Multiproxy paleosol evidence for a rain shadow effect linked to Miocene uplift of the North Patagonian Andes

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ABSTRACT

Climate changes related to orographic barrier uplift have been in the research spotlight over recent years. Several works have focused on the interactions between climatic and tectonic processes in order to understand the development of a rain shadow. Patagonia is an ideal region in which to analyze such interactions, since a Miocene climate change, from wetter to drier conditions, has mainly been associated with Andean uplift. In this work, we analyzed a succession of stacked paleosols recorded in a Miocene North Patagonian foreland basin in order to understand how the paleosol moisture regime related to the atmospheric humidity changes caused by the uplift of the Patagonian Andes. Based on macromorphological, micromorphological, and geochemical studies, and supported by a high-resolution chronostratigraphic scheme based on U–Pb geochronology and magnetostratigraphy, the paleosols were characterized with corresponding mean annual paleoprecipitation (MAP) and mean annual temperature values. Alfisol-like paleosols were identified at the base of the foreland infill (15–14.6 Ma) with a MAP of 1229 ± 108 mm/yr. The Andisol-like paleosols recognized in the middle section of the sequence (14.6–12.75 Ma) exhibited a MAP of 1053 ± 108 mm/yr, whereas the Aridisol-like paleosols occurring in the upper section of the infill (12.75–11.5 Ma) presented a MAP of 677 ± 108 mm/yr. The determined Miocene mean annual temperatures ($\sim 11 \pm 2.1$ °C)

were similar to the present-day values (11 °C). Based on the complete tectonic record of the Patagonian Andes, the observed decrease in MAP was assigned to the rain shadow effect created by the uplift of the North Patagonian Andes. Results indicate that although the process started around 19 Ma, the rain shadow effect was not effectively recorded before ca. 14.6 Ma.

INTRODUCTION

The development of orographic barriers causes significant climatic and ecological changes, not only on the windward side, where the regional winds prevail, but also on the leeward side. The uplift of a mountain range may constitute a topographic barrier to atmospheric circulation, thus producing the generation of an orographic rain shadow (Masek et al., 1994; Vandervoort et al., 1995; Stern et al., 1997; Harrison et al., 1998; Kleinert and Strecker, 2001; Roe, 2005; Hughes et al., 2009; Rohrmann et al., 2016; Meijers et al., 2018; Hoorn et al., 2018). This rain shadow effect is caused when air, forced to rise over a mountain range, is cooled adiabatically, generating clouds, condensation, and precipitation along the windward side of the orographic front. On the leeward side of the orographic barrier, the winds descend, and a decrease in relative moisture occurs, producing less precipitation (Roe, 2005; Hughes et al., 2009; Hoorn et al., 2018). The interplay between tectonic and climatic processes has been analyzed in several regions of the world, in particular, concerning the Himalayan and Andean range uplifts (e.g., Masek et al., 1994; Vandervoort et al., 1995; Stern et al., 1997; Harrison et al., 1998; Kleinert and Strecker, 2001; Strecker et al., 2007; Poulsen

et al., 2010; Mulch et al., 2010; Rohrmann et al., 2016). Although these coupled processes have constituted the main subject of several lines of research in recent years, few contributions have analyzed the temporal relationship between climate changes and an orographic barrier uplift in quantitative terms, in particular, addressing the relationship between the onset of the orographic barrier uplift and the rain shadow effect (Kleinert and Strecker, 2001; Roe, 2005; Blisniuk et al., 2005; Ortiz-Jaureguizar and Cladera, 2006; Hughes et al., 2009; Palazzesi and Barreda, 2012; Le Roux, 2012; Rohrmann et al., 2016; Raigemborn et al., 2018; Meijers et al., 2018; Hoorn et al., 2018; Pingel et al., 2019).

For the Andean range, it has been reported that one of the main phases of orogenic growth occurred during the Miocene (Ramos et al., 2002; Dávila and Astini, 2003; Strecker et al., 2007, 2012; Poulsen et al., 2010; Mulch et al., 2010; Ramos and Folguera, 2009; Hain et al., 2011; Bilmes et al., 2013; Del Papa et al., 2013; Rohrmann et al., 2016; Ciccioioli et al., 2018). Such works state that several Andean segments were then configured and evolved with dissimilar tectonic, volcanic, and geographic features. Some of these segments have been widely analyzed, and tectonically controlled climate changes have been established. In fact, several authors indicate that the orographic uplift of the Central Andes, in late Miocene times, triggered meaningful climate changes, from wetter to drier conditions (Kleinert and Strecker, 2001; Hartley, 2003; Strecker et al., 2007, 2012; Poulsen et al., 2010; Mulch et al., 2010; Rohrmann et al., 2016). Nevertheless, the interactions between tectonics and climate still remain less understood for other Andean segments. One of them is the Patagonian Andes (Fig. 1), which currently

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Figure 1. (A) Main tectonic structures and Neogene depocenters of the Patagonian broken foreland (masl—meters above sea level). The Paso del Sapo Basin is located within the Andean retroarc, adjacent to the North Patagonian Andes, modified from Bilmes et al. (2013). (B) Geological map of Paso del Sapo Basin (see location in Fig. 1A), indicating main inverted normal faults, and Neogene units that constitute the sedimentary infill.

constitutes a worldwide topographic barrier to atmospheric circulation, causing one of the most drastic orographic rain shadows on Earth (Blisniuk et al., 2005). This mountain range uplift occurred mainly during early to middle Miocene times (19–10 Ma; Burdigalian to Tortonian; Folguera and Ramos, 2011; Orts et al., 2012; Bilmes et al., 2013; Gianni et al., 2015; Echaurren et al., 2016; Folguera et al., 2018; Franzese et al., 2018; Bucher et al., 2019a, 2019b; López et al., 2019). Partially synchronous with mountain uplift, significant climate and ecological changes are recorded in several regions of the Patagonian Andean foreland (Blisniuk et al., 2005; Ortiz-Jaureguizar and Cladera, 2006; Palazzesi and Barreda, 2012; Le Roux, 2012; Raigemborn et al., 2018). Although these studies, based on faunal turnover, suggest evidence of climate changes during the Miocene, no quantitative analysis of such processes in the region has established a relationship between the onset of the orographic barrier uplift and the rain shadow effect. Knowing that multiproxy analysis of paleosols is a powerful tool for establishing paleoclimatic conditions in continental basins (e.g., Retallack, 1985, 1988, 1993, 1994; Atkinson, 1986; Zaleha, 1997; Kleinert and Strecker, 2001;

Hoorn et al., 2018), in this work, we applied this technique to a paleosol succession recorded in a Miocene Andean retroarc foreland basin: the Paso del Sapo Basin, which has a continuous geological record between 15 and 11.5 Ma (Bucher et al., 2018, 2019b). The aim of this study was to report new quantitative data on the paleoclimate of northern Patagonia in order to elucidate the relationship between the processes associated with the history of Andean uplift in northern Patagonia and the rain shadow effect.

GEOLOGICAL BACKGROUND

The Patagonian Andes developed as a response to the subduction of different oceanic plates beneath the South American plate in the last ~100 m.y. (e.g., Folguera and Ramos, 2011). As a result, contractional deformation was not continuous over time, and two main phases can be recognized. The first contractional phase was constrained mainly to Cretaceous times, whereas the second one is assigned to the Neogene, mainly during Miocene times (Folguera and Ramos, 2011; García Morabito et al., 2011; Orts et al., 2012; Bilmes et al., 2013; Gianni et al., 2015; Folguera et al., 2015, 2018; Echaurren et al.,

2016; Gianni et al., 2017; Bucher et al., 2018, 2019b). Both contractional phases are recorded in the Main Cordillera of the Patagonian Andes, as well as in the foreland region, revealing that the anatomy of the Andean uplift and Andean foreland zones is that of a broken foreland system, known as the Patagonian broken foreland (Bilmes et al., 2013). This broken foreland system is located between 40°S and 48°S; it is ~1000 km in length, extending in width 300 km toward the east (Fig. 1). Within the Patagonian broken foreland, there are several intermontane basins floored by basement, which are isolated or partially connected to each other (i.e., Catán Lil Basin, Piedra del Águila Basin, Collón Cura Basin, Gastre Basin, Ñirihuau Basin, and Paso del Sapo Basin; Fig. 1A). The Paso del Sapo Basin (Bucher et al., 2018, 2019a) is located in the Andean foreland region of the Patagonian broken foreland, between 42°25'S and 42°55'S and 69°21'W and 69°46'W (Fig. 1; Bilmes et al., 2013; Bucher et al., 2018, 2019a). It is defined as a small, intermontane basin with an ~300-m-thick infill, assigned to the Miocene contractional phase by structural, geochronologic, and magnetostratigraphic analyses (Bucher et al., 2018, 2019a, 2019b). The tectonic activity that led to

the formation of the Paso del Sapo Basin has been constrained from 15 to 11.5 Ma, showing a tectonic climax at the beginning of its evolution (15–14 Ma), and a deceleration in the vertical growth of tectonic structures at 12.75 Ma (Bucher et al., 2019a, 2019b).

The Neogene infill of the Paso del Sapo Basin is grouped into two synorogenic continental units (Fig. 1B; Bucher et al., 2018): the La Pava Formation (Nullo, 1978) and the Collón Cura Formation (Yrigoyen, 1969). Both units show evidence of syntectonic sedimentation and can be related to the Miocene contractional event that gave rise to the Paso del Sapo Basin and the Patagonian Andes (Bucher, 2018; Bucher et al., 2018, 2019a). The La Pava Formation is characterized by fine- to coarse-grained volcanoclastic deposits accumulated in alluvial-fluvial environments with well-developed paleosols (Nullo, 1978; Bilmes et al., 2013; Bellosi et al., 2014; Bucher et al., 2018). The Collón Cura Formation is characterized by volcanoclastic siltstones and sandstones, and it has less mature paleosols than the La Pava Formation. The Collón Cura Formation has been differentiated into a lower and an upper section based on clearly recognizable paleoenvironmental variations. The lower section accumulated in a lacustrine-deltaic system, whereas the upper section was deposited in alluvial-fluvial environments (Ramos et al., 2011; Bilmes et al., 2014; Bellosi et al., 2014; Bucher et al., 2018, 2019a). Recently, an accurate chronostratigraphic scheme was defined (Fig. 2; Bucher et al., 2019b). The age of this synorogenic Miocene succession was constrained based on magnetostratigraphic data calibrated with two U-Pb geochronological ages, resulting in an age of 15 Ma to 11.5 Ma. In this scheme, the La Pava Formation was constrained to 15–14.6 Ma (Langhian), whereas the lower and the upper sections of the Collón Cura Formation were constrained to 14.6–12.75 Ma (Langhian–early Serravallian) and 12.75–11.5 Ma (Serravallian–early Tortonian), respectively (Fig. 2; Bucher et al., 2019b).

METHODS

The analysis of paleosols in the field was carried out by means of the recognition and description of paleosol horizons and their macroscopic pedofeatures (e.g., thickness, nodules, cutans, rhizoliths, and slickensides) and bioturbation characteristics (Soil Survey Staff, 1975, 1998; Retallack, 2001). In order to rule out modern alterations and contamination, the soils were analyzed at a minimum depth of 30 cm. In order to characterize the type and intensity of the main pedogenic processes, a detailed micro-morphological analysis on 30 thin sections was

Figure 2. Geochronological time scheme for the Paso del Sapo Basin. From left to right: Geomagnetic polarity time scale (Gradstein et al., 2012), interpreted polarity, sedimentary column, and absolute ages for La Pava Formation (15–14.6 Ma), lower section of Collón Cura Formation (14.6–12.75 Ma), and upper section of Collón Cura Formation (12.75–11.5 Ma). Green stars indicate paleosols levels analyzed in this work. Yellow stars show U/Pb radiometric ages. Figure taken from Bucher et al. (2019b). M—mudstones; S—sandstones; C—conglomerates.

conducted to determine mean grain size, mineralogical composition, porosity, microstructure, microfibrils, groundmass, and pedofeatures (Bullock et al., 1985; Stoops, 2003).

A geochemical major-element analysis was performed on 23 B-horizons samples (Hyland and Sheldon, 2016; Supplementary Material¹). Sampled levels were chosen by discarding samples with diagenetic and modern modification. The geochemical studies involved whole-

rock analyses and were conducted at the ALS Patagonia laboratory using the ME-XRF26 method. The major-oxide results were processed to obtain weight percent to molar percent. The data set obtained was used to discriminate different geochemical indexes and paleoclimatic proxies. Paleosols indexes were calculated as follows: (1) for weathering processes, the chemical index of alteration minus potassium (CIA-K; Eq. 1) and the CALMAG index (Eq. 2; Maynard, 1992; Nordt and Dreise, 2010); (2) for clayey-ness, the Al/Si ratio (Sheldon and Tabor, 2009); (3) for salinization, the SAL index (Retallack, 2001); (4) for oxidation, the total iron/alumina ratio (Retallack, 2001); (5) for hydrolysis of

¹GSA Data Repository item 2020036, Tables DR1, DR2, and color versions of all figures, is available at <http://www.geosociety.org/datarepository/2020> or by request to editing@geosociety.org.

alkaline earth elements and loss of exchangeable bases, the $\Sigma\text{Bases}/\text{Al}$ ratio (Retallack, 2001); and (6) for changes in paleosol provenance, the Ti/Al ratio (Sheldon, 2006).

$$\text{CIA} - K = 100 * \left[\frac{\text{Al}}{\text{Al} + \text{Ca} + \text{Na}} \right]. \quad (1)$$

$$\text{CALMAG} = 100 * \left[\frac{\text{Al}}{\text{Al} + \text{Ca} + \text{Mg}} \right]. \quad (2)$$

$$\text{SAL} = \frac{\text{Na} + \text{K}}{\text{Al}}. \quad (3)$$

To calculate mean annual paleoprecipitation (MAP) values, three different proxies were used: (1) the $\Sigma\text{Bases}/\text{Al}$ ratio (Eq. 4; Sheldon et al., 2002); (2) the CIA-K index (Eq. 5; Maynard, 1992; Sheldon et al., 2002); and (3) the CALMAG index, developed specifically for Vertisols or soils with vertic pedofeatures (Eq. 6; Nordt and Dreise, 2010).

$$\text{MAP} = -259.3 \times \ln(\Sigma\text{Bases} / \text{Al}) + 759;$$

with standard error of ± 235 mm / yr,

where $\Sigma\text{Bases} = \text{Ca} + \text{Mg} + \text{Na} + \text{K}$.

$$\text{MAP} = 221.12^{0.0197 \times (\text{CIA} - K)},$$

with standard error of $\pm 181 \frac{\text{mm}}{\text{yr}}$. (5)

$$\text{MAP} = 22.69 \times \text{CALMAG} - 435.8;$$

with standard error of ± 108 mm / yr. (6)

In order to calculate mean annual temperature (MAT) values, two different proxies were used: (1) the SAL index (Eq. 7; Sheldon et al., 2002; Gallagher and Sheldon, 2013); and (2) the PWI index (Eq. 8; Gallagher and Sheldon, 2013).

$$\text{MAT} = -18.516 \times \text{SAL} + 17.298;$$

with standard error of ± 4.4 °C (7)

$$\text{MAT} = 2.74 \ln(\text{PWI}) + 21.39;$$

with standard error of 2.1 °C

where

$$\text{PWI} = 100 \times \left[\frac{(4.2 \times \text{Na}) + (1.66 \times \text{Mg})}{(5.54 \times \text{K}) + (2.05 \times \text{Ca})} \right]. \quad (8)$$

RESULTS

The macromorphological and micromorphological analyses allowed us to recognize

different kinds of pedotypes. Based on pedotype abundance, three paleosol successions were recognized. The La Pava Formation (15.0–14.6 Ma) is composed of pedotype I paleosols, developed on alluvial to fluvial plains (Fig. 2). The lower Collón Cura Formation (14.6–12.75 Ma) is characterized by paleosols of pedotype II, related to delta plain environments (Fig. 2). Finally, the upper Collón Cura Formation (12.75–11.5 Ma) comprises a succession of pedotype III paleosols, associated with alluvial to fluvial plains (Fig. 2).

Paleosols

Succession of Pedotype I (La Pava Formation)

Description. Pedotype I is ~2 m thick and is composed of well-developed pale-brown to very pale-brown horizons (Figs. 3A, 3B, and 3C). The typical profile is constituted by A, Bt (B horizon with illuvial accumulation of clay), B/C, and, in some cases, Bg horizons (B horizon with hydromorphic features). The most typical pedofeatures (Figs. 3A, 3B, and 3C) are well-structured Bt horizons with angular to subangular peds and abundant cutans, Fe-Mn nodules, rhizoliths, and invertebrate trace fossils. The coarse fraction of these paleosols is composed of silt to medium-sand particles of quartz, feldspars, and acidic volcanoclastic components. Volcanic-derived clasts show a moderate degree of alteration and include

Figure 3. Pedotype I (Alfisol-like) features. (A) Typical profile of the Alfisol-like paleosols including horizonation, ped structure, macromorphological and micromorphological pedofeatures, and trace fossils. (B) Field picture of Alfisol-like paleosols showing the stacked pattern of the paleosol succession. (C) Field picture indicating the horizonation of one Alfisol-like paleosols, and detail of *Coprinisphaera* isp. trace fossil. (D) A horizon showing bioturbation channels. (E) Bt horizon with development of dense complete multilaminated clay infill and clay coatings. (F) Bg horizon showing Fe-Mn nodules. (G) B horizon showing cross-striated b-fabric characterized by two sets of birefringent clay orientated at an angle of ~80° (black arrows).

volcanic ash and pumiceous lapilli. The A horizons exhibit different kinds of voids, such as channels and chambers (Figs. 3A and 3D). In the A horizons, rhizoliths and invertebrate trace fossils (*Coprinisphaera* isp. and *Celliforma* isp.) are abundant. The A horizons are 10–20 cm thick and well structured, with laminar to granular peds; the Bt horizons have a thickness of 20–80 cm. The diagnostic pedofeatures are the presence of abundant laminated to microlaminated clay coatings in void walls and the presence of dense, yellowish brown, complete to incomplete illuvial dusty clay infills (Figs. 3A and 3E). The Bt horizons are well structured and show grain-striated and cross-striated b-fabric characterized by two sets of birefringent clay oriented at an angle of $\sim 80^\circ$ (Fig. 3G). The Bg horizons can be identified by abundant brownish Fe-Mn nodules and by the scarce presence of clay coatings (Fig. 3F). B/C horizons show poorly developed pedogenic features (e.g., scarce clay coatings) and do not preserve the sedimentary structure. As regards the geochemistry of the paleosols of pedotype I (see Supplementary Material [footnote 1]), average values of different indexes were calculated: The CALMAG weathering index is 71.34, the CIA-K weathering index is 68.75, the $Al/\Sigma Bases$ ratio is 1.47 on average, the Al/Si ratio is 0.13, the total iron/alumina (Fe/Al) ratio is 0.20, the Ti/Al ratio is 0.06, and the $(K + Na)/Al$ ratio is 0.32 (see Supplementary Material [footnote 1]).

Interpretation. The macromorphological and micromorphological analyses, together with the geochemical signature, suggest that the main pedogenic processes that occurred in the paleosols of pedotype I were clay illuviation (argilluviation), bioturbation, hydromorphism, and vertization (Bullock et al., 1985; Stoops, 2003; Tabor and Myers, 2015). The presence of thick laminated clay coatings covering void walls and dense, complete to incomplete, clay infills in Bt horizons provides evidence of strong clay illuviation (Kühn et al., 2010; Varela et al., 2016, 2018). These results, as well as the well-defined banding of the clay coatings, suggest climate seasonality (Retallack, 2001; Varela et al., 2018). In turn, the high Al/Si ratio indicates clay enrichment in B horizons by illuviation. Bioturbation is evidenced by rhizoliths, channels and chambers, and *Coprinisphaera* isp. and *Celliforma* isp. trace fossils (Stolt and Lindbo, 2010; Sarzetti et al., 2017). Fe-Mn nodules are indicative of hydromorphic processes related to poor drainage conditions, and the low Fe/Al ratio (<0.25) reflects their intensity (Lindbo et al., 2010). Vertization is inferred because of the presence of grain-striated and cross-striated b-fabric (Kovda and Mermut, 2010). This pedotype shows high CALMAG and CIA-K weathering index values, suggesting moderate to strong weathering processes (Sheldon and Tabor, 2009). The $\Sigma Bases/Al$ ratio (~ 1.50) indicates a moderate loss of

exchangeable bases. The mean values determined for the Ba/Sr ratio (1.63) indicate leaching that peaked in these paleosols (Sheldon and Tabor, 2009). The Ti/Al ratio (0.06) is consistent with the volcanoclastic to epiclastic parental sediments (Sheldon and Tabor, 2009), and the low values of the $(K + Na)/Al$ ratio suggest minimum salinization. Based on the micromorphological and macromorphological description, in addition to the geochemical index, the paleosol succession of pedotype I corresponds to modern Alfisols (Soil Survey Staff, 1975, 1998; Retallack, 2001), in agreement with previous works (Belloso et al., 2014; Sarzetti et al., 2017).

Succession of Pedotype II (Lower Collón Cura Formation)

Description. The succession of pedotype II is represented by paleosols with a thickness of 0.4–1 m and pale-yellow to pale-brown color. It is characterized by poorly developed paleosols, and, in general, primary lamination is preserved (Figs. 4A and 4B). Pedotype II shows a typical profile of A-B-B/C horizons. Bt (B horizon with illuvial accumulation of clay) and Btss (B horizon with illuvial accumulation of clay and slickensides) horizons are present in some cases. When Bt and Btss horizons appear, the most relevant morphological features are well-structured B horizons, composed of subangular peds, and the presence of clay coatings and slickensides

Figure 4. Pedotype II (Andisols-like) features. (A) Typical profile of the Andisol-like paleosols including horizonation, ped structure, macromorphological and micromorphological pedofeatures, and trace fossils. (B) Field picture indicating the horizonation of one Andisol-like paleosol. (C) Detail of slickensides developed in B horizons. (D) A horizon with typical bioturbation channels. (E) Bt horizon showing clay coatings. (F) Detail of clay coatings developed in Bt horizon. (G) B horizon showing cross-striated b-fabric characterized by two sets of birefringent clay orientated at an angle of $\sim 80^\circ$ (black arrows).

in Btss horizons (Figs. 4A, 4B, and 4C). Trace fossils (*Coprinisphaera* isp. and *Celliforma* isp.) are common in this pedotype. The coarse fraction of these soils is mainly composed of silt particles with isolated fine-sand grains. The silt fraction is dominated by volcanic ash and volcanoclastic crystals of quartz and feldspars, whereas the sandy fraction consists of pumiceous clasts, quartz, and feldspars. A horizons (Figs. 4A and 4C) are 5–15 cm thick and are well structured, with granular to subangular peds. The frequency of occurrence of rhizoliths and different kinds of voids, such as channels and chambers, is high. B horizons show a thickness of 20–40 cm. The presence of laminated to microlaminated clay coatings and dense, complete to incomplete, illuvial dusty clay infills determines the occurrence of Bt horizons (Figs. 4A, 4E, and 4F). B horizons may also contain brownish to blackish Fe-Mn nodules. Grain-striated and cross-striated b-fabrics, characterized by two sets of birefringent clays oriented at an angle of $\sim 80^\circ$, are also present (Fig. 4G); scarce pedogenic carbonates can be recognized. B/C horizons exhibit thicknesses of 20–50 cm and are characterized by the presence of scarce morphological features, such as clay coatings, subangular peds, and slickensides, and by the preservation of depositional structures. The geochemistry of the paleosols of pedotype II (see Supplementary Material [footnote 1]) shows average values of 65.49

for the CALMAG weathering index, 65.21 for the CIA-K weathering index, 1.10 for the Al/ Σ Bases ratio, 0.15 for the Al/Si ratio, 0.20 for the total iron/alumina (Fe/Al) ratio, 0.05 for the Ti/Al ratio, and 0.40 for the (K + Na)/Al index (Supplementary Material [footnote 1]).

Interpretation. Along with the geochemical signature, the macromorphological and micromorphological analyses indicate that bioturbation, argilluviation, vertization, hydromorphism, and calcification were the main pedogenic processes in these paleosols (Bullock et al., 1985; Stoops, 2003; Tabor and Myers, 2015). Bioturbation, in particular, is one of the most significant processes in pedotype II and can be recognized by the presence of rhizoliths (Stolt and Lindbo, 2010) and trace fossils (*Coprinisphaera* isp. and *Celliforma* isp.; Sarzetti et al., 2017). Argilluviation (Kühn et al., 2010) is evidenced by clay coatings and microlaminated infills, and by the high values of the Al/Si ratio. Slickensides and cross-striated b-fabric indicate vertization processes (Kovda and Mermut, 2010; Imbellone et al., 2009). Fe-Mn nodules suggest hydromorphism in these soils, and the low Fe/Al ratio (<0.25) records the intensity of this process. Calcification can be recognized by the presence of pedogenic calcium carbonates (Lindbo et al., 2010). The CALMAG and CIA-K weathering indexes show a moderate to low degree

of weathering, with mean values of 65.49 and 65.21, respectively (Sheldon and Tabor, 2009). The Al/ Σ Bases ratio (1.10) indicates a moderate loss of exchangeable bases, whereas the value of the Ba/Sr ratio (1.35) suggests moderate leaching (Sheldon and Tabor, 2009). As regards the Ti/Al ratio (0.05), it is consistent with parental sediments of a volcanoclastic to epiclastic nature (Sheldon and Tabor, 2009), while the high values of the (K + Na)/Al ratio are evidence of salinization. The occurrence of poorly developed paleosol profiles and the preservation of the primary structure, in addition to the pedofeatures described above, reflect that the paleosols of pedotype II are comparable with modern Andisols, with variable argillic and vertic features (Soil Survey Staff, 1975, 1998; Retallack, 2001), as stated in previous works (Bellosi et al., 2014; Sarzetti et al., 2017).

Succession of Pedotype III (Upper Collón Cura Formation)

Description. The paleosols of pedotype III (Fig. 5) are 0.5–2 m thick and characterized by pale-brown to whitish colors. A typical horizon profile is A, Btk (B horizon with illuvial accumulation of clay and calcium carbonates), Bk (B horizon with accumulation of calcium carbonates), and B/C (Figs. 5A, 5B, and 5C). The coarse fraction is mainly composed of silt to

Figure 5. Pedotype III (Aridisols-like) features. (A) Typical profile of the Aridisols-like paleosols including horization, ped structure, macromorphological and micromorphological pedofeatures, and trace fossils. **(B)** Field picture of Aridisols-like paleosols showing the stacked pattern and the horization of the paleosol succession. **(C)** Field picture of typical Btk horizon showing the abundance of calcic carbonate, and detail of calcic rhizoliths. **(D)** Channel and chamber with clay-coated fragments (Aif), and calcic carbonates in the groundmass (Cp). **(E)** Btk horizon with illuvial clay coatings (Ai), illuvial carbonate coatings (Ci), and calcic carbonate in the groundmass (Cp). **(F)** Bk horizon showing the presence of calcitic crystallitic b-fabric determined by micritic calcite. **(G)** Bk horizon with calcic carbonate in the groundmass and filling voids.

medium-sand particles of quartz, feldspars, volcanic ash, pumiceous clasts, and volcanic lithic grains. A horizons have a thickness of 10–15 cm. The presence of different kinds of voids is common, including channels and chambers, as well as the presence of rhizoliths and invertebrate trace fossils (*Coprinisphaera* isp. and *Celliforma* isp.). A horizons are well structured, with granular to subangular peds. Btk horizons (Figs. 5C, 5D, and 5E) show the presence of laminated to microlaminated, complete to incomplete, yellowish brown illuvial dusty clay infills. These horizons have a thickness of 20–50 cm, are well structured with subangular peds (Figs. 5C, 5D, 5E), and show grain-striated b-fabric (Fig. 5F). Bk horizons are 30–90 cm thick. Calcium carbonates are present as complete to incomplete infills and in the groundmass. Micritic calcite determines a calcitic crystallitic b-fabric (Figs. 5F and 5G). B/C horizons are characterized by the presence of scarce pedogenic features, such as calcium carbonate accumulation and rhizoliths, and by the preservation of depositional structures (e.g., diffuse stratification). From a geochemical perspective, the paleosols of pedotype III (see Supplementary Material [footnote 1]) show average values of 49.04 for the CALMAG weathering index, 50.32 for the CIA-K weathering index, 0.73 for the $Al/\Sigma Bases$ ratio, 0.15 for the Al/Si ratio, 0.22 for the total iron/alumina (Fe/Al) ratio, 0.06 for the Ti/Al ratio, and 0.38 for the $(K + Na)/Al$ index (Supplementary Material [footnote 1]).

Interpretation. The macromorphological and micromorphological features suggest that the main pedogenic processes that affected these paleosols were bioturbation, calcification, and, subordinately, argilluviation and vertization (Bullock et al., 1985; Stoops, 2003; Tabor and Myers, 2015). Bioturbation is evidenced by the presence of rhizoliths and trace fossils (*Coprinisphaera* isp. and *Celliforma* isp.; Stolt and Lindbo, 2010; Sarzetti et al., 2017). Calcification was the most important process in these paleosols, and it is indicated by the presence of calcium carbonates, such as calcitic crystallitic b-fabric and calcium carbonate coatings. Argilluviation was identified (Kühn et al., 2010) in the Btk horizons through the presence of laminated to multilaminated clay coatings and by the Al/Si values. The presence of grain-striated b-fabric in B horizons indicated the occurrence of vertization (Kovda and Mermut, 2010). These paleosols show the lowest CALMAG and CIA-K weathering index values (49.04 and 50.32, respectively), suggesting that the weathering was relatively weak (Sheldon and Tabor, 2009). The $Al/\Sigma Bases$ ratio (0.73) indicates a slight loss of exchangeable bases, and the low value of the Ba/Sr ratio (0.73) shows minimum leaching (Sheldon and

Tabor, 2009). The Ti/Al ratio (0.06) is consistent with volcanoclastic to epiclastic parental sediments (Sheldon and Tabor, 2009). Finally, the high relative values of the $(K + Na)/Al$ ratio indicate a salinization process. The pedogenic processes described here, the geochemical signature, and the presence of abundant pedogenic calcium carbonate are comparable with modern Aridisols (Soil Survey Staff, 1975, 1998; Retallack, 2001).

Paleotemperature and Paleoprecipitation Calculations for the Paso del Sapo Basin Paleosols

The parent material of the paleosols of the Paso del Sapo Basin show similar compositions for the different successions of pedotypes analyzed. Volcanoclastic components are abundant, and a mixed composition with subordinate epiclastic components is representative of the whole paleosol succession of the Paso del Sapo Basin. Since there is no meaningful evidence of composition changes, and the Ti/Al ratio remains more or less unchanged throughout the entire succession, no provenance change is interpreted for these paleosols, and a quantitative paleoclimate reconstruction using Ti/Al ratio content can thus be applied (Fig. 6; Varela et al., 2018). Because the CALMAG proxy provides the most reliable MAP estimates for Vertisols and paleosols with vertic features (Nordt and Dreise, 2010; Adams et al., 2011), the MAP values calculated using the CALMAG index were preferred for this study.

The MAT values calculated for the paleosols that characterize the La Pava Formation (succession of pedotype I) show mean values of 12 ± 2.1 °C (PWI index) and 12 ± 4.4 °C (SAL index; Fig. 6). The calculated MAP data show the maximum rainfall for the paleosols analyzed here (Fig. 6), with values of 1229 ± 108 mm/yr (CALMAG index), 881 ± 182 mm/yr ($\Sigma Bases$ index), and 871 ± 235 mm/yr (CIA-K index). MAP shows significant differences between the lower and upper sections of the paleosol succession of pedotype I: MAP values of 1357 ± 108 mm/yr (CALMAG index), 934 ± 182 mm/yr ($\Sigma Bases$ index), and 970 ± 235 mm/yr (CIA-K index) were calculated for the lower section, and MAP values of 1081 ± 108 mm/yr (CALMAG index), 802 ± 182 mm/yr ($\Sigma Bases$ index), and 791 ± 235 mm/yr (CIA-K index) were calculated for the upper section (Fig. 6).

The calculated MAT values for the paleosol succession assigned to pedotype II, which characterizes the lower section of the Collón Cura Formation, are similar to those calculated for pedotype I (Fig. 6), with values of 11 ± 2.1 °C (PWI index) and 10 ± 4.4 °C (SAL index). A de-

crease was registered in the MAP with respect to the mean values calculated for the basal paleosols of pedotype I (Fig. 6): 1053 ± 108 mm/yr (CALMAG index), 782 ± 182 mm/yr ($\Sigma Bases$ index), and 801 ± 235 mm/yr (CIA-K index).

In the upper section of the Collón Cura Formation, the succession of pedotype III paleosols was analyzed. These paleosols showed the lowest paleoprecipitation values throughout the entire Paso del Sapo Basin infill, and similar paleotemperatures compared with previous paleosols (Fig. 6). Average values of 10 ± 2.1 °C (PWI index) and 10 ± 4.4 °C (SAL index) were calculated for the MAT (Fig. 6), and average values of 677 ± 108 mm/yr (CALMAG index), 663 ± 182 mm/yr ($\Sigma Bases$ index), and 613 ± 235 mm/yr (CIA-K index) were calculated for the MAP (Fig. 6). It is important to note that significant differences in MAP values were observed between the lower and upper sections of the paleosol succession. Average MAP values of 513 ± 108 mm/yr (CALMAG index), 602 ± 182 mm/yr ($\Sigma Bases$ index), and 522 ± 235 mm/yr (CIA-K index) were calculated for the lower section of this paleosol succession (Fig. 6). Regarding the upper section, average MAP values of 841 ± 108 mm/yr (CALMAG index), 723 ± 182 mm/yr ($\Sigma Bases$ index), and 703 ± 235 mm/yr (CIA-K index) were determined (Fig. 6).

PALEOCLIMATE RECONSTRUCTION FOR THE LEEWARD SIDE OF THE NORTH PATAGONIAN ANDES

Based on macromorphological and micromorphological pedofeatures, and geochemical proxies obtained in this work, together with a recent high-resolution chronostratigraphic scheme for the Paso del Sapo Basin infill (Bucher et al., 2019b), the paleoclimatic conditions and main climate changes during its evolution could be established. Alfisol-like paleosols that characterize the lower section of the synorogenic sequence (La Pava Formation) were constrained to 15–14.6 Ma (Bucher et al., 2019b). Evidence such as well-developed, multilaminated clay coatings indicates strong seasonality conditions for these paleosols (Varela et al., 2018). MAT and MAP were determined to be 1229 ± 108 mm and 12 ± 2.1 °C, respectively (Fig. 6; CALMAG and PWI indexes). Hence, this Alfisol-like paleosol succession was assigned to a humid subtropical (temperate) climate (Zhang et al., 2016). Upward in the section, the synorogenic sequence continues with the lower section of the Collón Cura Formation, an Andisol-like paleosol succession constrained to 14.6–12.75 Ma (Bucher et al., 2019b), which has seasonal conditions with mean values of

Figure 6. Mean annual temperature (MAT) and mean annual precipitation (MAP) calculation curves for paleosols of the Paso del Sapo Basin. Shaded areas of each curve correspond to the standard error of each climofunction. The values of MAP and MAT in the boxes show the mean values for each pedotype and correspond with the CALMAG and PWI indexes, respectively. From left to right: Integrated section of the Paso del Sapo Basin infill indicating the paleosol level analyzed (green arrows), Ti/Al ratio, MAP, and MAT. M—mudstones; S—sandstones; C—conglomerates.

1053 ± 108 mm/yr and 11 ± 2.1 °C for MAT and MAP, respectively (Fig. 6; by using CALMAG and PWI indexes). These features make it possible to assign a humid subtropical (temper-

ate) climate to this succession of Andisol-like paleosols as well (Zhang et al., 2016). For the uppermost part of the synorogenic sequence (upper section of the Collón Cura Formation),

a succession of Aridisols-like paleosols was constrained to 12.75–11.5 Ma (Bucher et al., 2019b). For this paleosol succession, mean values of MAT and MAP of 677 ± 108 mm/yr and 10 ± 2.1 °C, respectively, were determined (Fig. 6; CALMAG and PWI indexes). These values indicate a humid subtropical (temperate) climate (Zhang et al., 2016) with dry seasons, marked by the abundance of pedogenic calcium carbonates (Retallack, 2001; Tabor and Myers, 2015). Following the criteria of Köppen (1923), the succession may be assigned to a dry cold steppe climate. It is important to highlight that at 12.75–12.5 Ma, the MAP shows minimum mean values (~600 mm/yr) at the base of the upper Collón Cura Formation (Fig. 6).

During the evolution of the synorogenic sequence, the mean MAT values ranged from 13 ± 2.1 °C to 10 ± 2.1 °C (PWI index; Fig. 6). Taking into account the uncertainties of the method applied, we consider these values as statistically equivalents, and the MAT could be considered constant in the entire succession. The mean MAP values decreased from 1357 ± 108 mm/yr to 677 ± 108 mm/yr (Fig. 6; CALMAG index). This meaningful decrease by >600 ± 108 mm/yr suggests an aridization trend, in agreement with previous qualitative data (Bellosi et al., 2014; Sarzetti et al., 2017).

The MAP and MAT values calculated for the Andean foreland region in the Paso del Sapo Basin can be compared with mean precipitation and temperature values currently recorded on windward and leeward locations of the Andes. The paleosols determined at the base of the synorogenic sequence (lower section of the La Pava Formation) have mean MAP and MAT values of 1357 ± 108 mm/yr (CALMAG index) and 13 ± 2.1 °C (PWI index), respectively. These values are similar or slightly lower than the present-day values of the windward regions of the North Patagonian Andes (e.g., Puerto Varas and Osorno localities; Fig. 7), which show MAP and MAT values of around 1350–1750 mm/yr and 11 °C, respectively (Climate Watch, 2018). It is important to highlight that the higher MAP values registered currently on the windward regions of the Andean range may be related to an increase in rainfall caused by their position on the uplifted flank (a comparison with lowland sites would have been preferable, but there are not better low-elevation localities with available data to analyze and compare). In turn, modern analogues for the uppermost section of the synorogenic sequence (upper section of the Collón Cura Formation) with mean MAP and MAT values of 677 mm/yr and 10 °C, respectively, are currently found on the leeward side of the North Patagonian Andes at similar latitudes (e.g., El Maiten or Esquel localities; Fig. 7), with MAP

LINKING PALEOCLIMATIC CHANGES WITH THE NORTH PATAGONIAN ANDEAN UPLIFT

The timing of the contractional deformation related to the Miocene uplift of the North Patagonian Andes has been geochronologically constrained between 19 and 10 Ma (Folguera and Ramos, 2011; Orts et al., 2012; Bilmes et al., 2013; Ramos et al., 2015; Gianni et al., 2015, 2017; Echaurren et al., 2016; Folguera et al., 2018; Bucher, 2018; López et al., 2019; Bucher et al., 2019b). During this time, deformation was not a continuous process. At the onset, the deformational front was restricted to the Andean domain ca. 19–18 Ma. The deformational front migrated eastward between 17 and 15 Ma, with the development of continuous deformation in the Andean and foreland regions between 16 and 10 Ma (Folguera and Ramos, 2011; Orts et al., 2012; Bilmes et al., 2013; Ramos et al., 2015; Gianni et al., 2015, 2017; Echaurren et al., 2016; Folguera et al., 2018; Bucher, 2018; López et al., 2019; Bucher et al., 2019b). After 10 Ma, low-temperature thermochronology data, used to constrain the exhumation history for the Patagonian Andes, indicate that the topographic growth of the North Patagonian Andes has been controlled by high glacial denudation rates since 7 Ma (Thomson, 2002; Thomson et al., 2010).

The regional Miocene climate change of the Andean foreland has been recorded by several authors from significant ecological and biological fluctuations, which they regard as being partially synchronous with uplift of the Patagonian Andes (Blisniuk et al., 2005; Ortiz-Jaureguizar and Cladera, 2006; Palazzesi and Barreda, 2012; Le Roux, 2012; Raigemborn et al., 2018). These works suggest that the climate change from temperate-warm and humid to colder and drier conditions occurred in response to both regional and global causes. On the one hand, these works support the hypothesis that the regional cause of the aridization of the Andean foreland was related to the configuration of an orographic barrier as a result of the Patagonian Andes uplift. Thus, the uplift of the mountain range blocked moisture from the west and caused a rain shadow to the east (on the leeward side of the mountains), producing more arid conditions. This process reduced seasonal precipitation, favoring grassland expansion, which gave rise to a new biome: the Patagonian steppe (“*Estepa Patagónica*”; e.g., Palazzesi and Barreda, 2012; Raigemborn et al., 2018). On the other hand, partially synchronous global climate changes are registered in several regions of the world, including the Patagonian region (Blisniuk et al., 2005; Ortiz-Jaureguizar and Cladera, 2006; Palazzesi and Barreda, 2012; Le Roux, 2012; Raigemborn et al., 2018). The

Figure 7. Tectonic-climate evolution for the North Patagonian Andean and foreland region. (A) Map of the studied area showing the present-day topographic relief and the geographic locations of the localities used as modern analogues (masl—meters above sea level). (B) Cross sections (see location of X-X' in A), indicating main climatic conditions and changes for the different stages in the Miocene to present Andean uplift. Miocene paleoaltitudes are roughly estimated, using the minimum height of 1 km for the onset of a rain shadow effect (see Blisniuk et al., 2005). MAP—mean annual precipitation; MAT—mean annual temperature.

and MAT values of 600–650 mm/yr and 9 °C (Climate Watch, 2018), respectively. It is important to emphasize that the Paso del Sapo region

at present shows mean values of MAP and MAT of 208 mm/yr and 11 °C (Climate Watch, 2018; Fig. 7), respectively.

mid-Miocene climatic optimum occurred between 17 and 15 Ma (Flower and Kennett, 1994; Zachos et al., 2001; Böhme, 2003; Lewis et al., 2007; Sun and Zhang, 2008; You et al., 2009), and subsequent cooling is recorded from 16–15 to 10 Ma (Flower and Kennett, 1994; Zachos et al., 2001; Böhme, 2003; Lewis et al., 2007; Sun and Zhang, 2008; You et al., 2009). This cooler phase was associated with rapid Antarctic ice-sheet growth and major biogeographic changes (Zachos et al., 2001). Quantitative data provided by You et al. (2009) indicate that the mid-Miocene MAT variations were around 3 °C. So far, among the few works focusing on the timing for the aridization of the Patagonian Andean foreland region, some suggest a gradual aridization process between ca. 16.5 Ma to 10 Ma (Riccardi and Rolleri, 1980; Blisniuk et al., 2005; Ortiz-Jaureguizar and Cladera, 2006; Le Roux, 2012; Palazzesi and Barreda, 2012). In turn, data from fossil pollen indicate that between 10 and 6 Ma, annual precipitation values were rarely below ~600 mm/yr (Palazzesi et al., 2014).

The initial paleosols of the studied sequence (ca. 15–14.6 Ma) indicate paleoclimatic conditions similar to modern climate conditions for the localities on the windward side of the North Patagonian Andes. It is remarkable that these higher modern MAP values windward may be due to the effect of the orographic barrier, which increases precipitation (Roe, 2005; Hughes et al., 2009; Hoorn et al., 2018). Modern analogues also indicate that similar climate conditions for the upper section of the synorogenic sequence, constrained from 14.6 to 11.5 Ma, may be found on the leeward side of the North Patagonian Andean segment. In addition, the present-day climate conditions for the studied region indicate that the aridization continues, decreasing from ~400 mm/yr since 11.5 Ma (Fig. 7). Taking into account data presented by Palazzesi et al. (2014), the decrease in the MAP values from 600 to 200 mm/yr occurred after 6 Ma. In addition to the thermochronological data presented by Thomson (2002) and Thomson et al. (2010), which suggest that the topographic growth of the North Patagonian Andes has been limited since 7 Ma, the processes that generated the final stage of aridization (from ~600 to ~200 mm/yr) have been related not only to the Andean uplift, but also to other processes (i.e., global climate changes), as has already been proposed by Palazzesi et al. (2014).

The data presented in this paper indicate constant values of MAT and a meaningful decrease in MAP (>600 mm/yr) between 15 and 11.5 Ma for the North Patagonian Andean foreland. Taking into account the stable MAT, the meaningful decrease in MAP values (from ~1350 mm/yr to ~700 mm/yr) has been interpreted as mainly

related to the growth and uplift of the Patagonian Andes and the generation of a rain shadow, probably enhanced by the global changes (Fig. 7). At a lower scale, the minimum MAP values of ~600 mm/yr calculated in this region at 12.75–12.5 Ma, and the abrupt MAP decrease by ~300 mm/yr constrained between 15 and 14.5 Ma, could be interpreted as high-frequency climate variations not related to the Andean uplift (e.g., high-frequency climatic cycles registered in the Ebro Basin; Valero et al., 2014).

Is important to note that our results do not show significant variations in the MAT, as was proposed by several authors for this time period, either as a response to the global cooling phase registered after 16–15 Ma (after the mid-Miocene climatic optimum; Flower and Kennett, 1994; Zachos et al., 2001; Böhme, 2003; Lewis et al., 2007; Sun and Zhang, 2008; You et al., 2009), or related to the cooling associated with the uplift of the orographic barrier. We interpret that this issue is related to the uncertainties of the method applied, because the errors of the indexes (2.1 °C for PWI index and 4.4 °C for the SAL index) are higher than the absolute mid-Miocene MAT variations (around 3 °C following simulations made by You et al., 2009). Therefore, MAT estimations using PWI and SAL indexes are limited, and other proxies should be used in order to obtain a better constraint on the MAT variations for the analyzed succession.

The new paleoclimatic quantitative data set indicates that the onset of the rain shadow effect at this latitude may be constrained to ca. 14.6 Ma, in partial agreement with previous works (Riccardi and Rolleri, 1980; Blisniuk et al., 2005; Ortiz-Jaureguizar and Cladera, 2006; Palazzesi and Barreda, 2012; Le Roux, 2012). As a final assessment, we stress that the beginning of the uplift of the Patagonian Andes, constrained to 19–18 Ma, would have had a time delay of ~4 m.y. until rain shadow generation. This indicates that, during the first phase of mountain growth (Fig. 7; 19–18 to 14.6 Ma), the North Patagonian Andes were not high enough to generate recordable effects of the orographic barrier (1 km of height according to Blisniuk et al., 2005). However, the global climate changes registered in Miocene times and the lack of quantitative paleoclimate data between 19 and 15 Ma suggest that more detailed studies should be carried out in order to demonstrate this proposed time delay.

The tectonically controlled climate change addressed in this work reflects the interactions between tectonics and climate in the North Patagonian Andean foreland. The onset of the orographic uplift and the generation of a rain shadow are constrained to early–middle Miocene times. Similar processes have been addressed in several works for the Central Andes, indicating a later

(late Miocene) rain shadow generation related to orographic barrier uplift (Kleinert and Strecker, 2001; Sobel and Strecker, 2003; Strecker et al., 2007, 2012; Poulsen et al., 2010; Mulch et al., 2010; Rohrmann et al., 2016). Likewise, it is important to note that the rainfall shadow resulting from the growth of the Andes has continued to evolve until today.

CONCLUSIONS

A continuous synorogenic paleosol succession from the North Patagonian foreland region was analyzed at the Paso del Sapo Basin. Based on a multiproxy analysis of paleosols, which included macromorphological, micromorphological, and geochemical studies, supported by a high-resolution chronostratigraphic scheme, three different types of paleosol successions were defined. Alfisol-like paleosols that characterize the lower section of the synorogenic sequence (La Pava Formation) were constrained to 15–14.6 Ma, and they show MAP and MAT values of 1229 ± 108 mm and 12 ± 2.1 °C, respectively. Middle Andisol-like paleosols (lower Collón Cura Formation), constrained from 14.6 to 12.75 Ma, exhibit mean values of MAP and MAT of 1053 ± 108 mm/yr and 11 ± 2.1 °C, respectively. Finally, values of 677 ± 108 mm/yr and 10 ± 2.1 °C for the MAT and MAP variables, respectively, were calculated for the Aridisol-like paleosols that characterize the uppermost section of the synorogenic sequence (upper Collón Cura Formation), constrained to 12.75–11.5 Ma. Our results, combined with regional tectonic and climate contributions, indicate that although the onset of uplifting of the North Patagonian Andes has been constrained to 19–18 Ma, the rain shadow effect is not recorded before ca. 14.6 Ma.

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