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# 0.3 byr of drainage stability along the Palaeozoic palaeo-Pacific Gondwana margin; a detrital zircon study

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**Abstract:** The palaeo-Pacific margin of Gondwana in the present-day south–central Andes is marked by tectonic activity related to subduction and terrane accretion. We present detrital zircon U–Pb data encompassing the Palaeozoic era in northern Chile and northwestern Argentina. Cathodoluminescence images reveal dominantly magmatic zircon barely affected by abrasion and displaying only one growth phase. The main age clusters for these zircon grains are Ediacaran to Palaeozoic with an additional peak at 1.3–0.9 Ga and they can be correlated with ‘Grenvillian’ age, and the Brasiliano, Pampean, and Famatinian orogenies. The zircon data reveal main transport from the nearby Ordovician Famatinian arc and related rocks. The Silurian sandstone units are more comparable with Cambrian units, with Brasiliano and Transamazonian ages (2.2–1.9 Ga) being more common, because the Silurian deposits were situated within or east of the (extinct) Famatinian arc. Hence, the arc acted as a transport barrier throughout Palaeozoic time. The complete suite of zircon ages does not record the accretions of exotic terranes or the Palaeozoic glacial periods. We conclude that the transport system along the palaeoPacific margin of Gondwana remained stable for *c*. 0.3 byr and that provenance data do not necessarily reflect the interior of a continent. Hence, inherited geomorphological features must be taken into account when detrital mineral ages are interpreted.

**Supplementary material:** U–Pb data, CL images, and detailed geological maps are available at [www.geolsoc.org.uk/ SUP18796.](http://www.geolsoc.org.uk/SUP18796)

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The South American margin of Gondwana was tectonically active during long intervals of the Palaeozoic era (e.g. Bahlburg & Hervé 1997; Chew *et al*. 2007; Hervé *et al*. 2013). Intense magmatism accompanied the evolution of an Ordovician magmatic arc (e.g. Pankhurst *et al*. 1998; Coira *et al*. 1999; Chew *et al*. 2007). After relative tectonic and magmatic quiescence during Silurian to Devonian time (Bahlburg & Hervé 1997), arc magmatism reoccurred during the Carboniferous to Permian periods (e.g. Breitkreuz *et al*. 1992; Lucassen *et al*. 1999; Chew *et al*. 2007). Furthermore, two microplates are assumed to have collided with the central proto-Pacific South American margin during the Palaeozoic era: the Cuyania–Precordillera terrane during Middle to Late Ordovician time (e.g. Ramos *et al*. 1986; Keller 1999; Thomas & Astini 2003) and the Chilenia terrane during Middle Devonian time (e.g. Ramos *et al*. 1986; Willner *et al*. 2011). The two collisions and their timing are still under debate (e.g. Keller 1999; Finney 2007; Gleason *et al*. 2007; Bahlburg *et al*. 2009). During this time the climate varied and included two periods of glaciation during the Late Ordovician to early Silurian and Late Devonian to Late Carboniferous (e.g. Díaz-Martínez & Grahn 2007; Gulbranson *et al*. 2010). These glacial events are also recorded in other parts of Gondwana (e.g. Ghienne *et al*. 2007; Fielding *et al*. 2008; Isbell *et al*. 2008; Van Staden *et al*. 2010). This Palaeozoic evolution of the South American margin of Gondwana makes it highly suitable as a sedimentological tracer for tectonic and climatic changes (see Cawood *et al*. 2012).

Detrital zircon U–Pb age distributions often are used to track tectonic activity in the near or far surroundings of sedimentary basins and to trace continental block configurations (e.g. Augustsson *et al*. 2006; Gerdes & Zeh 2006; Cawood *et al*. 2012). Similarly, variation in zircon data from stratigraphic series in a more or less confined area may reflect tectonic activities, as illustrated by, for instance, Sircombe (1999) and Bahlburg *et al*. (2009). Furthermore, zircon age distributions for syn-glacial sedimentary units may be differentiated from those preceding and following the glaciation (e.g. Xiao *et al*. 2012; Gärtner *et al*. 2014), which is due to changes in the size and position of the catchment area as a cause of the glaciation. Numerous studies on Palaeozoic strata along the present-day south and central Andes correlate sedimentary rocks and sedimentation with different phases of igneous activity (e.g. Hervé *et al*. 2003; Augustsson *et al*. 2006; Chew *et al*. 2007; Willner *et al*. 2008; Bahlburg *et al*. 2009), but these studies concentrate on limited age intervals along different parts of the Gondwana margin. This study encompasses the complete Palaeozoic era along this margin.

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The goals of this study are to (1) track the geomorphological evolution for the complete Palaeozoic era along the palaeo-Pacific margin in the present-day south–central Andes and (2) connect it to the tectonic evolution in the area. On a broader scale, (3) we question the validity of using zircon data as a tracer for tectonic variations through time without consideration of the geomorphological surroundings of the studied basins. We show how the presence of an erosional barrier may lead to zircon age variations without imposing tectonic activity and how the erosional barrier causes age distributions to remain stable despite variations in tectonic activity. (4) We also test the possibility that climatic effects may affect age distributions (see Cawood *et al*. 2012). The goals are reached by the study of detrital zircon age distributions of Palaeozoic sedimentary units along the present-day south–central Andes in northern Chile and northwestern Argentina between 22 and 25°S. We use detrital U–Pb zircon age distributions of Cambrian (Augustsson *et al*. 2011) and Ordovician to Permian sandstone to identify their source areas. The data are combined with cathodoluminescence (CL) and backscattered electron (BSE) imaging to reveal the magmatic or metamorphic origin of the zircon grains and their abrasional regime (zoning, shape, degree of roundness; e.g. Corfu *et al*. 2003; Augustsson *et al*. 2006).

## Geological setting

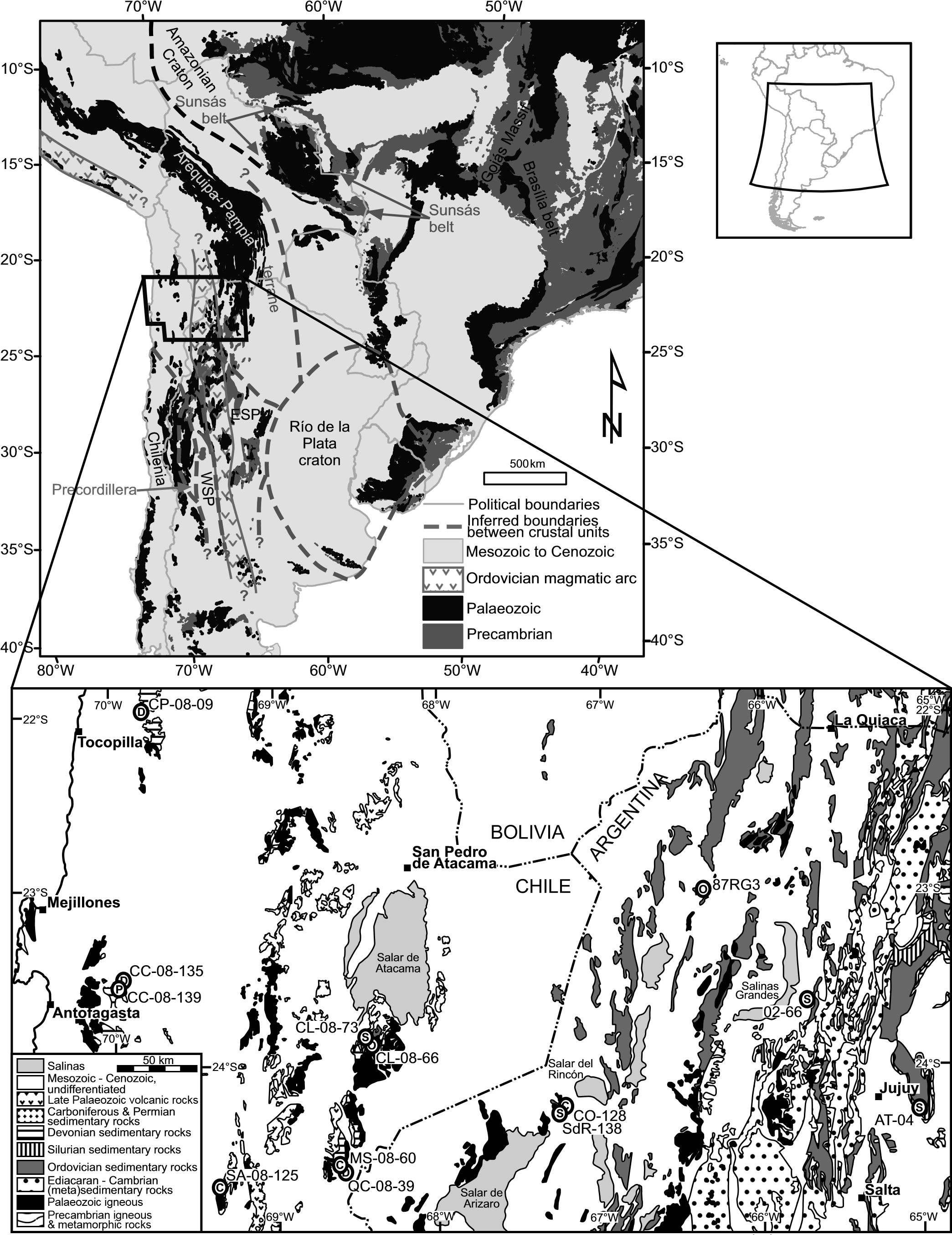
The Ediacaran to Palaeozoic development of the south–central Andean continental margin started after the amalgamation of Gondwana during Neoproterozoic time and the corresponding Brasiliano (Pan-African) orogeny between *c*. 750 and 500 Ma, a phase of intense granitoid magmatism and metamorphism in South America (e.g. Basei *et al*. 2010; Saalmann *et al*. 2011). In northwestern Argentina, the Early Cambrian Pampean orogeny took place along the Gondwanan margin (Rapela *et al*. 1998). It can be traced southwards at least into northern Patagonia, and possibly even further south (Pankhurst *et al*. 2003; Rapalini *et al*. 2013). The Pampean orogeny was followed by intrusion of Cambrian, postorogenic granitoid bodies such as the Santa Rosa de Tastil and Cañaní batholiths (Augustsson *et al*. 2011; Escayola *et al*. 2011). Subsequently, the magmatic arc of an Ordovician orogeny (Fig. 1) possibly extended from Venezuela in the north to southern Patagonia in the south (Pankhurst *et al*. 2003; Chew *et al*. 2007; Horton *et al*. 2010; Rapalini *et al*. 2013). In the south–central Andes, the corresponding Famatinian orogeny had been initiated by Late Cambrian time and arc magmatism continued at least until middle Ordovician time (Fig. 2; e.g. Pankhurst *et al*. 1998; Sims *et al*. 1998; Mulcahy *et al*. 2007; Rapela *et al*. 2007). The orogeny was accompanied by low- to high-grade metamorphism (e.g. Damm *et al*. 1990; Lucassen *et al*. 2011). The following relative tectonic and metamorphic quiescence is attributed to a subductionfree period, which lasted until the Carboniferous period in the south–central Andes (e.g. Bahlburg & Hervé 1997; Bahlburg *et al*. 2009) and may have extended from northern Peru to southern Chile (e.g. Augustsson *et al*. 2006; Chew *et al*. 2007). Indications exist for post-orogenic Devonian igneous activity in the Eastern Sierras Pampeanas (Grosse *et al*. 2009, and references therein). At the same time arc magmatism took place further SE in northern Patagonia (Pankhurst *et al*. 2006; Hervé *et al*. 2013). The reinitiation of subduction in the south–central Andes is marked by increased intrusive and extrusive calc-alkaline activity particularly around the Carboniferous–Permian transition, but possibly already during Early Carboniferous time (Fig. 2; Breitkreuz *et al*. 1992; Lucassen *et al*. 1999; Bahlburg *et al*. 2009). The Permian period was a time of intense igneous activity in northern Chile and northwestern Argentina (e.g. Donato & Vergani 1985; Damm *et al*. 1990; Breitkreuz & Zeil 1994; Lucassen *et al*. 1999). Furthermore, both in southern and central Chile and Argentina and in Peru Permian calc-alkaline igneous rocks occur (e.g. Carlier *et al*. 1982; Llambías 1999; Pankhurst *et al*. 2006; Kleiman & Japas 2009; Rocha-Campos *et al*. 2011). At least in the south they have been related to an igneous arc position (Llambías 1999; Kleiman & Japas 2009).

The different plate-tectonic settings resulted in a variety of depositional regimes in the study area. Late Neoproterozoic to Palaeozoic deposition starts with turbiditic slope sediment of the Ediacaran to Early Cambrian Puncoviscana Formation and equivalents, extending from southernmost Bolivia to central Argentina (Fig. 1; e.g. Ježek 1990; Augustsson *et al*. 2011). It has been interpreted to represent a peri-arc basin (Kraemer *et al*. 1995; Keppie & Bahlburg 1999; Zimmermann 2005; Escayola *et al*. 2011) or passive margin deposition (Ježek 1990; Kraemer *et al*. 1995; PiñánLlamas & Simpson 2006) and was metamorphosed, deformed, and uplifted during the Pampean orogeny (Do Campo *et al*. 1994; Piñán-Llamas & Simpson 2006; Rapela *et al*. 2007). The Puncoviscana Formation is overlain unconformably by the tidaldominated Middle to Upper Cambrian siliciclastic Mesón Group, which extends from southern Bolivia to northwestern Argentina (Kumpa & Sanchez 1988; Augustsson *et al*. 2011). This unit occurred within an extensional basin either along a passive continental margin or in an intracontinental setting (e.g. Dalziel & Forsythe 1985; Sánchez & Salfity 1999).

The emergence of the Famatinian arc resulted in the formation of the marine Lower Ordovician Cordón de Lila Igneous and Sedimentary Complex (CISL, Complejo Ígneo-Sedimentario del Cordón de Lila; Niemeyer 1989) in northern Chile in a forearc to intra-arc setting (e.g. Niemeyer 1989; Zimmermann *et al*. 2010). Siliciclastic rocks are particularly common in the lower part of the CISL. Here, turbiditic sandstone is intercalated with rhyolitic bodies and tholeiitic basalt (e.g. Breitkreuz *et al*. 1989; Niemeyer 1989). Bimodal volcanic activity here and further east in Argentina represents the volcanic arc situated on a thin continental crust (Breitkreuz *et al*. 1989; Zimmermann *et al*. 2010). A retro-arc foreland basin developed to the east during or soon after peak volcanic activity during Middle Ordovician time (Bahlburg 1991). The Ordovician basin extended north into Bolivia (e. g., Egenhoff 2007), with the Ordovician successions being deformed during the Famatinian orogeny (e.g. Niemeyer 1989). Metamorphism took place *c*. 470–420 Ma (Middle to Late Ordovician time) in northern Chile and northwestern Argentina (Lucassen *et al*. 2011). In this study, deposits of the foreland basin are represented by the Middle Ordovician Lower Turbidite System of the Puna Turbidite System in northwestern Argentina (Bahlburg 1990; Bahlburg *et al*. 1990), which is mainly composed of marine, turbiditic sandstone and pelite (Bahlburg 1990).

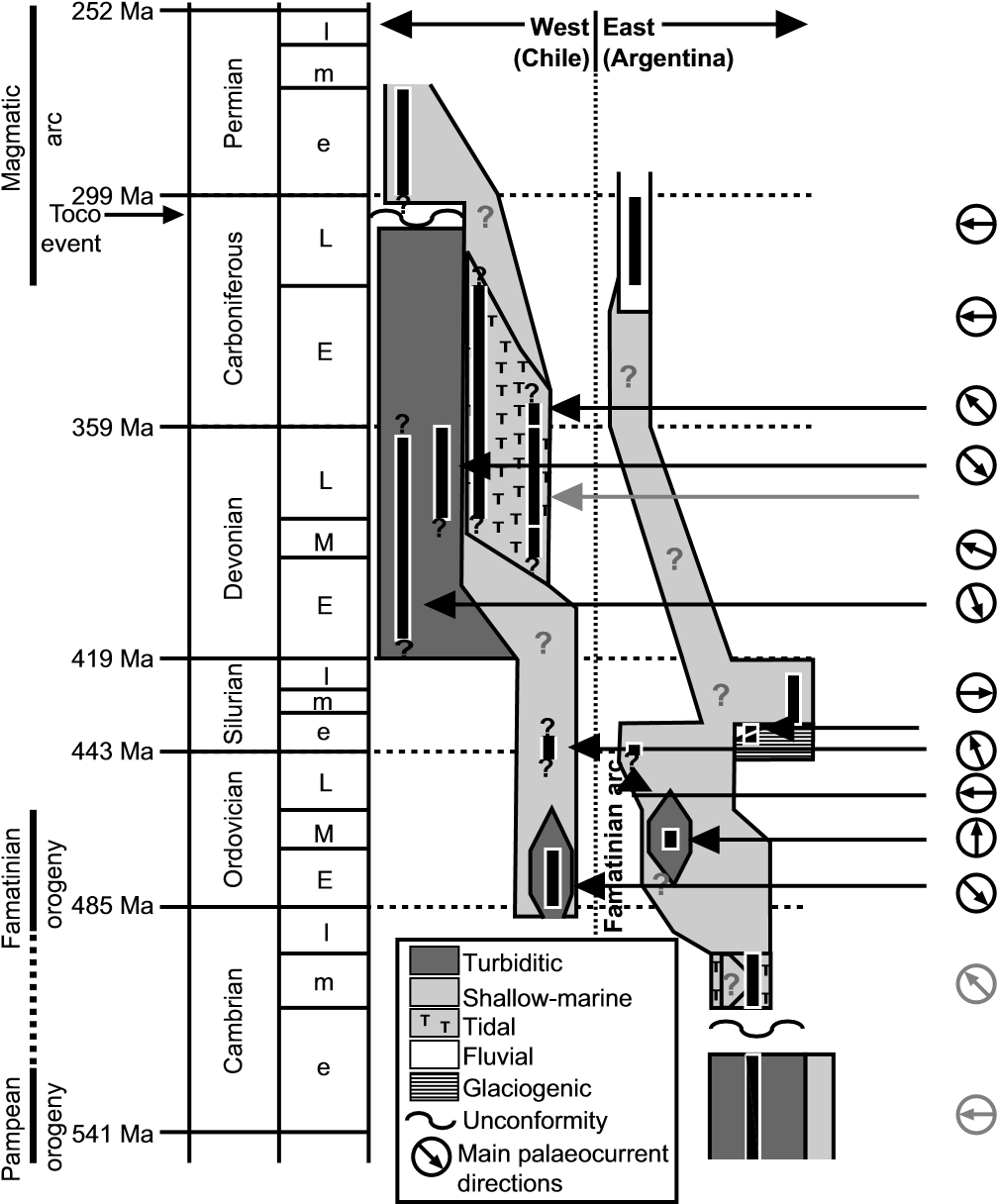
The development of a passive continental margin caused stable shelf conditions for shallow-water sedimentation during Silurian time (e.g. Bahlburg & Hervé 1997; Astini *et al*. 2004; Niemeyer *et al*. 2010). The Quebrada Ancha Formation at Cordón de Lila in Chile within the belt of Ordovician magmatic rocks marks the westernmost occurrence of Silurian successions in northern Chile and northwestern Argentina (Fig. 1). It consists mainly of shallowmarine, sub-tidal, quartz-rich sandstone with a thick conglomeratic base (Niemeyer *et al*. 2010). It is coeval with the Salar del Rincón Formation in westernmost Argentina, a shallow-marine siliciclastic unit with quartz-rich sandstone (Donato & Vergani 1985; Benedetto & Sánchez 1990) and with a depositional age around the Ordovician–Silurian transition (Isaacson *et al*. 1976; Rubinstein & Vaccari 2004). In northwestern Argentina the Late Ordovician to Early Silurian glacial period is documented by the existence of the glacial lower member of the Zapla Formation (e.g. Grahn & Gutiérrez 2001; Astini *et al*. 2004) for which coeval equivalents exist in both Bolivia and Peru (e.g. Díaz-Martínez & Grahn 2007). The glacial deposits are succeeded by the alluvial to fluvial upper member of the Zapla Formation and with the conformably overlying shallow-marine, sub-tidal Lipeón Formation in the Eastern Cordillera (e.g. Andreis *et al*. 1982; Moya & Monteros 1999). The Silurian units in Argentina were deposited on the extinct Famatinian arc, on its eastern fringe, or to the east of the former arc.

During Devonian time, kilometre-thick marine, turbiditic sequences formed along the tectonically passive margin of Gondwana, west of the former Famatinian arc (Bahlburg &



**Fig. 1.** Geological framework of central South America and the study area with sampling sites (circles with letters indicating the age). Only crustal units that are relevant for this study are shown. The thick, dotted lines mark assumed boundaries between crustal units. The terranes are drawn in their present positions; only Precordillera and Chilenia possibly were accreted during Palaeozoic time. It should be noted that disagreement exists on whether the Western Sierras Pampeanas belongs to the Cuyania–Precordillera terrane or to the Arequipa–Pampia terrane (see Kay *et al*. 1996; Thomas *et al*. 2012) and that the Pampean orogenic belt extends southwards from the Eastern Sierras Pampeanas into Patagonia (Pankhurst *et al*. 2003; Rapalini *et al*.

2013). The Puncoviscana Formation and the Mesón group are represented by the Ediacaran to Cambrian units in the lower map. ESP, Eastern Sierras Pampeanas; WSP, Western Sierras Pampeanas; O, Ordovician; S, Silurian; D, Devonian; C, Carboniferous; P, Permian. The general map is modified after CGMW (2001) and includes information from Rapela *et al*. (2007, 2010), Bahlburg *et al*. (2009) and Pankhurst *et al*. (2014). The detailed map is modified after Reutter *et al*. (1994).



Breitkreuz 1993; Bahlburg & Hervé 1997; Niemeyer *et al*. 1997*a*). The turbidite systems extended between 22 and 32°S. In this study they are represented by the Sierra del Tigre and El Toco formations. They partly are coeval with platform deposits to the east. Sandstone and pelite of the Middle Devonian to Lower Carboniferous Zorritas Formation in easternmost Chile represent the transition from tidal to deltaic conditions (Niemeyer *et al*. 1997*b*; Isaacson & Dutro 1999). Towards the end of the Carboniferous period, the deposits in the Cordillera de la Costa in Chile (El Toco and Sierra del Tigre formations) were affected by folding during the Toco tectonic event (Bahlburg & Hervé 1997; Niemeyer *et al*. 1997*a*).

The Sierra Argomedo Formation of poorly defined Devonian to Carboniferous age represents the shallow-marine part of a basin of pre- to post-Toco event age (Breitkreuz 1985; Marinovic *et al*. 1995). Detrital zircon dating indicates an Early Carboniferous maximal depositional age for the upper part of the Sierra Argomedo Formation (this study). It mainly contains pelite and some sandstone with metre-thick volcanic or subvolcanic intercalations that are interpreted as flows or sills (Breitkreuz 1985; Marinovic *et al*. 1995).

After emergence of the Carboniferous–Permian arc, a continental depositional regime prevailed in westernmost Argentina with the fluvial Cerro Oscuro Formation (Donato & Vergani 1985; Galli *et al*. 2010). It is separated from the Late Ordovician to early Silurian Salar del Rincón Formation by an angular unconformity and is composed mainly of red sandstone and conglomerate (Donato & Vergani 1985). An angular unconformity, caused by the Toco event, also is present between the Devonian Sierra del Tigre Formation and the lower Permian Cerro de Cuevitas Formation in western Chile (Niemeyer *et al*. 1997*a*). The Cerro de Cuevitas Formation is composed of shallow-marine, coastal mudstone and sandstone in the lower part and of limestone and tuff layers in the upper part (Niemeyer *et al*. 1997*a*).

## Methods

Sixty to 170 detrital zircon grains with average very fine to fine sand size (60–130 µm; Table 1) from each of 13 sandstone samples

Cerro de Cuevitas Fm. (CC-08-139)

Cerro Oscuro Fm. (CO-128)

Sierra Argomedo Fm. (SA-08-125)

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| --- | --- |
| Zorritas Fm. (Upper Mbr; MS-08-60)  El Toco Fm. (CP-08-09)  Zorritas Fm. (Middle Mbr)  Zorritas Fm. (Lower Mbr; QC-08-39)  Sierra del Tigre Fm. (CC-08-135)  Lipeón Fm. (AT-04)  Zapla Fm. (02-66)  Quebrada Ancha Fm. (CL-08-73)  Salar del Rincón Fm. (SdR-138)  Lower Turbidite System (Puna  (Turbidite System; 87RG3) Complejo Ígneo-Sedimenario del  Cordón de Lila (CISL; CL-08-66)  Mesón Group  Puncoviscana Fm. | **Fig. 2.** Schematic stratigraphy at 22–25°S including relative east–west positions of sampling sites, main palaeocurrent directions and depositional setting of those sedimentary units in the study area that are referred to in the text (black columns with white frames). Question marks represent age uncertainties. Names of studied units are given in black. References to these units are given in Table 1. Additional references: Puncoviscana Formation: Ježek (1990), Augustsson *et al*. (2011); Mesón Group: Kumpa & Sanchez (1988), Augustsson *et al*. (2011); lower member of the Zapla Formation: Astini *et al*. (2004); Middle Member of the Zorritas Formation:  Rubinstein *et al*. (1996), Niemeyer *et al*.  (1997*b*); time scale from Gradstein *et al*.  (2012). |

from Ordovician to Permian units (Table 2) were randomly selected for CL and BSE imaging, and for laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) analysis at the Institut für Mineralogie, Münster, using a Thermo-Fisher Element 2 coupled to a New Wave UP193HE ArF Excimer laser. The samples were crushed manually in a metal mortar with a flat bottom and without using grinding movements. The crushed material was sieved to a grain size of <250 µm. Heavy minerals were separated using a sodium polytungstate liquid with a density of 3.0 g cm−3. A random selection of the zircon grains were handpicked and transferred to 1 inch epoxy mounts. All zircons in the picked fraction of each sample were mounted. No selection was made regarding grain size, shape, or colour.

CL and BSE imaging was carried out using a JEOL 840 with SE detector, a Centaurus BSE/CL-detector system and an Oxford INCA EDX system. The images were used to investigate morphology and growth zoning in the zircon grains, particularly to find suitable ablation spots for the U–Pb analysis, preferably in zircon rims and in areas free of inclusions. Oscillatory and sector zoning was used as an indication for a magmatic origin. Irregularly zoned and homogeneous grains were considered metamorphic (see Corfu *et al*. 2003). The roundness scheme of Powers (1953) was used in a simplified manner with the classes euhedral (comparative to very angular and angular), subangular to subrounded and rounded (comparative to rounded and well-rounded). Grains with length/ width ratios of 1.3–1.8 are considered oval. Grains with higher and lower ratios are defined as elongated and round, respectively (Fig. 3).

The LA-ICP-MS setup and analytical procedure of Kooijman *et al*. (2012) were followed with the exception that total ablation time was 55 s, which included 20 s with the shutter closed to measure the background. Each spot was pre-ablated with three laser shots using a 45 µm beam to remove surface common Pb. Washout times were set to 45 s (with the exception of no pre-ablation and 90 s wash-out for measurements 01 to 40 in SdR-138). A spot size of 35 or 25 µm, depending on the size of the zircon zones, was used. We used a measurement scheme in which five unknowns

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| **Table 1.** *Zircon characteristics*  Roundness (%) Mean grain  Sample  *n*  Growth zoning (%)  Grain shape (%)  length (µm) Subangular to   |  |  |  |  |  |  |  |  |  |  |  | | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | |  |  | Magmatic | Metamorphic | Euhedral | subrounded | Rounded | Elongated | Oval | Round |  | | CL-08-66 | 100 | 90 | 10 | 8 | 60 | 32 | 24 | 48 | 28 | 130 | | 87RG3 | 90 | 94 | 6 | 19 | 66 | 15 | 36 | 46 | 18 | 110 | | SdR-138 | 49 | 98 | 2 | 20 | 78 | 2 | 47 | 35 | 18 | 90 | | CL-08-73 | 109 | 82 | 18 | 25 | 72 | 3 | 23 | 49 | 28 | 70 | | 02-66 | 66 | 83 | 17 | 8 | 75 | 17 | 27 | 49 | 24 | 80 | | AT-04 | 40 | 75 | 25 | 3 | 92 | 5 | 28 | 40 | 32 | 130 | | CC-08-135 | 124 | 92 | 8 | 21 | 68 | 11 | 28 | 52 | 20 | 80 | | QC-08-39 | 123 | 93 | 7 | 19 | 62 | 19 | 33 | 38 | 29 | 110 | | CP-08-09 | 108 | 93 | 7 | 28 | 64 | 8 | 29 | 45 | 26 | 70 | | SA-08-125 | 140 | 86 | 14 | 21 | 70 | 9 | 30 | 45 | 25 | 90 | | MS-08-60 | 83 | 84 | 16 | 17 | 70 | 13 | 35 | 44 | 21 | 60 | | CO-128 | 33 | 85 | 15 | 24 | 61 | 15 | 42 | 33 | 24 | 120 | | CC-08-139 | 124 | 81 | 19 | 10 | 68 | 22 | 25 | 48 | 27 | 80 | |

were bracketed by two standards to monitor instrumental mass bias and elemental fractionation induced by the laser. GJ-1 zircon (Jackson *et al*. 2004) was used as reference material and the 91500 zircon (Wiedenbeck *et al*. 1995) was measured as unknown to monitor precision and accuracy. Reproducibilities are 1.7% for 207Pb/206Pb and 2.6% for 206Pb/238U for the measuring period. They correspond to values obtained by Kooijman *et al*. (2012).

An in-house Excel® spreadsheet (Kooijman *et al*. 2012) was used to process the data. Common-lead correction was applied in cases for which 206Pb/204Pb was less than 2000. Ages with a discordance of more than 10% were not used in data plots and summary tables or for interpretation. 206Pb/238U dates were used for ages <1000 Ma, whereas 207Pb/206Pb dates were used for ages >1000 Ma. Reported uncertainties are always given at the 2σ level. Of the ages determined, 50–90% were considered concordant; that is, at least 66 ages per sandstone (with the exceptions of SdR-138, AT-40, and CO-128, with 49, 40, and 33 concordant ages, respectively; Table 3). Zircon grains may display ages that are too young owing to Pb loss after deposition (Nelson 2001). Therefore, only age clusters (i.e. no single ages) are considered significant for the interpretation. For further evaluation of the data, Isoplot 3.75 of Ludwig (2012) was used. This includes calculations of weighted average ages for the main zircon age accumulations. The number of grains used for each of the calculations is given in parentheses in the text, as well as in Table 3.

## Results

### Morphology and zoning

Zircon grains having oscillatory and sector zoning, typical for magmatic growth, are dominant in all sandstone samples (75– 98%; Table 1) and particularly in the Ordovician Puna Turbidite

System (87RG3), the Silurian Salar del Rincón Formation (SdR138), and the Devonian units (CC-08-135, QC-08-39, CP-08-09; >90%). Most grains are subangular to subrounded (60 to >90% in single samples). Euhedral zircon grains make up 8–28%, with the largest amounts (24–28%) in the Upper Ordovician to lower Silurian Quebrada Ancha Formation (CL-08-73), the Devonian El Toco Formation (CP-08-09) and the Carboniferous Cerro Oscuro Formation (CO-128; Table 1). The lowest amounts (up to 10%) occur in the Ordovician CISL (CL-08-66), the Silurian Lipeón and Zapla formations (AT-04, 02-66) and the Permian Cerro de Cuevitas Formation (CC-08-139). Rounded zircon grains make up only 9–15% in most samples (Table 1). Grains from the Devonian Sierra del Tigre and the Permian Cerro de Cuevitas formations from the same sampling locality are particularly similar: they have the same mean length of 80 µm and are dominated by magmatic (>80%), subangular to subrounded (70%) and oval grains (50%), but the younger unit has a slightly higher proportion of metamorphic, rounded and round grains. Most grains in all analysed sandstone samples display only one growth zoning regardless of the roundness of the grains. No systematic correlation of morphology or zoning with grain size was found. Hence, few stratigraphic and geographical trends were recognized. A negative correlation occurs between zircon age and the amount of euhedral grains (>20% for ages <520 Ma; <10% for ages >1.8 Ga).

### U–Pb ages

Of the zircon grains with concordant ages, 89% have 2σ age uncertainties of 5% or below, equivalent of a maximum of 25 Ma for grains dated at 500 Ma. All analysed sandstone samples contain 45–80% zircon grains with Ediacaran to Palaeozoic age that can be related to the Brasiliano (*c*. 750–500 Ma; e.g. Basei *et al*. 2010), Pampean (Early Cambrian, probably ending at 520–490 Ma; e.g.

Rapela *et al*. 1998), and Famatinian (probably starting at 510– 490 Ma, ending at *c*. 460 Ma; e.g. Pankhurst *et al*. 1998) magmatic phases and to post-Famatinian zircon growth (Fig. 4, Table 3). The highest abundances were recorded for the Devonian unit of the Zorritas Formation (QC-08-39) and the Carboniferous Cerro Oscuro Formation (CO-128) and the lowest for the Silurian Zapla Formation (02-66).

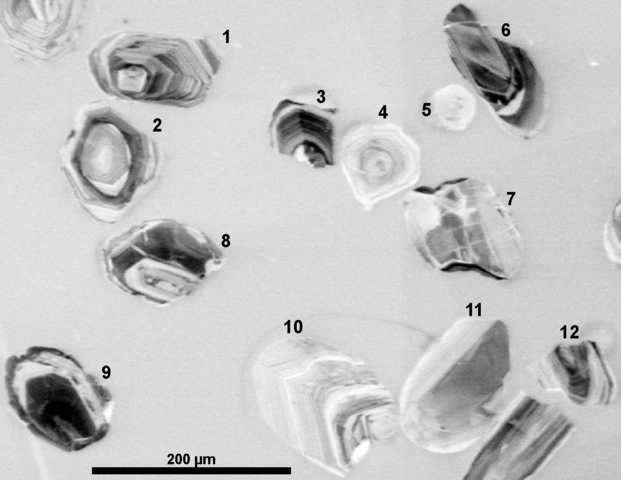
The most common pre-Brasiliano ages are ‘Grenvillian’

(1.3–0.9 Ga; Fig. 4). The abundance is 10–30% in all samples (with the exception of AT-04 from the Silurian Lipeón Formation, with only one grain). The highest values occur for the Devonian Sierra del Tigre Formation and the Permian Cerro de Cuevitas Formation (CC-08-135, CC-08-139). Most spectra have a distinct peak at *c*. 1.2 or *c*. 1.1 Ga (*n* = 3–8 were used for the age calculations). Exceptions are the latest Ordovician to Silurian Salar del Rincón, Quebrada Ancha, and Lipeón formations (SdR-138, CL-08-73, AT-04). Sandstone of the mainly Silurian units to the east in Argentina (Salar del Rincón, Zapla, and Lipeón formations; SdR-138, 02-66, AT-04) also contains 10–20% Transamazonian-age zircon grains (2.2–1.9 Ga); the Salar del Rincón Formation sandstone (SdR-138) has a pronounced peak at *c*. 2.1 Ga (*n* = 5; Figs 4 and 5). In other sandstone samples Transamazonian ages are rare or absent. Other age clusters occur at 900–730 Ma (particularly in the Ordovician Puna Turbidite System, the Silurian Lipeón Formation and the Carboniferous part

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| **Table 2.** *Unit descriptions and sampling locations*   |  |  |  |  |  |  |  |  |  |  | | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | | Sample | Geological unit | Age | Dating objects | Thickness | Depositional setting | Current directions | Reference | Sampling location | Coordinates | | CL-08-66 | CISL (Cordón de Lila Igneous  and Sedimentary  Complex), lower part | Early Ordovician (to *c*. 470 Ma) | Zircon in lava | 3 km | Marine,  turbiditic, forearc to intra-arc | SE | 1 | Cordón de Lila, Chile | 23°50’05”S, 68°22’10”W | | 87RG3 | Puna Turbidite  System: Lower  Turbidite System, lower part | (Dapingian to) early Darriwilian  (Middle  Ordovician) | Graptolites | *c*. 3 km | Deep-marine, turbiditic, retro-arc | N | 2 | Río Grande, Argentina | *c*. 23°02’S,  66°26’W | | SdR-138 | Salar del Rincón  Fm | Late Hirnantian to  Rhuddanian (latest  Ordovician to earliest Silurian) | Spores, brachiopods | 120 m | Shallowmarine | W | 3, 4, 5 | Cerro Oscuro, Argentina | 24°08’43”S, 67°15’04”W | | CL-08-73 | Quebrada Ancha  Fm, middle mbr, lower part | Hirnantian to  Aeronian (latest  Ordovician to early Silurian) | Brachiopods | 67 m | Shallowmarine, subtidal | NW–N | 6 | Cordón de Lila, Chile | 23°46’29”S, 68°24’11”W | | 02-66 | Zapla Fm, upper mbr | Aeronian to early Chitinozoans  Telychian (early Silurian) | | <39 m | Fluvial channels of  coastal alluvial fans | – | 7, 8 | Los Colorados, Argentina | *c*. 23°42’S,  65°41’W | | AT-04 | Lipeón Fm, lower Late Llandovery to  part Ludlow (early to  late Silurian) | | Chitinozoans, bivalves | 1.6 km | Shallowmarine, subtidal | E | 5, 7, 9 | Zapla, Argentina | 24°14’25”S, 65°04’55”W | | CC-08-135 | Sierra del Tigre Fm, Devonian uppermost part | | Brachiopods | 2 km | Open-marine, turbiditic | S–SE | 10, 11 | Cerro de Cuevitas, Chile | 23°31’45”S, 69°57’37”W | | QC-08-39 | Zorritas Fm, Lower Eifelian to early  Mbr, upper part Givetian (Middle  Devonian) | | Brachiopods | 1.3 km | Shallowmarine, tidal | NW–W | 12, 13 | Quebrada 24°35’35”S,  Chucalaque, 68°35’04”W Chile | | | CP-08-09 | El Toco Fm Late Devonian | | Plants | ≥2.3 km | Deep-marine  shelf, turbiditic | S–E | 11, 14 | Cerro Puntillas, 21°59’10”S,  Chile 69°48’25”W | | | SA-08-125 | Sierra Argomedo Late Devonian  Fm, upper part to Early  Carboniferous | | Zircon grains, bivalves, intrusion | ≥1.2 km | Shallowmarine, shoreface | W | 15 | Sierra Aspera, 24°43’24”S,  Chile 69°21’00”W | | | MS-08-60 | Zorritas Fm, Upper Early Tournaisian  Mbr, lower part (Early  Carboniferous) | | Brachiopods | 40 m | Marine, deltaic distributary  channels | NW | 12, 16 | Quebrada Agua 24°33’35”S,  Dulce/Zorra, 68°39’00”W Chile | | | CO-128 | Cerro Oscuro Fm Bashkirian  to middle  Gzehlian (Late  Carboniferous) | | Plants | 310 m | Braided river, retro-arc | W | 3, 17 | Cerro Oscuro, 24°08’43”S,  Argentina W67°15’04”W | | | CC-08-139 | Cerro de Cuevitas (Latest  Fm, lower part Carboniferous to) early Permian | | Invertebrates | 460 m | Shallowmarine, coastal | – | 10 | Cerro de 23°31’49”S, Cuevitas, Chile 69°57’31”W | |   References: 1, Niemeyer (1989), Zimmermann *et al*. (2010); 2, Bahlburg (1990), Bahlburg *et al*. (1990); 3, Donato & Vergani (1985); 4, Isaacson *et al*. (1976), Benedetto &  Sánchez (1990), Rubinstein & Vaccari (2004); 5, Abre *et al*. (2005); 6, Navarro *et al*. (2006), Niemeyer *et al*. (2010); 7, Grahn & Gutiérrez (2001); 8, Moya & Monteros (1999); 9, Andreis *et al*. (1982), Aceñolaza *et al*. (1999), Malanca *et al*. (2010); 10, Niemeyer *et al*. (1997*a*); 11, Bahlburg & Breitkreuz (1993); 12, Niemeyer *et al*. (1997*b*); 13, Boucot *et al*. (2008); 14, Breitkreuz (1986), Moisan *et al*. (2011); 15, Breitkreuz (1985), Marinovic *et al*. (1995), this study; 16, Isaacson & Dutro (1999); 17, Galli *et al*. (2010). |

of the Zorritas Formation; 87RG3, AT-04, MS-08-60; 10% each) and 1.9–1.3 Ga (particularly in the Ordovician CISL, the Silurian Zapla Formation and the Permian Cerro de Cuevitas Formation: CL-08-66, 02-66, CC-08-139; 10–12% each), but without distinct peaks (Fig. 4).

Regarding the younger ages, Pampean and Famatinian age peaks occur at *c*. 520 and *c*. 490 Ma in the Ordovician units (*n* = 5–7; CISL and Puna Turbidite Complex; CL-08-66, 87RG3; Fig. 4, Table 3). A peak at *c*. 640 Ma for the CISL marks a higher abundance of Brasiliano ages than for the Puna Turbidite Complex. The mainly Silurian units from Argentina (Salar del Rincón, Lipeón, and Zapla formations; SdR-138, AT-04, 02-66) have distinct Brasiliano-age peaks at *c*. 680, *c*. 630 or 620, and 600–570 Ma (*n* = 3–7; Fig. 4, Table 3). Pampean and Famatinian ages occur but are less common. The coeval deposits to the west in Chile (Quebrada Ancha Formation; CL-08-73) are more similar to the Ordovician sandstone samples with a Pampean or Famatinian age peak at *c*. 510 Ma (*n* = 7). They also contain abundant zircon grains formed from synsedimentary igneous activity (peak at *c*. 440 Ma; *n* = 6 were used for the age calculation; Fig. 4, Table 3). The Devonian deposits (Sierra del Tigre, El Toco, and Zorritas formations; CC-08-135, CP-08-09; QC-08-39) are dominated by Pampean to Famatinian age clusters at 530–510 and *c*. 470 Ma, as well as at 440–430 Ma, (*n* = 4–10; Fig. 4, Table 3). The Sierra del Tigre Formation also contains Brasilianoage zircon grains with a main peak at *c*. 610 Ma (*n* = 7; Fig. 4, Table 3). Similar to the Devonian units, the Carboniferous upper part of the Sierra Argomedo Formation in Chile (SA-08-125) is dominated by Pampean or Famatinian ages with a main age cluster at *c*. 490 Ma (*n* = 7; Fig. 4, Table 3). Brasiliano and post-Famatinian ages occur with peaks at *c*. 580 and *c*. 380 Ma (*n* = 12 and 4, respectively). The youngest age accumulation, which represents a maximal depositional age, is at 323 ± 5 Ma (*n* = 3, mean square weighted deviation = 0.52, probability = 59%). The age population of the Carboniferous part of the Zorritas Formation in Chile (MS-08-60)



**Fig. 3.** Cathodoluminescence image from QC-08-39 illustrating the variation in zircon grain shape, degree of roundness, and zoning. The colours are inverted to illustrate the zoning more clearly. Grain 1 is euhedral, grains 2–9 and 12 are subangular to subrounded, and grains 10 and 11 are rounded. Grains 1 and 6 are elongated, grains 8–11 are oval, and all others are round. All grains are oscillatory zoned, except for grains 6, 9 and 11, which have an irregularly zoned or homogeneous, thick rim.

is similar but with different distribution. Brasiliano, Silurian, and Devonian ages are most frequent with main peaks at *c*. 620, *c*. 420 and *c*. 380 Ma (*n* = 4–6; Fig. 4, Table 3). The proportion of Pampean or Famatinian zircon ages is smaller (peak at *c*. 510 Ma, *n* = 4). The age spectrum of the Argentine Cerro Oscuro Formation (CO-128; *n* = 33) is dominated by Famatinian ages with main peaks at *c*. 500

**Table 3.** *Summary of zircon U–Pb data*

and *c*. 470 Ma (*n* = 3 and 4; Fig. 4, Table 3). The age spectrum for the Permian unit (Cerro de Cuevitas Formation; CC-08-139) is dominated by Brasiliano (*c*. 600 Ma, *n* = 6) and Famatinian peaks (*c*. 500 and *c*. 470 Ma; *n* = 7 and 5; Fig. 4, Table 3). Hence, the spectrum is rather similar to that from the Devonian Sierra del Tigre Formation (CC-08-135) from the same sampling locality in the Cordillera de la Costa in Chile. The youngest age peaks always are in accordance with the estimated depositional ages or they are older (Table 3).

## Implications

The large abundance of Late Cambrian to Ordovician ages implies that the Famatinian magmatic arc acted as a major source for detrital material throughout Palaeozoic time (Figs 1 and 6). The age peaks correlate with the timing for Famatinian activity (Table 3), because both peak magmatic activity and medium- to high-grade metamorphism were reached at 490–460 Ma in the Puna region of northwesternmost Argentina (Lucassen *et al*. 2011; Hongn *et al*. 2014), in the Sierras Pampeanas further south (Fig. 1; e.g. Pankhurst *et al*. 2000; Astini & Dávila 2004; Dahlquist *et al*. 2008), and in the equivalent arc in Peru to the north (e.g. Chew *et al*. 2007). The time frame is consistent with graptolite-stratigraphic evidence from NW Argentina (Erdtmann, cited by Breitkreuz 1986; Bahlburg *et al*. 1990; Zimmermann *et al*. 1999). Furthermore, Pampean to post-Pampean ages can be explained by magmatic bodies of equivalent age along the Famatinian arc in northwesternmost Argentina (see Augustsson *et al*. 2011; Escayola *et al*. 2011). Hence, short eastern and western transport from the arc was dominant even long after its magmatic activity ceased. This is supported by the large amount of magmatic zircon grains that dominantly display only one growth zone and the significant amount of euhedral grains that correlate with young ages.

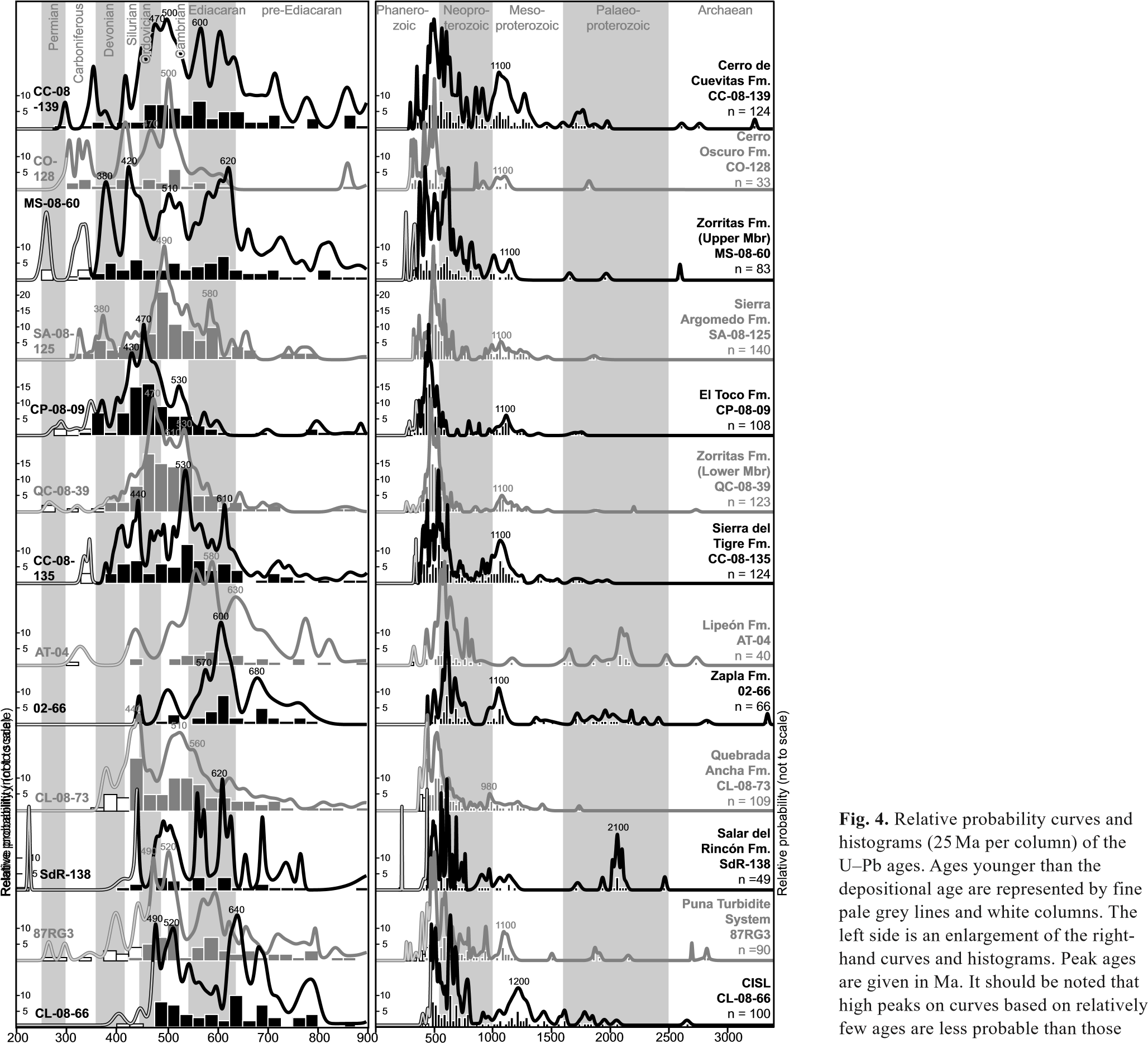
Main age peaks

No. of No. of Brasiliano– Pampean– Post- Youngest age Youngest zircons concordant Pre-Rodinian Rodinian Pampean Famatinian Famatinian accumulation age peak

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| --- | --- | --- | --- | --- | --- | --- | --- | --- | --- |
| Sample | dated | ages | (Ga) | (Ga) | (Ma) | (Ma) | (Ma) | (Ma)\* | (Ma)† |
| CL-08-66 | 138 | 100 |  | 1.2 (*n* = 5) | 640 (*n* = 8) | 520 (*n* = 7), 490 (*n* = 5) |  | 480 (*n* = 5) | 500 (*n* = 5) |
| 87RG3 | 145 | 90 |  | 1.1 (*n* = 4) |  | 520 (*n* = 5), 490 (*n* = 6) |  | 470 (*n* = 5) | 470 (*n* = 3) |
| SdR-138 | 90 | 49 | 2.1 (*n* = 5) |  | 620 (*n* = 3) |  |  | 490 (*n* = 4) | 610 (*n* = 4) |
| CL-08-73 | 148 | 109 |  | 0.98 (*n* = 5) | 560 (*n* = 5) | 510 (*n* = 7) | 440 (*n* = 6) | 430 (*n* = 9) | 440 (*n* = 7) |
| 02-66 | 99 | 66 |  | 1.1 (*n* = 4) | 680 (*n* = 5), 600 (*n* = 7), 570 (*n* = 4) |  |  | 500 (*n* = 3) | 500 (*n* = 4) |
| AT-04 | 66 | 40 |  |  | 630 (*n* = 4), 580 (*n* = 4) |  |  | 550 (*n* = 4) | 580 (*n* = 5) |
| CC-08-135 | 140 | 124 |  | 1.1 (*n* = 8) | 610 (*n* = 7) | 530 (*n* = 6) | 440 (*n* = 5) | 410 (*n* = 3) | 410 (*n* = 4) |
| QC-08-39 | 153 | 123 |  | 1.1 (*n* = 5) |  | 530 (*n* = 10),  510 (*n* = 7), 470 (*n* = 10) |  | 390 (*n* = 3) | 460 (*n* = 4) |
| CP-08-09 | 137 | 108 |  | 1.1 (*n* = 6) |  | 530 (*n* = 4), 470 (*n* = 10) | 430 (*n* = 8) | 350 (*n* = 6) | 350 (*n* = 4) |
| SA-08-125 | 153 | 140 |  | 1.1 (*n* = 6) | 580 (*n* = 7) | 490 (*n* = 12) | 380 (*n* = 4) | 320 (*n* = 3) | 480 (*n* = 6) |
| MS-08-60 | 169 | 83 |  | 1.1 (*n* = 4) | 620 (*n* = 5) | 510 (*n* = 4) | 380 (*n* = 4), 420 (*n* = 6) | 330 (*n* = 3) | 380 (*n* = 5) |
| CO-128 | 57 | 33 |  | 1.1 (*n* = 3) |  | 500 (*n* = 4), 470 (*n* = 3) | | 410 (*n* = 3) | 410 (*n* = 3) |
| CC-08-139 | 171 | 124 |  | 1.1 (*n* = 6) | 600 (*n* = 6) | 500 (*n* = 7), 470 (*n* = 5) | | 350 (*n* = 3) | 450 (*n* = 3) |

\*The youngest concentration of detrital ages with at least three ages.

†The youngest peaks were calculated with zircon ages with a maximum of 5% discordance, with at least three ages, and with at least 50% probability for the resulting ages. The 95% confidence level of the ages is up to a maximum of 12 Ma. Mean square weighted deviations always are <1.0. Also, ages younger than the estimated depositional age were considered for the calculations. Despite this, the youngest age peak never falls below the estimated depositional age.



Age (Ma) Age (Ma) based on many ages.

In addition to the Famatinian-arc and Pampean-age sources, material of ‘Grenvillian’ and Brasiliano ages also may have had local sources. To the north, crystalline rocks of ‘Grenvillian’ age exist both in the mainly Bolivian–Peruvian Sunsás belt (Fig. 1) and in the Arequipa–Pampia terrane (e.g. Loewy *et al*. 2004; Santos *et al*. 2008; Casquet *et al*. 2010). Diamictite of possible Neoproterozoic age in northernmost Chile (Sierra Limón Verde) also is dominated by zircon of ‘Grenvillian’ age (Morandé *et al*. 2012). To the south, Cambrian orthogneiss at *c*. 26°S in the Puna contains abundant inherited zircon of both ‘Grenvillian’ and *c*. 650 Ma age (Escayola *et al*. 2011). Furthermore, in the Western Sierras Pampeanas (Fig. 1), intrusive activity and metamorphism are recorded at 1.2–1.0 Ga (e.g. Vujovich *et al*. 2004; Rapela *et al*. 2010; Thomas *et al*. 2012). The crystalline ‘Grenvillian-age’ rocks may originally have been connected in a north–south-extending belt (e.g. Casquet *et al*. 2010; Ramos 2010). Escayola *et al*. (2007) also suggested a north–south-directed magmatic arc of Neoproterozoic age based on a back-arc-related ophiolite dated at 647 ± 77 Ma in the Eastern Sierras Pampeanas (whole-rock Sm–Nd).

Local ‘Grenvillian-age’ sources are supported by the Devonian El Toco Formation and Permo-Carboniferous units in northern Chile (not studied here), which contain detrital zircon grains of *c*. 1 Ga age with a mantle-influenced Hf-isotope signature (positive εHf values; Bahlburg *et al*. 2009). This is equivalent to similar Nd-isotope signatures (positive εNd values) of *c*. 1 Ga felsic and mafic rocks in the Western Sierras Pampeanas (e.g. Kay *et al*. 1996; Rapela *et al*. 2010) and gneiss at Quebrada Choja at 21°N in northern Chile (Loewy *et al*. 2004). Ordovician to Devonian sandstone from southern Peru and northwestern Bolivia contains abundant zircon of ‘Grenvillian’ age with a mantle εHf signature (Reimann *et al*. 2010), and therefore equivalent parent rocks probably have been present in the central Andean region. Contrary to this, the ‘Grenvillian-age’ material in the Sunsás belt and in the Peruvian part of the Arequipa–Pampia terrane mainly has crustal Nd-isotope signatures (negative εNd; Santos *et al*. 2008; Casquet *et al*. 2010), thus precluding these rocks as potential source rocks.

Early Carboniferous rhyolite (350-340 Ma) occurs at 27–28°S in the southern Puna (Martina *et al*. 2011). Hence, also post-Famatinian zircon grains may have originated from local source rocks.

It is plausible that the local sources were complemented by transport from the south throughout Palaeozoic time (Fig. 6). The Eastern Sierras Pampeanas (Fig. 1) contain abundant igneous and

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| **Fig. 5.** Relative probability curves for U–Pb ages of zircon grains from the sandstone samples of the mainly Silurian SdR-138, AT-04, and 02-66 compared with the ages from all other sandstone samples from this study at the top and for the Puncoviscana Formation and the Mesón Group at the bottom (Adams *et al*.  500  3000  2500  2000  1500  1000  Age (Ma)  Relative probability  (  not to scale  )  Relative probability (not to scale)  500  400  60  700  0  900  80  10001100  0  1200  Age (Ma)  Silurian  (  N = 3; n  = 153)  Others  (  N = 9; n  = 993)  Ediacaran to lower Cambrian  Puncoviscana Fm.  N = 11; n  (  = 724)  Middle to upper Cambrian  Mesón Group  N = 7; n  = 474)  (  2008, 2011; Augustsson *et al*. 2011). Only pre-depositional ages are included for the Ordovician to Permian strata. The left curves are enlargements of the righthand curves.    **Fig. 6.** Schematic evolutionary model for the detrital transportation paths along the continental Palaeo-Pacific margin of Gondwana from middle to late Cambrian to early Permian time. Transport additional to the arc source itself may have taken place from the south. The area from the Sierras Pampeanas in the south to the study area in the Puna in the north is shown. The basin types refer to the Puna area. A-P, Arequipa–Pampia terrane; Ch, Chilenia; Pc, Precordillera; RdP, Río de la Plata craton. |

metamorphic rocks of pre-Pampean to Pampean age (600–520 Ma; Siegesmund *et al*. 2010, and references therein; Iannizzotto *et al*. 2013, and references therein). Middle Devonian to Early Carboniferous (*c*. 390–330 Ma) igneous activity interpreted to be post-orogenic or extension-related is also recorded (Grosse *et al*. 2009, and references therein; Alasino *et al*. 2012). Furthermore,

shear-related metamorphic activity took place during Early Devonian time at *c*. 400 Ma (Höckenreiner *et al*. 2003). Post‘Grenvillian’ to pre-Brasiliano ages also may be explained by north-directed transport, because A-type orthogneiss, representing anorogenic igneous activity related to Rodinia break-up, in the Western Sierras Pampeanas has been dated at 774 ± 6 Ma (sensitive high-resolution ion microprobe U–Pb; Baldo *et al*. 2006). Therefore, the Sierras Pampeanas may have contributed zircon throughout the Palaeozoic era. Hence, either Brasiliano and postFamatinian rocks within a few hundred kilometres fed the depositional basins or (additional) southern longshore drift transported material in shallow-marine areas from the Eastern and Western Sierras Pampeanas before it was deposited in shallow or deeper marine environments (depending on the facies of the various studied sedimentary units). Longshore drift is in accordance with the scarcity of pre-Famatinian ages in the only analysed fluvial sandstone, the Carboniferous Cerro Oscuro Formation. Here, apparently no or little transport from Brasiliano-dominated areas took place. Dominant north-directed currents also are in accordance with the transportation paths for Ordovician retro-arc foreland basin deposits in the southern Puna (Zimmermann & Bahlburg 2003).

Longshore drift from far in the north is unlikely because of (1) the crustal Nd-isotope signatures for the ‘Grenvillian-age’ material in the Sunsás belt and the Peruvian part of the Arequipa–Pampia terrane (Santos *et al*. 2008; Casquet *et al*. 2010); also, (2) sources in the north would assume selective transport of detritus of specific ages only, because igneous and metamorphic rocks in northernmost Chile, Bolivia and Peru mainly are of 1.5–0.9 Ga and Palaeoproterozoic age (*c*. 2.0–1.7 Ga) in addition to Ordovician arc intrusive rocks (Damm *et al*. 1990; Wörner *et al*. 2000; Loewy *et al*. 2004; Santos *et al*. 2008; Casquet *et al*. 2010). We also rule out the area of the Brasília belt in central Brazil with the Goiás Massif and the Goiás–Crixás Archean Block with main ages of >2.5, 1.8–1.5, and 0.8–0.6 Ga (Fig. 1; e.g. Pimentel *et al*. 1999; Queiroz *et al*. 2000; Laux *et al*. 2005), because such zircon ages contributed little to the sedimentary rocks in this study; hence, the continent interior probably was shielded by the arc.

Previous studies have postulated main transport from the Famatinian arc for the Ordovician and Devonian units studied here because of the presence of volcanic detritus, palaeocurrent directions, and mafic geochemical influences (Breitkreuz 1985; Niemeyer 1989; Bahlburg 1990; Bahlburg & Breitkreuz 1993; Niemeyer *et al*. 1997*b*; Pérez *et al*. 2006; Zimmermann *et al*. 2010). Long-lived local transport is in accordance with the Ordovician–Silurian age peak that is particularly distinct in some of the Silurian and Devonian sedimentary units, because granitoid bodies related to the igneous Faja Eruptiva Oriental in northwestern Argentina probably are of *c*. 440 Ma age (Bahlburg *et al*. 2014). The granitoid furthermore contains inherited zircon that dominantly is of ‘Grenvillian’ and Brasiliano to Famatinian age (Bahlburg *et al*. 2014). This indicates the possibility of main detrital transport directly from the Famatinian arc, including recycled, igneous material.

Volcanic detritus in the Late Carboniferous, continental Cerro Oscuro Formation has earlier been assumed to be derived from the younger Carboniferous–Permian arc (Galli *et al*. 2010), but the zircon ages indicate that the extinct Famatinian arc was a major source. Hence, recycling of older material took place at the latest after emergence of the younger arc. Accordingly, the zircon ages in the lower Permian Cerro de Cuevitas Formation can be explained by recycling of the Sierra del Tigre Formation owing to the similarities in age distribution. The grain size and morphology, with the high amounts of rounded and round grains in the Permian unit, also are in accordance with recycling. Hence, the dominance of Famatinian-age grains indicates continuous transport from the same, nearby geographical region. Detritus from the Sierra del Tigre Formation cannot have been eroded and transported directly to the coastal depositional environment of the Cerro de Cuevitas Formation because of the angular unconformity separating the two units. Instead, detritus probably was deposited after a temporary storage phase as syntectonic, continental sediment. Reworking of this sediment (which has not been found in the region) led to renewed transport of the detritus. Finally, it was deposited in the ocean as the Cerro de Cuevitas Formation.

The turbiditic Devonian El Toco and Sierra del Tigre formations have dominant palaeocurrent directions that indicate axial transport from the north in addition to the Famatinian arc source directly to the east (Fig. 2; Bahlburg & Breitkreuz 1993; Niemeyer *et al*. 1997*a*), although no significant difference in zircon age spectra exists in comparison with the other Palaeozoic units. Such transport can be explained either by (1) main detrital supply from the arc in combination with local sources of older and younger age, or by (2) main transport from the arc and additional north-directed longshore drift, with reworking of detritus temporarily stored in the shoreface before turbiditic mass movements caused axial transport towards the south (Fig. 6). Because of the scarcity of local sources of plausible pre-Pampean and post-Famatinian age, we cannot distinguish between these two models.

### Detrital source in the east during Silurian time

The limited occurrence of zircon grains with ‘Grenvillian’, Pampean, and Famatinian ages from the Silurian units in Argentina, deposited in a central position and to the east of the former arc, points to less input from the Famatinian arc and related rocks. The abundance of Transamazonian ages indicates that the main transportation path may have emanated from the Río de la Plata craton in the east (Figs 1 and 6), where igneous and metamorphic rocks of 2.2–2.0 Ga age are common (e.g. Santos *et al*. 2003; Rapela *et al*. 2007). This is in agreement with the lower amount of euhedral and magmatic grains in the Zapla and Lipeón formations, because this may indicate a longer transport from areas with more zircon-bearing metamorphic rocks than for other units. The western border of the Río de la Plata craton is estimated to be situated directly east of the Eastern Sierras Pampeanas and possibly further east at the latitudes of the southern Puna (Fig. 1; Rapela *et al*. 2007). Whereas the main transport from the Famatinian arc probably involves firstcycle sedimentation owing to the short distance (at least for the older units), longer transport, such as from the Río de la Plata craton, may imply at least one recycling phase along the path.

Transamazonian-age zircon occurs not only in crystalline rocks of the Río de la Plata craton, but also in small amounts in the metasedimentary Ediacaran to Cambrian Puncoviscana Formation and equivalents and in the discordantly overlying siliciclastic Cambrian Mesón Group. The Puncoviscana Formation is dominated by detrital zircon with ages of 1.2–1.0 Ga and 670–610 Ma and commonly also 550–520 Ma, as well as some material of *c*. 2 Ga age (Fig. 5; Adams *et al*. 2008, 2011). The Mesón Group contains zircon mainly of 700–500 Ma, usually 2.2–2.0 Ga and occasionally also of 1.1–0.9 Ga age (Fig. 5; Adams *et al*. 2011; Augustsson *et al*. 2011). As a result of deformation and uplift during the Pampean orogeny, the Puncoviscana Formation was eroded. The resulting detritus was probably deposited in now buried or eroded sedimentary units, rather than in the Mesón Group, which is dominated by first-cycle material (Augustsson *et al*. 2011). The amount of Transamazonian zircon grains in the Puncoviscana Formation is much smaller than in the Silurian units in Argentina (Fig. 5). Therefore, it is unlikely to have shed large amounts of detritus into the Silurian basins. Hence, the Cambrian and Silurian deposits in Argentina represent a different transport system than the one to the west of the Famatinian arc. Ordovician strata in the area of the Eastern Sierras Pampeanas also contain Transamazonian-age zircon grains (La Cébila and Suri formations; Rapela *et al*. 2007; Verdecchia *et al*. 2011). Apart from these occurrences, Transamazonian-age zircon is scarce in Palaeozoic strata along the south–central Andes (Bahlburg *et al*. 2009), as well as in Devonian to Permian units at 26–32°S in Chile (Willner *et al*. 2008), and in upper Palaeozoic to Triassic strata at 31–52°S (Hervé *et al*. 2003, 2013; Augustsson *et al*. 2006).

### The Famatinian arc as an erosional barrier

The association of Famatinian arc detritus with ‘Grenvillian-age’ zircon grains and the scarcity of Transamazonian-age grains for units deposited west of or in a central position in the extinct arc indicate that the Famatinian arc acted both as a sediment source and as a transport barrier. Because of the large north–south extent of the Ordovician arc and its position along the margin of Gondwana it is unsurprising that it was a barrier for transport from the interior of the continent towards the western coastal area. During times of active tectonism, sediment from the interior of the continent may have been trapped on the continental side of foreland basins. The arc remained as a combined detritus feeder and detritus barrier from the time of its formation until at least the emergence of the younger magmatic arc during Carboniferous time. Continuous arc erosion is in accordance with estimated exhumation of rocks with Ordovician peak-metamorphic ages and Precambrian protoliths in northern Chile from the Ordovician period until late Palaeozic time (Lucassen *et al*. 2000). It also agrees with slow exhumation rates of <1 mm a−1 until Early Carboniferous time in the Sierras Pampeanas (de los Hoyos *et al*. 2011).

The only exception to the feeder–barrier function of the extinct Famatinian arc may have occurred during the Silurian period.

Among the Silurian units only the Quebrada Ancha Formation in Chile, with a dominance of Pampean and Famatinian zircon ages, was deposited in a western–central position in relation to the former Famatinian arc. It contains palaeotransport directions corresponding to transport from the arc (Navarro *et al*. 2006). Both the occurrence of Famatinian arc detritus to the west and the Palaeozoic exhumation estimates are in agreement with continued erosion of the fossil arc during Silurian time (Fig. 6; Lucassen *et al*. 2000; de los Hoyos *et al*. 2011). Therefore, we assume that the local transportation paths west of the arc continued throughout Palaeozoic time. In this case, the Quebrada Ancha Formation in Chile probably represents only a fragment of a much more extensive, unpreserved north–south-stretching (Late Ordovician to) Silurian sedimentary belt.

### Continent collisions

The Western Sierras Pampeanas have been assumed to be part of the allochthonous Cuyania–Precordillera (e.g. Kay *et al*. 1996; Thomas *et al*. 2012) or of an autochthonous Arequipa–Pampia terrane (e.g. Casquet *et al*. 2006; Rapela *et al*. 2010; Mulcahy *et al*. 2011). If the southern transport model, which involves zircon transport from the Western Sierras Pampeanas from late Cambrian and throughout Palaeozoic time, is correct, it implies that the Western Sierras Pampeanas were attached to Gondwana, supporting the existence of Arequipa–Pampia (Bahlburg *et al*. 2000).

The apparent change in transportation path of Silurian age appeared after the assumed docking of the Cuyania–Precordillera terrane west of the Sierras Pampeanas during Middle to Late Ordovician time (Figs 1 and 6; e.g. Ramos *et al*. 1986; Thomas & Astini 2003). Volcanic ash in the Precordillera proper dated at *c*. 470 Ma has been taken as evidence for a position in the vicinity of the Famatinian arc during Early to Middle Ordovician time (Fanning *et al*. 2004; Thompson *et al*. 2012). Despite this assumption, the arc did not have an imprint on the detrital zircon record in the Palaeozoic sedimentary units of the Precordillera or Cuyania. Instead, ages of 1.5–1.0 Ga are prevalent in Cambrian to Silurian units (Finney *et al*. 2004, 2005; Gleason *et al*. 2007; Naipauer *et al*. 2010; Abre *et al*. 2012). The scarcity of detrital zircon grains dated at *c*. 1.5–1.2 Ga in the units of this study indicates that transport to the basins further north did not take place from the Cuyania–Precordillera. However, we do not expect sediment transport to have crossed the collisional zone. Instead, transport from the collisional zone itself and from the Gondwana hinterland is plausible. Therefore, one could expect that the collision would cause transport of zircon of similar age populations to those in our results, but with different age proportions; however, this is not the case.

The Chilenia terrane is estimated to have collided with Gondwana along the western margin of Cuyania–Precordillera during Middle Devonian time (Fig. 1; e.g. Ramos *et al*. 1986; Willner *et al*. 2011). If collision took place, the main source area for the successions investigated here (the Famatinian arc) and the Sierras Pampeanas south of the study area should have been affected morphologically by continent collision twice during Palaeozoic time. We are not aware of any exposures from Chilenia proper, but Carboniferous to Permian accretionary wedge deposits in the area are marked by ‘Grenvillian’, Brasiliano, Pampean, and Famatinian detrital zircon ages (Willner *et al*. 2008; Bahlburg *et al*. 2009; Álvarez *et al*. 2011), similar to zircon ages from Gondwana. Therefore we cannot deduce any potential transport from Chilenia into the studied siliciclastic units. Hence, we neither record zircon input from the exotic continents themselves nor can we differentiate zircon transport from Gondwana versus the collisional zones. Generally, it is noteworthy that two continent collisions (Cuyania–Precordillera and Chilenia) did not necessarily affect zircon age populations in subsequent rocks.

### Glacial periods

The Silurian shift in apparent transportation paths coincides with the Late Ordovician to early Silurian ice age that caused glaciation in several parts of Gondwana (e.g. Díaz-Martínez & Grahn 2007; Ghienne *et al*. 2007; Van Staden *et al*. 2010). The glacial lower member of the Zapla Formation in northwestern Argentina contains polished and striated clasts (e.g. Grahn & Gutiérrez 2001; Astini *et al*. 2004). Astini *et al*. (2004) observed greywacke fragments and other sedimentary clasts, which they interpreted to derive from regional deposits. Also, in Bolivia, equivalent glaciogenic deposits contain clasts of local origin (Schönian & Egenhoff 2007). Schönian & Egenhoff (2007) postulated a regional, lowlatitude ice sheet with its centre in present-day northwestern Argentina, to the east of the extinct Famatinian arc. The ice sheet may also have transported detritus from the interior of the continent and thereby could have facilitated mixing with local material in the postglacial units of the Zapla and Lipeón formations. The glacial deposits of the Zapla Formation are chemically more mature than the younger Silurian strata (Abre *et al*. 2005). Hence, major reworking of the glacially transported sediment into the Zapla and Lipeón formations is unlikely.

Marked climatic changes occurred repeatedly during Palaeozoic time, as the South American part of Gondwana was affected by glaciation during both Late Ordovician to early Silurian and Late Devonian to Late Carboniferous times. Similarly, both the palaeomagnetic and palaeoclimatic apparent polar wander paths for South America indicate continent movement roughly from low to high to low latitudes during Palaeozoic time (Scotese & Barrett 1990; Tomezzoli 2009; Font *et al*. 2012). Despite these climatic differences, little age variation is recorded in the zircon age spectra. Hence, the climatic changes seem to have had little or no effect on the drainage configurations.

## Conclusions

In this study we document that there is very little evidence for transport of detrital material from the interior of Gondwana to the western margin of the super-continent. The scarcity of Transamazonian-age zircon grains in all but the Cambrian and Silurian units may give the impression that, after emergence of the Ordovician Famatinian arc, transport from the continent interior towards its western (south–central Andean) margin took place during Silurian time only. This biased view most probably is caused by the sampling of sedimentary rocks in different plate-tectonic and geographical positions, particularly in relation to the Ordovician magmatic arc. The shallow-water Silurian units with continent-interior material were the only ones that were situated in a palaeogeographical position that may have promoted input of large amounts of corresponding detritus. The arc blocked further sediment transport to the western margin. Therefore we postulate that transport from the interior of Gondwana continued throughout Palaeozoic time towards the west (Fig. 6), despite its scarcity in the Palaeozoic sedimentary units of the present-day central and southern Andes.

The tectonic situation throughout the Palaeozoic era was such that the transportation paths along the continental margin west of the Famatinian arc were affected only to a small extent. Hence, most of the studied Palaeozoic sedimentary record indicates transport from the arc, directly or after recycling of uplifted sedimentary units. This suggests that the arc and associated post-arc sedimentary units were eroded throughout Palaeozoic time since the formation of the arc. A present-day analogue is the back-arc basin of the Sea of Japan, which traps sediment from the Asian continent, whereas the fore-arc in the North Pacific Ocean is fed from the Northeastern Japan Arc (e.g. McLennan *et al*. 1990; Otosaka *et al*. 2004). Hence, it cannot be expected that material from the interior of the continent is found along the Palaeozoic active continental proto-Pacific margin.

The difference in age spectra for Silurian units indicates that care must be taken when interpreting zircon age populations from a sedimentary succession in a confined area. Direct comparison of sedimentary units in a temporal perspective may be invalid if the units occur in basins that were situated in different tectonic settings or in different positions in relation to main erosional barriers (but in similar tectonic settings). In such cases, the following should be noted.

1. Variations in age spectra cannot necessarily be taken as indicators of changes in transportation paths that are related to tectonic activity.
2. Stability in age distributions cannot necessarily be taken as evidence for a stable tectonic situation if detritus is trapped in other basins or if the studied basin is shielded by an erosional barrier from the tectonically active areas. Similarly, Krippner & Bahlburg (2013) studied Pleistocene Rhine sand and demonstrated that a complete orogenic cycle (the Alps) can be overlooked with zircon age data.
3. Renewed tectonic activity in the dominant source areas may cause recycling of first-cycle sedimentary rocks with zircon age distributions that are indistinguishable from those of younger, poly-cycle deposits. Hence, inherited geomorphological structures need to be taken into account when detrital grain ages are interpreted.
4. Despite reports of climate-related variation in zircon age distributions (e.g. Xiao *et al*. 2012; Gärtner *et al*. 2014), this study also demonstrates that ice-age periods do not necessarily cause variations in detrital age spectra.

Our data illustrate that zircon age spectra for sediment supplied from different and distant source areas may be similar and nonunique. The age spectra of the Silurian units in the south–central Andes resemble zircon populations of Early Palaeozoic sandstone and quartzite along the north Gondwana margin now exposed in Germany (see Gerdes & Zeh 2006; Bahlburg *et al*. 2010). The zircon grains from the German strata were interpreted to originate from the West African craton, Cadomia, and the Arabian–Nubian Shield (Gerdes & Zeh 2006; Bahlburg *et al*. 2010); that is, Gondwana areas that were remote from the South American western margin, but with rocks of ages that caused similar detrital zircon age distributions. This demonstrates that detrital zircon age populations do not necessarily reflect one specific source area. Similarly, Andersen (2014) stressed the similarity in both U–Pb age distributions and Hf-isotope composition of detrital zircon from Baltica and Laurentia.

Finally, we conclude that the transport system along the western margin of Gondwana remained stable at least throughout Palaeozoic time (i.e. for *c*. 0.3 byr) and that palaeogeographical reconstructions based on provenance data do not necessarily reflect the geological composition in the interior of the continent. Hence, the results of provenance studies may represent a much smaller part of the continent than anticipated, even for periods of tectonic quiescence.

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