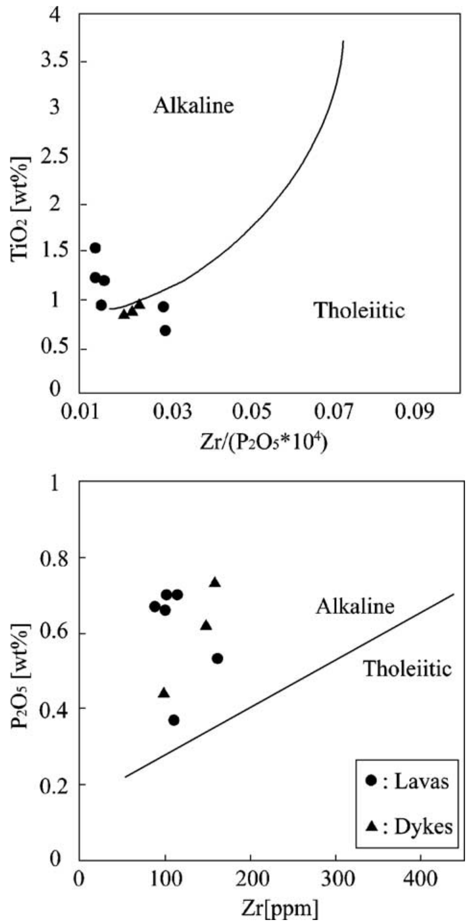

Fig. 7. Winchester and Floyd (1976) discriminating diagrams for basalts. Totola Formation mafic lavas and related dykes.

content as two of the samples (PC 33, PC 40) fall in the tephrite–basanite and tephriphonolite fields of the TAS classification diagram. Despite the fresh appearance in hand-specimen and thin section, the moderate to high loss on ignition values (LOI>2 wt.%) suggests some alkali remobilization by alteration. However, the alkaline affinity of the lavas and dykes is confirmed by their relatively high TiO_2 and P_20_5 content as shown in discrimination diagrams for basalts of Winchester and Floyd (1976, Fig. 7). The lavas and dykes are enriched in incompatible elements (K, Ba, Rb, Sr), when compared to MORB values (Fig. 8). The chondrite-normalized REE patterns present a gentle slope (La/Yb=7–13) and no Eu anomalies. Light REE are only slightly enriched in relation to the heavy REE (La/Sm=3–6, Sm/Yb=2–3, Fig. 9).

We have acquired four new K/Ar ages for the previously undated Totola lavas and five age data for related dykes and intrusives (Table 1). Two of these correspond to concordant K/Ar whole rock ages of 66.0±2.0 and 65.0±2.0 Ma from lava flows forming part of the eastern limb of the Totola Syncline (samples PC 26, PC 27, Fig. 5) which are consistent with a new age of 64.0±2.0 Ma, obtained by Basso and Mpodozis (in preparation). These ages are older than the K/Ar (whole rock), 61.0±2.0 Ma age obtained for a basaltic-andesitic lava flow sampled further south, near Cerro Oscuro (PC 28, Fig. 5). We interpret this number as a minimum value because one andesitic dyke (PC 29) intruding the Totola lavas nearby yielded 66.0±2.0 Ma (whole rock) age. Other

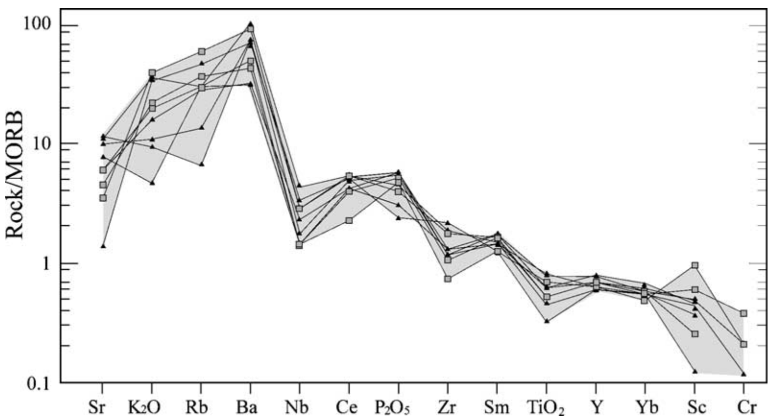

Fig. 8. MORB-normalized spidergram. Totola Formation mafic lavas and related dykes.

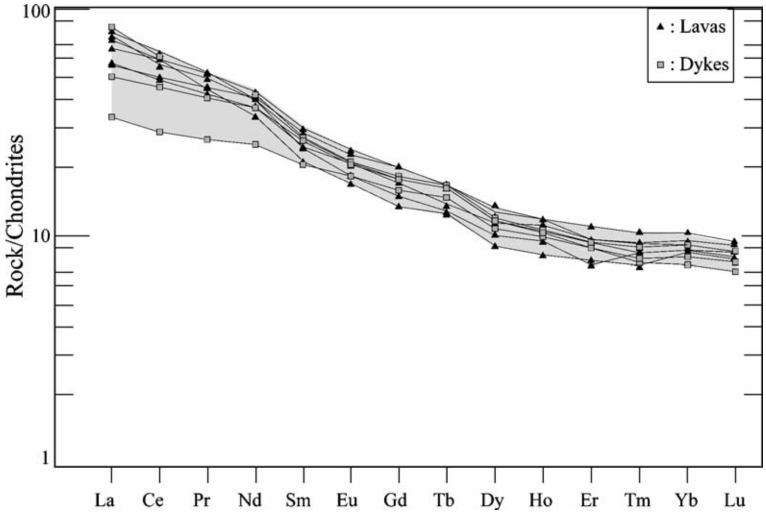

Fig. 9. Chondrite-normalized REE patterns. Totola Formation mafic lavas and related dykes.

K/Ar (biotite) ages from dykes intruding the Tonel Formation north of Quebrada Tonel (PC 39: 68.0±2.0 Ma) and the Purilactis Formation southeast of Cerro Quimal (PC 33: 66.0±2.0 Ma) fall in the same general range indicating the main period of volcanism and dyke emplacement occurred most probably during the Maastrichtian-Danian interval. Only one andesitic dyke (PC 40) intruding the middle member of the Tonel Formation at Quebrada El Salto turned out to be younger (61.0±3.0 Ma, whole rock, see Fig. 5). The larger intrusive bodies also gave similar ages to the Totola Formation. Ramírez and Gardeweg (1982) reported two (biotite) ages of 66.4±1.4 and 64.4±1.4 Ma for the monzonite to granodiorite Cerro Quimal stock for which Andriessen and Reutter (1994) obtained a slightly younger (biotite) age of 63.3±0.7 Ma. We produced two other hornblende ages of 70.2±2.0 and 63.0±2.0 Ma from andesite porphyries (PC 17, PC 52) outcropping 10 km to the south of Cerro Totola and another (K/Ar) biotite age of 68.0±2.0 Ma for diorites that crop out southeast of Cerro Quimal (Basso and Mpodozis, in preparation).

13. Relationships with the Purilactis Formation and local correlations

As it is shown in Fig. 3, in the area between Cerro Quimal and Cerro Negro (Fig. 5) the Totola Formation unconformably overlies the Tonel Formation and the Licán Member of the Purilactis Formation. No direct relations exist between the Totola volcanics and the Barros Arana and/or the upper members of the Purilactis Formation. Hartley et al. (1992a) and Flint et al. (1993) (independently of nomenclature differences) considered that the strata we include in the Totola and Barros Arana Formations could be time-equivalent units. Also, in principle, nothing precludes a hypothetical correlation between the Totola Formation and the Upper Purilactis Formation, including the Vizcachita member. However, although we are aware that this is not a totally solved problem, as the Totola Formation overlies in angular unconformity the Tonel and the Licán units, and no angular unconformity exists between the different units in the Barros Arana area, our preferred interpretation is to consider the Totola Formation as younger than the Barros Arana Formation (see Fig. 3).

Equivalent units to the Totola Formation are almost certainly present in the subsurface of the Salar, forming part the oldest units recognized in the Toconao 1 well. According to Muñoz et al. (1997) at 2980 m below surface, the well penetrated red claystone and siltstone and minor shallow marine carry foraminifera and pollen of Senonian age. Although this strongly folded unit was considered by Muñoz et al. (1997) to unconformably cover a “Paleozoic basement” unit of andesitic lavas, tuffs and sandstones and minor limestone intercalations, review of the seismic lines and original well reports does not prove any definite unconformity and, contrary to described, any Paleozoic diagnostic fossil assemblage. Strata akin to the Totola Formation also outcrop further north along strike with the structural axis of the Barros Arana Syncline, some 100 km NNE of Cerro Quimal where they have been described as the Lomas Negras Formation (Lahsen, 1969; Marinovic and Lahsen, 1984). This unit (base unexposed) crops out in an erosional window below Miocene–Pliocene volcanics and includes 500 m of "andesitic" lavas, interbedded with multicolored conglomerates, sandstone, mudstone and thin beds (1–3 m) of marine, fossiliferous, (oolitic) limestones including bivalves (*Brachiodontes* sp.), foraminifera, fish bones, trace fossils and plant remains (Marinovic and Lahsen, 1984). It can be considered a direct equivalent to the Totola Formation as it shown by a $^{39}Ar/^{40}Ar$ (biotite) plateau age of 66.6±1.2 Ma for a lapilli tuff near the base of the formation reported by Hammerschmidt et al. (1992).

14. Post Purilactis Paleogene sequences

14.1. The Paleocene–Eocene "*Orange*" Formation

Folded Mesozoic units of the Purilactis Group and volcanics rocks of the Cerro Totola Formation (and associated intrusives) forming part of the Barros Arana Syncline are overlain by a clastic sequence up to 900 m thick, with a characteristic orange color ("Orange" Formation). The sequence outcrops mainly to the south of Cerro Totola and in the area around Cerro Negro, where it is deformed by a complex set of tight (tens to hundreds of meters wavelength) upright folds (Fig. 5). The base of the Orange Formation is exposed north of the road connecting the Salar de Atacama with the Gaby porphyry copper deposit where remarkable angular unconformity (D3, Fig. 3) separate shallow dipping strata of the Orange Formation from deformed, subvertical sandstones of the middle member of the Tonel Formation (Fig. 5). This fining upward succession was first recognized as an independent unit by Arriagada (1999) and Mpodozis et al. (1999). The lowermost levels of the formation are constituted by roughly 400 m thick, coarse red conglomerates with clasts up to 1 m in diameter, intercalated with coarse reddish, laminated cross bedded, arkosic sandstones. Clasts are mainly derived from the Cordillera de Domeyko basement (granitoids, rhyolitic porphyries, mafic lavas) but dacitic and porphyritic andesite clasts similar to the Late Cretaceous and Paleocene volcanics outcropping west of the Cordillera de Domeyko (Cinchado and Quebrada Mala Formations, Montaño, 1976; Marinovic and García, 1999) are also found. The proximal alluvial fan facies association of the lower member grades upward into 500 m of "orange" channelized and sheet-flood medium- to fine-grained sandstones showing both planar and cross bedded stratification which alternate with thin mudstone layers and evaporites (gypsum) that became increasingly thick towards the upper levels of the formation (saline lake and/or playa facies association).

14.1.1. Age and tectonic setting

The Orange Formation is equivalent to a thick (up to 1400 m) sedimentary sequence (Unit H, Jordan et al., submitted for publication), which in the Salar subsurface onlaps folded Late Cretaceous strata and fills the intervening synclinal depressions. In the Toconao 1 borehole Unit H is formed by volcanoclastic sandstones interbedded with conglomerates and claystones (Muñoz et al., 1997, 2002; Jordan et al., submitted for publication). Some 70 km south of Cerro Quimal (Pan de Azúcar, region) andesitic lava horizons interbedded in the Formation have been dated at 57.9±1.9 and 58.0±3 Ma (K/Ar, whole rock, Gardeweg et al., 1994) suggesting a Late Paleocene age. At Loma Amarilla (Fig. 5) the unit is covered unconformably by Eocene–Oligocene strata (Loma Amarilla Formation, Fig. 3). The strong angular unconformity at the base of the Orange Formation and the rapid upward change from proximal alluvial fan to playa lake facies is consistent with the unit being a post-tectonic sedimentary sequence, which accumulated after a new and intense pulse of compressional deformation, which probably occurred during the early Paleocene (see Discussion below).

14.2. The Eocene–Early Oligocene Loma Amarilla Formation

A thick sequence of mostly unconsolidated, coarse, reddish to gray, gravels and conglomerates, identified here as the Loma Amarilla Formation, forms most of the El Bordo ridgeline from Cuesta de Barros Arana to Cerro Negro (Figs. 4 and 5) where it unconformably overlie almost all the older units. Earlier workers considered most of this sequence to be of Miocene age (Ramírez and Gardeweg, 1982; Marinovic and Lahsen, 1984; Hartley et al., 1992a). The Loma Amarilla gravels fill a very irregular topography carved in the Cretaceous and Paleocene–Eocene sequences, to the west of the El Bordo Escarpment is sculptured by a gently west dipping pediplain surface above which rest remains of Late Miocene (10–8 Ma) ignimbrites are found (Naranjo et al., 1994; Mpodozis et al., 2000). North of Cerro Quimal, subhorizontal Loma Amarilla strata onlap Paleozoic and Purilactis outcrops although the same gravels appear almost subvertical, east of the El Bordo ridgeline. Progressive internal unconformities visible at the northern termination of the Loma Amarilla syncline and along the El Litio-Baquedano road (Fig. 5) indicate the syntectonic character of the sequence. The exposed thickness of the Loma Amarilla Formation can surpass 2500 m. Its lower levels include (Loma Amarilla–Cerro Negro area) up to 200 m of white tuffs, coarse conglomerates with tuffaceous matrix and reworked coarse pyroclastic flows with abundant, porphyritic dacitic clasts (Fig. 3). The thick upper beds are a monotonous succession of massive, poorly bedded coarse conglomerates, unconsolidated gravels, and minor sandstone. Conglomerates contain clasts from 5 to 50 cm in diameter, including subrounded boulders of Paleozoic granitoids, rhyolites, dacite and andesite. According to Hartley et al. (1992a), who carefully studied a detailed section south of Cerro Negro, they represent an association of proximal alluvial fan facies including hyperconcentrated debris flow deposits.

14.2.1. Age and tectonic setting

The only horizon suitable for dating in the gravel-rich Loma Amarilla Formation is the tuffaceous basal horizon where Ramírez and Gardeweg (1982) first obtained a K/Ar plagioclase age of 39.9±3.0 Ma. This value was later refined by Hammerschmidt et al. (1992) who obtained two $^{39}Ar/^{40}Ar$ plateau (biotite) ages of 43.8±0.5 (Pu-1) and 44.2±0.9 Ma (Pu-13) which indicate that accumulation of the Loma Amarilla Formation began during the middle Eocene. As a complement, we obtained a new K/Ar age of 59.1±2 Ma (BF-451) for an amphibole-bearing andesite clast from a volcanoclastic conglomerates near the base of the formation. This value is similar to the ages reported for the lavas interbedded within the Orange Formation south of the Salar basin by Gardeweg et al. (1994) and the numerous 63 to 55 Ma K/Ar ages informed for lavas of the Cinchado Formation exposed west of the Cordillera de Domeyko (Marinovic and García, 1999) indicating the sedimentary reworking of Paleocene volcanic units. Another (porphyritic) clast from volcanoclastic conglomerates, 100 m above the basal contact of the formation yielded a younger (K/Ar, hornblende) age of 33.9±0.3 Ma. Straightforward acceptance of these data implies that the entire Late Eocene is traversed in the first 100 m of he formation although the very low potassium content (0.36% K_2O) of the analyzed material casts doubt over the accuracy of the age. It is noteworthy that this value falls in the range of the total degassing $^{39}Ar/^{40}Ar$ biotite ages (32.05+0.6; 36.4±1.0 Ma) of samples Pu-1 and Pu-13 reported by Hammerschmidt et al. (1992) which may well indicate a regional event of Ar loss, probably initiated in the Miocene, according to the 20–10 Ma age of the lowest temperatures steps in both samples.

Studies of seismic lines in the Salar allow Loma Amarilla strata to be compared to seismic sequence J in the Salar subsurface (Jordan et al., submitted for publication), which, in the Toconao 1 borehole are represented by 600 m of multicolored sandstone and conglomerates (Muñoz et al., 1997). Sequence J overlies (Paleocene–Eocene?) sequence H above a prominent erosional unconformity and is overlain by seismic sequence K, which can be correlated with the Oligocene–Lower Miocene Paciencia Group. This latter unit, which crops out extensively in the Cordillera de la Sal (Fig. 4), includes thin ash layers whose ages span from ca. 28 to 20 Ma (Marinovic and Lahsen, 1984; Mpodozis et al., 2000). These age constraints indicate that the Loma Amarilla Formation was deposited during the Late Eocene–Oligocene interval being no older than 44 Ma and no younger than 28 Ma. The Loma Amarilla Formation is herein interpreted as a thick sedimentary blanket, syntectonic with the Eocene “Incaic” deformation that affected large tracts of Northern Chile approximately between 45 and 35 Ma.

15. Discussion

15.1. Origins: what kind of basin?

Initiation of compressional uplift associated with east-verging, high-angle reverse faults and fault-propagation folds along the eastern edge of the Cordillera de Domeyko in the El Bordo Escarpment region triggered the accumulation of the syntectonic sediments of the Tonel Formation. These new findings seem to disprove the extensional hypothesis for the origin of the Salar de Atacama basin presented by Macellari et al. (1991), Hartley et al. (1992a) and Flint et al. (1993) and better support contraction-dominated models as suggested by Charrier and Muñoz (1994) and Muñoz et al. (1997) who considered that the Salar de Atacama formed as a foreland basin related to Cretaceous east-vergent thrusting. More recently, McQuarrie et al. (2005) suggested that during the Cretaceous "a fold-thrust belt existed in the western portion of the Central Andes as early as Cretaceous to early Paleocene time". According to these models, east-directed thrusting along the Cordillera de Domeyko should have produced a thrust-loaded foredeep in which sediments of the lower Purilactis Group may have been deposited (see also Horton et al., 2001).

The Cordillera de Domeyko is not, however, a thin-skinned thrust belt. Detailed regional mapping (Mpodozis et al., 1993a; Marinovic et al., 1995; Cornejo and Mpodozis, 1996) indicates that this tectonic domain which is a complexly deformed basement involved belt cored by Late Paleozoic volcanic and intrusives that was first uplifted during the Late Cretaceous inversion of the Jurassic–Early Cretaceous back arc basin of Northern Chile (Mpodozis and Ramos, 1990; Amilibia et al., 2000; Tomlinson et al., 2001). Since then it has been a feature of positive relief that was reactivated by Tertiary episodes of deformation (Paleocene, Eocene–Oligocene, Miocene) including the transpressional Eocene–Oligocene "Incaic" event that formed the Domeyko Fault System (Maksaev, 1990; Reutter et al., 1991; Tomlinson et al., 1994, 1997; Tomlinson and Blanco, 1997; Cornejo et al., 1997; Maksaev and Zentilli, 1999). Whereas crustal thickening and associated lithospheric flexure are the primary causes of subsidence in foreland basins adjacent to thin-skinned thrust belts, subsidence in basins adjacent to reverse-fault bounded basement blocks seems to require more than simple flexural response to thickening (Jordan, 1995). The Salar de Atacama basin has been a subsiding region not only during the Cretaceous, but also during most of the Cenozoic (see Flint et al., 1993). Even today the basin is topographically more than 1 km below the theoretical altitude to be expected if the local crustal column were in isostatic equilibrium (Yuan et al., 2002). Götze and Krause (2002) speculated that the unusual CAGH gravimetric anomaly outlines an early Paleozoic (?) mafic to ultramafic rock body at mid-crustal (10–38 km) level below the Atacama basin. Although it is possible that the mass excess of this body may help to explain the protracted history of basin subsidence during the Late Cretaceous and Tertiary we still do not know how important this feature may be.

The Cordillera de Domeyko runs N–S for more than 800 km between 21° and 28°S, including its northern extension of the Sierra de Moreno–Sierra del Medio ((Figs. 1, 2, 10)). Late Cretaceous compressional deformation has also been documented in the Sierra de Moreno (Fig. 10), although the gross geometry of the system is different than that of the Salar de Atacama segment. The Sierra de Moreno constitutes an N–S elongate basement block bounded on both sides by high-angle reverse faults. The system of west-verging, high-angle reverse faults along its western edge (Sierra de Moreno Fault system, SMFS, Tomlinson et al., 2001) constitutes the dominant structural system and carries the Paleozoic basement over deformed Jurassic–Early Cretaceous continental and marine sedimentary rocks (Ladino et al., 1997; Tomlinson et al., 2001, Fig. 10). Geochronological constraints on the age of inception of uplift and shortening have been presented for the Sierra de Moreno by Ladino et al. (1997, 1999) and Tomlinson et al. (2001) who, studying syntectonic intrusions and sedimentary rocks, demonstrated that west-verging thrusting initiated there between 109 and 83 Ma.

Syntectonic continental sediments of eastern provenance (Tambillos Formation), equivalent to the Lower Purilactis sequence, accumulated in a "foreland" position unconformably above deformed Jurassic–Lower Cretaceous strata in the SMFS

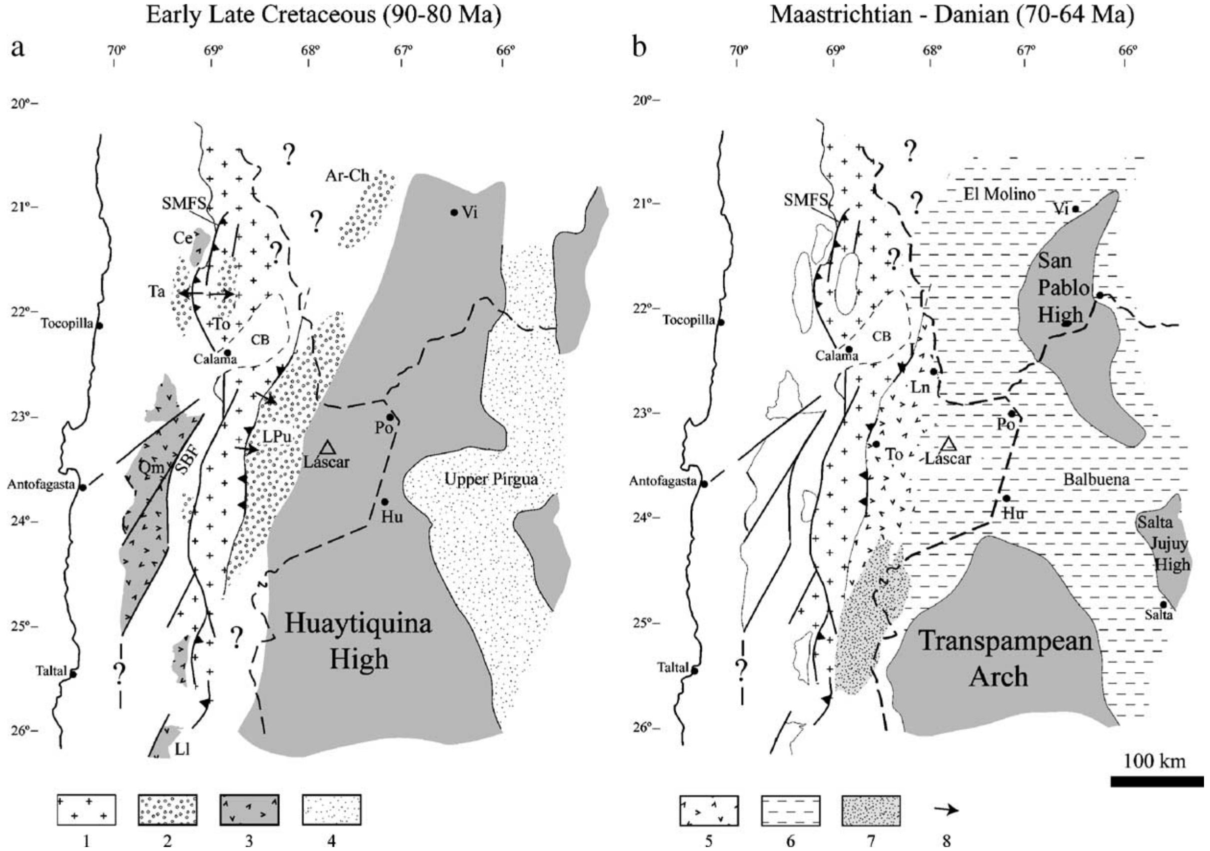

Fig. 10. Late Cretaceous–Early Tertiary paleogeography of the Western Central Andes (21–26°S). (1) Cordillera de Domeyko basement, (2) Late Cretaceous syntectonic deposits, (3) Arc volcanics and intra-arc basin sediments, (4) Continental sediments of the Salta Rift System, (5) Alkaline mafic lavas of the Totola Formation, (6) Shallow marine deposits of the El Molino–Yacoraite sea, (7) Continental sequences of the Sierra de Almeida region, (8) Sediment provenance direction, Ce: Cerro Empexa Fm., Ta: Tambillos Fm., To: Tolar Fm., Qm: Quebrada Mala Fm., Ll: Llanta Fm., LPu: Lower Purilactis Group, Ar-Ch: Airofilla-Chaunaca Formations, SMFS: Sierra de Moreno Fault System, SBF: Sierra El Buitre Fault, CB: Calama basin, Po: Poquis, Hu: Huaitiquina, Vi: Vilque well, Ln: Lomas Negras, To: Cerro Totola. Discussion in text.

footwall (Bogdanic, 1991; Tomlinson et al., 2001). At the same time, continental conglomerates and sandstones of the Tolar Formation accumulated east of the basement uplift in a separate basin (Fig. 10). Paleocurrent analysis (Bogdanic, 1991; Tomlinson et al., 2001) indicates an eastern provenance for the Tambillos clastics and a western provenance for the Tolar. A 109±4 Ma (K/Ar, biotite) age obtained for an andesitic clast from the basal conglomerates of the Tolar Formation (Quebrada Quinchamale) and another (biotite) age of 77±3 Ma for a tuff 460 m above the base (Sierra del Medio) establish the Late Cretaceous age for the sequence (Ladino et al., 1997; Tomlinson et al., 2001).

15.2. Interactions with the Late Cretaceous intra-arc basins of northern Chile

The Late Cretaceous sequences deposited in the pre-Andean basins are not the only units of this age in northern Chile. West of the Cordillera de Domeyko, volcanic and sedimentary units such as the Cerro Empexa Formation, west of the Sierra de Moreno block (Tomlinson et al., 2001), the Quebrada Mala Formation, west of the Salar de Atacama basin (Montaño, 1976; Marinovic et al., 1996; Marinovic and García, 1999; Cortés, 2000) and the Llanta Formation, further south, west of the Pedernales basin (Cornejo et al., 1993, 1997), accumulated in a north–

south series of discontinuous fault-bounded basins (Fig. 10). The Quebrada Mala Formation, which unconformably overlies Late Jurassic–Early Cretaceous sediments and early Late Cretaceous (98–94 Ma) volcanic rocks, is a very thick (3700 m) unit of conglomerates and sandstones, interbedded with calc-alkaline andesitic lavas and welded tuffs dated (K/Ar) between 86 and 66 Ma (Williams, 1992; Marinovic and García, 1999; Cortés, 2000). It was deposited in a small, rhomb-shaped basin bounded to the east by an NNE-trending, west-dipping, high-angle fault (Sierra El Buitre Fault) formed in the Late Cretaceous as a normal fault along the southeastern basin edge (Marinovic and García, 1999; Figs. 2 and 10).

Thus there exists the potentially contradictory evidence of extension-related basins along the volcanic arc and coeval shortening in the Cordillera de Domeyko. Nevertheless, the discontinuous nature of the intra-arc basins, thick volcanic and sedimentary infill, and the orientation of related faults is consistent with the basins being formed in a (right-lateral?) strike-slip setting as Arévalo et al. (1994) and Arévalo (1999) suggested for the Late Cretaceous Hornitos basin in the Copiapó region (26–27°S). Coexistence of zones of margin-parallel strike-slip faulting and margin-perpendicular shortening is not, however, rare in active continental margins where plate convergence is oblique, as has been discussed in recent tectonic models for the Late Cenozoic southern Andes (i.e. Cembrano et al., 2002; Folgera et al., 2003). Theoretical models such as that of Saint Blanquant et al. (1998) show that oblique plate convergence along active margins may produce a strain-partitioned transpressional structural array characterized by arc-parallel strike-slip faults, which may be associated with local pull-apart basins, along the thermally weakened arc. Strike-slip motion may coexist with arc-orthogonal shortening and thrusting in the back arc region (“magma facilitated strike-slip partitioning”). This model, which has been verified for the active Sumatra arc (see Mount and Suppe, 1992; Tikoff and Teysier, 1994), may be applicable to the tectonic conditions that prevailed in Northern Chile during the Late Cretaceous (Fig. 10).

15.3. Regional comparisons of basins

The Salar de Atacama basin is located at the western slope of the vast Altiplano-Puna plateau where a thick sequence of sedimentary rocks accumulated during the Cretaceous and Tertiary, both in Bolivia (Sempere et al., 1997; Horton et al., 2001) and northwestern Argentina (Allmendinger et al., 1997; Salfity and Marquillas, 1999; Coutand et al., 2001). Studies on the timing, distribution and provenance of sediments have led some authors to suggest a western source for the Cretaceous sediments outcropping in southwestern Bolivia (Sempere et al., 1997; Horton et al., 2001; McQuarrie et al., 2005). The Cordillera de Domeyko must be considered to be of the western source terrain while the Salar de Atacama basin and the string of “pre-Andean” basins to 27°S in Chile (Mpodozis and Clavero, 2002, Fig. 1) may correspond with the more proximal depocenters of the Bolivia and northwestern Argentina Basin System. However, as the connection between the northern Chile depocenters and the rest of the Altiplano-Puna basins was cut by the onset of volcanism along the Western Cordillera at 26 Ma (Coira et al., 1982; Jordan and Gardeweg, 1988; Kay et al., 1999) correlations between the two now separated domains have been difficult to establish.

15.3.1. Connections with the Salta Rift Basins and the Huaytiquina High

The Purilactis basin appears to have been a narrow trough, separated on its east by one or more basement ridges from the Salta Rift Basin. Directly east of the El Bordo Escarpment, chronologically equivalent units to the Lower Purilactis Group have not been found in the deep Toconao 1 well drilled in the center of the Salar de Atacama basin. As we mentioned earlier, this well proved a volcanic sequence correlated to the Totola Formation, with thin layers of shallow-water marine limestones bearing Late Cretaceous (“Senonian”) foraminifera (Muñoz et al., 1997; Muñoz and Townsed, 1997). The well was drilled over the northern extension of the Cordón de Lila, a basement ridge formed by Ordovician granitoids, Late Devonian–Early Carboniferous sediments and Permian igneous complexes, which protrudes as a peninsula at the southern end of the Salar de Atacama (Ramírez and Gardeweg, 1982, Fig. 2). South of the Salar, at Cerros Colorados and Quebrada Guanaqueros, the Paleozoic units are unconformably covered by 500 m of fine-grained red sandstones and shales interbedded with thin limestones layers Campanian to Maastrichtian dinosaur (*Titanosauridade*) bones (Pajonales Formation, Salinas et al., 1991). Such observations indicate that Lower Purilactis Group accumulated in a narrow, NS-trending depocenter between the Cordillera de Domeyko and a basement massif (Huaytiquina High), which separated the basin from the Cretaceous depocenters of the Salta rift System to the east (Salfity et al., 1985, see Fig. 10).

The oldest Mesozoic deposits known over the Huaytiquina High, in Chile, correspond to a section of marine sandstones, oolitic limestones and tuffs (Quebrada Blanca de Poquis Formation, Gardeweg and Ramírez, 1985) carrying Late Cretaceous foraminifera (*Hedbergella*) and lies unconformably on Ordovician strata in the Zapaleri region, near the triple boundary between Chile, Argentina and Bolivia (Fig. 10). Further south, in the vicinity of the Chile-Argentine frontier at Huaytiquina (Fig. 10), 100 m of Late Cretaceous carbonate-cemented sandstones, conglomerates and impure limestones attributed to the Yacoraite Formation (Marquillas and Salfity, 1988; Salfity and Marquillas, 1994, 1999) also overlie Ordovician sediments (Donato and Vergani, 1987; Marquillas et al., 1997). Moreover, Yacoraite limestones seem to be an important component of the basement of the Late Cenozoic CVZ volcanic arc, as carbonate skarn xenoliths with geochemical signatures akin to the Yacoraite limestones have been found in recent lavas erupted from the Lascar volcano (Matthews et al., 1996, 1997). The westernmost extension of this marine incursion is to be found in the marine sediments of the Lomas Negras Formation (Salfity et al., 1985) and the limestones interbedded with the alkaline volcanic rocks of the Totola Formation, which seem to have covered a large portion of the Purilactis basin (Fig. 10).

The latest Cretaceous marine incursion over the Huaytiquina High linked, for the first time, the Purilactis basin to the Salta Rift System of northwestern Argentina (Marquillas and Salfity, 1988; Comínguez and Ramos, 1995; Salfity and Marquillas, 1999). Volcanism seems to have terminated in the Chilean intra-arc basins near the end of the Cretaceous whereas the alkaline nature of the Totola lavas points to more extensional conditions in the Purilactis basin favoring the incursion of the Yacoraite–El Molino sea over the former basement highs (see Fig. 10). Yet, models presented for the Salta rift speculate that Yacoraite limestones accumulated during the thermal subsidence stage of the rift system (Comínguez and Ramos, 1995), a disputable hypothesis as the onset of rifting occurred more than 60 Ma before deposition of the Maastrichtian limestones. The oldest fill unit of the Salta basin is the Pirgua Subgroup (Fig. 11), which Salfity et al. (1985) considered as a counterpart of the Tonel and Purilactis Formations. This correlation is at least partly inconsistent with paleomagnetic data, which place the beginning of accumulation of the Tonel Formation in the interval 119–84 Ma, most probably in the upper part of that time slice. The oldest Pirgua unit, the coarse, proximal alluvial-fan, synrift conglomerates of the La Yesera Formation has interbedded alkaline lavas whose K/Ar ages varies from 128 to 96 Ma (Viramonte and Escayola, 1999; Viramonte et al., 1999), and seems to have formed during a mid Cretaceous extension event associated to continental-scale processes leading to the opening of the South Atlantic (Uliana and Biddle, 1988; Salfity and Marquillas, 1994; Viramonte et al., 1999). This extensional episode coincides with a generalized period of intra-arc extension in Chile (Aberg et al., 1984; Mpodozis and Allmendinger, 1993) that affected large areas of the Andean margin (Ramos and Alemán, 2000), before the compressional deformation associated with the initiation of sedimentation in the Purilactis basin.

The Upper Pirgua sequences (Los Blanquitos and Las Curtiembres Formations and the 78–76 Ma Las Conchas basalt, Salfity and Marquillas, 1999; Viramonte et al., 1999) are, probably, time-equivalents of the Lower Purilactis Group succession (Fig. 11). The abrupt facies change from the coarse alluvial fan conglomerates of the La Yesera synrift to the fine-grained lower energy facies of the Las Curtiembres and Los Blanquitos sediments which accumulated on a subtler relief due to the filling of the basin and to less active tectonics (Comínguez and Ramos, 1995) led Salfity and Marquillas (1999) to propose that at ca. 89 Ma the Salta basins were affected by a tectonic episode (cessation or slow-down of active rifting?). We consider that this change could be a far-field effect of the initiation of compression in the Cordillera de Domeyko.

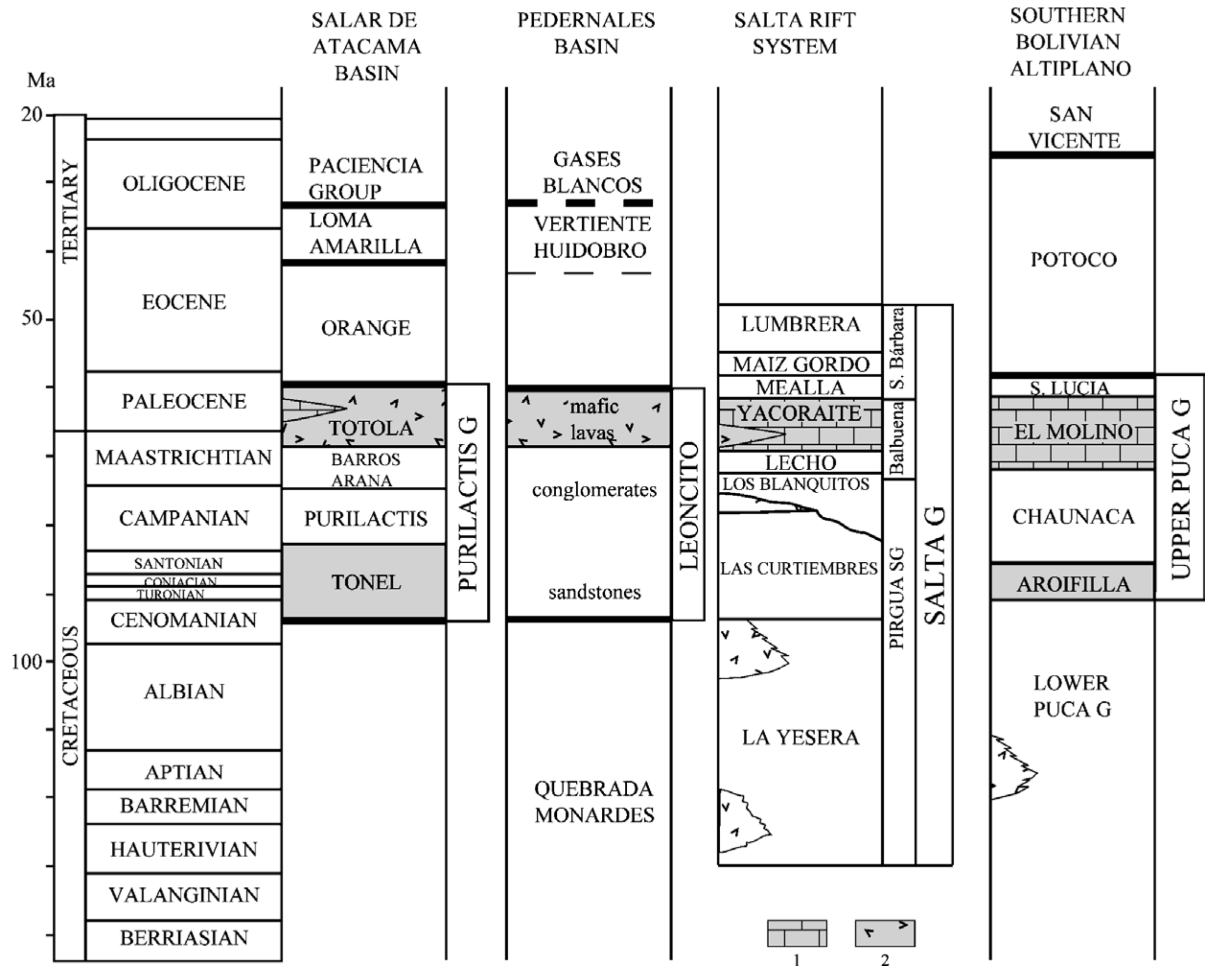

Fig. 11. Correlation table of Cretaceous–Paleogene units from the southwestern Central Andes (20–26°S). Thick black lines represent regionally important unconformities. (1) Marine carbonates, (2) volcanics.

15.4. Relationships with other Chilean pre-Andean basins and the Bolivian Altiplano

Continental Cretaceous basins that can be compared with the Salar de Atacama (Purilactis) basin also occur south of the Salar de Atacama between 24° and 27°S (Punta Negra and Pedernales-Maricunga basins, Fig. 1). These are located between the Cordillera de Domeyko and the CVZ arc to the east as is the Salar de Atacama basin. In the Punta Negra basin, the thick Tertiary volcanic and sedimentary cover precludes observation of the older parts of the sedimentary fill. In the Maricunga-Pedernales basin (Fig. 1), a thick sequence of continental clastics (Leoncito Formation) accumulated in strong angular unconformity over Early Cretaceous sandstones (Quebrada Monardes Formation) (Cornejo et al., 1993, 1998; Mpodozis and Clavero, 2002). The lower levels of the Leoncito Formation consist of more than 2000 m of fine-grained laminated sandstones (Tonel equivalent?, Fig. 11) that grade upwards to 800 m of sandstones and coarse conglomerates (proximal alluvial fan facies) with reworked Jurassic fossils, limestone clasts and Paleozoic granitoid boulders more than 1 m in diameter. The exposed top of the Formation includes 150–200 m of vesicular mafic lavas. Mafic dykes, thought to be feeders of those volcanics, have been dated (K/Ar) at 64 Ma (Cornejo et al., 1993, 1998).

The Leoncito sedimentary episode reflects the onset of uplift and unroofing of the Cordillera de Domeyko during the Late Cretaceous (Mpodozis and Clavero, 2002) and its basal unconformity above Lower Cretaceous rocks places a good constrain on the age of the beginning of sedimentation in the pre-Andean basins of northern Chile.

Northwest of the Salar de Atacama, the NNE trending active deforming zone of the El Bordo Escarpment and Cordillera de La Sal (Fig. 1) pass below the NNW trending active Andean arc to emerge in Bolivia as the Uyuni-Khenayani Fault Zone (Baby et al., 1990; Welsink et al., 1995; Elger and Onken, 2002). There, and also further to the northeast (Sevaruyo region), Cretaceous to Paleogene sediments of the Puca Group unconformably overlie Paleozoic strata (Lohman and Branisa, 1962; Cherroni, 1974; Sempere, 1995; Sempere et al., 1990, 1997; Welsink et al., 1995). The Kimmeridgian-Albian Lower Puca is a sequence of continental conglomerates and red sandstones, accumulated, in an extensional setting similar to that of the Lower Pirgua subgroup. That extensional phase culminated with a Cenomanian-Turonian marine episode (Miraflores Formation, Fig. 11). The Upper Puca Cycle started at 89 Ma with the deposition of red mudstones, fine-grained laminated sandstones, evaporites (gypsum beds) and pink-brown gypsiferous mudstones (Aroifilla Formation, latest Turonian-Coniacian) deposited, like the Tonel Formation, in a distal alluvial or desiccated lacustrine (playa-lake) environment (Sempere et al., 1997). The overlying Santonian-Campanian (86–73 Ma) Chaunaca Formation, inaugurated by a short-lived restricted marine episode, is also an association of playa-lake facies dominated by red-brown mudstones. In the Central Altiplano domain the Chaunaca mudstones possibly intermingle to the west with more proximal sandstones and mudstones of western provenance (Coroma Formation) that may be compared with the Purilactis Formation.

The Upper Puca cycle terminated with the 100 to 500 m thick Maastrichtian-Danian (73–60 Ma) El Molino Formation, whose shallow to restricted marine facies (carbonates, black shales and mudstones) alternate with tuffaceous horizons dated (^{39}Ar/^{40}Ar) at 71.6 and 72.5 Ma (Sempere et al., 1997). As in Northern Chile, the marine transgression advanced over the high ground east of the basin axis as illustrated in the Vilque-X1 well (Fig. 10) to the east of the Uyuni-Khenayani fault zone, where El Molino limestones directly overlie Silurian? strata (Welsink et al., 1995).

Sempere et al. (1997) suggested that at ~89 Ma (Late Turonian–Early Coniacian), the Early Cretaceous extensional conditions prevailing in the southern Altiplano began to change towards a more compressional setting and considered that the accumulation of the younger Upper Aroifilla and Chaunaca Formations occurred in a distal foreland basin related to compression occurring to the west, i.e. in Chile. Horton and DeCelles (1997) and McQuarrie et al. (2005) also proposed that Late Cretaceous sedimentation in the Altiplano took place in an early foreland basin related to initial Andean shortening. The more complex model presented by McQuarrie et al. (2005) envisaged that the upper Puca sedimentation occurred to the east of an elevated forebulge which separated the Bolivian basin from a western “foredeep” close to the zone of active deformation where Purilactis Group equivalents may have been deposited (see also Horton et al., 2001). However, we do not see conclusive evidence for the existence of a forebulge, as the Tonel Formation crops out directly along strike with the Upper Puca sequences (Aroifilla-Chaunaca) of the Uyuni-Khenayani Fault zone (Fig. 10). An alternative proposition may be that the Bolivian basins were during the Coniacian-Campanian connected along strike with the Purilactis depocenter and also stretched further south possibly trough the Pedernales basin. This western realm, bounded to the west by the Cordillera de Domeyko, was separated by the Huaytiquina High from the Salta Rift System. Only during the Maastrichtian when the Yacoraite–El Molino sea advanced over the basement highs did the two domains become depositionally connected (Fig. 10).

15.5. Post Purilactis Paleogene tectonics

The (D3) unconformity at the base of the Orange Formation marks the end of accumulation of the Purilactis Group. We interpret the fining-upward Orange Formation whose facies change from coarse conglomerates to sandstones and evaporites, as sedimentary strata post-dating a regional pulse of compressional deformation, which affected large expanses of northern Chile in the earliest Paleocene and produced renewed uplift in the Cordillera de Domeyko as indicated by an apatite fission track age of 64.5±0.9 Ma for the Cerro Quimal intrusive reported by Andriessen and Reutter (1994) which trace rapid cooling, probably related to exhumation and uplift shortly after pluton emplacement at around 66 to 64 Ma. West of the Cordillera de Domeyko, the Late Cretaceous Quebrada Mala basin, in the Central Depression of the Antofagasta region (Figs. 2 and 10) was inverted before the accumulation of Paleocene lavas and tuffs of the Cinchado Formation (Marinovic et al., 1996; Marinovic and García, 1999). The basal Paleocene unconformity has been also documented at the subsurface of the Salar de Atacama basin, where Seismic Unit H onlaps over a system of folds affecting strata comparable to the Totola Formation (Muñoz et al., 2002; Jordan et al., submitted for publication). The same situation has been documented further south, in the El Salvador-Copiapó region (26–28°S) were a regional unconformity separates Paleocene volcanics from Late Cretaceous strata (Cornejo et al., 1993, 1999; Arévalo et al., 1994; Iriarte et al., 1996).

The Paleocene was a time of extensive volcanism in the Central Depression both in Antofagasta and Copiapó-El Salvador (Arévalo et al., 1994; Cornejo and Mpodozis, 1996; Marinovic et al., 1996). Geochemical data indicate that the continental Andean crust in northern Chile was thickened as a result of the early Paleocene deformation episode as Paleocene–Eocene magmas evolved at higher pressures than during the Late Cretaceous in a crust already able to stabilize hornblende (Cornejo et al., 1999; Cornejo and Matthews, 2001).

The early Paleocene deformation seems to be have affected, primarily, the western Andean domain. To the east of the Salar de Atacama basin, in the Zapaleri region (Fig. 2) continental sandstones of the Chofjias Formation (Gardeweg and Ramírez, 1985), which are possible distal equivalents of the Orange Formation conformably overlie the Upper Cretaceous marine strata. Reduced sediment thickness and the presence of numerous paleosol horizons characterize the Paleocene in the southern Bolivian Altiplano (Santa Lucía, lowermost Potoco and Impora Formations, Sempere et al., 1997), suggesting a period of low sedimentation rates. This is also recorded in the thin Paleocene lacustrine sequences of the Santa Barbara Subgroup (Mealla and Maíz Gordo Formations), which concordantly overlie the Yacoraite limestones in the western Puna and the Salta basins (Salfity and Marquillas, 1999).

During the Eocene–Early Oligocene gravels were shed eastward from the Cordillera de Domeyko, into the Salar basin, forming the synorogenic Loma Amarilla conglomerates. The Loma Amarilla synorogenic sediments are associated with the Eocene "Incaic" deformation (Coira et al., 1982), when the whole Cordillera de Domeyko was affected by left-lateral transpression which led to localized thrusting, strike-slip faulting, clockwise block rotations, major uplift and porphyry copper emplacement (Maksaev, 1990; Reutter et al., 1991; Cornejo et al., 1997; Mpodozis et al., 1993a,b; Arriagada et al., 2000). Very rapid uplift and important erosion was suggested by Alpers and Brimhall (1988) to explain the relation between porphyry copper emplacement and supergene enrichment at La Escondida mine. Maksaev and Zentilli (1999), modeling fission track (apatite) data, indicate that at least 4 to 5 km of rocks were exhumed from the Cordillera de Domeyko at rates as high as 100 to 200 m/my during the "Incaic" deformation episode between 50 and 30 Ma.

The onset of the Incaic deformation near 45 Ma occurred at the beginning of a period of very high rates of oblique convergence between the Farallon and South American plates (Pilger, 1984; Pardo-Casas and Molnar, 1987). At the same time there was also an eastward migration of the magmatic front from the Central Depression to the Cordillera de Domeyko axis. However, there was very little Eocene volcanism and magmatism is represented only by discontinuous clusters of small-volume, shallow-level stocks, including copper-rich porphyry stocks emplaced between 42 and 37 Ma along the traces of the Domeyko Fault System (Maksaev, 1990; Sillitoe, 1992; Cornejo et al., 1997, 1999). Extreme transpressional conditions prevailed during this period. These conditions seems to have prevented easy magma ascent, favoring the ponding of magmas at the base of the crust which led, in turn, to ductile transpressional failure of the lower crust and localized tectonic thickening below the Cordillera de Domeyko (Mpodozis et al., 1993a,b; Yañez et al., 1994; Tomlinson et al., 1994; Arriagada et al., 2000).

Steep REE patterns of Eocene–Early Oligocene intrusives (Maksaev, 1990; Cornejo et al., 1999; Richards et al., 2001) indicate the progressive stabilization of garnet in the source region of the Eocene magmas, which is consistent with the formation of a significant crustal root below the Cordillera de Domeyko at that time (Cornejo et al., 1999). Rapid isostatic rebound triggered by the creation of the crustal root may explain the rapid uplift recorded in the Cordillera de Domeyko during Eocene–Early Oligocene times. The regional distribution of the Loma Amarilla Formation indicates that sedimentation was controlled by the formation of a regional uplifted area centered along the topographic axis of the Cordillera de Domeyko (Maksaev and Zentilli, 1999) from which sediments were shed to the west and east to almost completely conceal the previous structural landscape along the El Bordo Escarpment (Mpodozis et al., 2000; Arriagada et al., 2002; Figs. 4 and 5).

Thick (up to 6000 m) Late Eocene to Oligocene continental sequences also cover large tracts of the southern Altiplano in Bolivia (i.e. Potoco and Tiahuanacu Formations, Sempere et al., 1997; Lamb and Hoke, 1997; Horton et al., 2001; Rochat, 2002) and the Puna, in Argentina (Geste and Quiñoas Formations, Jordan and Alonso, 1987; Allmendinger et al., 1997; Kraemer et al., 1999; Coutand et al., 2001; Voss, 2003). Provenance studies and the discontinuities existing between depocenters indicate that sedimentation was associated not only with uplift in the western Andes but also with the uplift of the Eastern Cordillera block in Bolivia (Lamb and Hoke, 1997) and discrete blocks in the Argentine Puna (Boll and Hernández, 1986; Kraemer et al., 1999; Coutand et al., 2001). The distributed nature of the deformation, which affected a very wide swath of the central Andes, including the westernmost Sierras Pampeanas at 27°S (Coughlin et al., 1998), may be an indication that the Eocene (Incaic) orogeny was associated with a shallowing of the subducting slab as has been suggested for southern Peru (Sandeman et al., 1995; James and Sacks, 1999).

16. Conclusions

The Salar de Atacama basin is a deep depocenter, which exhibits a well-preserved Mesozoic to Cenozoic sedimentary record of the earliest tectonics events, which led to the formation of the Central Andes. New field and geochronological data support the presentation of an updated stratigraphic division for the early basin fill (Purilactis Group). This scheme better reflects the geodynamics and allows the development of basin history. Unlike previous hypotheses we consider that the basin was formed not as an extensional basin, nor a classic foreland basin produced by lithospheric flexure. Instead it was created in the early Late Cretaceous as a consequence of inversion of the Mesozoic back arc basin of northern Chile. Inversion led to the formation of the proto-Cordillera de Domeyko as a narrow, thick-skinned, basement range bounded by high-angle reverse faults and blind reverse faults, which shift vergence along strike. Syntectonic sediments were deposited in basins both to the west (Tambillos Formation) and east of the Cordillera de Domeyko (Purilactis Group in the Salar de Atacama Basin).

The extremely thick Cretaceous to Paleogene fill of the Salar de Atacama (Purilactis) basin seems to be, in some sort, related to the presence, below the Salar de Atacama basin of inherited (Paleozoic) anomalously dense crustal body, which originated the CAGH anomaly. Repeated episodes of uplift related to moderate amounts of shortening along the eastern border the Cordillera de Domeyko led to the accumulation of syntectonic sediments of the Tonel, Purilactis and Barros Arana Formations during the Late Cretaceous. During the same time interval, but closer to the continental margin, in the central part of the Antofagasta region, volcanism and sedimentation seems to have occurred in basins controlled by strike-slip faults along thermally weakened active Andean arc. The prevalent tectonic regime during the Late Cretaceous in Northern Chile seems to have been controlled by oblique plate convergence which led to the partitioning of regional strain into a margin-parallel component, along the arc, and a margin-perpendicular component in the back arc. Back arc compression favored the inversion of the former Jurassic–Early Cretaceous back arc basin and the accumulation of syntectonic sediments in the Salar de Atacama basin.

The Salar de Atacama basin is the westernmost Cretaceous–Paleogene depocenter of the Altiplano Puna domain and events recorded in its sedimentary fill are consistent with the reported history of more distal areas of northwestern Argentina and south-eastern Bolivia. Nevertheless, comparison with the Salta Rift System of northwestern Argentina shows that the Salta Rift System was formed during Early Cretaceous extension, also recorded in Bolivia (Lower Puca) and along the Chilean active margin. It pre-dates the Purilactis basin. Far-field effects of the Early Late Cretaceous inversion of the northern Chile back arc basin and initial uplift of the Cordillera de Domeyko have been recognized both in the southern Bolivian Altiplano and the Salta Rift System.

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