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Conodont geothermometry of the lower Paleozoic from the Precordillera (Cuyania terrane), northwestern Argentina

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ABSTRACT

The thermal history of the Precordillera terrane of northwestern Argentina has been constrained by the conodont colour alteration index (CAI) in combination with previously published paleothermal data (e.g., illite crystallinity and clay mineral assemblages). The pattern of paleotemperatures displays an increase in paleotemperatures to the west and south of the basin. This configuration shows a gradual and continuous transition from diagenesis to low-grade metamorphism, which is apparently not controlled by any of the morpho-structural subdivisions of the Precordillera (i.e., Western, Central, or Eastern). According to our results, the lower Paleozoic sedimentary burial played a secondary role in the heating of the Precordillera. Instead, the predominant component was loading by thrust sheets, which reflects the effects of the Devonian collision of Chilenia, particularly, in the Western Precordillera. Conversely, our paleothermometric data from the easternmost exposures of the Precordillera do not evidence anomalies referable to any of the accretionary events that contributed to the early Paleozoic building of the southern proto-Andean margin of Gondwana. Instead, the expected thermally altered conodonts from the Cuyania accretion are represented by metamorphosed conodont elements transported to the deeper settings of the west. The CAI data also suggest that overburden depth varied from ca. 3.6 km in the shelf region of the Eastern Precordillera to ca. 12 km in the slope to rise deposits of the Western Precordillera, thus providing constrains for the palinspastic restoration across the orogen. On the other hand, the smooth increase of peak paleotemperatures to the south of the Precordillera is associated with the exposure of deeper crustal levels at that sector, probably related to larger shortening due to stronger collisional effects, or alternatively, a weaker mechanical response of its elastic lithosphere.

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RESUMEN

La historia térmica del terreno de la Precordillera del noroeste de Argentina se ha analizado mediante el Índice de Alteración del Color (CAI) de conodontes en combinación con otros parámetros paleotermométricos publicados con anterioridad (e.g., cristalinidad de illita y asociaciones paragenéticas de arcillas). El patrón de paleotemperaturas exhibe un incremento hacia sur y oeste de la cuenca. Esta configuración muestra una transición gradual y continua desde la diagénesis al metamórfismo de bajo grado, la cual aparentemente no esta controlada por las subdivisiones morfoestructurales de la Precordillera (i.e., Occidental, Central, u Oriental). De acuerdo a nuestros resultados, la cubierta sedimentaria del Paleozoico inferior constituyó un control secundario en el calentamiento de la Precordillera, a diferencia de la carga tectónica que representa el componente principal. Esta refleja los efectos de la colisión devónica de Chilenia que se concentran, particularmente, en la Precordillera Occidental. La información paleotermométrica de los afloramientos más orientales de la Precordillera no presenta anomalías térmicas que puedan referirse a procesos de acreción. No obstante, se reconocen conodontes con alteración térmica, procedentes de paleoambientes orientales y transportados a los sectores occidentales más profundos de la cuenca como resultado de la colisión de Cuyania con el margen proto-Andino de Gondwana en el Paleozoico inferior. Los datos CAI también sugieren que la profundidad de soterramiento varía desde ca. 3.6 km en el sector de la plataforma carbonática de la Precordillera Oriental a ca. 12 km en los

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depósitos de talud a pie de talud de la Precordillera Occidental. Por otra parte, el leve incremento de paleotemperaturas hacia el sur de la Precordillera se relacionaría con la exposición de niveles corticales más profundos en ese sector. Esto, a su vez, podría vincularse a un mayor acortamiento debido a efectos colisionales importantes o a una respuesta mecánica más débil de la litósfera elástica, entre otras causas posibles.

1. Introduction

The Precordillera of NW Argentina plays a key role in a number of West-Gondwanan tectonic reconstructions from the early Paleozoic, where it stands at the external zone of the Famatinian orogen (Pankhurst and Rapela, 1998; Keller, 1999; Quernardelle and Ramos, 1999). Major changes in its basin configuration and paleobiogeographic affinities were classically related to stages of rifting, drifting and collision of the Precordillera (as part of the Cuyania composite terrane) with the South American margin (Ramos et al., 1986, 1998; Astini et al., 1995; Thomas and Astini, 2003). Following accretion of Chilenia to its proto-Gondwana margin leaded to foreland basin rearrangements and to a Siluro–Devonian thermal overprint (Ramos et al., 1986; Astini, 1996). The Chilenia and the Cuyania terranes were finally amalgamated in the Carboniferous by continental sedimentation punctuated by short transgressive periods. Continental, glacigenic and marine upper Paleozoic rocks accumulated in the Paganzo and Río Blanco foreland basins, with increasing dominance of offshore facies to the west. Locally, middle to late Triassic sequences linked to the Cuyo basin deposited in angular unconformity on a NW-trending asymmetric rift, in westernmost Precordillera. This basin was heavily controlled by the basement fabric and developed over the ancient suture zone with Chilenia. Alternatively, the Precordillera acted as a positive structural element for most of the Mesozoic. According to most supported hypothesis, the tectonic evolution of the Precordillera remained relatively stable until Andean triggering of the thin-skinned fold and thrust belt (Jordan et al., 1983, 1993; Ramos et al., 2002).

The thermal maturation history of the Precordilleran basin is a critical parameter through which the Cuyania terrane can be evaluated, but has often remained overlooked. Fortunately, conodonts are abundant and diverse in the lower Paleozoic sedimentary sequences of the Precordillera. These tooth-like apatitic microfossils can provide, with a precise biostratigraphic control, important information on the thermal maturation of the sedimentary rocks in which they occur. The latter property is based on the original investigations of Epstein et al. (1977) and Rejebian et al. (1987), who established the conodont Color Alteration Index (CAI) and constrained their values according to heat and exposure time. This paper examines the thermal distribution patterns of the Precordilleran basin, founded on CAI values and complemented with previously published studies on metamorphic indexes, in order to asses the geotectonic evolution of the Cuyania terrane and, additionally, provide essential clues for future hydrocarbon explorations in the region.

2. Geological framework

The Precordillera has typically been defined as a high-level fold and thrust belt that extends between 29° and 33° S in the Andean foothills (Allmendinger et al., 1990; von Gosen, 1992). The Bermejo and Jocolí basins separate the Precordillera from fault-bounded crystalline basement uplifts of the Sierras Pampeanas to the east, whereas the Uspallata–Calingasta–Iglesia basin covers the structural boundary with the Cordillera Frontal to the west (Fig. 1). It is arranged in 4–6 major NS trending belts mostly composed of Cambrian to Carboniferous rocks, depending on the latitude.

An Eastern, Central, and Western domains have classically been distinguished following structural and stratigraphical criteria. The Eastern and Central Precordillera involves an important passive margin carbonate platform, Cambro-Ordovician in age, which is covered by siliciclastic foreland deposits. The Western Precordillera exhibits deeper environments, with slope to ocean floor deposits, which include pillow lavas and mafic–ultramafic bodies in the westernmost sections. It is affected by a very low-grade metamorphism that locally reaches green-schists facies and shows evidence of a complex deformation and metamorphism during the Ordovician (Alonso et al., 2008; Voldman et al., 2009) and Silurian to Devonian times (von Gosen, 1992; Buggish et al., 1994; Astini, 1996; Davis et al., 1999; Gerbi et al., 2002; Robinson et al., 2005).

On the other hand, the igneous rocks from the Precordillera are limited to the lower Paleozoic mafic–ultramafic ophiolitic bodies, Carboniferous acid plutons and scarce and isolated Neogene volcanic centres with related shallow intrusions associated with the shallowing of the subduction and the foreland migration of the volcanic arc (Ramos et al., 2002).

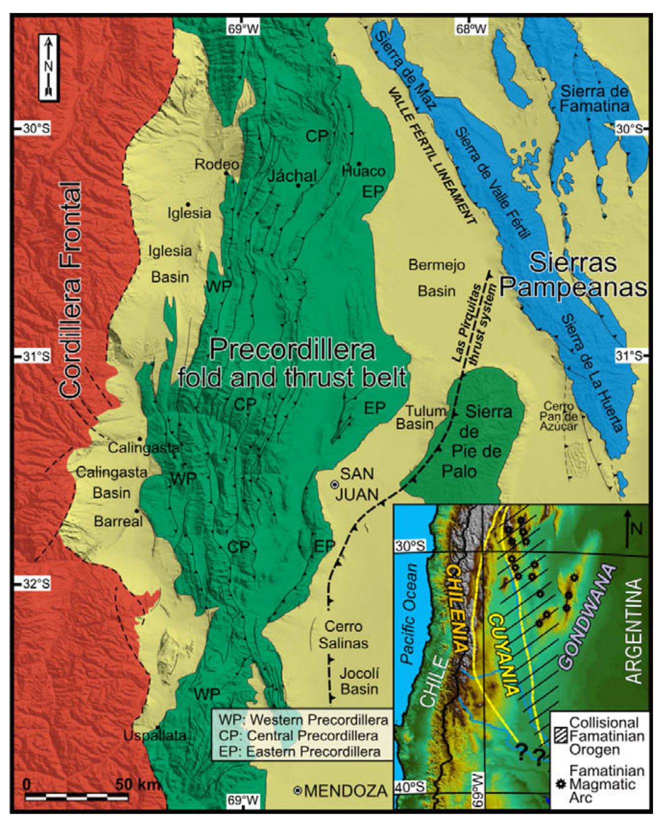

Fig. 1. Location map and major geologic structures of the Argentine Precordillera thrust and fold belt, as part of the Cuyania terrane, superposed on shaded relief map derived from Shuttle Radar Topography Mission (SRTM).

2.1. Pre–Andean history and deformations

The proto-Andean margin of Gondwana was subjected to important tectono-stratigraphic events in early to middle Paleozoic times, included in the Famatinian Orogeny, and related to the accretion of the Cuyania and Chilenia terranes (Ramos et al., 1986; Astini et al., 1995; Astini, 1996; Davis et al., 1999). Several lines of evidence had been interpreted as the onset of deformation related to the accretion of Cuyania in Mid-Late Ordovician times (Ocloyic orogeny): (1) drowning of the carbonate platform and introduction of a synorogenic clastic wedge in the basin (Astini et al., 1995); (2) peak Ocloyic metamorphism (ca. 470–460 Ma) of crystalline rocks and their metasedimentary cover from the Sierras Pampeanas and Famatina (Baldo et al., 2001; Casquet et al., 2001); (3) cessation of subduction-related Famatinian magmatic arc activity (Ramos et al., 1998; Coira et al., 1999; Thomas and Astini, 2003); and (4) the vestiges of a west vergent Ocloyic thin-skinned thrust-belt overprinted by the Andean deformation at the Guandacol River area (Thomas and Astini, 2007), among others (Fig. 3). Earlier proposals about Cuyania as an accreted fragment (Dalla Salda et al., 1992), or as a marginal plateau (Dalziel, 1997), left behind after a continent–continent collision in the Ordovician, are difficult to reconcile with the faunal, isotopic, paleomagnetic, geochronological and stratigraphic evidences. However, alternative hypotheses consider that the Cuyania terrane rifted during the Ordovician and collided in the Devonian as a native block from Gondwana (Aceñolaza et al., 2002; Finney, 2007) or as a marginal plateau adjacent to Laurentia (Keller, 1999).

A recent contribution recorded an Ordovician metamorphic event by means of conodont paleothermometry in the slope sedimentary sequences of the Western Precordillera (Voldman et al., 2009). Allochthonous conodonts from reworked deposits probably sourced from the eastern carbonate platform and autochthonous conodonts from the enclosing matrix allowed constraining the metamorphic age between the *Paltodus deltifer* and the *Lenodus variabilis* zones (latest Tremadocian – earliest Darriwilian, 480–465 Ma). From an assessment of basin thermal history, conodont CAI data is in agreement with an Ordovician instead of Devonian collision for the Cuyania terrane, which is meant consistent with a drifted microcontinent model. Another recent study compares the paleobiogeographic evolution of the conodont faunas from the Precordillera with those from Laurentia and other continents by multivariate analysis, verifying the hypothesis of "the Precordillera as a Laurentian derived terrane" as not conclusive (Albanesi and Bergström, in press), in contrast with prevailing arguments (e.g., Benedetto, 2004).

In the western slope facies of the Precordillera, pillow lavas and mafic–ultramafic sills are preserved in the northern sectors together with distal turbidites and hemipelagites, whereas ultramafic rocks dominate in the southern sections (Ramos et al., 2000). The ophiolitic nature of these rocks is fundamental to interpret the closing of an ocean basin that extended between the Cuyania and Chilenia terranes (Ramos et al., 1986; Davis et al., 1999). However, the geodynamic context regarding the Chilenia collision continues being a matter of debate, such as the tectonic mechanism for the accretion, the timing (Astini, 1996; Davis et al., 1999) or even its very existence (Dalziel et al., 1994; Willner et al., 2008). The collision of Chilenia renewed deformation within the Cuyanian foreland (Ramos, 2004), and probably induced reactivation of its eastern suture (Astini, 1996).

Geochronologic dating of metamorphic rocks from the Precordillera and adjacent blocks (San Rafael Block, Western Sierras Pampeanas and Cordillera Frontal) suggest that the final stage of docking of Chilenia (Chanic orogeny) probably occurred during the Mid-Late Devonian (Cucchi, 1971; Buggish et al., 1994; Basei et al., 1998; Ramos et al., 1998; Davis et al., 1999; Tickyj et al., 2001; Varela et al., 2003). Compressive deformation probably extended into the early Carboniferous and is responsible for the Protoprecordillera fold and thrust belt, a NS-trending structural high, which separated the western basins of Río Blanco–Calingasta–Uspallata from the eastern basin of Paganzo (Ramos et al., 1986; Limarino et al., 2006).

Nevertheless, the role of the Protoprecordillera in the upper Paleozoic paleogeography is controversial. The occurrence of upper Paleozoic marine deposits in the Central Precordillera, the style of deformation and the presence of igneous-metamorphic sources derived-sediments in the western region downplay the relevance of this structural high (González Bonorino, 1990; von Gosen, 1992), and preclude a large tectonic thickening. The Protoprecordillera progressively loosed relief until its final collapse after post-oro-

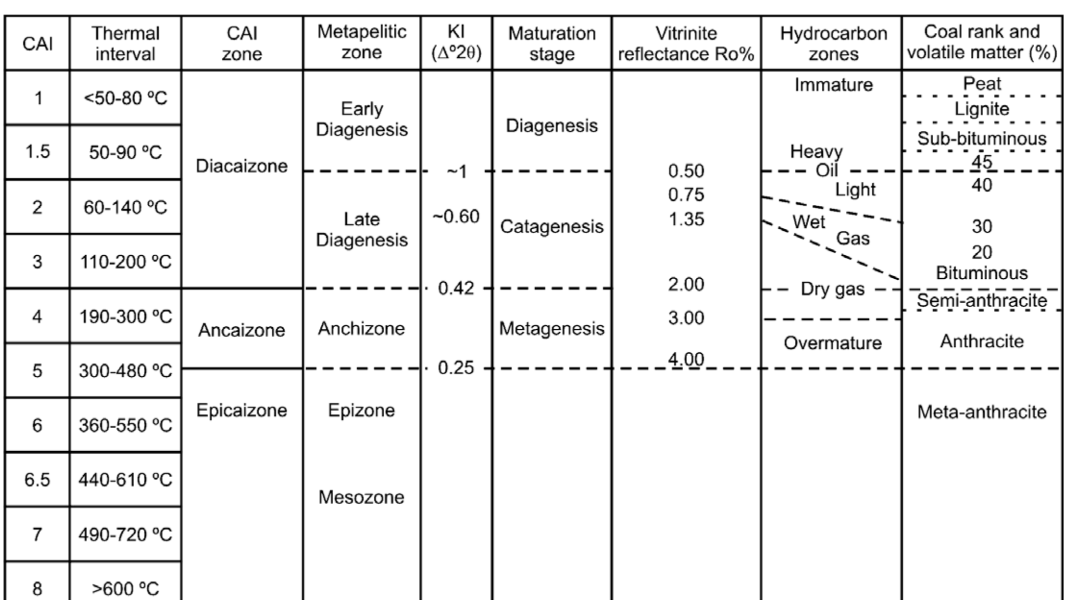

Fig. 2. Conodont CAI values contrasted with other low-grade metamorphism indexes. Adapted after Voldman et al. (2008).

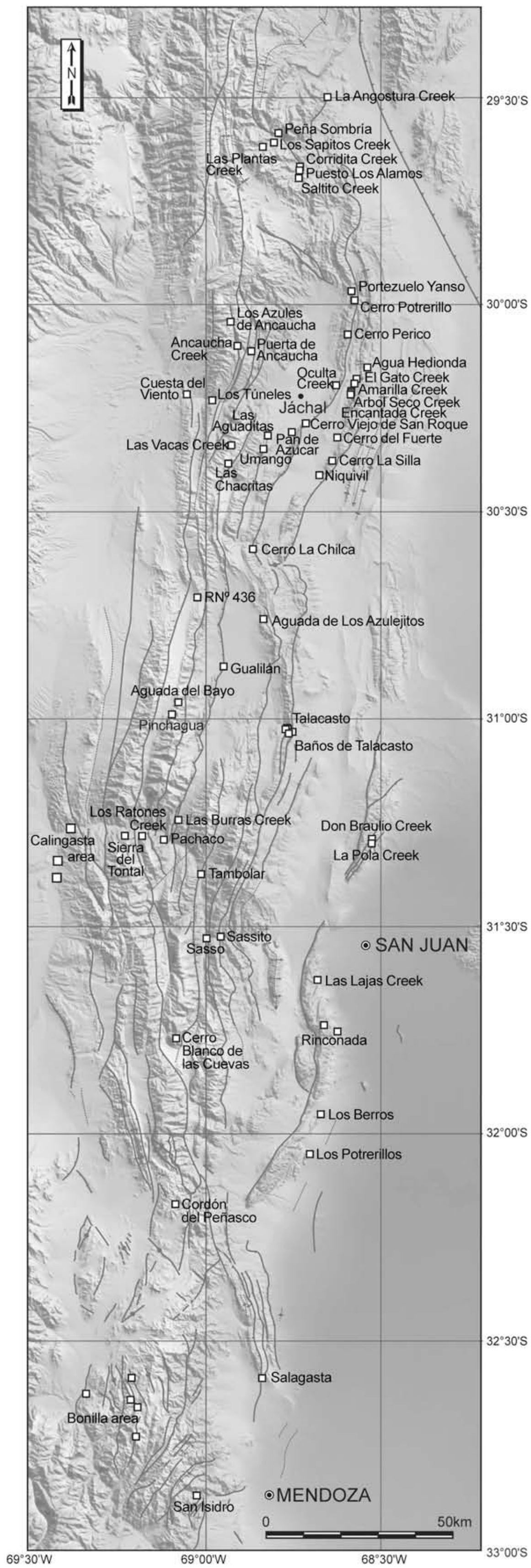

Fig. 3. Localities with paleothermometric records following Tables 1–3.

genic extension or flexural subsidence during the Neopaleozoic deformation.

Subsequently, a new geodynamic regime driven by the Gondwana breakup since the early Permian was accompanied by an important tectonic activity (Gondwanic cycle; Ramos, 1999). In the Andean margin of the Río Blanco–Calingasta–Uspallata basins, siliciclastic sedimentation was replaced by very thick sequences of acid volcanic and volcaniclastic rocks that belong to the Choiyoi Group (Llambías, 1999, and references therein). Conversely, this volcanism was not present in the Paganzo basin; in turn, the tectonism appears reflected in local unconformities and important changes in the paleocurrent patterns (Limarino et al., 2006).

2.2. Neogene deformation

Andean shortening resulted in contractional and transpressive deformation of the Precordillera. Neogene to present Andean continental-deposits lay unconformably in a series of successor and intramontaneous basins, some of them active as piggy-back basins within the thrust belt (Beer et al., 1990; Zapata and Allmendinger, 1996a,b). The basal orogenic sequences were deposited between 21 and 18 Ma in a foreland setting that originally extended all over the Precordillera, in response to active crustal shortening and thickening of the Frontal Cordillera to the west (Jordan et al., 1993; Ramos, 1999; Verges et al., 2001; Milana et al., 2003). Progressive thrusting of the Precordillera steps forwards into a broken foreland basin with the uplift of Sierras Pampeanas (Jordan et al., 1993), with very minor strike-slip components (Siame et al., 2006).

The northern and southern areas of the Precordillera show the more complex structures, which resulted from tampering the Neogene to Present Andean deformation with inherited structural anisotropies and paleogeographic influences (Cortés et al., 2005). For example, the Triassic Río Blanco half-grabens were tectonically inverted, with reactivation of its NW-trending Paleozoic structures. Alternatively, structural studies simplified the Precordillera as a typical fold and thrust belt, shortened near 70–100% disregarding the variable strata thicknesses and the pre-Andean deformations (e.g., Allmendinger et al., 1990; von Gosen, 1992; Cristallini and Ramos, 2000; Cardozo and Jordan, 2001).

3. The conodont color alteration index (CAI)

It has long been known the change of color in conodont elements exposed to heat since the pioneer experiment of Ellison (1944). Epstein et al. (1977) and Rejebian et al. (1987) quantified the conodont CAI and calibrated it by means of laboratory procedures and empirical data. These authors demonstrated that the change of color in conodont elements is progressive, accumulative and irreversible. Unweathered conodonts are yellow to light brown and have a smooth surficial texture (CAI 1). When exposed to heat, minute organic matter lamellas included in the conodont structure progressively coalificate and volatilize, though transforming the color of the elements. The microfossils gradually darken until they become black at approximately 300 °C (CAI 1.5–5). At higher temperatures, conodonts recrystalize and release carbon and water from their structure. In this stage, they progressively change from black to gray (CAI 6), to white (CAI 7) and translucent (CAI 8) (Rejebian et al., 1987). Finally, conodont disintegration starts above 600 °C.

The conodont colour alteration sequence follows the Arrhenius reactions, therefore responds directly to temperature and duration of heating. Other variables such as dolomitization, hydrothermalism, contact metamorphism and diagenesis are usually accompanied by indicative textural alterations of the conodont surfaces, such as patinas or coarse recrystallization (cf., Königshof, 2003; Voldman et al., 2008; Blanco-Ferrera et al., 2009). Organic-rich rocks, such as bituminous shales, can eventually influence the

Table 1
Conodont CAI values from the Precordillera (this study). Localities follow the NS-trending as shown in the map of Fig. 3.

Location	Formation (level)	CAI	Conodont zone (age)
Peña Sombría (Guandacol River area)	San Juan (top)	4.5	*Oepikodus evae* (Early Ordovician)
Los Sapitos Creek (Guandacol River area)	San Juan (top)	2	*Tripodus laevis* (Middle Ordovician)
Las Plantas Creek (Gualcamayo River area)	Gualcamayo (middle part of Lower Member)	2.5–3.5	*Tripodus laevis* (Middle Ordovician)
Corridita Creek (Gualcamayo River area)	San Juan (top)	4–5	*Microzarkodina parva* (Middle Ordovician)
Corridita Creek (Gualcamayo River area)	Gualcamayo (middle part and top of Lower Member)	4–5	*Microzarkodina parva* (Middle Ordovician)
Puesto Los Álamos (Gualcamayo River area)	San Juan (top) and Gualcamayo (top of Lower Member)	3	*Baltoniodus navis – M. parva* (Middle Ordovician)
Saltito Creek (Gualcamayo River area)	Gualcamayo (top of Lower Member)	2.5	*Microzarkodina parva* (Middle Ordovician)
Portezuelo Yanzo	San Juan	2	*Oepikodus evae* (Early Ordovician)
Cerro Potrerillo	San Juan	2	*Lenodus variabilis* (Middle Ordovician)
Los Azules (Ancaucha Creek)	Yerba Loca (middle part)	4–4.5	*Lenodus variabilis* (Middle Ordovician)
Cerro Perico	San Juan	2	*Lenodus variabilis* (Middle Ordovician)
Pantanito (Ancaucha Creek)	Yerba Loca (middle part)	3	*Lenodus variabilis* (Middle Ordovician)
Puerta de Ancaucha	Los Sombreros (top)	3–4	*Lenodus variabilis* (Middle Ordovician)
Agua Hedionda	San Juan (top)	2	*Oepikodus evae* (Early Ordovician)
Los Gatos Creek	San Juan (top)	1.5–2	*Lenodus variabilis* (Middle Ordovician)
Oculta Creek	San Juan (top)	2,5	*Lenodus variabilis* (Middle Ordovician)
Amarilla Creek	San Juan (top)	1.5–2	*Lenodus variabilis* (Middle Ordovician)
Árbol Seco Creek	Los Azules (top of Upper Member)	2	*Amorphognathus tvaerensis* (Late Ordovician)
Encantada Creek	Los Azules (top of Middle Member)	1.5	*Pygodus serra* (Middle Ordovician)
Jáchal River	Los Sombreros (top)	3	*Lenodus variabilis* (Middle Ordovician)
Cerro Pan de Azúcar	San Juan (lower part)	3	*Oepikodus evae* (Early Ordovician)
Las Aguaditas	Las Aguaditas (base, lower, upper and top)	3	*Eoplacognathus pseudoplanus – Amorphognathus tvaerensis* (Middle-Late Ordovician)
Umango	San Juan (middle part and top)	2.5	*?Tripodus laevis* (Middle Ordovician)
Cerro La Silla	La Silla (top) and San Juan (base)	1.5	*Paltodus deltifer* (Early Ordovician)
Las Chacritas	Las Chacritas (base and top)	2.5–3	*Eoplacognathus suecicus* (Middle Ordovician)
Niquivil	San Juan (upper part)	1.5–2	*Oepikodus intermedius* (Middle Ordovician)
La Chilca	San Juan (upper part)	2.5	*?Lenodus variabilis* (Middle Ordovician)
Sierra de la Invernada (RN° 436)	Sierra de la Invernada (top)	2	*Eoplacognathus suecicus* (Middle Ordovician)
Sierra de la Invernada (Aguada del Bayo)	Sierra de la Invernada (middle part)	2	*Lenodus variabilis* (Middle Ordovician)
Sierra de la Invernada (Pinchagua Creek)	Sierra de la Invernada (middle part)	2.5	*Pygodus serra* (Middle Ordovician)
Talacasto (Poblete Norte Creek)	San Juan (top)	2.5–3	*?Lenodus variabilis* (Middle Ordovician)
Talacasto (RN° 436)	San Juan (base and middle part)	3	*?Lenodus variabilis* (Middle Ordovician)
Talacasto (Poblete Sur Creek)	San Juan (top)	2.5–3	*?Lenodus variabilis* (Middle Ordovician)
Baños de Talacasto	San Juan (top)	2,5	*Lenodus variabilis* (Middle Ordovician)
Talacasto (Ancha Creek)	San Juan (top) and Los Espejos (middle part)	2.5–3	*Lenodus variabilis* (Middle Ordovician) – *Kokelella v. variabilis* (Ludlow)
Las Burras Creek	San Juan (middle part)	4	*?Oepikodus intermedius* (Middle Ordovician)
Los Ratones Creek	San Juan olistolith	4	*?Oepikodus evae* (Early Ordovician)
Villicum (Don Braulio Creek)	San Juan (top), Gualcamayo (top of Middle Member), La Cantera (base)	3	*Lenodus variabilis* (Middle Ordovician) – *Pygodus serra* (Middle Ordovician)
Pachaco	San Juan (middle and top)	4–4.5	*Oepikodus intermedius* (Middle Ordovician)
Villicum (La Pola Creek)	San Juan (top), Gualcamayo (top of Middle Member)	3	*Lenodus variabilis* (Middle Ordovician)
Tambolar	San Juan (middle part)	4	*Oepikodus evae* (Early Ordovician)
Sassito River	Sassito (middle part)	3	*?Amorphognathus superbus* (Late Ordovician)
Sasso River	San Juan (top)	4	*Oepikodus evae* (Early Ordovician)
Rinconada	San Juan (top)	3	*Lenodus variabilis* (Middle Ordovician)
Cerro Blanco de las Cuevas	San Juan (top)	4	*Prioniodus elegans* (Early Ordovician)
Cordón del Peñasco	Alcaparrosa	5	Late Ordovician
Salagasta	San Juan (top)	3	*?Oepikodus evae* (Early Ordovician)
San Isidro	Empozada (middle part)	3	*Lenodus variabilis* (Middle Ordovician)

CAI due to infiltering of hydrocarbons or pyrite in the laminated conodont structure under reducing conditions or by enhaced thermal maturation related to the radioactive decay of Uranium. Nevertheless, this phenomenon hardly exceeds 0.5 CAI units and it is best observed at lower values (CAI 1.5–2) (Legall et al., 1981; Helsen, 1997). The influence of tectonics on CAI values is probably restricted to the variation of overburden levels, making it possible to use conodonts as geothermometers within orogenic belts (e.g., Rasmussen and Smith, 2001). However, under increasing conditions of regional dynamothermal metamorphism, where besides temperature, fluid pressure and oriented stress can play significant roles, recrystallization and ductile deformation of conodonts could be accelerated, in accordance with the metamorphism and ductile deformation experienced by the host rock (Teichmüller, 1987; Sudar and Kovács, 2006). A correlation between CAI and other common low-grade metamorphic indexes is shown in Fig. 2.

Despite conodonts can be found in diverse types of rocks (e.g., limestones, marbles, shales, cherts and sandstones), where possible, studied specimens come from unweathered calcareous

Table 2
Conodont CAI values from the Precordillera (from other authors).

Location	Formation	CAI
Cerro Viejo de San Roque	San Juan	2–3
Las Vacas Creek	San Juan	3–4
Gualilán	San Juan	3
Las Lajas Creek	San Juan	3
Los Potrerillos	San Juan	2–3
Los Berros	San Juan	2–3
Cerro del Fuerte	Los Espejos	1.5–2

Note: All CAI values are after Keller et al. (1993), except for Cerro del Fuerte, studied by Lehnert et al. (1999), not verified in our study.

Table 3
Published thermal data from the Precordillera.

Location			Formation/unit	Depositional age	Source	Estimated paleotemperature (°C)	Reference
Section	Lat.	Long.					
La Angostura Creek	−29.486	−68.659	La Flecha (top)	Sunwaptan, Late Cambrian	TAI 4	170–220	Sial et al. (2008)
Jáchal River	−30.205	−69.048	Yerba Loca	Middle-Late Ordovician	Metabasite petrography + thermodynamic modelling	250–350	Robinson et al. (2005)
Aguada de Los Azulejitos	−30.749	−68.834	La Chilca (base)	Late Llandovery – early Wenlock, early–middle Silurian	TAI 3.5	170–200	Pöthe de Baldis (1987)
Aguada de Los Azulejitos	−30.749	−68.834	Los Espejos (base)	Gorstian, late Silurian	TAI 3	96–120	Pöthe de Baldis (2000)
Cerro Viejo de Huaco	−30.177	−68.580	Los Azules (lower – middle members)	Middle–Late Ordovician	Clay mineral assemblages	50–80	Cingolani et al. (1997)
Cerro Viejo de Huaco	−30.177	−68.580	San Juan	Middle Ordovician	Clay mineral assemblages	50–80	Cingolani et al. (1997)
San Juan River	−31.254	−69.375	Don Polo	Ordovician (410–425 Ma metamorphic age)	Microfabric + IC + whole rock (phyllite) K–Ar dating	300–400	Buggish et al. (1994)
San Juan River	−31.254	−69.375	Alcaparrosa	Late Ordovician	Metabasite petrography + thermodynamic modelling	250–350	Robinson et al. (2005)
Sierra del Tontal	−31.272	−69.223	Alcaparrosa	Late Ordovician	Petrography + chlorite geothermometry	239–287	Rubinstein et al. (1998)
Sierra de Alcaparrosa	−31.332	−69.412	Alcaparrosa	Late Ordovician	Petrography + chlorite geothermometry	272–304	Rubinstein et al. (1998)
San Juan River	−31.373	−69.415	East of Calingasta	Ordovician (410–425 Ma metamorphic age)	Microfabric + IC + whole rock (phyllite) K–Ar dating	300–400	Buggish et al. (1994)
Cerro Bonilla area	−32.583	−69.207	Bonilla Group	Pre-Carboniferous (377 Ma metamorphic age)	White mica Ar–Ar dating	350–420	Davis et al. (1999)
East of Uspallata	−32.621	−69.337	Early Paleozoic	Cambrian–Ordovician (410-425 Ma metamorphic age)	Microfabric + IC + whole rock (phyllite) K–Ar dating	300–400	Buggish et al. (1994)
Cerro Bonilla area	−32.636	−69.210	Bonilla Group	Pre-Carboniferous (384 Ma metamorphic age)	White mica Ar–Ar dating	350–420	Davis et al. (1999)
Cerro Bonilla area	−32.654	−69.190	Bonilla Group	Cambrian – Early Ordovician (396–404 Ma metamorphic age)	Microfabric + IC + whole rock (phyllite) K–Ar dating	300–400	Buggish et al. (1994)
Cerro Bonilla area	−32.725	−69.195	Bonilla Group	Cambrian – Early Ordovician (365–405 Ma metamorphic age)	Petrography + felspar K–Ar dating	150–350	Varela (1973) (corrected by Linares (1977)); Etcheverría and Koukharsky (2002)

Notes: TAI, Thermal Alteration Index; IC, Illite crystallinity, Lat.–Long. values are approximate, after published papers.

rocks, in order to avoid any 'lithologic' influence. Apart from previously published data (Voldman and Albanesi, 2005, and references therein), 70 newly conodont samples (total rock weight, 250 kg), which correspond to about 70 localities and sections of the Precordillera, produced thousands of Cambrian to Silurian conodont elements (Fig. 3, Table 1). The studied conodonts were recovered following the standard techniques, as summarized by Stone (1987). CAI values were determined on the outer basal margins and finest sections of the elements by direct comparison with a standard conodont CAI set provided by A. Harris (USGS) (Epstein et al., 1977; Rejebian et al., 1987). Modal CAI sample values were transformed into temperatures according to the schemes proposed by Epstein et al. (1977) and Rejebian et al. (1987). Since the time factor diminishes its importance for burial durations larger than 50 Ma, we assigned the theoretical temperature interval to the determined CAI values (Table 1). The studied conodont specimens are housed in the Museo de Paleontología, Universidad Nacional de Córdoba, under repository code CORD-MP. Additionally, previously published paleothermal data from the Precordillera are incorporated for further information (Tables 2 and 3).

4. Conodont CAI values and the Andean Precordilleran structure

CAI values of the studied conodonts could differ from one locality or section to another. Nevertheless, the CAI variability within sampled levels was usually low (one CAI degree at the most). This fact and the generally good preservation shown by the microfossils suggest that our collections were not affected by contact metamorphism or hydrothermalism (Rejebian et al., 1987).

The geographic position of the studied localities was restored, assuming an homogeneous shortening of ~70% for the Central and Western Precordillera, between 29° 30′ and 31° 30′ S (Allmendinger et al., 1990; Cristallini and Ramos, 2000; Cardozo and Jordan, 2001). Both tectonostratigraphic domains constitute a westward thickening thrust stack, telescoped onto Neogene foreland deposits by an eastward tectonic transport. At 32° S, the shortening of the Precordillera reduces to 85 km (~55%), while at 33° 20′ S the Precordillera experienced less than 30 km of shortening, having the Andean deformation been concentrated in the Principal and Frontal Cordilleras (Ramos et al., 1996).

The interpreted depth of the décollement levels varies between 15 km at 30° 15′ S (70% of shortening; Allmendinger et al., 1990) to 5–6 km at 31° 30′ S (71% of shortening; Cristallini and Ramos, 2000) and 29° 30′ S (70% of shortening; Cardozo and Jordan, 2001). According to this premise, the Western and Central Precordillera display a near uniform regional cross-section structure, despite the spatial variability caused by ancient fabric controls along-strike thrust geometries (Fig. 4). In contrast, the Eastern Precordillera displays an opposite west-verging thick-skinned structure, characterized by reverse faults and fault-propagation folds, genetically linked to the basement uplifts of the Sierras Pampeanas (Zapata and Allmendinger, 1996a,b; Meigs et al., 2006; Vergés et al., 2007). On the basis of geological cross-sections and palinsp-

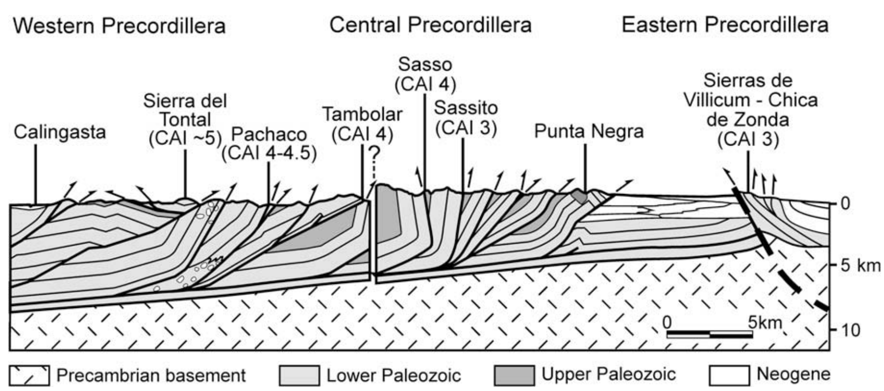

Fig. 4. Schematic structural cross-section of the Precordillera at ~31°30′S latitude showing the thin-skinned deformation of the Western and Central Precordillera and the thick-skinned deformation front at the Eastern Precordillera (modified after Cristallini and Ramos, 2000).

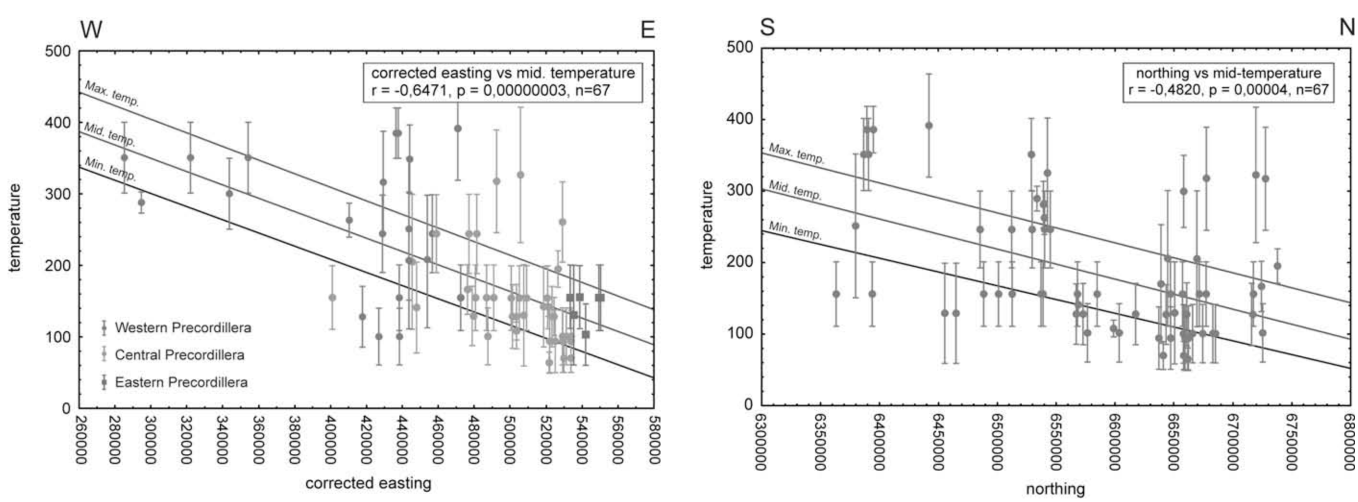

Fig. 5. Graphics of lineal regressions for recorded paleotemperatures versus distance along N–S and W–E transects in the Precordillera.

astic restorations, the Eastern Precordillera and the Bermejo basin accommodated ~17 km of shortening at 30° S (Zapata and Allmendinger, 1996b), and the Eastern Precordillera itself, 14 km at 31° 30′ S (Cristallini and Ramos, 2000) of Neogene deformation.

The maximum Andean contraction of the Precordillera developed where the basement was not involved in the compressive deformation, such as in the San Juan River section. Conversely, the shortening was considerably reduced where the Precordillera presents inverted Mesozoic rifts (Cortés et al., 2005). The spatially corrected distribution of CAI temperatures along with previously published paleothermometric data from the Precordillera indicates a gradual increase of paleotemperatures to the west and, in a less prominent way, to the south of the basin (Fig. 5).

5. Interpretation and discussion of the thermal data from the Precordillera

In contrast with the upper Paleozoic rocks from Western Precordillera, the pre-Carboniferous units display a complex deformation and a low-grade metamorphic overprint, as recognized in our field works and previous studies (Cucchi, 1971; Ramos et al., 1986; von Gosen, 1992; Buggish et al., 1994; Davis et al., 1999; Gerbi et al., 2002). Alvarez-Marron et al. (2006) recognized only one pre-Neogene deformation within the Paleozoic thrust stack at the Jáchal River section (30° S). This deformation, Devonian in age at most, is expressed by low-grade metamorphism, cleavage and kilometer-scale foldings to the west (Western Allochthon), and a low-angle basal Carboniferous unconformity to the east (Frontal Unit). In a first regional approach, Keller et al. (1993) recognized a progressive increase of paleotemperatures to the west of the basin, which they attributed to a westerly heat source; i.e., the Siluro-Devonian metamorphism. They preliminary noticed, on the basis of conodont CAI values, that the distribution of diagenesis, anchizone and epizone domains does not follow any morpho-structural subdivisions of the Precordillera. Our richer CAI data base clearly discards an abrupt transition between the unmetarmophosed parts of the foreland thrust belt of the Central Precordillera and the low-grade metamorphic internal parts from the Western Precordillera (Fig. 3).

According to Robinson et al. (2005), the mafic rocks from the Western Precordillera display a paragenetic association composed of prehnite, pumpellyte, actinolite, in the lack of zeolites/laumontite, which is indicative of the higher levels of sub-greenschists metamorphism (Table 3). The mentioned authors constrained the metamorphic conditions to ca. 250–350 °C and 2–3 kbar, with paleogeothermal gradients of ~30–35 °C/Km. Thereby, ruling out the possibility that the low-grade metamorphism affecting the Western Precordillera developed in an ocean-floor or extensional setting, with anomalously high geothermal gradients (diastathermal metamorphism). On the other hand, Voldman et al. (2008) observed at the Ancaucha Creek that the Ordovician mafic intrusions from Western Precordillera produced very restricted thermal anomalies, which could slightly contribute to a regional increment of the heat flux. A similar phenomenon was probably caused by the Miocene volcanic centres (e.g., Cerro Blanco, Cerro La Sal), which host carbonate xenoliths with Ordovician contact-metamorphosed conodonts of CAI 6; i.e., >500 °C. Additionally, there are no anomalous CAI values in our samples referable to frictional heating, which generally appears limited to a few cm from the fault planes (cf., d'Alessio et al., 2003). We have neither found evidences of heat transport by fluid migration along the fault planes (anomalous CAI values with indicative conodont textural alterations, cf., Rejebian et al., 1987), which is usually referred to the lower Permian San Rafaelic remagnetization (e.g., Rapalini and Astini, 2005, and references therein).

There is no manifest variation in CAI values in relation to their stratigraphic age in our samples (Ordovician to Silurian), suggesting that the lower Paleozoic stratigraphic overburden played a relatively small role in the heating of the Precordillera. The present-day heat flux of the Precordillera reflects cooler to normal geothermal gradients (43–60 mW/m^2, Ruiz and Introcaso, 2004), which are typical of mature passive margins and foreland basins (Allen and Allen, 2005). Therefore, a 30 °C/km geothermal gradient results plausible to calculate burial depths in the Precordillera, taking into account its long-lasting geotectonic evolution and its present Andean foreland position. The estimate is consistent with those from the Appalachian basin, where the CAI paleothermometric technique was originally calibrated by Epstein et al. (1977). According to the 30 °C/km geothermal gradient and the estimated paleotemperatures, overburden ranged from ~12 km to ~3.6 km in the Western and the Eastern Precordillera, respectively, for sampled levels. This is in accordance with Alonso et al. (2005) palinspastic reconstruction of the San Juan River section, who suggested that the uplift of the Western Precordillera corresponds with thrust faults detached at 10–15 km depth.

Tectonic overburden enhanced the thermal maturation of the Western Precordillera. The consistent and uninterrupted gradient of peak paleotemperatures that increases from east to west was undisrupted by the Andean thrusting. Hence, the great late Tertiary brittle contraction could not obliterate the initial distribution pattern of peak paleotemperatures resulting from the Chanic low-grade metamorphism. On the other hand, recent structural studies suggest that the Siluro-Devonian diastrophism had an east-vergent style of deformation, similar to the Andean one (Davis et al., 1999; Gerbi et al., 2002; Alonso et al., 2005; Alvarez-Marron et al., 2006). The pre-Carboniferous east-vergent style of deformation of the Western Precordillera is compatible with the consumption of a narrow ocean basin below the Chilenia terrane, with tectonic thickening at the leading edge (deep ocean basinal successions) of the Precordillera and emplacement of ophiolitic bodies within the suture zone (Ramos et al., 1986, 2000; Astini et al., 1995; Davis et al., 1999; Gerbi et al., 2002).

Conodont CAI values from the easternmost exposures of the Precordillera do not show anomalies referable to an Ocloyic westerly propagation of nappes or to the inception of a synorogenic clastic wedge within the Precordilleran foreland (present coordinates) (cf., Thomas and Astini, 2007). Nevertheless, in the eastern part of the terrane, the Sierra de Pie de Palo (Fig. 1) records top-to-the-west imbricate ductile thrusting along the Las Pirquitas thrust system, which places Grenville-age basement rocks over the metacarbonate-bearing Caucete Group (Ramos et al., 1998). These latter rocks are Cambrian–Ordovician? in age and isotopically similar to Precordilleran platform deposits and correlative carbonates of the Cerro Salinas, to the west of the Tulum Valley (Galindo et al., 2004, Naipauer et al., 2005). The erosion of these rocks ("phantom units") probably supplied detritus and metamorphosed conodonts to the deeper settings of the Western Precordillera, as pointed out by paleocurrent and provenance studies, reflecting the Ordovician collisional process (Voldman et al., 2009, and references therein).

Considering our hypothesis, conodont colors recorded in those transported microfossils suggest maximum burial temperatures of 300 °C for the Cambro-Ordovician strata of the Tulum region, so that all potential for oil and gas generation has been lost. Moreover, extensive faulting and deformation imply that seals were probably inadequate for retaining any lower Paleozoic hydrocarbon (cf., Zambrano and Suvires, 2008). A more favorable situation is found in the Huaco region, where CAI ranges in values between 1.5 and 3, and there are reports of liquid hydrocarbons filling fractures (Baldis and Beresi, 1990). In addition, CAI 3 values recorded in the Eastern and Central Precordillera indicate that the carbonate platform and its lower Paleozoic cover still have a thermal potential for gas production. The remaining CAI data indicate that a commercial accumulation of lower Paleozoic oil and gas beneath the Precordillera is rather unlikely.

Considering the paleogeographic position of the Precordillera in the Cuyania terrane, the NS-correlation of paleotemperatures (see Fig. 5b) was affected by a long history of oblique convergence and subduction beneath the South American plate, which resulted in crustal-scale partitioning deformation between strike slips and thrust motions (cf., Siame et al., 2006). Notwithstanding that, there is a smooth increment of peak paleotemperatures to the south. This pattern could be explained considering the exposure of deeper crustal levels at this sector of the Precordillera. Locally, recorded paleotemperature values from the northernmost Precordillera, in proximities to the Valle Fértil Lineament (Guandacol River area), could be reflecting the advance of the Ocloyic crustal stacking. The along-strike variation of the exhumed levels could reflect a stronger Chanic collision to the south of the Precordillera, or alternatively, a weaker mechanical response of its elastic lithosphere. In any case, there are no evidences of continuity of the Ocloyic thrustbelt to the southern Precordillera (although isolated outcrops of metamorphosed carbonates to the south of Mendoza River, west of Potrerillos, could be adscribed to it, which would reinforce the former hypothesis). According to von Gosen (1995), during the Neogene the internal parts of the southern Precordillera were transferred passively further eastward than the northern sectors along one important thrust fault zone that obliquely cuts through the orogen. However, given that thrusting alone can not tectonically exhume deepest crustal rocks (Platt, 1993), a climate influence (e.g., Ordovician and/or Carboniferous deglaciation) could not be overlooked in the exhumation processes of the Precordillera.

6. Conclusions

Thermal analysis has been conducted in the Precordillera of NW Argentina based on new collection of conodont CAI values in association with previously published paleothermal data. The distributional pattern of paleotemperatures from the Precordillera is assumed to increase to the west and south of the basin. Our conodont CAI collection statistically verifies this proposal. The trend of increasing paleotemperatures to the west reveals a gradual and continuous transition from diagenesis in the eastern sectors of the Precordillera to low-grade metamorphism in the internal parts of the foreland thrust and belt (Western Precordillera), without control of any tectono-stratigraphic subdivisions of the Precordillera (i.e., Eastern, Central or Western). The distribution pattern of paleotemperatures and the timing of its development indicate that stratigraphic burial was not a main control in the heating of the Precordillera. Moreover, elevated CAI values from the Western Precordillera are due to nappe stacking, which resulted from the Devonian collision of Chilenia, under normal geothermal gradients. Those geothermal gradients and the recorded paleotemperatures predict burial depths of ca. 12 km for the slope to rise deposits of the Western Precordillera to ca. 3.6 km for the shelf region in the Eastern Precordillera, providing constraints for the structural sections across the orogen. Alternatively, conodont CAI values from the Eastern Precordillera resulted unperturbed by the accretionary processes that affected the Gondwanan margin. The expected thermally altered conodonts from the Cuyania accretion are represented by metamorphosed conodont elements transported to the deeper settings of the basin. The smooth southward increase of paleotemperatures throughout the Precordillera may reflect stronger collisions to the south of the Precordillera, which exposed deeper crustal levels. However, a weaker mechanical response of its elastic lithosphere or climate influence should not be disregarded.

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