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History of crustal growth and recycling at the Pacific convergent margin of South America at latitudes 29°–36° S revealed by a U–Pb and Lu–Hf isotope study of detrital zircon from late Paleozoic accretionary systems

A.P. Willner [a,c,*], A. Gerdes [b], H.-J. Massonne [a]

[a] Institut für Mineralogie und Kristallchemie, Universität Stuttgart, Azenbergstr. 18, D-70174 Stuttgart, Germany
[b] Institut für Geowissenschaften, Mineralogie, J.W.Goethe Universität, D-60438 Frankfurt, Germany
[c] Institut für Geologie, Mineralogie und Geophysik, Ruhr-Universität, D-44780 Bochum, Germany

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ABSTRACT

Detrital zircon provides a powerful archive of continental growth and recycling processes. We have tested this by a combined laser ablation ICP-MS U–Pb and Lu–Hf analysis of homogeneous growth domains in detrital zircon from late Paleozoic coastal accretionary systems in central Chile and the collisional Guarguaráz Complex in W Argentina. Because detritus from a large part of W Gondwana is present here, the data delineate the crustal evolution of southern South America at its Paleopacific margin, consistent with known data in the source regions.

Zircon in the Guarguaráz Complex mainly displays an U–Pb age cluster at 0.93–1.46 Ga, similar to zircon in sediments of the adjacent allochthonous Cuyania Terrane. By contrast, zircon from the coastal accretionary systems shows a mixed provenance: Age clusters at 363–722 Ma are typical for zircon grown during the Braziliano, Pampean, Famatinian and post-Famatinian orogenic episodes east of Cuyania. An age spectrum at 1.00–1.39 Ga is interpreted as a mixture of zircon from Cuyania and several sources further east. Minor age clusters between 1.46 and 3.20 Ga suggest recycling of material from cratons within W Gondwana.

The youngest age cluster (294–346 Ma) in the coastal accretionary prisms reflects a so far unknown local magmatic event, also represented by rhyolite and leucogranite pebbles. It sets time marks for the accretion history: Maximum depositional ages of most accreted metasediments are Middle to Upper Carboniferous. A change of the accretion mode occurred before 308 Ma, when also a concomitant retrowedge basin formed. Initial Hf-isotope compositions reveal at least three juvenile crust-forming periods in southern South America characterised by three major periods of juvenile magma production at 2.7–3.4 Ga, 1.9–2.3 Ga and 0.8–1.5 Ga. The $^{176}Hf/^{177}Hf$ of Mesoproterozoic zircon from the coastal accretionary systems is consistent with extensive crustal recycling and addition of some juvenile, mantle-derived magma, while that of zircon from the Guarguaráz Complex has a largely juvenile crustal signature. Zircon with Pampean, Famatinian and Braziliano ages (<660 Ma) originated from recycled crust of variable age, which is, however, mainly Mesoproterozoic. By contrast, the Carboniferous magmatic event shows less variable and more radiogenic $^{176}Hf/^{177}Hf$, pointing to a mean early Neoproterozoic crustal residence. This zircon is unlikely to have crystallized from melts of metasediments of the accretionary systems, but probably derived from a more juvenile crust in their backstop system.

1. Introduction

During the last decade detrital zircon has been intensively used as a powerful archive to study the variation of mechanisms for growth and recycling of continental crust through time. Because zircon survives multiple erosion-transport processes as well as even high grade metamorphic events, it can preserve magmatic and metamorphic crystallization ages recorded by the U–Pb system. This method can be combined with Lu–Hf analyses on the same grains to distinguish magmatic episodes that added juvenile mantle material to continental crust from those episodes that merely recycled existing crust (Kinny and Maas, 2003; Gerdes and Zeh, 2006; Augustsson et al., 2006; Scherer et al., 2007). This is due to the fact that during crystallization of zircon extreme fractionation of Lu–Hf ratios occurs between zircon and co-precipitating phases, such that zircon preserves the overall Hf-isotope signature of the magma, and hence its source.

The Pacific margin of South America provides an excellent example to test this method comparing the signature of detrital zircon mixed from continent-wide sources with known data from potential source

* Corresponding author. Institut für Geologie, Mineralogie und Geophysik, Abteilung Petrologie, Ruhr-Universität, D-44780 Bochum, Germany. Tel.: +49 234 322 4577; fax: +49 234 321 4433.
E-mail address: arne.willner@rub.de (A.P. Willner).

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regions. In particular, this margin has been active throughout Phanerozoic times showing diverse mechanisms of growth and recycling of continental crust through time. Hence, successive lateral growth of the South American convergent continent margin at latitudes 29°–36° S between 570 and 220 Ma involved (1) the deposition of vast masses of greywacke, their accretion, deformation and incorporation into wide magmatic arcs during the Early Paleozoic Pampean and Famatinian orogenies in Argentina after the assemblage of West-Gondwana, (2) the late Ordovician collision of the allochthonous microplate "Cuyania" exposed in the Argentine Precordillera, (3) the possible Devonian collision of the hypothetical terrane "Chilenia", and (4) the deposition of mainly greywacke on this microplate and to the W of it as well as its accretion to the continental margin and partial incorporation into magmatic arcs during Late Paleozoic times (Figs. 1 and 2; e.g. Ramos 2000). This variable multicyclic history of continental growth should be recorded by the detritus within the late Paleozoic accretionary system, which represents the last increment of lateral growth of the South American margin involving material eroded from W Gondwana.

During the lateral growth of South America zircon crystallized during successive episodes of calc-alkaline granitoid intrusions, recrystallized during accompanying high temperature (HT) metamorphism in the magmatic arcs, and was recycled by various episodes of deposition and accretion of siliciclastic sediments. Contrasting crustal recycling and crustal growth by juvenile magma addition during development of the late Paleozoic and Mesozoic magmatic arcs in central Chile was well documented by Lucassen et al. (2004). Thus, it is of interest to investigate, how these processes evolved at earlier times at the same convergent margin.

We combined cathodoluminescence (CL) imaging with LA-ICP-MS U–Pb and Lu–Hf isotope analyses of homogeneous growth domains in detrital zircon from metasediments of the late Carboniferous accretionary systems at the Pacific margin of southern South America to address the following questions:

- Can a pattern of crustal growth and recycling processes be shown for the Paleopacific margin of southern South America?
- How do these processes contrast with those which formed the allochthonous "Cuyania" terrane?
- Can we discriminate between major sources of provenance?
- Is it possible to provide evidence of the existence of the so far only hypothetical "Chilenia" terrane as the potential "backstop" of the coastal accretionary systems?
- What is the age and the nature of the first magmatic event after the emplacement of "Cuyania" in relation to accretionary processes farther west?
- What are the depositional ages of various tectonic units of the late Paleozoic coastal accretionary systems of central Chile in order to establish their relative growth pattern?

2. Geological setting

The lateral growth of the Pacific margin of South America during Paleozoic times started after and during amalgamation of Western Gondwana by the Neoproterozoic Brasiliano-Pan-African orogeny (Vaughan and Pankhurst, 2008; Fig. 1). This collision event resulted in major granitoid formation and HT metamorphism within mobile belts mainly at 520–650 Ma (Trouw et al., 2000; Basei et al., 2000). An

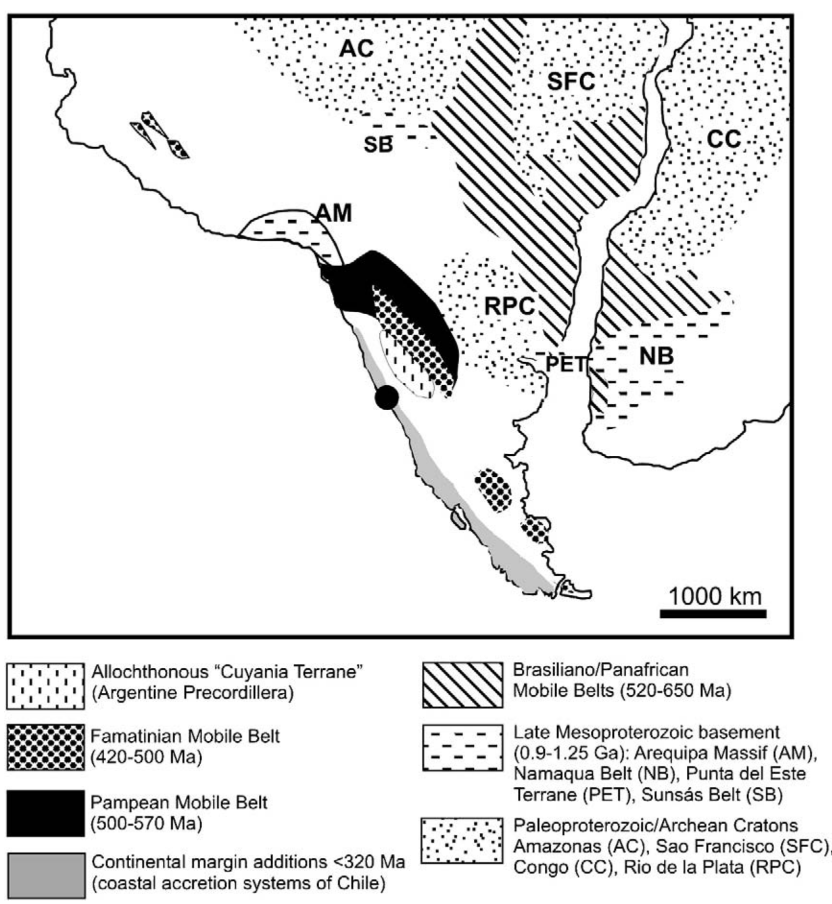


Fig. 1. Potential source areas for metagreywacke in the late Paleozoic accretionary systems of the Coastal Cordillera in Central Chile. All basement complexes shown include granitoids and high grade metamorphic rocks. They reflect the successive lateral continental growth along the western margin of Gondwana. Based on Geological Map of Gondwana 1:10,000,000 (De Wit et al., 1988). The circle marks the sampling areas of this study.

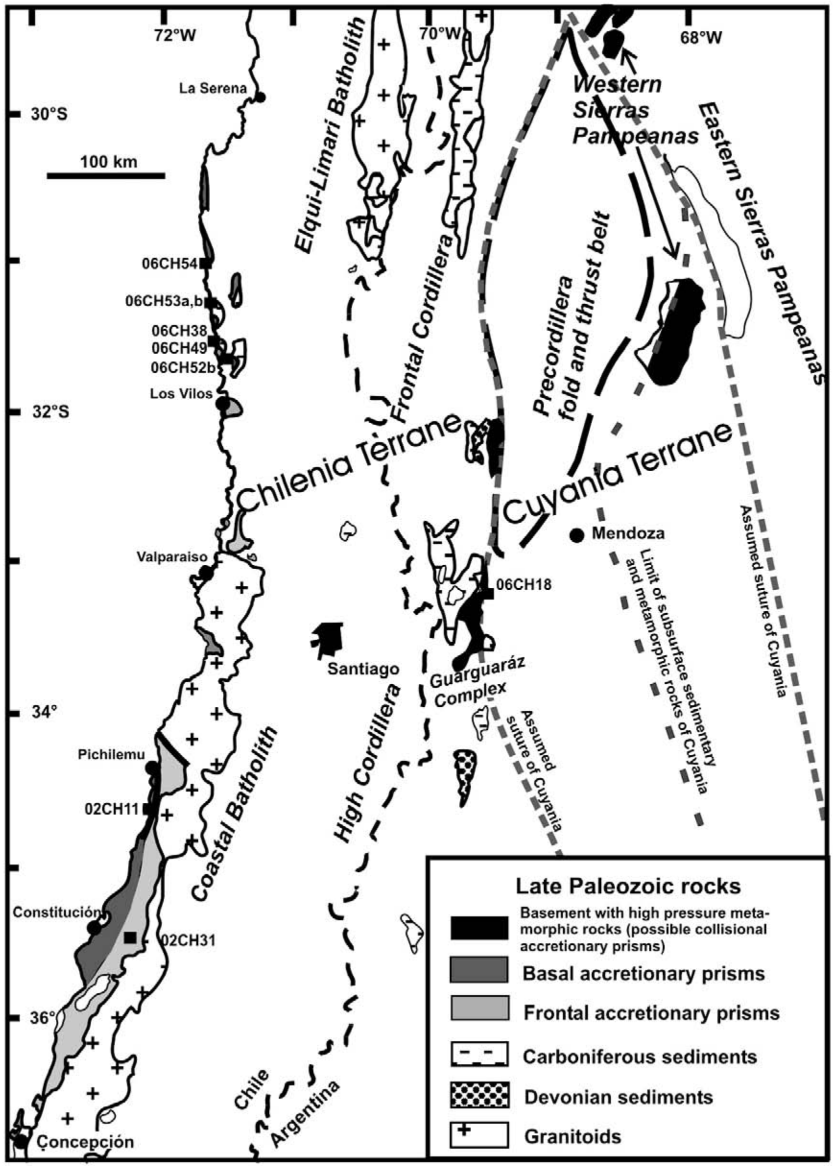


Fig. 2. Compilation of tectonic environments during late Paleozoic times at lat. 29–37° S including sample locations. Information is taken from Caminos (1979), Mpodozis and Kay (1992), Ramos (2000), Rebelledo and Charrier (1994) and Willner et al. (2005). The suture of the Cuyania Terrane is proposed by Ramos (2000) on the basis of serpentinite occurrences and rock occurrences under sedimentary cover known from drillholes.

important magmatic/metamorphic Paleoproterozoic episode leading to potential zircon formation in the amalgamated adjacent cratons is represented by the Transamazonian orogeny (1.9–2.4 Ga) that is particularly dominant in the Rio de la Plata Craton, the westernmost and southernmost craton of South America (Rapela et al., 2007). Mesoproterozoic magmatic/metamorphic overprints at 0.9–1.4 Ga (equivalent to the Grenvillian orogeny) are restricted to the Arequipa Massif of S Peru (Martignole and Martelat, 2003), the Sunsás Belt of W Brazil (Schwartz and Gromet, 2004; Chew et al., 2007) and the Punta del Este Terrane of Uruguay (Basei et al., 2000; Fig. 1). On the western flank of the accreted Western Gondwana a passive Neoproterozoic to early Cambrian margin developed, which was filled with predominantly turbiditic sediments (Puncoviscana Formation and equivalents exposed in the Argentine Eastern Sierras Pampeanas; Ježek and Miller, 1985; Willner et al., 1987). According to Rapela et al. (2007) the Puncoviscana sediments were probably not deposited west of the Rio de la Plata Craton, but further south and emplaced by dextral strike slip during their accretion to the continental margin. Two distinct orogenies along the western continental margin overprinted the previously passive margin during early Paleozoic times: at 520–570 Ma the siliciclastic sediments were accreted and partly incorporated into a magmatic arc (Pampean Mobile Belt; Rapela, 2000). At 490–420 Ma a wide magmatic arc including medium to high-grade metamorphic rocks formed in the western and central part of the Eastern Sierras Pampeanas (Famatinian Mobile Belt; Rapela, 2000). Recently, this prominent continental margin orogen was proved to extend southward towards the North Patagonian and Deseado Massifs (Pankhurst et al., 2006) as well as northward to the Eastern Cordillera of Peru (Chew et al., 2007). Local posttectonic granitoids intruded until ~350 Ma.

Whereas the Pampean mobile belt extends to the present margin in northern Chile (Lucassen et al., 2000), the allochthonous Cuyania Terrane exposed in the Argentine Precordillera was accreted to South America at 29°–36° S about 460–435 Ma ago (Ramos 2000; Casquet et al., 2001; Fig. 2). This terrane is characterized by exposed Lower Paleozoic shelf sediments deposited at a different continental margin. The derivation of the Cuyania Terrane is currently controversially discussed: Astini et al. (1995) and Kay et al. (1996) favour a Laurentian origin, whereas Finney et al. (2005) and Gleason et al. (2007) argue for derivation from Gondwana with a long-side displacement along the western Gondwana margin. Areas with high pressure (HP) metamorphism exceeding 10 kbar possibly representing collisional accretionary prisms were detected on both flanks of the Cuyania Terrane: Sierra Pie de Palo (Casquet et al., 2001) and Sierra de Umango (González et al., 2005) forming the Western Sierras Pampeanas in the east and the Guarguaráz Complex of the Argentine Frontal Cordillera (Lopez and Gregori, 2004; Massonne and Calderón, 2008) in the west. However, the Western Sierras Pampeanas are considered part of the Cuyania Terrane by Ramos (2000, 2004), whereas Rapela et al. (2007) regard them as a separate terrane and prolongation of the Mesoproterozoic Arequipa Massif emplaced during the Pampean orogeny. On the other hand, the Guarguaráz Complex consists of clastic metasediments, marble, amphibolite and ultramafic rocks. Lopez and Gregori (2004) cite K/Ar and Rb/Sr cooling ages at 445-317 Ma.

After the collision of Cuyania with Gondwana extensive Devonian and Carboniferous siliciclastic sediments (mainly turbidites) were deposited west of the Argentine Precordillera (Caminos, 1979; Fig. 2). These sediments also represent the bulk mass of material constituting the accretionary systems of the Chilean Coastal Cordillera (Hervé et al., 2007). Large parts of the Late Paleozoic sediments in central Chile are believed to be underlain by a further allochthonous terrane named "Chilenia" (Ramos, 2000, 2004) which is not exposed. The existence and nature of this terrane is currently also controversially discussed.

Cyclic mass flow by continuous accretion in central Chile occurred for about 100 Ma between 320 and 220 Ma (Willner et al., 2005). During this time interval the accretion mode changed from frontal accretion (Eastern Series) to basal accretion (Western Series; Richter et al., 2007; Figs. 2 and 3), i.e. from predominant horizontal shortening at shallow depths to subhorizontal flattening in deeper levels of the accretionary prisms. A late Paleozoic magmatic arc developed concomitantly with the basal accretion of the HP/low temperature (LT) rocks of the Western Series and was accompanied by HT/low pressure (LP) metamorphism around 300 Ma that overprinted the Eastern Series. Hence, a true paired metamorphic belt constitutes the last and westernmost addition to the Pacific margin of South America. There are two major branches of the late Paleozoic magmatic arc, the Elqui–Limarí Batholith (28°30′–31° S) and the Coastal Batholith (33°–39° S). Ages of calc-alkaline intrusions in the Elqui–Limari Batholith range from 310±8 to 256±10 Ma (Rb–Sr wr isochrons; Mpodozis and Kay, 1992; Pankhurst et al., 1996). In the Coastal Batholith intrusions occurred from 308±15 to 257±1 Ma (Rb–Sr wr isochrons, U/Pb zircons, Hervé et al., 1988; Willner et al., 2005) with the main pulse at the earliest stage. Carboniferous intrusion ages are also known from the Argentine Frontal Cordillera (Caminos, 1979; Llambias and Sato, 1990). There is a notable lateral offset of strike between both batholiths at lat. 32°30′ S (Fig. 2) presumably due to a strike–slip fault: in contrast to the Coastal Batholith, which intrudes the oldest part of the accretionary wedge, a wide fore-arc basin exists to the west of the Elqui–Limari Batholith. Whereas units are largely coherent in the segment south of lat. 32°30′ S, they are strongly dissected by postaccretionary faulting in the north (Rebelledo and Charrier, 1994). When accretion ended during Triassic times, a major change of the subduction geometry occurred to an extensional regime with basin formation, unconformable deposition of sediments on the accretionary system and extensive rhyolitic volcanism (Ramos, 2000; Willner et al., 2005).

3. Analytical methods

3.1. U–Pb dating

Between 50 and 100 detrital zircon grains from each of six metagreywacke samples as well as four to 14 zircon grains from three leucogranite/rhyolite pebbles each were dated. A total of 506 U–Pb ages were obtained by LA-ICP-MS techniques at Johann Wolfgang Goethe University Frankfurt (JWG) using a Thermo-Scientific Element II sector field ICP-MS coupled to a New Wave UP213 UV laser system equipped with a teardrop-shaped, lower-volume laser cell (Gerdes and Zeh, 2006; Janoušek et al., 2006; Zeh et al., 2007; Slama et al., 2008; Gerdes and Zeh, 2008). Laser spots (30 µm for U–Th–Pb) selection was guided by internal structures as seen in CL images of the mounted and polished grains. Data were corrected for laser-induced elemental fractionation and mass bias by normalization to the reference zircon GJ-1 ($^{206}Pb/^{238}U=0.0986\pm0.0004$; $^{207}Pb/^{206}Pb=0.06016$; JWG ID-TIMS). Prior to this, the drift in inter-elemental fractionation (Pb/U) during sample ablation was corrected by applying a linear regression through all measured ratios. Correction for

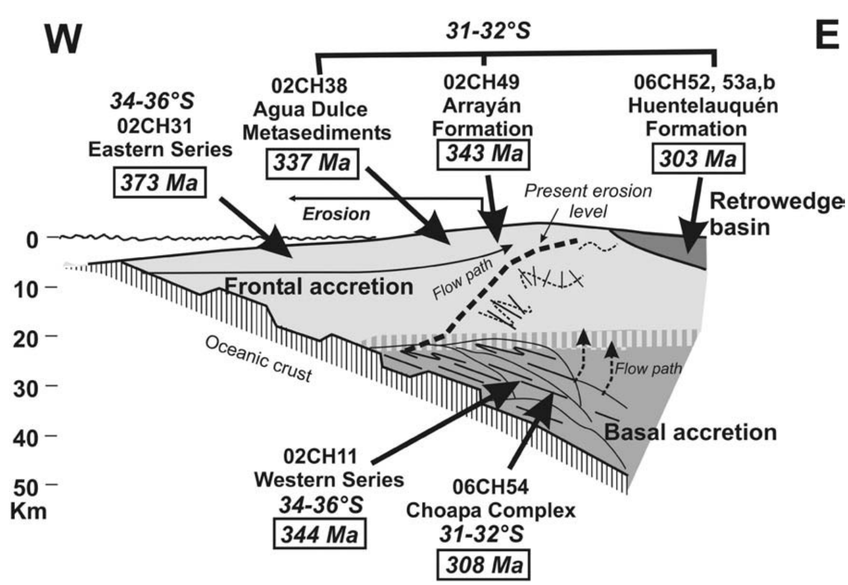


Fig. 3. Principle structure of the Late Paleozoic accretionary systems in central Chile at ~300 Ma. During their formation a continuous change of the accretion mode from frontal to basal accretion occurred (Richter et al., 2007). Frontal accretion is characterized by horizontal particle flow paths and shortening, folding of bedding (stippled lines) and by subvertical foliation (solid lines). Basal accretion is characterised by vertical shortening and particle flow paths as well as subhorizontal foliation, intrafolial folds and duplex structures. Rocks of a retrowedge/forearc basin are only exposed at lat. 31–32° S (Rebelledo and Charrier, 1994). Sampled local formations are assigned to these tectonic environments. The maximum depositional ages derived from this study are inserted in boxes.

common-Pb was done whenever the corrected $^{207}Pb/^{206}Pb$ was outside the internal errors of the uncorrected ratio and was based on the interference- and background-corrected ^{204}Pb signal and the terrestrial Pb evolution model of Stacey and Kramers (1975). Reported uncertainties (2σ) were propagated by quadratic addition of the external reproducibility (2SD; standard deviation) obtained from the standard zircon GJ-1 (~1.2% and ~1.4% for the $^{207}Pb/^{206}Pb$ and $^{206}Pb/^{238}U$, respectively; $n=12$) during individual analytical session and the within-run precision of each analyses (2SE; standard error). The accuracy of the method was verified by each nine analyses of reference zircon 91500 (1064.8±4.3 Ma, MSWD of concordance and equivalence=0.86), Plešovice (338.6±1.9 Ma, $MSWD_{C+E}=0.78$), and Temora (415.9±2.3 Ma, $MSWD_{C+E}=1.1$). The results are within uncertainty of commonly accepted values (Wiedenbeck et al., 1995; Black et al., 2003; Slama et al., 2008). Further 27 U–Pb ages were added to the data set that had previously been measured with the SHRIMP II at Curtin University in Perth, Australia, including seven that were re-measured for comparison of the methods. Results are represented in Figs. 5–7 and in Tables E1–E7 in the online electronic archive.

Table A1 in the Appendix shows the results of the seven zircons analysed by both, LA-ICPMS and SHRIMP. In 6 zircons the $^{206}Pb/^{238}U$ and the $^{207}Pb/^{206}Pb$ ages agree between both methods within 1σ uncertainty. For one zircon SHRIMP analysis yields a ca. 10% higher $^{206}Pb/^{238}U$ (9% reverse concordant) and for another a ca. 50% higher $^{207}Pb/^{206}Pb$ (40% discordant). In both cases the LA-ICP-MS $^{206}Pb/^{238}U$ and $^{207}Pb/^{206}Pb$ ages appear to be more consistent (98–93% concordant). Thus in general the comparison between both methods is very good which was also shown earlier based on a larger data set by Gerdes and Zeh (2006).

For interpretation of the detrital zircon ages only concordant or nearly concordant (<10% discordant) data were considered. For the probability and frequency plots the $^{207}Pb/^{206}Pb$ ages are used for all zircon grains >1 Ga and the $^{238}U/^{206}Pb$ ages for all zircon <1 Ga.

3.2. Hf isotope analyses

We selected 138 dated zircon grains for Lu–Hf isotope analyses to cover all previously identified age clusters. Analyses (40 µm spots) were performed with a Thermo-Scientific Neptune multi-collector ICP-MS at JWG using the same laser and ablation cell as described above (Gerdes and Zeh, 2006, 2008). To correct for isobaric interferences of Lu and Yb on mass 176 the isotopes ^{172}Yb, ^{173}Yb and ^{175}Lu were simultaneously monitored. The ^{176}Yb and ^{176}Lu were calculated using a $^{176}Yb/^{173}Yb$ of 0.796218 (Chu et al., 2002) and $^{176}Lu/^{175}Lu$ of 0.02658 (JWG in-house value). The instrumental mass bias for Hf isotopes was corrected using an exponential law and a $^{179}Hf/^{177}Hf$ value of 0.7325. In case of Yb isotopes the mass bias was corrected using the Hf mass bias of the individual integration step multiplied by a daily βHf/βYb offset factor (Slama et al., 2008; Gerdes and Zeh, 2008). All zircon LA MC ICP-MS analyses were adjusted relative to the JMC 475 $^{176}Hf/^{177}Hf$ ratio of 0.282160 and the reported uncertainties (2σ) were propagated by quadratic addition of the external reproducibility of GJ-1 (2 SD, $n=17$) and the within-run precision of each analyses (2 SE). The external reproducibility (2SD; $n>50$) over more than 6 month of reference zircon 91500, GJ-1, and Plešovice ($^{176}Hf/^{177}Hf$=0.282298±0.000026, 0.282003±0.000018, and 0.282482±0.000015, respectively) at JWG is about 0.005–0.009% (<1 epsilon unit). Results are represented in Fig. 8 and in Table E8 of the online electronic archive.

3.3. Cathodoluminescence imaging

We used a CL detector of a CAMECA SX50 electron microprobe (EMP) at Ruhr–Universität Bochum and a CAMECA SX100 EMP at Universität Stuttgart to select homogeneous growth zones in individual zircon grains for analysis, to detect inherited cores and criteria for a magmatic or metamorphic origin. Although metamorphic zircon, which usually represents late or postmagmatic recrystallization phenomena in HT terrains often cannot unambiguously be distinguished from magmatic zircon, we used the following criteria to discriminate metamorphic zircon: heterogeneous patchy zoning with sharp and curved sectoral boundaries (Fig. 4A, C, D, E), zones of dissolution and re-growth propagating through the zircon crystal (Fig. 4A, B, C), discordant planar and non-planar sectoral overgrowth phenomena (Fig. 4B, C, D, E), convolute zoning and blurred former oscillatory zoning (Fig. 4A, B, C; Vavra et al., 1999; Corfu et al., 2003; Kinny and Maas, 2003; Harley et al., 2007). Yet most analyzed zircon grains show perfect concentric, narrowly spaced oscillatory growth zoning characteristic of magmatic origin (Fig. 4F, J, K). Also sectoral zoning (Fig. 4F, K), resorption phenomena (resorption embayment Fig. 4F; rounded cores Fig. 4F, J) as well as xenocryst cores (Fig. 4G, H) are frequent. Often the ratio Th/U<0.1 is used for discrimination between magmatic and metamorphic zircon (Rubatto, 2002). However, in our study only 3 grains assigned to a metamorphic origin show such a low ratio, whereas 13 grains with clear oscillatory magmatic zonation would also fulfil this criterion. In this paper the assignment to a metamorphic origin is descriptive only and is taken as a hint at the presence of high grade metamorphic rocks in the source regions. The results do not conflict with any of the data interpretations presented below. Nevertheless, the classification must be handled with great caution, because it cannot be assessed whether partly domains with relic magmatic ages were measured or initial $^{176}Hf/^{177}Hf$ ratios are preserved. Both alternatives appear likely present.

Finally seven whole rock analyses of leucogranite and rhyolite pebbles were achieved at Ruhr–Universität Bochum by X-ray fluorescence spectroscopy. The Fe^{2+}/Fe^{3+} ratio and the water content were determined by standard wet chemical methods. Data are given in Table E8 of the electronic data base.

4. Geological and petrographical characteristics of samples

Metasediments for zircon separation were taken from different units within the accretionary system of north-central Chile considering the contrasting setting in the domains north and south of 32°30′S (Figs. 2 and 3; Table A2).

(1) Sample 02CH11 is a metagreywacke from the Western Series. This unit was overprinted by HP-LT metamorphism (Willner, 2005) and interpreted as a basal accretionary complex (Richter et al., 2007). Metapsammopelites in this unit are monotonous and consist of entirely recrystallized quartz, albite, white mica and chlorite. A metamorphic banding formed as transposition foliation subparallel to the former bedding. The heavy mineral spectrum, consisting of zircon, tourmaline, apatite, ilmenite and magnetite with occasional rutile and epidote, is considerably reduced due to multiple recycling. Contents of the stable heavy minerals zircon and tourmaline vary considerably among samples. Sample 06CH54 is a metagreywacke with identical characteristics from the Choapa Metamorphic Complex (Rebelledo and Charrier, 1994). This complex can be considered as an equivalent of the Western Series north of 32°30′S due to comparable tectonic structures such as subhorizontal transposition foliation produced by basal accretion and abundant metabasite intercalations.

(2) Three metagreywacke samples were taken from structurally higher levels of the accretionary systems: 02CH31 from the Eastern Series, 02CH38 from the Agua Dulce Metaturbidites, 02CH49 from the Arrayán Formation. The latter two sedimentary sequences (see Rebelledo and Charrier, 1994) appear as equivalents to the Eastern Series due to similar tectonic structures as large scale chevron folding of bedding planes and lack of a transposition foliation.

Sample 02CH31 has similar characteristics as 02CH11 except for biotite grown at the expense of chlorite (+ K-feldspar) due to the HT overprint of the Eastern Series (Willner, 2005). Samples 02CH38 and 02CH49 are fine-grained greywackes with the original clastic texture preserved with a very low grade metamorphic overprint. However, sample 02CH38 shows a somewhat higher degree of recrystallization of the matrix compared to 02CH49. Mainly angular clasts of quartz,

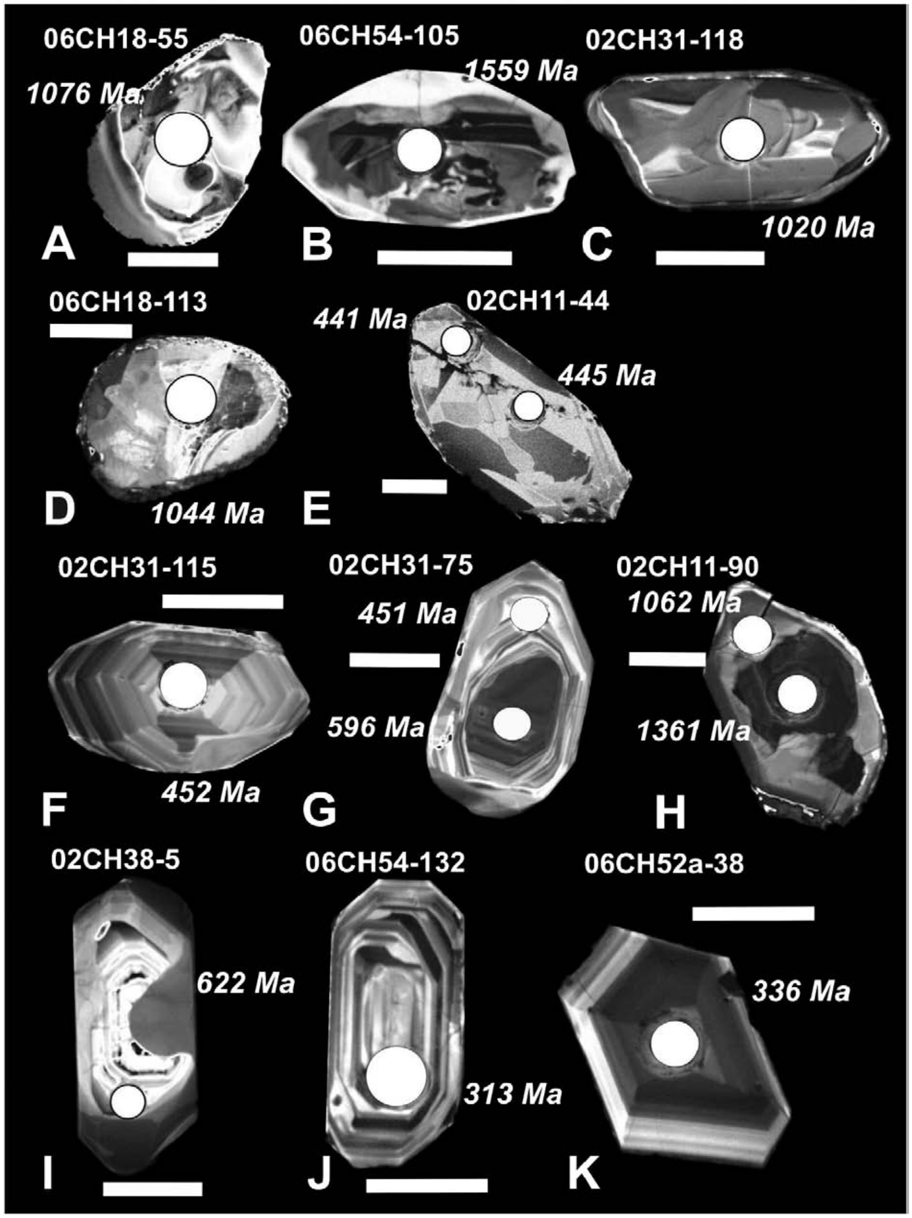


Fig. 4. Representative CL images of (1) metamorphic zircon A-E and (2) magmatic zircon F-K. For explanation see text. All scale bars are 50 μm.

albite, white mica and chlorite are observed in all three samples. Tourmaline, zircon, apatite, epidote, ilmenite and magnetite represent the heavy mineral spectrum. In coarser grained samples from the Arrayán Formation rhyolite, feldpar–quartz aggregates, heavily altered basalt, and shale are observed as lithoclasts. Whereas the depositional age for the rocks of the Eastern Series and the Agua Dulce Metaturbidites is as yet unknown, a Devonian to Lower Carboniferous biostratigraphical age is assigned to the Arrayán Formation (Rebelledo and Charrier, 1994).

(3) The Arrayán Formation is unconformably overlain by the Huentelauquén Formation with an Upper Carboniferous to Permian biostratigraphical age (Rebelledo and Charrier, 1994). Sediments of the Huentelauquén Formation including sandstone, slate, limestone and conglomerate are considered to be neritic. Conglomerate intercalations show conspicuous contents of rhyolite and leucogranite pebbles (60–70%) apart from subordinate pebbles of vein quartz, sandstone, slate and some phyllite. Locally, either rhyolite or leucogranite dominates. Pebbles with sizes up to 15 cm are subangular to rounded suggesting they originated from a nearby homogeneous source.

Zircon was separated from two leucogranite pebbles (06CH52a, 06CH53a) and a rhyolite clast (06CH53b) of the Huentelauquén Formation. These clasts are remarkably similar and genetically linked. Medium to coarse grained leucogranites are characterized by predominant quartz-K-feldspar symplectites filling spaces between idiomorphic to hypidiomorphic plagioclase and K-feldspar crystals, feldspar–quartz aggregates, and idiomophic quartz crystals with resorption embayments. Biotite and titanite are rare accessories. Secondary epidote and serizite formed within plagioclase crystals. Epidote–quartz filled fissures are abundant. The leucogranite may be interpreted as a subvolcanic equivalent of the rhyolite. Rhyolite pebbles are characterized by idiomorphic phenocrysts of quartz with resorption embayments and feldspar phenocrysts in a groundmass with pronounced fluidal fabric. The groundmass is strongly serizitized, whereas both epidote and serizite replace plagioclase. Also quartz–epidote filled fissures are abundant.

(4) Furthermore, zircon was separated from mica-schist sample 06CH18 from a presumable collisional accretionary prism (Guarguaráz Complex; Lopez and Gregori, 2004) at the western suture of the

"Cuyania Terrane" for comparison and to gain further hints at its nature (Figs. 2 and 3). This coarse-grained rock contains the mineral assemblage biotite – white mica – plagioclase – quartz. Heavy minerals are magnetite, zircon and tourmaline. According to Massonne and Calderón (2008) the medium grade rocks of the Guarguaráz Complex were overprinted at 14 kbar/450 °C. The depositional age of the protolith is as yet unknown.

5. Results

5.1. Precambrian zircon from the Guarguaráz complex

Detrital zircon of the HP collisional Guarguaráz Complex (06CH18; 90 analyses) defines a broad age cluster at 0.94–1.45 Ga (95% of the population) with a predominant maximum at 1.09 Ga and minor maxima at 0.94, 1.23, 1.36 and 1.42 Ga (Figs. 5 and 6). All ages are $<10\%$ discordant. Four zircons yielded ages >1.42 Ga (1.8, 1.9, 2.5 and 2.8 Ga). Phanerozoic ages are missing. The youngest concordant zircon was dated at 555 ± 8 and 581 ± 8 Ma and represents a poorly defined magmatic event at the end of Neoproterozoic times after a considerable age gap. These ages define a maximum deposition age of the Guarguaráz metasediments. Considering a presumable Devonian age of metamorphism (Lopez and Gregori, 2004) the sedimentation age of the Guarguaráz metasediments most likely is Lower Paleozoic.

Metamorphic zircon is remarkably rare in the Mesoproterozoic zircon population (3% of the population; Fig. 4A, D). Also inherited xenocryst cores were not detected suggesting a low degree of crustal recycling. Instead, rounded resorbed cores or cores with resorption embayments of the same magmatic event are frequent. All magmatic zircon grains typically show oscillatory zoning with strongly contrasting CL brightness. The degree of rounding of the Mesoproterozoic zircon grains is remarkably low (19% idiomorphic – 52% edge rounded – 27% subrounded – 2% rounded). 74% of the zircon (0.1–0.6 mm lengths) is represented by types P2, P3, P4 and P5 of Pupin (1980), whereas 26% fall into classes S15, S20 and S25.

5.2. Precambrian/early Paleozoic zircon from the coastal accretionary systems

Detrital zircon in samples from the late Paleozoic coastal accretionary systems shows similar age pattern (Figs. 5 and 6). This is consistent with an overall comparable range of morphological characteristics (0.05–0.3 mm lengths, mean 0.1–0.15 mm; 3–18% idiomorphic — 31–51% edge rounded — 31–43% subrounded — 4–15% rounded; morphological types after Pupin (1980): P5 — 46%, P4 — 28%, P3 — 19%; S15 — 5%, S20 — 1%, S25 — 1%). The age maxima are represented by zircon of magmatic as well as metamorphic origin. The latter comprises 12% of the population.

About 24% of the detrital zircon ages fall in the range of 360–550 Ma with pronounced maxima at 520–547 Ma, 423–488 Ma and 363–404 Ma. However, dominance of the maxima strongly varies regionally. Age clusters at 556–722 and 779–982 comprise another 21% of the analyses. Zircon with inherited xenocrystic cores was detected in the first four populations (Table 1; Fig. 4G) proving crustal recycling during these events.

A second major age cluster (26%) defined by detrital zircon of mainly magmatic origin occurs at 0.99–1.39 Ga (dominant maxima at 0.99–1.11, minor maxima at 1.18–1.39 Ga; Figs. 5 and 6). This cluster is similar to the Mesoproterozoic Precordilleran age spectrum in sample 06CH18. However, four grains with inherited xenocrystic cores were detected in this population (Table 1) contrasting to sample 06CH18. These zircon grains partly indicate crustal recycling during Mesoproterozoic times.

Minor age clusters (12%) of older detrital zircons with random distribution within the late Paleozoic accretionary systems are represented by age maxima at 1.46–1.79 Ga, 1.87–2.22 Ga (Transamazonian event), 2.43–2.79 Ga and 2.95–3.20 Ga (Figs. 5 and 6). The oldest recorded zircon age is 3.37 ± 0.14 Ga. Two grains with xenocrystic cores (Table 1) indicate that similar crustal recycling processes occurred during Paleoproterozoic times.

5.3. Carboniferous zircon from the coastal accretionary systems

All studied samples except 06CH18 contain a population of idiomorphic zircon with ages in the range of 294–346 Ma (Figs. 4, 5, and 7). This group invariably includes detrital zircon with the same morphological and CL characteristics as zircon separated from the leucogranite and rhyolite pebbles of the Huentelauquén Formation. These grains (0.1–0.4 mm length) show restricted morphological variation comprising mainly the classes P4 and P5 according to Pupin (1980) with minor contribution of classes S20 and S25. Perfect concentric oscillatory zoning with strongly contrasting CL brightness is conspicuous (Fig. 4J, K). Quartz and rutile inclusions are common and xenocryst cores are lacking with one exception among detrital zircon (Table 1). Cores with rounded resorption surfaces (Fig. 4J) or resorption embayments as well as sector zoning (Fig. 4K) only occur occasionally. Three magmatic pebbles from the Huentelauquén Formation were dated:

Leucogranite pebble 06CH52a exhibits a $^{206}Pb/^{238}U$ age range of subconcordant zircon between 315 and 337 Ma. Ages partly do not overlap within the 2σ-errors (Figs. 5 and 7). This suggests disturbance of the U–Pb system due to lead loss in some of the analysed domains. Because all pebbles show strong evidence of a postmagmatic hydrothermal overprint, such an effect is not unlikely. The $^{207}Pb/^{206}Pb$ mean age of all analyses is 326 ± 12 Ma. This age is similar to a weighted age average of 332 ± 4 Ma, where the three youngest ages are excluded. The latter age is interpreted as crystallization age of the leucogranite.

Zircon of leucogranite pebble 06CH53a yields a $^{206}Pb/^{238}U$ age range of 346–289 Ma (Figs. 5 and 7). Zircon of leucogranite pebble 06CH53a yields a $^{206}Pb/^{238}U$ age range of 346–289 Ma (Figs. 5 and 7). The weighted average gives an unreasonable high MSWD (mean square of weighted deviates) of 15 even when the oldest analyses are excluded (Fig. 7). Only the three younger grains are equivalent with a concordia age of 294 ± 4 Ma. The remaining three analyses possibly contain some inheritance. This view is supported by the Hf isotope composition ($\varepsilon Hf_{(t)}$ of 1.4 and −3.1, respectively), which is distinct to that of the three younger grains (0.1 ± 0.5; see Section 5.1; Table E6 in the electronic data base). Thus, the leucogranite most likely crystallized during late Carboniferous times.

Four dated zircon grains of rhyolite pebble 06CH53b with a $^{206}Pb/^{238}U$ age range of 289–305 Ma yield a weighted average of 298 ± 11 Ma and a concordia age of 299 ± 46 (Figs. 5 and 7). The somewhat elevated MSWD is related to one analysis, which probably was affected by some Pb-loss. The concordia age overlaps with that of leucogranite 06CH53b and defines a maximum deposition age for the conglomerate of 303 Ma consistent with the Upper Carboniferous–Lower Permian biostratigraphical age of the Huentelauquén Formation (Rebelledo and Charrier, 1994).

The age range of the pebbles is also represented by the youngest detrital zircon population in sample 06CH54 (299–346 Ma; weighted means at 333 ± 3 Ma and 304 ± 7 Ma; Figs. 5 and 7). This sample represents the basally accreted Choapa Metamorphic Complex and contains a relatively large number of this Carboniferous zircon (29% of the population). The younger maximum at 304 ± 7 Ma overlaps with the concordia ages obtained from zircon of the leucogranite and rhyolite pebbles and is similar to the age of the youngest zircon at 298 ± 10 Ma. Hence, a similar maximum deposition age of 308 Ma can be inferred for the protolith of this sample, which coincides with that of the Huentelauquén Formation.

In contrast, Carboniferous zircon is less frequent (9% of the population) in sample 02CH49 from the Arrayán Formation, has a relative uniform $^{206}Pb/^{238}U$ and is distinctly older (331–346 Ma; weighted mean at 340 ± 5 Ma; Figs. 5 and 7). The age of the youngest

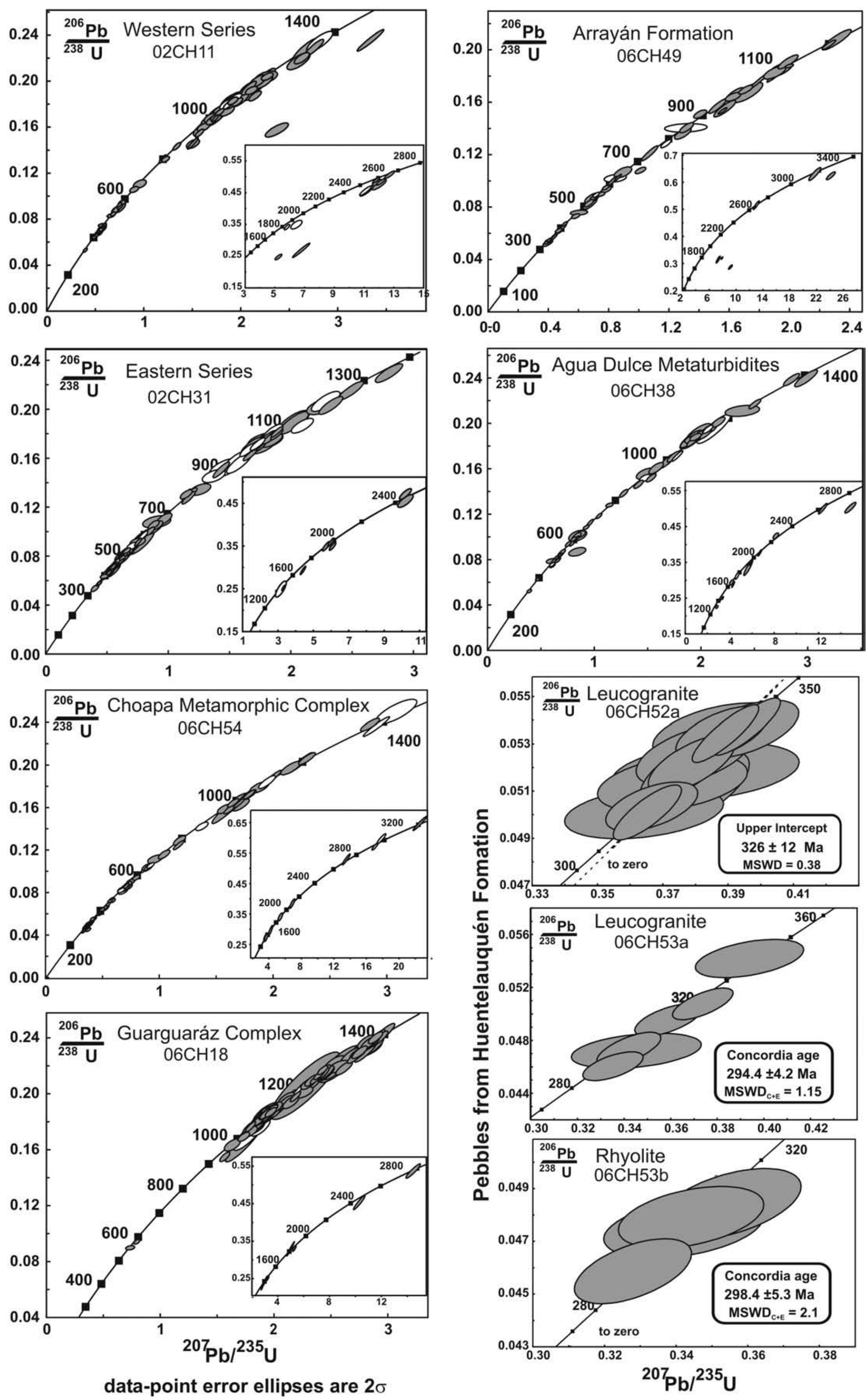


Fig. 5. Concordia plots. Dark grey: magmatic zircon; white: metamorphic zircon. Background data are available in Tables E1–E7 in the electronic data base.

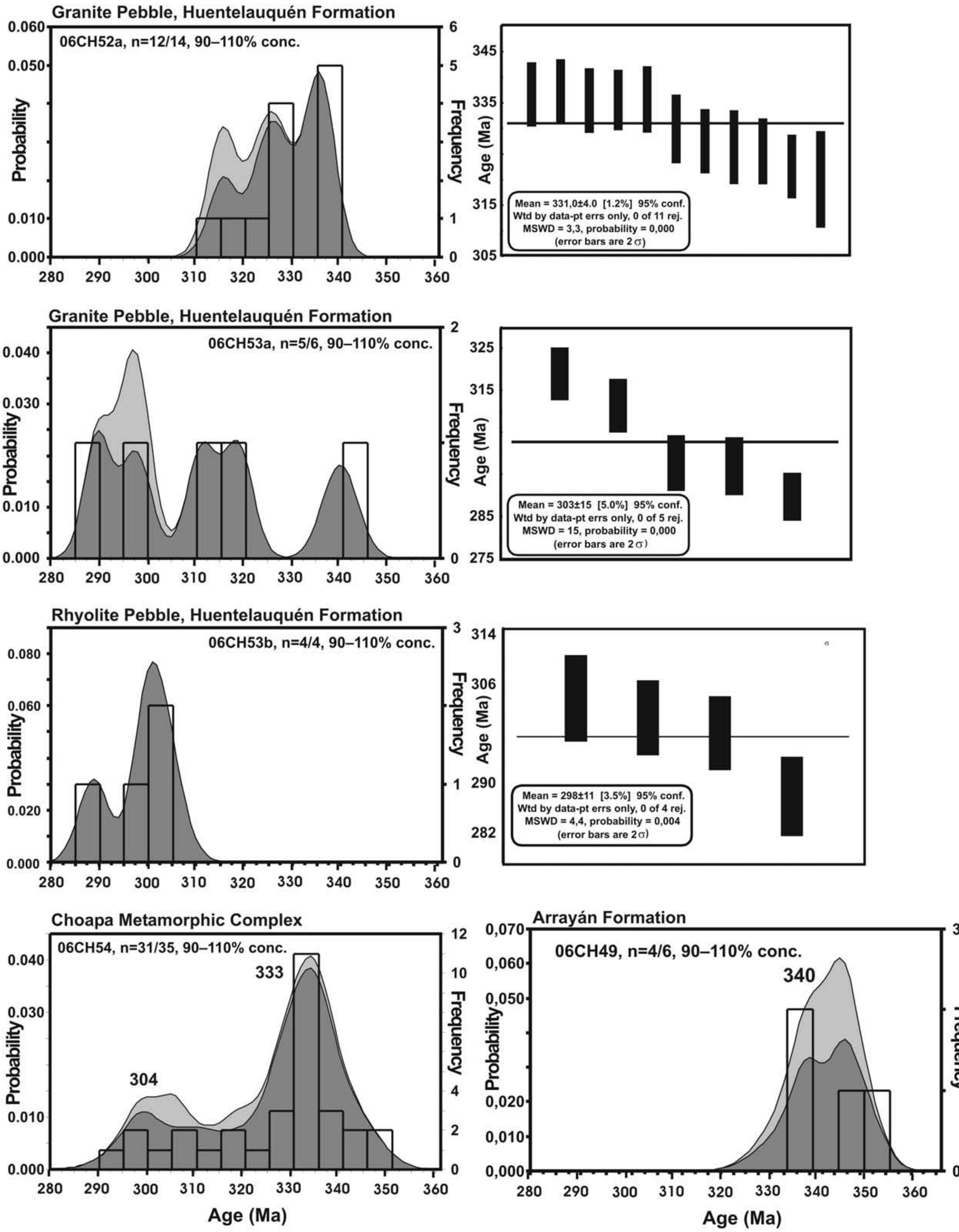


Fig. 6. Probability plots and weighted averages of Carboniferous zircon from leucogranite and rhyolite pebbles of the Huentelauquén Formation and probability plots of equivalent youngest detrital zircon population in the Choapa Metamorphic Complex and the Arrayán Formation. Background data are available in Tables E1–E7 in the electronic data base.

grain at 331 ± 12 Ma is similar to the weighted mean. The inferred maximum deposition age of 343 Ma is significantly older than that of the metasediments in the Choapa Metamorphic Complex, but consistent with a Lower Carboniferous biostratigraphical age of a part of the hosting Arrayán Formation in the frontally accreted upper level of the accretionary system (Rebelledo and Charrier, 1994).

The youngest ages of magmatic origin of the remaining three samples from the coastal accretionary systems are constrained by two detrital magmatic zircon grains in sample 02CH31 (362 ± 11 Ma; 364 ± 10 Ma), but only one in samples 02CH11 (335 ± 9 Ma), and 02CH38 (330 ± 7 Ma) yielding vaguely defined maximum depositional ages of 373 Ma, 344 Ma, and 337 Ma, respectively. Considering the late Carboniferous

Table 1
List of ages of inherited xenocryst cores and host zircon

Sample	Grain	Rim	2σ	Core	2σ
02CH11	97	335	9	444	15
02CH49	12A	409	11	553	9
02CH11	129	424	12	1999	27
02CH31	37	439	17	883	63
02CH11	75	445	10	2660	17
02CH31	75	446	13	596	18
06CH54	2	470	12	1397	57
02CH31	88	488	9	565	17
02CH49	13A	538	20	825	20
02CH11	111	559	22	1152	49
02CH31	125	565	16	647	18
02CH31	135	599	18	1037	31
02CH49	11E	1019	38	2680	16
02CH11	116	1049	49	1201	71
02CH11	90	1062	58	1361	26
02CH31	27	1088	54	2483	38
02CH38	37	1671	50	2939	21
02CH38	44	2220	31	2649	15

age of metamorphism (Willner et al., 2005), we can infer a Carboniferous sedimentation at least of the host formations of samples 02CH11 (Western Series) and 02CH38 (Agua Dulce Metaturbidites). This is even likely, when some possible lead loss in these grains is taken into account. All these zircon grains most probably originated during a local Carboniferous magmatic activity at 340–295 Ma. Only in sample 02CH31 from the Eastern Series such zircon is missing. It is worth noting that of all Carboniferous grains only one from sample 02CH11 (Table 1) contains an inherited core, which has an age of 444 Ma.

5.4. Lu–Hf isotope compositions

We selected 138 zircon grains covering all measured age populations for Lu–Hf analysis. These zircon grains have measured $^{176}Lu/^{177}Hf$ ratios of 0.003–0.0002 and present-day $^{176}Hf/^{177}Hf$ ratios of 0.2806–0.2827. This corresponds to $\varepsilon Hf_{(t)}$ values of −20 to +13, which reveal zircon protoliths derived from recycling of old crust as well as from juvenile sources (Fig. 8). Initial $^{176}Hf/^{177}Hf$ expressed as $\varepsilon Hf_{(t)}$ represent the Hf-isotope signature at the time of zircon crystallization reflecting directly the evolution stage of the crust from which it derived. This can be linked to distinct crustal evolution trends. Zircon with positive $\varepsilon Hf_{(t)}$ similar to the depleted mantle originated from juvenile crust, whereas zircon with negative $\varepsilon Hf_{(t)}$ crystallized from old recycled crust. Summarizing the information of all studied zircon, three episodes of juvenile crust formation can be distinguished by overlap of $\varepsilon Hf_{(t)}$ with the depleted mantle evolution array as well as from crustal evolution trends shown as shaded areas in Fig. 8: an Archean episode at 2.7–3.3 Ga, a Transamazonian episode at 1.9–2.3 Ga and a Mesoproterozoic episode at 0.8–1.5 Ga. Mesoproterozoic zircon from the Guarguaráz Complex (1.05–1.40 Ga) has positive $\varepsilon Hf_{(t)}$ of 0.6–13.3 except for only two grains at 1.4 Ga with εHf −2.4 and −5.9 (9% of the population). This suggests a predominantly juvenile crust in the source area. Zircon crystals with negative values appear to reflect recycled Transamazonian crust similar to the youngest grain from this complex at 0.55 Ga (εHf −10). By contrast, Meso- to Neoproterozoic detrital zircon from the coastal accretionary systems (0.7–1.4 Ga) shows a broad spectrum of $\varepsilon Hf_{(t)}$ (−7.8 to +9.9) including 37% of the grains having negative values. This suggests mixing of zircon derived from juvenile and recycled crust. In average, however, the zircon grains point to a middle Mesoproterozic crustal residence time.

Detrital zircon derived from the Famatinian, Pampean and Brasiliano Mobile Belts (0.4–0.7 Ga) have $\varepsilon Hf_{(t)}$ ranging from −20.2 to +5.8; 84% of the grains have negative values. No juvenile crust seems to have been formed during this period. Zircon rather derived from various recycled crustal sources, which are Transamazonian as well as Mesoproterozoic (trend MC2 in Fig. 8).

The Carboniferous magmatic zircon population including those of the rhyolite/granite pebbles in the Huentelauquén Formation shows – compared with the Lower Paleozoic zircon – a more juvenile signature with a rather restricted range of $\varepsilon Hf_{(t)}$ (−3.2 to +4.1) and late Mesoproterozoic to early Neoproterozoic crustal residence time (T_{DM}Hf of 0.95–1.3 Ga). Carboniferous grains originated either from mixing of juvenile, mantle-derived magma with older crust or more likely from recycling of crustal material considerably more juvenile than that represented by early Paleozoic–Precambrian detrital zircon from the coastal accretionary systems.

6. Provenance, crustal growth and geodynamic implications

6.1. Archean and Paleoproterozoic evolution

Detrital zircon in the coastal accretionary systems of central Chile dates back to Archean times of crustal formation in South America. The earliest age cluster at 2.95–3.20 Ga is as yet unknown from detrital zircon in the sediments of the Precordillera or in the metasediments of the NW–Argentine basement, but was detected in the late Paleozoic accretion systems in Patagonia (Augustsson et al., 2006). This Archean zircon must have derived – presumably by multiple recycling – from older cratons within W Gondwana such as the Amazonas, São Francisco, Congo and Rio de la Plata cratons. Most likely the latter craton is the nearest source, from where such ages are known (Rapela et al., 2007).

The earliest information provided by the εHf-values of the detrital zircon is a distinct Archean crustal evolution trend suggesting formation of new crust in W Gondwana at around 2.7–3.4 Ga. A likely area for formation of this Archean juvenile crust is the 3.1–3.4 Ga old greenstone belt of the Nico Pérez Terrane in the eastern Rio de la Plata Craton (Hartmann et al., 2001). Somewhat more abundant is the age cluster at 2.4–2.8 Ga in the coastal accretionary prisms as well as in the Guarguaráz Complex. εHf-values of these zircon grains plot on the Archean crustal evolution trend. Such ages are as well represented by some detrital zircon of the Eastern Sierras Pampeanas basement (Adams et al., 2008). This recycled source is confirmed by the presence of xenocryst cores within zircon of Mesoproterozoic and Famatinian age. The Rio de la Plata Craton, however, may also be a near direct source, where 2.5–2.6 Ga old meta-trondhjemites and mafic granulites are known (Rapela et al., 2007).

A few Transamazonian ages (1.9–2.2 Ga) are represented in zircon of the coastal accretionary systems, but not in the Guarguaráz Complex. Corresponding grains partly also occur in rocks of the Eastern Sierras Pampeanas basement (Adams et al., 2008) and as xenocryst in one Famatinian zircon grain. A direct source, however, is again the Rio de la Plata craton, where such ages dominate (Rapela et al., 2007). Hafnium isotope composition of few younger zircon, i.e. of Famatinian, Pampean, Brasiliano and Mesoproterozoic origin, partly point to recycling of crust formed during the Transamazonian cycle. εNd-data suggest formation of juvenile crust during the Transamazonian cycle in the Rio de la Plata craton (Rapela et al., 2007). The recorded Transamazonian crustal evolution path from the Hf isotopes (Fig. 8) is consistent with Nd model ages of 1.6–2.1 Ga characterizing the metasediments of the late Paleozoic accretionary system in south-central Chile (Lucassen et al., 2004) that are equivalent to those studied here. It should also be noted that some detrital zircon from the Guarguaráz complex falls on the Transamazonian crustal evolution path.

6.2. Mesoproterozoic to early Paleozoic evolution

The prominent age clusters at 1.0–1.4 Ga in the zircon of the coastal accretionary systems and that of 1.0–1.5 Ga in the Guarguaráz Complex reflect the importance of a Mesoproterozoic event on crustal growth and recycling processes in southern South America, which is in agreement with previous studies on detrital zircon of metasediments in the Argentine Sierras Pampeanas (Schwartz and Gromet, 2004; Rapela et al., 2007; Adams et al., 2008) and of Ordovician sediments from Southern Bolivia (Egenhoff and Lucassen, 2003). The Mesoproterozoic age spectrum in the metasediments of the Guarguaráz Complex is very similar to that from the Lower Paleozoic sediments of the Argentine Precordillera (Finney et al., 2005; Gleason et al., 2007) which constitute the allochthonous Cuyania Terrane. This similarity is further corroborated by the above derived Lower Paleozoic sedimentation age and former lithological comparisons by Lopez and Gregori (2004). On the other hand, the Mesoproterozoic age cluster in the zircon of the coastal accretionary systems matches that of the Precordillera (Gleason et al., 2007), but also considerable input possibly came from recycling metagreywackes in NW-Argentina, where a pronounced age maximum of 1.05–1.10 Ga was detected (Schwartz and Gromet, 2004; Rapela et al., 2007; Adams et al., 2008). A direct source of the Mesoproterozoic components in South America could have been the Arequipa Massif, the Sunsás Belt and the Punta del Este Terrane. The Western Sierras Pampeanas may also be a direct South American source, if it is not related to the Cuyania Terrane, as suggested by Rapela et al. (2007).

The high positive εHf-values of the Mesoproterozoic zircon from the Guarguaráz Complex mainly point to a distinct long-term (1.1–1.5 Ga) event of juvenile Mesoproterozoic crust formation in the close and apparently uniform source region of the sediments of the Cuyania Terrane. Recycled zircon, also from Transamazonian crust, is minor. This may not be significant, because we studied one sample only. However, the basement below the sediments of the Argentine Precordillera possibly represents such crust. This was detected by Kay et al. (1996) who had studied xenoliths brought up in Precordilleran Miocene volcanic rocks. Here U/Pb protolith ages of amphibolite and orthogneiss are at 1.1 Ga. Calculated depleted mantle Nd model ages range between 0.8 and 1.6 Ga. Kay et al. (1996) interpreted the Precordilleran basement as formed in an oceanic arc-back arc environment near a continental margin.

εHf-values of Mesoproterozoic zircon in the coastal accretionary systems also point to formation of juvenile crust at 1.1–1.5 Ga, as well as to considerable crustal recycling. Evidence for this recycling also came from four zircon grains with xenocryst cores (Table 1). The juvenile crustal signature may reflect partial input from the Cuyania Terrane, but formation of juvenile crust at 1.4–1.5 Ga has also been reported from a South American source, the Mesoproterozoic Sunsás Belt in W Brazil (Condie et al., 2005). Furthermore Augustsson et al. (2006) discovered a ~1 Ga old detrital zircon population in Patagonia that originated from juvenile crust. Hence, the same event of Mesoproterozoic juvenile crustal formation presumably took place in the source region of the Cuyania sediments as well as in South America itself. However, it seems that more detritus from recycled crust was contributed by South American sources.

Late Proterozoic zircon (556–722 Ma) from the sediments of the coastal accretionary systems derived – presumably partly by recycling – from the Brasiliano orogeny to the east of the Pampean orogen, whereas maxima at 779–982 Ma could correspond to the youngest detrital zircon ages from the Argentine Precordillera (Gleason et al., 2007). The Braziliano age cluster overlaps with the youngest detrital zircon ages observed in the early Paleozoic basement of Argentina (Schwartz and Gromet, 2004; Rapela et al., 2007; Adams et al., 2008).

The provenance of the predominant late Paleozoic zircon from the sediments of the coastal accretionary systems (360–550 Ma) is evident. Such ages are known from magmatic and metamorphic zircons grown during the Pampean and Famatinian orogenies in the Lower Paleozoic mobile belt of NW-Argentina, but not from Precordillera sediments (Gleason et al., 2007). Age clusters at 510–547 Ma represent the Pampean orogeny in the Eastern Sierras Pampeanas, those at 423–488 Ma can unequivocally be attributed to the Famatinian orogeny further west and those at 363–404 Ma presumably to post-Famatinian intrusions.

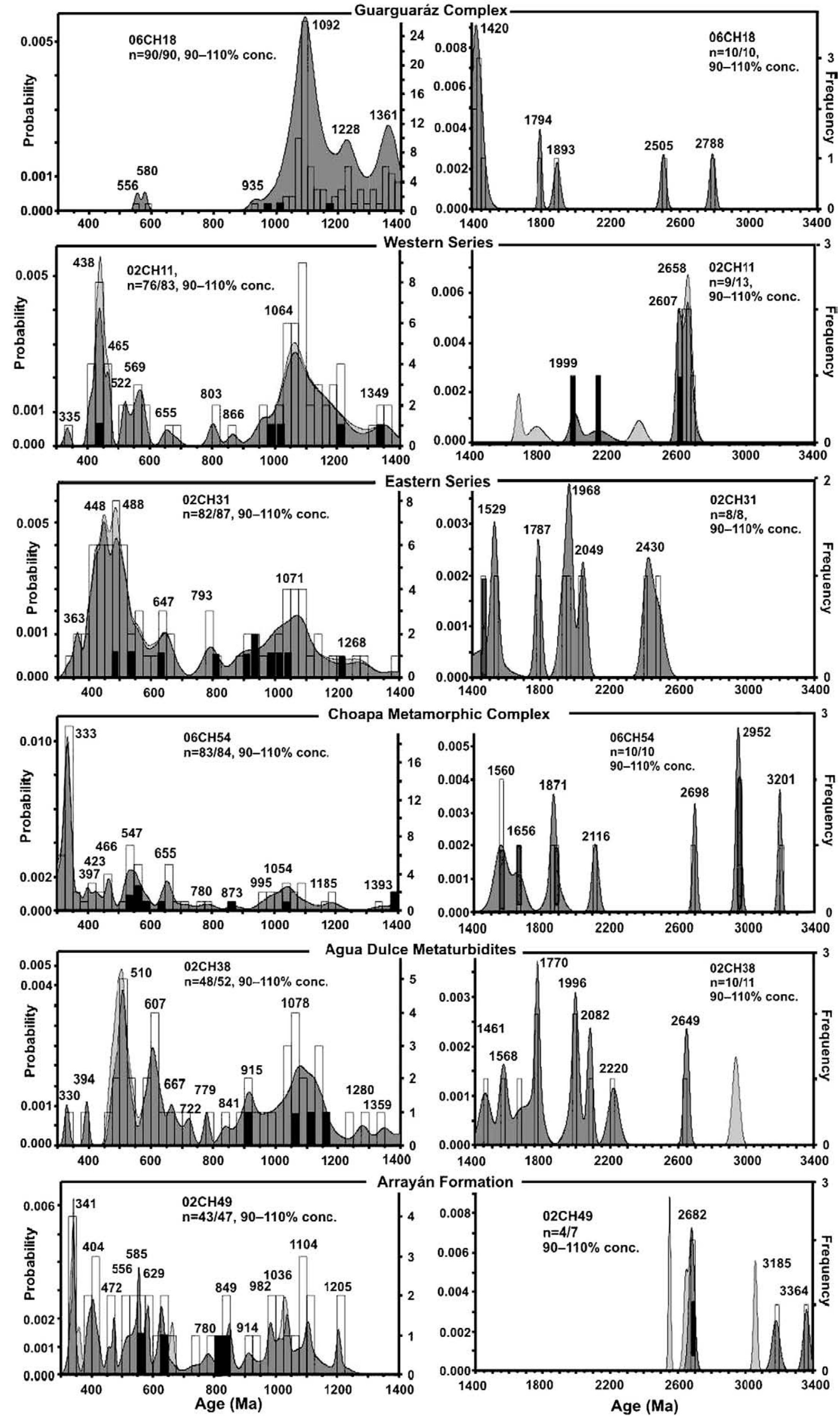
Few zircon of possible metamorphic origin within the Pampean, Famatinian and Brasiliano age clusters is consistent with the existence of some high grade metamorphic rocks in these three source regions. The predominantly negative $\varepsilon Hf_{(t)}$ of Neoproterozoic and Lower Paleozoic zircon grains indicate crustal recycling of Transamazonian as well as Mesoproterozoic crust (Fig. 8). This is confirmed by eleven grains with xenocryst cores. It furthermore gives some hints at the nature of the crust below the extended metasediments of the Eastern Sierras Pampeanas that also contains Mesoproterozoic rocks. No juvenile crust seems to have been formed after 660 Ga. By contrast, three grains with ages of 688, 724 and 838 Ma have high positive $\varepsilon Hf_{(t)}$ of 5.8, 7.7 and 6.0, respectively. These data point to derivation from a crust with a highly juvenile component, evidently the youngest juvenile crust produced in South America which can possibly be related to the break-up of the Rodinia supercontinent (see Rapela et al., 2007, for summary).

6.3. The Chilenia terrane and the evolution of the coastal accretionary prisms

The age spectrum of the detrital zircon from the Guarguaráz Complex and the assumed Lower Paleozoic depositional age of its metasediments suggest that mainly sediments derived from the Cuyania Terrane were subducted within its western suture zone forming the HP collisional Guarguaráz Complex (Massonne and Calderón, 2008). This contrasts with earlier suggestions by Lopez and Gregori (2004) that the Guarguaráz Complex would represent the accretion complex corresponding to the Famatinian magmatic arc. In this case, Paleozoic detrital zircon would have been incorporated in accreted marginal sediments as it is in the coastal accretionary systems of central Chile. It rules out the possibility that the latest lateral growth of the Pacific margin of South America after emplacement of the Cuyania Terrane started by accretion of marginal sediments at its western flank. Instead, it supports the collision with a further microcontinent to the west of Cuyania (Davis et al., 1999; Ramos 2000; Gerbi et al., 2002; Ramos, 2004), the “Chilenia Terrane”.

There are no direct outcrop areas of this hypothetical terrane. However, our zircon studies point to a magmatic event that ought to have originated within the Chilenia Terrane: The youngest detrital zircon in the coastal accretionary systems at 30°–32° S and the dated subvolcanic leucogranite/rhyolite pebbles of the Huentelauquén Fomation define a distinct magmatic event within an age range between 346±7 and 294±4 Ma, which by far exceeds known maximum crystallization ages of the late Paleozoic magmatic arc (305–310 Ma; Hervé et al., 1988; Mpodozis and Kay 1992; Willner et al., 2005) as well as the maximum age of accretion in north-central Chile of ~320 Ma (Willner et al., 2005). The source of the Carboniferous zircon must be local and close to the coastal accretionary prism according to the characteristics of the magmatic pebbles of the conglomerate of the Huentelauquén Formation (presence of subangular boulders and dominance of the clast assemblage by magmatic pebbles of one source). Most likely magmatism occurred within the basin of the Carboniferous sediments, i.e. in the area of the present Argentine Frontal Cordillera and/or to the west of it and, thus, within the hypothetical Chilenia Terrane. On the other hand, for some of the youngest detrital zircon a derivation from a late-Famatinian source as far as to the east of the present Argentine Precordillera cannot be completely excluded.

Fig. 7. U/Pb age spectra (combined probability/histogram plots) of Precambrian and early Paleozoic detrital zircon. Ages that are >90% concordant are represented by dark grey shading in the spectra, those with <90% concordance by light grey shading. Metamorphic zircon is marked black in the histograms, magmatic zircon white. Background data are available in Tables E1–E7 in the electronic data base.


Guarguaráz Complex
06CH18
n=90/90, 90–110% conc.
556
580
935
1092
1228
1361
06CH18
n=10/10,
90–110% conc.
1420
1794
1893
2505
2788
Western Series
02CH11,
n=76/83, 90–110% conc.
335
438
465
522
569
655
803
866
1064
1349
02CH11
n=9/13,
90–110% conc.
1999
2607
2658
Eastern Series
02CH31
n=82/87, 90–110% conc.
363
448
488
647
793
1071
1268
02CH31
n=8/8,
90–110% conc.
1529
1787
1968
2049
2430
Choapa Metamorphic Complex
06CH54
n=83/84, 90–110% conc.
333
397
423
466
547
655
780
873
995
1054
1185
1393
06CH54
n=10/10
90–110% conc.
1560
1656
1871
2116
2698
2952
3201
Agua Dulce Metaturbidites
02CH38
n=48/52, 90–110% conc.
330
394
510
607
667
722
779
841
915
1078
1280
1359
02CH38
n=10/11
90–110% conc.
1461
1568
1770
1996
2082
2220
2649
Arrayán Formation
02CH49
n=43/47, 90–110% conc.
341
404
472
556
585
629
780
849
914
982
1036
1104
1205
02CH49
n=4/7
90–110% conc.
2682
3185
3364
Probability
Frequency
Age (Ma)

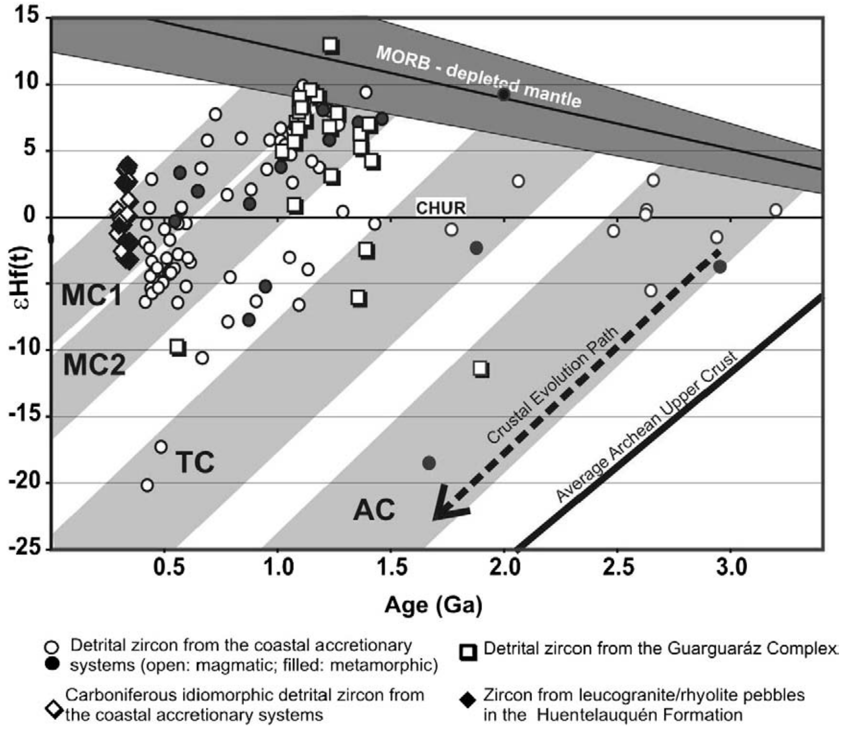


Fig. 8. Comparative $\varepsilon Hf_{(t)}$ evolution diagram for single detrital grains. $\varepsilon Hf_{(t)}$ was calculated using a decay constant of 1.865×10^{-10} (Scherer et al., 2001), a CHUR $^{176}Lu/^{177}Lu$ and $^{176}Hf/^{177}Hf$ ratio of 0.0332 and 0.282772, respectively (Blichert-Toft and Albarede, 1997) and the ages obtained for the respective zircon cores. The depleted mantle array was calculated using data for MOR-basalts (Patchett et al., 1981) and the crustal evolution path assuming a crustal $^{176}Lu/^{177}Lu$ ratio of 0.0093 (average of granitoid data from Vervoort and Patchett, 1996; see also Scherer et al., 2007). 2σ-errors generally are equal or smaller than the size of symbols. Shaded areas represent crustal evolution trends of zircon that might originate from common crustal domains (AC Archean crust; TC Transamazonian crust; MC Mesoproterozoic crust). The evolution trend of average Archean upper crust was calculated by Condie et al. (2005) assuming a crustal $^{176}Lu/^{177}Lu$ ratio of 0.013. Background data are available in Table E8 in the electronic data base.

The restricted range of Hf isotope signatures of the Carboniferous zircon population suggests the existence of a distinct basement below Devonian to Carboniferous sediments in the west of the Argentine Precordillera. It can be interpreted as due to crustal recycling. However, this recycled crust is different from recycled crust represented by the source of the early Paleozoic detrital zircon. Negative $\varepsilon Hf_{(t)}$-values of the detrital Carboniferous magmatic zircon population do not overlap with the Transamazonian and the early Mesoproterozoic crustal evolution trends (trends TC and MC2 in Fig. 8), but with a trend evolving from a relatively young juvenile crust formed at 0.8–1.2 Ga (trend MC1 in Fig. 8). Generation by partial melting of a late Mesoproterozoic crust appears most likely. Alternatively, melting of Carboniferous sediments of the coastal accretionary systems themselves appears unlikely. In this case more negative εHf-values reflecting higher model ages must appear due to the high content of older crustal components in the detritus. This was furthermore shown by Lucassen et al. (2004) to have occurred under the late Paleozoic magmatic arc in south-central Chile: the Upper Paleozoic sediments have Nd model ages of 1.6–2.1 Ga similar to the produced melts. An alternative interpretation of the difference between the signatures of the Carboniferous and the early Paleozoic zircon in our study is to suggest a homogeneous mixture of old recycled crustal components with juvenile components in a high temperature granitic magma. Such melt could be represented by the Carboniferous leucogranite and rhyolite pebbles. However, six of seven bulk analyses of these rocks show that they are corundum-normative and, thus, rather S-type granitoids (see Table E9 in the electronic data set). On the other hand, the proposed ~0.8–1.2 Ga old crust is partly similar to the crust below the adjacent Cuyania Terrane, where Kay et al. (1996) yielded depleted mantle Nd model ages between 0.8 and 1.6 Ga from xenoliths.

Some chemical characteristics of the leucogranite and rhyolite pebbles point to similarity with volcanic arc and within-plate granite as shown in the Rb/Y+Nb discrimination diagram of Pearce et al. (1984), where the Carboniferous granitoid magma follows a similar trend as calc-alkaline granitoids from the Famatinian magmatic arc (Fig. 9; Table E9 in the electronic data set). Some authors had already proposed a west-dipping subduction zone before collision of the Chilenia and Cuyania Terranes by structural evidence observed in the basement of the Argentine Frontal Cordillera (Davis et al., 1999; Gerbi et al., 2002). In this scenario the Carboniferous granitoids could have represented the corresponding magmatic arc. Nevertheless, the Carboniferous event sets important time marks for the evolution of the late Paleozoic coastal accretionary systems of Chile:

- All metasediments of the coastal accretionary systems except those of the Eastern Series are shown to be Carboniferous being deposited after 344 Ma (maximum depositional age of the Western Series). This age could also be the minimum age of the accretion of the Chilenia Terrane to the continental margin. However, Devonian sediments are already overlying the hypothetical Chilenia basement in Argentina (Caminos, 1979). These findings point to a short time of existence of a passive continental margin. Bahlburg and Hervé (1997) postulated passive margin conditions north of 27°S for even 100 Ma (~410–310 Ma) due to absence of magmatic and tectonic activity. Due to the emplacement of the Chilenia Terrane this time should have been shorter in central Chile.
- The youngest maximum depositional ages of the Choapa Metamorphic Complex (308 Ma) and the Huentelauquén Formation (303 Ma) coincide. This further proves that the Huentelauquén Formation was deposited in a retrowedge basin. The significantly younger maximum depositional age of the basally accreted Choapa Metamorphic Complex compared to those of the frontally accreted Agua Dulce Metaturbidites (337 Ma) and Arrayán Formation

(343 Ma) corroborates the observation of Richter et al. (2007) that the mode of accretion changed from frontal to basal in central Chile through time. Furthermore it is implied that the change of the accretion mode occurred before the deposition of the protoliths of the Choapa Metamorphic Complex, i.e. before 308 Ma. This corroborates the finding of Willner et al. (2005) that basal accretion was ongoing in central Chile at ~305 Ma.

7. Conclusions

Isotopic signatures of detrital zircon from the late Paleozoic coastal accretionary systems of central Chile unravel multiple crustal growth and recycling processes in the interior of South America. Direct contribution of the basement of the hypothetical Chilenia Terrane to the detritus of the Carboniferous sediments can be excluded. Up to the present Argentine Frontal Cordillera the whole area east of the coastal accretionary systems was covered by Carboniferous sediments suggesting large-scale Carboniferous subsidence. There was no erosional barrier at that time. Large river systems supplied detritus from extensive areas in the interior of the continent to the continental margin (Fig. 10).

Three major crustal evolution trends starting with three episodes of juvenile crustal formation and evolving towards more recycled crust were detected by studying εHf-variation of all detrital zircon populations: an Archean trend starting with juvenile crustal formation at 2.7–3.4 Ga, a Transamazonian trend (1.9–2.4 Ga) and a Mesoproterozoic trend (0.8–1.5 Ga). These trends can also be related to proposed source regions which are: the Brasiliano orogen in southern Brazil, the Rio de la Plata Craton, the Pampean and Famatinian orogens in Argentina, several small outcrop areas of Mesoproterozoic rocks in southern South America as well as the allochthonous Cuyania Terrane. The detected three episodes of juvenile crustal formation also coincide with global peaks of juvenile magma productivity interpreted as rapid crustal growth due to origin of mantle superplumes (Stein and Hofmann, 1994). Mesoproterozoic zircon is notably present, generally with a juvenile crustal signature in the Guarguaráz Complex of the Argentine Frontal Cordillera and with a mixed juvenile and recycled signature in the Chilean late Paleozoic coastal accretionary prisms. However, unequivocal discrimination between a Cuyania and a South American source appears to be impossible.

On the other hand, Paleozoic zircon that crystallized from arc-related calc-alkaline magmas generally represents recycled crust. Yet there is a considerable difference between zircon of Brasiliano, Pampean and Famatinian origin (360–660 Ma) and zircon of a newly detected Carboniferous magmatic event (298–346 Ma). Whereas the early Paleozoic zircon represents a considerable range of recycled crust (Transamazonian as well as Mesoproterozoic), the Carboniferous zircon crystallized from a relatively young late Mesoproterozoic crust. This crust most likely occurred west of the Cuyania Terrane and may represent the basement of the Chilenia Terrane that is supposed to have collided with Cuyania previously accreted to Gondwana. Existence of this terrane is further confirmed by the presence of a presumable collisional accretionary prism, the Guarguaráz Complex. Here detrital zircon shows that its metasediments are most likely equivalents to the sediments of Cuyania that were subducted and subjected to HP metamorphism. However, the age of the HP-metamorphism has still to be confirmed. The possibility that the late Paleozoic coastal accretionary systems are also present under the Carboniferous basin up to the margin of Cuyania can be ruled out. The Carboniferous subduction related magmatic event at 294–346 Ma could represent an earlier magmatic arc related to a supposed westward subduction under the Chilenia Terrane (Davis et al., 1999).

The detrital zircon of the Carboniferous magmatic event sets time marks for the evolution of the coastal accretionary prisms themselves. With exception of the Eastern Series deposition of the accreted sediments took place after 344 Ma, i.e. shortly before and during late Paleozoic accretion starting at ~320 Ma in central Chile (Willner et al., 2005). The significantly older frontally accreted metasediments compared to the basally accreted metasediments in the coastal accretionary system indicate that the marked change of the accretion mode must have occurred before 308 Ma. Later sedimentation of the basally accreted metasediments of the Choapa Metamorphic Complex occurred at the same time as those of the less deformed Huentelauquén Formation forming part of a retrowedge basin that extends as far as to the present Argentine Frontal Cordillera.

After the end of accretion processes at ~220 Ma in central Chile lateral growth of the Pacific continental margin ended (Willner et al., 2005), but intensive magmatic activity in the Mesozoic magmatic arc was accompanied by considerable addition of juvenile magma from the mantle (Lucassen et al., 2004). With this event crustal growth ended at the Pacific margin of southern South America.

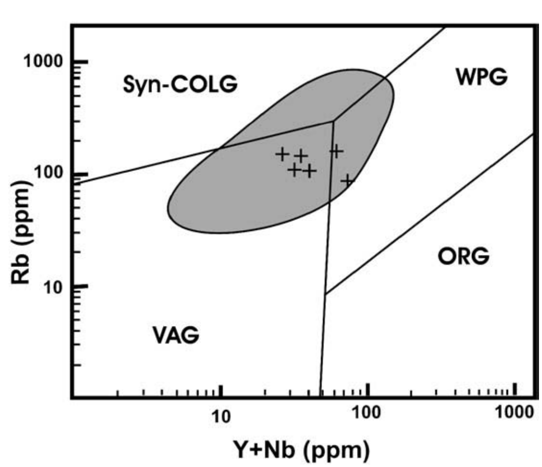


Fig. 9. Rb versus Y+Nb plot for composition of leucogranite and rhyolite pebbles in the conglomerate of the Huentelauquén Formation (crosses). Fields after Pearce et al. (1984): VAG — volcanic arc granite; ORG — ocean ridge granite; WPG — within plate granite; syn-COLG — syncollisional granite. The field is for the distribution of Famatinian granitoids after Quenardelle and Ramos (1999). Background data are available in Table E9 in the electronic data base.

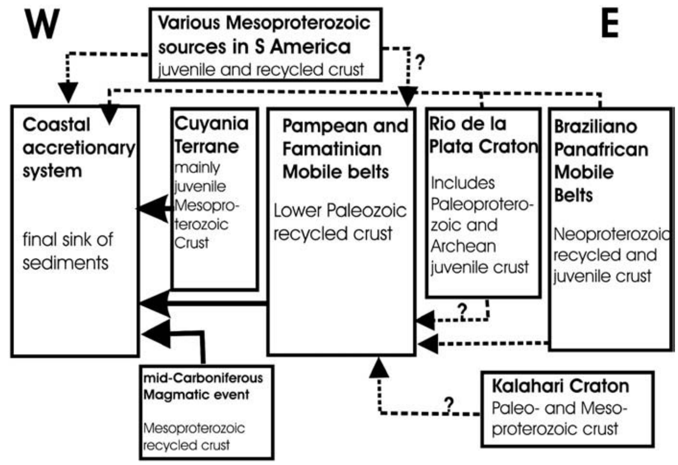


Fig. 10. Summary of provenance relationships and derivation of recycled versus juvenile crust in Southern South America.

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Appendix A

Comparison of results measured by LA-ICPMS versus SHRIMP

Sample-grain	Measurements by LA-ICPMS at Frankfurt University					Measurements by SHRIMP at Curtin University Perth				
	$^{206}Pb/^{238}U$	1σ (Ma)	$^{207}Pb/^{206}Pb$	1σ (Ma)	conc%	$^{206}Pb/^{238}U$	1σ (Ma)	$^{207}Pb/^{206}Pb$	1σ (Ma)	conc%
02CH49-4b	331	6	360	48	93	341	4	568	49	60
02CH49-5b	992	12	964	18	103	982	8	977	21	100
02CH49-5a	1096	12	1090	25	101	1110	12	1125	31	99
02CH49-9d	390	5	380	37	103	394	6	393	16	100
02CH49-9c	850	9	864	18	98	947	15	874	99	108
02CH49-10b	388	5	409	34	95	394	3	407	23	97
02CH49-14a	631	6	628	10	100	623	10	572	19	109

Localities of selected samples

Sample	Location	Latitude	Longitude	Rock type	Unit/Formation
02CH11	Coast N′ Bucalemu	34°34.20′ S	72°04.05′ W	Metagreywacke	Western Series
02CH31	Huinganes station	35°25.10′ S	72°19.50′ W	Metagreywacke	Eastern Series
06CH54	Playa La Cebada	30°58.24′ S	71°39.08′ W	Metagreywacke	Coapa Metamorphic Complex
02CH38	SE Punta Tomás	31°31.22′ S	71°34.34′ W	Metagreywacke	Agua Dulce Metaturbidites
02CH49	SE Punta Tomás	31°31.44′ S	71°34.05′ W	Metagreywacke	Arrayán Formation
06CH52a	Road cut S′ Huentelauquén	31"35.08′ S	71°31.92′ W	Granite pebble Conglomerate	Huentelauquén Formation
06CH53a	Cliff W′Mantos de Hornillo	31°09.22′ S	71°39.78′ W	Granite pebble/ Conglomerate	Huentelauquén Formation
06CH53b	Cliff W′Mantos de Hornillo	31°09.22′ S	71°39.78′ W	Rhyolite pebble/ Conglomerate	Huentelauquén Formation
06CH18	S′Refugio Don Domingo	33°23.02′ S	69°27.00′ W	Garnet micaschist	Guarguaráz Complex

Appendix B. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.chemgeo.2008.04.016.

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