The Late Paleozoic tectonometamorphic evolution of Patagonia revisited: Insights from the pressure-temperature-deformation-time (P-T-D-t) path of the Gondwanide basement of the North Patagonian Cordillera (Argentina)

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Key points

- Monazite ages of ca. 300 Ma record peak conditions of 650 °C and 11 kbar achieved during prograde metamorphism and progressive deformation

- Late Paleozoic crustal thickening of Gondwanide basement coeval with forearc metamorphism resulted from a transpressional advancing orogen

- Low-grade metamorphism and transtension at ca. 170 Ma and hydrothermal activity at ca. 100-80 Ma during the onset of Andean transpression
Abstract

Combined field structural analysis with in situ EPMA (electron probe microanalysis) Th-U-Pb monazite dating, petrologic and microstructural data provide a reconstruction of the pressure-temperature-deformation-time (P-T-D-t) path of the Gondwanide basement of the North Patagonian Cordillera. For samples from the Challhuaco hill, the timing of development of the metamorphic S2 foliation and associated L2 lineation and tight to isoclinal F2 folds is constrained by monazite ages of 299 ± 8 and 302 ± 16 Ma during peak metamorphic conditions of ca. 650 °C and 11 kbar, achieved during prograde metamorphism and progressive deformation. Metamorphism and deformation of metamorphic complexes of the North Patagonian Andes seem to record Late Paleozoic crustal thickening and are coeval with metamorphism of accretionary complexes exposed further west in Chile, suggesting a coupled Late Devonian-Carboniferous evolution. Instead of the result of continental collision, the Gondwanide orogeny might thus be essentially linked to transpression due to advancing subduction along the proto-Pacific margin of Gondwana. On the other hand, a second generation of monazite ages of 171 ± 9 and 170 ± 7 Ma constrain the timing of low-grade metamorphism related to kink band and F3 open fold development during Jurassic transtension and emplacement of granitoids. Finally, a Cretaceous overprint, likely resulting from hydrothermal processes, is recorded by monazite ages of 110 ± 10 and 80 ± 20 Ma, which might be coeval with deformation along low-grade shear zones during the onset of Andean transpression.

Keywords: EPMA Th-U-Pb monazite dating, in situ petrochronology, garnet metapelite thermobarometry, Upper Paleozoic, Terra Australis, southwestern Gondwana
1. Introduction

The Late Paleozoic Gondwanide orogen was early recognized in central Argentina by Keidel (1938 and references therein), allowing to establish one of the pioneer geologic correlations across the Atlantic Ocean. However, the tectonic significance of this key orogenic event is still under debate and has been alternatively attributed to subduction, collisional or accretional tectonics (e.g., von Gosen, 2003; Pankhurst et al., 2006; Gregori et al., 2008, 2016; Ramos, 2008; Varela et al., 2015; Vizán et al., 2015, 2017; Hervé et al., 2016; Heredia et al., 2017, 2018). These proposals have in turn led to contrasting hypotheses about the origin of the North Patagonian Massif, for which an allochthonous, autochthonous or para-autochthonous origin has been inferred (Ramos, 1984, 2008; Dalla Salda et al., 1992; Pankhurst et al., 2006, 2014; Gregori et al., 2008; González et al., 2011, 2012, 2018; Rapalini et al., 2010, 2013; Ramos & Naipauer, 2014; Castillo et al., 2017; Heredia et al., 2018).

In the particular case of northwestern Patagonia (Argentina), most contributions focused on the Late Paleozoic magmatism (Cerredo & López de Luchi, 1998; Pankhurst et al., 2006; Urraza et al., 2011; Varela et al., 2005, 2015; Castillo et al., 2017), whereas constraints on the timing and characteristics of contemporaneous deformation and metamorphism are scarce (Fig. 1; Giacosa et al., 2004; von Gosen & Loske, 2004; García-Sanseguido et al., 2009; von Gosen, 2009; Martínez et al., 2012; Urraza et al., 2015). In addition, no clear correlations have been so far established between the igneous-metamorphic basement of the North Patagonian Cordillera/Massif and accretionary complexes exposed further west in Chile in terms of their structural and metamorphic evolution, though the latter record coeval Late Paleozoic sedimentation and subsequent deformation and metamorphism (Martin et al., 1999; Willner et al., 2004; Kato et al., 2008; Hervé et al., 2013). Recently, Hervé et al. (2016, 2018) identified juvenile Devonian magmatism in the accretionary complex of Chile south of 40 °S and thus inferred the Late Devonian-Early Carboniferous accretion of an island arc complex.
along the southwestern margin of Gondwana. This hypothesis contrasts significantly with the proposal of Martínez et al. (2012), who suggested the collision of the Chilenia Terrane at ca. 390 Ma, well-documented north of 40 °S (Willner et al., 2011).

On the other hand, a correct understanding of the coupling between deformation and metamorphism is critical for the reconstruction of P-T-D-t paths of deformed metamorphic rocks (Williams & Jercinovic, 2012) and their significance for orogenic evolution. In this sense, combined in situ petrologic and microstructural data allow to link deformation mechanisms, metamorphic reactions and macroscopic structural features and to infer the relative timing of different mineral associations and microstructures (Johnson, 1999; Marsh et al., 2009; Hobbs et al., 2011; Skrzypek et al., 2011; Goncalves et al., 2012; Yakymchuk & Godin, 2012; Oriolo et al., 2018). The absolute timing of deformation and metamorphism, in turn, can be constrained by interpreting in situ geochronologic data in the light of microstructures, which provide insights into mineral equilibria, deformation mechanisms and metamorphic reactions (Mulch & Cosca, 2004; Tchato et al., 2009; Williams & Jercinovic, 2012; Villa et al., 2014; Gasser et al., 2015; Oriolo et al., 2016a, 2018; Schulz, 2017; Bosse & Villa, 2019). Furthermore, the integration of structural, microstructural, petrologic and geochronologic data permits reconstructing the crustal architecture of orogens and contributes on the understanding of their tectonic and geodynamic evolution (e.g., Goscombe & Gray, 2009; Raimondo et al., 2010; Endo & Wallis, 2017).

In this contribution, new structural, microstructural, petrologic and EPMA Th-U-Pb monazite data from gneisses and schists of basement rocks of the North Patagonian Cordillera (southwestern Argentina) are shown, providing a reconstruction of the P-T-D-t path of the Gondwanide basement of the study area. Based on these results, a revised model for the tectonic evolution of northern Patagonia is presented, providing insights into the coupled evolution of the igneous-metamorphic basement of North Patagonian Cordillera/Massif and accretionary
complexes of Chile, and the Late Paleozoic configuration of the southwestern Gondwana margin. In addition, geochronologic and structural constraints on the Mesozoic tectonic evolution of the North Patagonian Cordillera are included as well, based on the Jurassic-Cretaceous record of basement rocks.

2. Regional setting

The North Patagonian Cordillera (southwestern Argentina) is located at the southern segment of the Central Andes and is bounded to the west and east by the Central Valley of Chile and the North Patagonian Massif, respectively (Fig. 1). This morphostructural unit comprises an Andean thick-skinned fold and thrust belt and is essentially made up of Paleozoic igneous-metamorphic rocks, Mesozoic granitoids and subordinated volcano-sedimentary rocks, and Cenozoic volcano-sedimentary sequences and granites (González Díaz, 1982; Rapela et al., 1988; Dalla Salda et al., 1991; Giacosa et al., 2001, 2005; Giacosa & Heredia, 2004; Castro et al., 2011a; Bechis et al., 2014a; Iannelli et al., 2017).

In the study area, located to the south of the Nahuel Huapi lake (Fig. 2), the basement is considered to be part of the Colohuincul Complex (Dalla Salda et al., 1991) and comprises mostly paragneisses and schists with subordinated intercalations of amphibolites, metarhyolites, felsic orthogneisses and foliated intrusions (Dalla Salda et al., 1991; García-Sansegundo et al., 2011). A structural characterization was presented by Dalla Salda et al. (1991) and García-Sansegundo et al. (2009), though quantitative structural data are still scarce. Martínez et al. (2012) obtained EPMA Th-U-Pb monazite ages of 391.7 ± 4.0 and 350.4 ± 5.8 Ma for migmatitic paragneisses near the Brazo Tronador, interpreted as the timing of peak and retrograde metamorphic conditions, respectively. For these rocks, metamorphic conditions of ca. 612 °C and 4.9 kbar were determined. In addition, high-pressure metamorphism with peak
conditions of ca. 440 °C and 18 kbar were calculated for a garnet-bearing micaschist exposed at the southern margin of the Gutiérrez lake (Martínez et al., 2012). In this area, detrital zircons of metasedimentary rocks yield Early Paleozoic maximum sedimentation ages (Hervé et al., 2018).

Deformation and metamorphism of basement rocks of the North Patagonian Cordillera and the western North Patagonian Massif have been interpreted as the result of Late Paleozoic compression/transpression during the Gondwanide Orogeny (Giacosa et al., 2004; García-Sansegundo et al., 2009; von Gosen, 2009). Within this framework, rocks of the study area were assigned to a hinterland position (García-Sansegundo et al., 2009). In contrast, Martínez et al. (2012) inferred a forearc setting for these rocks, associated with an older, Devonian collisional event, as recorded north of 40 °S (Willner et al., 2011). The validity of this model was, however, questioned by Hervé et al. (2016, 2018). These authors proposed a Late Devonian-Early Carboniferous accretion of an island arc complex, considering igneous-metamorphic basement rocks of the North Patagonian Cordillera and the western North Patagonian Massif as remnants of a Devonian continental arc.

On the other hand, the Jurassic record of the study area is characterized by granitoids of the Cordilleran batholith yielding U-Pb SHRIMP zircon crystallization ages of ca. 176-160 Ma (Castro et al., 2011a) and scarce volcano-sedimentary sequences (Giacosa et al., 2001). Based on pluton fabrics, Castro et al. (2011a) inferred regional sinistral strike-slip deformation, thus contrasting with the Middle-Late Jurassic extensional regime recorded by extensional structures of volcanic rock depocenters of the North Patagonian Cordillera located further south (Echaurren et al., 2016, 2017).

Finally, Cretaceous to Cenozoic rocks of the North Patagonian Cordillera are intimately related to Andean tectonics, with two main Middle-Late Cretaceous and Miocene orogenic
phases (Orts et al., 2012; Gianni et al., 2018; Horton, 2018). In contrast to the Miocene transpressional phase, which is well-recorded by structural, kinematic and geochronologic data of volcano-sedimentary and granitic rocks (Diraison et al., 1998; Giacosa et al., 2001, 2005; Giacosa & Heredia, 2004; Orts et al., 2012; Bechis et al., 2014a), the Cretaceous Andean record of the study area is only restricted to few granitoids yielding K-Ar and Rb-Sr ages of ca. 120-80 Ma (González Díaz, 1982, and references therein). However, coeval thrusting was reported in adjacent regions of northwestern Patagonia (Orts et al., 2012; Echaurren et al., 2016, 2017; Gianni et al., 2018).

3. Methodology

Field mapping and structural analysis were applied in different exposures of the igneous-metamorphic basement of the North Patagonian Cordillera south of the Nahuel Huapi Lake (Fig. 2), where samples were also collected for microstructural, geochronologic and thermobarometric analysis. EPMA Th-U-Pb monazite data combined with SEM-based automated mineralogic methods were obtained in three samples of metasedimentary rocks (BA 20-17, BA 22-17 and BA 11-18). Sample descriptions and locations are presented in Text S1 in the Supporting Information, whereas analytical details are provided in Text S2 in the Supporting Information. In addition, thermobarometry was applied in sample BA 22-17 in order to quantify pressure-temperature metamorphic conditions (Text S2 in the Supporting Information; Powell & Holland, 1994, 2008).
4. Results

4.1. Structure

The most conspicuous macroscopic fabric element is a metamorphic \( S_2 \) foliation (Fig. 2), though relics of bedding planes \( S_0 \) and a metamorphic \( S_1 \) foliation are occasionally observed as well (Figs. 3a, 3b). \( S_1 \) represents an axial plane foliation of isoclinal \( F_1 \) folds affecting \( S_0 \) planes, whereas \( S_2 \) represents an axial plane foliation of tight to isoclinal \( F_2 \) folds that fold \( S_1 \) (Fig. 3c), which is clearly observed at the microscale (see Section 4.2). Rootless \( F_2 \) hinges are common, being defined by folded quartz segregations, and in some cases, \( F_2 \) folds are superimposed to isoclinal \( F_1 \) folds (Fig. 3d). The orientation of \( S_2 \) planes is variable, despite showing a dominant NNW-SSE to WNW-ESE strike (Fig. 3b). A stretching lineation \( L_2 \) defined by shape-preferred orientation of micas and/or staurolite is associated with \( S_2 \), exhibiting subhorizontal plunges towards NNW-SSE to WNW-ESE, which is parallel to the strike of \( S_2 \) planes in most cases (Fig. 2).

Kink bands and late gentle to open \( F_3 \) folds affecting \( S_2 \) are also present exhibiting, in many cases, a kink geometry (Fig. 3e, 3g). Though variable, \( F_3 \) axial planes strike dominantly NE-SW and exhibit subvertical to steep dip towards SE. Fold hinge lines, in turn, show moderate plunge towards ENE and E. A spaced \( S_3 \) foliation is occasionally observed as an axial plane foliation of \( F_3 \). On the other hand, a mylonitic \( S_3 \)’ foliation is also developed within low-grade shear zones (Fig. 3f), comprising mostly ultramylonites, phyllonites and pseudotachylites. Field relationships between kink bands and shear zones are not exposed. In some cases, a crenulation cleavage is associated with the mylonitic foliation, being localized within shear zone lozenges (Fig. 3h), whereas a cataclastic overprint is observable as well.

The schematic structural profile of Figure 4 summarizes the relationship of different structural elements for the western margin of the Jakob lake (Fig. 2). In this case, \( S_2 \) planes
strike dominantly NNW-SSE and dip steeply towards ENE, whereas $L_2$ lineations plunge gently towards SSE. Tight to isoclinal $F_2$ folds have subvertical hinge lines. Conjugated subvertical NE-SW- and NW-SE-striking kink bands exhibit subvertical axes. In a similar way, subvertical low-grade ductile to brittle-ductile $S_3$' shear zones strike dominantly NE-SW to NNE-SSW and show a dextral sense of shearing.

4.2. Microstructures

Schists are mostly made up of muscovite + biotite + quartz + plagioclase and characterized by the $S_2$ foliation, which typically comprises a spaced schistosity. Schistosity domains are mostly made up of muscovite and subordinated biotite, whereas microlithons encompass quartz and, to a minor extent, plagioclase. $S_2$ is commonly observed as a crenulation schistosity, due to the presence of relics of folded $S_1$ planes in the microlithons of $S_2$ (Fig. 5a). In turn, $S_1$ is characterized by muscovite + quartz + biotite ± plagioclase. In addition, garnet and plagioclase porphyroblasts are frequently observed in the metapelitic schists (Fig. 5b). Staurolite, K-feldspar and sillimanite are occasionally present as well. Monazite, rutile, tourmaline, zircon, apatite and titanite are accessory minerals.

Muscovite and biotite present shape-preferred orientation parallel to $S_1$ and $S_2$ planes, whereas quartz shows granoblastic microstructures and, in some cases, chessboard extinction (Fig. 5c). Garnet and plagioclase porphyroblasts commonly exhibit inclusion trails defining folded $S_1$ planes and strain shadows (muscovite + quartz + biotite + plagioclase ± garnet) of $S_2$ planes (Figs. 5b, 5d), suggesting a syn-$S_2$ crystallization during folding (i.e., $F_2$) of $S_1$. Some syn-$S_2$ garnet porphyroblasts are also revealed by the presence of an internal rotated foliation that resembles $S_2$ (Fig. 5e). In a similar way, a syn-$S_2$ growth of plagioclase porphyroblasts is indicated by trails of inclusions that are parallel to $S_2$ (Fig. 5f).
Chlorite and fine-grained white mica are retrograde minerals replacing garnet, biotite and plagioclase. Both minerals form fine-grained aggregates with shape-preferred orientation parallel to \( S_2 \), which is locally affected by mikrokinking or crenulation (Fig. 5g; Text S1 in the Supporting Information). When present, \( S_3 \) corresponds to a crenulation cleavage.

Finally, ultramylonites and phyllonites of \( S_3 \)' shear zones are made up of scarce porphyroclasts of plagioclase, muscovite and opaque minerals immersed in a fine-grained matrix of quartz + chlorite + muscovite + biotite + opaque minerals. In the matrix, quartz is mostly restricted to granoblastic monomineralic layers, indicating post-shearing recovery, whereas mica layers typically exhibit grain shape-preferred orientation and local concentration of opaque minerals in trails parallel to the mylonitic foliation (Fig. 5h), suggesting the role of dissolution-precipitation. Microstructures suggest that strain was mostly accommodated along mica-rich layers, with dissolution-precipitation as the main deformation mechanism, though subordinated bulging recrystallization was locally observed in quartz. However, evidence of cataclasis overprinting mylonitic features are common, with microfractures filled with fine-grained epidote + chlorite + white mica + quartz.

4.3. Mineral chemistry and thermobarometry

Garnet, white mica, biotite, plagioclase and chlorite compositions in the garnet-bearing metapelite of sample BA 22-17 were determined. Analytic methods and results are provided in Text S1 in the Supporting Information and Table S1 in the Supporting Information, respectively.

Compositional zonation of garnets of up to 7 mm diameter were determined using the GXMAP mode of automated SEM (Fig. 6a). The garnet mineral chemical evolution was then detailed by EPMA analyses along profiles. In the porphyroblasts, the spessartine (Mn) contents
decrease from ~9 to 1 mol % and the pyrope (Mg) contents increase from 3 to 11 mol % from core to rim, corresponding to a typical prograde zonation (Spear, 1993). The grossular (Ca) contents decrease from 28 to 10 mol %, whereas almandine increases from 60 to 77 mol %. In addition, a decrease of $X_{Ca}$ from 0.27 to 0.05 and an increase of $X_{Mg}$ from 0.03 to 0.14 are observed from cores to rims. Garnet outer rims display no significant increase of Mn and Ca. Unaltered biotite are rarely preserved in the micaschists and could be analyzed only when enclosed in plagioclase and quartz. According to their angular structural orientation in reference to the main foliation $S_2$, these biotites should be considered as $S_1$ biotites. They have a $X_{Mg}$ of ca. 0.46, with Ti of ca. 0.13 and $^{IV}Al$ of ca. 0.18 (p.f.u.). Biotites in the main foliation $S_2$ are mostly altered into chlorite, with totals below 92 wt. %. Their $X_{Mg}$ is ca. 0.52, with similar Ti and $^{IV}Al$ as the $S_1$ biotites. As biotite, the muscovite appears in the $S_1$ microstructural position, mostly oblique to the main foliation $S_2$. Their $Si^{4+}$ contents around 3.18 (p.f.u.) do not differ among different microstructural positions, similarly to Na (ca. 0.26-0.28 p.f.u.) and Mg (ca. 0.06-0.08 p.f.u.) contents. On the other hand, chlorite appears in various microstructural positions (Text S1 in the Supporting Information). Along the margins of the garnet, replacement of garnet by chlorite is observed, with $X_{Mg}$ of 0.52 at $^{IV}Al$ of 2.58 and $^{VI}Al$ of 2.75 (p.f.u.). Along $S_2$ planes, the chlorite replaces the biotite at variable degrees, which led to $X_{Mg}$ of 0.38-0.52, $^{IV}Al$ of 2.23-2.58 and $^{VI}Al$ of 2.73 (p.f.u.). Inclusions of plagioclase in the garnet porphyroblast are unzoned oligoclase and have anorthite contents between ca. 15 and 20 mol %. In the matrix along the foliation and next to the garnet margins, the small oligoclases are also unzoned but with lower anorthite contents between 11 and 15 mol %. Large plagioclase porphyroblasts in $S_2$ microlithons, which enclose crenulated $S_1$, show the same variation in anorthite contents. In general, a trend toward slightly lower Ca contents during plagioclase crystallization can be defined.

The garnet crystallized with a low-variance assemblage of biotite, muscovite,
plagioclase and quartz. These phases occur as aligned inclusions in the garnet porphyroblasts, and also along S\(_1\) and S\(_2\) foliations. Rutile and ilmenite also occur in garnet porphyroblasts and in the foliated matrix. In terms of the garnet \(X_{\text{Mg}}-X_{\text{Ca}}\) evolution (Spear, 1993), the garnet zonation semiquantitatively corresponds to a prograde metamorphism, with an initial increase of both pressure and temperature. The strong decrease of Ca toward the garnet rim signals subsequently decreasing pressure. Such a relative \(\Delta P/\Delta T\) trend appears to be mainly preserved when uncertainties about the corresponding plagioclase and biotite are considered (Spear, 1993).

Pressure and temperature for crystallization of garnet core were calculated by the calibrations of Holdaway (2000) and Wu (2015), involving enclosed plagioclase An\(_{17}\) and mica in the relic S\(_1\) foliation. This yielded ca. 400-450 °C at 7 kbar, with a slight dependency of the chosen biotite and plagioclase composition. A minimum error of \(\pm 50 \, ^\circ\text{C}\) and \(\pm 1\) kbar has to be considered for each equilibrium calculation. Calculations for the middle parts of the garnet zonation profile with increased Mg contents, considering the composition of S\(_1\) biotites, led to higher temperatures and pressures. Peak conditions were calculated from garnet inner rims with 10 mol % grossular combined with matrix An\(_{15}\) plagioclase, yielding ca. 650 °C at 11 kbar (Fig. 6b). A significant decrease of the grossular (Ca) content to 5 mol % towards the outer garnet rim resulted in a pressure drop to ca. 8 kbar for the final stage of the garnet crystallization during nearly isothermal conditions. The single P-T estimates from the zoned garnet define a clockwise P-T path within the kyanite stability field (Fig. 6b). Alternative calculations with other thermobarometers and the avPT routine (Powell and Holland, 1994) yielded comparable results. After the P-T record of the garnet-bearing assemblage, a retrograde evolution with further decrease of pressure and temperature can be expected, as recorded by temperature of ca. 335 °C calculated from chlorite thermometry. As the P-T path enters the monazite stability field for low-Ca bulk-rock compositions during decreasing pressure (Spear, 2010), monazites
might start to grow at or shortly after peak conditions.

4.4. Monazite dating and mineral chemistry

In the case of garnet micaschists of the Challhuaco hill, a population of high-Th+U monazites in samples BA 20-17 and BA 22-17 yield nearly identical Late Pennsylvanian ages of 299 ± 8 and 302 ± 16 Ma, respectively (Fig. 7). Likewise, sample BA 20-17 presents a low-Th+U monazite age population of 80 ± 20 Ma, whereas a low-Th+U monazite age population of sample BA 22-17 exhibits an age of 171 ± 9 Ma. The latter is comparable to the Middle Jurassic age recorded by sample BA 11-18, collected at the Jakob lake, which also yields a Cretaceous age of 110 ± 10 Ma. Further analytic results are provided in Table S2 in the Supporting Information. Carboniferous monazites are typically large and almost lack in sutured grain boundaries. In contrast, Jurassic and Cretaceous monazites are smaller and show severely sutured grain boundaries and sponge-like microstructures, being also associated with fine-grained chlorite and white mica (Fig. 7). The lack of Carboniferous monazites in sample BA 11-18 can be explained by the Ca bulk rock composition. The calculated values obtained from the automated SEM modal analysis of this sample indicate a considerably higher bulk rock Ca content (1.3 wt. %) than samples BA 20-17 and BA 22-17 (0.7 and 0.52 wt. %, respectively).

Monazite chemical composition support a discrimination of age populations (Fig. 8). Carboniferous monazites yield relatively homogeneous Y$_2$O$_3$ contents, with most values between ca. 0.8 and 1.2 wt. % (Fig. 8a). In spite of their similar ages, Jurassic monazites of samples BA 22-17 and BA 11-18 show differences, with Y$_2$O$_3$ contents that are lower or higher than a threshold value of ca. 1 wt. %, respectively. The Y$_2$O$_3$ content of Cretaceous monazites, in turn, varies between ca. 0.2 and 1.2 wt. %. On the other hand, Carboniferous monazites present comparable $X_{YPO_4}$ and $X_{GdPO_4}$ contents between ca. 0.018 and 0.024 (Fig. 8b). All
Jurassic monazites show $X_{GdPO_4}$ contents higher than 0.016, though sample BA 11-18 shows $X_{YPO_4}$ contents higher than those of sample BA 22-17. $X_{YPO_4}$ and $X_{GdPO_4}$ of the latter are relatively similar to those of Cretaceous monazites. Carboniferous and Jurassic monazites of sample BA 11-18 exhibit a positive correlation in the Th + U vs Ca plot, with Ca contents mostly higher than 0.1, as in the case of Cretaceous monazites (Fig. 8c). In contrast, Jurassic monazites from sample BA 22-17 yield Ca contents lower than 0.1. Both Carboniferous and Jurassic monazites of the Jakob lake show low Si contents (<0.15), whereas Cretaceous and Jurassic monazites of sample BA 22-17 exhibit values up to ca. 0.25 (Fig. 8d). The latter also show a positive correlation between Th + U and Si. Carboniferous monazites show the highest UO$_2$ contents (>0.62 %). Likewise, relatively high UO$_2$ contents are observed for Jurassic monazites of sample BA 11-18 (ca. 0.24-0.8 wt. %), thus showing a significant difference with low UO$_2$ contents of Cretaceous and Jurassic monazites of sample BA 22-17 (Fig. 8e). In sum, Carboniferous and Jurassic ages are the most abundant groups in the monazite age distribution pattern, though a subordinated Cretaceous population is also well-recorded (Fig. 8f).

5. Discussion and interpretation

5.1. P-T-D-t path of the Gondwanide basement

The ubiquitous S$_2$ foliation shows a dominant WNW- ESE to NNW-SSE strike (Figs. 2, 3), as reported by Dalla Salda et al. (1991) and García-Sansegundo et al. (2009). In a similar way, stretching lineations L$_2$ exhibit subhorizontal to gentle plunge to the NNW-SSE or WNW- ESE (Figs. 2, 4). Structural data point to bulk inclined transpression (Jones et al., 2004), with partitioning into strike-slip- and contraction-dominated domains suggested by differences in the orientation of S$_2$, L$_2$ and F$_2$ (Figs. 2, 4). Subvertical to steeply dipping S$_2$ and F$_2$ axial planes, steeply dipping F$_2$ hinges and gently plunging L$_2$ lineations (e.g., Jakob lake and Paso de las
Nubes) may record the former, whereas the latter may be registered by moderately to gently
dipping $S_2$ and $F_2$ axial planes, and gently dipping $F_2$ hinges subparallel to gently plunging $L_2$
lineations (e.g., Ñireco and Challhuaco hills). Alternatively, such differences may arise from
subsequent modifications due to Mesozoic and Cenozoic tectonic processes, which
significantly overprinted the area (e.g., Giacosa & Heredia, 2004; Castro et al., 2011a; Bechis
et al., 2014a). Local variations are observed in the structural profile between the Témanos and
Jakob lakes (Fig. 4), where significant post-Paleozoic deformation is absent. In the Paso de las
Nubes area (Fig. 2), on the other hand, the presence of Jurassic volcano-sedimentary rocks
overlying the basement suggests the lack of post-Jurassic tilting, thus indicating a relative
preservation of the orientation of basement fabrics (i.e., subvertical to steeply dipping $S_2$ and
subhorizontal to gently plunging $L_2$), though vertical axis rotations cannot be discarded.

The garnet-bearing mineral assemblage of the Challhuaco hill schist records a first
section of a prograde P-T path under conditions outside the monazite stability field, even for
low bulk rock Ca compositions (Fig. 6b). Crystallization of monazite can thus be expected at
peak metamorphic conditions (ca. 650 °C and 11 kbar) and mostly during subsequent
de compression. In consequence, the Carboniferous monazite population will mark the
minimum age for most of the prograde syn-$S_1$-$S_2$ amphibolite facies event. Since monazites lie
parallel to $S_2$, the timing of $S_2$-$L_2$ and associated $F_2$ fabrics is constrained by EPMA Th-U-Pb
monazite ages of 299 ± 8 and 302 ± 16 Ma (Fig. 6), roughly coeval or slightly younger than
peak metamorphic conditions of ca. 650 °C and 11 kbar. Since $S_1$ and $S_2$ share structural and
microstructural features (e.g., isoclinal folding), it can thus be inferred that $S_1$-$F_1$ may be
Carboniferous in age as well. According to microstructural evidence (Fig. 6a), $S_1$-$F_1$ may be
related to prograde metamorphism developed prior to peak metamorphic conditions, being the
evolution from $S_1$-$F_1$ to $S_2$-$F_2$ the result of progressive transpressional deformation (Carreras
& Druguet, 2018; Fossen et al., 2018).
On the other hand, late deformation fabrics (i.e., kink bands, $F_3$ open folds, shear zones) record low-grade metamorphic conditions. Kink bands and open folds are commonly associated with widespread crystallization of fine-grained chlorite and white mica (Section 4.2; Text S1 in the Supporting Information; García-Sansegundo et al., 2009), for which mean temperature conditions of ca. 335 °C were calculated from chlorite thermometry (Section 4.3).

In shear zones, mineral associations and microstructures suggest peak deformation conditions of ca. 350-300 °C, though evidence of late cataclasite and pseudotachylite development suggests further deformation below brittle-ductile transition conditions up to ca. 200-150 °C (Fagereng & Toy, 2011; Wallis et al., 2015; Wehrens et al., 2016). Despite being previously interpreted as the result of the final stages of the Gondwanide orogeny (García-Sansegundo et al., 2009), new evidence suggests a post-Paleozoic age for all these younger structures (see below). The pervasive crystallization of new monazite with low Th + U contents during a wide time span during the Mesozoic suggests that associated deformation fabrics under retrograde metamorphic conditions might take place at considerably lower pressures than Carboniferous events (Fig. 6b).

Comparison of P-T-D-t data obtained herein with previous P-T-t data from Martínez et al. (2012) suggests some differences among basement blocks in the study area in terms of P-T evolution and timing of metamorphism. Martínez et al. (2012) reported EPMA Th-U-Pb monazite ages of 391.7 ± 4.0 and 350.4 ± 5.8 Ma at the Brazo Tronador, for which metamorphic conditions of ca. 612 °C and 4.9 kbar were determined. In addition, conditions of ca. 440 °C and 18 kbar were determined for schists at the southern margin of the Gutiérrez lake (Martínez et al., 2012), but the timing of this high-pressure metamorphic event still remains elusive. Instead of a common P-T-D-t path (Dalla Salda et al., 1991; García-Sansegundo et al., 2009), basement blocks south of the Nahuel Huapi lake thus record different parts of a more complex long-term tectonometamorphic evolution, extending at least from ca. 390 to 300 Ma (see also
Section 5.2). Hence, these rocks underwent not only deformation and metamorphism related to the Gondwanide orogeny, which is restricted to the Carboniferous-Permian (e.g., Pankhurst et al., 2006; Heredia et al., 2017, 2018), but also to pre-Gondwanide Devonian tectonometamorphic processes.

5.2. Implications for the Late Paleozoic evolution of northern Patagonia

Basement rocks exposed to the south of the Nahuel Huapi lake have been traditionally ascribed to the Colohuincul Complex (Dalla Salda et al., 1991; Giacosa et al., 2001; García-Sansegundo et al., 2009; Martínez et al., 2012), which was defined by Turner (1965) to nucleate isolated outcrops of low-grade schists, quartzites and phyllites exposed in the northernmost North Patagonian Cordillera and the Southern Neuquén Precordillera at ca. 39° S. Both lithotypes and metamorphic grade in the *locus typicus* of the Colohuincul Complex contrast significantly with those of the study area (Sections 2, 4.3; García-Sansegundo et al., 2009; Martínez et al., 2012). In addition, metasedimentary rocks of the Colohuincul Complex were correlated with those of the Cushamen Formation (Cingolani et al., 2011, and references therein), which may have a late Carboniferous deposition age (Hervé et al., 2005; Marcos et al., 2018), being thus incompatible with Middle Devonian to Carboniferous ages of metamorphism reported for the study area (Section 4.4; Martínez et al., 2012). For this reason, the Bariloche Complex is proposed as a new stratigraphic unit to include basement rocks exposed to the south of the Nahuel Huapi lake, which record Devonian-Carboniferous metamorphism and Early Paleozoic maximum deposition ages (Hervé et al., 2018).

On the other hand, EPMA Th-U-Pb monazite ages of ca. 300 Ma (Section 4.4) constrain the timing of deformation and metamorphism linked to the development of ubiquitous NNW-SSE- to WNW-ESE-striking $S_2$ and associated $L_2$ and $F_2$ fabrics (Section 5.1). Comparable
NW-SE- to NNW-SSE-striking fabrics were also reported for medium-grade basement rocks of the western North Patagonian Massif, where foliated late Carboniferous intrusions occur as well (Cerredo & López de Luchi, 1998; Giacosa et al., 2004; Pankhurst et al., 2006; von Gosen, 2009; García-Sansegundo et al., 2009; Renda et al., 2017). Furthermore, metasedimentary rocks recording Upper Devonian-Carboniferous low-grade metamorphism of the Northern and Southern Neuquén Precordillera (Fig. 1) also exhibit NW-SE- to NNW-SSE-striking foliations (Franzese, 1993, 1995; Zappettini et al., 2012).

The timing of metamorphism and associated deformation of the basement in northwestern Patagonia is mostly constrained at ca. 390-300 Ma (Fig. 1), being thus comparable to the Devonian to Lower Permian ages of associated intrusions (Fig. 9a; Varela et al., 2005; Pankhurst et al., 2006; López de Luchi & Cerredo, 2008). EPMA monazite ages of 391.7 ± 4.0 and 350.4 ± 5.8 Ma were obtained in migmatitic paragneisses near the Brazo Tronador (Martínez et al., 2012), whereas a U-Pb titanite age of 380 ± 2 Ma was reported for a calc-silicate gneiss close to Piedra del Águila (Lucassen et al., 2004). Late Devonian metamorphism is evidenced by comparable ages of 360 ± 7 (EPMA Th-U-Pb monazite; Urraza et al., 2008) and ca. 359 Ma (U-Pb titanite data; Varela et al., 2005) reported further north in the North Patagonian Cordillera and Southern Neuquén Precordillera, respectively. In addition, Devonian to Carboniferous metamorphic overgrowths were reported in zircon grains from metasedimentary rocks at the Gutiérrez lake margin (Hervé et al., 2018) and El Maitén gneiss (Pankhurst et al., 2006), whereas whole-rock K-Ar ages of ca. 380-300 Ma roughly constrain the timing of greenschist facies metamorphism of the Piedra Santa Complex in the Southern Neuquén Precordillera (Franzese, 1995).

Geochemical data of Devonian to Permian magmatic rocks spatially associated with basement inliers show a progressive increase of Sr/Y from the Devonian to the Lower Permian (Varela et al., 2005, 2015; Pankhurst et al., 2006), which may result from increasing crustal
thickness (Fig. 9a; Chapman et al., 2015; Chiaradia, 2015; Profeta et al., 2015). This is consistent with the crustal thickening event suggested by Cerredo & López de Luchi (1998) for this Late Paleozoic magmatism. P-T-t data of the basement show a similar trend, with younger ages of metamorphism for higher pressure peak metamorphic conditions (Sections 4, 5.1; Martínez et al., 2012; Serra-Varela et al., 2017, 2018), although this trend could alternatively be explained by differential exhumation. Nevertheless, a progressive crustal thickening is further supported by the prograde P-T path of the Challhuaco sample (Fig. 6). Peak P-T conditions and the relatively low geothermal gradient recorded during prograde metamorphism (ca. 10-20 °C/km, Fig. 6) together with coeval intrusions at ca. 330-300 Ma located immediately south (Varela et al., 2005, 2015; Pankhurst et al., 2006) suggest a position close to the arc-forearc transition in a Carboniferous continental arc setting for the Challhuaco schists (Fig. 9b).

Ages of metamorphism recorded in accretionary complexes of Chile between ca. 390-300 Ma are coeval with ages of metamorphism recorded in the metamorphic basement of the North Patagonian Cordillera and the western North Patagonian Massif (Figs. 1, 9a), accounting for a common evolution of all these areas, further supported by similarities in the orientation of NNW-SSE- to WNW- to ESE-striking metamorphic foliations related to transpression (Fig. 2; Martin et al., 1999; Duhart et al., 2001). In the high P/T Western Series at ca. 41°S (Fig. 1), a 40Ar/39Ar hornblende plateau age of 361.0 ± 1.7 Ma in a garnet-bearing amphibolite of Los Pábilos boulders provides a minimum age for eclogite to epidote-amphibolite facies metamorphism at ca. 555 °C and >13 kbar (Kato et al., 2008), comparable to peak metamorphic conditions of ca. 600-760 °C and 11.0-16.5 kbar estimated by Willner et al. (2004) for these rocks. The age of subsequent retrograde blueschist facies metamorphism at ca. 350-500 °C and 10-14 kbar is constrained at 305.3 ± 3.2 and 296.6 ± 4.7 Ma by Rb-Sr mineral isochron data from garnet-bearing amphibolites and schists, respectively (Willner et al., 2004). A slightly
older $^{40}$Ar/$^{39}$Ar muscovite age of 325.0 ± 1.1 Ma was also reported for this event (Kato et al., 2008). On the other hand, K-Ar and $^{40}$Ar/$^{39}$Ar muscovite ages of ca. 390-370 Ma from borehole samples of schists, further supported by an $^{40}$Ar/$^{39}$Ar hornblende inverse isochron age of 359.3 ± 4.4 Ma of an undeformed granitic intrusion, indicate Devonian metamorphism for the Llanquihue Basement Complex (Fig. 1; McDonough apud Quezada, 2015; Hervé et al., 2016).

The tectonometamorphic evolution of the Gondwanide basement and associated intrusions seem thus to record progressive crustal thickening related to a Late Paleozoic continental magmatic arc (Varela et al., 2015; Hervé et al., 2016). However, Martínez et al. (2012) suggested a Devonian collision of the Chilenia terrane as the most likely trigger for deformation and metamorphism of rocks of the North Patagonian Cordillera. This interpretation is based on the peak metamorphic conditions of ca. 440 °C and 18 kbar calculated for a garnet-bearing micaschist exposed at the southern margin of the Gutiérrez lake and the EPMA Th-U-Pb monazite age of 391.7 ± 4.0 Ma recorded by a migmatitic paragneiss of the Brazo Tronador. Since the latter also yielded peak metamorphic conditions of ca. 612 °C and 4.9 kbar, the Devonian monazite age cannot be ascribed to high-P/low-T metamorphism but, instead, seem to be associated with low-P/high-T conditions. In contrast to the collisional model of Martínez et al. (2012), Hervé et al. (2016, 2018) proposed the Late Devonian-Early Carboniferous accretion of an island arc complex. Nevertheless, the proximity of the latter with respect to the continental arc (Hervé et al., 2018) makes the presence of two adjacent subduction zones unlikely. Alternatively, the development of a Carboniferous advancing accretionary orogen might be a likely trigger for the Gondwanide orogeny (Fig. 9b), since it might increase the interplate coupling, thus leading to crustal thickening and a relative stabilization of the margin (e.g., Cawood et al., 2009). Therefore, the Gondwanide orogeny might be essentially linked to the evolution of a transpressional advancing orogen, developed along the continental margin of southwestern Gondwana.
5.3. *The basement record of Mesozoic tectonomagmatic events*

Monazite has a relatively high closure temperature for Pb diffusion, being >800 °C for grain sizes larger than 10 μm and cooling rates of 10 °C/My (Cherniak et al., 2004). Th diffusion is approximately three times slower than Pb diffusion (Cherniak & Pyle, 2008) and, therefore, the Th-U-Pb system should remain closed in monazite under low-grade metamorphic conditions. Nevertheless, monazite commonly records low-grade metamorphic and hydrothermal events (e.g., Poitrasson et al., 1996; Teufel & Heinrich, 1997; Just et al., 2010; Oyhantçabal et al., 2012; Bergemann et al., 2017). Such disturbances of the Th-U-Pb system seem to arise from mechanisms such as neocrystallization, recrystallization and dissolution/reprecipitation, instead from diffusive exchange (Cherniak et al., 2004; Gardés et al., 2006; Cherniak & Pyle, 2008; Kelsey et al., 2008; Harlov et al., 2011; Oriolo et al., 2018).

Though Jurassic magmatism is widespread in the North Patagonian Cordillera (Castro et al., 2011a; Echaurren et al., 2017), evidence of coeval metamorphism and associated deformation is still scarce. In the study area, the igneous-metamorphic basement hosts the Jurassic Cordilleran batholith, which yields U-Pb SHRIMP zircon crystallization ages of ca. 176-160 Ma (Castro et al., 2011a). The only Jurassic record reported so far for basement rocks is a EPMA Th-U-Pb monazite age of 169.6 ± 6.7 Ma for a micaschist of the Catedral hill, which was attributed to contact metamorphism caused by Jurassic magmatism (Martínez et al., 2012), possibly related to a close tonalitic intrusion. Interestingly, monazite data (Section 4.4) record comparable Middle Jurassic ages of 171 ± 9 and 170 ± 7 Ma. Microstructural evidence shows that Jurassic monazites are spatially associated with retrograde micas, which show shape-preferred orientation parallel to S₂ and are affected by F₃ microfolds and microkink bands (Figs. 5f, 7). Therefore, monazite ages are interpreted to record the timing of Jurassic low-grade metamorphism and deformation constrained at ca. 335 °C by chlorite thermometry. Though
Jurassic granitoids are absent in the Jurassic sample locations, the influence of magmatism in these ages cannot be discarded.

The here obtained monazite ages are similar to the U-Pb SHRIMP zircon age of 175.9 ± 4.9 Ma reported for a granulite xenolith enclosed in Paleogene basalts to the south of the North Patagonian Massif (Castro et al., 2011b), whereas slightly older Early Jurassic intraplate deformation was documented by K-Ar and Ar/Ar mica data of Paleozoic granitoids in the northeastern North Patagonian Massif (Martínez Dopico et al., 2017). Recently, extension-related very low-grade metamorphism was recognized in the Southern and Northern Neuquén Precordillera (Suárez & González, 2018). Further evidence suggesting a Jurassic age for low-grade metamorphism and deformation is provided by late deformation fabrics, so far considered to be related to the Gondwanide orogeny (García-Sansegundo et al., 2009). Sample BA 11-18 was collected close to conjugated kink bands (Fig. 4), which are also microscopically observed (Fig. 5f, Text S1 in the Supporting Information). Considering the orientation and kinematics of both sets of kink bands (e.g., Cobbold et al., 1971; Carreras et al., 2013), a NNW-SSE shortening direction (present coordinates) can be inferred, further supported by magmatic fabrics of Jurassic plutons (Castro et al., 2011a). Comparable sets of conjugated brittle structures related to Jurassic deformation were reported to the southeast in the Deseado Massif (Reimer et al., 1996; Japas et al., 2013). Likewise, Jurassic volcanism, ore deposits and sedimentation recorded elsewhere in Patagonia have been mostly attributed to NE-SW to ENE-WSW extension and NW-SE to NNW-SSE shortening (Pankhurst et al., 2000; Rapela et al., 2005; Silvestro & Zubiri, 2008; Giacosa et al., 2010; Páez et al., 2011; Naipauer et al., 2012; Japas et al., 2017; Bechis et al., 2014b; Navarrete et al., 2016, 2018).

Considering the NNW to NW convergence direction along the proto-Andean paleo-Pacific margin determined for the ca. 180-160 Ma period (Müller et al., 2016) and all aforementioned structural evidence, a regional strike-slip-dominated transtensional regime can
thus be inferred for Patagonia in the Jurassic (Fig. 10a; e.g., Giacosa et al., 2010; Castro et al., 2011a; Japas et al., 2013). In such a setting, both kink bands and F3 open folds can be explained (Section 4.1, Fig. 10a; Fossen et al., 2013). Besides, the extensional component might favor the development of a thermal anomaly, leading to a relatively high geothermal gradient (Castro et al., 2011b; Suárez and González, 2018) that may explain the obtained monazite ages (Fig. 6b).

Cretaceous monazite ages of 110 ± 10 and 80 ± 20 Ma are associated with randomly oriented fine-grained aggregates of chlorite + white mica + opaque minerals ± epidote, which in some cases are located along microveins. In this context, monazite ages seem to record the timing of hydrothermal activity, possibly related to Cretaceous granitoids in the North Patagonian Cordillera suggested by K-Ar and Rb-Sr ages of ca. 120-80 Ma (González Díaz, 1982, and references therein).

On the other hand, dextral NE-SW- to NNE-SSW-striking subvertical S3’ shear zones are comparable in terms of orientation and kinematics to dextral NE-SW- to NNE-SSW-striking subvertical faults related to Andean Neogene transpression in the North Patagonian Cordillera (Diraison et al., 1998). However, deformation conditions estimated for shear zones (Sections 4.2, 5.1) indicate that they underwent deformation in a deeper crustal level than Neogene faults and, therefore, a more likely Cretaceous age is inferred for shear zones. This is further supported by middle Cretaceous deformation associated with ENE-NE shortening directions documented in northwestern Patagonia, where coeval tectonic activity along NNW-SSE- to NNE-SSW-striking thrusts was also reported (Orts et al., 2012; Echaurren et al., 2016, 2017; Gianni et al., 2018). Hence, shear zones might result from an early Andean Cretaceous dextral transpressional event (Fig. 10b), probably associated with coeval magmatism and hydrothermal events recorded by monazite ages, as the latter match the onset of Andean tectonics at ca. 100-95 Ma (Somoza and Zaffarana, 2008). Nevertheless, further geologic,
geochronologic and structural data are still necessary to assess the evolution of Cretaceous tectonomagmatic and hydrothermal processes.

6. Conclusions

Combined field structural data with *in situ* EPMA Th-U-Pb monazite, petrologic and microstructural data allow reconstructing the P-T-D-t path of the Gondwanide basement of the North Patagonian Cordillera. For samples from the Challhuaco hill, the timing of development of the metamorphic S2 foliation and associated L2 lineations and F2 folds is constrained by monazite ages of 299 ± 8 and 302 ± 16 Ma during or shortly after peak metamorphic conditions of ca. 650 °C and 11 kbar, whereas S1-F1 fabrics might be related to prograde metamorphism developed prior to these peak conditions. Within this framework, the evolution from S1-F1 to S2-F2 seem to be the result of progressive transpressional deformation.

The prograde P-T path recorded by samples of the Challhuaco hill together with regional evidence suggests Late Paleozoic crustal thickening. Contemporaneous metamorphism and deformation recorded by metamorphic complexes of northwestern Patagonia (Argentina) and accretionary complexes exposed further west (Chile) point to a common evolution of both areas, most likely resulting from Carboniferous transpression due to advancing subduction. Instead of the result of collision of large continental blocks, the Gondwanide orogeny might thus be essentially linked to geodynamic processes associated with subduction along the proto-Pacific margin of Gondwana.

On the other hand, monazite ages of 171 ± 9 and 170 ± 7 Ma indicate a Middle Jurassic low-grade metamorphic overprint coeval with development of kink bands and F3 open folds. This Middle Jurassic deformation event was contemporaneous with the emplacement of Cordilleran granitoids in the North Patagonian Cordillera and might result from a regional
transtensional event, ubiquitously recorded in Patagonia. Finally, monazite ages of 110 ± 10 and 80 ± 20 Ma are interpreted to result from hydrothermal processes, possibly related to Cretaceous magmatism. The timing of deformation along low-grade dextral NE-SW- to NNE-SSW-striking subvertical shear zones seems to be coeval, thus recording middle Cretaceous dextral transpression associated with the onset of Andean tectonics.

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Fig. 1. Sketch map showing main exposures of Paleozoic basement rocks in northwestern Patagonia (modified after Cingolani et al., 2011; Hervé et al., 2011, 2013; Serra-Varela et al., 2018). Inset shows area of the map in South America, whereas the yellow rectangle indicates location of the map of Figure 2. Main morphostructural units are indicated with dotted white lines (modified after Ramos et al., 2011; Bechis et al., 2014): Coastal Cordillera, Central Valley, North Patagonian Cordillera, North Patagonian Massif, Southern Neuquén Precordillera (SNP) and Northern Neuquén Precordillera (NNP). Location of Los Pábilos boulders (LP) and borehole geochronologic data of the Llanquihue Basement Complex (LBC) is shown with dotted lines (Hervé et al., 2016). Geochronologic constraints on the timing of metamorphism of the metamorphic basement of Argentina are also indicated (1: EPMA Th-U-Pb monazite, Urraza et al., 2008; 2: U-Pb titanite, Lucassen et al., 2004; 3: U-Pb titanite, Varela et al., 2005; 4: EPMA Th-U-Pb monazite, Martínez et al., 2012).
Fig. 2. Geologic map of the study area (modified after González Bonorino, 1973; Greco, 1975; García-Sansseungdo et al., 2009; Bechis et al., 2014). Monazite geochronologic data (italics) of Martínez et al., (2012) and sample locations (bold) are shown. Lower hemisphere equal area projection of poles of $S_2$ (black dots) and $L_2$ (blue dots) are also presented. Contour intervals at 2% per 1% area.
Fig. 3. (a) Bedding planes ($S_0$) defined by alternation of metapelites (dark grey) and metawackes (light grey). Lower hemisphere equal area projections of $S_0$ poles. Contour intervals at 2 % per 1 % area. (b) NNW-SSE-striking $S_2$ foliation in paragneisses. (c) Axial plane (AP) of isoclinal $F_2$ folds. (d) Axial plane traces of $F_2$ folds overprinting $F_1$ folds. (e) Dextral kink band cross-cutting $S_2$ planes. (f) Mylonitic $S_3'$ foliation in a dextral shear zone. (g) Example of $F_3$ open kink fold. Lower hemisphere equal area projections of poles of $F_3$ axial planes (AP) and axes (FA). Contour intervals at 2 % per 1 % area. Red line shows the orientation of the $S_2$ foliation. (h) Development of crenulation cleavage (yellow lines) within a dextral shear zone lozenge, which corresponds to a contractional overstep (*sensu* Ponce et al., 2013). Red line shows the orientation of the $S_2$ foliation, whereas green lines indicate mylonitic $S_3'$ foliation.
Fig. 4. Schematic structural profile at the western margin of the Jakob lake. Lower hemisphere equal area projections of poles of different structural elements are also shown. Contour intervals at 2% per 1% area. In the case of F2 folds and kink bands (KB), circles and squares represent axial planes and axes, respectively. Rootless S2 folds defined by quartz (Qtz) segregations are schematically illustrated.
Fig. 5. Photomicrographs of gneisses and schists of the study area. (a) Crenulation of $S_1$ defined by inclusions of opaque minerals in plagioclase (Pl) and muscovite crystals (cross-polarized light). $S_1$ is preserved within microlithons of the $S_2$ crenulation schistosity. (b) Garnet (Grt) porphyroblast with $S_2$ strain shadows (cross-polarized light). (c) Chessboard extinction in quartz (cross-polarized light). (d) Relics of $S_1$ crenulation within a garnet porphyroblast, which show chlorite replacement along rims (plain-polarized light). (e) Relics of rotated $S_2$ foliation within garnet porphyroblasts (cross-polarized light). (f) Inclusion trails of opaque minerals within plagioclase porphyroblast that define an internal foliation, which is subparallel to $S_2$ in the matrix (cross-polarized light). (g) $S_2$ foliation replaced by retrograde white mica and chlorite and crosscut by microkink bands (cross-polarized light). (h) Ultramylonite showing plagioclase porphyroclast and monomineralic layers of granoblastic quartz in a fine-grained matrix (cross-polarized light). In the matrix, quartz is mostly granoblastic, whereas mica layers exhibit grain shape-preferred orientation and local concentration of opaque minerals in trails (yellow arrows) parallel to the mylonitic foliation.
Fig. 6. (a) Map of energy dispersive X-ray (EDX) spectra (GXMAP) of sample BA 22-17 from the Challhuaco hill. Spectra from zoned garnet (Grt) with variable Fe, Mg, Mn and Ca contents in normalized element wt. % are labelled with different colors. EDX spectra from muscovite (Ms), albite (Ab), chlorite (Chl), quartz (Qtz) and apatite (Ap) are also indicated by distinct colors. Locations of analytical profile (1 to 35) and traces of garnet internal foliation $S_1$ and external foliation $S_2$ are marked as well. Positions of some monazites along $S_2$ are shown by circles. (b) P-T path of sample BA 22-17. Pressure estimations were obtained with the garnet-biotite-muscovite-plagioclase barometer (Wu, 2015), whereas temperature was calculated using garnet-biotite thermometer (Holdaway, 2000). Colors indicate garnet zonation (blue: core, yellow: middle, green: rim). Stability fields of monazite and allanite at different bulk rock contents as a function of Ca molar content (dotted lines) are shown together with the xenotime stability field (Janots et al., 2007; Spear, 2010). Stability fields of $\text{Al}_2\text{SiO}_5$ polymorphs are included as well. Mnz: monazite, Aln: allanite, Xtm: xenotime. Red arrows indicate the evolution of Carboniferous $S_1$-$F_1$ to $S_2$-$F_2$ fabrics related to the P-T path, whereas the brown arrow shows a schematic trajectory to explain Jurassic and Cretaceous monazite growth.
Fig. 7. Th-U-Pb chemical model ages of monazite (a, d, g). Total ThO$_2^*$ vs PbO (wt. %) isochrones diagrams. ThO$_2^*$ is ThO$_2$ + UO$_2$ equivalents expressed as ThO$_2$. Regression lines with the coefficient of determination $R^2$ are forced through zero (Suzuki et al., 1994; Montel et al., 1996). Weighted average ages (Ma) with MSWD and minimal 2σ error are calculated from single analyses according to Ludwig (2001). The symbols mark analyses belong to distinct monazite age populations (ca. 30-70 Ma, BA 20-17; ca. 100-200 Ma, BA 11-18; ca. 120-250 Ma, BA 22-18; ca. 280-330 Ma, BA 20-17 and BA 22-17) that define isochrones. White square symbols mark data of low-Th monazites, not considered for isochrone weighted mean ages. Backscattered electron images (BSE) of monazite (b, c, e, f, h, i), showing single Th-U-Pb ages (Ma) and weighted averages with 2σ error calculated from several analyses within a grain. (a) BA 20-17, garnet micaschist, Challhuaco hill. (b) Monazite (Mnz) with Carboniferous ages. (c) Sponge-like monazite after strong alteration, situated in mica-rich domain with muscovite (Ms) and chlorite (Chl); weighted average age is calculated from low-Th analyses. (d) BA 20-17, garnet micaschist, Challhuaco hill. (e) Small monazite grain with low Th-contents in the upper part. Weighted average is calculated from lower part with 1-6 wt. % Th. (f) Small Jurassic monazite with weak oscillatory zonation pattern. (g) BA 11-18, paragneiss, Jakob lake. (h) Strongly sutured monazite grain surrounded by chlorite. (i) Small and strongly sutured monazite grains with muscovite and chlorite. Only monazite with total >98.5 wt. % of oxides are considered for weighted average age.
Fig. 8. Mineral chemistry and distributions of monazite Th-U-Pb chemical ages (see also Fig. 7). (a) Age vs Y$_2$O$_3$, showing elevated Y$_2$O$_3$ contents in sample BA 11-18. (b) $X_{GdPO_4}$ vs $X_{YPO_4}$ (mole fractions). Maximum $X_{YPO_4}$ in garnet (Grt) mineral zone according to Pyle et al. (2001). (c) Th+U vs Ca (mole fractions). (d) Th+U vs Si (mole fractions). (e) Age vs UO$_2$. (f) Histogram of monazite ages in all samples (BA 20-17, BA 22-17 and BA 11-18). Diverse distribution of Carboniferous, Jurassic and minor Cretaceous monazite populations.
Fig. 9. (a) Compilation of Sr/Y and U-Pb zircon geochronologic data (Varela et al., 2005, 2015; Pankhurst et al., 2006) of intrusions of Argentina (blue), recording increasing crustal thickness of the Late Paleozoic continental arc. Equivalent crustal thickness values were estimated using the equation of Chapman et al. (2015). Data from a granitic clast (green) of a conglomerate in Chile, which is inferred to derive from the Devonian continental arc in Argentina (Quezada, 2015; Hervé et al., 2016), are included as well. The timing of metamorphism of metamorphic complexes of Argentina (red; Lucassen et al., 2004; Varela et al., 2005; Urraza et al., 2008; Martínez et al., 2012; this work) and accretionary prism complexes of Chile (orange; Willner et al., 2004; Kato et al., 2008; McDonough apud Quezada, 2015) is also indicated. (b) Schematic Carboniferous tectonic setting for the region south of ca. 40 °S (modified after Kato et al., 2008; Martínez et al., 2012; Hervé et al., 2013, 2016). Continental arc magmatism is recorded by Carboniferous intrusions distributed between the North Patagonian Cordillera and the western North Patagonian Massif (Varela et al., 2005, 2015; Pankhurst et al., 2006). In the accretionary prism (Chile), deposition of the Western Series sedimentary protoliths takes place (Hervé et al., 2013), whereas the Los Pábilos boulders undergo exhumation and retrograde blueschist facies metamorphism (Willner et al., 2004; Kato et al., 2008). The red star shows the location of the Challhuaco hill samples. See Section 5.2 for further details.
Fig. 10. Schematic Jurassic (a) and Cretaceous (b) structural evolution of the study area. Maximum shortening (blue arrows) and extension (red arrows) directions together with strike-slip component (yellow arrows) are indicated. See Section 5.3 for further details. (a) Emplacement of Jurassic magmatism and development of conjugated kink bands and F3 open folds in a strike-slip-dominated transtensional setting (modified after Castro et al., 2011a). (b) Thrusting and dextral shearing related to dextral transpression.