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THE CONLARA METAMORPHIC COMPLEX: LITHOLOGY, PROVENANCE, METAMORPHIC CONSTRAINTS ON THE METABASIC ROCKS, AND CHIME MONAZITE DATING

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ABSTRACT

The Conlara Metamorphic Complex, the easternmost complex of the Sierra de San Luis, is a key unit to understand the relationship between the late Proterozoic-Early Cambrian Pampean and the Upper Cambrian-Middle Ordovician Famatinian orogenies of the Eastern Sierras Pampeanas. The Conlara Metamorphic Complex extends to the east to the foothills of the Sierra de Comechingones and to the west up the Río Guzmán shear zone. The main rock types of the CMC are metaclastic and metaigneous rocks that are intruded by Ordovician and Devonian granitoids. The metaclastic units comprise fine to medium-grained metagreywackes and scarce metapelites with lesser amounts of tourmaline schists and tourmalinites whereas the metaigneous rocks encompass basic and granitoids rocks. The former occur as rare amphibolite interlayered within the metasedimentary rocks. The granitic component corresponds to a series of orthogneisses and migmatites (stromatite and diatexite). The CMC is divided in four groups based on the dominant lithological associations: San Martin and La Cocha correspond mainly to schists and some gneisses and Santa Rosa and San Felipe encompass mainly paragneisses, migmatites and orthogneisses. The Conlara Metamoprphic Complex underwent a polyphase metamorphic evolution. The penetrative D_2-S_2 foliation was affected by upright, generally isoclinal, N-NE trending D_3 folds that control the NNE outcrop patterns of the different groups. An earlier, relic S_1 is preserved in microlithons. Discontinuous high-T shear zones within the schists and migmatites are related with D_4 whereas some fine-grained discontinuous shear bands attest for a D_5 deformation phase. Geochemistry of both non-migmatitic metaclastic units and amphibolites suggest that the Conlara Metamorphic Complex represents an arc related basin. Maximun depositional ages indicate a pre- 570 Ma deposition of the sediments. An ample interval between sedimentation and granite emplacement in the already metamorphic complex is indicated by the 497 ±8 Ma age of El Peñon granite. D_1-D_2 history took place at 564 ± 21 Ma as indicated by one PbSL age calculated for the M_2 garnet of La Cocha Group. D_3 is constrained by the pervasively solid-state deformed Early Ordovician granitoids which exhibits folded xenoliths of the D_1-D_2 deformed metaclastic rocks. Pressure-temperature pseudosections were calculated for one amphibolite using the geologically realistic system MnNCKFMASHTO (MnO-Na_2O-CaO-K_2O-FeO-MgO-Al_2O_3-SiO_2-H_2O-TiO_2-Fe_2O_3). Peak metamophic conditions (M_2) indicate 6 kbar and 620 °C. Late chlorite on the rims and in cracks of garnet, along with titanite rims on ilmenite and matrix plagioclase breaking down to albite suggests that the P-T path moved back down. Monazite analyses yield isochron Th–U–Pb ages ranging from 446 to 418 Ma. The oldest age of 446 ± 5 Ma correspond to a migmatite from the Santa Rosa Group. Monazites in samples from the La Cocha and the San Martin group crystallised at decreasing temperatures, followed by the 418 ± 10 Ma low-Y_2O_3 monazites in one sample of the la Cocha Group that was also obtained from a migmatite, and would likely mark a later stage of a retrograde metamorphism New CHIME monazite ages presented here likely represent post-peak fluid assisted recrystallisation that are similar to amphibole and muscovite cooling ages. Therefore the monazite ages may represent a re-equilibration of the monazite on the cooling path of the basement complex.

Keywords: lithology-geochemistry metaclastic rocks-geochemistry and metamorphism amphibolite-CHIME monazite dating-Conlara Metamorphic Complex

1. INTRODUCTION

The Conlara Metamorphic Complex, the easternmost complex of the Sierra de San Luis, is a key unit to understand the relationship between the late Proterozoic-Early Cambrian Pampean and the Upper Cambrian-Middle Ordovician Famatinian orogenies of the Eastern Sierras Pampeanas (Fig.1a,b). The protholits of the Neoproterozoic -early Ordovician metamorphic complexes that make up the Sierras were deposited along the southwestern margin of Gondwana. Provenance of the metaclastic units relates the complex to arc basins (López de Luchi et al., 2003) with maximum depositional ages older than ca. 580 Ma (Steenken et al., 2006) exhibiting no Pampean inheritance. U-Pb garnet age constrains the peak metamorphism to 564 ± 21Ma (Siegesmund et al., 2010) and separates its evolution from the rest of the complexes of the Sierra de San Luis that underwent only the Famatinian orogeny. In addition among the metaclastic units the Conlara Metamorphic Complex encompasses banded schists (i.e., La Cocha Group) (Fig.1b), a particular unit that appears in the northeastern corner of the Sierra de Córdoba (i.e. Tuclame Formation), in the Sierra de Ancasti (i.e. Ancasti Formation) and also in the Puncoviscana Formation (Steenken et al., 2006, Drobe et al., 2009) located north of the Sierras Pampeanas.

The Sierra de San Luis which comprises a Cambro-Ordovician metamorphic basement intruded by Ordovician and Devonian plutons (Fig 1a, b), and consists of three major NNE trending metamorphic complexes: the Nogolí (NMC), Pringles (PMC), and Conlara (CMC) complexes (Sims et al.,, 1997, 1998). The complexes are separated by the two narrow phyllite belts of the San Luis Formation (SLF) (Prozzi and Ramos 1988). The basement complexes of the Eastern Sierras Pampeanas were structured by three orogenic events: the late Proterozoic-Early Cambrian Pampean, the Upper Cambrian-Middle Ordovician Famatinian and the Middle -Upper Devonian Achalian. These events were interpreted to be the result of subduction and continental collision of several terranes along the margin of Gondwana. The opening of Carboniferous continental basins marked the end for the ductile deformation and the compressive tectonic movements (Ramos et al., 1986, Ramos 1988, Kraemer et al., 1995; Sims et al., 1997,1998; Rapela et al., 1998, 2016, Steenken et al., 2004, 2006, 2008; Otamendi et al., 2020, and references therein).

Although investigations on the magmatic, metamorphic and structural evolution, provenance, geodynamic setting of the protoliths of the metamorphic basement and metalogenesis of the CMC were performed (López de Luchi 1986, Ortiz Suárez 1988, 1996, Ortiz Suárez et al., 1992; Llambias et al., 1996, 1998; Sims et al., 1997, 1998; von Gosen 1998; López de Luchi et al., 2003; 2007, 2008, 2009, 2018; Steenken et al., 2004, 2006, 2008, and references therein), few studies have focused on a comprehensive study of the Conlara Metamorphic Complex.

The aims of this paper are (1) to present a comprehensive lithological characterization of the Conlara Metamorphic Complex, (2) provide P-T constraints for the some units by analysing the relationship between metamorphism and deformation, (3) to present CHIME monazite ages for different units of the CMC and discuss time constraints for its evolution, (4) to decipher the significance of the CMC in relation to the rest of the Eastern Sierras Pampeanas

2. GEOLOGICAL SETTING

The main metamorphic complexes of the Sierra de San Luis (Fig. 1b) are mainly built up by greenschist to granulite facies metaclastic and metaigneous rocks. The NMC comprises amphibolite facies gneisses, felsic orthogneisses and minor mafic gneisses and is intruded by monzonites and granites (Sims et al., 1997, 1998, González et al., 2004, Steenken, et al., 2006). A SHRIMP ^{208}Pb/^{232}Th monazite age of 478± 4 Ma was reported for a migmatite from the northern part of the complex (Steenken et al., 2006). The PMC is made up by greenschist facies to amphibolite facies gneisses, mica-schists, amphibolite intruded by mafic and felsic plutons. Locally, granulite facies which develops in the vicinity of the numerous Late Cambrian back-arc mafic bodies (Sims et al., 1997; Hauzenberger et al., 2001, Steenken et al., 2006) is associated with a near isobaric P-T path and was dated as Late Cambrian (Steenken et al., 2008) based on a concordant U/Pb zircon ages of 498±10 Ma. The San Luis Formation (Prozzi and Ramo,s 1988) consists of meta-quartz-arenites and phyllites, minor black shales, scarce conglomerates, and acidic volcanic rocks (Hack et al., 1991, von Gosen 1998). Available radiometric ages for acidic metavolcanic layers yielded a U-Pb zircon age of 529 ± 12 Ma while detrital zircon grains indicated a maximum deposition age of ca. 515 Ma (Perón Orillo et al., 2019 and references therein).

The Conlara Metamorphic Complex (CMC), the easternmost metamorphic complex of the Sierra de San Luis is limited to the west by the Río Guzmán shear zone whereas its eastern margin to the Sierra de Comechingones (Sims et al., 1997) is controlled by the Las Lajas and Guacha Corral shear zones (Siegesmund et al., 2010) (Fig. 1b). The main rock types of the CMC are metaclastic and metaigneous rocks that are intruded by Ordovician and Devonian granitoids. These rocks were studied since decades from the pioneer studies of Pastore and González (1954), up to more recent studies by Drobe et al.(2009, 2011), López de Luchi (1986), López de Luchi and Cerredo (2001), López de Luchi et al. (2003, 2008, 2009, 2018), Morosini et al. (2019), Ortiz Suárez (1988,1996), Ortiz Suárez et al. (2009), Siegesmund et al. (2010) and Steenken et al. (2004, 2006, 2008) and references therein.

The Conlara Complex was originally defined by Sims et al. (1997) to encompass the sector of the Sierra de San Luis south of 32º 40´ and east of the Guzmán shear zone, including the metamorphic basement of the El Morro, Tilisarao, and La Estanzuela hills, as well as the southern tip of the Sierra de Comechingones. These authors consider that the complex is mainly made up of medium to high grade metapelitic and metapsammitic gneisses and schists, minor calc-silicates, amphibolites and marbles and well-developed pegmatites. Later, the Conlara Complex was renamed as Conlara Metamorphic Complex (López de Luchi and Cerredo, 2001, Steenken et al., 2004, 2006) and extended to include the metamorphic rocks north of the Renca batholith, thus encompassing banded schists and fine grained schists.

The main rock types of the CMC are schists and banded schists, biotite gneisses, migmatites and orthogneisses. Tourmalinites, tourmaline-schists, calc-silicatic rocks related with W-mineralization and amphibolites are also locally recognized. The mineral assemblages of the metaclastic rocks are characterized by the paucity of diagnostic paragenesis, in fact biotite and rare relic garnet are the only AFM phase present (López de Luchi et al., 2008). López de Luchi and Cerredo (2001) proposed a polyphase deformational evolution of the metamorphic units that encompasses at least three ductile deformation phases which produced foliations, banding, folding and localized high-temperature shear zones. The penetrative, dominant foliation, $S_{2,}$ locally associated with banding in some metaclastic schists is related to the second deformation (D_2). A relic S_1 is preserved in microlithons as biotite polygonal arcs or as thin folded leucocratic veins with S_2 axial plane cleavage. The penetrative S_2 foliation is in turn affected by upright, generally isoclinal, D_3 folds with N-NE/S-SE axial planes and an associated S_3. Thin (≤ 1-2mm) biotite-quartz mylonitic shear zones related to D_3 are occasionally found. Late Cambrian and Early Ordovician large discrete granitoid (e.g., El Peñón Granite) and gabbroid (e.g., Las Cañas, Morosini et al., 2019) plutons and sheet-like stocks (e.g., Río de la Carpa Granite; Llambías et al., 1996) intruded the CMC during $D_{3,}$ after the peak metamorphism. Therefore, they are not considered as part of the prograde path of the Complex.

Geochemical studies of the metaclastic rocks of the CMC showed that the protoliths were mainly greywackes, semipelites and subordinated pelites with compositional features typical of Phanaerozoic active margin settings. Immobile trace elements such as La, Th, and Sc contents and ratios suggest continental island arc settings (López de Luchi et al., 2003). Steenken et al. (2004, 2006) and Drobe et al. (2009, 2011) found isotopic and geochemical similarities between the CMC and the Puncoviscana Formation, the Early Cambrian key unit of the Eastern Sierras Pampeanas to the north of Sierras de Córdoba.

The maximum depositional age of the protolith of the Conlara Metamorphic Complex is at ~586 Ma (Steenken et al., 2006). However, the youngest detrital zircon of the dataset yielded 520 ± 6 Ma. More recently, Rapela et al. (2018) found the same Ediacaran detrital cluster and a very small population of Late Cambrian zircon ages. The emplacement of the El Peñón Granite (497 ±8 Ma; U-Pb zircon age), that hosts enclaves from the D_2-banded schists of the CMC, constrain the peak metamorphic conditions to the Pampean Orogeny (Steenken et al., 2006, 2008).

3. LITHOLOGY OF THE CONLARA METAMORPHIC COMPLEX

The metamorphic complexes of CMC comprise fine to medium-grained metagreywackes and scarce metapelites that make up schists, gneisses and migmatites with lesser amounts of tourmaline schists and tourmalinites and metaigneous rocks that encompass basic and granitoids rocks. The former occur as rare amphibolite interlayered (likely emplaced as sills within the sedimentary pile) within the metasedimentary levels. The granitic component corresponds to a series of orthogneisses and migmatites which underwent all the prograde metamorphic evolution of the CMC being emplaced between D_1-D_2 (López de Luchi 1986, López de Luchi et al., 2003, 2008). In several areas gneisses transitionally grade into migmatites due to an increase in leucosomes. Migmatites differ on the mineral paragenesis and the different types of relationship between leucosome/mesosome/paleosome from heterogeneous stromatitic type, with leucosome rarely exceeding 30-40% compared to diatexite. In this study we consider as migmatites rocks in which evidence of a melt phase is recognized independently of the identification of melt-producing reactions In satellite images gneiss and migmatites are recognized as uniform medium gray areas in which granitoid as well as pegmatites might be observed. Pastore and González (1954) included these rocks within the gneissic micacites and mixed rocks while other authors named them migmatites (Gordillo and Lencinas, 1979).

The CMC is divided in four groups (Fig. 1a) based on the dominant lithological associations: San Martín (Fig 2a, 3a) and La Cocha (Fig. 2b,3b) correspond mainly to metaclastic lithologies and Santa Rosa (Fig. 2c, 3c) and San Felipe (Fig. 2d, 3d) encompass mainly paragneisses, migmatites and orthogneisses. The La Cocha Group is partially equivalent to Las Aguadas Metamorphic Complex of Ortiz Suárez (1988, 1996). The overall arrangement of these four groups could be likely controlled by a D_3 related major structure (fold?) governed by N-NE axis (López de Luchi et al., 2008)

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3.1.THE SAN MARTÍN GROUP

The San Martín Group (SMG) is composed of fine to medium grained schists (Fig. 2a, 3a) that are predominantly metapelites with subordinate metapsammites and scarce tourmalinites. The SMG appears as two NNE belts at both sides of central belt of the migmatites of the Santa Rosa Group (Fig. 1a) and hosts pegmatites, Ordovician and Devonian plutons (Las Chacras, El Telarillo and El Hornito plutons, Fig. 1a). It is noteworthy the profuse development of thin pegmatite sills as well as fine to coarse grained garnet±tourmaline leucogranites and locally biotite granitoids like the Rio de la Carpa Granite Contacts with the La Cocha group are covered whereas an increase of psammitic layer leads to gneisses and migmatites along the border of the western belt. The eastern belt (Fig. 1a) is mainly made up of biotite+quartz +muscovite +plagioclase(oligoclase) fine grained schist (Fig. a, b) moderately to highly interlayered with almost pure quartz and tonalite to granite layers of variable thickness. In the western belt (Fig.

1a) north of the Las Chacras-Potrerillos Batolith, the dominant mineral paragenesis is quartz + biotite+muscovite+plagioclase (oligoclase-albite) ±tourmaline±apatite and subordinately biotite +quartz+muscovite ± tourmaline and biotite +quartz+plagioclase ± garnet. Martínez et al. (2015) describe scarce thin amphibolites and calco-silicatic layers in this belt. Immediately south of Quines thin tourmalinites layer are found (Fig. 4c). Some fine grained schists (Fig. 4d) appears interlayered with the banded schist of the La Cocha Group or in scarce occurrences as discontinuous rafts in the migmatites of the Santa Rosa Group.

These fine grained schists (Fig. 4a, b) are characterized by a continuous S_2 foliation defined by biotite and muscovite flakes set in a granoblastic matrix of fine-grained (up to 0.1 mm) granoblastic quartz with variable content of sodium- rich oligoclase. In other cases, the continuous S_2 foliation is made up only of parallel biotite flakes. Muscovite also forms blocky, randomly-oriented porphyroblasts up to 2 mm across. In some localities fibrolite/sillimanite appears inside the muscovite blasts as aggregates that are crenulated and probably correspond to a preexisting parallel foliation that was overprinted during an event in which sillimanite was retrogressed. The local preservation of a centimeter-scale primary sedimentary compositional layering defined by varying biotite/plagioclase ratio leads to a metamorphic banding of alternating metapelites and metapsammites with different biotite content. The S_2/S_3 foliation (López de Luchi, 1986, López de Luchi and Cerredo, 2001) is approximately parallel to this banding but along strike the metapsammitic layers are boudinaged. Locally this foliation is deformed into weak crenulations (Fig. 4b) at a high angle to this banding, and the biotite shows kink-bands.

In some cases the fine-grained schists show an anastomosing foliation defined by oriented biotite, muscovite and chlorite along with quartz-plagioclase (An_{18}) polygonal aggregates.

The psammitic (0.3-0.4 mm) fine-grained schist layers also show a continuous variable defined foliation -oriented green biotite and scarce muscovite- along with quartz and plagioclase (An_{26-33}). Locally oriented brown biotite (occasionally very large up to 5mm) and polygonal quartz-plagioclase aggregates define the foliation. In most of the samples of the psammitic fine-grained schist conspicuous idioblastic up to 0.18mm apatite crystals are present.

Martínez et al. (2015) describe a transition from west to east from the fine grained biotite schist to gneiss north of Villa Praga. The transition is accomplished by an increment of “metapsammitic” rocks with variable content of plagioclase and high contents of biotite and quartz. In some examples microcline accompanies plagioclase modal increase whereas apatite is widespread. North of San Martín on the road to Quines an alternance of metapelitic and metapsammitic layers is observed.

A distinctive feature of the San Martín Group is the profuse development of granitoid leucosomes either fine grained to pegmatitic granitic sills or fine to medium grained tonalitic to granodiorite (Fig. 3a). Granites are made up by quartz,plagioclase,-microcline, muscovite with garnet, apatite and tourmaline as accessory minerals. The tonalitic-granodiorite layers are made up of quartz-plagioclase-biotite-muscovite. Some thin, discontinuous, partly discordant leucocratic veins composed of recrystallized equant quartz aggregate are also present.

The dominant structure is a NNE-trending east dipping foliation (S_2) that is locally folded by decametric to metric-scale open to tight (F_2) folds (Fig. 2a, 3a) with development of a discontinuous S_3 axial plane foliation (Fig. 4d). Two folding phases are depicted. S_2 results from the isoclinal (F_1) folding of a S_1 fabric that is recognized by the development of boudins, intrafolial folds or unrooted hooks of metapsammites and leucocratic veins. F_2 folds affects Ms-Grt- tourmaline granitoids that were emplaced with clear cut contacts parallel to S_2 (Fig. 3a). Aggregates of biotite-muscovite-chlorite wrapped by the S_3 foliation could replace former aluminous porphyroblast that would have grown predating S_3 development. Large muscovite plates (Fig. 4a) overprint the S_2 foliation and turned into mica fishes due to F_3 folds. Crenulated sillimanite aggregates inside the muscovite blasts also suggest that F_2 folding as well as S_3 development was associated to retrogression of the peak metamorphic assemblages. Along the limbs of the F_2 folds granitoids are boudinaged and stretched (Fig. 3a). These sills are folded and thicker in the eastern flank of the folds which is consistent with the western vergence of the axial plane of the last folding phase. Some thin, discontinuous, partly discordant leucocratic and discrete tonalitic leucocratic layers draw open folds whose axial planes are parallel to the S_3 foliation are recognized.

A late shearing is responsible for the development of plagioclase sigma clasts, muscovite fishes and kinks in biotite and muscovite. Associated with this shearing, very fine-grained (0.04 mm) quartz and very tiny, oriented biotite-chlorite microbelts are developed. The geometry and grain-size of micro-shear could suggest extensional crenulation cleavage.

3.1.2. LA COCHA GROUP

The La Cocha Group (LCG) forms a roughly submeridian belt within the north central area of the CMC (Fig. 1a). The dominant rock type are banded schists (Fig. 2b, 3b) locally interlayered with fine-grained schists, occasionally accompanied by thin layers of tourmalinites, tourmaline schists and amphibolites The LCG hosts both Ordovician and Devonian granitoids. The former are represented by El Peñón Granite, El Salado (Ortíz Suarez, 1988,1996), Los Alanices and La Tapera pluton (López de Luchi, 1986, López de Luchi and Cerredo, 2001) among others and a well developed suite of huge pegmatites, whereas the Renca batholith is the intruding Devonian unit (Fig. 1a).

3.1.2.1 **Banded schists**

The most impressive outcrop feature of these biotite-schists is represented by a metamorphic banding composed of 0.5-2 cm alternating whitish quartz-plagioclase-rich and gray biotite-rich layers which is affected by upright, generally isoclinal, D_3 folds associated with a steeply-SE dipping NE-striking S_3 foliation.

S_2 foliation is represented by the metamorphic banding/layering which is parallel to S_3 along F_3 limbs and normal to S_3 in F_3 hinge zones (Fig. 5a). The present foliation S_3 results from the folding of the S_2 foliation by D_3 folds. S_2 foliation that is represented by the metamorphic banding/layering (Fig 5a, b) is parallel to S_3 along F_3 limbs and normal to S_3 in F_3 hinge zones (Fig. 5b).. An earlier, relic S_1 (Fig. 5a) is preserved in some leucocratic domains as biotite microfolds. There is a progressive increase in grain-size from S_1 biotite, to polygonal plagioclase-quartz aggregates and biotite defining the S_2 foliation and banding.

Mica rich domains are made up by green biotite, plagioclase, scarce quartz plus opaque minerals, apatite and zircon. In some sector a pre- S_3 garnet (Fig. 5c) locally surrounded by plagioclase is observed. The garnet contains symmetrical S-shaped curved trails of inclusions (epidote, quartz, ilmenite), suggesting syntectonic growth. Garnet is partially retrogressed along grain boundaries to chlorite and to sericite-chlorite mainly along cracks.

Quartz rich domains are made up of plagioclase ($An_{26/30}$), quartz and some biotite. Relic garnet is locally found in these domains. Plagioclase is xenomorphic. Relic F_2 microfolds are defined by the relic biotite 1 in the quartz domains (Fig. 5a). Quartz blasts are flattened at axial ratios of ~1:2 with grain boundaries denoting migration process. Chlorite appears either in small flakes associated with biotite or in large clots that would result from the retrogression of garnet parallel to the S_2 fabric.

Muscovite blasts (Fig. 5d) in plates larger than the average grain size are observed in most of the outcrops. These blasts define muscovite fishes or are undeformed according to their location in the F_3 folds. In some sectors rutile, up to 0.7 mm is observed. Apatite with pleochroic cores and rutile are common accessory minerals.

3.1.2.2. **Fine grained schists**

Different textural types of fine-grained schist (Fig. 4d) are recognized when interlayered with the banded schists. One type of fine-grained schist has a spaced foliation with alternating phyllosilicate-rich and feldspar-rich layers. The former is composed of oriented greenish brown biotite and muscovite and polygonal quartz-plagioclase. S_2 is defined by oriented biotite and muscovite which alternate with quartz-plagioclase (An_{16-20}) rich layers. The quartz rich domains are composed of coarser grained oligoclase (An_{25}) and quartz aggregates. Minor tourmaline and idioblastic apatite with pleochroic cores (blue to red) are accessory minerals in variable amount. Some biotite$_1$ polygonal arcs remain enclosed among S_2 muscovite or biotite. Robust, slightly folded or undeformed post-foliation ramdomly distributed muscovite is also observed.

Another type of fine grained schist is characterized by a continuous foliation defined by biotite and muscovite flakes set in a granoblastic matrix of quartz and scarce plagioclase. Oversized biotite is observed in some examples whereas muscovite blasts are abundant. In some localities aggregates of quartz, sericite and chlorite that may replace former andalusite porphyroblasts are wrapped by the S_3 fabric. Maisterrena (1984) mentioned similar aggregates associated with staurolite porphyroblasts.

3.1.3.TOURMALINITES AND TOURMALINE SCHISTS IN THE LA COCHA AND SAN MARTÍN GROUPS

The tourmalinites represent a conspicuous albeit scarce lithology within the La Cocha and San Martín groups. The tourmalinites are fine grained rocks which occur as 0.5-2 m thick layers within the sequence of fine grained muscovite-biotite schists associated with the banded schists. Unfortunately, the poor outcrop conditions, make it difficult to estimate precisely the along strike length. No clear relation was observed so far with scheelite-bearing calc-silicate rocks in the studied outcrops. Tourmaline-rich rocks also occur in contact with lenticular granitic and pegmatite bodies.

The tourmalinites exhibit a well developed planar fabric at outcrop scale. They are fine grained, (usually <1 mm), gray to black with a silky luster and variable amounts of tourmaline (up to >90% by volume). The tourmaline crystals are generally aligned. Textural variations are common. Some of the tourmalinites show a very thin (< 1mm) layering in which tourmaline-rich layers alternate with quartz-muscovite granoblastic layers. Quartz and variable amounts of plagioclase, muscovite and biotite are the most abundant minerals associated with zoned tourmaline in the tourmalinites. Some layers contain appreciable amounts of apatite (up to 5% by volume) and rutile (up to 2% by volume). Small grains of zircon and monazite are accessory minerals. Muscovite and some opaque mineral sheets are oriented in the foliation plane. The layering is coplanar to the main S_2/S_3 foliation. Some tourmalines show a distinctive optical zoning with subhedral to euhedral, reddish-orange brown (20- 30 μm) cores (probably detrital tourmaline) and bluish green rims. Apatite forms aggregates of subhedral grains or is isolated. In some thin sections lenses or veins of coarse grained deformed quartz are observed.

Tourmaline rich biotite-muscovite schists are characterized by variable concentrations of tourmaline, normally less than 10% by volume. Coexisting major minerals are mica (muscovite±biotite), quartz and plagioclase. Accessory minerals include titanite, rutile, and zircon. Tourmaline occurs normally as: fine to very fine-grained (<200 μm in length), subhedral to anhedral crystals disseminated within a foliated quartz-micaceous matrix.

3.1.4 THE SANTA ROSA GROUP

The Santa Rosa Group (SRG) displays a large N-S extension, in two belts one located along the eastern border of the Sierra and another central belt of discontinuous outcrops that extends all along the CMC, partially enveloping the SMG and LCG (Fig. 1a).The SRG encompasses medium grained biotite gneisses with leucocratic quartz and granite veins and stromatitic migmatites (Fig. 2c, 3c). The stromatitic migmatites locally show centimetric up to 5 m long amphibolite lenses with sharp contacts and a penetrative foliation parallel to that of the migmatite. These mafic rocks are usually injected by leucocratic veins that are folded with axial planes parallel to those of the host migmatite. In some cases coarse grained pegmatites dikes are also observed.

3.1.4.1 **Gneisses**

Medium to coarse grained dark grey biotite gneisses are heterogeneous and grade into stromatites. The grain size of the gneiss increases towards the south of the group but it is always finer than the locally accompanying stromatites. The coarsening of the gneiss is coupled with a decreasing intensity of foliation development. The S_2 schistosity/foliation is continuous and slightly anastomosed (Fig. 6a) and is defined by the preferred orientation of biotite or spaced and defined by the alternance of plagioclase rich and biotite rich domains Locally the tightly folding of the S_1 fabric is observed (Fig. 6a) therefore the dominant S_2 spaced or continuousfoliation results from the transposition of S_1. Quartz plagioclase veins of up to 1 cm thick are parallel to S_2. Locally in high-strain domains S_2 foliation is overprinted by a discontinuous S_3 locally in high strain domains.

Gneisses are medium- to coarse-grained with granoblastic-lepidoblastic textures and are composed of the mineral assemblage quartz+plagioclase+biotite±apatite±zircon±rutilo ±magnetite (Fig. 6a). Accessory minerals include apatite and zircon. Muscovite blasts overgrow the folded foliation. Quartz is xenoblastic and shows evidence of grain boundary migration recrystallization and chessboard patterns. Plagioclase normally oligoclase is xenoblastic to subidioblastic. Subgrains and glide twinning are observed. Biotite is greenish brown, subhedral and define the S_2 fabric. In some gneisses, tourmaline as disseminated subhedral to anhedral crystals may constitute up to 2-3% by volume.

3.1.4.12. **Stromatites**

Stromatitic migmatites (Fig. 2c, 3c) are associated with the gneisses and appear as isolated outcrops to the north of the Renca batholith especially near its NE contact or in tectonic contact with the banded schists of the La Cocha Group along the contact of the eastern belt (Steenken et al., 2006). Elsewhere the type of contacts either with the schists or the gneisses is not easy to solve since they are partially covered. Stromatites are coarser grained than the biotite-gneiss. Leucosome content is variable at the outcrop scale (Fig. 2c) and in most cases its felsic minerals are coarser grained than the mafic phases of the mesosome (Fig. 3c). The leucosomes are generally rimmed by thin, discontinuous melanosomes and range from concordant through slightly to strongly discordant to the S_2 foliation (Fig. 2c). The concordant leucosomes that constitute up to 20–30 %vol and exhibit a weakly developed and slightly wavy planar fabrics, are variably pytgmatically- or isoclinally-folded recording a pre-S_2 timing (Fig. 2c). In places, younger leucosomes cut the deformed leucosomes (Fig. 3c).

The stromatitic migmatites are tonalitic and composed of quartz, plagioclase (oligoclase-andesine), biotite and muscovite together with tourmaline, apatite, zircon and scarce sillimanite as accessory minerals. Rocks are coarse to medium grained and characterized by the alternance of 0.5–5 cm thick layers of medium- to coarse-grained (Fig. 2c, 3c), strongly recrystallized granitic leucosome and biotite rich mesosome with some layers of paleosome (gneiss). Textures vary from granoblastic for the leucosome to grano-lepidoblastic for the mesosome/paleosome. In most of the localities there is no preservation of D_1 structures. D_2 fabrics overprint D_1 (Fig. 6b).

In the leucosome, plagioclase is mostly fresh with diffuse normal zonation and occasionally deformational twins. Quartz is anhedral, limpid with undulose extinction to parallel subgrains and locally grain boundary migration textures. Chessboard patters in the central part of relict crystals document high temperatures sub-solidus deformation. Biotite is light brown/honey-yellow to greenish brown and partially replaced by tiny muscovite and epidote. Xenomorphic plagioclase is mostly undeformed but shows intense sericitisation.The melanosome is made up of quartz, plagioclase and biotite. Biotite together with elongated quartz crystals and scarce muscovite define the S_2 foliation. Fibrolite is occasionally present associated with biotite and also as relic fibrolite inside muscovite blasts that overprint the S_2 foliation (Fig. 6c). Unmelted biotitic schists are interlayered in the migmatitic areas. In the Santa Rosa Group discrete shear belts related with the green-schist D_4 overprint (López de Luchi et al., 2008) either do not generate a penetrative fabric (Fig. 6d) controlling retrograde assemblage of chlorite and sericite or in highly strained areas define an anastomosed foliation by muscovite-epidote or muscovite-biotite (small)-epidote assemblages.

3.1.5.THE SAN FELIPE GROUP

The San Felipe Group (SFG) shows a more restricted geographical distribution mainly within the southern half of the CMC (Fig 1a), especifically in the Sierra de San Felipe and surroundings and in a wider area from south east of Paso Grande up to La Totora Granite (Fig.1a). This group partially hosts the Devonian La Totora and Renca batholiths. These rocks that are highly heterogeneous and grade from orthogneiss to diatexite and to stromatite with granitic leucosomes (Fig. 2d, 3d), differ from the Santa Rosa Group because diatexites are more abundant and exhibit retrograded cordierite, garnet and sillimanite. Rafts of biotite gneiss are often preserved in the stromatitic type. In cases where the spaced foliation S_2 of the biotite gneiss exhibits the tightly folded S1 foliation, the main foliation of the migmatites wrap around them suggesting a syn-D_2 timing of the leucosome development. Leucosome layers of these heterogeneous migmatites have variable width (centimetric to decimetric) and form an interconnected net of veins that thins out along the strike or are oblique to the foliation. When the leucosome become thicker (up to 10 cm), they are oblique to the metamorphic layering and may eventually coalesce to form discordant dikes and concordant sills that grade into the granite orthogneiss. Locally, the melts appear to have migrated along foliation planes and accumulated within shear bands or in boudin necks suggesting that melting and/or melt movement in the rocks was coeval with deformation.

In the stromatites, discrete light-colored leucosomes vary from rounded blebs less than five millimeters across to pods stretching several centimeters, elongated parallel to the dominant foliation (Fig. 2d, 3d). Many of the leucosomes are pytgmatically- or isoclinally-folded. The diatexites show also variable textures from rocks in which distinct leucosome /melanosome boundaries are recognized. i.e. schlieren diatexite to rocks with poorly-defined boundaries, i.e. massive diatexites that grade into orthogneisses. These diatexites often possess a very consistent grain size for both the felsic and mafic minerals (Fig. 7a). Although in other cases felsic and mafic minerals are evenly distributed, localized concentrations of mafic minerals are present.. Lenses of granitic orthogneiss exhibit irregular contacts against the stromatitic migmatites

The leucosomes show a medium-grained blastic texture and lobated intergranular contacts consisting of quartz, plagioclase and biotite and locally sillimanite and K feldspar, with apatite, zircon and Fe-Ti oxides as accessory minerals (Fig. 7a). Quartz appears in large, equidimensional amoeboidal crystals that exhibits grain boundary migration recrystallization and includes biotite and apatite. Contacts with the K-feldspar crystals are lobated. Myrmekitic intergrowths are present in the borders in contact with the plagioclase. Sillimanite (Fig. 7b) appears as thin prisms inside muscovite blasts that overgrow the S_2 foliation.

The mesosome is made up of lepidoblastic and granoblastic layers. The micaceous layers are made up of somewhat oversized biotite blasts that include plagioclase and opaque minerals (Fig. 7c). The granoblastic layers are composed of quartz, plagioclase, biotite and very scarce cordierite and accessory minerals. Quartz shows large grains with lobated edges on the quartz-feldspathic microlithons, but also quartz ribbons are locally parallel to the biotite layers. Plagioclase is found as relictic porphyroclasts and shows tapered twins. In some examples recrystallization of plagioclase into mosaics of pseudopolygonal individuals with triple junctions at 120º is observed. Cordierite occurs in irregular interstitices as small crystal with scarce or no pinnitization.The melanosome is represented by medium grained schlieren composed of biotite with lepidoblastic texture, scarce sillimanite and garnet. Discrete shear zones (Fig. 7d) with a recrystallized quartz mosaic and tiny flakes of biotite appear in some localities

3.1.5.1 **Orthogneisses**

Discrete orthogneissic bodies rarely reach some tens- meters, more often they form decimeter sized layers interbedded with the host migmatites. These rocks corresponds to the earlier granites of López de Luchi et al. (2008). Rocks are slightly pinkish light to medium grey equigranular medium to coarse granitoids that range from tonalite to syenogranite with a predominance of monzogranite. Minerals are quartz, microcline, plagioclase, biotite, muscovite and in some cases garnet, scarce sillimanite and cordierite. Accessory minerals include allanite, apatite, tourmaline and zircon. Biotite is homogeneously distributed and defines a foliation that is parallel to the S_2 fabric of the migmatites.

Rocks are typically banded with irregular distribution of S_2 oriented biotite which locally forms mm to cm large clots, plagioclase is slightly more sodic than in host schists, rarely garnet occurs surrounded by retrograde post- S_2 plagioclase coronas (Fig. 7e). Minor amounts of S_2 aluminum silicate (generally sillimanite) locally occur in association with albite indicating that conditions of first sillimanite isograde (upper amphibolite facies) have been attained.

Quartz is anhedral and exhibits static recrystallization indicated by polygonal aggregates of equant quartz crystals whereas in other sectors grain boundary migration recrystallization and chess board patterns are observed. Anhedral to subhedral plagioclase (oligoclase-andesine) exhibit borders corroded by quartz, glide twinning and locally subgrains. Inclusions of biotite, drop-like quartz and scarce K-feldspar are seen (Fig. 7f). Locally some needle sillimanite is observed in relation with muscovite blasts. K-feldspar, microcline is anhedral, locally interstitial and show flame perthites. Fine grained inclusions of plagioclase were observed. Myrmekite intergrowths are common between coarse grained plagioclase and K-feldspar. Biotite appears in discontinuous rafts or as small grains inside the feldspars. In areas in which mylonites are developed, K-feldspar and plagioclase that exhibit subgrains are wrapped by a S_3 surface. A discontinuous foliation is defined by the alignment of biotite rafts.

3.2. PROVENANCE OF THE METACLASTIC UNITS

Samples and analytical techniques

The metamorphic rocks that were studied for provenance comprise schist of the La Cocha and San Martin groups with metamorphic assemblages that correspond to quartz-andesine/oligoclase-biotite±muscovite ±garnet±rutile and scarce chlorite.

Twenty four whole rock major and trace elements data are presented (Table 1). This study is based on eleven new whole rock major and trace element chemical analysis, our own published data (López de Luchi et al., 2003, Drobe et al., 2009, 2011) and three samples from sillimanite schists from Sierra del Morro taken from Sims et al. (1997). Samples were selected considering no evidence of veining or open-system behaviour. Data were recalculated to an anhydrous base and plotted on classification, provenance and tectonic discrimination diagrams. See Appendix 1 for analytical procedures.

3.2.1 WHOLE ROCK GEOCHEMISTRY

Whole rock chemical composition of (meta)sediments coupled with detrital zircon ages, Nd and Sr isotopic composition might allow to track the source and tectonic setting of metaclastic rocks (e.g. Floyd and Leveridge, 1987; McLennan et al., 1990, 1993; Nesbitt et al., 1996, Condie et al., 1995, 2001, López de Luchi et al., 2003, Drobe et al., 2009, 2011; Cawood et al., 2012, Rudi et al., 2016; Cisterna et al., 2018). Major elements have been largely used in provenance analysis but they are less reliable (McLennan et al., 1990; Nesbit et al., 1996) than trace elements like the HFSE and REE that remains stable during weathering, diagenesis and metamorphism and makes them far more suitable for the discussion of the provenance and the tectonic settings of the basins where clastic sediments were deposited (e.g. Bathia 1983; Taylor and McLennan 1985; McLennan et al., 1993; Roser and Korsch, 1986,1988; Roser et al., 1996).

3.2.1.1 **Classification and source weathering**

In the classification of Herron (1988) the metaclastic rocks from the San Martín and La Cocha groups are mainly wackes, some of them next to the limit between shale and wacke. The three samples of Sierra del Morro are shale (Fig. 8a). Fe_2O_3/K_2O ratio for the wackes is bracketed between 0.1 and 0.5 which approximately corresponds to the values of this ratio for tonalite to monzogranite. In a sedimentary environment, Fe-Mg minerals are less stable than felsic minerals therefore the increase in Fe_2O_3/K_2O could indicate higher lithic content and/ or larger contribution of mafic sources,and probably proximity to the source (Herron, 1988). The trend of increasing Fe_2O_3/K_2O with SiO_2/Al_2O_3, opposite to the that of differentiation in igneous rocks (Fig. 8a), could relate to weathering and transportation.
The chemical index of alteration (CIA) Nesbitt and Young (1982) is a useful parameter to quantify the weathering degree of source rocks. The analysed sample collection shows CIA (Fig. 8b, Table 1) values of 54 to 64 that indicate an overall low-moderate weathering. One very fine grained schist of the La Cocha Group and one fine grained biotite-muscovite schist and the sillimanite schist of the San Martín Group show higher values between 76 to 84 which would be consistent with a shale protolith. Apatite and carbonate are very low or even absent in the selected samples and thus do not affect the CIA significantly. Our data are plotted in the ternary diagram Al_2O_3, $CaO^* + Na_2O$ and K_2O (Fig. 8b), to evaluate the deviation that the metaclastic rocks show from the original source (Nesbitt and Young, 1982; Nesbitt et al., 1996). Since the trend for each of the units is close to parallelism to the A-CN axis, no significant modification of the $(CaO^* + Na_2O)/K_2O$ ratio from that of the source is observed. The intersection of the weathering trend defined by the studied samples with the feldspar join suggests a source rocks with variable plagioclase/K-feldspar that would be granodiorite for the La Cocha Group and

granodiorite-granite for the San Martin Group. This type of source agrees with the values for the Fe_2O_3/K_2O ratio for granodiorite and monzogranite.

3.2.1.2 **Provenance and tectonic setting**

Different trace elements are used to analyse both provenance and tectonic setting. Sc, Th, Zr, REE, and high field strength trace elements (HFSE) are reliable chemical proxies in provenance and tectonic setting analysis of sediments and metasedimentary rocks due to their insolubility and immobile character during transport, diagenesis, weathering, and metamorphism (McLennan et al., 1990, 1993).

The contrasting compatibility of Th and Sc in mantle-derived rocks makes the Th/Sc ratio sensitive to average source composition and to identify a mafic component in the (meta)-clastic material (Taylor and MacLennan, 1985). Different average composition for upper continental crust (UCC) indicate a value from 0.8 to 1 (Taylor and McLennan, 1985; McLennan et al., 1990; McLennan, 2001). The fine grained schists of the SMG exhibit Th/Sc ratios (Fig. 8c, Table 1) from 0.6-0.7 except in two samples (M155 and A89-05) with values of 1.4 and 0.9 respectively that corresponds to rocks with higher CaO and P_2O_5 than the rest. The banded schists of the LCG show mainly UCC values from 0.8 to 1.4. In this case the three samples with higher Th/Sc, A70-04, MR10 and A10-06 are higher in CaO and in two of them also in P2O5 than the rest. Th/Sc values for the sillimanite schist of the Sierra del Morro are the lowest. Zr/Sc ratio that is a good indicator of sediment recycling, is higher in sands and fine grained sediments related to passive margin and collisional related basin than in those from active margin basins due of the higher abundance of heavier minerals, such as zircon in the sands (McLennan et al., 1990, Mc Lennan 2001). Values of this ratio are on average higher in the La Cocha Group, i.e. 19-44 than in the San Martín Group, 5-26 and in the sillimanite schist of Sierra del Morro (Table 1, Fig.8c). Therefore, based on Th/Sc ratio a source with a mafic component somewhat less recycled and more akin to active margin related basins is suggested for the SMG and the schist of Sierra del Morro. In any case, the wide range of variation in both parameters indicates either a wide spectra of source compositions, lithological variability or an unstable tectonic environment. High La contents are characteristic of intermediate to basic sources whereas Th is higher in felsic rocks. La/Th ratio is on average lower in the fine grained schists than in the banded schists that is characterized by values close to upper crustal values of 2.82 (Taylor and McLennan 1985) (Table 1). Co/Th (Table 1) mostly below 1.27 (the higher value for felsic igneous rocks) for the La Cocha Group suggest also felsic sources but values on average higher for the San Martín Group might suggest the involvement of a mafic component.

Traditional discrimination diagrams for characterizing the tectonic setting of sedimentary basins based on major-element compositions do not incorporate a coherent statistical treatment of the data (Verma et al., 2013 and references therein). Nevertheless since CIA values suggest a trend of normal weathering, our dataset is evaluated in the conventional TiO_2 vs Fe_2O_{3t}+MgO (Bhatia, 1983) as well as using the robust statistical approach based on the discriminant functions (Table 1) proposed by Verma et al. (2013).

In the traditional plots (Bhatia, 1983; Bhatia and Crook, 1986) “oceanic island arcs” are sedimentary basins adjacent to oceanic island arcs or island arcs partly formed on thin continental crust with sediments mainly derived from the calc-alkaline or tholeiitic arc, and continental island arc basins correspond to the sedimentary basins adjacent to island arcs formed on a well-developed continental crust or on thin continental margins. Intra-arc, back-arc, and fore-arc basins of this setting are dissected continental arc fragments detached from the mainland and are fed by sediments mainly derived from felsic volcanic rocks.. Active continental margins are developed on or adjacent to a thick continental crust composed of rocks of older fold belts. Sedimentation corresponds to retro-arc foreland marginal basins and oblique-slip basins sourced on granite-gneisses and siliceous volcanics of the uplifted basement.

In the TiO_2 vs Fe_2O_3t+MgO (Fig. 8d, Table 1) samples mostly plot in the fields of oceanic and continental island arc whereas the discriminant functions (Table 1) (Verma et al., 2013) indicate arc setting which implies active volcanism.

Ti/Zr and La/Sc ratios (Bhatia and Crook, 1986) are helpful in the identification of the relative contribution of magmatic vs. recycled sources and can discriminate successfully between different tectonic environments. Ti and Sc represents the abundance of mafic phases and are used to evaluate the volcanic vs. mantelic fingerprint in the source and in consequence the maturity of magmatic arcs (Bathia and Crook, 1986). Zirconium (Hf and Th) is the main constituent of zircon, the content of which increases due to rework in sedimentary rocks. La/Sc ratios are controlled by the mafic or felsic composition of the source; input from mafic and ultramafic source areas results in an enrichment of Sc. In the Ti/Zr vs. La/Sc diagram (Fig. 8e) (Bhatia and Crook, 1986), samples of the La Cocha Group plot in an area of overlapping between “active continental margin” and “continental island arc” in a trend similar to the recent deep sea turbidites deposited at continental arc margins (Mc Lennan et al., 1990) as already indicated by (López de Luchi et al., 2003). In the plot of $(La/Yb)_N$ vs Eu/Eu* (Fig. 8f) most of the samples are located in the back -arc and continental arc fields.

In synthesis, major and trace elements values and ratios suggest an arc related setting with a major contribution from felsic sources. REE patterns are generally fractionated, with $(La/Yb)_N$ values of 6.87 to 9.70 for the La Cocha Group and more variable from 5.07 to 12.89 for the San Martín Group (only three data are available). Significant enrichment in LREE (Table 1) and the distinctive negative Eu/Eu* coupled with a relatively flat $(Gd/Yb)_N< 2$) HREE pattern suggest derivation from upper crustal rocks composed of felsic components (McLennan et al.,1990).

3.3. LITHOLOGY AND GEOCHEMISTRY OF THE METABASIC UNITS: AMPHIBOLITES

3.3.1.**Field and lithological features**

Amphibolites which are relatively scarce in the CMC, appear as concordant layers or lenses of dark green to dark grey medium-to fine-grained variably foliated rocks mostly injected by concordant and discordant leucocratic veins. The fine grained amphibolites show a well defined foliation and lineation and form concordant layers (up 70 cm thickness) or lenses (up 3 m long). In the scarce outcrops inside the schists of the LCG, amphibole porphyroblasts and biotite define a planar/linear fabric in a matrix composed of epidote, hornblende, actinolite, plagioclase, ilmenite and titanite. Most of the minerals are anhedral. Generally, the amphibole exhibits a light green core and a dark green border. In the migmatites and gneisses, amphibolites are medium grained, hornblende and plagioclase are the main mineral constituents; whereas quartz, biotite, epidote, titanite and ilmenite are minor phases together with occasionally zircon and apatite. Chlorite, intergrown with or replacing biotite was developed along the foliation planes. Garnet was found in some amphibolite layers located close to the contact between the banded schist of the LCG and the SRG. In some cases a fine banding is defined by grey milimetric layers consisting of plagioclase and scarce quartz and dark green layers of amphibole and scarce yellow-green epidote-clinozoisite. Textures are nematogranoblastic with polygonal to lobate intergranular contacts or nematoblastic and lepidoblastic when biotite is present. Leucocratic fine to medium grain veins consist of quartz, plagioclase and scarce biotite, while coarse grain veins exhibit quartz, microcline, plagioclase and biotite.

Regionally some amphibolites are associated with calc-silicatic rocks. Delakowitz et al. (1991a, b) described in the western flank of the Sierra del Morro. i.e. in the present Santa Rosa Group, an association of (banded) amphibolites, epidote-hornblende schists, marbles, (scheelite-bearing) calc-silicates, (tourmaline-bearing) mica schists and, sporadically, hornblende schists emplaced in gneisses and migmatites.. No garnet was mentioned in these amphibolites of Sierra del Morro. The granoblastic calc-silicate layers are made up by alternating 0.2 mm to 1 mm bands of different grain size composed of clinoamphibole, garnet, titanite, quartz, apatite, and opaque minerals.

3.3.2. **Geochemistry**

Five whole-rock major and trace elements compositions of the amphibolites are presented. Two data (21-3, P16) belong to garnet bearing amphibolites whereas samples P22, P68 and P69 are hornblende-plagioclase amphibolites. See Appendix 1 for analytical procedures.

The samples are chemically classified as high-Mg tholeiitic basalts (Fig. 9a) based on major elements (Jensen, 1976) and as basalts (Fig. 9b) based on trace elements ratios, i.e. Zr/Ti vs Nb/Y (Pearce, 1996) diagrams respectively.

Major element chemistry (Table 2) contents show a limited range for major elements, i.e. SiO_2 47.4 to 50.4 wt.%, Al_2O_3 14.2 to 15.5 wt.% with moderate Fe_2O_3 (9.7 to10.7 wt.%), CaO (11.5 to 13.8 wt.%) and MgO (5.8 to 7 wt.%) concentrations, low MnO (0.14 to 0.17 wt.%) and P_2O_5 (0.07 to 0.21 wt.%) contents. The total alkali contents (Na_2O+K_2O) for these rocks is below 3.2 wt.%. Mg# from 53 to 58 indicates an evolved signature. Abundances of transitional trace elements (Ni, Cr, and Co) are low whereas Rb, Ba, Sr, Pb and LREE, high field strength elements like Nb and Ta show moderate contents.. Primitive mantle normalized REE patterns (Fig. 9c) show a negative slope with $(La/Yb)_N$: 4.1 to 5.7, $(La/Sm)_N$: 2.1 to 2.9 and (ii) moderate $(Gd/Yb)_N$ = 1.2 to 1.4. Primitive mantle□normalized incompatible trace element patterns show (Fig. 9d) positive peaks at U, Ta, Pb, and Nd and negative Nb anomalies.

Tectonic setting discriminant diagrams based mainly in the less mobile elements indicate MORB or island-arc basaltic precursors for these amphibolites. Plots indicate MORB in the MnO-TiO_2-P_2O_5 (Mullen, 1983) diagram (Fig. 10a), E-MORB in the Zr-Nb-Y diagram (Meschede, 1986) (Fig. 10b), MORB -IAT in the Zr-Ti-Y (Pearce and Cann, 1973) (Fig. 10c) diagram. As samples show a LILE, Pb, U and LREE enrichment combined with MORB signature the most likely signature is comparable with E-MORB (Gale et al., 2013 and references therein) or with a back-arc basalts signature (Pearce and Stern, 2006). FeO*/MgO 1.3 to 1.6 for a TiO_2 range from 1.7 to 1.9 wt.% and Ba/Nb between 17 to 18 and Nb/Yb between 3.4 to 4.1 (Table 2) are an indication of back arc basalts (Pearce and Stern, 2006; Gale et al., 2013).

3.4. METAMORPHISM OF THE METABASIC UNITS

Pressure-temperature pseudosections were calculated for sample 21-3 using the software package Theriak/Domino (De Capitani and Petrakakis, 2010) and the database of Holland and Powell (1998) for the geologically realistic system MnNCKFMASHTO (MnO-Na_2O-CaO-K_2O-FeO-MgO-Al_2O_3-SiO_2-H_2O-TiO_2-Fe_2O_3). The bulk composition of the sample was determined by XRF analysis and is given in Table 2.

Sample 21-3 that belongs to the Santa Rosa Group near the contact with the La Cocha Group immediately north of the Renca Batholith (Table 2). contains (Fig. 11a, b) garnet, amphibole, plagioclase, quartz, epidote and chlorite with accessory ilmenite, titanite, rutile, zircon and apatite. Garnet (Fig. 11a) forms porphyroblasts (up to 2 mm) with inclusions of epidote, plagioclase, quartz and ilmenite and is commonly fractured with chlorite occurring along the cracks. Compositionally, garnet (Fig. 11c) has some minor zoning with elevated XSps (=Mn/(Fe^{2+}+Mg+Ca+Mn)) in the core (0.36 to 0.3), lower Xalm (=Fe^{2+}/(Fe^{2+}+Mg+Ca+Mn)) in the core (0.34-0.40) and relatively flat Xpyr (=Mg/(Fe^{2+}+Mg+Ca+Mn)) and Xgrs(=Ca/(Fe^{2+}+Mg+Ca+Mn)) contents (0.08 to 0.09 and 0.24 to 0.2 respectively). Where ilmenite (with MnO of 10.2 to 12.8 wt%, and Fe_2O_3 of 1.0 to 4.1 wt%) occurs in the matrix, it is surrounded by a thin rim of titanite (Fig. 11b inset). The bulk of the matrix consists of equigranular amphibole (up to 1 mm in size), plagioclase and quartz (both usually up to 500 µm). Plagioclase has been extensively altered to sericite and where analyses were obtainable, it gives *X*ab (=Na/(Na+Ca+K) of 0.88 to 0.92. The muscovite that grows inside the plagioclase grains gives XK(=K/(K+Ca+Na)) of 0.93 to 0.94 and has Si p.f.u of 3.16 to 3.18. Amphibole has ragged grain edges, and commonly epidote and chlorite (Fig. 11a) occur on the edges of amphibole grains. Amphibole is classified as tschermakitic hornblende (one analysis was magnesio hornblende), has XMg (=Mg/(Mg+Fe^{2+}) values of 0.60 to 0.63, and contains inclusions of quartz, ilmenite and titanite. A single rutile was found within a titanite inclusion in amphibole. Two different varieties of *X*epidote are present and occur as inclusions within garnet and amphibole and within the matrix. They have different Xps (=Fe^{3+}/(Fe^{3+}+Al)) of 0.09 to 0.12 or 0.19 to 0.25. In some cases, separate grains show different composition while in other situations, a grain has a core of one composition that transitionally passes to the other. Chlorite occurs in the matrix as coarse grains growing on the rims of amphibole, it also occurs on large cracks within garnet grains. Chlorite has X_{Mg} of 0.53 to 0.63 (Suppl.- Mat. 1).

3.4.1. **Pressure-Temperature conditions**

For sample 21-3, the mineral assemblages and field observations indicate peak metamorphic conditions to be sub-solidus. For this reason, H_2O was set in excess. Compositional isopleths were calculated for garnet and plagioclase to further constrain the peak P-T conditions (X_{Alm}, $_{Sps}$, X_{Pyr}, X_{Grs} and X_{Ab}). The amount of Fe_2O_3 present in the bulk composition was determined based on recalculated compositions and modes of the Fe_2O_3 bearing minerals present (Droop, 1987). A T-MFe_2O_3 diagram was also calculated to see the effect higher or lower Fe^{3+} content has on the modelled mineral assemblages (see Supplementary files).

The following mixing models and notations were applied: garnet, White et al. (2005); plagioclase, Holland and Powell (2003); biotite, White et al. (2005); white mica, Coggon and Holland (2002); ilmenite, White et al. (2005); chlorite, a combination of Mahar et al. (1997) and Holland and Powell (1998); epidote, Holland and Powell (1998); talc, Holland and Powell (1998); clinoamphibole, Diener et al. (2007); clinopyroxene, Green et al. (2007); magnetite-spinel, White et al. (2002); orthopyroxene, White et al. (2002). Rutile and quartz are also included as pure phases.

3.4.2. **P-T modelling results**

Based on the sample petrology, the stable peak assemblage is garnet + amphibole + ilmenite + quartz + plagioclase + epidote + titanite. Due to the presence of Fe^{3+} bearing minerals, a T-MFe_2O_3 diagram was calculated initially (see supplementary figure). This was conducted at 6 kbars because this is the pressure indicated by the amphibole compositions. However this diagram showed large stability fields for clinopyroxene which is not observed in the sample. Potentially this results from an excess of calcium in the whole rock composition due to the presence of apatite. Despite being a minor component in the samples (only 1 to 2 %), the large proportion of calcium in apatite (which is not considered in the modelling) can cause excess Ca in the modelled composition. In order to evaluate this, a T-MCa diagram was made (see supplementary figure). This diagram uses a moderate Fe^{3+} value because this stabilises ilmenite at slightly lower temperatures. The right hand side of the T-MCa diagram is based on the Ca content of the measured bulk composition. Moving left, with a Ca decrease, clinopyroxene stability moves up temperature. The interpreted peak assemblage field for the sample is present on this diagram (field in bold), the reduced value of Ca was used to create a new P-T diagram for the sample. The interpreted peak field is present on this diagram, occurring at 5 to 6.5 kbar and 500 to 650 °C (Fig. 12). Garnet compositional isopleths were conducted in order to provide additional information about the P-T evolution. However, these do not intercept. Potentially this may be a result of an inappropriate Ca or Fe^{3+} value used in the composition. Isopleths were also calculated for the T-MCa and TMFe3 diagrams and varying these values does not produce more suitable compositions. Another possibility is that the garnet has experienced some diffusional re-equilibration at peak conditions, since the cations diffuse at different rates, offsetting the intercept. The XAlm and Xsps values do occur within the peak field, although they do not overlap. Both additionally show the same trend with the P-T evolution moving up T and P. The garnet information, along with the changes in mineral assemblage define the P-T evolution for the sample. These changes in particular are the presence of albite inclusions in garnet, and titanite inclusions in amphibole; this corresponds to the early portion of the evolution being in an albite + titanite bearing field. Garnet growth and the presence of ilmenite (plus a rutile inclusion in ilmenite) suggests an evolution involving garnet growth into the peak field. While we have no constraints on the peak T conditions, the garnet seems to have experienced diffusional re-equilibration, which only occurs in the time frame of a metamorphic event at temperatures above 600 °C and on small garnet grains such as those present in this sample (Carlson, 2006; Caddick et al., 2010). So we suggest peak conditions were 6 kbar and 620 °C (limited only by the clinopyroxene present field). This is confirmed by geothermobarometry in sample 21-3 where tschermarkite coexists with plagioclase, epidote and Ti bearing phases. The calibration of amphibole geothermobarometer by Zen and Schluz (2004) yielded a narrow range of P-T values between 600-623 °C at 6.0 kbar. Late chlorite on the rims and in cracks of garnet, along with titanite rims on ilmenite and matrix plagioclase breaking down to albite suggests that the P-T path moved back down T (see P-T Figure 12).

3.5. ELECTRON MICROPROBE (EPMA) MONAZITE DATING

Monazite dating was performed on four samples: one migmatite, P7, of the Santa Rosa Group, two, 19-9 and 65-04, from the banded schists of the La Cocha Group and one 104, from the San Martín Group.

3.5.1 ANALYTICAL METHODS

3.5.1.1 **SEM-based automated mineralogy (MLA)**

Automated mineralogical methods (e.g. Fandrich et al., 2007), based on a scanning electron microscope SEM Quanta 600-FEG-MLA by FEI Company, equipped with Bruker Dual X-Flash energy dispersive spectrometers for EDX analyses were applied to complete thin sections of coarse- and fine-grained biotite-plagioclase paragneisses, as described in Schulz (2017). Acceleration voltage of the focussed electron beam was set at 25 kV, which corresponds to beam current of 10 nA. A software package for mineral liberation analysis (MLA version 2.9.0.7 by FEI Company) was used for the automated steerage of the electron beam for EDX identification of mineral grains and collection of numerous EDX spectra. The following measurement routines were applied:

XMOD is a single spot point counting for mineral mode analysis, based on ~ 10^5 EDX spectra at a 10 µm step size in a thin section. Mineral modes in wt% (Table 3) have been recalculated by introducing average densities listed in mineral databases. A calculated chemical assay was derived from the mineral mode, densities, and the mineral chemical compositions. For mica and plagioclase the representative compositions from EDX measurements were introduced. The calculated assays (Table 3) thus represent an approximation to the bulk rock composition in a given thin section plane and may differ from the bulk composition obtained from a larger volume of rock.

The SPL (Sparse Phase Lineup) search routine applies a backscattered electron (BSE) grey colour value trigger for the detection of minerals of interest which subsequently are analysed by single spot EDS spectrum (Schulz et al. 2020). This enables the detection and identification of rare phases as monazite, xenotime and zircon and their surrounding minerals. With a line-up function one receives a catalogue of all monazite and xenotime grains with their intermineral relationships in a sample. It also allows the quantification by which minerals monazite is locked (Table 3). This was used to select monazite grains for detailed investigation in backscattered electron imaging under the SEM, and quantitative WDS analysis with the electron microprobe (EPMA).

3.5.1.2. **Electron microprobe (EPMA) monazite dating**

Electron microprobe Th-U-Pb dating is based on the observation that common Pb in monazite (LREE, Th)PO_4 is negligible when compared to radiogenic Pb resulting from the decay of Th and U (Montel et al., 1996). Electron microprobe analysis of the bulk Th, U and Pb concentrations in monazite, at a constant $^{238}U/^{235}U$, allows for the calculation of a chemical model age (CHIME) with a considerable error (Montel et al., 1996; Pyle et al., 2005; Jercinovic et al., 2008; Suzuki and Kato 2008; Spear et al., 2009). A protocol for the monazite analysis with a JEOL JXA-8230 electron microprobe hosted at the Institute of Material Science of the TU Bergakademie Freiberg/Saxony, Germany, has been developed and adopted. The electron beam was set at 20 kV acceleration voltage, 100 nA beam current, and 5 µm beam diameter. For analyses of monazite <10 µm in diameter, the beam diameter was set to 3 µm. The Mα1 lines of Th and Pb and the Mβ1 lines for U of a PETL crystal in a spectrometer with a capsuled Xe proportional counter were selected for monazite analysis. Orthophosphates of the Smithsonian Institution (Jarosewich and Boatner 1991; Donovan et al., 2003) were used as reference materials for calibration of REE and Y, ignoring the low residual Pb (< 0.5 wt%) in some of the crystals. Calibration of Pb was carried out on a natural crocoite. The U was calibrated on a metal reference material. The Th was calibrated on a reference monazite labelled as Madmon, with validated special ThO_2*-PbO characteristics (Schulz et al., 2007; Schulz and Schüssler 2013; Schulz et al., 2019). The Madmon reference monazite was also used for offline re-calibration of ThO_2, as well as for the control of age data. Instrument drift during the 2 analytical sessions of 24 hrs was controlled by measurements on the Madmon reference monazite and was negligeable for Th, Pb and U. Analyses with Al and totals outside the range from 98 – 101.5 wt % were not considered for further presentation.

Interference of YLγ on the PbMα line was corrected by linear extrapolation as proposed by Montel et al. (1996). An interference of ThMγ on UMβ was also corrected. The number of single analyses varies with the grain size of the monazite, e.g. 2-5 analyses in grains of 30 - 40 µm. In most cases the small grain sizes of the monazite (<20 µm) allowed only one single analysis per grain. Monazite chemical ages were first calculated using the methods of Montel et al. (1996). A 1σ error deduced from the counting statistics (JEOL error) and an error $\varepsilon_{Pb} = \sqrt{(Cts/s_{PEAK} + Cts/s_{BKG})}/(Cts/s_{PEAK} - Cts/s_{BKG})$ was propagated to an error in Pb element %. For Pb the error in element% is ~0.004 (recalculated from the JEOL error) or ~0.001 (recalculated from ε_{Pb}) for the reference monazite Madmon (~0.25 wt% Pb). We applied an error in Pb element% of 0.004 to all analyses, which propagates for the reference monazite Madmon with ~506 Ma typically to ±16 Ma (2σ), and for Ordovician monazites (~0.09 wt% Pb; 2 - 4 wt% Th) to 30 - 40 Ma (2σ). Ages were further determined using the ThO_2*–PbO isochron method (CHIME) of Suzuki et al. (1994) and Montel et al. (1996) where ThO_2* is the sum of the measured ThO_2 plus ThO_2 equivalent to the measured UO_2. This is based on the slope of a regression line in ThO_2* vs PbO coordinates forced through zero. As the calculation of the regression line provides underestimation of the error, weighted average ages for monazite populations were calculated from the single analyses defining the regression line using Isoplot 3.0 (Ludwig, 2001). In all analysed samples, the model ages obtained by the isochron and the weighted average methods coincide within the error. The age data are interpreted as the time of closure for the Th-U-Pb system of monazite during growth or recrystallization in the course of metamorphism.

3.5.2. Monazite ages and mineral chemistry

La Cocha Group:Sample 65-04 is a fine grained banded biotite-plagioclase schist with 2 mm thick quartz-rich planar layers alternating with 1 mm thick biotite- and plagioclase-rich layers (Table 3). In some of the layers large muscovite blades are oriented oblique to the layering. The sample contains many small monazite grains (25 µm diameter at MD50, Table 3), which mostly allowed only a single analysis per grain. In this sample significantly most monazite grains are completely locked or in contact with epidote. When a double corona of inner apatite and outer allanite (epidote-group mineral) surrounds a monazite, this provides a hint to a retrogressive replacement of monazite (Finger et al. 1998). Such corona structures are not observed here. Therefore it is concluded that monazite crystallized from epidote. The weighted average age by the monazite analyses which define the isochron is at 418 ± 10 Ma
(Fig. 13a).

La Cocha Group: Sample 19-09 is also a banded biotite plagioclase paragneiss with alternating biotite- and quartz-rich layers. In the biotite-rich layers appear small round garnet porphyroblasts with 0.2 mm in diameter. The thin section is cut perpendicular to the fold axis of a similar F_3 fold that affects S_2. Biotite blades show preferential orientation parallel to the axial plane foliation of the fold, whereas some large muscovite blades are decussate in reference to the layering and the axial plane foliation. Muscovite is therefore interpreted to have crystallised post-folding. Monazite grains are mostly small (33 µm diameter at MD50, Table 3). Some of the monazite grains are partly locked by apatite. In a rare single case, a monazite grain is in contact to xenotime. The weighted average age given by the monazite analyses which define the isochron is at 434 ± 12 Ma (Fig. 13b).

San Martín Group: Sample 104 is a fine grained banded biotite plagioclase paragneiss with alternating biotite-plagioclase- and quartz-rich layers. Large muscovite porphyroblasts overgrow the foliation which is defined by preferentially oriented biotite flakes. There are also some decussate biotite flakes oriented perpendicular to the foliation. Monazite grains are slightly larger than in samples 65-04 and 19-09, with 38 µm diameter at MD50 (Table 3). Some monazite grains are in contact with large apatite grains and one case of a monazite-xenotime contact was observed. The weighted average age given by the monazite analyses which define the isochron is at 431 ± 8 Ma (Fig. 13c).

<u>San Felipe Group</u>: Sample P7 is a medium-grained biotite plagioclase paragneiss composed of a biotite-rich and a plagioclase-quartz layer. Muscovite appears in fine-grained aggregates. Some of the biotites are retrograded into chlorite (Table 3). The monazite grain sizes are the largest among the sample suite and are 60 µm diameter at MD50 (Table 3). This allowed to analyse some grains by multiple single spots. Some of the monazite grains display sutured grain boundaries and are surrounded by a corona of apatite with tiny thorite grains. This can be interpreted as a retrogressive replacement of monazite (Broska and Siman, 1998; Finger et al., 1998; Krenn and Finger, 2007; Budzyń et al., 2011; 2017).The weighted average age given by the monazite analyses which define the isochron is at 446 ± 5 Ma (Fig. 13d).

In XGdPO$_4$ vs XYPO$_4$ coordinates, the monazites are subdivided into two groups with different XYPO$_4$, ranging between the garnet zone at 0.03 and the sillimanite zone at 0.06 XYPO$_4$ as indicated in Pyle et al. (2001), at uniform XGdPO$_4$ between 0.02 and 0.03 (Fig. 14a). In the age vs Y_2O_3 plot, monazite in samples 65-04 and 19-09 have similar Y_2O_3 contents at around 1.4 wt%. In sample 104 the monazite Y_2O_3 contents are around 2 wt%. In sample P7 the monazite Y_2O_3 are 2 - 3 wt% (Fig. 14b). In the monazite XHREE+Y vs XLREE diagram (Fig. 14c) the four studied samples display overlap of data clusters and follow a common linear trend (Fig. 14c).The dominant cheralite exchange (Th or U + Ca = 2 REE) is typical for systems with elevated Ca in bulk rock composition (Spear and Pyle, 2002). Most of the monazites plot along the cheralite substitution trend and only some single analyses in samples 104 and 65-04 deviate (Fig. 14d) from this trend.

<u>4. DISCUSSION</u>

4.1 .PROVENANCE AND TECTONIC SETTING OF THE METACLASTIC UNITS

Geochemical criteria for the combined analysis of new and already published data of the metaclastic units of the Conlara Metamorphic Complex provide significant additional constraints on the sedimentary provenance and tectonic setting of the protoliths (López de Luchi et al., 2003). The metaclastic units (Fig. 8a) of the La Cocha and San Martín groups are greywackes and shales. The lower Th/Sc and Zr/Sc ratios (0.59 to 0.67 and 5 to 26 respectively) indicate a source with a mafic and less recycled component for the SMG whereas the higher Th/Sc and Zr/Sc ratios (0.99 to 1.41 and 19 to 44 respectively) in the LCG (Table 1, Fig.8c) suggest felsic and recycled sources. Lower La /Th and higher Co/Th (Table 1) ratios also point towards the involvement of a mafic component in the SMG protoliths. REE for the CMC (Table 1) shows a moderately negative Eu anomaly associated with low Gd_N/Yb_N being the fine grained schist of the San Martin Group those with the lowest values and the smaller Eu anomaly. All these results are typical for recent turbidites (McLennan et al., 1990). Negative Eu anomalies (Fig. 8f) also demonstrate intracrustal differentiation of the magmatic precursors by processes involving separation of plagioclase, such as partial melting or fractional crystallization.

Most of the inferences about the tectonic setting of the metagreywackes point to a continental island arc, active margin or back-arc basins (Fig. 8 d, e, f). The samples plot in areas of overlap of the fields for trailing edges, back arc and continental arc (Fig. 8f) indicating the development of a back-arc basin which evolved with the progressive erosion of the source area. Uplifted old basement and arc-related detritus must be the end members of the mixtures. The mafic rocks that are interlayered with the migmatites exhibit a chemical signature indicative of a back-arc setting which is akin to the inferred tectonic setting for the sedimentary basin.

Cawood et al. (2012) used the shape of the frequency plots (Fig.15a) and cumulative proportion curves (Fig. 15b) based on the difference between the measured crystallization age for a detrital zircon grain and the depositional age of the succession in which it occurs, to infer the tectonic setting of the basins. Although the number of spots fulfill the statistical requirements only for the two samples taken from Rapela et al (2016), all the curves are similar and indicate collisional to convergent settings. Actually in Cawood et al. (2012) back- arc basins are characterized by detritus with ages that approximate the depositional age of the samples together with those derived from adjoining cratons which make them similar to his proposal for foreland basins

4.2. RELATIONSHIP BETWEEN METAMORPHISM-DEFORMATION-MAGMATISM

The metaclastic rocks of the CMC are metagreywackes and scarce metapelites with lesser amounts of tourmaline schists and turmalinites. López de Luchi (1986) proposed a polyphase metamorphic evolution. The penetrative, dominant foliation, $S_{2,}$ locally associated with banding in some metaclastic schists is related to the second deformation (D_2). An earlier, relic S_1 is preserved in microlithons as biotite polygonal arcs or as thin folded leucocratic veins with S_2 axial plane cleavage (Fig. 5a).The penetrative S_2 foliation is in turn affected by upright, generally isoclinal, N-NE trending D_3 folds that control the patterns of the roughly NNE trending belts of the different groups that constitute the Complex (Fig. 1a). Non penetrative, discontinuous high-T shear zones within the schists and migmatites are related with D_4 whereas some fine-grained discontinuous shear bands that are especially conspicuous in the Sierra de San Felipe (Fig. 1a) attest for a D_5 deformation phase.

Deformation/metamorphism relationships are consistent for the different components of the CMC with peak metamorphic conditions being nearly synchronous with D_2 in the metaclastic units. Plagioclase coronas around garnet both in banded schists and in the migmatites might indicate a decompression path related to D_3. As Morosini et al. (2019) indicated that the 470 Ma Las Cañas Plutonic Complex was emplaced at shallow crustal levels in a marginal basin that was structured and metamorphosed previously to the Famatinian magmatism, an exhumation path would have developed predating its emplacement. Biotite-quartz mylonitic shear zones are related with D_4 whereas D_5 discontinuous shear bands are defined by sericite-quartz ± chlorite layers. The LCG and SMG are characterized by oriented biotite-quartz-plagioclase±muscovite ± garnet ± scarce fibrolite assemblages. In contrast the Santa Rosa and San Felipe migmatites and gneisses bear the non-diagnostic biotite (S_2 oriented)-plagioclase assemblage, with garnet and sillimanite rarely occurring as prograde phases and garnet displaying post S_2 plagioclase replacement. Muscovite blasts overprinting S_2 planes (Fig. 5c) are widely distributed on mica rich domains of all the rock types.

Preliminary calculations based on Ti in Biotite and Garnet-Biotite thermometers indicate temperatures from 529 to 608 °C and barometric constraints based on Garnet-Plagioclase-Muscovite-Biotite barometer of 5.4± 0.5 kbar for the La Cocha Group and from 582 to 755 °C and 6.5 ± 1 kbar for the San Felipe Group migmatites (López de Luchi et al., 2008, 2009). Morosini et al. (2019) obtained average P-T conditions for the metamorphic climax of the banded schists from~5.5 ± 1.2 kbar and 680 ± 37 °C and calculated a temperature gradient ~32.3 °C/km that corresponds to a low pressure series in the middle-amphibolite facies. Calculations on one garnet amphibolite, that appear as part of a disrupted layer along the S_2 foliation of the migmatites, suggest that the peak conditions were 6 kbar and 620 °C (Fig. 12)which are similar to the P-T calculated by Lopez de Luchi et al. (2008) for the San Felipe Group migmatites.

P-T calculations on the metaclastic rocks were performed using garnet that grew during S_2 development (Fig. 11a), therefore this data would correspond to the P-T constraints during D_2 development. In the migmatites and amphibolite S_2-D_2 as indicated by the P-T calculations suggest that these rocks attained M_2 at deeper levels of the crust than the La Cocha Group.

P-T values for D_3 can be roughly framed in the LCG by the calculated pressure of crystallization based on the Al-in hornblende barometer of ~4.45 ± 1.10 kbar for the intruding 470 Ma Las Cañas Pluton (Morosini et al., 2019). These results imply that the host rock, i.e. the La Cocha Group,was uplifted~4 km before this pluton was emplaced. Growth of andalusite blasts (Ortiz Suárez, 1988) might correspond to this stage of uplift. Isothermal decompression between D_2 and D_3 in the banded schist of the La Cocha Group could be indicated by the lack of thermal aureole around the La Tapera Pluton (López de Luchi and Cerredo, 2001) and the Las Cañas Pluton (Morosini et al., 2019). D_3 folded S_2 and controlled the emplacement of the syn-kinematic Ordovician granitoids. The average estimated temperature for D_5 based on the Chlorite thermometer would be 360°C (López de Luchi et al., 2009). In contrast biotite growth in D_4 shear zones may suggest higher temperatures. The present distribution of the groups that make up the CMC as NNE-SSW trending belts could result from the differential exhumation of the basement complexes related with D_4.

No clear evidence of syn-D_2 Pampean (?) granitoid magmatism was found in the CMC except for some orthogneisses of the San Felipe Group. The Ordovician magmatism which intrude the already metamorphosed CMC (López de Luchi et al., 2007) corresponds to the Ordovician Granodiorite-Granite (OGGS) and Tonalite (OTS) suites. The low temperature OGGS, like El Peñón Granite or Los Alanices granites are considered to represent partial melts from the CMC suggesting that a partial melting event may have occurred at or before 497 Ma (López de Luchi et al., 2007). The only evidence of in situ dehydration (?) melting is the growth of sillimanite in the migmatites which could be due to the prograde muscovite out reaction and the presence of cordierite. The back reaction in which muscovite blast that overprint the S_2/S_3 foliation include sillimanite/fibrolite is also present (Fig. 6c). The development of biotite rich mesosome might also suggest a water present melting event (Weinberg and Hasalova, 2015). Migmatitic rocks exhibit at least two deformation events (Fig. 6, 7). The main S_2 foliation is folded along which profuse leucogranitic leucosomes are recognized. Stromatites and diatexite are locally observed around some Ordovician granitoids that intrude otherwise non migmatitic rocks (López de Luchi, 1986, Ortíz Suárez, 1988, 1996, Steenken et al., 2006, López de Luchi et al., 2007).

Therefore if partial melting and associated migmatization are related to the low-T Ordovician granitoids (López de Luchi et al., 2007), either one or two discrete metamorphic events could have developed. In the latter case migmatization would have overprinted a first Pampean or older (?) metamorphism preserved not only in the non-migmatitic rocks like the La Cocha Group but also in the metamorphic enclaves inside the Ordovician granites. Alternatively if only one metamorphic event is recorded for the entire complex migmatization would be developed in rocks of suitable composition probably similar to the San Martin Group where profuse leucosome development is locally present whereas the psammitic rocks are not affected by this process.

López de Luchi et al. (2007) proposed that the emplacement of the OGGS was coeval with folding and shear zone development associated to D_3 whereas Morosini et al. (2019) proposed that Las Cañas Pluton emplacement was controlled by a shear zone during the exhumation of the CMC. The Ordovician granitoids may represent melts migrated from their sources probably structurally controlled. Probably D_3 was active in the interval 500- 470 Ma and controlled not only the ascent of crustal derived granitoid magmas sourced on the CMC but also that of the I-type magmas like La Tapera or Las Cañas plutons during the exhumation of the complex. The association of gabbroic rocks in the Las Cañas Pluton might explain a hot thermal regime and the slow cooling of the complex. The present distribution of belts in which in situ partial melting may be envisaged alternates with belts in which granitoids either S or I type are intruded, results from an event younger that the migmatization.

4.3. MAXIMUM DEPOSITIONAL AGES (MDA) BASED ON DETRITAL ZIRCON ANALYSIS

Four detrital zircon age data for the CMC were compiled from the literature: A73-05 (Drobe et al., 2009) corresponds to the banded schists of the La Cocha Group at the type locality, CON16002 (Rapela et al., 2016) was taken from a banded schist that crops out in or is in contact with the Santa Rosa Group migmatites; A25-01 (Steenken et al., 2006) corresponds to a fine grained biotite-muscovite schist from the San Martín Group and FSL 16004 (Rapela et al 2016) is another fine grained schist located 3 km north of A25-01 and corresponds to the same outcrop than sample A09-06 (Table 1). Datasets with associated errors higher than 10% were not included. Age data were filtered accepting values less than 10% discordant and common Pb<5%; $^{207}Pb/^{206}Pb$ ages are used for ages >800 Ma. The sample size and the strategy of sampling as well as other factors like the calculations, errors and standard exert a strong influence of the final calculated maximum depositional ages. Age pick (Gehrels et al., 2006) and detrital zircon spectra using Isoplot v 4.1 were used to calculate MDA. In any case no record of detrital zircon of Pampean ages is observed. The spectra of all the samples are quite similar with MDA based on Age Pick from 598 to 553 Ma being the San Martín Group schist those the older detrital ages (Fig. 15a, b) Detrital zircon age of the CMC rocks (Fig. 15a), exhibit a similar shape-spectra and relative proportions of inherited zircon ages. In the banded schist of the La Cocha Group there is a range of younger ages from 800 to 550, with the younger peaks at 600 to 570 Ma. A significant part of the pattern lies between 1200 and 900 Ma which is typical of Grenvillian ages. Isolated Palaeoproterozoic single ages are present. Rapela et al. (2016) reported small Ordovician peak at ca. 490 Ma obtained from some zircon rims but there is an interval from 553 Ma to 487 Ma with no detrital record in his results for the banded schist. It is important to mention that the above mentioned sample is located inside migmatites and that granitoids intruding the banded schist cover an interval from ca. 497 to 470 Ma. Spectra for the San Martin Group schist exhibit dominant peaks from 631 to 587 Ma and from 1060 to 920 Ma together with a possible input from sources at 780/820 Ma in sample FSL 16004, which also shows a single grain age ca. 2700 Ma as indicated by Rapela et al. (2016). As it can be observed (Fig 16 a, b) MDA based on age pick are older than 550 Ma. In consequence, these metaclastic rocks do not show Pampean detritus input, which suggests that the deposition predates the tectonic uplift and subsequent erosion of the Pampean orogen. Thermo-barometric data from Sierra de Comechingones document a minimum of 20 km of unroofing of the Pampean basement during the early Cambrian (530-510 Ma) (Rapela et al., 1998; Otamendi et al., 2004).

4.4..TIME CONSTRAINTS FOR THE METAMORPHIC EVOLUTION OF THE CMC

The metamorphic evolution of the Conlara Metamorphic Complex in an absolute chronological framework is still an open question. Maximum depositional ages of ca.580 Ma and Ordovician crustal derived granitoid intrusions on the already metamorphic Complex indicate an ample interval between sedimentation and granitoid intrusion. Available data so far indicates that the La Cocha Group preserves a Pampean? or older metamorphism as indicated by the Pb / Pb garnet age of 564 ± 21 Ma (Siegesmund et al., 2010). Peak metamorphic conditions have also been assigned also to the Famatinian orogeny (Whitmeyer and Simpson, 2004) based on loosely constrained U-Pb monazite age of ca. 480 Ma obtained for migmatite probably located in the Santa Rosa Group that is referred to as a personal communication without any published analytical information. Interestingly in the banded schist that is preserved as resisters or mega enclaves inside the migmatites of the Santa Rosa Group, Rapela et al. (2016) found rims of around 480 Ma in detrital zircons, an age that is coincident with the age of the Ordovician granitoids.

Specifically for the La Cocha Group the lower limit for the metamorphic peak was established at 497 ± 8 Ma (Steenken et al., 2005) based on the age of a leucogranitic facies of the El Peñon Granite which intrudes during D_3 with clear cut contacts with the La Cocha Group after the M_1 (Llaneza and Ortiz Suárez, 2000), or M_2 (Steenken et al., 2005) metamorphic climax. This upper limit is further reinforced by a U-Pb SHRIMP zircon age of 470 Ma ± 8 Ma of an amphibole tonalite facies of the Las Cañas Plutonic Complex (Morosini et al., 2019) and by the Rb/Sr whole rock age of 460.2 ± 39.4 Ma (MSWD: 4.04, initial ratio: 0.7075, Rb/Sr whole rock errorchron) for the La Tapera Granite (Fig 17 a,b) (López de Luchi and Cerredo, 2001), which also intrudes the La Cocha Group. The analysis of the structural evolution of the Conlara Metamorphic Complex was first considered by López de Luchi (1984, 1986), Llaneza and Ortíz Suárez (2000) and López de Luchi and Cerredo (2001). According to these studies, the banding of the banded schist-gneisses is parallel to S_2 an axial-plane foliation that results from D_2. Since the pervasively solid-state deformed granitoid intrusions (Fig. 17c) such as the 460.2±39.4 Ma La Tapera Granite (López de Luchi, 1986, Lopez de Luchi and Cerredo, 2001), the El Salado granodiorite and the 497 ± 8 Ma El Peñón Granite (Llaneza and Ortiz Suárez, 2000, Steenken et al., 2005) include folded xenoliths of the banded schist and were intruded during $D_{3,}$ their late Cambrian emplacement would mark the earliest stage of Famatinian compression within the Complex. Microstructures in El Peñon denote continuity from magmatic to high-temperature solid-state deformation, indicating the synkinematic emplacement of the pluton with respect to D_3. Moreover as the emplacement of the 470 ± 8 Ma Las Cañas Plutonic Complex is considered to have occurred during the exhumation of the Complex this D_3 event might also be responsible for this exhumation.

The cooling history of the Conlara Metamorphic Complex after the Famatinian magmatism is recorded by K-Ar ages between 440 and 420 Ma on dark and light micas (Steenken et al., 2008). Rb/Sr mica-WR isochrons indicate that cooling in the Conlara Metamorphic Complex below the 500°C isotherm occurred at about 439 Ma (Steenken et al., 2004).

Consequently, D_1 – D_2 history may have taken place during the Pampean Orogeny (see Steenken et al., 2005) which contrasts with the assumption that the entire evolution of the Sierra de San Luis could be related to post-Pampean events (Whitmeyer and Simpson 2004). D_3 would be constrained between 497-470 Ma whereas the younger D_4 (López de Luchi et al., 2008) will be bracketed between ca 450-420 Ma based on K-Ar muscovite ages.

If the peak metamorphism is Pampean or even older as would be suggested by the age calculated on the garnet of the banded schist of the La Cocha Group, the entire complex must have remained at temperatures above 500 ºC for about 100 Ma. On the contrary, if only one metamorphic event of Ordovician age has occurred, it remains difficult to reconcile the 570 Ma MDA of the protoliths of the metaclastic rocks with a ca. 480-470 Ma metamorphic event since the El Peñón Granite clearly intruded the already metamorphic La Cocha Group and exhibit metamorphic enclaves.

Based on the P-T calculations on the metaclastic rocks an almost isothermal (?) uplift of ca. 4 km is suggested for the interval Late Ediacaran/Early Cambrian-Early Ordovician from a peak pressure, calculated either 5.4 ± 0.5 kbar or 5.50 ± 1.20 kbar for D_2 for the metaclastic rocks to ca. 4.45 ± 1.1 kbar calculated for D_3 based on Al-in hornblende barometry from one facies of the Las Cañas Pluton.

If migmatization occurred in the interval 500-480 Ma as may be suggested by the zircon rims in the sample of the banded schist of Rapela et al. (2016), the age of the crustal derived granitoids and the regional comparisons (for example Larrovere et al., 2012), this event should have occurred at deeper crustal levels than those currently exposed by the La Cocha Group. Granite emplacement in this non migmatitic group that records ca 4 km uplift, was controlled by D_3 in the late Cambrian-early Ordovician. Thermo-barometric data from Sierra de Comechingones document a minimum of 20 km of unroofing of the Pampean basement during the early Cambrian (530 to 510 Ma) (Rapela et al., 1998; Otamendi et al., 2004).

4.5. **SIGNIFICANCE OF THE MONAZITE AGES**

The monazites are interpreted to have crystallised after a medium-grade regional metamorphic event. Sample P7 is a migmatite of the Santa Rosa Group taken in the south of the Renca Batholith has the highest monazite Y_2O_3 contents. When the monazite-xenotime thermometers which relate *X*HREE+Y to monazite crystallisation temperatures, are applied, maximal conditions of 600 - 670 °C (Heinrich et al., 1997) and 520 to 590 °C (Pyle et al., 2001) can be estimated for this sample. The sample 104 from the San Martin Group exhibits contents between those of the migmatite and the two samples, i.e. 19-09 and 65-04, of the banded schists of the La Cocha Group which show similar Y_2O_3 contents(Fig. 14b). For sample 65-04 with the lowest monazite Y_2O_3 and *X*HREE+Y, the monazite crystallization temperatures are considerably lower, at 500 to 560 °C (Heinrich et al., 1997) and 400 to 470 °C (Pyle et al., 2001).

In the study area, temperatures of ca 600 °C were assigned to the host metaclastic rocks of the 470 Ma Las Cañas Complex (Morosini et al., 2019) whereas in the interval between 448 ± 10 Ma to 439 ± 7 Ma the basement was at around 500 to 400 °C, as indicated by the K-Ar amphibole cooling age for one amphibolite and K-Ar muscovite cooling ages for pegmatites (Steenken et al., 2006, 2008). Younger muscovite cooling ages down to 416 ± 4 Ma in pegmatites emplaced in the southern sector of the Santa Rosa Group suggest that a temperature of more than 400 °C may have persisted even into the Silurian.

Monazite analyses yield isochron Th–U–Pb ages ranging from 446 to 418 Ma. A trend of decreasing Y_2O_3 with decreasing age is apparent (Fig. 13, 14). The oldest age of 446 ± 5 Ma corresponds to sample P7, located only 500 m south of the locality where a K-Ar amphibole age, of 439 ± 7 Ma was obtained. Therefore the monazite age of this sample may represent a re-equilibration of the monazite on the cooling path of the basement complex. The monazites in samples 19-09 from the La Cocha Group and 104 from the San Martín Group yielded isochron ages of 434 ± 12 Ma and 431 ± 8 Ma respectively. The low-Y_2O_3 monazites of the sample 65-04 displays an isochron at 418 ± 10 Ma and would likely mark a late stage of a monazite growth. On the other hand the observation of a high temperature monazite with apatite coronas exclusively in sample P7 suggests a retrogression path. These results indicate ages younger than the 497 to 470 Ma interval of intrusive magmatic activity and the younger record of zircon growth at suprasolidus conditions (480 Ma) for rims of a banded schist inside the migmatites of the Santa Rosa Group. Numerous studies indicate that age resetting in metamorphic monazite can be due to post-peak dissolution–reprecipitation (Taylor et al., 2014 and references therein) or heterogeneous annealing of the monazite crystal lattice (Skrzypek et al., 2018). The stability field of metamorphic monazite is controlled by bulk rock Ca and Al, and the allanite-monazite univariant equilibrium is shifted to lower temperatures when bulk rock Ca decreases (Spear 2010). The calculated assay of Ca is lowest in the sample P7 with the oldest monazite isochron and higher in the sample 65-04 with the youngest monazite isochron (Fig. 13). This would support the interpretation that the metamorphic monazite crystallized during decompression subsequent to Tmax along a clockwise P-T evolution with sample P7 giving a minimal for P max. In consequence the calculated CHIME ages would represent minimum estimates for the timing of peak metamorphism and are interpreted as cooling ages from the prograde high T peak metamorphism probably resulting from post-peak fluid assisted recrystallization.

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4.6. **REGIONAL CORRELATIONS OF THE CONLARA METAMORPHIC COMPLEX**

Regional correlations were based in lithologies, metamorphism, magmatism, crystallization, metamorphic or cooling ages and detrital zircon pattern. Considering all the metaclastic units of the Eastern Sierras Pampeanas, only the Conlara Metamorphic Complex in the Sierra de San Luis was correlated with the Puncoviscana Formation that is interpreted as a lower grade metamorphic equivalent of the widespread greenschist, amphibolite and granulite facies metasedimentary rocks, more intensely affected by the Pampean orogeny (Rapela et al., 1998, 2016; Steenken et al., 2006, 2008; Escayola et al., 2007; Drobe et al., 2009, 2011: Siegesmund et al., 2010; Casquet et al.,2018).

The Pringles Metamorphic Complex and the San Luis Formation, the two units immediately to the west of the CMC (Fig. 1), exhibit a younger sedimentary record (Steenken et al., 2006, 2008, Drobe et al., 2009, 2011, López de Luchi et al., 2018, Perón Orrillo et al., 2019 and references therein) with a Cambrian zircon population that reflects erosion from the early Cambrian rocks, which are widespread in the Sierras de Córdoba and Comechingones.

The metaclastic rocks of the Conlara Metamorphic Complex and probably more specifically the banded schist of the La Cocha Group were correlated based on lithological and deformational features with the Ancasti Formation, the Tuclame schist and finally with similar banded lower grade rocks of the Puncoviscana Formation (Steenken et al., 2006, Martino et al., 2009 and references therein)

In Figure 16 (a,b) selected pattern of detrital zircon ages from Sierras de Córdoba, Sierra de Ancasti and Sierra Brava are included. Data for the Tuclame Formation are not sufficient for the calculation of a normalized plot or age pick. Although detrital pattern for the Sierras de Córdoba metaclastic rocks are comparable to those of CMC, in our analysis we focus on the lithologies that are similar to the groups of the CMC from which zircon data is available. The ages (single direct evaporation ages) of individual zircon grains of the Tuclame Formation (Schwartz and Gromet, 2004) vary from ∼1900 to ∼600 Ma, with groupings between 750 to 550 Ma, around 850 Ma and between 1050 to 950 Ma which made them comparable to the La Cocha Group. Martino et al. (2009) considered that the banded schists of the Tuclame Formation are equivalent to the banded schist of the La Cocha group and calculated 557 ± 25 °C and 3.9 ± 1 kbar for the peak metamorphism. The calculated pressure is even within error lower than the one obtained by López de Luchi et al. (2008, 2009) and Morosini et al. (2019) which might imply higher thermal gradients for this section of banded schists.

Banded schists were also recognized and studied in the Quilmes, Ambato, Aconquija and Ancasti mountain ranges, where in most cases two metamorphic events with a likely climax of Famatinan age is described. Available age constraints for the Conlara Metamorphic Complex as well as detrital zircon patterns indicate a strong similarity with the Sierra de Ancasti units even in the regional distribution of the units with migmatites flanking banded schists.

The Sierra de Ancasti (Murra et al., 2011 and references therein) is made up by three main units: the eastern flank corresponds to the Sierra Brava Complex in which paragneiss and migmatite make up the Jumeal member and marble, schist and amphibolite make up the Calera member. The central sector is mostly formed by banded schist of the Ancasti Formation. In the western flank of the sierra, metasediments of the Ancasti Formation prograde to gneiss and migmatite of the El Portezuelo Metamorphic-Igneous Complex. The banded schists of the La Cocha Group can be correlated with the Ancasti Formation whereas the Santa Rosa and San Felipe migmatites might be correlated with the El Portezuelo Metamorphic-Igneous Complex.

The Ancasti Formation records a low-pressure M_{-1} metamorphism that was overprinted by a syn-deformational M_{-2} event of medium grade. The M_{-1} metamorphism was assigned to the Pampean orogen through a Rb-Sr mineral isochron of 524 ± 28 Ma (Knüver and Miller ,1981, Bachmann and Grauert, 1987). Marbles cropping out in the banded schists exhibit Ediacaran depositional age of 570-590 Ma based on the trend of $^{87}Sr/^{86}Sr$ in Neoproterozoic seawater (Murra et al., 2011). Ordovician I-type granitoids with crystallization ages of 471 ± 5 and 466 ± 5 Ma were emplaced in this unit (Dahlquist et al., 2012) and show metamorphic enclaves.

The El Portezuelo Metamorphic-Igneous Complex is composed of migmatites that evolved from amphibolite to granulite metamorphic facies, reaching thermal peak conditions of 670 °C and 5.3 kbar for garnet migmatites to 740°C and 4.7 kbar for cordierite migmatites (Larrovere et al., 2011). U-Pb geochronology on monazite grains within the leucosome records the time of migmatization between 477 and 470 Ma. In the Sierra de Quilmes, located north of Sierra de Ancasti, Buttner et al. (2005) indicated that the monazite and titanite U–Pb data constrain the metamorphic peak in migmatites and calc-silicate rocks at or slightly prior to ∼470 Ma. It is remarkable that migmatization in the El Portezuelo Complex and in Sierra de Quilmes yielded the same age as the main magmatism and actually, within errors, the same age as the magmatism of the CMC. In these metamorphic units granitoids intrude an already metamorphic host that is also preserved as rafts inside migmatites.

We suggest that the Conlara Metamorphic Complex might record two events an older late Ediacaran to Early Cambrian metamorphic event and an Ordovician partial melting high T/P event that at the present level of exposure is recorded as D_3 controlled granitoid magmatism with S3 related biotite in the La Cocha Group and probably as prograde sillimanite in gneisses and migmatites. Larrovere et al. (2011) proposed the existence of a regional mid-crustal high thermal zone during lower-medium Ordovician times that will be expressed as the migmatite terranes in Sierra de Ancasti, Sierra de Ambato and Sierra de Aconquija. We propose that the migmatites of the Conlara Metamorphic Complex could correspond to the same event that overprinted a former Pampean event preserved in the non migmatitic rocks.

5 <u>FINAL REMARKS</u>

The Conlara Metamorphic Complex, the easternmost complex of the Sierra de San Luis, is a key unit to understand the relationship between the late Proterozoic-Early Cambrian Pampean and the Upper Cambrian-Middle Ordovician Famatinian orogenies of the Eastern Sierras Pampeanas. is divided in four groups based on the dominant lithological associations: San Martin and La Cocha correspond mainly to schists and some gneisses and Santa Rosa and San Felipe encompass mainly gneisses, migmatites and orthogneisses. The Conlara Metamorphic Complex underwent a polyphase metamorphic evolution. The penetrative D_2-S_2 foliation was affected by upright, generally isoclinal, N-NE trending D_3 folds that control the NNE outcrop patterns of the different groups. An earlier, relic S_1 is preserved in microlithons. Discontinuous high-T shear zones within the schists and migmatites are related with D_4 whereas some fine-grained discontinuous shear bands attest for a D_5 deformation phase (Fig. 18). Geochemistry of both non-migmatitic metaclastic units and amphibolites suggest that the Conlara Metamorphic Complex represents an arc related basin in which MDA indicate a pre- 570 Ma deposition of the sediments (Steenken et al., 2006, Drobe et al., 2009). The 497-470 Ordovician magmatism indicates an ample interval between sedimentation and emplacement of granitoids in the already metamorphic complex as indicated by both clear cut contacts and D_1-D_2 deformed enclaves of metaclastic host. No reliable records of either Pampean aged detritus or magmatic ages belonging to the peak of the Pampean Orogeny were so far found. The garnet PbSL age of 564 ± 21 Ma for M_2 metamorphism of the banded schists of the CMC would so far suggest that D_1-D_2 history might have occurred at very early stages of the Pampean orogeny.

Pressure-temperature pseudosections were calculated for one amphibolite of the Santa Rosa group using the geologically realistic system MnNCKFMASHTO (MnO-Na_2O-CaO-K_2O-FeO-MgO-Al_2O_3-SiO_2-H_2O-TiO_2-Fe_2O_3). Peak metamophic conditions (M_2) indicate 6 kbar and 620 °C. The absence of clinopyroxene limits the maximun temperature. Late chlorite on the rims and in cracks of garnet, along with titanite rims on ilmenite and matrix plagioclase breaking down to albite suggests that the P-T path moved back down T.

D_3 is constrained by the syn-kinematic emplacement of the pervasively solid-state deformed granitoids like the Early Ordovician El Peñón and La Tapera plutons which bear xenoliths of the host banded schists, show clear cut contacts with their host and display a folded cartographic pattern with D_3 NE- trending axial planes. This evidence argues against the assumption that the entire metamorphic evolution of the Sierra de San Luis could be assigned to post-Pampean events.

D_4 deformation is bracketed between the 450-420 cooling ages of muscovite booklets of pegmatoids and amphibole age in one amphibolite.

Monazite analyses yield isochron Th–U–Pb ages ranging from 446 to 418 Ma. The oldest age of 446 ± 5 Ma correspond to a migmatite from the Santa Rosa Group. Monazites in samples from the La Cocha and the San Martin group crystallized at decreasing temperatures. The younger isochron age of 418 ± 10 Ma corresponds to low-Y_2O_3 monazites in one sample of the La Cocha Group. New CHIME monazite ages presented here likely represent post-peak fluid assisted recrystallisation that are similar to amphibole and muscovite cooling ages. Therefore the monazite ages may represent a re-equilibration of the monazite on the cooling path of the basement complex.

The restricted low greenschist facies D_5 overprint is comparable (both in strike and composition) to the mylonitic paragenesis characteristic of Río Guzmán Shear Zone, the Devonian age of which closely matches the regional biotite cooling ages of the CMC and would have been responsible for final unroofing of the complex (Fig. 18).

Regional comparisons suggest a strong similarity with the Ancasti Formation as well as with the El Portezuelo Fm even in the spatial dsitribution of the units. Further work on the metamorphism and age dating are necessary to accurately constrain the tectonic evolution of the CMC since the age of metamorphism and crustal-derived magmatism would indicate high temperature in the crust over a time span of almost 100 Ma

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CAPTIONS

Appendix 1. Analytical procedures

Supplementary Material 1 EMPA data on minerals from sample 21-3

Supplementary Material 2.Figure TXCa for sample 21-3

Supplementary Material 3 Figure TXFe3 for sample 21-3

Supplementary Material 4 **Online Resource.** Selected electron microprobe analyses of metamorphic monazite from biotite plagioclase paragneisses. Data are given in wt%. Th_{Suz} and ThO_2* are calculated from Th and U after Suzuki et al., (1994). Monazite ages from single analyses are given with 2σ error, see text. Mnz - monazite single grain; Data from reference monazite Madmon (Schulz and Schüssler 2013) is weighted average of 20 single analyses performed during the sessions on the samples. The monazites were analysed with a JEOL JXA-8230 at Institute of Material Science of the TU Bergakademie Freiberg/Saxony, Germany.

Table 1: Major and minor element contents of the metaclastic rocks of the Conlara Metamorphic Complex. Major elements are expressed in % weight, minor elements in ppm. Analytical techniques are presented in the text. Location of the samples are indicated.

Table 2: Major and minor element contents of selected amphibolites of the Conlara Metamorphic Complex. Major elements are expressed in % weight, minor elements in ppm. Analytical techniques are presented in the text. Location of the samples are indicated.

Table 3**:** Results from automated SEM Mineral Liberation Analysis (MLA) of petrographic thin sections of monazite-bearing biotite plagioclase paragneiss samples. Modal mineralogy of major phases, from XMOD point counting routine with ~10^5 EDX analyses per thin section. Ca and Al bulk rock compositions in wt%, as calculated assay from modal mineralogy. Area, number of grains and grainsize as MD50 (grainsize at 50 wt% of cumulative grainsize distribution) in µm for zircon (Zrn), monazite (Mnz) and xenotime (Xtm) from automated SEM Spare Phase Lineup(SPL) routine. The monazite grains are in contact (locked) with diverse other phases, as allanite and epidote (Aln + Epi), apatite (Ap), biotite (Biotite), ilmenite (Ilm), plagioclase (Pl) and quartz (Quartz). These microstructural relationships are reported as ternary locking in %. Note high degree of ternary locking of monazite with allanite and epidote in sample 65-04.

Figure 1 a) Geological map of de Conlara Metamorphic Complex includig the location of the monazite samples and the amphibolite sample; b) Geological sketch of the Eastern Sierras Pampeanas with the location of the study area.

Figure 2: Field photografies of the diagnostic features of the Conlara Metamorphic Complex groups: a) San Martin Group, eastern belt; note the relitic psammitic pelitic bedding at the upper left corner of the image and the folded pegmatite dike b) La Cocha Group inmediatelly east of La Cocha hamlet: note that apart from the banding a difuse relictic bedding is observed; c) Santa Rosa group inmediately west of San Pablo d) San Felipe Group south of the Renca Batholith.

Figure 3: Field images of the Conlara Metamorphic Complex groups: a) detail of the F2 folds that affect S2 foliation in the San Martin Group fine grained schist west of Las Lagunas; b) Detailed outcrop features of the La Cocha Group. Note the folding and the crenulation of the quartz-plagioclase veins as well as melt migration towards the hinges of the F_3 folds c) Detail of the texture of the Santa Rosa group migmatites d) Detail of diatexite of San Felipe Group south of the Renca Batholith. An amphibolite lense is observed.

Figure 4: Microscopic images of the San Martin Group: a) fine grained biotite schist. Flattening of the quartz with the major axis parallel to the S_2 foliation is present. A muscovite blast overgrowing the S_2 foliation is observed; b) biotite-muscovite fine grained schist that is affected by F_2 related crenulations; c) tourmaline schist, nematoblastic tourmaline-biotite layers alternate with quartz -scarce plagioclase layers;

d) fine grained biotite-muscovite schist interlayered with the banded schist in the La Cocha Group. Abbreviations from Whitney and Evans (2010).

Figure 5: Microscopic images of the La Cocha Group: a) spaced foliation in biotite muscovite banded schist S1 appears a relic folded continuous foliation whereas S_2 is parallel to the banding defimed by mica-rich and quartz domains; b) spaced folation defined by banding in the hinge of the F_2 folds in which a S_3 is developed paralell to the axial plane. In the limbs S_3 is parallel to S_2; c) muscovite blast overgrowing S_2/S_3.
Abbreviations from Whitney and Evans (2010).

Figure 6. Microscopic imagesof the Santa Rosa Group: a) F_{1+N} and F_{2+N} in gneiss b) prograde sillimanite in the mesosome of a stromatite c) folded fibrolite sillimantic layers inside a muscovite blast that overgrows the lepidoblastic biotite folia. This sample is located near the eastern contact of the central belt; d) microshears made up by recristallized quartz mosaic in a leucosome. Abbreviations from Whitney and Evans (2010).

Figure 7. Microscopic images of the San Felipe Group: a) texture of the biotite rich mesosome. Note the lobate borders between plagioclase and the muscovite blast; b) detail of the muscovite blast in which fibrolitic sillimanite is present.Note also some fibrolite inside the plagiocase in contact with the muscovite blast; c) biotite rich mesosome in stromatite Note the lobate borders between plagioclase and quartz; d) detail of microshear in the diatexite. Note the pressure shadows tails at both sides of the plagioclase blast. Tiny recristallized biotite is associated with the mosaics of polygonal quartz e) general texture of the orthogneiss. Discontinuous biotite-muscovite folia are parallel to flattened quartz aggregates. f) garnet in a plagioclase and alkaline feldspar aggregates. Lobate boundaries as well as grain boundary migration indicate high temperature deformation. Abbreviations from Whitney and Evans (2010).

Figure 8: Metaclastic units of the San Martin and La Cocha groups a) Chemical classification of metaclastic rocks (Herron 1988). Average compositions of magmatic rocks were taken from Albarede (2003); b, c source maturity: b) Chemical index of alteration (CIA, (Nesbitt and Young, 1982): [Al_2O_3m/ (Al_2O_3m+ CaOm* +Na2Om +K2Om)] x 100). The arrows represent ideal weathering trends of granite (right hand side), granodiorite (central arrow) and tonalite (left hand side). A nearly parallel development compared to the ideal weathering trend is observed for the whole data collection. Pl: Plagioclase, -kaolinite (Knl), gibbsite (Gib), chlorite (Chl), muscovite (Ms), Ksp: Potassium feldspar. c) Th/Sc vs. Zr/Sc diagram after McLennan *et al.*, (1993), Zr/Sc is a measure of upper crystal reworking whereas Th/Sc measures mafic vs felsic input. Numbers identify the mean values for 1, Ocean Island Arc (OIA), 2, Continental Island Arc (CIA), 3, Active Continental Margin (ACM) and 4, Passive Margin (PM) according to Bathia and Crook (1986); d, e, f, Geotectonic setting discriminant diagrams. d) TiO_2 vs. Fe_2O_3T + MgO (Bhatia 1983), e) Ti/Zr vs. La/Sc (Bhatia and Crook 1986); f)$(La/Yb)_N$ vs. Eu/Eu* diagram. Boxes correspond to the compositions reported for Mc Lennan et al., (1990) for sedimentary basins of known tectonic setting.

Figure 9: Amphibolite of the southern sector of the Conlara Metamorphic complex. a) Major element classification of Jensen (1976). Major elements expressed in cations, b) trace element based classification (Pearce 1996); d-c) Primitive mantle normalized (Sun and Donough 1995) multielement plots e) REE f) trace element. Plots are set at a similar y-scale to compare the enrichment regarding the primitive mantle model. MORB, MORB+BAB and BAB averages from Gale et al., 2013.

Figure 10: a-d) Major and Trace-elements-based tectonic setting discrimination plots a) MnO-TiO2-P2O5 diagram (Mullen 1983). Although the samples plot in the MORB field, amphibolies from CMC exhibit higher TiO2 than MORB or BAB; b) Nb-Zr-Y diagram Meschede (1986). Samples of the amphibolites are enriched in Nb regarding MORB and BAB which would represent the arc component; c) Zr-Ti-Sr plot (Pearce and Cann 1973Amphibolie samples plot in between the MORB and island arc tholeiite; d) V-Ti-Sc plot (Vermeesch 2006). Samples plot in the MORB field in a central position next to the limit between different fields. Note that four independent parameters and ratios suggest that the amphibolites of the CMC have a MORB signature combined with HFSE and LILE enrichment which suggest a subduction or crustal component.

Figure 11: a) Photomicrograph of sample 21_3 showing mineral relationships between garnet and the matrix; b) Photomicrograph showing the main mineral assemblage of the sample, the inset shows the

growth of a titanite halo on matrix ilmenite; c) Mineral chemical zoning of a garnet porphyroblast, garnet is 300 um in size.

Figure 12: Pseudosection: Final P-T diagram for sample 21-3 with a reduced CaO content from the original whole rock composition. Additional diagrams show garnet compositional isopleths for X_{Alm}, $X_{Grs,}$ X_{Sps} and X_{Pyr}. The X_{Pyr}, X_{Alm} and X_{Sps} composition found in the garnet is given as the red fields on the P-T pseudosection and the stability field of albite is the blue region on the P-T diagram. The interpreted P-T path is given by the dashed arrow.

Figure 13: (a)-(d) Th-U-Pb chemical model ages of monazite (Mnz) in biotite-plagioclase paragneisses. Total ThO_2* vs PbO (wt%) isochron diagrams. ThO_2* is ThO_2 + $_{UO2}$ 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 in Ma with MSWD and minimal error of 2σ are calculated from the single analyses belonging to an isochron according to Ludwig (2001). The symbols mark analyses belonging to a monazite age population and defining isochrons. Al and Ca are bulk rock compositions in wt% from Table 1.

Figure 14: Mineral chemistry of monazite and distributions of monazite Th-U-Pb chemical ages. (a) Diagram $XGdPO_4$ vs $XYPO_4$, with compositions above the garnet (Grt), staurolite (Sta), sillimanite (Sil) metamorphic mineral zones as indicated by Pyle et al. (2001). (b) Monazite Y_2O_3 vs age; note position of sample 65-04 with the youngest monazite isochron (c) Monazite LREE and HREE+Y compositions in mole fractions, plotting along a common trend (d) Th+U vs Ca diagram with a common cheralite substitution trend of the monazites in the paragneiss samples.

Figure 15: Detrital zircon provenance of the metaclastic rocks of the La Cocha and San Martin groups. Sample data were taken from Drobe et al. (2009), Rapela et al. (2016), Steenken et al. (2006): a) Probability Density Plots (Frequency vs age). Note the similarity in the detrital peak ages and shape of remnant peaks between each group despite the different number of zircons. Cryogenian age peak are higher in the La Cocha Group samples; b)Cumulative frequency proportion vs the upper crustal residence time for detrital zircons (crystallization minus depositional age), and tectonic setting of the basins as proposed by Cawood et al., (2005). Independently of the number of zircons that are insufficient for samples A73-05 and A25-01 the shape is similar and correspond to collisional related basins. Samples 16002 and A73-05 are banded schist from the La Cocha Group; 16004 and A25-01 belong to fine grained schists of the San Martín Group.

Figure 16: Comparison of the detrital zircon provenance of the metaclastic rocks of the La Cocha and San Martin groups of the Conlara Metamorphic Complex with some metaclastic units of Sierra de Córdoba, Sierra de Ancasti and Sierra Brava: a) Normalized zircon plots (Gehrels 2014). Note that although the dominant peaks are of Brasiliano and Grenvillian ages, the relative proportion of each zircon population changes. References : (1) Iannizzotto et al., 2013; (2) Escayola et al., 2007;(3) Rapela et al., 2016 ;(4) Drobe et al., 2009;(5) Steenken et al., 2006;(6) Rapela et al., 2007;b) Age picks (Gehrels et al., 2006). Ediacaran, Tonian and Stenian (Grenvillian ages) peaks are dominant whereas Cryogenian ages are abundant in the metaclastic units located along the eastern sector of the Eastern Sierras Pampeanas of Brasiliano and Grenvillian ages.

Figure 17: La Tapera granite, images show the relationship between D_3 and granitoid emplacement and textures: a) sharp contact between the granitoid and the banded schist of the La Cocha Group, note both one metamorphic enclave and the parallelism between the contact and the penetrative foliation in the granite; b) photomicrography of the same contact in which the parallelism between biotite flakes and the limit between the two units can be observed, note that the S_2 banding is sharply cut c) Photomicrography of submagmatic and subsolidus high-temperature microstructures, profuse fine-grained quartz-plagioclase mosaics composed of equant, strain-free grains and subgrains in plagioclase.

Figure 18: Summary of the tectono metamorphic evolution of the Conlara Metamorphic Complex. See text for explanations.

-The Conlara Metamorphic Complex, the easternmost complex of the Sierra de San Luis, is a key unit to understand the relationship between the late Proterozoic-Early Cambrian Pampean and the Upper Cambrian-Middle Ordovician Famatinian orogenies of the Eastern Sierras Pampeanas.

- The CMC is divided in four groups based on the dominant lithological associations: San Martin and La Cocha correspond mainly to schists and some gneisses and Santa Rosa and San Felipe encompass mainly paragneisses, migmatites and orthogneisses.

-The Conlara Metamoprphic Complex underwent a polyphase metamorphic evolution starting at the Neoproterozoic and culminating in the early Ordovician.

-Peak metamorphic conditions (M_2) indicate 6 kbar and 620 °C. Late chlorite on the rims and in cracks of garnet, along with titanite rims on ilmenite and matrix plagioclase breaking down to albite suggests that the P-T path moved back down.

- CHIME monazite analyses yield isochron Th–U–Pb ages ranging from 446 to 418 Ma which suggest post-peak fluid assisted recrystallisation and are similar to amphibole and muscovite cooling ages. Therefore the monazite ages may represent a re-equilibration of the monazite on the cooling path of the basement complex.