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Early Paleozoic accretionary orogenies in NW Argentina: Growth of West Gondwana

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ABSTRACT

Two Early Paleozoic orogenic cycles in the Gondwana margin of NW Argentina were driven by subduction, and interrupted by collision of continental ribbons of Laurentian affinity. Subduction started at the passive margin of Gondwana, possibly in response to the end of amalgamation of the supercontinent. The passive margin was underlain by a hyper-extended continent-ocean transition, and low-angle subduction gave rise to a wide and hot fore-arc region during the Pampean orogenic cycle. The arrival of the first continent ribbon ended the Pampean cycle, steepened the subduction, further heating the fore-arc and restarting arc magmatism, after a magmatic lull, 250–300 km trenchwards, initiating the Famatinian orogenic cycle. This cycle started with a 30 Ma-long period of extension and marine sedimentation, followed by the arrival of the second continent ribbon and inversion the back-arc, initiating a ~20 Ma-long shortening event culminating with the shutting down of the arc. This event gave rise to a 300 km-wide, low-topography, hot orogeny. Thus, the many peculiarities of the two Paleozoic orogens of NW Argentina result from a subduction history that efficiently transferred heat to a 300–400 km-wide belt of turbidite-dominated sediments, that was first part of the Pampean fore-arc and then part of the Famatinian retroarc, developed at the extended continental margin of Gondwana. The two orogenies with continued high heat flux lasted ~110 Ma, giving rise to two calc-alkaline arcs separated by a 300 km belt of high-T – low-P migmatites and peraluminous granites formed by anatexis of sediments deposited on the passive margin as well as those deposited in the wide Pampean fore-arc. These turbidite-dominated sequences were metamorphosed and melted to form a continental crystalline basement and accreted to the cratonic margin, adding 500 km of crystalline rocks to the margin of Gondwana, at the same time that a similar process was happening in eastern Australia along the same continental margin. Interestingly this wide accretionary orogen has now become the region where the Andean system developed a wide orogeny above flat-slab subduction.

1. Introduction

The Paleozoic Orogens of NW Argentina were part of the Terra Australis Orogen, TAO, along the southern margin of the Gondwana supercontinent and linked with the Cape Basin in South Africa and the Ross-Delamerian Orogens of eastern Australia (Fig. 1) (Cawood, 2005; Foden et al., 2006; Schwartz et al., 2008). The TAO resulted from a broad reorganization of plate movements that started at 570 Ma with the final amalgamation of Gondwana (Cawood and Buchan, 2007). This caused subduction initiation along the passive margins that had resulted from the break-up of Rodinia and the opening of the Pacific and Iapetus oceans. Subduction intensely reworked passive margin sediments and ultimately led to the accretion to the continent of large volumes of metamorphosed turbidites and their basement comprised of continent-oceanic transition or ocean floor (Collins, 2002b).

In the West Gondwana margin exposed in NW Argentina, subduction of the Iapetus ocean led to two orogenies during the Early Paleozoic resulting from the accretion of two continental ribbons of Laurentian affinity (Cawood, 2005; Dalla Salda et al., 1992a; Dalla Salda et al., 1992b; Dalla Salda et al., 1998; Ramos, 2004; Ramos et al., 1986). These two orogenic cycles took place in close succession where E-dipping subduction gave rise first to the Pampean orogenic cycle followed by the Famatinian cycle (e.g. Ramos et al., 1998; Rapela et al., 1998a). These cycles reworked a vast volume of turbidites with minor carbonates and volcanic rocks deposited between the Ediacaran and the Early Cambrian that form the Puncoviscana sequence. Reworking was driven by the development of two separate magmatic arcs, and a regional high-T– low P metamorphism, and deformation, leading to what

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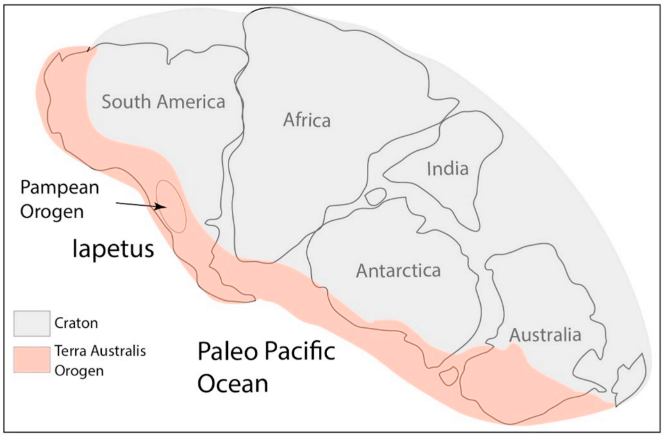

Fig. 1. The Terra Australis Orogen in the early Paleozoic, modified after Cawood (2005).

Lucassen and Becchio (2003) described as "monotonous gneisses and migmatites".

The regional prevalence in NW Argentina of variably metamorphosed turbidite sequences, similar to one another, has meant that investigations of detrital zircon ages have been instrumental in the understanding the tectonic and sedimentary evolution of the region. However, its importance has also lead to instances of over-interpretation, where simple explanations, such as changes in sediment source or issues with the statistical representativeness of the data (see Andersen, 2005) have been overlooked in favour of explanations involving significant tectonic changes.

This paper reviews the literature covering these two orogenic cycles in order to delineate the tectonic evolution of NW Argentina in the Early Paleozoic, focusing on the temporal evolution of sedimentation, metamorphism, magmatism and deformation. Here, we have avoided whenever possible using references to conference abstracts or theses, focusing on published articles instead. The paper first briefly describes the position of NW Argentina within Gondwana in the Paleozoic, and then defines the regional morphotectonic subdivision of the area necessary for contextualizing the literature. It then sets-up the regional geology, starting by a review and discussion regarding the nature of the all-important Puncoviscana sequence, followed by a summary of the Pampean and the Famatinian orogenic cycles. The main features and peculiarities of the orogens are then discussed, with particular focus on their paired magmatic belts and their intense, long-lived, high T – low P metamorphic history, and the impact of heat on the nature of the orogens. We finish with a comparison with the contemporaneous Delamerian-Lachlan Orogens in Australia, at the other end of the TAO, and with the Altai Orogen, part of the Central Asian Orogenic Belt.

2. Three morphotectonic subdivisions

NW Argentina can be divided into three main morphotectonics regions: the Puna Plateau, and the Western and Eastern Sierras Pampeanas (Figs. 2 and 5). The plateau is a geomorphological feature with internal subdivisions based on geology and geomorphology. The Sierras Pampeanas are a series of N-S trending mountains uplifted by Tertiary Andean reverse faults and their division is based on significant geological differences between the west and the east.

2.1. Puna Plateau

The Plateau has been divided into two geologically distinct blocks knowns in the literature as the Western and Eastern Eruptive Belts, or the Faja Eruptiva Occidental and Oriental, respectively (Fig. 2). Here we group under this morphotectonic unit, the series of mountains that form the Cordillera Oriental. These mountains lie at the eastern edge of the Plateau but are grouped here with the Plateau because of similarities in their geological evolution and geographic position (Fig. 2). The term "eruptive" used in the literature to describe these belts is somewhat misleading because both eruptive belts are dominated by plutonic rocks. However, we maintain the original terminology here in order to maintain a direct link with the literature. We will generally refer to them in their abbreviated form: WEB, EEB, and CO, respectively.

The exposed basement in the Plateau (Fig. 3) comprises mainly Late Ediacaran to Early Paleozoic sedimentary rocks of the Puncoviscana sequence, overlain by Late Cambrian to Ordovician marine sedimentary basins. The main calc-alkaline Famatinian arc trends NW through the Plateau and defines the WEB (Bahlburg, 1998; Coira et al., 1999; Pankhurst et al., 1998). The EEB is a subparallel magmatic belt, 400 km long from ~ 26°S to 22°S (Méndez et al., 1973), of dominantly peraluminous granitic rocks and migmatites. The CO to the east, is a region where high-T – low P variably metamorphosed rocks of these sedimentary sequences are intruded by Early Cambrian granitic batholiths of the Pampean arc in the north, and by Early Ordovician plutons derived from crustal anatexis related to the Famatinian Orogen in the south (Fig. 4) (e.g. Büttner et al., 2005; Einhorn et al., 2015; Hauser et al., 2011; Hongn et al., 2014; Pearson et al., 2012; Sola et al., 2013).

2.2. Western and Eastern Sierras Pampeanas: Laurentia in the west, Gondwana in the east

The regional subdivision into Western and Eastern Sierras Pampeanas (WSP and ESP, respectively) was first introduced by Caminos (1979). Ramos et al. (1986) suggested that the Precordillera, part of the Western Sierras Pampeanas, was a "suspect" terrane, and later Dalla Salda et al. (1992a, 1992b and 1998) argued that these terranes were derived from Laurentia. The position of the boundary between them varies in the literature. We follow Rapela et al. (2016) who divided the two based on terrane origin: the Western Sierras Pampeanas (WSP) comprising rocks with Paleoproterozoic and Grenvillean zircon ages and geological affinities with Laurentia; and the Eastern Sierras Pampeanas (ESP) dominated by (meta)sedimentary rocks with detrital zircon ages similar to those found in Gondwana.

The WSP forms a narrow band between the ESP and the Precordillera/Cuyania terrane, an exotic block later accreted to the WSP (Ramos, 2004). The WSP comprises the Sierra de Pie de Palo, and the Sierras de Maz-Espinal-Umango and Sierra de Toro Negro (Fig. 5), and is thought to connect northwards with the Arequipa–Antofalla and the Rio Apa blocks, to form a Paleoproterozoic block, known as the MARA block (Casquet et al., 2005; Casquest et al., 2006; Rapela et al., 2007; Rapela et al., 2016).

The rocks with Laurentian affinities (Dalla Salda et al., 1992a, 1992b and 1998), that comprise the WSP record a prolonged magmatic evolution during the Grenvillean period (1330 to 1030 Ma), including the intrusion of anorthosite–mangerite–charnockite–granite (AMCG) complexes, followed by the intrusion of A-type granites at ~840 and 770 Ma, and carbonatite complexes at ca. 570 Ma (Rapela et al., 2016). The sedimentary rocks associated with this basement have detrital zircon ages indicative of Grenvillean sources (Fig. 6), with a major age peak between 1330 and 1030 Ma, and a minor peak at 1900–1800 Ma (see also Casquet et al., 2008; Naipauer et al., 2010; Varela et al., 2011; Vujovich et al., 2004). This contrasts with detrital zircon age groups in (meta-)sedimentary rocks of the Eastern Sierras Pampeanas with peaks at ~ 1.0 Ga and 0.6 Ga, common throughout Gondwana (Fig. 6) (e.g. Miller et al., 2011). The WSP has also been referred to as the Pampean terrane, a term related to an earlier interpretation that it was a para-utochtonous ribbon of Gondwana, pulled away from and returned to the continental margin (Forsythe et al., 1993; Ramos et al., 2010; Rapela et al., 1998a). Some authors, such as Otamendi (pers. commun. 2018), remain cautious regarding the Laurentian affinities of the WSP.

Mulcahy et al. (2011 and 2014) determined that the rocks with

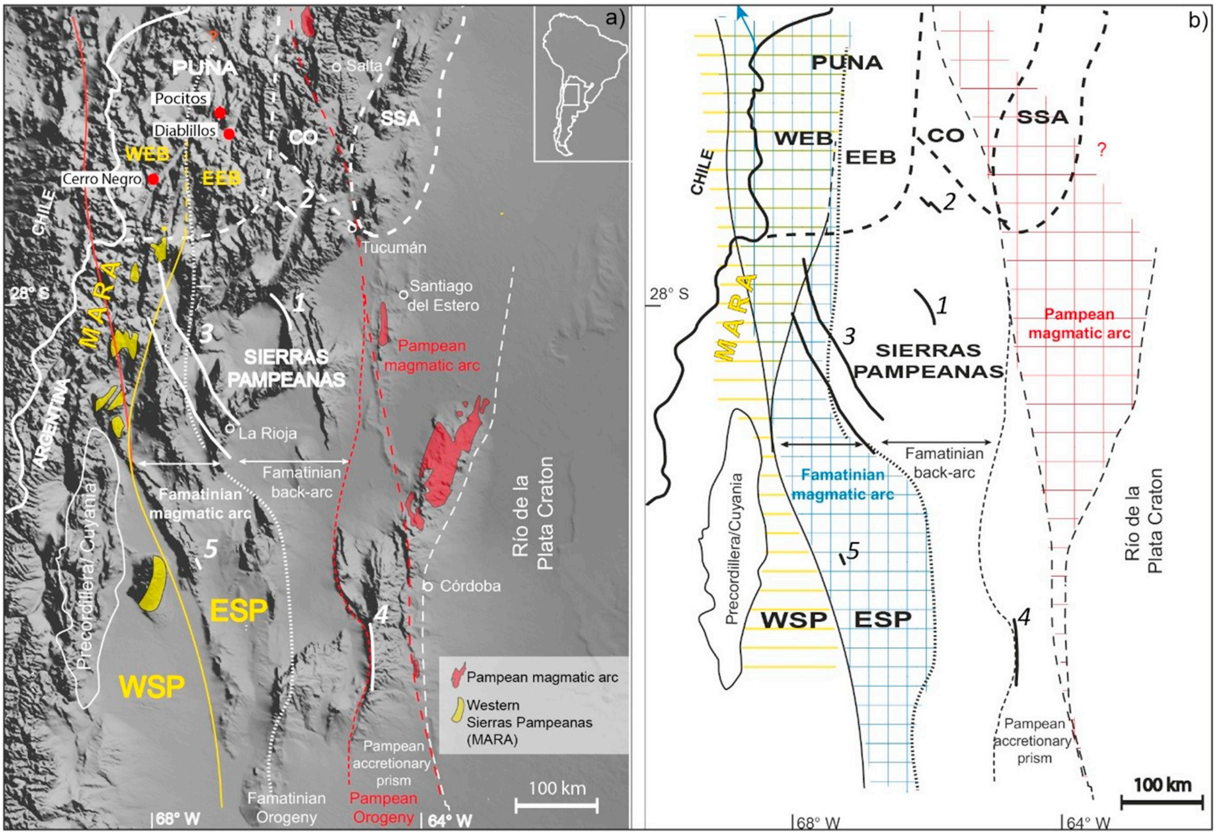

Fig. 2. a) DEM image indicating subdivision of NW Argentina into the Puna Plateau, including the Western and Eastern Eruptive Belts (WEB and EEB, respectively) and Cordillera Oriental (CO), and into the Western and Eastern Sierras Pampeanas (WSP and ESP, respectively). We have included the Cordillera Oriental as part of the broader Puna because of geological similarity. The Pampean arc rocks are in red, and outcrops of the Western Sierras Pampeanas with Laurentian affinity are in yellow. Note that there are no outcrops of rocks with Laurentian affinity in the Puna. The Famatinian arc is west of the Famatinian back-arc defining a broad N-S band. Also indicated are outcrop positions and names of localitie mentioned in the text. b) Schematic regional morphotectonic divisions. Note that the Famatinian arc crosses from the Eastern to the Western Sierras Pampeanas in the north. SSB is System Santa Bárbara; MARA is a Paleoproterozoic exotic block that includes the Western Sierras Pampeanas (WSP), as well as the Arequipa–Antofalla and the Rio Apa blocks to the north not shown in figure (Rapela et al., 2007). Numbered lines in a) and b) correspond to shear zones mentioned in the text: 1-La Chilca, 2- Pichao, 3- TIPA, 4-Guacha Corral, and 5- Arenosa Shear Zone. Inspired by figure in Dahlquist et al. (2016).

Fig. 3. Recumbent Famatinian-age folds in the Puna Plateau. This mountain exposes a sequence of metamorphic rocks including a thick amphibolite layer. In the foreground a Quaternary basalt lava flow some 20 m in height.

Laurentian affinities in the Sierra de Pie de Palo underwent anatexis and major crustal shortening during the Famatinian Orogeny (see also Casquet et al., 2012b). This implies that this block had already been accreted to the margin the ESP and both deformed together at the margin of Gondwana. This shortening event was a result of the docking of the Precordillera/Cuyania terrane to the Gondwana margin. This is the second block with Laurentian affinities and Grenvillean ages (Dalla Salda et al., 1992a and b; Naipauer et al., 2010; Ramos, 2004; Ramos et al., 1986; Ramos et al., 2010; Rapela et al., 2016; Thomas and Astini, 1996; van Staal and Hatcher, 2010; van Staal et al., 2011).

The Precordillera/Cuyania terrane was defined using biostratigraphy, geochronology and other lines of evidence (see review in Ramos, 2004). This terrane is essentially comprised of Cambrian–Ordovician platformal carbonates with Laurentian fossils. There is no exposed basement, except for the 1244 ± 42 Ma tonalite-trondhjemite pluton described by Sato et al. (2004). More recently, Rapela et al. (2016) suggested that a metamorphic mafic-ultramafic complex with Grenvillean ages, outcropping to the east of the Precordillera, could represent the basement to the limestones (see also Rapela et al., 2010). The Precordillera/Cuyania terrane records Famatinian deformation and metamorphism (Sato et al., 2004). Because of anatexis during the Famatinian cycle, Mulcahy et al. (2014) concluded that the Sierra de Pie de Palo, was part of MARA and could not be the basement to rocks in the Precordillera/Cuyania terrane, as had previously been interpreted (e.g. Vujovich et al., 2004). Galindo et al. (2004) reached the same conclusion using carbonate isotope chemistry instead, but interpreted the Sierra de Pie de Palo to be (para-)autochtonous to the South American continent.

The Eastern Sierras Pampeanas, in common with the Puna Plateau, comprises Late Ediacaran to Early Paleozoic sedimentary rocks of the Puncoviscana sequence. The basement to the Puncoviscana sequence is not exposed in either the Puna or the Sierras Pampeanas. Metamorphism tends to be dominantly high-T – low-P (Buchan- or Barrovian-type) (e.g. Finch et al., 2017; Larrovere et al., 2011; Lucassen

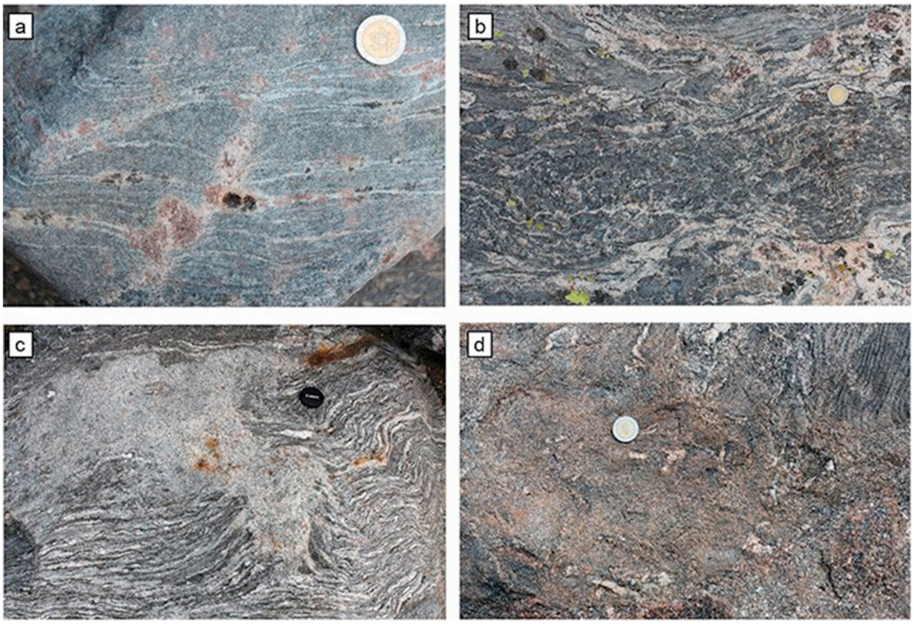

Fig. 4. Migmatites and granites from Sierra de Quilmes (see Fig. 5 for location). a) Granulite-facies paragneiss migmatite with foliation-parallel leucosomes with small garnet grains and orthopyroxene, connected to a high-angle leucosome containing large (> 2 cm) poikiloblastic garnet; b) residual migmatite with a central layer rich in cordierite and only small volumes of leucosome, surrounded by paragneisses with voluminous leucosomes and garnet. The cordieritite in the centre is interpreted to represent residue left behind by melt extraction; c) stromatic migmatite where layer-parallel leucosomes link seamlessly to a grey granite. This is interpreted to record the process of magma extraction and granite generation; d) anatectic granite with a raft of stromatic migmatite in the upper right. Granite has elongated white blocks of leucosome rimmed by melanosome indicating that solidified stromatic migmatites were broken up and included in the granitic magma.

and Becchio, 2003; Lucassen et al., 2011; Sola et al., 2017) and at high-grade these are gneisses, schists, migmatites, and less commonly, marbles and metabasites (e.g. Hongn and Riller, 2007). Above the Puncoviscana sequence, and separated by an angular unconformity, there are a number of deformed and metamorphosed younger Cambrian and Ordovician shallow marine basins (e.g. Astini, 2008; Bahlburg, 1998; Cisterna et al., 2017; Zimmermann and Bahlburg, 2003).

The Puncoviscana sequence, the effective regional basement, has two well-defined detrital zircon ages: a Late Neoproterozoic – Early Cambrian group (680–570 Ma), not present in the Laurentian rocks of the Western Sierras Pampeanas (WSP), and a Mesoproterozoic peak between 1150 and 850 Ma, concentrated between 1080 and 970 Ma, that partially overlaps with the range of ages in the WSP (Rapela et al., 2016). A third, Paleo- to Mesoproterozoic peak (2400–1400 Ma) is occasionally present (Adams et al., 2011; Aparicio González et al., 2014; Escayola et al., 2011; Hauser et al., 2011; Ježek et al., 1985; Miller et al., 2011; Schwartz and Gromet, 2004). In many samples, a young peak of Pampean ages (550-510 Ma) can also be recognized (Adams et al., 2011; Adams et al., 2008; Rapela et al., 2016; Sims et al., 1998), which indicates that some of the sediments included in this sequence formed contemporaneously with and as a result of erosion of the Pampean arc. These detrital zircon ages place the sediment sources in Gondwana (e.g. Ježek et al., 1985; Rapela et al., 2016), possibly the Sunsás Belt of western Brazil (e.g. Adams et al., 2008; Einhorn et al., 2015), or the Dom Feliciano Belt of southern Brazil and Uruguay, or the Natal–Namaqua belt and the East African orogen (Rapela et al., 2016).

The Puncoviscana sequence records a general southward increase in metamorphic grade (Fig. 1 in Lucassen et al., 2011), from low-grade close to the border of Bolivia in the north (Coira et al., 1999), to high-grade, granulite facies metamorphism south of Salta (Fig. 5) (Büttner et al., 2005; Escayola et al., 2011; Moya, 2015; Zimmermann, 2005). Some high-temperature rocks have been metamorphosed at pressures equivalent to 25–30 km, exposed in deep sections of the Famatinian magmatic arc in the Sierras Pampeanas (Cristofolini et al., 2012; Ducea et al., 2010; Tibaldi et al., 2013) whereas others melted at shallow crustal levels (3 kb, Sola et al., 2017). There is also a southward increase in the intensity of Late Ordovician folding (Bahlburg, 1998) and a deepening of exposures of Ordovician volcano-sedimentary sequences (Cristofolini et al., 2014; Otamendi et al., 2017). Deeper exposures in the south crop out in the area where the Nazca plate today subducts at a low angle under South America (Gilbert et al., 2006), and the southern end of the Sierras Pampeanas coincides with the southern end of the current flat subduction. This suggests the possibility that there has been more intense uplift and erosion in the south.

The nature and origin of the Puncoviscana sequence is fundamental for understanding the Paleozoic evolution of NW Argentina, not only because it comprises the regional effective basement but also because it is the main country rocks to both the Pampean and Famatinian magmatic arcs, and the most likely source of the abundant crustal-derived magmas. Crucially, its current definition and problematic regional correlations hamper understanding of the tectonic evolution of NW Argentina.

3. Puncoviscana sequence: Ediacaran to Cambrian sedimentation

3.1. Definition

Several terms are used to define the Puncoviscana sequence in the literature, such as Puncoviscana Group, Basin, Complex or "Formation senso lato" (Aceñolaza and Aceñolaza, 2007). The Puncoviscana sequence was first defined by Turner (1960), and extends over 800 km north-south and 150 km east-west (Ramos, 2008) and is at least 3000 m thick (Aceñolaza and Aceñolaza, 2007; Adams et al., 2011), with no exposed base or basement. These sediments were derived from the east and their nature has been widely discussed (e.g. Aceñolaza and Toselli, 2009; Aceñolaza and Aceñolaza, 2007; Ježek et al., 1985; Omarini et al., 1999; Zimmermann, 2005). According to Aceñolaza and Aceñolaza (2007) the Puncoviscana sequence formed at a continental margin "consisting mostly of shelf, shelf-edge and slope deposits, with a general deepening trend towards the west" (see Zimmermann, 2005, for a different view). Like many others, Aceñolaza and Aceñolaza (2007) considered the sediment sources to be the cratonic parts of southwest Gondwana in the east. The sediments consist of (Omarini et al., 1999): "a) monotonous outcrops of coarse-grained turbiditic sandstones of flysch type, b) thick and monotonous strata of argillites, siltstones and sandstones, c) minor massive strata of diamictites and polymictic conglomerates, d) isolated but thick massive strata of shallow-water micritic limestones."

High-grade meta-turbiditic rocks in much of the Sierras Pampeanas have the same detrital zircons as low-grade Puncoviscana sequence rocks, and a number of preserved primary sedimentary features that support the interpretation that they are part of the Puncoviscana

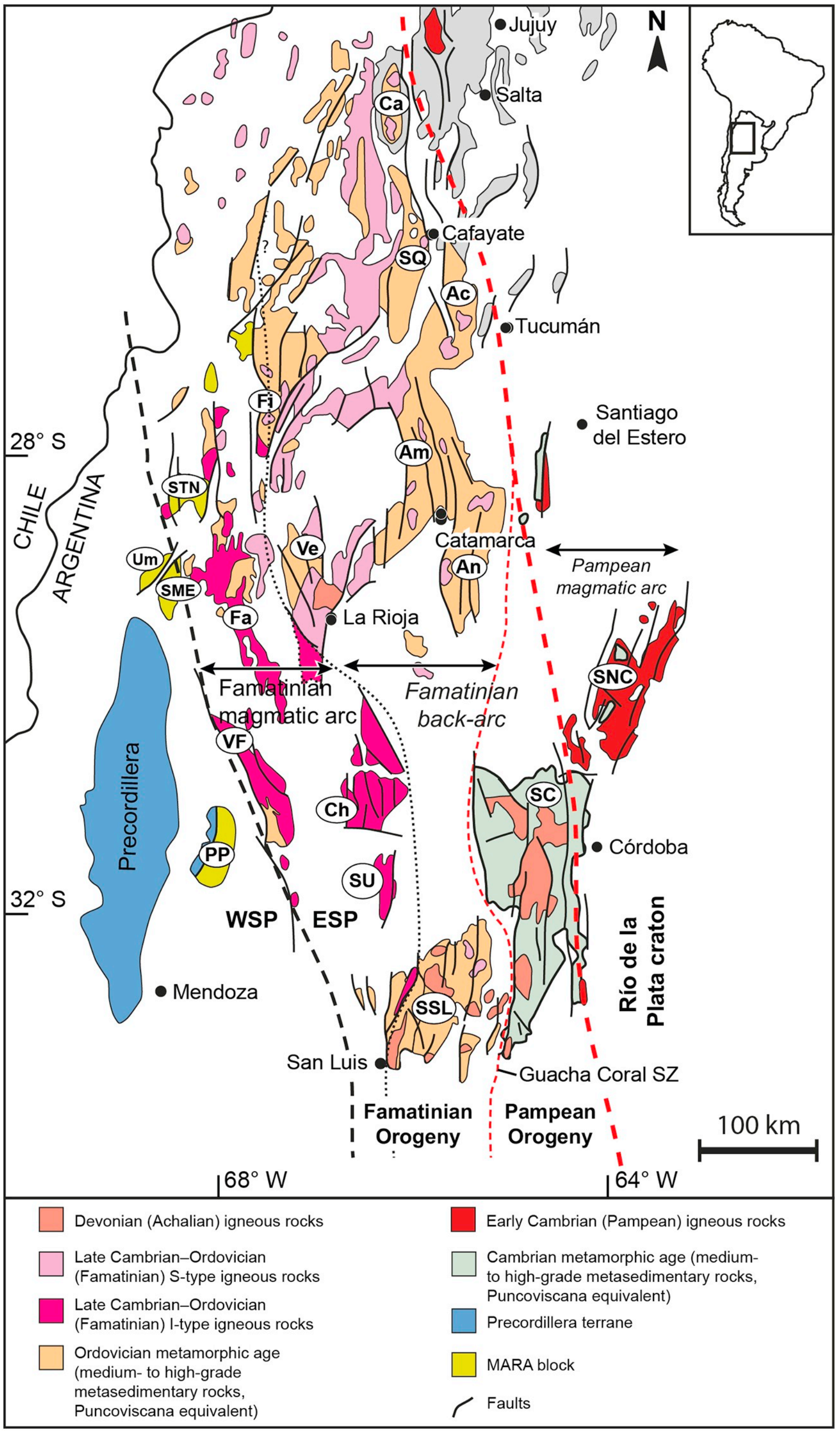

Fig. 5. Schematic geological map of NW of Argentina including the main mountain ranges mentioned in the text and geological subdivisions. Modified after Dahlquist et al. (2016), Rapela et al. (2016), Steenken et al. (2006) and Semenov and Weinberg (2017). Mountain range abbreviations: Sierra de Aconquija (Ac), Sierra de Fiambala (Fi), Sierra de Ambato (Am), Sierra de Ancasti (An), Sierra de Cachi (Ca), Sierra de Velasco (Ve), Sierra de Famatina (Fa), Sierra de Maz-Espinal (SME),Sierra Norte de Córdoba (SNC), Sierra de Valle Fértil (VF), Sierra de Chepes (Ch), Sierra de Ulapes (SU), Sierra de Pie de Palo (PP), Sierras de Córdoba (SC), Sierra de Quilmes (SQ), Sierra Norte de Córdoba (SNC), Sierra de San Luis (SSL), Sierra de Toro Negro (STN), Sierra de Umango (Um).

sequence (Fig. 7) (Aceñolaza and Miller, 1982; Franz and Lucassen, 2001; Willner et al., 1987). This is an important basis for regional interpretations and is reflected in most regional geological maps (e.g. Zimmermann, 2005). However, Collo and Astini (2008) noted that the turbidite deposits of Late Neoproterozoic to Cambrian age that have traditionally been brought together under the term Puncoviscana sequence, include different deposits with similar features. These similarities impede differentiating between them based on sedimentological, post-depositional or geochemical features (see also Buatois and Mángano, 2003; Mángano and Buatois, 2004; Mon and Hongn, 1991). This is further hampered by metamorphism and deformation, including isoclinal folding and a long history of thrusting.

Determining the range of deposition ages of the Puncoviscana sequence is complicated by the vast volume of sediments and relatively few fossils, and by the natural limits of detrital zircon ages (e.g. Miller et al., 2011; Rapela et al., 2016). Sediments were deposited between > 600 and 520 Ma (Late Neoproterozoic to Cambrian) (Drobe et al., 2009; Lork et al., 1991; Omarini et al., 1999; Rapela et al., 1998b; Schwartz and Gromet, 2004; Sims et al., 1998). Its maximum age is unconstrained because, as mentioned, its base does not crop out (Ježek et al., 1985), and could be as old as 650 Ma (Aceñolaza and Aceñolaza, 2007; Omarini et al., 1999). This long history of deposition is a point of contention, and there are questions whether these rocks are predominantly Neoproterozoic (Omarini et al., 1999) or Early Cambrian

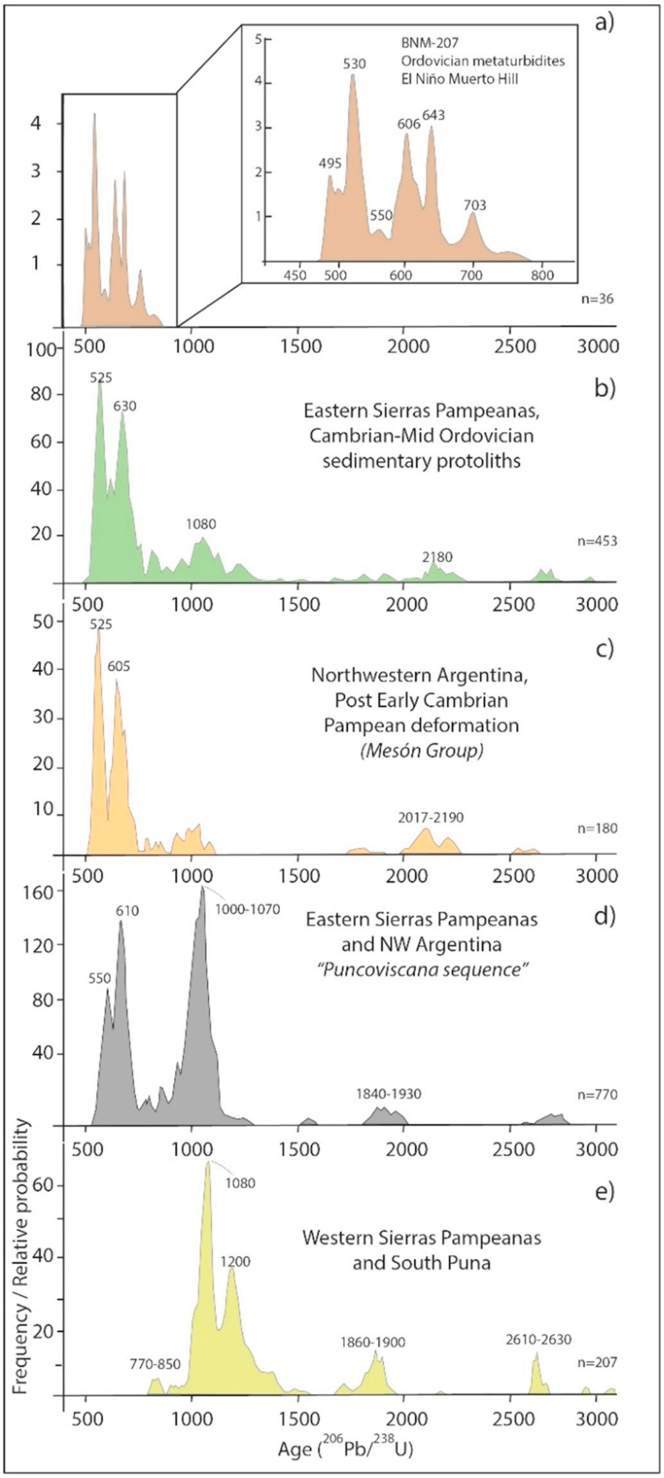

Fig. 6. Typical patterns of detrital zircon U-Pb ages in different sedimentary packages in NW Argentina. a) sample BNM-207 from Hauser et al. (2011); b-e) from Rapela et al. (2016). Note differences between typical Puncoviscana sequence in (d) and the Western Sierras Pampeanas (e), indicative of its Laurentian affinities. The latter lacks the well-defined Brasiliano peak centred at around 600 Ma so characteristic of the Puncoviscana sequence, and it has a peak at 1200 Ma with a tail towards 1300 Ma. Ordovician sediments in (a) include Famatinian-age detritic zircons (see also Fig. 4 in Einhorn et al., 2015; and Fig. 4 in Collo et al., 2009).

(Durand and Aceñolaza, 1990; Escayola et al., 2011).

Aceñolaza and Toselli (2009) found the occurrence of Ediacaran soft-bodied fossils in the older levels of the sequence in the north, whereas to the south other fossils indicate Early Cambrian sedimentation ages. In some localities there is ~30 Ma gap between the youngest detrital zircons and deposition age, including the type locality for the Puncoviscana sequence (Escayola et al., 2011). In most localities, detrital zircons of Pampean age have been found in Puncoviscana sedimentary rocks (Drobe et al., 2009; Hauser et al., 2011; Lork et al., 1991; Pearson et al., 2012; Rapela et al., 2016) suggesting that sediments were recycled from the contemporaneous Pampean volcanic arc. The timing of the end of Puncoviscana sedimentation is significant because, as we will see below, it marks the transition from the Pampean and Famatinian orogenic cycles. The youngest zircon ages marking the maximum deposition age of Puncoviscana sediments vary from 520 Ma (Adams et al., 2011), to 517 Ma (Aparicio González et al., 2014) or even as young as 509 Ma (an age peak defined by four concordant analyses in Pearson et al., 2012).

A number of authors equate the duration of deposition of the Puncoviscana sequence with that of the Pampean Orogeny (Einhorn et al., 2015; Hauser et al., 2011). Escayola et al. (2011) proposed that deposition started coevally with the Pampean arc volcanism at ~540 Ma and ended before the deposition of the overlying, unconformable Cambrian Mesón Group, which has a maximum deposition age of 513 ± 2 Ma, the concordia age defined by the four youngest zircon ages reported in Aparicio González et al. (2014). This age coincides with the youngest zircons in the Puncoviscana sequence and does not allow for a clear distinction.

Interlayered felsic volcanic rocks in the type locality of the Puncoviscana sequence, close to Bolivia, yielded a zircon age of 537 ± 1 Ma (Escayola et al., 2011). Nearby, a 523.7 ± 0.8 Ma pluton (TIMS zircon U-Pb age) intruded the Puncoviscana sequence after it was folded, suggesting that deformation occurred between 537 and 524 Ma (Escayola et al., 2011) and also that Pampean magmatism and sediment deposition were contemporaneous. We will return to these points in the next subsection.

Paleocurrent directions in the Puncoviscana sedimentary rocks indicate sediment provenance from the east or southeast, where the Río de la Plata Craton currently lies (Rapela et al., 2007). However, these sedimentary rocks lack detrital Paleoproterozoic (2.3–2.0 Ga) zircons dominant in the Craton. Rapela et al. (2016) argued therefore that the Puncoviscana sequence and the Craton were brought side-by-side by dextral lateral movement during the Pampean cycle (see also Miller et al., 2011; Rapela et al., 2007; Rapela et al., 2016). We note however that while many samples do lack Paleoproterozoic detrital zircons, they are occasionally present, such as in all Puncoviscana samples investigated by Adams et al. (2008).

3.2. Tectonic setting of deposition

There is a lack of consensus regarding the tectonic setting of deposition of the Puncoviscana sequence (Quenardelle and Ramos, 1999). This is most likely a result of unresolved internal complexities of the sedimentary sequence, added to the difficulty of separating and correlating turbidite sequences (Collo and Astini, 2008) and a lack of crystalline basement exposure. The depositional setting has been interpreted to be: a) a passive margin (Aceñolaza et al., 2002; Adams et al., 2011; Ježek et al., 1985; Rapela et al., 1998b; Schwartz et al., 2008), b) a rift between the margin of Gondwana and an outboard terrane (Omarini et al., 1999; Ramos, 2008), c) a foreland basin (Keppie and Bahlburg, 1999; Kraemer et al., 1995), or d) a fore-arc basin (Einhorn et al., 2015; Hauser et al., 2011; Rapela et al., 2007), evolving into a foreland basin during collision with an exotic block (Pampia and the Antofalla terranes in Escayola et al., 2011).

Although these sediments could have initially been deposited over a continental crust (Lucassen and Franz, 2005), extension may have led to formation of an ocean floor, at least in the southern part of the Sierras Pampeanas (Baldo et al., 1996; Rapela et al., 1998b), where the basin is deeper. Rapela et al. (1998b) noted that amphibolites

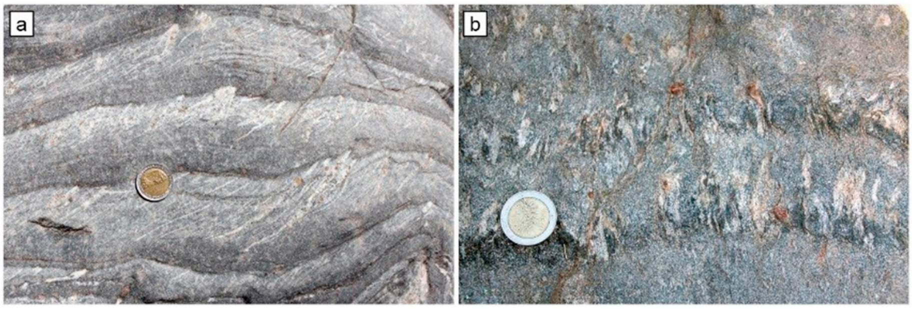

Fig. 7. Graded bedding preserved in Puncoviscana turbidites that have undergone anatexis. The darker layers are richer in Bt and represent pelitic layers that grade to lighter grey psammitic sections. In both cases, graded bedding indicates younging downwards. In a) small veinlets of leucosome are oriented at an angle to subhorizontal layering and are interpreted to represent melt extracted from the darker, pelitic layers. In b) leucosomes are preferentially preserved in the more fertile pelitic layers and are oriented vertically, parallel to an axial planar foliation at high-angle to primary layering (horizontal). Note centimetric garnet in pelitic layers in (b).

interbedded with Puncoviscana sedimentary rocks, have MORB signatures, suggesting that sedimentation was associated with rifting to form an oceanic crust. Omarini et al. (1999) using geochemical data of volcanic rocks that included alkaline rocks within the Puncoviscana sequence proposed a geodynamic model reflecting a progressive opening of a continental rift followed by its closure, reflected by volcanic arc magmatism. This evolution was challenged by regional considerations (Franz and Lucassen, 2001), and by geochronology that established that the volcanic rocks were Ordovician and the alkaline dykes were much younger, yielding Mesozoic ages (Hauser et al., 2010).

Escayola et al. (2011) proposed an alternative depositional setting for the Puncoviscana sediments. They built on the overlap between the time of sediment deposition, deformation and magmatic intrusion at around 540–530 Ma to argue that the Puncoviscana sediments were deposited in the fore-arc and/or trench of the west-facing Pampean arc before ~530 Ma. This was followed by *syn*-collision foreland sedimentation when a continental ribbon, originally rifted from Laurentia (the MARA block), arrived at the trench at ~530 Ma.

This kind of unresolved complexity of the Puncoviscana sequence is best revealed by the exposures in the north, close to the city of Salta (Aparicio González et al., 2014; Moya, 1998). Here, the Puncoviscana sequence comprises different formations with variable deformation intensities separated by an internal unconformity (Aparicio González et al., 2014; Moya, 1998). Have they all been deposited in the same basin and similar tectonic setting? Is part of the sequence pre- and part *syn*-orogenic?

Collo et al. (2009) argued for the need to separate pre- and *syn*-orogenic turbidites. They used detrital zircons and found that the turbidite sequences containing Cambrian zircons from the Pampean arc lie unconformably above turbidites of the Puncoviscana sequence that lack these zircons. They argued that the younger turbidites should not be included in the older, pre-Pampean Puncoviscana sequence (their Fig. 5, see also Rapela et al., 2016). As argued in the next section, we agree with the need to separate the two groups of turbidites but note that detrital zircons may not be a sufficient tool.

3.3. Terminology issues

As is clear from the above summary, the term Puncoviscana sequence brings together Ediacaran and Cambrian (up to ~ 520–510 Ma) sedimentary rocks that could not have been deposited in the same basin. This is because the sequence includes sediments deposited in at least two distinct tectonic settings: before initiation of the Pampean arc magmatism, and during arc activity. The earliest sediments were likely deposited in a continental passive margin basin (e.g. Ježek et al., 1985; Omarini et al., 1999; Rapela et al., 1998b). Later sediments are *syn*-orogenic and recycle Pampean arc zircons. These were likely deposited initially in the fore-arc basin of the Pampean arc, which became a foreland basin after the arrival of the MARA continental ribbon towards the end of the Pampean cycle (e.g. Collo et al., 2009; Einhorn et al., 2015; Escayola et al., 2011; Hauser et al., 2011; Omarini et al., 1999).

While there are inherent difficulties in telling apart variably deformed meta-turbiditic sequences, the usage of the term Puncoviscana sequence masks the distribution of rocks deposited in the different basins and underlies the difficulties in understanding regional tectonics. The Puncoviscana sequence should therefore be treated as two (or more) separate sequences, as previously argued (Aparicio González et al., 2014; Collo et al., 2009; Escayola et al., 2011; Hongn et al., 2014; Moya, 1998). In this paper we will use the informal term *late* Puncoviscana sequence when referring to the *syn*-orogenic turbidites deposited during the Pampean cycle.

3.4. Nature of the basement: non-volcanic hyper-extended continental margins or MARA

As previously mentioned, the lack of exposed basement to the Puncoviscana sequence impacts on interpretations of the depositional setting and the role of basement on subsequent tectonic development. Xenoliths found in Cretaceous basanite magmas in the Salta Rift (Lucassen et al., 1999) show a lower crust of dominantly granitic composition, with only rare mafic xenoliths. The granitic xenoliths have identical Nd and Sr isotopic composition to rocks of the Puncoviscana sequence, and the scarcity of mafic xenoliths (< 5% of the xenolith population) suggests the absence of a thick lower crust (Lucassen et al., 1999). This conclusion is supported by seismic studies: low seismic wave velocities in the lower crust (Wigger et al., 1994) indicate a missing lower crust of intermediate or mafic composition. Thus, a silica-rich crust of granitic composition appears to be in direct contact with the ~35–38 km-deep Moho across a well-defined seismic signal indicative of a sharp density contrast (e.g. Lucassen et al., 1999; Perarnau et al., 2012).

The passive margin of Gondwana in NW Argentina possibly comprised a transitional crust between the Proterozoic continental crust in the east and an oceanic crust in the west (e.g. Otamendi et al., 2012; Piñán-Llamas and Simpson, 2006). We suggest that this transitional crust was a non-volcanic hyper-extended continental margins (Manatschal, 2004; Manatschal and Bernoulli, 1999). This type of margin is characterized by a thinned continental upper crust and uplifted sub-continental lithospheric mantle, lacking a mafic lower crust (Lavier and Manatschal, 2006). Mafic lower crust would be expected if the crust was either an extended continental crust overprinted by extensional volcanism or an oceanic crust. In hyper-extended margins, the mafic lower crust is intensely thinned, with remnants of continental upper crust overlying an uplifted lithospheric mantle. The high density of this basement buried under the Puncoviscana sequence would help prevent it from being exposed, and would maintain a low topography (O'Halloran and Rey, 1999), thus explaining the dominance of marine sedimentation throughout both the Pampean and Famatinian orogenic cycles (see Table 1).

Non-volcanic hyper-extended continental margins are typically 200 km wide, and are ultimately split when the continents separate, with most of the transitional zone attached to one of the continents. Given that the Puncoviscana sequence covers a width of > 300 km, a hyper-extended continental margin would only comprise part of the basement, the rest probably transitioning to an oceanic crust in the west

Table 1
Evolution of NW Argentina during the Pampean and Famatinian Orogenies (see schematic representation in Fig. 8.

Time scale	Orogeny: tectonic setting	Magmatism and metamorphism	Sedimentation
Ordovician – Late (440–460)	**Famatinian – Oclóyic Phase** • Hot orogen: low topography, 300 km-wide, anomalously wide mylonitic thrusts control widening of orogeny towards foreland • Exotic Precordillera/ Cuyania block of Laurentian affinity approaches subduction zone: shortening starts (Oclóyic phase), Famatinian-fore-arc becomes foreland	**Arc magmatism + back-arc anatexis** • ***End of Famatinian magmatism*** • Continued paired magmatic activity with cyclical crustal anatexis	
Ordovician – Middle, Early (460–490)	**Famatinian – Extensional Phase** • High-T / low P metamorphism in wide back-arc: cyclical melting • High heat flux in back-arc with heat inherited from fore-arc processing during Pampean cycle, and high crustal heat production of Puncoviscana sequence • Extension and marine sedimentation in back-arc setting • Magmatic arc 250-300 km close to trench, outboard (west) of the Pampean Arc: steep subduction	**Arc magmatism + back-arc anatexis; Basaltic magmatism** • 470-460 Ma peak magmatic activity in arc and crustal anatexis in back-arc • **End of back-arc basaltic magmatism** Arenigian-Llanvirnian, ~470Ma • **Paired magmatic belts**: calc-alkaline arc outboard (e.g. Western Eruptive Belt) and peraluminous/migmatite belt in the back-arc/foreland (e.g. Eastern Eruptive Belt) • New magmatic arc established close to trench	• ***End of marine sedimentation*** • Peak subsidence at ~470-460 Ma • **Western Eruptive Belt** proximal to arc: Tolar Chico, Tolillar, Diablo, Falda Ciénaga Formations • **Eastern Eruptive Belt** back-arc: Santa Victoria Group
	Iruya unconformity		*Iruya unconformity*
Cambrian (~500)	• **Magmatic Lull ~515-500 Ma** • Renewed marine transgression	• Decreased intensity of magmatic activity during subduction reorganization	• Mesón Group (Puna) • Negro Peinado/Achavil Fms. • La Cébila/Ambato Complexes
	Tilcara unconformity		*Tilcara unconformity*
Cambrian – Neoproterozoic (510–550)	**Pampean** • Steepening of subducting slab due to convergence deceleration leads to: a) inflow of asthenosphere into fore-arc b) high T-low P metamorphism: extensive fore-arc metamorphism and anatexis c) wide orogeny expanding to fore-arc maintaining relatively low topography • Arrival of exotic MARA block of Laurentian affinity at subduction margin: major shortening event – Pampean fore-arc becomes foreland • Widening of orogeny towards fore-arc • Low-angle subduction and fore-arc sedimentation on hyper-extended transitional continent-ocean boundary	**Arc magmatism** • 525-520 Ma isothermal decompression, anatexis, climax of the orogeny • **Paired magmatic belts**: Pampean calc-alkaline arc inboard, peraluminous granitic belt outboard in a vast fore-arc • Peak magmatic arc activity ~535 Ma • Arc is established far from trench: indicative of low-angle subduction	• ***End of Puncoviscana sedimentation*** • *Late* Puncoviscana sequence sedimentation in wide forearc recycling Pampean arc • Puncoviscana sequence passive margin sedimentation starts in Neoproterozoic – unconstrained maximum age

and to a more typical continental crust in the east. Thus, the lack of basement outcrops could be explained by a tectonically thickened package of Puncoviscana sediments above a thin layer of extended continental crust of granitic composition, directly overlying lithospheric mantle rocks.

An alternative model is that rocks of the MARA block form the basement to the Puncoviscana sequence (Fig. 3 in Casquet et al., 2018). This model suggests that the MARA block was underthrust by an upper plate consisting of Puncoviscana sedimentary rocks, separated by an ophiolite defined by lenses of mafic-ultramafic rocks. This model is based on the findings in Murra et al. (2016), who argued that the metasedimentary rocks that comprise the Sierras de Córdoba (Fig. 3) are in fact two distinct packages of broadly similar ages but different detrital zircon populations. One package has detrital zircon ages typical of the Puncoviscana sequence. The other has detrital zircon ages of Laurentian affinity, and was interpreted to be part of the MARA block. This package consists of marbles and metasiliciclastic rocks of Ediacaran to early Cambrian age (ca. 630 and 540 Ma) and, like the Puncoviscana sequence, underwent high-grade metamorphism during the Pampean orogeny in the early Cambrian. Murra et al. (2016) dated fifty-six spots by U-Pb (SHRIMP) in zircons from a single migmatite sample interlayered with marbles (shown in their Fig. 5). This sample lacks detrital zircon ages between 650 and 570 Ma. Because this is the most common zircon age group in sediments of the Puncoviscana sequence, their absence was used as evidence for a Laurentian affinity (compare Figs. 6d and e).

The model proposed by Casquet et al. (2018) with the MARA block as the underthrusted basement to the Puncoviscana sequence builds on the evidence of a single sample but is further supported by the spatial distribution of mafic-ultramafic blocks, interpreted as the ophiolites separating the terranes. While being an appealing model, it has some difficulties. One is the need to thrust the Puncoviscana sequence as a coherent block, a few hundred kilometres over the MARA basement. Another is that peraluminous granites of Famatinian age intruding the Puncoviscana sequence or Ordovician rocks, typically carry inherited zircons that define the Puncoviscana Gondwanan signature (e.g. Bahlburg et al., 2016), unlikely to be the case if the basement was part of the MARA block. Yet another difficulty is the interpretation that the mafic-ultramafic sequence represents ophiolites tectonically emplaced above the marbles of the lower plate. In the Sierras de Córdoba, the marbles interpreted to be part of MARA are closely associated with the mafic-ultramafic rocks (Fig. 2 in Murra et al., 2016, and Fig. 1 in Tibaldi et al., 2008) and in places Mg-rich, forsterite-bearing marble is intercalated with ultramafic rocks (Anzil and Martino, 2012). This proximity suggests that the sequence including marbles and the mafic-ultramafic rocks are of volcano-sedimentary origin, possibly developed at the extensional margin of Gondwana either immediately before or during deposition of the Puncoviscana sequence. In this case, the lack of ~600 Ma detrital zircons in the sample investigated by Murra et al. could be due to a restricted, local sediment source. In this regard, it is interesting to note that the mafic rocks were classified as OIBs (Tibaldi et al., 2008) and previously interpreted to be related to the subduction of mid-ocean ridges during the Pampean Orogen.

It seems that the literature consistently returns to the early interpretation that this mafic-ultramafic sequence represents ophiolites. However, the mafic rocks vary in composition and timing, and their association to ultramafic rocks is not always clear. Further to that, the spatial association between the ultramafic rocks and marbles suggests that these rocks do not represent an ophiolite sequence.

4. Paleozoic orogenic cycles in NW Argentina

Having introduced the morphotectonic subdivisions and reviewed the Puncoviscana sequence and the constraints on the nature of the basement, we now focus on the two Early Paleozoic orogenic cycles in the geological record. The Pampean orogenic cycle (Pampean cycle for short) was the first and was associated with a major shortening event related to arc magmatism and anatexis of the Puncoviscana sequence. The arc forms a discontinuous belt of magmatic rocks from the Sierras de Córdoba in the south, to the Cordillera Oriental in the north (Figs. 2 and 5). The fore-arc to the east underwent high-temperature – low to medium-pressure anatexis in the Sierras de Córdoba (e.g. Otamendi et al., 2004) and in the Western Eruptive Belt of the Puna (Lucassen and Becchio, 2003; Lucassen et al., 2000).

This cycle was most intensely studied in the Sierras de Córdoba (e.g. Martino et al., 2003; Otamendi et al., 2004; Pankhurst et al., 2000; Rapela et al., 1998b; Schwartz et al., 2008; Simpson et al., 2003) where its products have been only lightly overprinted by Famatinian intrusions and metamorphism. The Pampean cycle lasted between ~555 and 520 Ma (Adams et al., 2011; Rapela et al., 2016; Schwartz et al., 2008), and was followed by a period of quiescence between 520 and 515 and 505–490 Ma (e.g. Einhorn et al., 2015; Pankhurst et al., 1998; Ramos, 2008), and the start of the Famatinian orogenic cycle thereafter. The Famatinian cycle gave rise to the Famatinian arc, located between 200 and 350 km outboard (west) of the Pampean arc (e.g. Saavedra et al., 1998) and magmatic activity lasted from ~505 to 440 Ma (Bahlburg et al., 2016; Baldo et al., 1997; DeCelles et al., 2011; Ducea et al., 2017; Pankhurst et al., 1998). The Famatinian cycle was accompanied by intense back-arc extension and anatexis (Wolfram, 2017) with minor volumes of back-arc tholeiitic basalts (Bahlburg et al., 2016; Coira et al., 2009a; Hauser et al., 2008). A shortening event then followed, known as the Oclóyic phase of the Famatinian cycle.

The Famatinian cycle affected a belt at least 400 km wide, from the Western Sierras Pampeanas (Casquet et al., 2008; Rapela et al., 2016) to the Pampean arc in the east, overprinting the Pampean fore-arc (see maps in Dahlquist et al., 2016; Rapela et al., 2016). The Famatinian back-arc underwent intense high-T – low-P anatexis forming vast regions of peraluminous granites and migmatites (e.g. Coira et al., 1999; Lucassen and Becchio, 2003; Viramonte et al., 2007). Both the Pampean and Famatinian orogenic cycles follow a broadly similar evolution starting with subduction towards the east under Gondwana, and ending with the arrival of exotic continental ribbons with Laurentian affinities, that cause major crustal shortening.

The terminology used in the literature referring to different tectonic settings can be confusing: the Pampean arc was associated with a fore-arc region to the west (Escayola et al., 2011), but upon collision with the exotic ribbon, the fore-arc became the foreland of the Pampean Orogen (Astini, 2008; Escayola et al., 2011). During the Famatinian cycle, Famatinian back-arc extension overprinted rocks deposited in the fore-arc of the Pampean cycle, and this back-arc became itself a foreland when the arrival of the second exotic block triggered the Oclóyic shortening phase.

5. Pampean orogenic cycle

The Pampean orogenic cycle was the result of the Early to Mid-Cambrian east-directed subduction under the margin of W Gondwana followed by terrane accretion (Fig. 8) (Aceñolaza and Toselli, 2009; Omarini et al., 1999; Pankhurst et al., 1998; Rapela et al., 1998b). The orogenic cycle lasted between 555 and 515 Ma (Rapela et al., 2016; Schwartz et al., 2008) and is sometimes called the Tilcarian-Pampean or Tilcarian Orogeny (Aceñolaza and Toselli, 2009; Escayola et al., 2011; Omarini et al., 1999). Some authors define the Tilcarian Orogeny as the main period of deformation and magmatism closing the Pampean cycle (Adams et al., 2011). Tilcara refers to the angular unconformity that separates the Puncoviscana sequence from overlying Late Cambrian-Early Ordovician quartzites (e.g. Mesón Group). The latter do not record this Early Cambrian deformation event. The Pampean cycle is associated with a magmatic arc, which today flanks the Río de la Plata Craton and is separated from it by an inferred fault called the Córdoba Fault (Fig. 8) (Peri et al., 2013). It is also associated with an extensive fore-arc to the west, which underwent deformation, metamorphism,

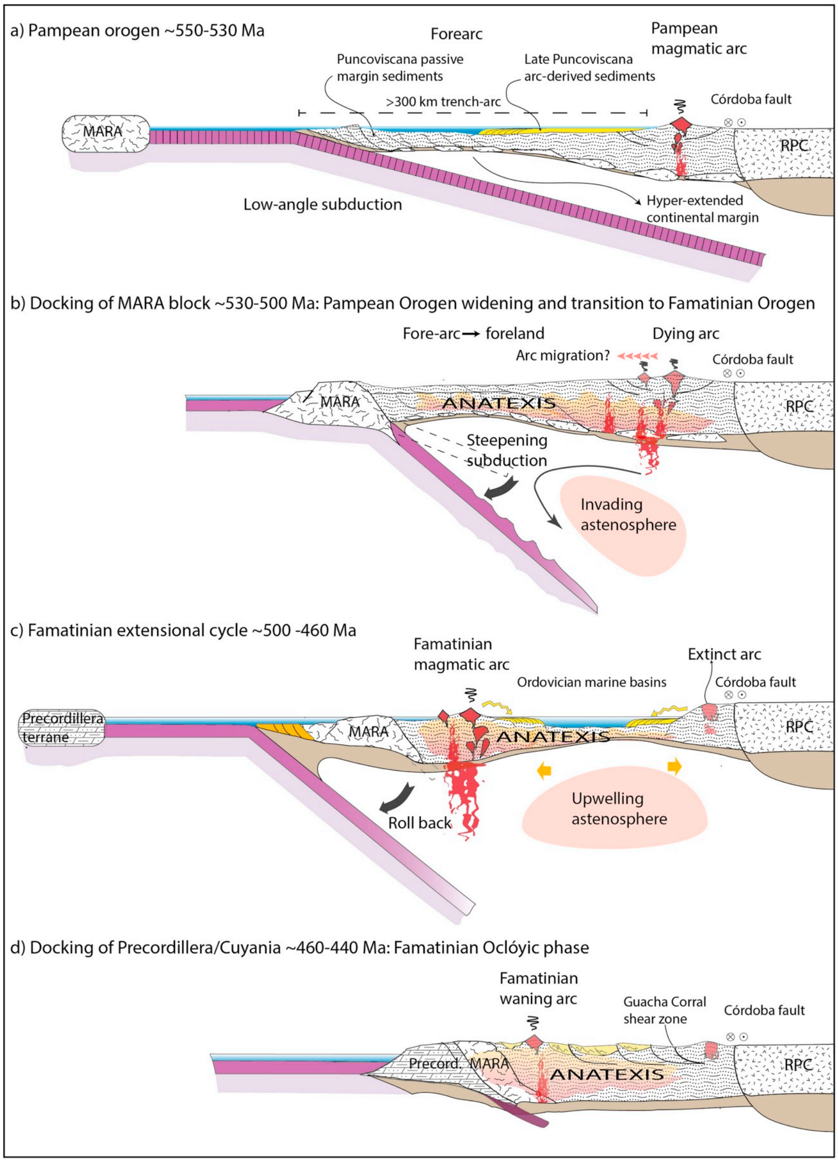

Fig. 8. a–b) Two steps in the evolution of the Pampean Orogen. It was initially characterized by low-angle subduction and a wide fore-arc developed over a previously hyper-extended continental passive margin overlain by a thick pile of the Puncoviscana passive margin turbidites. After the start of the Pampean Orogen, it received new sediments derived from the growing arc forming the *late* Puncoviscana sequence. The fore-arc was heated up possibly during the initiation of the subduction or subduction of a mid-ocean ridge (not shown) combined with gradual steepening of the subduction slab as the exotic Laurentian MARA block docked (b). This led to extensive Pampean anatexis in the fore-arc. Docking caused reorganization of the subduction system, imposing a magmatic lull preceding the ignition of the Famatinian cycle in (c). The Famatinian magmatic arc was closer to the trench and associated with widespread back-arc extension and deposition of the Ordovician marine sediments, and anatexis underneath. (d) Arrival of the second Laurentian terrane, the Precordillera/Cuyania block, triggered the start of the Oclóyic phase of basin inversion and growth of a wide orogenic belt that reached as far as the edge of the Pampean arc. This orogeny is controlled by trench-verging thrusts, possibly controlled by pre-existing Pampean fore-arc thrusts. See Table 1.

and anatexis that formed migmatites and peraluminous granites (Fig. 8a). This cycle affected the pre-existing passive margin Puncoviscana Basin (Aceñolaza and Toselli, 2009) and, as described above, it led to renewed sedimentation to form the *late* Puncoviscana sequence, that includes Pampean-age detrital zircons (e.g. Adams et al., 2011).

5.1. Duration and evolution: arrival of the exotic MARA block

The cycle started with calc-alkaline magmatism at 555 Ma (Schwartz et al., 2008) that forms a curved but broadly N-S trending belt along the Eastern Sierras Pampeanas. The arc is to the west of the Río de la Plata Craton in the southern section, and moves away from it northwards towards the Cordillera Oriental (Fig. 2) (Dahlquist et al., 2016; Hauser et al., 2011; Iannizzotto et al., 2013; Rapela et al., 1998b; von Gosen et al., 2014). Rapela et al. (1998b) found that development of the calc-alkaline Pampean arc at 540–530 Ma preceded regional low-P anatexis, which peaked at 520 Ma with the formation of peraluminous granites (Iannizzotto et al., 2013; Rapela et al., 2002; Rapela et al., 1998a; Rapela et al., 1998b). This is the timing of most intense deformation in the Sierras de Córdoba, and was accompanied by isothermal decompression (Otamendi et al., 2004; Rapela et al., 2016), marking a short period of rapid exhumation of the metamorphic complex (Baldo et al., 1997) in response to a rapid fore-arc heating and cooling event (Simpson et al., 2003). Rapid cooling at the end of the Pampean cycle is inferred from the broad similarity of U-Pb titanite and apatite ages, and $^{40}Ar/^{39}Ar$ muscovite cooling ages between 510 and 500 Ma (Simpson et al., 2003).

Metamorphic conditions within the peraluminous granite belt in the southern part of the Sierras de Córdoba reached temperatures above 800 °C at mid-crustal pressures of 8–9 kb, before decompression. The high temperatures were ascribed to a combination of burial of rocks with high contents of radioactive elements and intrusion of mafic magmas that were emplaced before and just after peak metamorphism (Otamendi et al., 2004). More recently, Do Campo et al. (2013) applied mica-chlorite geothermobarometry to rocks of the Puncoviscana sequence from the Cordillera Oriental and found consistent peak conditions of 9 kb at 250 °C. Similar metamorphism was not found in overlying rocks of the Mesón Group, constraining the age of metamorphism to the Pampean Orogen. This is the first record of high-P low-T metamorphism anywhere in either the Pampean or Famatinian Orogenies, and while requiring confirmation, it raises questions about the tectonic setting and evolution of the orogeny. Pampean-age metamorphism was also recorded in the western eruptive belt in the Puna Plateau, where high-T–low- to medium-P conditions (600–750 °C, and 4 to 8 kb)

(Lucassen et al., 2011) are associated with peraluminous magmatism (Escayola et al., 2011; Zimmermann et al., 2014). In this region, rocks were later overprinted by high-T Famatinian-age metamorphism (Lucassen et al., 2011).

Major changes in the Pampean orogenic cycle and its ultimate demise result from the accretion of the exotic MARA block to the subduction zone (Fig. 8). Early work referred to this block as the para-utochtonous Pampean terrane (Rapela et al., 1998a), but later work argued that this is an allochtonous Laurentian block comprising the Western Sierras Pampeanas, Rio Apa and the Arequipa-Antofalla block (Ramos et al., 2010; Rapela et al., 2007; Rapela et al., 2016). The acronym MARA refers to the Sierra de **M**az-**A**requipa-**R**io **A**pa blocks (Escayola et al., 2011; Rapela et al., 2016). In this paper, we will use the term MARA block to refer specifically to the ribbon comprising the Western Sierras Pampeanas and the Arequipa-Antofalla block, excluding the Rio Apa block.

5.2. Right-lateral displacement during the Pampean Orogen

The Córdoba Fault that separates the Pampean terrane from the Río de la Plata Craton is thought to link with the Transbrasiliano fault to the NE (Curto et al., 2014; Peri et al., 2013; Rapela et al., 2016). The nature of this boundary is a matter of some speculation because it is unexposed. Magnetotelluric studies find a sharp change in resistivity across the inferred fault (Favetto et al., 2008).

As mentioned in Section 3.1, Rapela et al. (2007) suggested that the lack of Paleoproterozoic detrital zircons in the Puncoviscana sequence indicated that it was deposited away from the Río de la Plata Craton and that the Córdoba Fault juxtaposed the two through right-lateral displacement. The dominance of Mesoproterozoic detrital zircons suggested to Schwartz and Gromet (2004)that Puncoviscana sedimentation took place in proximity to cratons with a strong Mesoproterozoic (1 Ga) signature, such as the Arequipa–Antofalla terrane, the Borborema Province and the Kalahari Craton. Verdecchia et al. (2011) compared two different Early Paleozoic sedimentary sequences and found that the older one lacked detrital Paleoproterozoic zircons, present in the younger sequence. This was interpreted to be a result of the juxtaposition between the Río de la Plata Craton and the sedimentary basin in the Mid- to Late Cambrian, after the main Pampean tectonothermal event at 530–520 Ma.

The inference of a dextral component of motion during the Pampean cycle is supported by the existence of dextral shear zones (Martino, 2003) and by paleomagnetic studies that indicate a large displacement between the Pampean terranes and diverse Gondwana cratons (Spagnuolo et al., 2012). On this basis, many authors interpret that the Pampean cycle was a result of dextral transpression.

5.3. Deformation and spatial distribution of Pampean-age fore-arc anatexis

Studies of Pampean deformation and metamorphism have focused on the Sierras de Córdoba and Cordillera Oriental (Hauser et al., 2011; Iannizzotto et al., 2013; Martino et al., 2010; Otamendi et al., 2006; Rapela et al., 1998a; Rapela et al., 1998b). In the Sierras de Córdoba shear zones evolved from an early stage of arc-parallel dextral shear zones into thrusts (Martino et al., 2003; Otamendi et al., 2004). Thrusting started during anatexis of the fore-arc at ~520 Ma, and was dominantly west-verging (Semenov and Weinberg, 2017). Some of these thrusts were reactivated during the Oclóyic phase of the Famatinian Orogeny, and in subsequent events (Martino et al., 2003). The anomalously wide Guacha Corral Shear Zone (lower right in Figs. 2 and 5; Bonalumi and Gigena, 1987; Martino et al., 1995) was one such Pampean thrust reactivated in the Famatinian (Semenov and Weinberg, 2017). This is a major shear zone, 10–15 km wide marking the boundary between these two orogenies (Martino, 2003; Simpson et al., 2003; Steenken et al., 2010; Whitmeyer and Simpson, 2003; Whitmeyer and Simpson, 2004): to the east, rocks record Pampean-age metamorphic peak, locally overprinted by Famatinian-age metamorphism and intrusions. To the west, in the footwall of the shear zone, granitic magmatism and peak metamorphism are Famatinian in age (Demartis et al., 2011; López de Luchi et al., 2003; Otamendi et al., 2004; Steenken et al., 2006; Whitmeyer and Simpson, 2003; Whitmeyer and Simpson, 2004), with no evidence of the Pampean Orogeny, other than metavolcanic rocks contemporaneous with sedimentation (Sollner et al., 2000) and detrital zircons (Sims et al., 1998). Semenov and Weinberg (2017) argued that this shear zone acted as a major detachment during the Pampean cycle that transferred shortening towards the fore-arc during an orogeny that grew wide rather than tall because of the hot nature of the lithosphere, unable to sustain high topography.

An important observation for the reconstruction of the Famatinian back-arc later in this paper is that high-grade Pampean-age metamorphism is absent from a wide belt between the Guacha Corral Shear Zone in the west, and the vicinity of the Famatinian arc in the east. Here, Pampean ages reappear, associated with crustal anatectic rocks and basaltic intrusions. Gallien et al. (2010) found 525 ± 9 Ma in zircon rims (U-Pb, LA-ICP-MS) within Famatinian migmatites close to the Famatinian arc in the Eastern Sierras Pampeanas. They suggested that this indicated Pampean age amphibolite facies metamorphism predating Famatinian anatexis. Lucassen and Becchio (2003) found migmatites in the Western Eruptive Belt (WEB) with metamorphic titanite ages indicative of Pampean anatexis between 530 and 510 Ma. In Cerro Negro, also in the WEB (location in Fig. 2a), Escayola et al. (2011) dated zircons (U-Pb, SHRIMP) from a two-mica granodioritic gneiss and found an age of 536.4 ± 5.3 Ma. This age is similar to the TIMS zircon age of 523.7 ± 0.8 Ma that they obtained for a sample of Hbl-Bt granodiorite, part of the main Pampean arc in the Cordillera Oriental, ~200 km across strike from Cerro Negro.

It is possible that there are as yet unknown domains that have undergone high-grade Pampean metamorphism in this broad intervening region dominated by Famatinian metamorphism. There have been reports of such metamorphic rocks in the Sierra de San Luis but this has been shown to be unlikely (Sims et al., 1998; Sollner et al., 2000; Steenken et al., 2006). The difficulty in dating metamorphism in NW Argentina lies in distinguishing detrital Pampean-age zircons (or titanite or monazite) from grains representing Pampean metamorphic growth. Sims et al. (1998) argued that contemporaneous growth of metamorphic zircon and monazite, would generate zircons with very low Th/U because of Th preference for monazite. They showed how Pampean-age zircon grains from metasedimentary rocks from the Sierra de San Luis (Nogoli and Pringles Metamorphic Complexes) had high Th/U values compared to the zircon rims of Famatinian age that were contemporaneous with monazite growth. They argued that the Pampean-age zircons were detrital grains originated from the arc, and the late Famatinian growth reflected metamorphism during which monazite and zircons grew simultaneously.

Zimmermann et al. (2014) dated zircons (U-Pb, LA-ICP-MS) from a sample of tholeiitic gabbro from the Western Eruptive Belt (WEB), part of a sequence of mafic-ultramafic intrusive lenses metamorphosed and tectonically emplaced into Ordovician sedimentary rocks that are either un-metamorphosed or weakly metamorphosed. These lenses have previously been interpreted as: a) ophiolites (e.g. Allmendinger et al., 1983; Bahlburg and Hervé, 1997; Blasco et al., 1996; Forsythe et al., 1993) and therefore a potential Ordovician suture zone (Blasco et al., 1996; Forsythe et al., 1993; Ramos et al., 2010), b) part of the Ordovician (Famatinian) arc magmatism (e.g., Coira et al., 2009a; Coira et al., 2009b), or c) magmatic rocks predating the Ordovician (Mon and Hongn, 1991). Zimmermann et al. (2014) found that they have primitive Nd isotope values and chemical features of supra-subduction magmas, and zircons yielded magma crystallization at 543.4 ± 7.2 Ma. Older zircons were interpreted to be derived from crustal assimilation. Thus, accepting the youngest zircon group as marking crystallization age rather than assimilated zircons, these lenses are supra-subduction magmatic rocks of the Pampean cycle, possibly part of the magmatism responsible for the heating of the fore-arc.

Even though there is only a relatively small number of Pampean age determinations in the fore-arc rocks exposed in the WEB, their age range this far west of the arc coincide with the duration of magmatism recorded in the arc, instead of recording only the waning stages of the Pampean cycle. This supports the inference that the fore-arc was hot throughout the Pampean cycle generating anatectic granites (Lucassen and Becchio, 2003) while being intruded by primitive magmas (Zimmermann et al., 2014).

5.4. Low-angle subduction and paired magmatic belts

The Pampean arc is currently > 300 km to the east of the edge of the MARA block (Fig. 8a). This is indicative of the large distance between the trench and the arc at Pampean times. An accurate estimate of this distance is unavailable as it requires removing the effects of deformation since the docking of MARA, including deformation related to the Famatinian, Achalian and Andean Orogenies. However, widespread crustal thickening recorded by the structures and dominant exposure of mid-crustal rocks, suggest that the current width is a minimum, from which we infer a low-angle subduction zone resulting from either rapid convergence velocities, or subduction of a young oceanic crust, or of an oceanic plateau (Cawood and Buchan, 2007).

The arrival of the MARA block is responsible for the widespread uplift and high T - low P anatexis around 525–520 Ma (Fig. 8b) (Rapela et al., 1998b; Rapela et al., 2016), giving rise to a paired magmatic belt with a calc-alkaline arc to the east, flanked to the west by a peraluminous granite belt derived from crustal melting. Schwartz et al. (2008) argued that an alternative to the accretion of MARA would be a low-angle subduction zone caused by the subduction of a mid-ocean ridge. This would have caused compression and the extinction of the calc-alkaline magmatism, contemporaneously with heating and development of the peraluminous magmatic belt in the fore-arc. This process is appealing because it explains both the mafic intrusions and anatexis in the fore-arc, as well as the flattening of subduction and widening of the fore-arc. Fore-arc heating would also explain the finding of Pampean-ages in the Western Eruptive Belt (WEB) in the Puna, 200 km west (outboard) of the Pampean arc (Escayola et al., 2011; Lucassen et al., 2011; Zimmermann et al., 2014). However, as noted above, this hypothesis has no support on the age distribution of Pampean magmatic rocks. As noted above, the timing of anatexis in the WEB is contemporaneous with the arc, covering most of the Pampean cycle. Further to that, ridge subduction would lead to a diachronous and short-lived deformation and basaltic magmatic event, which have not been documented (see discussion). A number of authors used regional considerations to argue against this model (Escayola et al., 2011; Rapela et al., 2016). One possibility is that subduction of a mid-ocean ridge occurred during the Pampean cycle rather than at the end. This would explain the unusually hot fore-arc throughout the Pampean history. The end of the Pampean cycle was, as postulated by many, marked the accretion of the MARA block.

6. Magmatic lull: late Cambrian – early Ordovician marine sedimentation

The Puncoviscana sequence is overlain by marine sedimentary sequences deposited in the Late Cambrian – Early Ordovician, between the end of the Pampean cycle at ~520 Ma, and before the start of the Famatinian orogenic cycle at ~500 Ma. This sedimentation seems to coincide with the magmatic lull previously mentioned, and gave rise to sedimentary sequences such as the Mesón Group in the Puna Plateau (Astini, 2008; Omarini et al., 1999), the Negro Peinado and Achavil Fms. of the Famatina Range (or Famatina system, Collo et al., 2009), and the La Cébila and Ambato Metamorphic Complexes in the Sierras de Velasco and Ambato (Rapela et al., 2016) (see Fig. 5 for location of different Sierras). The Puncoviscana sequence, deformed during the Pampean Orogen, is separated from these younger sediments by the Tilcara angular unconformity (Bahlburg and Breitkreuz, 1991; Ramos et al., 2010). These Late Cambrian – Early Ordovician sequences have a large peak of detrital zircons of Pampean age at ~520 Ma (Aparicio González et al., 2014; Rapela et al., 2016), and some of them have been rapidly buried during the Famatinian cycle to depths of > 20 km (Cristofolini et al., 2012; Gallien et al., 2010; Tibaldi et al., 2013). The map in Fig. 2 in Rapela et al. (2016) shows (meta)sedimentary sequences of this age covering vast sections of the Eastern Sierras Pampeanas, west of a broad belt of Puncoviscana sedimentary rocks, which they interpreted as part of the same depositional event as the Mesón Group.

The Mesón Group was first defined in the Cordillera Oriental (Kumpa and Sanchez, 1988; Sánchez and Salfity, 1999; Turner, 1960). It is dominated by quartzites deposited in shallow, near-coastal and tide-dominated environments (Aceñolaza, 2005; Sánchez and Salfity, 1999) that increase in thickness towards the north (Kumpa and Sanchez, 1988) reaching 3000 m near Bolivia (Astini, 2008). The quartzite is interlayered with within-plate basic alkaline sills and lavas, and shows significant mineral maturity (orthoquartzite), while lacking textural maturity, suggesting short transport (Astini, 2008). This author interprets its "deposition in retro-arc foreland basin", presumably as a result of the arrival of the MARA block transforming the Pampean fore-arc into a foreland. The interlayered alkaline volcanic rocks were interpreted to indicate rifting and intraplate extensional setting (Coira et al., 1999).

Adams et al. (2011) understood this sedimentary rocks as "the basal unit for the sedimentation of the Famatinian orogenic cycle in northwest Argentina". Astini (2008) recognized that the Mesón Group must have been deposited in the time gap between the end of Puncoviscana sedimentation (~520 Ma) and deposition of the ~490 Ma Santa Victoria Group shallow marine platformal sandstones and shales overlying it, above the Iruya unconformity (Astini, 2008; Turner and Méndez, 1975). Adams et al. (2011) using detrital zircons found a maximum deposition age for the Mesón Group of ~500 Ma compared to the 520–515 Ma found by Augustsson et al. (2011) which is closely aligned with the findings of Hongn et al. (2010) who suggested that extension and Mesón Group deposition would have started close to the time of intrusion of a ~526 Ma dacite porphyry.

The events occurring in this age bracket are crucial for understanding the switch from Pampean to Famatinian Orogen. Despite uncertainties and ambiguities in the available data, it seems that there is only a narrow time gap between the end of Puncoviscana sedimentation after 520 Ma (Adams et al., 2011) or after 509 Ma (Pearson et al., 2012), deformation leading to the Tilcara unconformity, sedimentation of the Mesón Group, renewed deformation and development of the Iruya unconformity and the start of deposition of the Santa Victoria Group at ~490 Ma.

In the Eastern Sierras Pampeanas, Rapela et al. (2016) compared the patterns of zircon age distribution of sediments deposited after the Pampean Orogeny with those of the Mesón Group in the Puna Plateau and found they are similar "supporting the idea that they are stratigraphically equivalent" (Fig. 6b,c). This conclusion was partly based on results from the Famatina Range (or Famatina system) where the Achavil Fm. had already been related to the Mesón Group (Collo and Astini, 2008; Collo et al., 2009). The Achavil Fm. comprises deformed, mostly shallow marine pelites, lacking evidence for contemporaneous active volcanism. It has a young detrital population of ca. 520 Ma and is overlain by the Volcancito Fm., with latest Cambrian (488 ± 1.7 Ma) fossils. The Achavil and the Volcancito Fms. are separated by an angular unconformity, the same Iruya unconformity already mentioned. The Volcancito Fm. is related to the base of the Santa Victoria Group in the Puna. Collo et al. (2009) interpreted the Achavil Fm. to result from a late-synorogenic peneplanization period at the end of the Pampean Orogen. These rocks were interpreted to be older than the Negro Peinado Fm. that crops out nearby, which has a similar detrital population but more intense deformation and higher P-T conditions of metamorphism. They argued that the Negro Peinado and Achavil Formations developed in the foreland region that resulted from the collision of the MARA block at the end of the Pampean cycle.

The La Cébila and Ambato Metamorphic Complexes in the Sierras de Ambato and Velasco, east of the Sierra de Famatina, have similar Late Cambrian to Early Ordovician depositional ages, and similar detrital zircon populations to the Achavil and Negro Peinado Fms. (Verdecchia et al., 2011).

The timing of deposition of the sediments comprising the three metamorphic complexes of the Sierra de San Luis is a little less clear. Dating of detrital zircons in samples from the two westernmost Nogoli and Pringles metamorphic complexes, yielded a concentrated age peak at ~530 Ma, followed by a broad group centred around 600 Ma, and a much smaller or absent concentration at 1 Ga, that is so well-defined in the Puncoviscana sequence (Sims et al., 1998; Steenken et al., 2006). This reduced Grenvillean-age peak is similar to that found in the Late Cambrian- Early Ordovician sediments and suggests a temporal link to rocks such as those of the Mesón Group. However, the low-grade metasedimentary rocks of the San Luis Fm., that is surrounded by the high-grade rocks of the Nogoli and Pringles metamorphic complexes, were deposited during the Pampean cycle contemporaneously with the deposition of the *late* Puncoviscana sediments. Deposition has been dated by the 529 ± 12 Ma zircons from metavolcanic rocks, contemporaneous with sedimentation (Sollner et al., 2000). This depositional age also contradicts a previous interpretation that the San Luis Fm. was deposited in the Ordovician above the Pringles metamorphic complex (Sims et al., 1997 cited in Sims et al., 1998). The easternmost Conlara metamorphic complex in the Sierra de San Luis also has detrital ages that can readily be related to the Puncoviscana Fm. and has a maximum depositional age of c. 590 Ma (Steenken et al., 2006), p. 968). In summary, the exact timing of deposition of the sediments in Sierra de San Luis remains only roughly constrained, and it is possible that some sediments relate to this period of deposition in the Late Cambrian- Early Ordovician.

All these Late Cambrian - Early Ordovician sedimentary sequences lack evidence of contemporaneous volcanism suggesting that they were deposited in a magmatic lull. This is also indicated by a detailed study of zircon inheritance where a lull was placed between ~515 Ma and 490 Ma (Einhorn et al., 2015). There is however some evidence for magmatism in this time range. For example, in the Cordillera Oriental the youngest intrusive phase in a batholith yielded ages that make it contemporaneous with sedimentation of either the Mesón Group or the base of the Santa Victoria Group (Hongn et al., 2010) blurring the gap between the two orogenies by intruding after cessation of Pampean shortening, and the start of renewed extension. Likewise, the Diablillos Intrusive Complex in the Puna (location in Fig. 2a) (Ortiz et al., 2017; Suzaño et al., 2015), could be interpreted as having crystallized at ~ 515 Ma, if taking the mean zircon U-Pb age, or more likely at ~490 Ma, if taking the youngest zircon ages, making it part of the early stages of the Famatinian cycle. There are other reports of ages of granitic rocks within this lull in the Cordillera Oriental dated at 511 ± 3 Ma (U/Pb zircon SHRIMP, Zappettini et al., 2008), and 507 ± 3 Ma (U/Pb zircon TIMS, Hongn et al., 2014).

We conclude based on: a) the reduced number of detrital zircons, b) the absence of volcanic and volcaniclastic rocks in the sedimentary sequences, and c) rare magmatic bodies, that the Late Cambrian - Early Ordovician was a period of magmatic quiescence. Together with the changes from shortening to extension and renewed marine sedimentation, this quiescence marks a period of reorganization of the subduction zone following the docking of the MARA block.

The evolution of Paleozoic NW Argentina before the start of the Famatinian Orogen can be summarized thus (Fig. 8a,b). The *late* Puncoviscana turbidites comprise sediments eroded from the exposed Pampean arc and deposited in a wide marine fore-arc. This fore-arc was being shortened contemporaneously with sediment deposition, initially perhaps as a result of relatively shallow subduction, followed by increased shortening intensity in response to the arrival of the MARA block. At this point the Pampean fore-arc became the foreland of the Orogeny and rocks were temporarily uplifted giving rise to the Tilcara unconformity. The region then subsided again, as recorded by renewed sedimentation of clean sandstones in a tidal marine setting (Mesón Group) and pelitic sediments in the Eastern Sierras Pampeanas (Achavil and La Cébila Fms and others) (see Fig. 6 for detrital zircon ages). These events must have occurred between ca. 520–515 Ma and 500–490 Ma, the time gap between the end of the Pampean and the start of Famatinian orogenic cycle. It is possible that some of these sediments recycled the earliest Famatinian zircons complicating precise determination of the timing of each event. These new sedimentary rocks were in turn deformed before deposition of Ordovician sedimentary rocks, giving rise to the younger Iruya angular unconformity (Astini, 2008; Collo et al., 2009), which must therefore have developed during the earliest stage of the Famatinian Orogen between 500 and 492 Ma, before the start of sedimentation of the Santa Victoria Group.

We note that the region is characterized by marine sedimentation during the Pampean through to the Famatinian, interrupted briefly by the Tilcara and Iruya unconformities, suggesting that the Pampean fore-arc region was never overly thickened despite the shortening record of the Pampean cycle. Alternatively, if the crust did thicken, subsequent thinning was relatively rapid so as to renew marine sedimentation in the Late Cambrian - Early Ordovician. In either case, a dense lithosphere might have prevented pronounced topography above sea level (O'Halloran and Rey, 1999).

7. Famatinian orogenic cycle

The Famatinian orogenic cycle was first defined in a congress abstract by Aceñolaza and Toselli (1976). It is responsible for a 1600 km-long magmatic arc in Argentina, striking roughly north-south between ~22°S and 38°S (Chernicoff et al., 2010). This is part of a much longer arc that resulted from east-dipping subduction under Gondwana that can be followed northwards to 15°S in Peru (Loewy et al., 2004). The cycle started with arc magmatism and back-arc extension and crustal anatexis (Fig. 8c). Mountain building started much later during the Oclóyic shortening phase, and gave rise to a zone of shortening 400 km-wide, from the Western Sierras Pampeanas in the west (Rapela et al., 2016), to the easternmost sections of the Sierras Pampeanas (Fig. 8d). Shortening started upon arrival of a second continental ribbon to the subduction zone that eventually caused the disruption of the entire system. Arc magmatism was most voluminous between ~480 and 465 Ma with a peak at ~470 Ma (Bellos et al., 2015; Castro et al., 2014; Ducea et al., 2017; Ducea et al., 2010; Mulcahy et al., 2014; Mulcahy et al., 2007; Otamendi et al., 2017; Pankhurst et al., 2000; Pankhurst et al., 1998; Quenardelle and Ramos, 1999) coincident with peak metamorphism and anatexis in the back-arc (Büttner et al., 2005; Finch et al., 2017; Sola et al., 2013; Sola et al., 2017). We start this section with an overview of the duration and geometry of the Famatinian belt, followed by subsections on basaltic and felsic magmatism, and ending with an overview of the contemporaneous sedimentary basins and how mountain building affected the belt.

7.1. Duration

The Famatinian cycle started at ~505–500 Ma (Table 1) with the growth of the magmatic arc, and development of an extensional back-arc that underwent anatexis at depth and was filled by Ordovician marine sediments at the surface (Fig. 9). This time is marked by the oldest zircon U-Pb ages in peraluminous magmatic rocks of the Famatinian back-arc (Bahlburg et al., 2016; Wolfram, 2017). Similar to the Pampean, the Famatinian orogenic cycle was also affected by the docking of a continental ribbon of Laurentian affinities, the Precordillera or Cuyania terrane, that outcrops in the south and is absent in

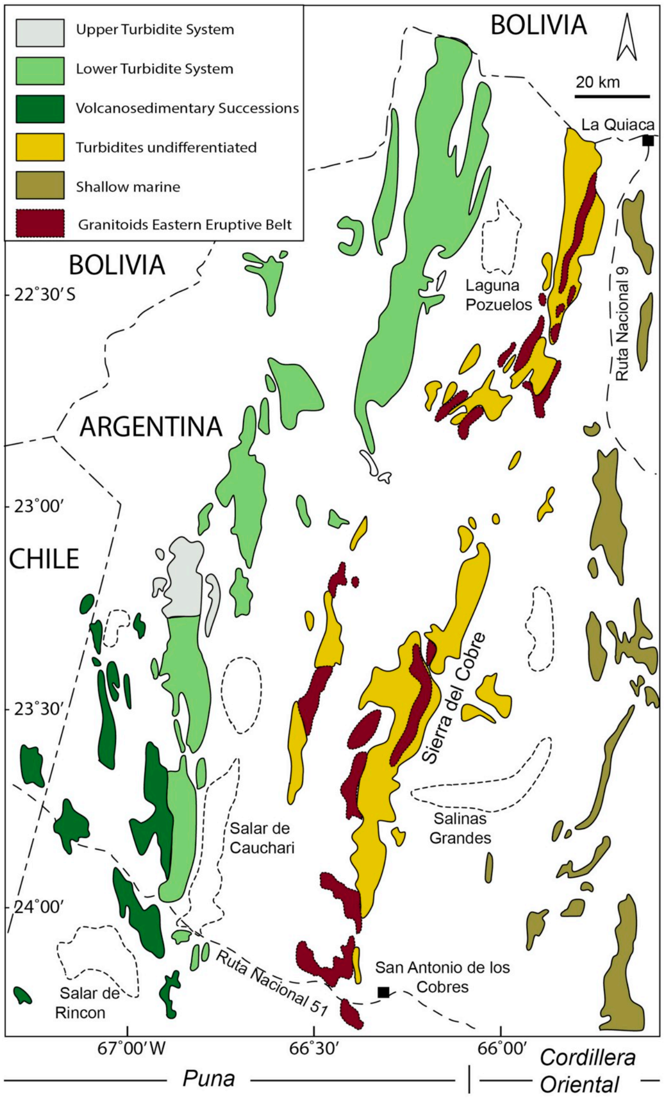

Fig. 9. Ordovician marine basins in north Puna (redrawn from Bahlburg, 1998) showing shallow marine sedimentation in the east deepening to the west.

the Puna (Fig. 2; Astini, 2008; Dalla Salda et al., 1992a; Dalziel et al., 1994; Ramos, 2004; Ramos, 2008; Thomas et al., 2015). The arrival of this ribbon triggered a phase of crustal shortening, known as the Oclóyic Orogeny or Oclóyic tectonic phase (Turner and Méndez, 1975), which may have started as early as 470 Ma or slightly later (between the Llanvirnian and Caradocian according to Ramos, 2004) (Table 1). The start of this event can be inferred from the closure of the Ordovician basins and their deformation, summarized in Section 7.5 "Ordovician marine basins". The Oclóyic tectonic phase is marked by folding and development of anomalously wide, intensely sheared, mylonitic-ultramylonitic thrust zones (Astini and Dávila, 2004; Finch et al., 2015a; Rapela et al., 1998a). It waned at around 440–435 Ma, as indicated by the end of peraluminous magmatism and crustal anatexis at 440 Ma (Bahlburg and Berndt, 2016; Bahlburg et al., 2016; Büttner et al., 2005; Mulcahy et al., 2014; Wolfram, 2017) and metamorphic titanite ages dated at 430–420 Ma (Lucassen and Becchio, 2003; Lucassen et al., 2011).

7.2. Geometry of the Famatinian magmatic belts

Similar to the Pampean cycle, Famatinian magmatism is characterized by two parallel belts: the calc-alkaline magmatic arc proper, characterized by hybrid rocks of andesitic composition with both mantle and crustal input, and the peraluminous granitic belt, dominated by magmas derived from recycled crust with only small volumes of basaltic rocks, defining bi-modal magmatism (Coira et al., 2009a). In contrast to the Pampean, where the peraluminous granitic belt is in the fore-arc, the Famatinian peraluminous belt is inboard (east) of the arc, in a back-arc/foreland position.

The nature of the two Famatinian magmatic belts has been widely studied (e.g. Alasino et al., 2016; Astini and Dávila, 2004; Bahlburg et al., 2016; Bellos et al., 2015; Dahlquist et al., 2008; Ducea et al., 2017; Lucassen and Franz, 2005; Ortiz et al., 2017; Otamendi et al., 2012; Otamendi et al., 2017; Otamendi et al., 2009; Pankhurst et al., 2000; Pankhurst et al., 1998; Rapela et al., 1998b; Saavedra et al., 1992; Saavedra et al., 1998; Sato et al., 2003; Stuart-Smith et al., 1999; Suzaño et al., 2017a; Tibaldi et al., 2013; Toselli et al., 1996; Toselli et al., 2002; Von Gosen et al., 2002; Zimmermann and Bahlburg, 2003). As recognized by Coira et al. (1999) geodynamic models for western Gondwana require a deeper understanding of the nature and distribution of magmatic rocks in this paired belt (double-arc in their terminology). Different interpretations have been put forward, including the presence of two contemporaneous and parallel subduction zones (Conti et al., 1996; Dalziel and Forsythe, 1985); an arc accompanied by back-arc magmatism (e.g. Coira et al., 2009a; Coira et al., 1999); or a collisional setting (summarized in Bahlburg and Berndt, 2016). Here, we summarize the geometry of the two belts and in other subsections summarize their compositional nature to conclude that it is most likely that the pair represents an arc and a back-arc.

In its southern section, the calc-alkaline magmatic arc trends ~NNW-SSE, parallel to the dominant regional fabric. Its most southern outcrops are exposed some 450 km south of the Sierra de San Luis (the southernmost mountain range in Fig. 5), in a small region of exposed metamorphosed quartz diorites and gabbros. These outcrops are related to a 270 km-long, positive aeromagnetic anomaly interpreted to mark the arc at depth (Chernicoff et al., 2010). This southern section is thrust over rocks of the Precordillera/Cuyania terrane, and truncated in the south by the northern margin of the Patagonia terrane (Fig. 11 in Chernicoff et al., 2010). Despite truncation, there is evidence for both Pampean and Famatinian age magmatism further south (Rapalini et al., 2013; Uriz et al., 2011). Its central part, in the Eastern Sierras Pampeanas (Fig. 5) is the best known section. Here it trends from south to north and into the Puna Plateau, where the arc trends NNW-SSE, and continues into northern Chile and onto the Arequipa block in Peru (Chew et al., 2016; Loewy et al., 2004; Ramos, 2008) and possibly to Colombia (Otamendi et al., 2017).

In the Eastern Sierras Pampeanas, the calc-alkaline magmatic arc comprises rocks ranging from gabbros to granites, forming a well-defined 80–150 km wide belt that is locally tilted exposing a nearly entire crustal section (Fig. 10). The peraluminous granitic belt to the east is accompanied by low-pressure migmatites and is exposed in an equally wide back-arc lacking rocks of intermediate composition (Figs. 2 and 5). North of 26°S in the Puna, the two belts are more formally defined forming the Western and Eastern Eruptive Belts or in Spanish: the Faja Eruptiva Occidental and Oriental, respectively (Figs. 2 and 5) (Méndez et al., 1973). The main calc-alkaline arc trends NNW across the Puna into northern Chile, and the peraluminous granitic belt trends N-S being 400 km-long, characterized by intense crustal anatexis and associated granites.

The 200–300 km-wide Ordovician peraluminous magmatic belt records paleo-depths that do not exceed ~25 km (Otamendi et al., 2012). Given its large width and relatively shallow depth of anatexis above a subduction zone, Lucassen and Franz (2005) suggested that this terrane could be "a fossil analogue of the wide zone of partially melted rocks at ~15– 25 km depth represented by the present-day seismic Altiplano Low-Velocity Zone". While the similarities are many, a key difference is that the low-velocity zone in the Altiplano developed within a very thick crustal section, unlikely to have existed during the Famatinian cycle as we will argue below.

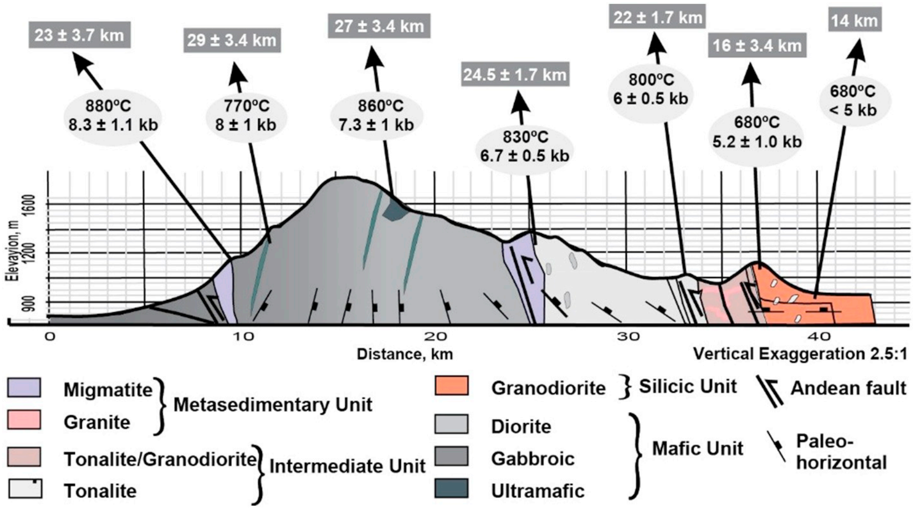

Fig. 10. Tilted section of the Famatinian arc exposed in the Sierra de Valle Fértil including P–T conditions and estimated depths (from Tibaldi et al., 2013).

7.3. Basalts: lithospheric source in the south, asthenospheric source in the north?

Away from the main calc-alkaline arc, in the Eastern Eruptive Belt (EEB) felsic magmatism in the Famatinian cycle is accompanied by mafic magmatism forming bimodal suites (Coira et al., 2009b), lacking andesites. These are juvenile basaltic magmas typical of back-arc magmatism (Coira et al., 2009a). A good example is the 80 km-long, 478 ± 6 Ma, mafic-ultramafic belt (La Jovita – Las Águilas in Sato et al., 2003; Sims et al., 1998) exposed in the Sierra de San Luis and dated by U-Pb zircon age. It comprises tholeiitic mafic-ultramafic rocks including pyroxenites, peridotites, dunites, gabbros and hornblendites (Sato et al., 2003). They have chemical compositions that resemble arc or back-arc tholeiitic rocks, and were temporally associated with peak Famatinian metamorphism. They host Ni-Cu mineralization and have evidence for having assimilated continental crust, such as the presence of graphite, phlogopite and magmatic hornblende (Skirrow and Sims, 1999).

Despite the presence of mafic rocks throughout the belt, a key feature the Famatinian cycle is that felsic crust is the dominant magma source with only minor additions from the mantle (Lucassen et al., 2000; Lucassen and Franz, 2005). Even the most mafic end-members include considerable continental input (Alasino et al., 2016; Ducea et al., 2017; Otamendi et al., 2012; Pankhurst et al., 2000; Suzaño et al., 2015; Suzaño et al., 2017a). Interpretations of the source of basaltic magmas has led to a regional divide, with significant consequences to our understanding of the thermal structure of the orogeny. In the Eastern Sierras Pampeanas, in the south, basaltic rocks are consistently interpreted as derived from melting of a Proterozoic subcrustal lithospheric mantle. Pankhurst et al. (2000) found for magmatic rocks of the Sierra de La Rioja, that isotopic values remained unchanged as a function of silica content, suggesting derivation from a single, isotopically homogeneous and evolved magma. However, in a number of other regions, including the Sierra de Valle Fértil shown in Fig. 10, there is a clear trend towards more juvenile isotopic values (Dahlquist and Galindo, 2004; Dahlquist et al., 2007; Ducea et al., 2010; Grosse et al., 2011; Pankhurst et al., 2000). Nevertheless, because even the most primitive of these rocks is isotopically evolved, it has been argued that the source is lithospheric (Dahlquist et al., 2008). More recently, Casquet et al. (2012a) found Famatinian-age amphibolites in the Valle Fértil with a depleted-mantle Nd isotopic signature, arguing that at least some of the evolved mafic magmas may have started out with a primitive signature and were hybridized by continental material.

In contrast, authors working in the Western Eruptive Belt in the Puna, in the north, have documented mafic end-members that approach juvenile mantle isotopic composition, trending towards more isotopically evolved with increased silica content. These suites were interpreted to indicate participation of asthenospheric melts and hybridization with continental magmas (Franz et al., 2006; Ortiz et al., 2017; Suzaño et al., 2017a). Franz et al. (2006) pointed out that given the 100 Ma time scale and the considerable size of the thermal event represented by the Paleozoic Orogens of NW Argentina, the thermal event must be driven by a shallow asthenosphere and its melting (Fig. 8). However, they also emphasized that the volumes of asthenospheric mafic magmas must have been relatively minor, as indicated by the absence of voluminous underplated basalts inferred from geophysics and xenoliths.

It is possible that differences between north and south reflect different mantle magma sources (asthenospheric versus lithospheric sources, respectively), however it is more likely that the mafic magmas in both regions were derived from the asthenosphere but their isotopic signature in the south was overwhelmed by widespread hybridization with crustal magmas (see Casquet et al., 2012a). In this case the N-S difference in isotopes of mafic magmas reflects different degrees of hybridization.

7.4. Felsic magmatism

Famatinian felsic magmatism is characterized by three main suites that are broadly contemporaneous (Pankhurst et al., 2000): calc-alkaline arc magmatism with considerable continental input (Otamendi et al., 2012; Otamendi et al., 2017; Pankhurst et al., 2000), S-type, peraluminous magmatism related to widespread anatexis of the Puncoviscana turbidites in back-arc setting (Alasino et al., 2014; Bahlburg et al., 2016; Lucassen and Becchio, 2003; Lucassen et al., 2011; Otamendi et al., 2004; Sola et al., 2013; Wolfram et al., 2018), and isotopically primitive TTGs (Baldo et al., 1997; Pankhurst et al., 2000; Rapela et al., 1998b; Sola et al., 2013) originated from melting at P > 15 kb.

A number of papers described the nature of the calc-alkaline rocks (e.g. Alasino et al., 2016; Bellos et al., 2015; Castro et al., 2014; Cisterna et al., 2004; Coira et al., 1999; Ducea et al., 2010; Ducea et al., 2015; Kleine et al., 2004; Lucassen and Franz, 2005; Otamendi et al., 2017; Pankhurst et al., 1998; Poma et al., 2004; Suzaño et al., 2015) as well as the peraluminous magmatic suites of the back-arc (Bahlburg et al., 2016; Becchio et al., 1999; Büttner et al., 2005; Viramonte et al., 2007; Wolfram et al., 2018). In this section, rather than reviewing the broad features of these rocks, we focus on aspects that make them particularly interesting for understanding the nature of crustal growth and recycling, or for understanding the evolution of the belt. We focus on: a) issues related to dating Famatinian magmatic rocks with wide age spread, b) multicyclic anatexis of the back-arc and implications for crustal differentiation (Wolfram et al., 2018), c) rapid rate of arc growth, and d) the significance of TTGs.

7.4.1. Dating Famatinian magmatic rocks: mean age or youngest age?

Magmatic and high-grade metamorphic rocks (migmatites) in the Famatinian arc and back-arc span the entire history of the Famatinian cycle from ~505 to 440 Ma (Fig. 6) (Table 1 in Bahlburg et al., 2016) and (Table 2 in Suzaño et al., 2017b). A significant issue with dating these rocks arises because a typical sample of granite or migmatite commonly yields a large spread of zircon and monazite along the concordia, sometimes covering the entire duration of the Famatinian Orogen (e.g. Finch et al., 2017; Ortiz et al., 2017; Pearson et al., 2012; Wolfram, 2017). The difficulties in interpreting such large spreads have been discussed by Lipman and Bachmann (2015). Here we focus on how this impacts on interpreting the evolution of the Famatinian cycle.

The first prominent peak of Famatinian ages in many samples is at ~505 Ma, marking the start of magmatism and anatexis (Fig. 11). This peak is followed by either several younger peaks or a continuous younging (Bahlburg et al., 2016; Wolfram, 2017). Pearson et al. (2012) found a large spread of zircon U-Pb ages in two-mica granitic rocks in the north (Sierra de Cachi in Fig. 5). Ages from two samples cover most of the duration of the Famatinian Orogen, and in one of them, the age range could be resolved into two age groups, ~485 and ~462 Ma, interpreted by the authors to represent igneous crystallization and metamorphic zircon growth, respectively.

Bahlburg et al. (2016) dated peraluminous igneous rocks in the back-arc in the Eastern Eruptive Belt (EEB), using LA-ICP-MS U-Pb zircon age determinations (Fig.11a). They found that the age range for each sample typically varies between 500 and 440 Ma, the duration of the Orogen, and define mean ages with large MSWD values centred at around 470 Ma. Taking the mean as the magma crystallization age would make the rocks older than the fossils in the sedimentary rocks hosting the granites. This contradiction led Bahlburg et al. (2016) to infer that the zircon population was mixed, and that magma crystallization age is given by the youngest age population or youngest individual spot in each sample. The youngest age for eight of their ten samples coincide at 444.9 ± 2.3 Ma, with two yielding older ages at 453.3 ± 9 Ma and 466.4 ± 6 Ma. Their results show that: a) each sample comprises a range of Famatinian zircon ages, b) the average age is meaningless, reflecting zircons grown at different times during the Famatinian cycle and recycled by later magmas, similar to what has been documented for Himalayan leucogranites (summarized in Weinberg, 2016), c) the youngest values reflect magma intrusion and crystallization, and d) there was a significant intrusion event at the end of the Famatinian cycle extending peraluminous magmatism to ca. 440 Ma in the back-arc, longer than previously considered.

Similarly, Wolfram et al. (2017) in the Sierra de Quilmes, also part of the back-arc, in a region without evidence for mafic or calc-alkaline magmatism, showed LA-ICP-MS zircon U-Pb yielded age spreads between ~505 Ma and 440 Ma. Their zircon ages defined five age groups, each marking an anatectic pulse, the first at the start of the Famatinian cycle at ~505 Ma and the last at the end at 440 Ma, indicating the cyclical nature of anatexis during the Famatinian (Fig. 11b).

Such spread is also found in some Famatinian calc-alkaline magmatic bodies (e.g. Ducea et al., 2010; Escayola et al., 2007; Otamendi et al., 2017; Suzaño et al., 2015), and raises the same questions about the timing of magma crystallization. The most common approach is to neglect outliers by assuming they are either inherited or metamorphic in origin, or affected by Pb-loss. The remaining ages are then used to determine a mean age, tending to disregard MSWD values even when they are larger than one, indicative of a possible mixed population.

The ~440 Ma limit for back-arc anatexis and magmatism established in the Puna (Bahlburg et al., 2016; Wolfram, 2017), contrasts with the ~465 Ma termination of magmatism inferred for the central section of the Famatinian arc in the Sierras Pampeanas (e.g. Dahlquist et al., 2008; Ducea et al., 2010; Otamendi et al., 2017; Pankhurst et al., 2000). However, many age determinations in this southern section are based on mean U-Pb spot ages, including samples with wide age spreads and MSWD > 1. This opens the possibility that arc magma crystallization was later than so far suggested.

7.4.2. Multicyclic crustal anatexis: evidence from migmatites

The five Famatinian age groups found by Wolfram (2017) in Fig. 11b were based on the investigation of four samples of anatectic Puncoviscana rocks of the back-arc exposed in Sierra de Quilmes, at the southern edge of the Cordillera Oriental where it transitions to the Sierras Pampeanas. Samples defined up to five zircon age populations within 505 and 440 Ma. She discussed the origin of the distribution and argued for zircon growth rather than the agency of Pb-loss or age mixing, because thesewould cause smearing of the age peaks along the concordia. She interpreted these peaks to indicate several melting pulses and discussed the possibility that they reflect cyclic heating and/or cyclic water influx into a hot region triggering melting events. This requires a long-lasting heat source to maintain the rocks close to their solidus for a long time, and could be a combination of back-arc extension and asthenospheric melting (Cisterna et al., 2017; Coira et al., 2009a), followed by the thickening of a crust comprising turbidites rich in HPEs. Wolfram (2017) argued that cyclic melting in the back-arc could signal pulsation in mantle heat transfer to the supra-subduction crust.

7.4.3. Rapid arc growth

Ducea et al. (2017) recently added high-precision CA-TIMS dating of zircons from six mafic rocks from the central section of the Famatinian arc stratigraphically below those previously dated. In so doing they obtained the age distribution across a 30 km vertical cross-section of the arc. Their high-precision results yielded a narrow range of ages between 472 and 467 Ma with individual age uncertainties of ~ 0.15 Ma (2σ). The narrow time slice constrained by these new results, combined with the less precise mean ages previously obtained (Ducea et al., 2010), suggest very rapid arc growth, forming an entire crustal section within 4 to 5 Ma. This implies a high magma addition rate of 300–400 $km^3\,km^{-1}\,M.yr.^{-1}$, similar to the fastest rates yet determined anywhere.

7.4.4. TTG: melting of underplated basalts?

Famatinian TTGs (Baldo et al., 1997; D'Eramo et al., 2014; D'Eramo et al., 2013; Galliski et al., 1990; Mendez et al., 2006; Pankhurst et al., 2000; Rapela et al., 1998b; Sola et al., 2013) form small plutons in the easternmost foreland of the Famatinian Orogen in the Sierras de Córdoba (D'Eramo et al., 2014; Pankhurst et al., 2000) and crop out in the Cordillera Oriental (Galliski et al., 1990; Lork et al., 1991; Sola et al., 2013). In the Sierras de Córdoba these plutons, of 3 to 10 km in radius, vary in composition from trondhjemites and tonalites in the east, to tonalites and granodiorites in the centre, to granodiorites and granites in the west to the north (D'Eramo et al., 2014).

Compositionally, these Na-rich rocks are significantly different from one another, with a number of them having geochemical features outside the range defining true TTGs (Fig. 12). The best candidates for true TTGs are the two bodies of trondhjemite from the eastern Sierras de Córdoba (the Güiraldes and La Fronda bodies in Rapela et al., 1998a). These have low K2O/Na2O < 0.6, low HREE content (Yb < 1.3) and elevated Sr/Y ratio. The granodiorite samples from the San Agustin body has many similarities, but has anomalously low La/Yb values, while granodiorite samples from the La Playa pluton are too potassic, with high Nb and low Sr/Y and La/Yb. The composition of the trondhjemite samples from the Güiraldes and La Fronda bodies indicate a range of possible melting pressures of basaltic source rocks above 15 kb (medium pressures in Fig. 12), which are higher than the original estimates 10–12 kb (Pankhurst et al., 2000; Rapela et al., 1998b).

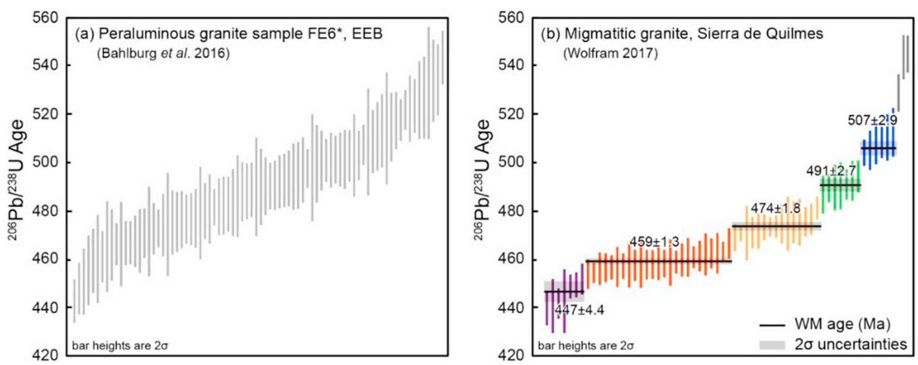

Fig. 11. Zircon U-Pb LA-ICP-MS spot ages and their uncertainties (vertical bars) for anatectic granite samples from the Famatinian back-arc region. Both samples (a) and (b) define age spreads covering the entire duration of the Famatinian cycle, from 505 to 440 Ma: a) peraluminous granite intrusion into Ordovician sedimentary rocks of the Eastern Eruptive Belt, sample WM_FE6* (Bahlburg et al., 2016). The sample defines a continuous age spread. b) Similar plot for a migmatitic granite from the Sierra de Quilmes (Wolfram, 2017) defining a number of age plateaus indicative of distinct zircon population ages.

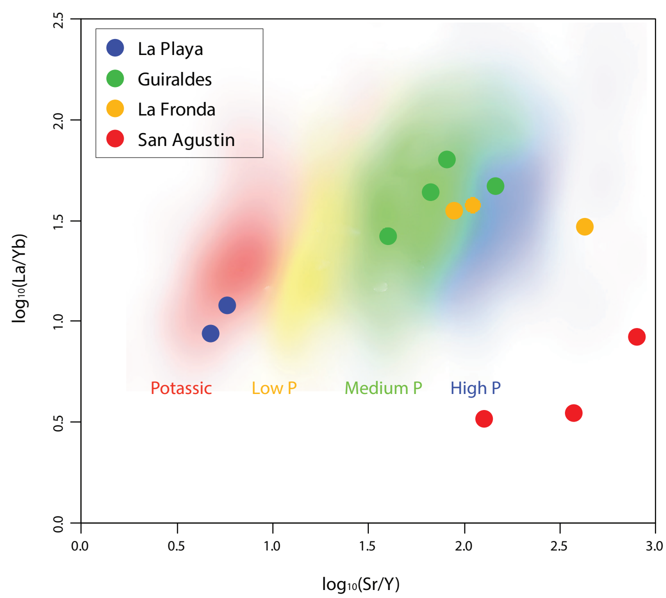

Fig. 12. TTG samples (from Pankhurst et al., 2000; Rapela et al., 1998b) plotted over the fields defined in Johnson et al. (2017). Trondhjemite samples from the Güiraldes and La Fronda plutons plot in the field for basalt anatexis at medium to high P (> 15 kbar). San Agustin granodiorites have many features common to TTGs but very low La/Yb values and high Sr/Y, whereas the La Playa samples plot in the potassic field.

Although isotopic data for the trondhjemite samples from the Güiraldes and La Fronda plutons are not available, the isotopic signature of the granodiorites from the La Playa and San Agustin bodies stand out when compared to all other available Famatinian granitic rocks (Pankhurst et al., 2000). They are isotopically primitive with $\varepsilon_{Nd(i)}$ between 0 and + 2 and $Sr_{(i)}$ close to 0.704. These rocks either have an asthenospheric source (Pankhurst et al., 2000), or represent anatexis of a juvenile sequence of basaltic rocks emplaced within the sub-crustal lithospheric mantle, in the garnet and/or hornblende stability field. In stark contrast, trondhjemites from the Cordillera Oriental in NW Argentina were found to be isotopically evolved, with values identical to the Puncoviscana sequence (Sola et al., 2013).

7.5. Ordovician marine basins: extensional environment to ~ 460 Ma

The Famatinian cycle was accompanied by widespread deposition of Ordovician marine sedimentary sequences across the width of the Puna and Eastern Sierras Pampeanas. Zimmermann et al. (2014) concluded that these sequences represent “a single active continental margin basin” and are key to constraining the Paleozoic tectonic evolution of the region. The picture that emerges from the review of the sedimentary record is that marine sedimentation dominated by turbidites defines an extensional period that started early in the Famatinian cycle at ~ 490 Ma and continued until 460 Ma (early Caradocian) (Table 1) (e.g. Astini, 2008; Cisterna et al., 2017; Hauser et al., 2011) suggesting that the region was below sea level over this ~30 Ma period. Sedimentation was accompanied by basaltic magmatism, which ended at 470 Ma, just before peak magmatism in the arc itself and maximum basin subsidence between 470 and 460 Ma. The end of sedimentation at ~460 Ma was accompanied by the beginning of shortening defining the start of the Oclóyic phase.

Early- to Mid-Ordovician marine sediments are found over all three domains of the Puna, with shallow-marine basins outcropping in the east (Cordillera Oriental, CO), and deeper basins in the west (Western Eruptive Belt, WEB; Fig. 9). Ordovician sedimentation in the Eastern Eruptive Belt (EEB) was accompanied by voluminous volcanism and peraluminous granitic intrusions, some of which are mingled with gabbros (Coira et al., 2009b; Kirschbaum et al., 2006), and some are the youngest Famatinian intrusions dated ~440 Ma (Bahlburg et al., 2016).

The Ordovician sequences have been subdivided into numerous formations (e.g. Astini, 2008; Astini and Dávila, 2004; Bahlburg and Hervé, 1997; Zimmermann and Bahlburg, 2003). Here we focus on the main features that help constrain tectonic evolution. The nature of sedimentation and *syn*-sedimentary volcanism define two sequences: in the WEB deposition was within or proximal to the arc, whereas in the EEB deposition was distal, in the back-arc.

7.5.1. Western Eruptive Belt (WEB): the arc signature in the basins

Zimmermann and Bahlburg (2003) and Bahlburg (1998) investigated the southern and northern part of the WEB, to which they refer to as south and north Puna, respectively. In the summary below, we use their Ordovician epochs. In the south WEB, the Tolar Chico, Tolillar, and Diablo Formations were deposited in the Tremadocian-Arenigian, between ~490 and 470 Ma, directly above the Puncoviscana sequence, separated by an unconformity (Zimmermann and Bahlburg, 2003). These marine sequences, including turbidites, were interpreted to be east of the Famatinian arc (Zimmermann and Bahlburg, 2003), but the transport direction recorded in all sedimentary sequences is from the S or SSW to N or NNE, along the axis of the basins (Zimmermann et al., 2002). These three formations mark a change from sources dominated by Puncoviscana rocks, to sources increasingly dominated by arc volcanic rocks with a dominance of andesitic volcaniclastic sediments interlayered with andesitic volcanic rocks. These rocks are now folded isoclinally with common west-vergence. Zimmerman and Bahlburg (2003) considered these formations to have been deposited in the back-arc proximal to the arc and some of these sediments were intruded by calc-alkaline plutonic rocks soon after their deposition. This is the case of the sedimentary rocks at Pocitos (location in Fig. 2) (Zimmermann and Bahlburg, 2003) intruded by the 476 ± 2 Ma Pocitos Intrusive Complex (Kleine et al., 2004).

The Early Ordovician sequence is overlain by the Llanvirnian-Llandeilian Falda Ciénaga Fm. (~470–460 Ma), a deep marine sequence of turbidites dominated by black shales with no evidence of volcanoclastic sediments (Zimmermann et al., 2002). It marks the deepening of the marine environment and peak subsidence while the longitudinal transport from south to north is maintained (Fig. 5 in Zimmermann et al., 2002).

In the north part of the WEB, Ordovician marine turbidites in the back-arc proximal to the arc are interlayered with lava flows related to the arc (Bahlburg, 1998). The basal sequence is the Arenigian Volcanosedimentary Successions (VS), which includes calc-alkaline basaltic, andesitic and rhyolitic volcanic rocks, interlayered with thick turbidites, indicative of proximity to the arc. This is followed by the Llanvirnian-Llandeilian Puna Turbidite Complex, which includes rare clasts of gneiss of unknown origin, and grains of Hbl, Tur, Px, suggesting erosion of high-grade metamorphic or igneous rocks. Like in the south, transport direction is along the axis of the basins but from north to south (Bahlburg, 1998). Similar to the contemporaneous Falda Ciénaga Fm. in the south, these sediments mark peak subsidence at Llanvirnian times (470–465 Ma) (see also Saavedra et al., 1998), before being folded in the Caradocian (Bahlburg, 1998). The intensity of Late Ordovician folding increases from N to S across the Puna (Bahlburg, 1998; Zimmermann and Bahlburg, 2003).

There are some similarities between sedimentation in the WEB and parts of the Eastern Sierras Pampeanas (Astini and Dávila, 2004). The early Ordovician stratigraphy predates Famatinian magmatic activity (Cisterna and Coira, 2014; Mángano and Buatois, 1996) and volcanogenic sediments appear in the Late Tremadoc. This reflects a similar change in the Late Tremadoc from the Tolar Chico Fm. to the Tolillar Fm. in the WEB (Zimmermann and Bahlburg, 2003). Thus, the Late Tremadoc marks the start of the volcanic record of the Famatinian arc and these volcanic rocks range in composition from basalts to rhyolites. In the Eastern Sierras Pampeanas, volcanic activity peaks in the Early to Mid-Llanvirnian (Saavedra et al., 1998), with thick volcanic packages above an angular unconformity dated to between the Early and Mid-Ordovician (~470 Ma) (Astini and Dávila, 2004). These thick packages coincide with peak subsidence in the WEB (Bahlburg, 1998; Zimmermann and Bahlburg, 2003) and coincide also with the end of back-arc basaltic volcanism, at the boundary between Arenigian and Llanvirnian, recorded in the Eastern Eruptive Belt (see below) (Coira et al., 2009b). This Early Llanvirnian angular unconformity recognized by Saavedra et al. (1998) was interpreted to mark a folding event and the start of shortening related to the Oclóyic phase (Astini and Dávila, 2004).

Thus, in both regions peak volcanic activity and basin subsidence occurred in the Llanvirnian, following the end of significant basaltic back-arc volcanism at ~470 Ma (Arenigian-Llanvirnian boundary). Shortening started close to the Arenigian-Llanvirnian boundary in the Eastern Eruptive Belt (Famatina Range, ~470 Ma) and later in the WEB, at the transition between the Llanvirnian-Caradocian (~460 Ma) (Bahlburg, 1998) where it follows the end of marine basin sedimentation. This angular unconformity suggests the onset of the Oclóyic shortening phase and the difference in age from south to north could either be apparent, related to uncertainty in the time constraints, or a time progression from south to north, or a number of tectonic switches between shortening and extension (Hongn and Riller, 2007). In any case, the onset of shortening was accompanied by the rapid burial and metamorphism of some Ordovician sedimentary rocks close to the Famatinian arc (Casquet et al., 2012a).

7.5.2. Eastern Eruptive Belt and Cordillera Oriental: the back-arc signature in the basins

The Tolar Chico, Tolillar, and Diablo formations of the WEB mentioned above, are temporally equivalent to the Santa Victoria Group in the Cordillera Oriental. This Group lies unconformably above the Cambrian – Late Ordovician Mesón Group (the Iruya unconformity described above, see Fig. 3 in Zimmermann and Bahlburg, 2003). The formation at the base of the Santa Victoria Group twas dated as very Late Cambrian (Furongian Stage 10 that begins at 492 Ma) (Esteban and Tortello, 2008; Zeballo and Albanesi, 2008), see also (Aceñolaza, 2005; Buatois and Mángano, 2005). Marine sedimentation continues to the Caradocian, at least to 460 Ma, ending with the start of the Oclóyic tectonic phase.

The tectonic setting of the basins in the WEB and those in the EEB and Cordillera Oriental can be differentiated by the nature of *syn*-depositional volcanic rocks. While the WEB basins have voluminous andesites (see Poma et al., 2004; Suzaño et al., 2015) and supra-subduction subalkaline basalts (Coira et al., 2009b), the EEB and Cordillera Oriental have bimodal magmatism, with peraluminous granitic magmas and alkaline basalts (Coira et al., 2009b), lacking andesites. The alkaline basalts have compositions intermediate between MORB and arc basalts, or a high-Ti OIB composition, typifying back-arc magmatism (Coira et al., 2009a). The EEB and Cordillera Oriental therefore record back-arc extension during the Tremadocian-Arenigian times.

Coira et al. (2009b) found that basaltic magmatism stopped at the end of the Arenigian, at ~470 Ma (in their Fig. 1). The tectonic significance of this remains obscure, but we note that it immediately precedes: a) peak subsidence recorded by the Llanvirnian turbidites in the WEB (e.g. Bahlburg, 1998), b) peak arc volcanic activity recorded by the volcano-sedimentary sequence in the central section of the Famatinian arc (Astini and Dávila, 2004; Saavedra et al., 1998), and c) coincides with the estimated peak in arc magmatism at ~470 Ma (e.g. Ducea et al., 2017), as well as regional metamorphism and anatexis of back-arc metasedimentary sequences (Wolfram, 2017). These changes preceding the onset of shortening, suggest a possible thermal peak of the system, leading to melting of the Puncoviscana sequence that impeded the passage of back-arc basalts after 470 Ma.

Hauser et al. (2011) found similar E-MORB/OIB basalts in the Cordillera Oriental. These basalts form peperites and are interlayered with deep marine sediments younger than 463 ± 11 Ma. Given the age uncertainty, these are either the same age or younger than those in the EEB. These OIBs have $\varepsilon_{Nd(i)} = -1$ to $+4$, indicative of a juvenile mantle magma that assimilated some crust, as indicated by the presence of detrital zircons. It is possible that these basalts are younger than Arenigian, suggesting the possibility of either diachronous magmatism evolving in time from the EEB to the Cordillera Oriental, or more simply that small volumes of back-arc basalts continued to find their way to the surface in the Cordillera Oriental.

The ~500–490 Ma Diablillos Intrusive Complex (location in Fig. 2a; Suzaño et al., 2015) lies close to the boundary between WEB and EEB. These rocks range from diorites to monzogranites, record magma mingling and have $^{87}Sr/^{86}Sr_{(i)}$ from 0.70446 to 0.71278, $\varepsilon_{Nd(i)} = -7$ to $+2.5$, and zircon $\varepsilon_{Hf(i)} = -3$ to $+3$ (Ortiz et al., 2017; Suzaño et al., 2015). Given their age, these rocks are most likely part of the main Famatinian arc. However, 20 km to the north, Viramonte et al. (2007) investigated a bimodal suite, including 485 ± 5 Ma basic tholeiitic metavolcanic rocks that have primitive $\varepsilon_{Nd(i)} = +2.3$ to $+2.5$ and crustal peraluminous granitic rocks dated at 462 ± 7 Ma and 475 ± 5 Ma, that have negative $\varepsilon_{Nd(i)}$ values. The comparison between these two locations, 20 km apart along the main N-S regional trend, shows that a region that was intruded by arc magmas, presumably proximal to the arc, became arc distal, dominated by bimodal magmatism. This in turn shows that the arc/back-arc boundary migrated in time and space. This point is reinforced by the findings of Cisterna et al. (2017), who described Arenigian MORB indicative of a back-arc, that are contemporaneous with and north of the intra-arc or inter-arc volcaniclastic sedimentary in the same mountain range described by Astini and Dávila (2004).

In summary, the Ordovician (meta-)sedimentary sequences preserved in the Puna suggest that marine sedimentation occurred throughout the first half of the Famatinian cycle. The evidence for short sediment transport paths suggests that there was significant topography from the early stages of arc magmatism, between the main Famatinian arc to the west and the neighbouring marine basins in the WEB (Zimmermann and Bahlburg, 2003), some of which include submarine

Famatinian andesitic arc volcanism (e.g. Mángano and Buatois, 1996). In contrast, the basins in the arc-distal EEB, lack evidence for arc magmatism. Marine sedimentation here was accompanied by bimodal magmatism defining an extensional marine back-arc setting that was transformed into a foreland when folding started, either at ~470 Ma as determined in the Eastern Sierras Pampeanas in the south, or at the start of the Caradocian at ~460 Ma in the north, marking the start of the Oclóyic tectonic phase of the Famatinian cycle. Famatinian peraluminous granite intrusions in the Eastern Eruptive Belt (EEB) and Cordillera Oriental continued during and after folding, ending at ~440 Ma (Bahlburg et al., 2016).

7.6. Oclóyic tectonic phase: development of major thrust zones

The Oclóyic Orogeny or Oclóyic tectonic phase of the Famatinian cycle (Astini and Dávila, 2004; Rapela et al., 1998a; Turner and Méndez, 1975) shortened a region > 300 km wide, from the Western Sierras Pampeanas (Casquet et al., 2008; Rapela et al., 2016) to the vicinity of the Pampean arc (Fig. 8d) (Semenov and Weinberg, 2017). As presented above, the start of the Oclóyic phase was some time in between ~470 Ma, determined in the south, and 460 Ma in the north, and is characterized by folding (Bahlburg et al., 2016) and regional thrusting, giving rise to some of the widest and most intensely sheared thrust zones in any orogen (Finch et al., 2015a; Finch et al., 2016; Semenov and Weinberg, 2017; Whitmeyer and Simpson, 2003).

Otamendi et al. (2017) argued that the arrival of the Precordillera/Cuyania block and the start of the Oclóyic phase occurred at 465 Ma, marked by the end of magmatic activity in the central portion of the Famatinian calc-alkaline arc in the Eastern Sierras Pampeanas. However, as discussed above, estimates of the timing of the end of arc magmatism could be an artefact of using mean zircon ages of samples with complex histories of zircon recycling. In this regard, we note that Castro et al. (2014) found zircon overgrowth at 450 Ma in the Famatinian arc in the Eastern Sierras Pampeanas, “revealing a protracted thermal history”. This is more clearly aligned with the duration of calc-alkaline arc magmatism in the Western Eruptive Belt, where it continued beyond 465 Ma, as marked by the 450–430 Ma U-Pb zircon magma crystallization ages in northern Chile (Loewy et al., 2004); the 460–450 Ma titanite ages for a granodiorite in Lucassen et al. (2011) and ages summarized in Fig. 3 in Ramos (2008). We therefore suggest that arc magmatism may have lasted as long as back-arc anatexis and that both ended roughly simultaneously at ~440 Ma (Table 1).

Otamendi et al. (2017) argued also that, because the Precordillera/Cuyania block does not extend northward towards the Puna, its arrival caused a westward rotation of the arc in the Puna. While there is weak change in the trend of the arc (see Fig. 2), we will argue below that the geological history of the central Famatinian arc and the Puna region are not that different and both underwent the Oclóyic phase between ~460–440 Ma.

The thrusts controlling crustal thickening during the Oclóyic phase have a number of features in common. The major shear zones are extremely wide, such as the 1 to 4 km-wide La Chilca Shear Zone (Larrovere et al., 2008), the 3 km-wide El Pichao Shear Zone (Finch et al., 2015a), and the extreme case of the 10–15 km-wide Guacha Corral Shear Zone (Martino, 2003; Otamendi et al., 2004; Semenov and Weinberg, 2017; Simpson et al., 2003; Whitmeyer and Simpson, 2003). They have been truncated by block movement due to Andean faulting, making it difficult to follow them across the landscape, however, the 2 km-wide TIPA (Tinogasta-Pituil-Antinaco) Shear Zone (López and Toselli, 1993) can be followed for over 200 km (Höckenreiner et al., 2003). The location and length of the shear zones are marked by numbers 1 to 5 in Fig. 2a,b.

These thrusts are collectively characterized by intense mylonitization (Fig. 13), including ultramylonites that can be up to 1 km-wide (Finch et al., 2016). They are dominantly W-verging thrusts, towards the trench (Fig. 8d) (Cristofolini et al., 2014), with only a small number of E-verging thrusts (e.g. Collo and Astini, 2008). Most significantly, none of these thrusts exhume high-pressure rocks, instead the hanging wall typically comprises cordierite-sillimanite-garnet migmatites, indicative of high-T – low-P metamorphism. Their hanging wall also typically records higher metamorphic conditions than rocks in their immediate footwall, indicating exhumation of deeper, hotter rocks (e.g. La Chilca Shear Zone, Larrovere et al., 2008; Otamendi et al., 2004). In both the El Pichao Shear Zone and the Guacha Corral Shear Zone, W-verging folding and thrusting started during anatexis (Finch et al., 2015a; Semenov and Weinberg, 2017). Büttner (2015) provided an alternative interpretation for the deformation in the Sierra de Quilmes. He suggested that top-to-W vergence represented an initial phase of normal movement associated with the extensional phase. Subsequent block rotation tilted the section, so that in its current position it records an apparent thrust movement. He argued that the El Pichao Shear Zone represents a later thrust, explaining emplacement of hot rocks over the cooler rocks of the footwall. While this is possible, Finch et al. (2015b) suggested instead that the two phases were part of a continuous top-to-west shortening event, explaining the fact that both phases have the same transport direction and the hot-over-cold geometry. Finch et al. (2017) dated monazite grains from within the El Pichao Shear Zone and from the footwall, and found an age difference of ~25 Ma (~460 and 435 Ma, respectively). They suggested that movement would have started at ~470 Ma while rocks were still partially molten and thrusting caused cooling of hanging wall migmatites, and heating of the footwall explaining the monazite age difference.

The NNW-SSE-trending TIPA Shear Zone cuts across Ordovician granitic rocks and is truncated by undeformed Carboniferous granites (Höckenreiner et al., 2003; Larrovere et al., 2016). The TIPA extends over several mountain ranges and in the south, it is one of a number of parallel shear zones comprising a broader system, and is a steep doubly-verging reverse shear zone, > 15 km thick, suggested to be the root of a thrust system (Larrovere et al., 2016). In the northern end, the TIPA is 2 km-wide with a reverse top-to-WSW motion (Höckenreiner et al., 2003). Rims of *syn*-tectonic garnets from this area yielded a Sm-Nd age of 402 ± 2 Ma interpreted to date mylonitization (Höckenreiner et al., 2003). This is much later than the typical age range for the Famatinian cycle, but a period of deformation around this time is also suggested by the 410 Ma titanite and zircon ages further south (Mulcahy et al., 2014). If this age is confirmed more widely, it suggests that movement either continued or was reactivated ~30–40 Ma after the end of Famatinian magmatic activity, possibly marking the arrival of yet another ribbon, the Chilenia terrane, inaugurating the Achalian cycle.

The Arenosa Creek Shear Zone is a small shear zone included in this review because of its consistent amphibole Ar-ages dating deformation. It shears a quartz-bearing metagabbro/diorite, initially at granulite facies temperatures, followed by amphibolite facies deformation. The shear zone is only 2.5 m-wide, with thrust-to-the-W shear sense, on a steeply dipping, NNE striking plane (Castro de Machuca et al., 2012). Amphiboles deformed at amphibolite facies yielded two well-defined Ar-Ar plateau ages at 441.9 ± 1.9 Ma and 438.7 ± 1.9 Ma, for a porphyroclast and for recrystallized amphiboles, respectively. These ages were interpreted as “the best estimates for the minimum age of the mylonitic event” (Castro de Machuca et al., 2012).

The Guacha Corral Shear Zone, started its history during the Pampean Orogen and was reactivated during the Famatinian cycle (Demartis et al., 2017) and was also active towards the end of the cycle (Semenov and Weinberg, 2017). In the process it thrusted Pampean-age migmatites west over greenschist facies rocks metamorphosed during the Ordovician Famatinian Orogen. The shear zone marks an important boundary with rocks in the hanging wall to the east having Cambrian peak metamorphism, and those in the footwall to the west, recording Ordovician peak metamorphism (Demartis et al., 2011; López de Luchi et al., 2003; Otamendi et al., 2004; Steenken et al., 2006; Whitmeyer and Simpson, 2004). This shear zone is part of a broader shear zone system (Simpson et al., 2003; Whitmeyer and Simpson, 2003) and part

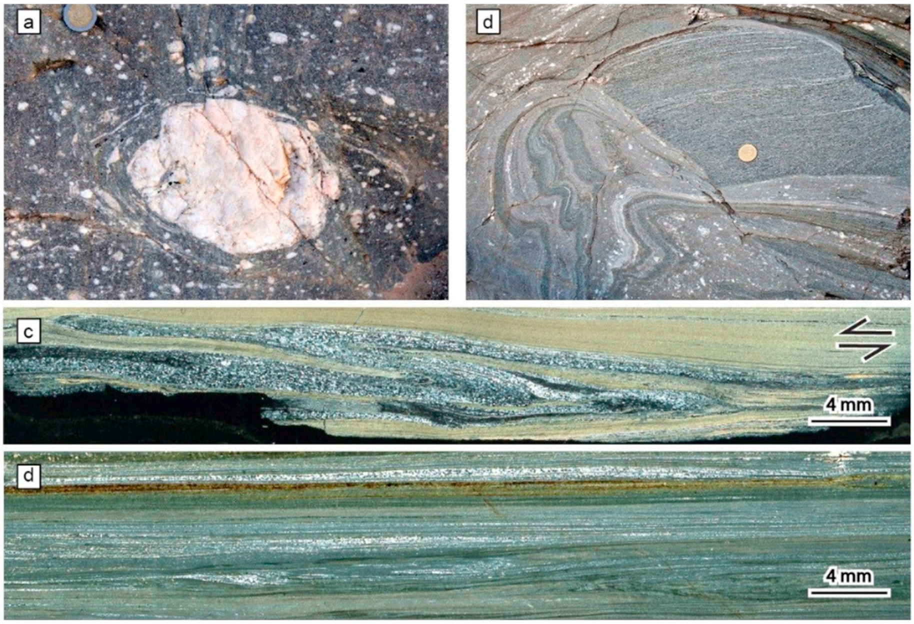

Fig. 13. a) Ultramylonite from the 3 km-wide El Pichao Shear Zone in Sierra de Quilmes (Finch et al., 2016), showing disaggregation of pegmatite intrusions and rotation of pegmatitic K-feldspar. This view is perpendicular to the stretching lineation. b) Angular clast of amphibolite in mylonitic, folded metasedimentary sequence including thin folded layers of amphibolite. From Sierra de Comechingones. c-d) Photomicrographs of ultramylonite from the 2.5 km-wide greenschist facies section of the Guacha Corral Shear Zone, Sierra de Comechingones, showing a shear fold under cross-polarized light (from Semenov and Weinberg, 2017). c) Recrystallized Qtz-Fsp-Ser layer defining recumbent fold in a Chl-Bt ultramylonite. (d) Chl-Bt ultramylonite showing extremely fine banding and transposed fold hinges.

of the easternmost shear system activated during the Famatinian, marking the eastern edge of this orogeny (Semenov and Weinberg, 2017).

Given the back-arc rifting early in the Famatinian cycle, it is plausible that some of these thrusts had an earlier history of normal movement. Suzaño (2015) argued that the dextral strike-slip El Peñón Shear Zone in the Puna Plateau was initially the boundary of an extensional zone, with an uplifted east side, providing sediments to the basin in the west. This zone sheared a *syn*- or pre-tectonic peraluminous granite with a wide spread of ages from 536 ± 2.9 to 486 ± 14 Ma, the latter considered to be the final granite intrusion age.

The Famatinian trench-verging system of potent thrusts was responsible for shortening and crustal thickening, and widened the Famatinian Orogeny eastwards, reaching the vicinity of the Pampean arc, 100 to 250 km inland from the Famatinian arc. These trenchward thrusts are in the foreland region of the orogeny and have therefore an unusual vergence. We suggest that this asymmetry is inherited from early-formed trench-verging thrusts in the fore-arc of the Pampean Orogen. This is best demonstrated by the reactivation of the Guacha Corral Shear Zone in the Famatinian cycle (Semenov and Weinberg, 2017). The existence of such early thrusts may have organized the geometry of subsequent deformation, forcing new Famatinian thrusts to follow the same vergence.

The anomalous widths of many Famatinian thrusts suggest they were long-lived, such as would be expected for the detachment soles of major thrust zones in widening orogens. This has been suggested for the Guacha Corral Shear Zone during the Pampean cycle (Semenov and Weinberg, 2017). A possible explanation for their width is shear heating. If each of the major shear zones were active for a few million years, moving at velocities of a few cm/y, approaching the velocity of plates, the balance between frictional heating and heat diffusion would give rise to such wide shear zones (Finch et al., 2015a).

8. Discussion

The picture that emerges from this review is summarized in Table 1 and Fig. 8. The Pampean arc starts with the activation of the passive margin of Gondwana. This arc developed > 350 km inland over a low-angle subduction zone. The fore-arc received sediments originating from the arc, forming the *late* Puncoviscana sequence. Contraction increased towards the end of the Pampean cycle with trenchward thrusting in potent shear zones, which widened the orogen towards the fore-arc. The arrival of the exotic MARA block is associated with this shortening and led to the demise of the Pampean cycle and reorganization of the system. This led to the Tilcara angular unconformity, between the Puncoviscana sequence and the Late Cambrian – Early Ordovician marine basins (e.g. Mesón Group, Negro Peinado, Achavil and La Cébila Fms.). This sedimentation lacks significant volcanic or volcaniclastic input, suggesting a lull in magmatism immediately preceding the start of the Famatinian cycle and its magmatism. The Famatinian cycle was marked by a significant trenchward arc migration, and widespread Ordovician marine sedimentation during back-arc extension above a second angular unconformity, the Iruya unconformity. Marine sedimentation ended with the start of the ~20–30 Ma long Oclóyic shortening phase at ~470–460 Ma, flagging the arrival of the second exotic block to the subduction zone, the Precordillera/Cuyania block. This phase gave rise to potent thrusts in the foreland region verging towards the trench and distributed over a 300 km wide orogen, reaching all the way inland to the western edge of the extinct Pampean arc.

8.1. Pampean fore-arc width and anatexis: asthenospheric invasion?

The current distance between the eastern edge of the MARA block and the Pampean arc is ~350 km and provides a first-order indication of the width of the Pampean fore-arc (Figs. 2, 5 and 8). A more precise value requires subtracting the effects of both the Pampean shortening, and the net deformation occurred during the Famatinian, Achalian and Andean Orogens. Such estimates are not available but given the dominance of major thrusts and evidence for crustal thickening, we infer that the current width of the fore-arc is likely a minimum. This suggests that the arc formed above a low-angle slab subduction, > 350 km from the trench.

Apart from its width, another puzzling aspect of the Pampean fore-arc region is that it underwent extensive anatexis during the orogeny. Fore-arcs are generally regions of low heat flow (Peacock, 1996), however there are examples where high temperatures have been recorded (Barnett et al., 1994). Ridge subduction has been used to explain Pampean fore-arc anatexis (Schwartz et al., 2008; Simpson et al., 2003) and could also account for a period of low-angle subduction accompanied by diachronous magmatism and metamorphism. While we cannot discount the possibility of ridge subduction, the region lacks some of the features that characterize this process, such as mid-ocean ridge basalts interacting with trench sediments, and subsequent alkali magmatism (Meneghini et al., 2014). Further to that and as seen above, the fore-arc underwent anatexis and mafic-ultramafic magmatism over a reasonably protracted time possibly corresponding to most of the duration of Pampean arc activity. Beyond ridge subduction, a number of other explanations have been suggested for other hot fore-arc regions: a) shear heating and water fluxing during initiation of subduction causing melting above the subducting slab (Barnett et al., 1994; Creixell et al., 2016), b) late trenchward migration of the arc, or c) a steepening of the subduction slab causing asthenosphere influx under the fore-arc region and basaltic intrusion into the fore-arc (Capitanio et al., 2007; Riel et al., 2013; Riel et al., 2016). We have no evidence in support of option (a). Options (b) and (c) could both be related to the arrival of the exotic block and the consequent steepening of the subduction zone. However, similar to ridge subduction, these options require that fore-arc magmatism and metamorphism post-date the end of arc magmatism in the hinterland. This diachronicity has been suggested (Rapela et al., 2002; Rapela et al., 1998b; Schwartz et al., 2008), but the spread of Pampean ages in the fore-arc discussed above seems to reject this.

Given current constraints, we suggest that the low-angle subduction during the Pampean cycle reflects cessation of shortening in the Pan-African orogens in the interior of Gondwana, and initiation of fast convergence and subduction at the passive margin (Cawood and Buchan, 2007). Prolonged anatexis in the fore-arc could be a combined result of thermal perturbation during subduction initiation (option (a) above), followed by a steepening of the subduction and asthenospheric invasion beneath the fore-arc upon arrival of the exotic block at the trench (option (c)) and possible arc migration (option (b)).

8.2. Extent of the Famatinian back-arc rift

Widespread Ordovician marine sedimentation during the Famatinian cycle from ~492 Ma to ~470–460 Ma suggests a long period of back-arc extension, with arc-proximal sequences in the Western Eruptive Belt containing andesites, and arc-distal sequences in the Eastern Eruptive Belt and Cordillera Oriental characterized by bimodal magmatism. The spatial distribution of Ordovician sediments can be used together with the distribution of Pampean metamorphic and magmatic rocks to constrain the geographical limits of the Famatinian back-arc rift zone (Figs. 2 and 5). As already described, rocks recording Pampean age metamorphism outcrop only in the vicinity of the Pampean arc (e.g. Sierras de Córdoba, east of the Guacha Corral Shear Zone, Figs. 2 and 5), and in the vicinity of the Famatinian arc (e.g. Western Eruptive Belt). These two regions are separated by a wide region where the Ordovician basins crop out interspersed with Puncoviscana basement that underwent high-grade metamorphism of Famatinian age. This geographical distribution suggests that the regions of Pampean metamorphism mark the eastern and western shoulders of the Famatinian back-arc rift, and the intervening region with its Ordovician basins and Famatinian metamorphism, marks the downthrown central rift region. In this central zone, high-grade Pampean metamorphic rocks were buried more deeply, and low-grade Pampean metamorphism of the Puncoviscana rocks was overprinted by high-grade Famatinian back-arc metamorphism, masking the Pampean signal. If this is so, the Guacha Corral Shear Zone could represent the reactivated edge of the Famatinian back-arc rift, and the current width of the back-arc, even after the Oclóyic shortening cycle, is ~ 150–200 km wide in the south, ~250–300 km in the centre, and 250 km in the north.

8.3. Paired magmatic belts: opposite polarity

The discussion above feeds into a discussion of another key feature of both the Pampean and Famatinian cycles: their paired magmatic belts. Even though these two orogenies resulted from east-dipping subduction zones, the paired magmatic belts have opposite polarity: the calc-alkaline Pampean arc is paired with a peraluminous, S-type magmatic belt in its fore-arc to the west (Schwartz et al., 2008), while the Famatinian arc is paired with a peraluminous magmatic belt in its back-arc to the east. Thus, the two calc-alkaline arcs, ~300 km apart, bound an intervening region that underwent extensive anatexis during both Pampean and Famatinian orogenic cycles lasting over 100 Ma.

8.4. Prolonged HT-LP metamorphism and anatexis

The starkest feature of the tectono-metamorphic history of the Pampean and Famatinian orogenic cycles is precisely this sustained high-temperature environment and the dominance of cordierite-bearing migmatites, indicative of Buchan facies series metamorphism (e.g. Lucassen and Becchio, 2003; Sola et al., 2017). In common with many orogenies world-wide, it is difficult to pin-point the precise source of heat for this long-lasting metamorphism. This was clearly expressed in Lucassen and Becchio (2003), who found a wide range in U-Pb titanite ages in high-grade rocks from the Western Eruptive Belt, where metamorphic conditions were 600–750 °C and ~5–7 kb. They found ages ranging from the Cambrian through to the Ordovician and Silurian interpreted as "the effect polyphase deformation with deformation-enhanced recrystallization of titanite and/or different thermal peaks during a longstanding, geographically fixed, high-T regime in the mid-crust of a continental magmatic arc".

As discussed above, the causes of fore-arc melting during the Pampean cycle are unclear but could be related to the establishment of the Pampean subduction zone, the subduction of an active ridge or a late steepening of the subducting slab and arc migration. In the Famatinian back-arc, peraluminous, S-type granites and migmatites have individual rock samples recording zircon age spreads spanning the entire duration of the Famatinian cycle, from 505 to 440 Ma (Bahlburg et al., 2016; Wolfram, 2017). This age range implies back-arc melting from the start of the orogeny and continuous high-temperatures for over 60 Ma. Further to that, several anatectic events have been inferred in a single location suggesting cyclical melting events (Fig. 11b; Wolfram, 2017).

The heat in the Famatinian back-arc was likely partly inherited from the Pampean cycle, boosted by the high heat flux of back-arc settings. Asthenospheric melting and heat advection with mafic magmas would have helped, as evidenced by back-arc basalts (Coira et al., 2009b; Kirschbaum et al., 2006), but their small volumes combined with the lack of evidence for considerable mafic magma underplating (Franz et al., 2006) suggest that this was only a minor addition. High heat production of the Puncoviscana sediments would also have contributed, particularly during crustal thickening of the Oclóyic phase. Wolfram et al. (2017) calculated geothermal curves assuming that the Famatinian back-arc crust comprised dominantly Puncoviscana sequence (Ducea et al., 2010; Franz et al., 2006; Lucassen et al., 1999; Perarnau et al., 2012). They used a large geochemical database to determine that the Puncoviscana sequence has an average $K_2O = 2.69$ wt%, Th = 13.4 ppm, U = 2.9 ppm, producing ~2 $\mu W\,m^{-3}$, ~ 20% more than the average upper continental crust composition (see also Lucassen et al., 2001). They found that this heat production could sustain relatively high crustal temperatures for crustal thicknesses of 30 or 40 km, but not sufficiently hot to trigger mid-crustal anatexis.

8.5. Hot and wide orogens

The sustained high temperature over 100 Ma of subduction through the Pampean and Famatinian cycles had an impact on the nature and width of the orogenies. This was recognized by Semenov and Weinberg (2017) who argued that high-temperature was responsible for the widening of the Pampean The same arguments apply to the hot Famatinian back-arc when it became a foreland during mountain building. Akin to the hot and wide Archean orogens (Chardon et al., 2009), or the Variscan (Franke, 2014), the crust during the Pampean and Famatinian Orogens was unable to sustain high topography forcing the orogenies to spread laterally (Cruden et al., 2006). Despite the potent thrusts across the orogenic belts, high pressure rocks have not been exhumed (e.g. Finch et al., 2015a; Larrovere et al., 2011; Otamendi et al., 2004), suggesting the thrusts were reasonably flat, long-lived decollement structures, with minor exhumation of the hanging wall, despite significant lateral movement (Semenov and Weinberg, 2017). Many of these features are typical of accretionary convergent plate margins (Collins, 2002a; Cruden et al., 2006).

8.6. Exotic terranes: The N and S divide

A feature of considerable controversy in the Paleozoic of NW Argentina is the northward continuation of MARA and the Precordillera/Cuyania exotic terranes accreted to the Gondwana margin (Figs. 2 and 8) (e.g. Lucassen et al., 2000). Rocks of Laurentian affinities that form the Western Sierras Pampeanas section of the MARA block are presumed to link directly with the Arequipa-Antofalla block in the north, running through the Puna Plateau, as if they were continuous. However, there are no rocks of Laurentian affinity in the Plateau. They disappear north of 27° only to reappear north of the Puna, in Bolivia and Peru. Similarly, the Precordillera/Cuyania terrane continues northwards only to 29°S (e.g. Thomas et al., 2015).

Their presence in the south and absence in the north has led some publications to argue that the Early Paleozoic orogenies in the Puna were purely accretionary (Lucassen and Franz, 2005; Zimmermann et al., 2014). This was also reflected recently in Otamendi et al. (2017) who suggested an anticlockwise rotation of the Famatinian arc in the north compared to the south. We note that this rotation is relatively minor and recent modelling suggests that a continental ribbon clogging a section of a subduction zone would cause intense crustal shortening of the over-riding plate in the clogged region, while triggering large-scale roll-back, rotation and extension of the over-riding plate where subduction remains unimpeded (Moresi et al., 2014). This would lead to an oroclinal bend with shortening in the south contemporaneous with widespread extension in the north. This is not the case. We emphasize instead the similarity between the Puna and Sierras Pampeanas, with: a) a contemporaneous 250–300 km trenchward jump in the position of the magmatic arc, b) extensive and broadly contemporaneous Cambrian to Ordovician marine sedimentation interrupted by angular unconformities, c) a similar record of shortening, during the Oclóyic cycle, and d) similar magmatic evolution and high-T – low-P metamorphism. Thus, the similarity of the two regions and the linearity of the two magmatic arcs suggest that either the exotic terranes continue to the north and are buried under the Puna Plateau or that there was significant coupling between the downgoing subduction slab and the overriding plate under the Puna, causing orogenies of similar magnitudes to those in the south.

8.6.1. Is MARA under the Puna Plateau?

The literature commonly refers to the MARA block in the Puna Plateau as if it were a physical body of exposed rocks linking the Western Sierras Pampeanas continuously with the Arequipa-Antofalla block to the north of Argentina. A review of the literature reveals that there are no such rocks exposed in the Western Eruptive Belt (WEB) in the Puna. The rocks that are exposed in the WEB (Fig. 4 of Ramos, 2008) comprise the Puncoviscana sequence (Lucassen et al., 2011), intruded by minor Pampean granitic rocks (Escayola et al., 2011), and overlain by Ordovician basins intruded by Famatinian-age magmas (Bahlburg, 1998; Zimmermann and Bahlburg, 2003; Zimmermann et al., 2010). This package is the same as in the rest of the Puna and Eastern Sierras Pampeanas and is part of the same terrane, not the Antofalla block of Loewy et al. (2004). This means that there is a break in continuity of rocks with Laurentian affinities between the Arequipa-Antofalla block and the Western Sierras Pampeanas (Fig. 2). As a consequence some authors argued that the accreted MARA block is missing under the Puna and that the Pampean Orogeny in this region was driven by coupling with the subducting plate (e.g. Lucassen and Becchio, 2003; Lucassen et al., 2011).

Götze and Krause (2002) using gravity, identified a dense body trending NW-SE parallel to and underlying the WEB, from Chile into Argentina. This is a body 400 km long, 100–120 km wide, at depths between 10 and 38 km, which they interpreted as the root zone of the Famatinian arc. Beck et al. (2015) instead suggested that this could be the southern continuation of the Antofalla block. Escayola et al. (2011) took another approach. They suggested that xenocrystic zircons in the Cerro Negro orthogneiss in the WEB (Fig. 2a), could be derived from a magma source region at depth with Laurentian affinity. However, the evidence is ambiguous because the two most significant xenocrystic zircon ages found, at 0.65 and 0.94–1.04 Ga, match those of the Puncoviscana sequence.

While geological or geophysical evidence for a basement of Laurentian affinity under the WEB is ambiguous, there are reasonable grounds to infer its presence below an overthrusted block of Gondwana. The strongest argument is that the rocks with Laurentian affinity that form the Western Sierras Pampeanas lie along strike from those of the Arequipa-Antofalla in Bolivia, with a gap in the WEB. Further to that, and as argued above, if the accreted continental ribbon was limited to the region south of 27°S, we should expect a significant anticlockwise rotation of the WEB, forming an orocline. This would also have impacted on the metamorphism and caused differences between the regions during the subsequent Famatinian cycle. However, this is not the case, suggesting that the colliding MARA block was most likely continuous from the Western Sierras Pampeanas to the Arequipa-Antofalla block, and that it is underthrusted by Gondwana rocks in the Puna. It is possible that this buried section was either thinner or denser than elsewhere, explaining its burial.

If this inference is correct, then the Ordovician Famatinian arc crosses the buried suture between the Eastern Sierras Pampeanas (ESP) and the MARA block under the Puna (Fig. 2) and then follows MARA northwards onto the Antofalla section (Loewy et al., 2004). As a final note on this theme, the general trend of the suture marked by the contact between the MARA block and Gondwana crust follows the general NNW-SSE trend of the ribbon, implying that the regional fabric, dominated by NNE-trending structures, reflects the bulk strain during the Famatinian Orogeny and not the original orientation of the ribbon accretion.

8.7. Comparison to Other Paleozoic Accretionary Orogens

8.7.1. Paleozoic Ross-Delamerian and Lachlan Orogenies of Eastern Australia

While the Pampean and Famatinian cycles were taking place in West Gondwana, at the other end of the Terra Australis Orogen, a similar set of events were taking place in eastern Australia and Antarctica. Here, the Cambrian Ross-Delamerian Orogen was followed by prolonged extension in the Lachlan Fold Belt that ended with the Late Ordovician-Early Silurian (450–440 Ma) contraction of the early Benambran deformation (Collins, 2002b). A number of authors have compared the Paleozoic orogenies of these two regions (Cawood and Buchan, 2007; Foden et al., 2006; Schwartz et al., 2008; Simpson et al., 2003). In both, the sequence of events marked a period of crustal growth and intense recycling and rejuvenation.

The Delamerian Orogen resulted from initiation of a landward subduction under the attenuated continental margin of the Gawler Craton (Foden et al., 2006). It started at 515 Ma and was linked to the Ross Orogen in Antarctica, which started earlier at 540 Ma. They both ended at 490 Ma, defining a cycle broadly contemporaneous with the Pampean cycle. Deformation affected the recently deposited Neoproterozoic-Cambrian passive margin sediments, but also the *syn*-tectonic Cambrian turbidites of the Kanmantoo Group, carrying detrital zircons derived from the active Ross Orogeny. This evolution is similar to that recorded by the Puncoviscana sequence of NW Argentina, except that the Puncoviscana has not been formally divided into a passive margin and a syn-orogenic sequence as discussed above.

At the end of the Delamerian Orogeny, immediately after an intensive period of crustal thickening, the area went into extension, possibly driven by roll-back of a west-verging subduction zone (Fergusson, 2003). This process formed an island arc and a back-arc a few thousand kilometres wide. This period coincides with the Famatinian cycle, and like it, a prolonged Ordovician extension ended with shortening that started at 455 Ma and continued during the Late Ordovician to Early Silurian Benambran Orogeny (Collins, 2002b). The Ross-Delamerian Orogeny provided sediments for the voluminous Cambro-Ordovician turbidite sequences deposited to the east, between the continental margin and the island arc. Like for NW Argentina, the basement of the turbidites has been widely debated (Fergusson, 2003) with only minor exposures of Cambrian basalts, and much discussion based on geophysics, and the nature of magmatism and zircon recycling. A significant difference is that eastern Australia lacks “material accreted from the paleo-Pacific plate” (Glen, 2013) and deformation is controlled by subduction dynamics rather than docking crustal blocks.

8.7.2. Paleozoic Altai Orogeny, Central Asian Orogenic Belt

The 420–360 Ma Devonian-Carboniferous Altai Orogeny, part of the Central Asian Orogenic Belt, represents an accretionary orogeny developed nearly contemporaneously and share many common features with both the Argentinian and Australian Paleozoic orogens at the margin of Gondwana.

The –Altai Orogeny reworked a vast accretionary wedge comprising an Early Paleozoic volcano-sedimentary unit stretching ~1500 km from Russia to central Mongolia. It was associated with the ~1800 km-long Cambro-Ordovician Ikh-Mongol arc system (Janoušek et al., 2018) and represents an accretionary system surrounding Precambrian Zavhan-Baydrag blocks (Jiang et al., 2017) that are part of the Mongolian Collage (recently reviewed by Xiao et al., 2017). The Chinese Altai comprises the voluminous Ordovician turbidites of the Habahe Group, as well as younger volcano-sedimentary sequences (Jiang et al., 2016; Xiao et al., 2009). Like in NW Argentina and eastern Australia, the basement to the Habahe Group is ill-defined, with Precambrian rocks neither exposed nor imaged remotely. Like the Puncoviscana sequence and the Kanmantoo Group, the Habahe Group underwent extensive regional high T – low P metamorphism soon after its deposition, which led to widespread anatexis and formation of Devonian migmatites and granites (Cai et al., 2011; Jiang et al., 2016).

The volcano-sedimentary crust forming the Chinese Altai was deformed during the Late Devonian. Similar to the contemporaneous evolution of eastern Australian, deformation was a result of Pacific-type supra-subduction tectonic switching (Jiang et al., 2016). This changed in the Permian when collision between the Chinese Altai and the Junggar arc occurred.

As a result of the accretionary processes, at the end of all three Paleozoic orogenies, the turbiditic sequences newly deposited in the different continental margins had undergone deformation and high-T – low P anatexis resulting in vast volumes of turbidites being thoroughly metamorphosed, melted and accreted to the continental margin either by the docking of extraneous continental blocks as in the case of NW Argentina, or by changes in subduction zone dynamics, such as in eastern Australia and the Altai.

Summary and Conclusions

The Pampean cycle was characterized by a magmatic arc formed a few hundred kilometres from the trench as a result of a low-angle subduction. It was separated from the trench by a vast fore-arc dominated by Puncoviscana sequence, initially deposited in a passive margin, but subsequently receiving sediments from the growing arc itself. Pampean arc magmatism peaked at 535–530 Ma accompanied by anatexis of the turbidites in the fore-arc which underwent high-T – low-P, reaching locally medium pressures. Fore-arc melting was possibly a result of processes linked to subduction initiation, ridge subduction or asthenosphere invasion below the fore-arc. The nature of the basement to the Puncoviscana sequence is unknown, but it most likely comprised a non-volcanic, hyper-extended continental margin transitioning to oceanic crust, as indicated by the lack of a mafic lower crust and lack of basement exposure.

The cycle ended at ~520 Ma with the docking of the MARA block, the first exotic block of Laurentian affinity to arrive at the subduction zone, intensifying crustal shortening and mountain building, with thrusts propagating into the fore-arc, widening the orogen. Deceleration of convergence caused steepening of the subducting slab and ingression of asthenospheric mantle causing renewed heating and minor mafic magmatism in the fore-arc. The hot crustal environment prevalent at this time, recorded by widespread low-P migmatization, suggests that crustal thickening was modest, and led to a wide rather than tall orogeny.

The arrival of MARA was followed by a tectono-magmatic lull accompanied by extension and renewed marine sedimentation forming Late Cambrian – Early Ordovician basins above the Tilcara unconformity (e.g. Mesón Group quartzites). After another brief phase of shortening and development of the Iruya angular unconformity, the Famatinian cycle started as a result of subduction reorganization.

The Famatinian cycle formed an arc 250–300 km trenchward from the Pampean arc and was initially accompanied by widespread Ordovician marine sedimentation in the form of turbidites similar in nature to the Puncoviscana sequence and recycling these older sediments as well as Pampean arc rocks. This sedimentation occurred during back-arc extension and bi-modal magmatism. The back-arc rift zone, currently varies in width from 100 km in the south to > 250 km in the north, measured from the edge of the Famatinian arc to the eastern side of the Pampean arc. The shoulders of this rift zone are marked by Pampean magmatic and metamorphic rocks found to the east and west of the Famatinian back-arc. After the waning of basaltic magmatism in the back-arc at ~470 Ma, sedimentation and extension were interrupted at ~470–460 Ma, with the start of the shortening Famatinian Oclóyic cycle, a response to the approaching Precordillera or Cuyania block, the second exotic block with Laurentian affinities. Its docking ultimately ended subduction at ca. 440 Ma, marking the end of the 70–60 Ma-long Famatinian cycle that started with extension and ended with shortening.

The Oclóyic cycle shortened the entire back-arc region all the way to the edge of the Pampean arc in the west. This transformed the back-arc into a foreland, and gave rise to some of the thickest thrust zones anywhere. Their trenchward vergence could be a result of the reactivation of trench-verging Pampean fore-arc thrusts. These thrusts typically exhume high-T – low-P, Crd-bearing migmatites, suggesting limited exhumation. The absence of rocks recording higher pressures combined with the 100 to 250 km width of the Famatinian foreland belt, suggest that the orogeny grew wide because it was hot and therefore unable to stack the crust. In which case the potent thrusts acted as flat detachments accommodating large horizontal displacements. A process has also inferred for the Pampean Orogeny.

One of the most significant features of the Famatinian cycle is the widespread peraluminous granitic magmatism resulting from migmatization of the fertile, recently deposited late Puncoviscana sequence. Crustal anatexis generated such large volumes of magma that it has led to the Eastern Eruptive Belt, part of the back-arc, to be considered a second arc. Voluminous crustal melting started during back-arc extension and continued into the shortening phase. Crustal melting also occurred along the arc, leading to intense hybridization of arc magmas, even the most mafic ones marked by their isotopic composition. Intense crustal anatexis resulted from a combination of factors: a) the existence of a fertile, recently deposited package of turbidites, b) inherited heat from a first phase of heating during the Pampean cycle, c) reheating during the Famatinian back-arc extension and minor mafic magmatism, and d) high internal heat production of the sedimentary package boosting temperatures during crustal thickening events. This evolution explains how the two orogens gave rise to paired magmatic belts, and explains why in the Pampean cycle it was the fore-arc that underwent anatexis whereas in the Famatinian cycle it was the back-arc/foreland, giving rise to opposite polarity of the paired belts.

The along-strike continuity between the two sections of the MARA block, the Western Sierras Pampeanas to the south, and the Arequipa-Antofalla block to the north, suggests that this Laurentian block has been thrusted under the Puncoviscana sequence in the Puna Plateau where no rocks with Laurentian affinity are exposed. Its underthrusting in the Puna could be a result of the MARA block being thinner and/or denser in this section than elsewhere. The Precordillera/Cuyania block is also missing north of 29ºS, but the Oclóyic phase, associated with its arrival is as intense in the Puna as in the Sierras Pampeanas suggesting that intense coupling between the subducting slab and the overriding plate must have existed at the time.

In summary, the Paleozoic accretionary orogens of NW Argentina started with the establishment of the low-angle Pampean subduction with a wide and hot fore-arc, which eventually steepened in response to the docking of MARA, establishing the Famatinian magmatic arc 250–300 km trenchwards, accompanied by a wide and hot extensional back-arc, later shortened as the foreland when the Precordillera/ Cuyania block docked at the margin of the continent. At this stage the system was too hot to allow significant crustal stacking giving rise instead to a wide orogen. This process reworked the vast pile of sediments that were metamorphosed into a wide belt of high-grade metasedimentary rocks and granites. Together with their hyper-extended continent-ocean transition, they were plastered back on to the craton margin adding a 500 km-wide section to the continental margin. Thus the Paleozoic orogens of NW Argentina are an important example of continental growth from the intense reworking of a passive margin covered by a large turbidite sequence (Gray et al., 2007) similar to contemporaneous orogens of eastern Australia. It is curious that NW Argentina is today an active region of flat-slab subduction and wide Andean Orogen deformation and more than three decades later we remain intrigued by “the possibility that pre-existing structures of the continental lithosphere may in some way help to modify the geometry of the subducted plate” (Allmendinger et al., 1983).

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