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Neogene Sedimentation and Tectonics in the Collón Curá Basin (Patagonian Andes of Argentina)

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Keywords

Collón Curá basin; Neogene Basin; $^{40}Ar/^{39}Ar$ dating; Intramontane basin; Patagonian Andes; plate kinematics; subduction.

ABSTRACT

We recall here recent developments concerning the filling and paleogeography of the Neogene Collón Curá basin in the foothill of the North Patagonian Andes. To better constrain its evolution, we bring two new $^{40}Ar/^{39}Ar$ on biotite datings of two ignimbrites that allow bracketing of the filling between 19.04 ± 0.72 and 5.06 ± 0.69 Ma. Like the other intramontane basins of the North Patagonian foothill, the Collón Curá basin initiated around 18 Ma at the time of the Quéchua phase. This corresponds in the area neither to an increase of the convergence rate of the Nazca and South America plates, nor to a significant change in the angle of convergence. We propose to look at this profound alteration of the regional stress regime in the light of the recent results of analog and numerical modeling of subduction. Folding of the slab above the mantle discontinuity at 660 km depth leads to the development of geometries that we retrieve in seismic tomography images in the region. Modeling shows that the overriding plate is coupled to this folding in depth. It suffers from changes in stress (from extensive to compressive stress regime) and topography. It is assumed that a more compressive stress regime will favor the reactivation of border faults of the basin and a regional uplift, hence agradation in basins. On the contrary, subsidence associated with a decrease

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in shortening will favor the backward erosion that will cannibalize these basins. The life duration of basins will be modulated by efficiency of earth surface processes proper to each watershed (catchment area, river slopes, lithology…). This internal cyclicity of subduction due to slab folding is freed from the kinematics of convergence. It is now necessary to look for the expression of this process in the sequential filling of the basins.

Introduction

As for the Neuquén Andes to the north (e.g. Branellec et al, 2015; Branellec et al, 2016), the Neogene reactivation of basement heterogeneities resulted in the North Patagonian Andes to the generation of an array of partitioned basins surrounded by basement highs (Figure 1; Mosquera and Ramos, 2006; García Morabito et al., 2011; García Morabito and Ramos, 2012; Huyghes et al., 2015a). Some of these basins have an Oligocene extensional history and they have been subsequently inverted during the Miocene (e.g. Ñirihuau basin, Paredes et al., 2009; Cura Mallín Basin, Folguera et al., 2006; Kilca-Aluminé Basin, Franzese et al., 2011; Auca Pan Basin, Ramos et al., 2015). The Collón Curá basin is one of these Miocene basins. It is a small basin (<5000 km^2) that currently contains no more than 350 m of Neogene sediments and these sediments are being excavated. It does not present the economic resources of its huge neighbor, the Neuquén basin. It is a fault-bounded intramontane basin integrated in the thrust belt (García Morabito et al., 2011) and formed behind an actively growing topography. Here is one of its geological interest, it is an important recorder of the Miocene deformation, erosion, and syntectonic deposition.

The stratigraphic framework is now laid out in broad outline as well as the main stages of its paleogeographic evolution (e.g. Huyghes et al., 2015a). We recall the main lines at the beginning of the paper. But recent discussions show that the chronological framework remains to be consolidated (Franzese et al., 2018). This is why we present here two new absolute dating of ignimbrites at the top and at the base of the outcropping Miocene sedimentary filling in order to better constrain the period of activity of the basin.

The Collón Curá basin is an intra-mountain basin above an active margin and tectonics has potentially played a determining role in its evolution. Significant progress has been made in recent years in understanding the dynamics of subduction when the slab is encountering the mantle discontinuity at 660 km in depth. Analog modeling first and then numerical approaches specified the folding of the slab above

this discontinuity. They characterized the durations of the process and evaluated how it forced shallow uplift and stress (e.g. Guillaume et al., 2009, Gibert et al., 2012, Cerpa et al., 2014, 2015). In parallel, seismic tomography imagery returned in the area geometries that resemble the modeled geometries (Pesicek et al., 2012). Slab break-off had already been invoked to comment the evolution of the basin (e.g. Ramos et al., 2014b). We propose to include it as an element of a broader scheme of analysis that could open (or not) a new way of interpretation of the regional dynamic.

We will therefore pursue three main goals here. Given the expectations of this issue, we will first focus on the Neogene infill and stratigraphic framework of the basin. We then consolidate the timing of the basin with two new dating of ignimbrites. At least, the final discussion is speculative. It proposes to extend interpretations already advanced on the link between subduction and the dynamics of the basin. One may wonder if this evolution is not partially decoupled in time from the global kinematics.

Regional Setting

We focus on an area located south of the Huincul Ridge that is a east-west inverted fault zone corresponding to a structural high formed after the Jurassic in the south of the Neuquén basin (Mosquera and Ramos, 2006; Grimaldi and Dorobek, 2011; Naipauer et al., 2012; Figure 1). This part can be divided into four main morpho-structural domains corresponding from west to east to (1) the Main Cordillera, (2) a series of several NNW-SSE oriented depocenters on the borders of the relief, (3) a nearly north-south continuous topographic high and (4) the Picun Leufu basin that corresponds to the main foreland.

The depocentres individualized east of the main cordillera correspond from north to south to the Kilca-Aluminé depocenter, the Auca Pan and Catán Lil depocenters and the Collón Curá basin, which is the most important one in the study area (Figure 1). The Kilca-Aluminé and Collón Curá depocenters have been interpreted as extensive depocenters during the Oligocene and lower Miocene and were filled at that time by volcanic and volcano-sedimentary deposits corresponding to the Auca Pan Formation in the Collón Curá Basin (Vergani et al., 1995; Franzese et al., 2011; García Morabito and Ramos, 2012). These basins have been inverted since the middle Miocene in response to the establishment of the Quechua compressive phase (Pardo-Casas & Molnar, 1987; Cobbold & Rossello, 2003; García Morabito et al., 2011; Huyghe et al., 2015a). Thus, due to this shift toward a compressive regime, these depocenters were isolated from the rest of the foreland by the inversion of Triassic normal faults (D'Elia et al., 2012) and the

uplift of basement blocks corresponding to the Sañico Massif, prolongated to the north by the Catán Lil anticline (Ramos et al., 2014b). Thus, the Collón Curá, KIlka and Catán Lil basins became – at least partially – endorheic during most of the Neogene and were filled by the erosional products of the main cordillera and by volcanic and volcano-sedimentary deposits (García Morabito et al., 2011; Huyghe et al., 2015a). In particular, the Collón Curá basin was partially isolated from the Picun Leufu basin from ~11 to 8 Ma due to the onset of the uplift of the Sañico massif and was probably fully endorheic until the Late Pliocene (~3 Ma) when the drainage network reconnected to the Picun Leufu basin (Huyghe et al., 2015a). After that, erosion predominated in the Collón Curá Basin and Huyghe et al. (2015b) suggested that compression was pushed to the east of the Sañico Massif near the Pampa Curacó Plateau during the Pliocene. Although Plio-Quaternary deformation processes have been documented in the Chihuidos and Anelo area farther north (Messager et al. 2010), a continuation of shortening up to the present day is still a matter of debate, particularly across the foreland of the North Patagonian Andes, where evidence of continuing shortening during the Quaternary is scarce and ambiguous.

The Neogene sedimentary filling is then unequally distributed in the study area. The Collón Curá basin with a maximum thickness of 500 m (García Morabito et al., 2011) is the main depocenter. The Picun Leufu area only preserves small relicts of Neogene syn-compressive deposits on topographic highs (Huyghe et al., 2015a).

North (the Cura Mallin and Agua Amarga basins) and south of the study area (the Nirihuau and Gastre basins) are the largest Neogene intermontane basins of the foothill and of the mountain interior (Figure 1). The Cura Mallin and the Nirihuau basins preserve above Oligocene extensional deposits, several kilometers of syn-synorogenic Neogene deposits (Bechis & Cristallini, 2005; Bechis et al., 2014 ; Paredes et al., 2009; Orts et al., 2012).

Miocene Depositional History of the Collón Curá Basin

At the bottom of the basin filling, the **Auca Pan Formation** is made up of andesitic rocks related to a magmatic event from 36 to 43°S that took place between 33 and 20 Ma in the retroarc zone of the foreland system (Figures 2 and 3; Turner, 1965, 1973; Rapela et al., 1983; Lagorio et al., 1998; Franzese et al., 2011; Ramos et al., 2014b). It was deposited in extensive depocenters at the feet of the main cordillera.

The **Collón Curá Formation.** The Collón Curá Fm is exposed along rivers of the eastern margin of the basin at the border of the Sañico Massif and along the Rio Limay in the southern part of the basin (Figure 2). The pioneering work by Roth (1899) identified the formation later named by Yrigoyen (1969). Based on its mammal fauna, Roth (1899) attributed it to Middle Miocene and Pascual and Odreman Rivas (1971) re-evaluated this attribution as Middle to Late Miocene.

Absolute ages came with K/Ar radiometric dating of pyroclastic deposits that yielded ages of 15.4±0.3 Myr and 14±0.3 Myr (Marshall et al., 1977) and 11±1 Myr (González Díaz & Nullo, 1980). At the base of the formation, the Pilcaniyeu Ignimbritic Member was dated at 15 Myr by Rabassa (1975) and 13.8±0.9 Myr by Mazzoni & Benvenuto (1990). Above it, Cazau et al. (1989) obtained an age between 11.5 and 10.7 Myr at the base of the clastic deposits.

In the Collón Curá basin, Giacosa et al. (2001) attributed to the Langhian and Serravalian the Collón Curá Fm. but the upper boundary is not consistently defined.

The Collón Curá Fm. is rarely exposed in the Collón Curá basin. To the north of the basin, it exhibits an angular unconformity with the underlying Paleogene Auca Pan Fm. The formation is thin, its basal part consists of a 20 m thick ignimbrite. This pyroclastic flow grades vertically into 5 m of sandstones with igneous rock pebbles. Sedimentary facies correspond to tuffaceous deposits, mostly reworked by fluvial processes, showing channels larger than 5 m, alternating with primary volcanosedimentary deposits (Figure 3, section C).

Caleufú Formation. The middle and upper Neogene Caleufú Formation consists of the Limay Chico and Alicura members (González Díaz et al., 1986) that comprises the vast majority of Neogene strata in the basin.

González Díaz et al. (1986) shed new light on the stratigraphic framework of the Collón Curá basin. They named Caleufú Fm the sediments overlying the tuffaceous material of the Collón Curá Fm that were formerly known as the Rio Negro Fm and the Alicura Fm (Dessanti, 1972; Nullo, 1979; González Díaz & Nullo, 1980). The Chimehuin Fm of Turner (1973) included the Collón Curá Fm and the grey sandstones (Galli, 1954) of the present Limay Chico Mb of González Díaz et al. (1986). The question was about the number of aggradational cycles recorded by these series. Invoking a single aggradational sequence from sedimentologic arguments, González Díaz et al. (1986) regrouped the Río Negro and Alicura Fm into a

single unit, the Caleufú Fm, which they divided into two members.

If Nullo (1979) ascribed the formation to the Middle Pliocene by analogy with the Río Negro Fm in its type locality, Nullo (1979) and González Díaz & Nullo (1980) nevertheless placed in the Pleistocene the Alicura Member. A partial chronologic framework is given by K/Ar dating of ignimbrites and basalt flows. Basalts at the base of the Limay Chico Mb yield an age of 20±4 Myr and two ignimbrites in the middle of the formation, ages of 14±1 Myr (González Díaz et al., 1990). The top of the Limay Chico Mb is given by another ignimbrite dated at 8±2 Myr (González Díaz et al., 1990). The 3±0.5 Myr Tipilihuqué basalt flow that covers the Alicura Member, gives the upper bound of the Caleufu Fm (González Díaz et al., 1990). But this framework remains to be consolidated especially for its basal part where these ages overlap those obtained for the Collón Curá Fm (e.g. Rabassa, 1975; Mazzoni & Benvenuto, 1990; or Cazau et al., 1989). Another question is whether or not the ignimbrite at the top of the Limay Chico Mb (8±2 Myr; González Díaz et al.,1990) corresponds to the Chimehuin Ignimbrite dated by Mazzoni & Rapela (1991). Moreover the age of 3±0.5 Myr obtained by González Díaz et al. (1990) does not strictly date the end of the agradation of the Alicura member. In the Las Coloradas basin which is located northeast of the Collón Curá basin, Franzese et al. (2018) dated to 8.5 Myr ($^{39}Ar/^{40}Ar$ age) the Tipilihuque basalt that toplaps the Miocene infill of this basin. In the field the lava flow does not overlap the Alicura member but its southernmost remnants are very close to the basin closure. So this arrangement raises questions about the actual age of the end of agradation in the Collón Curá basin. We will come back to this point later on in the paper.

The Caleufú Fm consists of siliciclastic and volcanosedimentary material. Dominant lithologies correspond to clay, sand and coarse sands in the lower part of the formation and grades up in the upper part where coarse conglomerates become more prevalent (Figure 3).

The Limay Chico Mb usually shows grey sandstones, tuffaceous deposits, conglomerates and claystones. Centimetre- to decimetre-sized clasts derived from Cenozoic and Mesozoic volcanic sources and from Paleozoic plutonic rocks. Pumice fragments with a greyish colour are abundant and frequent cross-bedding indicates a fluvial depositional environment. Laterally, the Limay Chico Mb. is made up of grey sandy pumice-rich fluvial deposits with the occurrence of red pebbles of igneous rocks, frequently at the base of fining-upward sequences.

Above the Limay Chico Mb in the middle of the basin, the Alicura Mb is composed of conglomerates with

sandstones and tuffaceous intercalations. Nature of clasts remains unchanged as well as lithologies. The transition with the Limay Chico Mb is difficult to locate (González Díaz et al., 1986). Red pebbles of igneous rocks are locally observed along the western border of the basin, frequently at the base of fining-upward sequences (Huyghe et al., 2015a).

A spectacular 130 m thick white ignimbrite is intercalated within the Caleufú Fm at the north east of the basin (Figure 3, section A). This unit is completely homogeneous and shows abundant quartz and pumice content. Another thinner ignimbrites (only a few meter thick) have been preserved in the basin particularly at the top of the series.

Pampa Curacó Formation

These facies are interpreted to have been deposited in fluvial environments, with some lacustrine events (Escosteguy & Franchi, 2010) and several paleogeographic settings have been proposed. It would record the growth of the orogenic front (González Díaz et al., 1986) or could have a fluvio- glacial origin (Gracia, 1958; Galli, 1969; González Díaz & Nullo, 1980). Alternatively, Flint & Fidalgo (1968) postulated that it corresponds to a pediment.

Time Bracketing of the Collón Curá Infill

The sedimentary filling of the basin where it can be observed, rests on different formations which extend from the Triassic in the south to the Oligocene elsewhere and even on the crystalline basement. This lack of stratigraphic concordance does not allow an easy assignment of the age of the basin. The authors have therefore endeavored to characterize their faunistic content (see above). Dating of tephra in conjunction with stratigraphic relations had been performed to better constrain the ages of geomorphic units.

These dates did not contradict the first assignments but they nevertheless raise in the current context some difficulties. The first is documentary. Most often published in symposium proceedings, they are hardly accessible for the time being to the international community. In addition, these often old datings do not always allow the reader to locate accurately the dated samples geographically and stratigraphically.

Authors implemented two methods, K-Ar ou $^{40}Ar/^{39}Ar$, most often (it seems to us) on whole-rock samples (e.g. Marshall et al., 1977; Mazzoni & Benvenuto, 1990). The current reader cannot evaluate if former workers neglected or not uncertainty in nucleogenic interference corrections in defining a plateau. This is

strictly valid only when a sample is compositionally homogenous (Renne, 2000). This is frequently untrue in the case of whole rock basalts and therefore interference corrections should be included in the definition of the plateau. Pyroclastic rocks too should be always suspected of containing xenocrystic contaminents and should never be analyzed as a whole-rock sample. Highly vesicular lavas, scorias, and pumices are problematic because their high surface to volume ratio causes excessive adsorption of atmospheric Ar which may overhelm $^{40}Ar^*$. So whole-rock K-Ar dates are increasingly suspected of being systematically too young (e.g. Mc Dougall et al., 1992; Renne et al., 1993).

We chose to date individual crystals of biotite sampled in ignimbrites of the Collón Curá and Caleufú formations. One of the most profound advantages of this approach is that pyroclastic material such as crystal-bearing tuffs can be dated without danger of biais from xenocrystaic contamination (e.g. Curtis, 1966; Lo Bello et al., 1987; Deino et al, 1990). Moreover $^{40}Ar/^{39}Ar$ incremental heating analysis of biotite presents significant advantages because the proportion of atmospheric 40Ar is typically reduced in higher temperature steps. For volcanic biotites, a $^{40}Ar/^{39}Ar$ plateau is generally a sufficient criterion for reliability.

So, we sampled two thick ignimbrites located at the base of the Collón Curá Formation (Figure 3; base of section C of Huyghe et al., 2015a; S 40°10.27', W 70°39.39') and at the top of the Alicura Member of the Caleufú Formation (top of section B of Huyghe et al., 2015a; S 40°07.121', W 70°55.212' location in Figure 2). We sampled 5 kg of these ignimbrites riches in pumices and biotites. Samples were taken at ~ 1 m depth to avoid potential surface alteration of the biotites. The plateau date has been calculated as the inverse variance weighted mean of individual plateau step ages. The uncertainty derived from error propagation (e.g. Renne et al, 1996).

The basal ignimbrite in the Collón Curá Formation yields an age of 19.04 ±0.72 Ma and the top one in the Caleufú Formation, an age of 5.06 ±0.69 Ma (Figures 4 and 5 and Table 1). Compared to the stratigraphic synthesis these ages provide new direct constraints of the history of the CCB. First, it moves the base of the Collón Curá Formation and the top of the Auca Pan Formation from 16 to 19 Ma compared to the stratigraphic framework proposed by Huyghe et al. (2015a) (Figure 4). The age boundary between these two formations was not accurately dated and this new age brings strong constrains about its age.

The age of the top of the sedimentary filling of the basin is subject to several controversies (see the synthesis of Huyghe et al., 2015a) and ranges between 8 and 3 Ma according to the authors and the dating method (Nullo, 1979; González Díaz and Nullo, 1980; González Díaz et al., 1990; Mazzoni & Rapela, 1991). Here, the absolute dating of the ignimbrite allows a time constrain of 5 Ma for the end of the filling of the basin (Figures 3 and 6).

These ages remain consistent with stratigraphic contains.

Tectonic and Stratigraphic Evolution

Huyghe et al. (2015a) considered four stages of paleogeographic evolution during the Neogene in this area where shortening is supposed to be low (<10 km, Ramos et al., 2014a).

Prior to the Neogene filling of the basin, Eocene and Oligocene intervals are interpreted as a phase of major erosion since Neogene deposits preserved to the east of the Sanico Massif often lie over Cretaceous or Jurassic sediments with a reduced angular discordance (<5°) and seem to fill ancient paleovalleys (Leanza et al., 1997; Hugo & Leanza, 2001). This evolution is characteristic of retro-arc foreland basins (Jordan, 1995).

At the end of the Oligocene or at the beginning of the Miocene, sedimentation began along a nearly N–S-oriented axis of maximum thickness deposition parallel to the Sañico Massif. The Collón Curá basin is supposed to have been individualized since the end of the Early Oligocene (Ramos et al., 2014a) but in the field, the oldest deposits are Early Miocene in age and are localized alteration products resulting from erosion of the Oligocene volcanic Auca Pan Fm. (Figure 6A; Huyghe et al., 2015a). The North Patagonian Massif (NPM) was probably not yet a prominent structure at that time.

In the Middle Miocene, a major volcanic phase supplied a large volume of volcaniclastic material to the Collón Curá Fm. (Galli, 1969) that extended in the basin and in a wide surrounding area to the south, to the east and up to the north of the NPM (Figure 6B). In the Picun Leufu basin, the same sedimentary material displays channelized structures characteristic of a single fluvial system showing that the Sañico Massif was probably not yet exhumed (Huyghe et al., 2015a).

Major change occurred during the Late Miocene and Early Pliocene. Progressive unconformities, with a maximum dip of 13°, are observed within the Caleufú Fm. along the western margin of the Sañico Massif in the central part of the Collón Curá basin (García Morabito et al., 2011). Many liquefaction structures,

interpreted as seismically induced and synsedimentary normal faults in the Limay Chico Mb. also show an intensification of tectonic activity (Huyghe et al., 2015a). In the same time, the Picun Leufu basin to the east is rather in non-deposition and possible erosion. This segmentation of the deposits is a consequence of the uplift of the Sañico Massif (García Morabito et al., 2011; Ramos et al., 2014a).

Thus, the Late Miocene Early Pliocene deposition is confined to an area of ca. 150 km by ca. 30 km in the Collón Curá basin with a thickness of tens of meters east of the Sañico Massif while sedimentary thickness reaches only few meters in the Renteria Plateau area to the east (Figure 2). Its connection with the rest of the foreland remains questionable at this time; the Collón Curá basin might have been isolated from the rest of the foreland basin. The deposition of the Limay Chico Member of the Caleufú Fm. in the Picun Leufu basin along the eastern border of the Sañico Massif suggests that it was rising and was eroded by rivers flowing to the east. Nevertheless, given the reduced volume of relicts of the formation in the Picun Leufu basin, if gateways existed during the filling of the Collón Curá basin, they were probably not active during long periods. This observation shows that the uplift of the Sañico Massif was faster than the filling of the Collón Curá basin, at least until the end of the deposition of the Caleufú Fm.

The deposition of the Alicura Mb. of the Caleufú Fm. during the Early Pliocene exhibits a coarsening compared to the Limay Chico Mb. This may reflect the end of the major filling of the Collón Curá basin (González Díaz et al., 1986) that was endoreic at this time (Huyghe et al., 2015a).

The Collón Curá basin underwent erosion during Late Pliocene and Early Pleistocene and the main depositional area moved to the east. It is questionable whether regressive erosion began to affect the Sañico Massif at this time and subsequently the Collón Curá basin, or if the activity of the surface processes in the basin allowed its overflowing towards the Picun Leufu basin. There is a lack of any sedimentological or morphological evidence to favor one or the other of these scenarios.

Discussion

Timing of the basin in the foreland setting

How the aggradational scheme of the basin highlights the mountain building processes in the area? The question spans the temporal space when considering whether shortening prevails over a single prolonged interval (~50–100 Myr) or over a series of punctuated events of shorter duration (<5–10 Myr) alternating with episodes of neutral or extensional modes (e.g. Steinmann, 1929). It spans the spatial space when

questioning if individual pulses are restricted or not to particular segments of the Andean margin (e.g. Groeber, 1951). It is why we choose to discuss the evolution of the Collón Curá basin compared to that of its close neighbors, the Catán Lil and Agua Amarga to the north and the Ñirihuau–Collón Curá basin to the south. These basins are the outermost Miocene intramountaine basins of the considered area.

Between 39°28'-39°48' S and 70°29'-70°44' W, the Catán Lil basin is a small intermontane basin (40x20 km) that accumulated 400 m of sediments during the Miocene (García Morabito et al., 2011; Figure 1). Developed between two thick-skinned anticlines, it nests above a Paleozoic crystalline basement where a Mesozoic continental–marine sedimentary succession forms the southermost part of the Neuquén Basin (Figure 2). Farther to the north, the Agua Amarga basin also developed above the Mesozoic infill of the Neuquén basin but to the north of the Huincul ridge, a strong inherited transverse structure that delimits the southern limit of the Neuquén basin (Figure 1). The Agua Amarga basin is larger (80x20 km) but not thicker than the Collón Curá basin and is located at the toe of the thin-skinned domain of the Agrio fold-and-thrust-belt and west of the thick-skinned Chihuidos anticlines. Compared to the northernmost basins, the Ñirihuau–Collón Curá basin at south of the Collón Curá basin, is larger (200 x 40 km; Figure 1). With more than 2000 m of Late Oligocene to Middle Miocene sediments deposited on top of the Oligocene volcanic rocks of the Ventana Formation (Rapela et al., 1988) and the igneous-metamorphic rocks of the Cushamen Formation (Volkheimer, 1964), it is also thicker (Spalletti, 1981; González Bonorino & González Bonorino, 1978). It has been alternately interpreted as a pull-apart basin (Spalletti & Dalla Salda, 1996), as an extensional basin in a back-arc setting (Mancini & Serna, 1989; Cazau et al., 1989) and as a foreland basin (Ramos & Cortés, 1984; Giacosa & Heredia, 1999 ; Bechis et al., 2014).

Did shortening prevail over a single prolonged interval or over a series of punctuated events of shorter duration? This makes necessary to set the chronological framework of these basins with the reserve imposed by the partial knowledge that we have of these basins. We dated two ignimbrites at the base and top of the sedimentary filling of the Collón Curá basin. At the bottom of the infill outcropping to the heart of the basin at the base of the Collón Curá Formation, we obtained an age of 19.4 ± 0.72 Ma (Figure 5). This dating allows revising the former ages by Rabassa (1975), Marshall et al. (1977) and Mazzoni & Benvenuto (1990) that yielded younger ages comprised between 13 and 16 Ma. So subsidence of the basin occurred a little bit earlier than the age given by Huyghe et al. (2015a). But this age does not correspond to the onset of sedimentation in the basin because the base of the basin is still buried.

The bottom of the Catan Lil basin is comprised between 16.1 and 13.8 Ma (Figure 7; Marshall et al., 1977; Rabassa, 1978; Rapela et al., 1988; Mazzoni & Benvenuto, 1990; Orts et al., 2012; Bilmes et al., 2013, 2014; Ramos et al., 2014a, 2015). The bottom of the sedimentary infill outcrops in the Agua Amarga basin. It is the Puesto Burgos Formation (Hugo & Leanza, 2001) late Oligocene-early Miocene in age (Figure 7). The sedimentary history began in the Ñirihuau basin at the same time with deposition of the marine Ñirihuau Formation (Roth, 1922 ; González Bonorino, 1973) that opened towards the Pacific (Figure 7). Cazau et al. (2005) and Paredes et al. (2009) revised the stratigraphic frame of the formation. More important here is the transition from lacustrine to fluvial deposition that occurred at the top of the Ñirihuau formation around 17 Ma and signed the onset of tectonic inversion (Paredes et al., 2009). So the onset of these basins as intra-mountainous basins occurred in a relatively narrow time bracket centered around 18 Ma (± 1 Ma).

Sampled at the top of the Collón Curá Formation, the upper ignimbrite gives us the end of aggradation at about ~5 Ma (Figure 5). On the basis of the bibliographic data, Huyghe et al. (2015a) previously assigned it to a little over 3 Ma. In the close Catán Lil basin, Franzese et al. (2018) dated plagioclases ($^{39}Ar/^{40}Ar$) of a lava flow of the basaltic plateau at top of the sedimentary infill that was assigned to the Plio-Quaternary (Cucchi & Leanza, 2005; Rojas Vera et al., 2014). They obtained an older age of ~9 Ma, in agreement with the $^{39}Ar/^{40}Ar$ age of the upper basaltic unit in the Kilca-Aluminé basin (Franzese et al., 2011, Figure 7). So aggradation in the basin ended earlier than expected and this allows these authors to postulate that shortening did not continue in the area after the end of the Miocene. Nevertheless our dating in the Collón Curá basin shows that volcanic activity continued until the Pliocene what supports the interpretations by Ramos et al. (2014a) and Huyghe et al. (2014, 2015). Franzese et al. (2018) did not sample the youngest flow of the basaltic plateau. Farther north in the Agua Amarga basin, the Pampa Encima Formation constitutes the last interval of aggradation before a drop of the base level of the Rio Neuquén and Rio Agrio (Hugo & Leanza , 2001; Messager et al., 2010). It was deposited during the Pliocene (Leanza and Hugo, 2001). Similarly to the Collón Curá basin, the final stage of the Ñirihuau basin contains pyroclastic and volcanoclastic materials reworked by fluvial and aeolian processes until Early-Late Miocene (Langhiane–Tortoniane; Giacosa et al., 2001). Comprised between ~10 and ~5 Ma, the time window spanning the onset of the excavation of these basins is larger.

The Miocene history of these basins is that of intra-mountainous basins. On the basis of bibliographic data,

the aggradation in these basins located between 37 and 40 ° S over more than 300 km began on a rather narrow window of time centered around 18 Ma. It coincides to the west with the closure of the Traigen oceanic basin between 20 and 15 Ma (Folguera et al., 2018). These features (timing and geographic distribution) suggest that we entered at this time under dynamic conditions that allowed the segmentation of the foreland by localized uplifts, it is the Quechua phase of the authors (e.g. Steinmann et al., 1929).

The end of the aggradation in these basins spreads over a longer period of 5 Ma. Field observations show active crustal folding of the Chihuidos and Sauzal folds east of Agua Amarga basin (Messager et al., 2010) and along the eastern border of the San Rafael block (northern Neuquén Basin) which is inverting since the Late Miocene (Costa et al., 2006, Folguera et al., 2018; Branellec et al, 2016). On the basis of these observations, the Quechua phase would continue untill the late Pleistocene at north of the Collón Curá basin.

The development along the Neuquén River of an alluvial fan at the outlet of the Agua Amarga basin shows that reconnection of the basin with its foreland has been achieved by backward erosion which then initiated the excavation of the basin. If backward erosion is of course controlled by the activity of the tectonic dam, it is also modulated by the intensity of surface processes determined by external forcing (climate, lithology, watershed geometry; e.g. Hilley and Strecker, 2005). These factors could therefore spread the timing of the basin excavation over time according to local characteristics.

However, Huyghe et al. (2015a) showed that the Collón Curá basin was still subsiding when the foreland, behind the tectonic dam, was already excavating. The vergence of the structure that initially isolated the basin from its foreland is directed towards the basin. Its lithostatic load could drive a flexural subsidence allowing for a time the activity of the basin.

The timing of initiation of these basins suggests that shortening controlled it at first order. The end of aggradation spreads on a broader time window probably as the result of a stack of regional and local forcings.

Kinematic forcing on basin evolution

Although the basins are distant from each other by several hundred kilometers, aggradation began in a narrow time window (18+/-1 Ma). This suggests that a lithospheric forcing probably controlled it and we examine here if the Atlantic spreading or if the conditions of the Pacific subduction can be linked to

shortening.

The growth of mid-oceanic ridges makes gravitational forces that could force the young and raised oceanic lithosphere to slide down the weaker asthenosphere and push on lithospheric material farther from the ridges (Meijer and Wortel, 1992). That is why the mid-Atlantic oceanic spreading could be a potential tectonic forcing.

The geological marker does not show a lot of demonstrative evidence of the Peruvian phase of shortening in the Collón Curá area because Cretaceous deposits are now eroded. But in the southern Neuquén basin, the older macroscopic record of inversion we observed in the field is the discordance of the Huitrìn Formation above the folded Avilé sandstones west of Chos Malal. This shows that the onset of inversion was Aptian in age. This is confirmed elsewhere in the basin by other observation or by seismic data (e.g. Cobbold and Rosello, 2003; Grimaldi and Dorobek, 2011) and now by thermochronology (e.g. Sánchez et al., 2017). This discontinuity follows shortly the onset of South Atlantic spreading at this latitude. Basing on the age of the oldest magnetic record along the Atlantic margin of South America (e.g. 133 Ma at 39°S and 120 Ma at 20°S; Meyer et al., 2017), kinematics models show that the Atlantic opening propagated northward (Setton et al., 2012; Müller et al., 2018). It is especially interesting to note that the onset of shortening in the Andes follows too this trend during the Cretaceous times: it begun at 125 Ma in the southern Neuquén basin and only at 113 Ma in Peru (10°S; Margirier et al., 2017). So the mid-Atlantic opening could play an operating forcing. The late Cretaceous inversion in Neuquén basin would be unrelated for Chen et al. (2019) to the westward motion of South America because the magnitude of displacement would be too low. But oceanic-opening and tectonic-shortening coincide in time and in space.

The correlation does not work in the Miocene time. Indeed aggradation in the basins (i.e. shortening, 18 Ma) occurred when the mid-Atlantic spreading rate was steady at 37 and 40°S (Figure 8). For the Pacific Ocean, the spreading rate decreased during this interval. It is not so surprising because ridge push is mostly active in lithosphere younger than 90 Ma, after which it has cooled enough to reach thermal equilibrium with older material and the slope of the lithosphere-asthenosphere boundary becomes effectively zero.

The tectonic regime changes may be associated with other parameters as plates velocities, plates

directions, the absolute movement of the trench, the age of the oceanic plate, the presence of sediments or the geometry of the oceanic plate (e.g., Ruff and Kanamori, 1983; Heuret et al., 2007; Schellart, 2008) that is to say to the dynamics of subduction.

The westward shift of South America could have forced the inversion of the basin. Plate-tectonic reconstructions show that South America did not stop migrating westward at the first order since the opening of the South Atlantic Ocean (Setton et al., 2012; Müller et al., 2018; Gurnis et al., 2018). For the period of time covered by the Collón Curá basin (0-18 Ma), is a change in the conditions of convergence likely to initiate the onset of the Quechua phase? For the considered time window, the South America plate continuously moved to the northwest and convergence angle between the Nazca and South America plates did not change more than 25° (Figure 9 and 10). The onset of aggradation in the basins is coeval with a decrease of the convergence rate between plates (Figure 10). It seems more "physical" to link the onset of shortening with an increase of the convergence rate. Also it could be more reasonable to interpret this inflexion in the convergence rate not as the cause of the occurrence of the Quechua episode but only as an evidence of the increase of the coupling between both plates.

The inversion that occurred at 40°S without strong kinematic change, leads to look for another explanations. The effect of the slab anchoring on the lower mantle interface and more generally the modality of the penetration of the slab into the mantle above it, remains an under-explored way of interpretation. The mantle retains a memory of subducted slabs, which body-wave tomography images in situ as domains of faster-than-average seismic wave velocities. The rupture with at least 5 m of slip of the M8.8 Maule earthquake in Chile in 2010 spreaded from 34 to 38°S (Lorito et al., 2011) and made it possible to image it for the latitudes we are interested in (Figure 11; Pesicek et al., 2012). The first order change in the slab geometry is located around 35°S with the transition from a horizontal Pampean flat slab (at north) to a southern slab with a moderate dip of ~30°. At the latitudes of our study area, this image does not show at the first order major differences between the north (section at 37°S) and south (section at 39°S; Figure 11A). The slab normally dips up to 200 km. The anomaly then bends towards the east and comes to horizontalization at a depth of 650 km. The slab, however, appears less continuous at 39°S and Pesicek et al. (2012) and Ramos et al. (2014) interpreted it as a result of slab tearing and detachment at ~200 km in depth.

Another fast anomaly stands almost vertically below the subduction trench and reaches depths greater than 1000 km (Figure 11A). This anomaly could also be interpreted as the remnant of a former slab diving in the mantle as observed along the Pacific margin of North (e.g. Sigloch & Mihalynuk, 2013; Sigloch & Mihalynuk, 2017) and South America (e.g. Pesicek et al., 2012; Chen et al., 2019). But its position may surprise. Since the South American plate shifted continuously westwards since 125 Ma (Figure 9; e.g. Seton et al., 2012), the trench shifted westwards too. So, we would be inclined to find the relics of the ancient slabs further to the east below the American plate as observed farther to the north by Chen et al. (2019). And the MITP08 tomography model too shows anomalies of seismic rates farter to the east that could be interpreted as remnants of former slab segments (Li et al., 2008; Hosseini et al., 2018).

In detail, the image of Pesicek et al. (2012) at 37 ° S is emblematic (Figure 11A). The Wadati-Benioff plane is associated with an anomaly of fast speeds which continues in depth up to 400 km. More than 200 km away from the trench, the slab then stumbles on a thick anomaly of 350 km which continues until a depth of 650 km . This one could correspond to a pile made by a slab folding on itself above the rheological discontinuity of the mantle (Figure 11B and 13; e.g. Guillaume et al., 2009; Gibert et al., 2012; Sigloch & Mihalynuk, 2013). The deep anomaly below the current slab shows an upward concave shape that must be included in this pattern (Figure 11A). We propose at 39°S a scheme that involves first the shallow rupture of the slab above 200 km in depth as invoked by Pesicek et al. (2012) and by Ramos et al. (2014b)(Figure 12B). The tomographic image shows another deeper anomaly in the lower mantle that requires a second rupture at the top of the pile in response to the pull of the slab diving into the lower mantle (Figure 12C). At 39 ° S, the image of Pesicek et al. (2012) does not necessarily show this second break up.

Numerical modeling shows that the way the slab deforms in depth depends strongly on the far-field boundary conditions i.e. on the individual values of the subducting and overriding plate velocities (Gibert et al., 2012; Cerpa et al., 2015). Two different styles of subduction are obtained once the subducting plate reaches the 660 km discontinuity. If the overriding plate moves toward the trench or not, the slab respectively lies forwards (case 1) or backwards (case 2) on the 660 km discontinuity (Figure 13). At the first order, the South America plate always shifted to the west (i.e. to the trench) since the Atlantic opening (e.g. Setton et al., 2012). But the absolute subduction velocity of the Nazca plate had always been greater (e.g. Pardo-Casas & Molnar, 1987; Müller et al., 2018). This ratio of subducting and overriding

plate velocities is compatible with a folding of the slab above the 650 km mantle discontinuity as explored by Gibert et al. (2012). And when folding occurs, the dip of the shallow slab changes.

Transient shallowing and steepening of the subducting slab has exerted an important influence on orogenic patterns (e.g. Ramos et al., 2002, 2014; Haschke et al., 2002, 2006; Kay et al., 2006; Kay & Coira, 2009; Ramos & Folguera, 2009; Chen et al., 2019) and might force the evolution of basins. Indeed, numerical modeling studies showed that the state of stress of the overriding plate is directly linked to the absolute velocity of the upper plate and the dynamic of the lithospheric mantle but also to the dip of the slab: upper plate shortening when the angle of the slab shallows, and extension when it steepens (Gibert et al., 2012; Cerpa et al, 2014). Thus a cyclic stress regime in the overriding plate due to slab folding in depth can be achieved with constant boundary conditions. Topography is also linked to the slab dip angle: when the slab has a shallow dip angle, the upper plate shortens and the topography is positive. In turn, when the slab dip angle steepens, the upper plate undergoes extension or less shortening and the topography becomes negative. Numerical modeling predicts for the far-field plate kinematics setting of the study area that these cycles will span about 20 Ma, depending of the mantle viscosity (Gibert et al., 2012 ; Cerpa et al., 2014; 2015) with alternating pulses of compression, of lesser compression or tension and episodes of regional uplift .

The seismic tomography shows also a vertical offset of about 200 km between both segments of the slab where the first rupture of our scheme occurred (Figure 11 and 12; Pesicek et al., 2012). This break-off might cause the basins to invert throughout an upward rebound of the remaining slab (Haschke et al., 2006; Ramos et al;, 2014). The authors do not yet agree on values of sinking rate of slabs in the mantle. But the interval of proposed values is narrow, comprised between 10 +/- 2 mm/yr (Sigloch & Mihalynuk, 2017) and 13 +/- 3 mm/yr (Butterworth et al., 2014). The distance between the two segments of the slab (i.e. 200 km) could yield a proxy of the onset of the slab break-off. We obtain here a value of 20 Ma compatible with the timing of basin aggradation.

Focal-mechanism solutions of intermediate depth and deep earthquakes under the Japan Sea and the East Asian margin (Zhao et al., 2009) show that the Pacific slab is under a compressional-stress regime in the depth range of 200 to 600 km. Such a stress regime in the slab might be caused by the slab meeting strong resistance at the 660-km discontinuity. In the same way, the asthenosphere under East China would be both hot and wet, being associated with the deep dehydration of a stagnant Pacific slab (Ichiki et al.,

2006). The velocity field in the fluid also changes during fold development (Cerpa et al., 2014). The convective circulation process in the upper mantle above the stagnant slab and the deep dehydration process of the slab would cause upwellings of hot and wet asthenospheric materials, leading to the formation of lithospheric attenuation as well as intraplate volcanoes (Zhao, 2012). This fits in the study area observations by Folguera et al. (2018) who modeled the elastic thickness (Te) of the study area. Showing the coincidence between lowest values of Te and emplacements of Plio–Quaternary magmatism, they speculated that heating would have weakened crustal rigidity. However, Te quickly increases toward the foreland and reaches its maximum in the North Patagonian massif and its continuation to the north in the Neuquén platform area where the lithosphere thickens (Tašárová, 2007).

Conclusion

The Collón Curá basin is one of these intramountainous basins of the Mio-Pliocene foothills of the Patagonian Andes. Growing within the wedgetop depozone of a wider Andean foreland basin, sediments accumulate above the active orogenic thrust wedge. Contrarily to other basins where accommodation is achieved by their subsidence, that of these intramontane basins is by the dynamics of the tectonic barrier which limits them (e.g. Hilley and Strecker, 2005). So the tectonic control is crucial for their development. Former studies focused on the surface mapping and logging of the formations in order to obtain a satisfactory stratigraphic frame. This must now be anchored on an absolute timing by multiplying absolute dating of the filling. This effort must be made in order to overcome apparent contradictions rose by the extrapolation of local data (e.g. Franzese et al., 2018).

Another important shortcoming concerns the subsurface data due to its lack of industrial potential. So understanding of its articulation with the Paleogene history of the area remains unclear.

If tectonics plays a determining role in the evolution of these basins, it seemed important to us to apprehend it in parallel with that of its neighbors. And it is striking to note that filling of these basins occurred around 18 Ma. This filling answered to a regional event that authors called the Quechua phase. However this episode did not correspond in the study area to a kinematic reorganization in terms of rates or orientation of the South America and Nazca plates (Figures 9 and 10). The kinematics would not explain to the first order this compressive episode.

In parallel to this, seismic tomography recently imaged anomalies of rapid velocities in the mantle that we interpret as the relics of ancient slabs (Pesicek et al., 2012). It shows slab segments laying above the 660 km-depth mantle discontinuity and slab folding as those obtained in analogic and numerical models (e.g. Guillaume et al., 2009; Gibert et al, 2012; Cerpa et al., 2014; Cerpa et al., 2015). This observation seems important to us because mechanically, the slab controls the transfer of stress and strain between this mantle discontinuity and the surface. Thus the state of stress of the overriding plate and topography too link directly with the dip of the slab. An important result of this interpretation is that a cyclic stress regime in the overriding plate can be achieved with constant boundary conditions. Thus the dynamics of subduction, independently of its boundary conditions, imposes on the surface a cyclicity which will be potentially inscribed in the sedimentary record.

If these basins are initiated at the same time, their excavation spreads itself over time. As a first approximation, this results from the efficiency of the surface processes which, by reconnecting the basin to its foreland, condemn it to excavation. This efficiency is determined by many parameters such as the size of catchments, the slope of the flows, the lithology of the grounds which are specific to a given basin and explains this spreading. But the development of slab folding, as it deposits on the 660 km depth boundary, periodically flatten the slab with a strong impact on the state of stress and topography. Upper plate shortening and uplift when the angle of the slab shallows, extension or lesser shortening and subsidence when it steepens (e.g. Gibert et al., 2012; Cerpa et al, 2014). The balance between local uplift (fault reactivation along the tectonic dam) and regional uplift (dynamic topography) through episodes of higher shortening, will also interfere with surface processes. The evolution of these basins should also be apprehended by the link to their close foreland that will generate or not space to accommodation. So a quantification of the regional vertical displacements is required to constrain well the dynamics of these basins.

Located in the hinterland of a volcanic arc, the combination of preserved landforms and abundant intercalated volcanic ashes may resolve the timing of filling and excavation, tectonic activity, and geomorphic evolution necessary to indentify a forcing of the internal dynamics of subduction on these depozones.

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FIGURE CAPTIONS

Figure 1: Neogene intra mountaine basins (yellow patches) of the North Patragonian foothill with main structures above the shaded topography. Inset with location of Figure 1 in South America.

Figure 2: Geological map of the southern Neuquén basin with location of the Collón Curá basin (location in Figure 1). Red dots are the sedimentary sections in Figure 3.

Figure 3: Sedimentary sections in the Collón Curá basin (from Huyghe et al., 2015a). Stars are the sampled ignimbrites for dating (40Ar/39Ar on biotite).

Figure 4: Synthesis of the ages of Cenozoic sedimentary formations in the southern part of the Neuquén basin. Dating results are from: 1: Barrio et al. (1989), 2: Pascual et al. (1984), 3: Rolleri et al. (1984), 4: Pascual and Odreman Rivas (1971), 5: Marshall et al. (1977), 6: Gonzalez Diaz and Nullo (1980), 7: Rabassa (1975), 8: Mazzoni and Buenvenuto (1990), 9: Cazau et al. (1989), 10: Gonzalez Diaz et al. (1990), 11: Ramos et al. (2014b) and 12: this study. Biostratigraphic ages are in blue and radiometric ages are in red.

Figure 5: 40Ar/39Ar age spectra for the two sampled ignimbrites at the outcropping base and at the top of the Collón Curá Basin. Ar-11-15 (top) is concordant, Ar-11-30 (bottom) is discordant.

Figure 6: Paleogeographic sketch map of the Collón Curá Basin and foreland from early Miocene (bottom) to Present (top).

Figure 7: Compilation of the ages of the sedimentary filling of the main intra montane basins between 38 and 41°S. Ages are from Roth (1922); González Bonorino (1973); Giacosa et al. (2001); Cazau et al. (2005) and Paredes et al. (2009) for the Ñirihuau Basin; from Marshall et al. (1977); Rabassa (1978); Rapela et al. (1983); Mazzoni and Benvenuto (1990); Cucchi y Leanza (2005); Orts et al. (2012); Bilmes et al. (2013, 2014); Ramos et al. (2014b, 2015); Rojas Vera et al. (2014) and Franzese et al. (2018) for the Catan Lil basin; Leanza & Hugo (2001) and Messager et al. (2010) for the Agua Amarga Basin, and this study for the

Collón Curá Basin. The bottom of the sedimentary infill does not outcrop in the Collón Curá Basin, so agradation began before the dated ignimbrite Ar-11-30 (around 19 Ma). The light grey triangles show the uncertainty on onset of agradation.

Figure 8: Oceanic spreading rates above the South Atlantic (dashed lines) and South Pacific (dotted lines) ridges at 37 and 40°S from Oligocene to Present (from Meyer et al., 2017).

Figure 9: Location of the South America plate with respect to the current geographical frame through time from Oligocene to Present. South America continuously moved to the northwest during this time interval (from Müller et al., 2018).

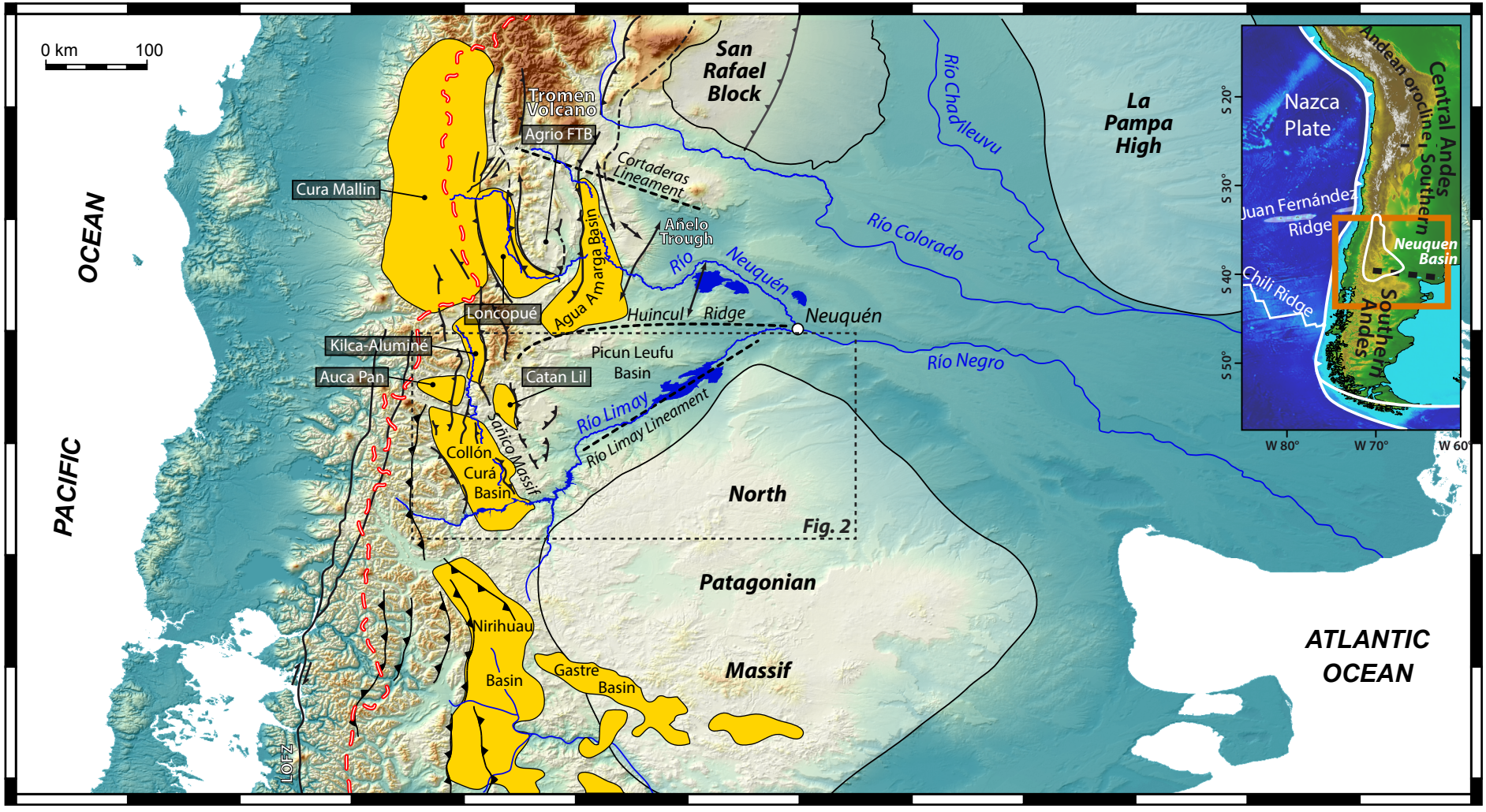
Figure 10: Convergence rates (red) and angle of convergence (violet) between the Nazca and South America plates from Oligocene to Present at the latitude of the Collón Curá basin (40°S; from Müller et al., 2018).

Figure 11: A) Seismic tomographic image of lithosphere and lower mantle at 37 and 39°S (from Pesicek et al., 2012). B) Analogical modeling of slab folding above the 660 km-depth mantle discontinuity (from Guillaume et al., 2009).

Figure 12: An interpretation of the seismic tomographic image by Pesicek et al. (2012). A) Stacking of slab folds above the 660 km-depth mantle discontinuity and its stagnation could force convective circulation process in the upper mantle and a deep dehydration process of the slab. The resulting upwellings of hot and wet asthenospheric materials could lead to lithospheric attenuation and intraplate volcanoes (Zhao, 2012). B) First shallow slab break-off with upward rebound of the remaining slab (Haschke et al., 2006). This break-off might cause the basins to invert. C) Second deep slab break-off at the top of the slab fold stack with sinking in the lower mantel.

Figure 13: Four characteristic modes of subduction toward the ocean (mode 2b) or the continent (other modes) and implying folding of the oceanic slab (mode 1a and 2a) or not (mode 1b and 2b) . The arrows show far-field absolute plates velocities (black arrows indicate trenchward motion). The symbol ⊗

indicates that the velocity is zéro (from Gibert et al., 2012).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Lineament
Cura Mallin
Añelo
Trough
Agua Amarga Basin
Río Chadileuvu
Río Colorado
Río
Neuquén
Huincul
Ridge
Neuquén
Loncopué
Kilca-Aluminé
Auca Pan
Picun Leufu
Basin
Catan Lil
Río Negro
Río Limay
Río Limay Lineament
Sánico Massif
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Fig. 2
Patagonian
Massif
Nirihuau
Basin
Gastre
Basin
LÓFZ
Nazca
Plate
Juan Fernández
Ridge
Chili Ridge
Andean orocline
Central Andes
Southern
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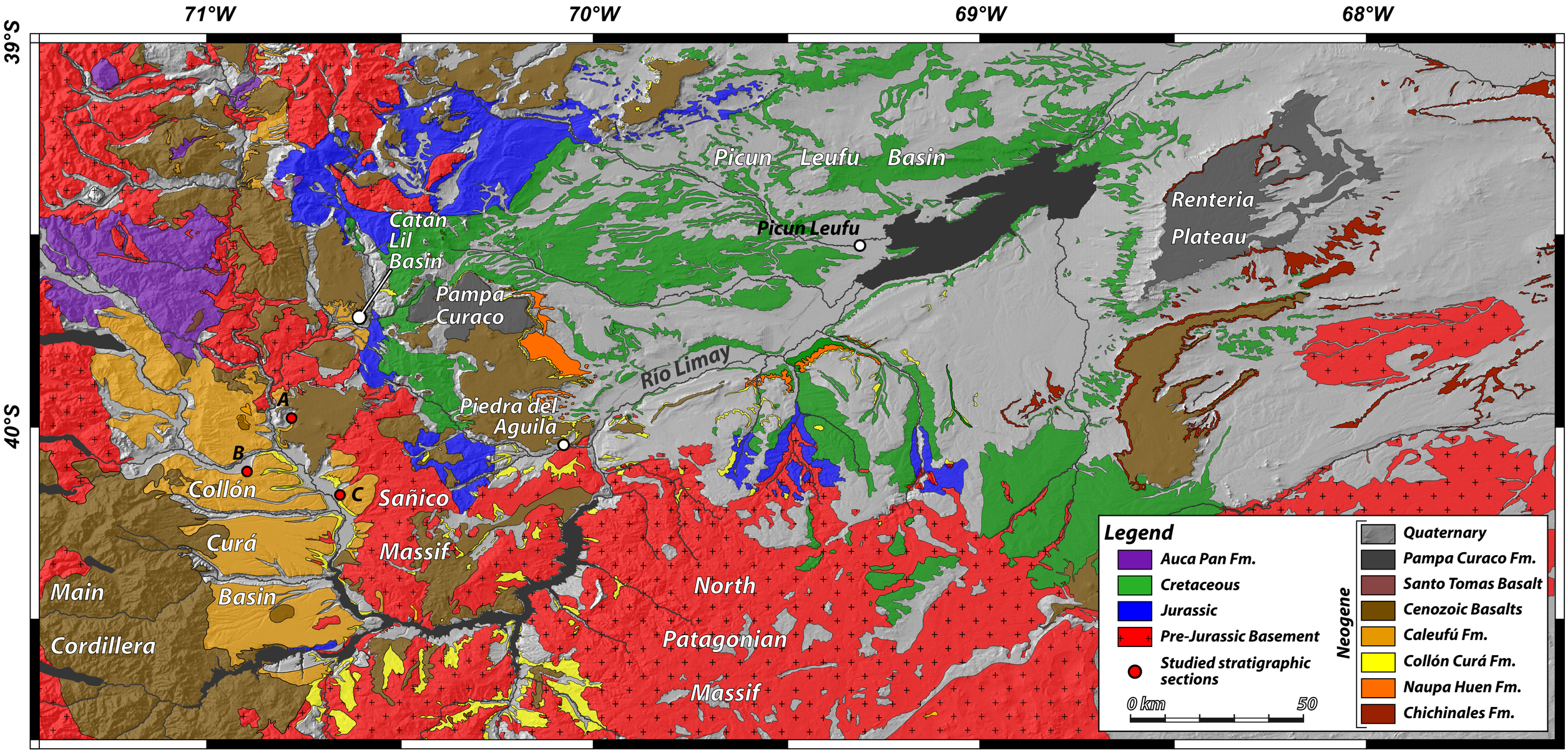
71°W
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39°S
40°S
Picun Leufu Basin
Picun Leufu
Renteria Plateau
Catán Lil Basin
Pampa Curaco
Río Limay
Piedra del Aguila
A
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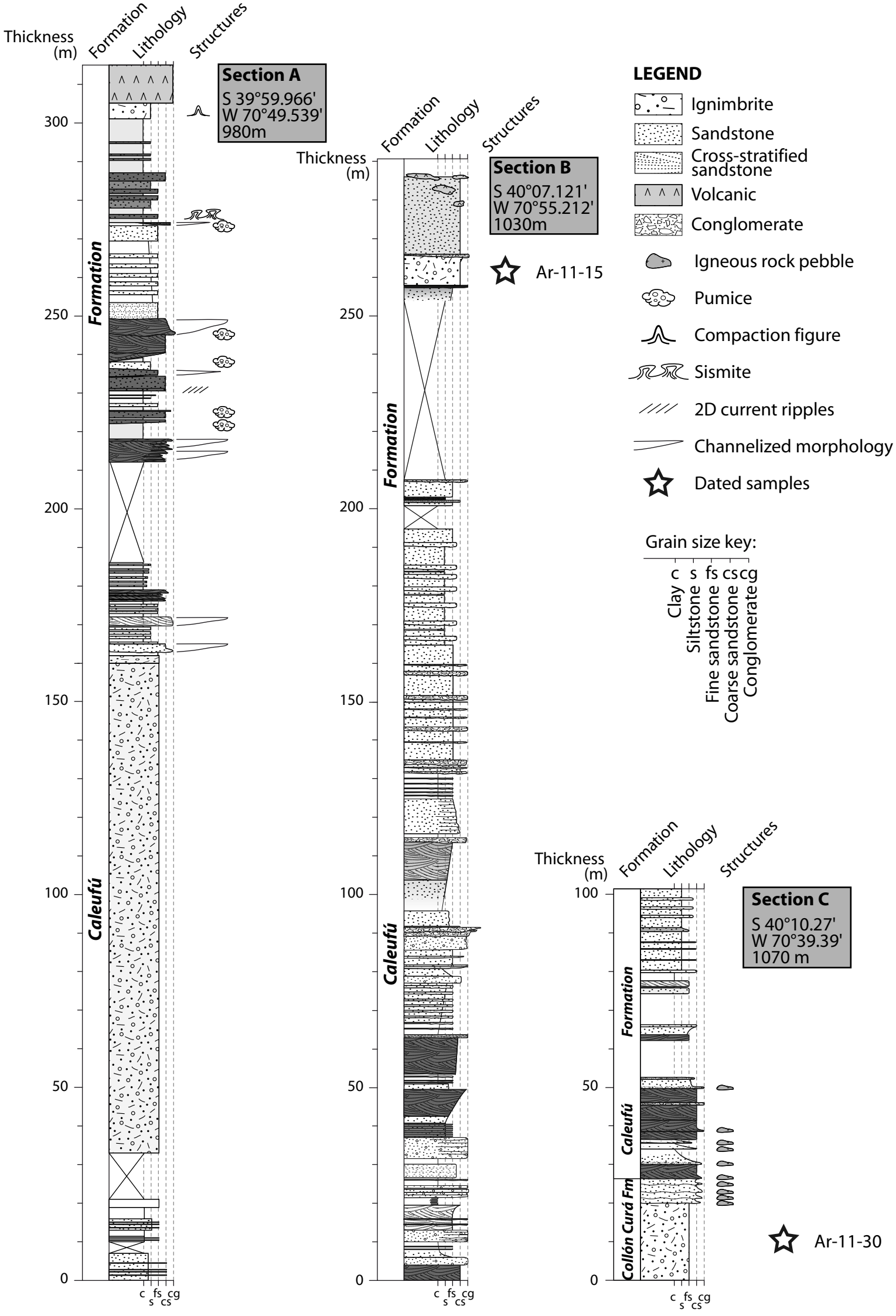
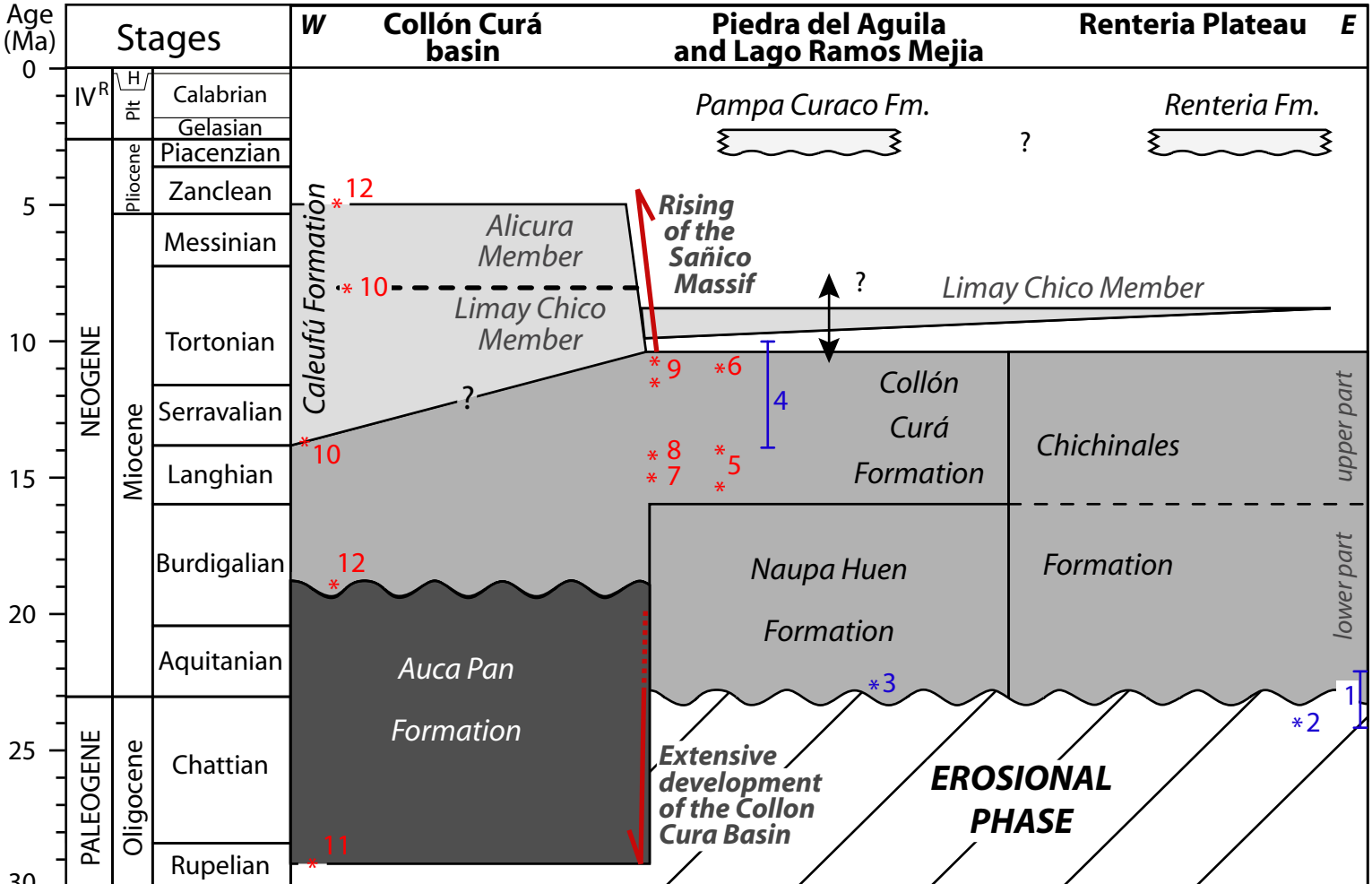
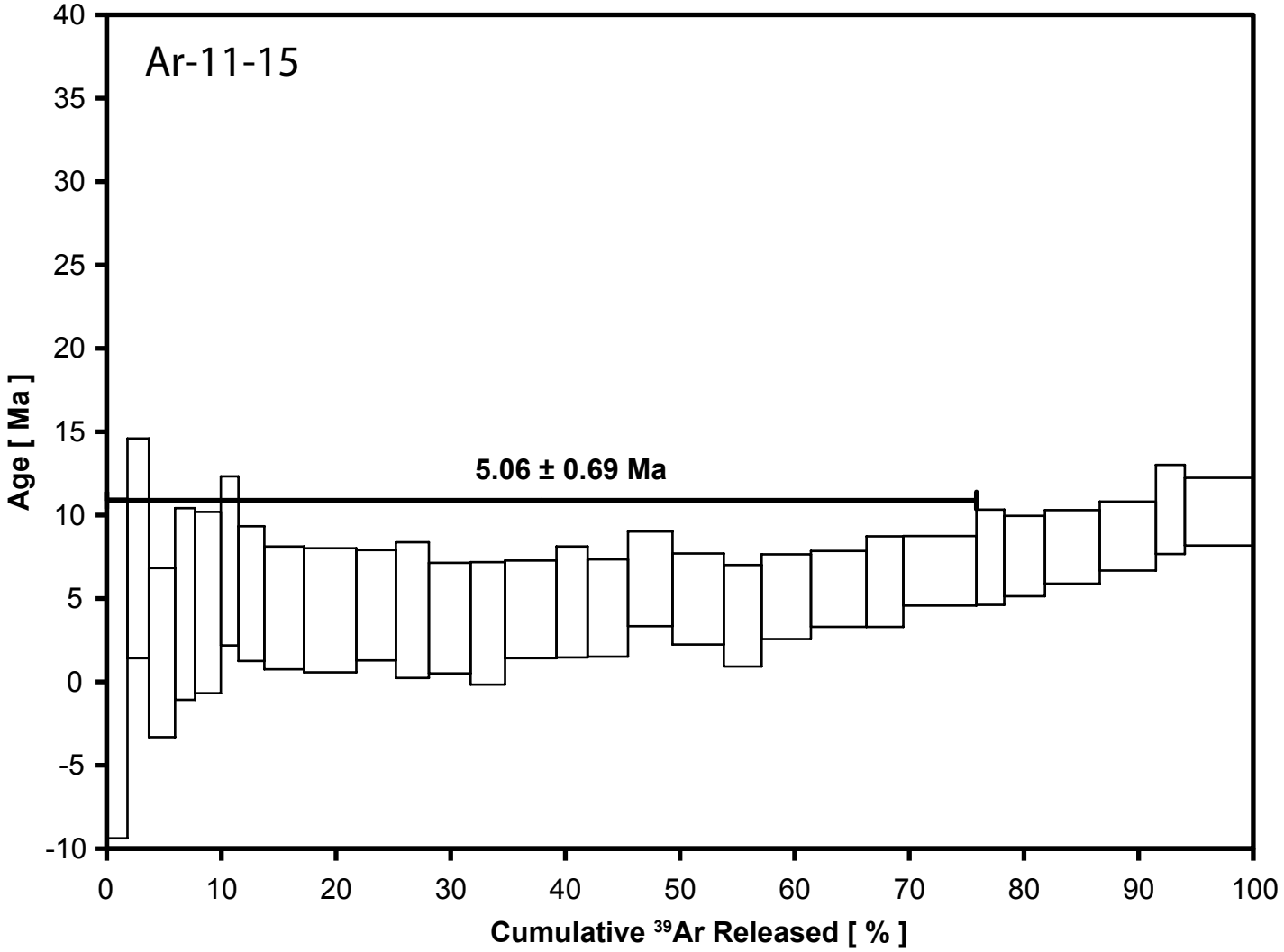
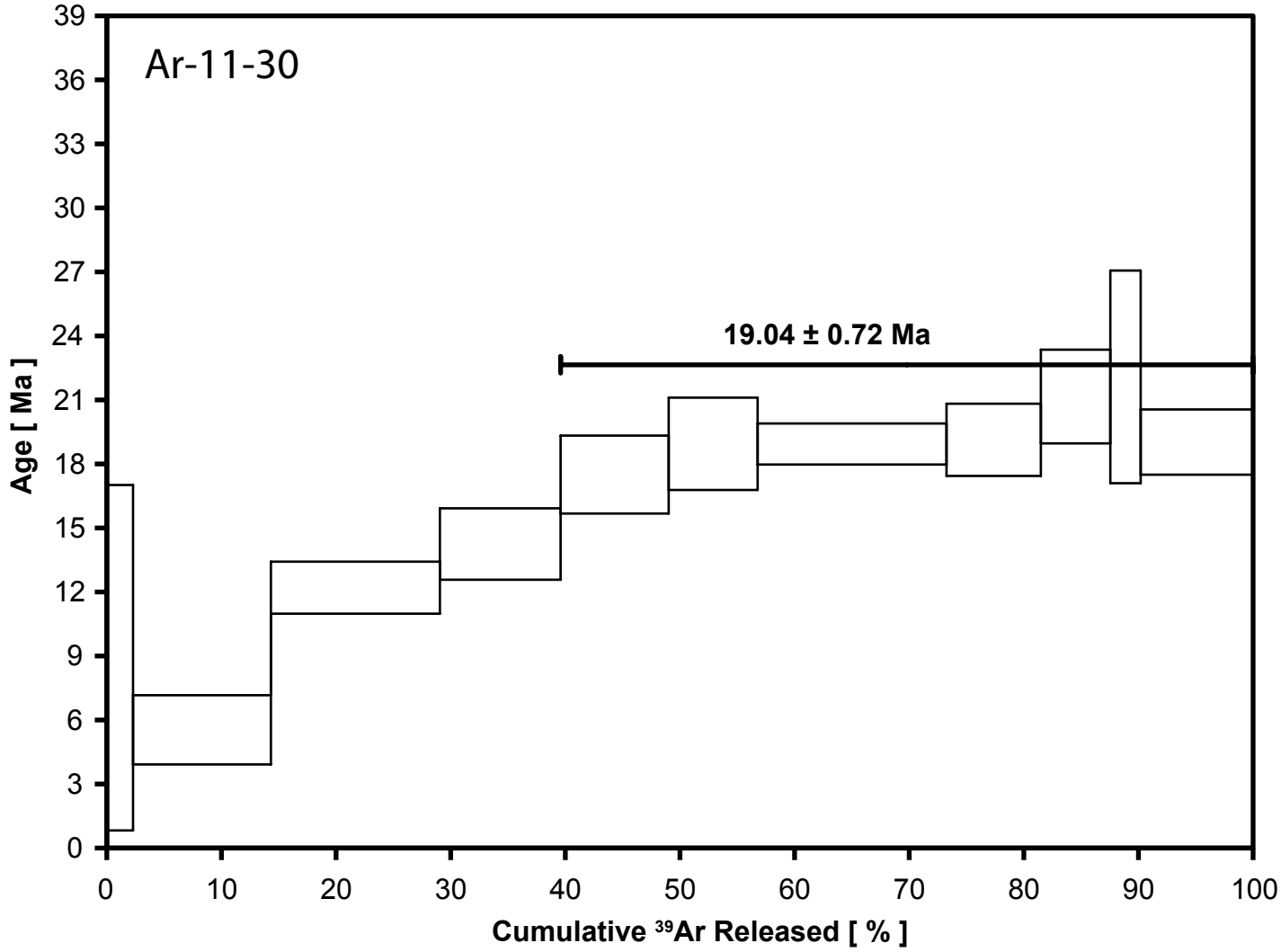
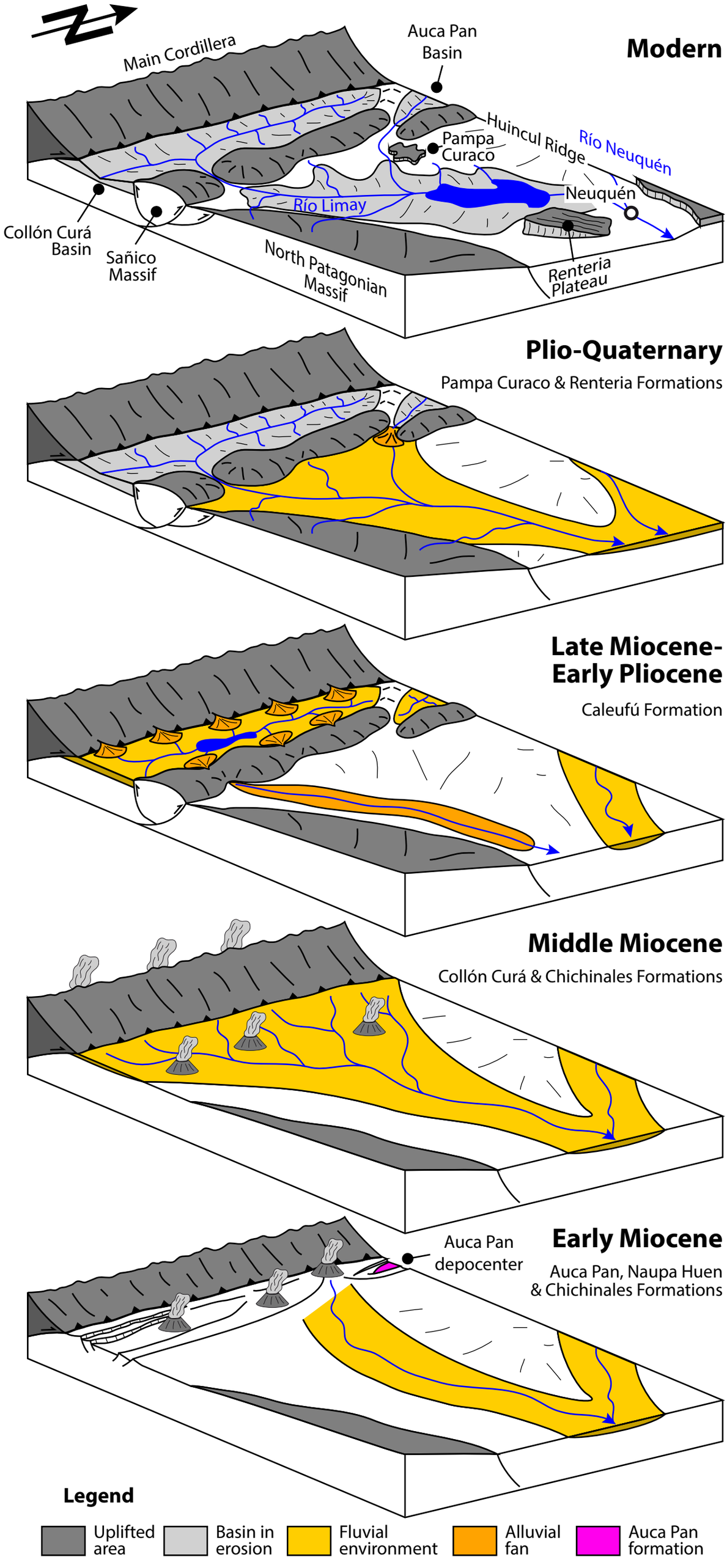
C
Collón
Sañico Massif
Curá Basin
Main Cordillera
North Patagonian Massif
Legend
Auca Pan Fm.
Cretaceous
Jurassic
Pre-Jurassic Basement
Studied stratigraphic sections
0 km
50
Neogene
Quaternary
Pampa Curaco Fm.
Santo Tomas Basalt
Cenozoic Basalts
Caleufú Fm.
Collón Curá Fm.
Naupa Huen Fm.
Chichinales Fm.

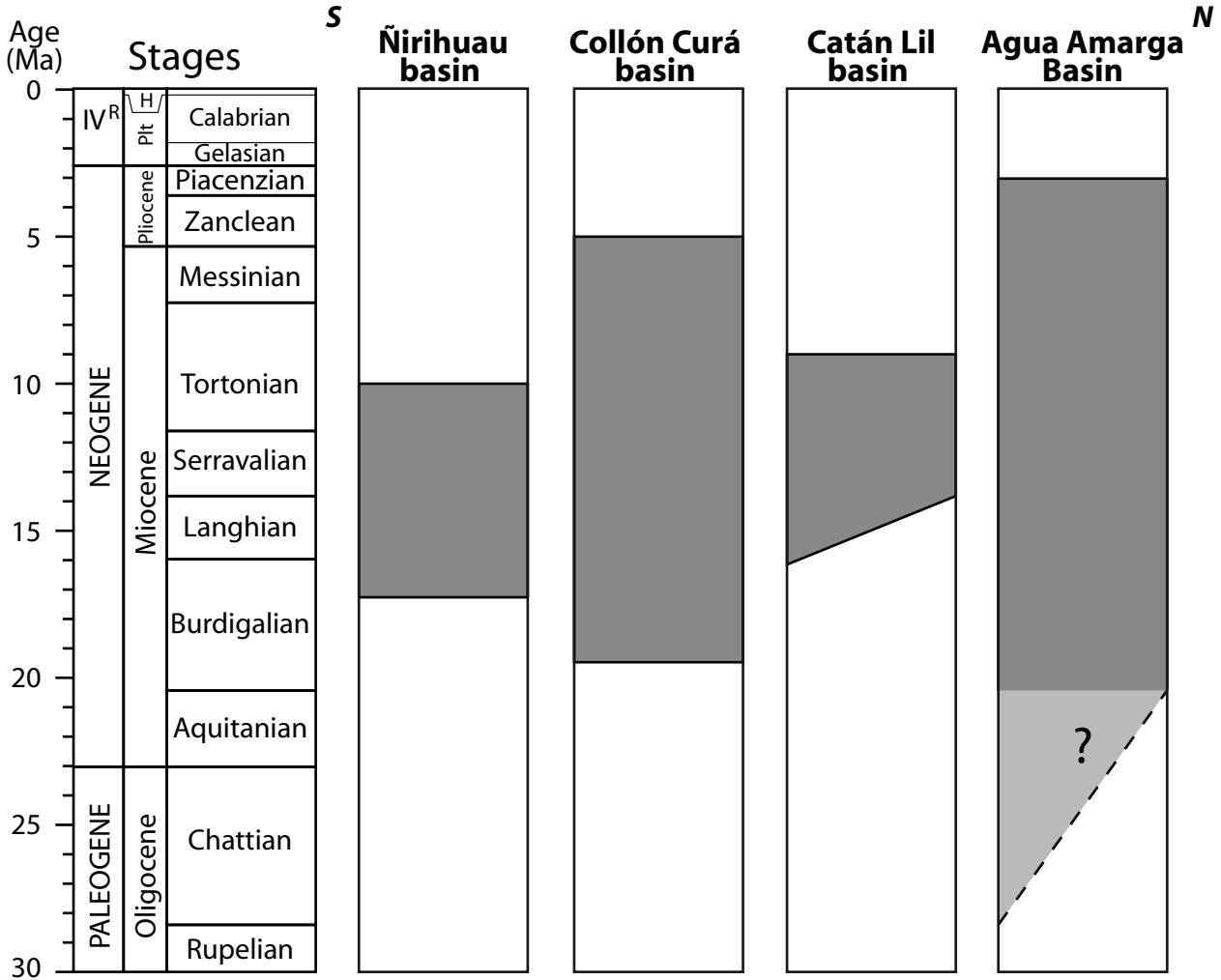

Thickness (m)
Formation
Lithology
Structures
Section A
S 39°59.966'
W 70°49.539'
980m
Section B
S 40°07.121'
W 70°55.212'
1030m
Section C
S 40°10.27'
W 70°39.39'
1070 m
Ar-11-15
Ar-11-30
Caleufú Formation
Caleufú
Collón Curá Fm
LEGEND
Ignimbrite
Sandstone
Cross-stratified sandstone
Volcanic
Conglomerate
Igneous rock pebble
Pumice
Compaction figure
Sismite
2D current ripples
Channelized morphology
Dated samples
Grain size key:
c s fs cs cg
Clay
Siltstone
Fine sandstone
Coarse sandstone
Conglomerate

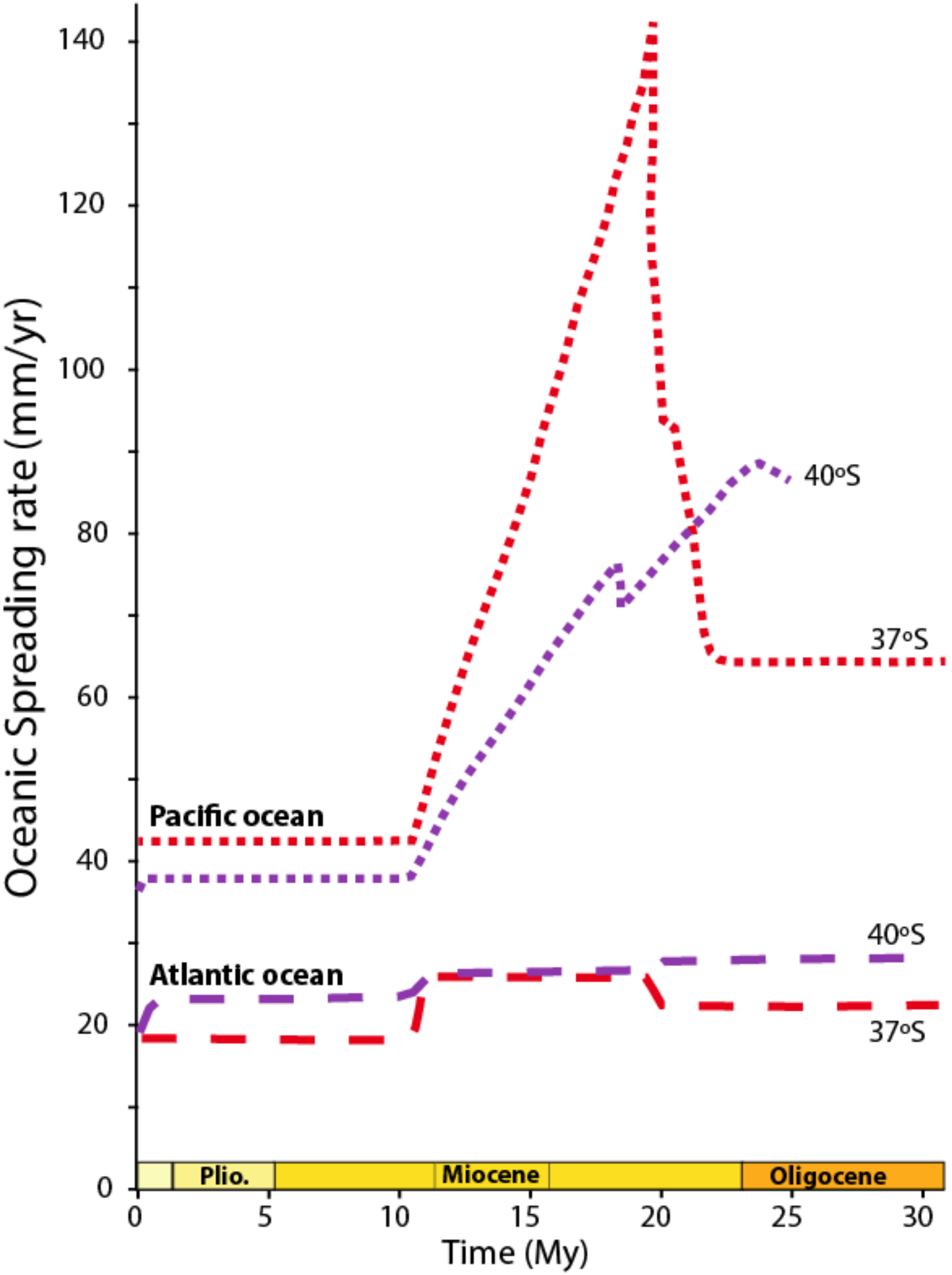

Age (Ma)
Stages
W
Collón Curá basin
Piedra del Aguila and Lago Ramos Mejia
Renteria Plateau
E
0
5
10
15
20
25
30
IV R
H
Plt
Calabrian
Gelasian
Piacenzian
Zanclean
Pliocene
Messinian
Tortonian
Serravalian
Langhian
Burdigalian
Aquitanian
Chattian
Rupelian
Miocene
Oligocene
NEOGENE
PALEOGENE
Pampa Curaco Fm.
?
Renteria Fm.
Caleufú Formation
Alicura Member
Limay Chico Member
Rising of the Sañico Massif
?
Limay Chico Member
?
Collón Curá Formation
Chichinales
upper part
Naupa Huen Formation
Formation
lower part
Auca Pan Formation
Extensive development of the Collon Cura Basin
EROSIONAL PHASE
12
10
10
12
11
9
6
8
7
5
4
3
2
1

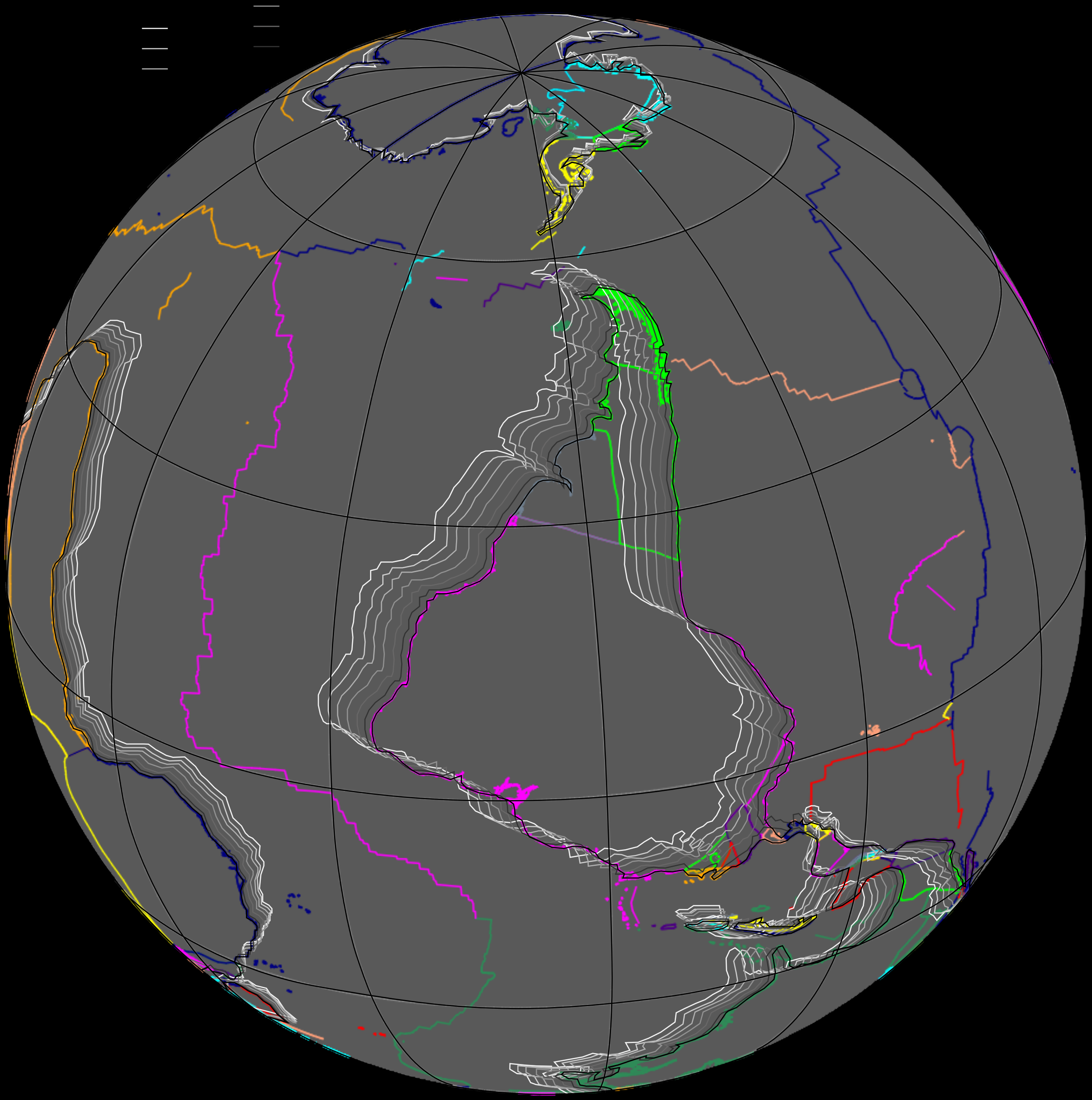

Ar-11-15
5.06 ± 0.69 Ma
Age [Ma]
Cumulative ^{39}Ar Released [%]
40
35
30
25
20
15
10
5
0
-5
-10
0
10
20
30
40
50
60
70
80
90
100

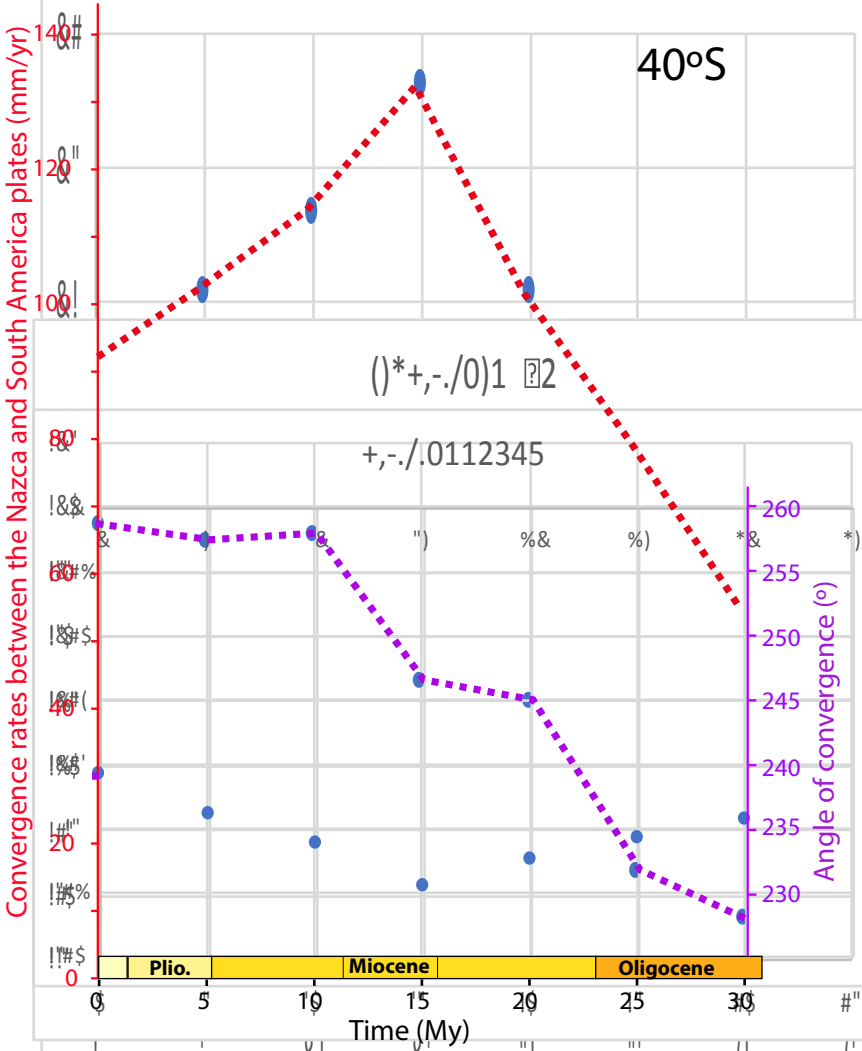

Ar-11-30
19.04 ± 0.72 Ma
Age [Ma]
Cumulative ^{39}Ar Released [%]
39
36
33
30
27
24
21
18
15
12
9
6
3
0
0
10
20
30
40
50
60
70
80
90
100

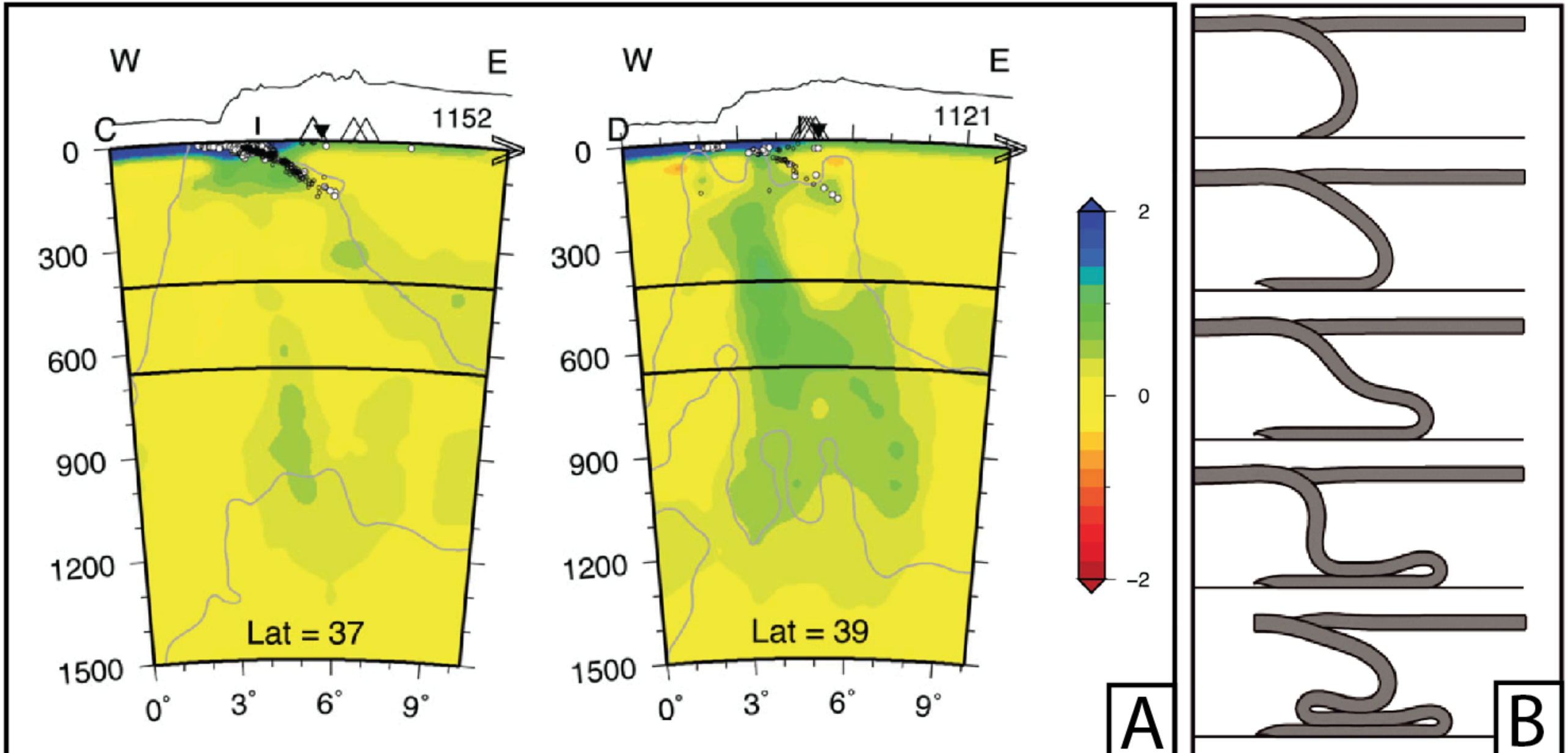

Modern
Main Cordillera
Auca Pan Basin
Huincul Ridge
Río Neuquén
Pampa Curaco
Neuquén
Río Limay
Collón Curá Basin
Sañico Massif
North Patagonian Massif
Renteria Plateau
Plio-Quaternary
Pampa Curaco & Renteria Formations
Late Miocene-Early Pliocene
Caleufú Formation
Middle Miocene
Collón Curá & Chichinales Formations
Early Miocene
Auca Pan depocenter
Auca Pan, Naupa Huen & Chichinales Formations
Legend
Uplifted area
Basin in erosion
Fluvial environment
Alluvial fan
Auca Pan formation

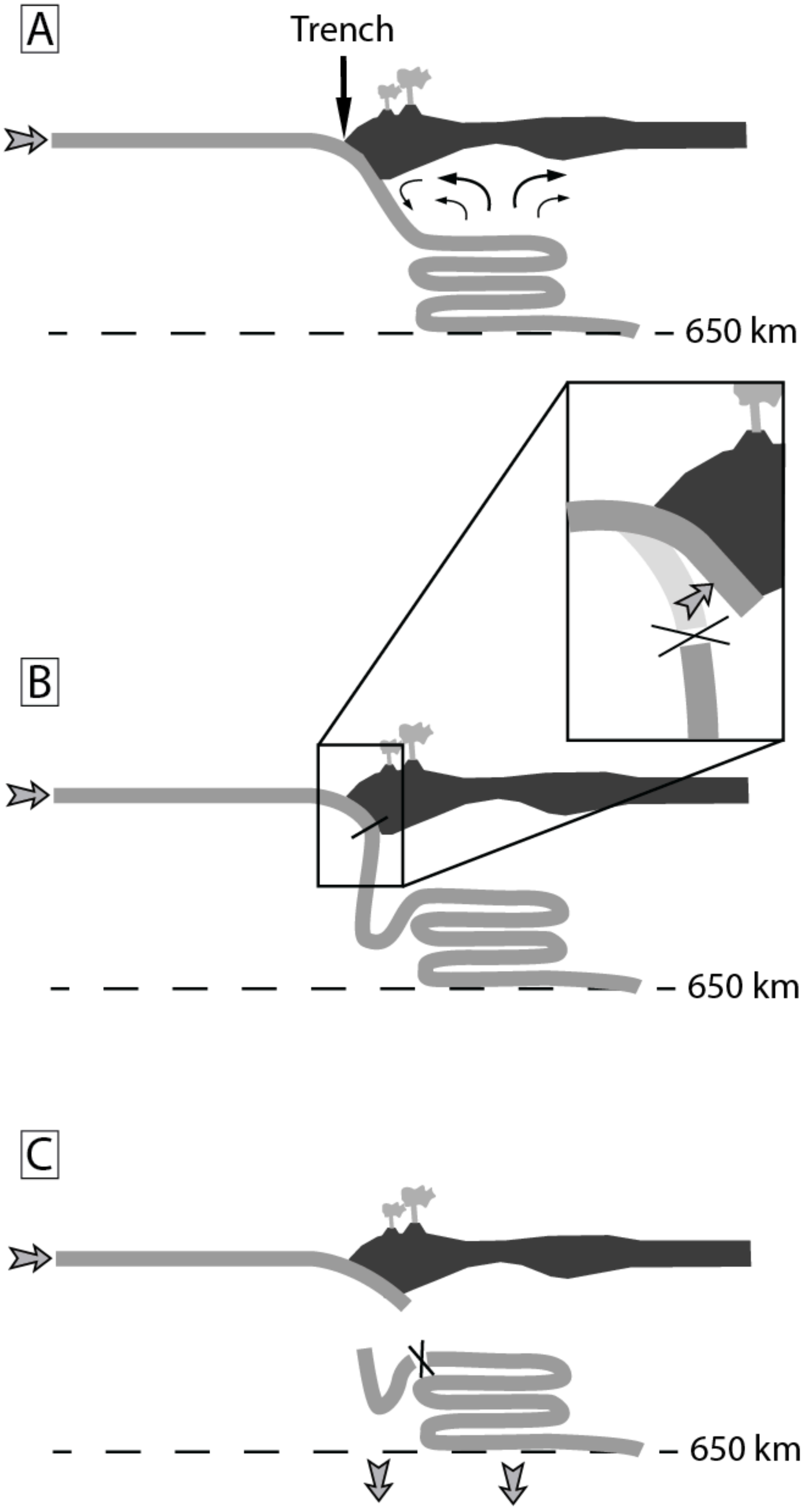

Age (Ma)
Stages
S
N
Ñirihuau basin
Collón Curá basin
Catán Lil basin
Agua Amarga Basin
0
5
10
15
20
25
30
IV R
H
Plt
Calabrian
Gelasian
Piacenzian
Zanclean
Pliocene
Messinian
Tortonian
Serravalian
Langhian
Burdigalian
Aquitanian
Miocene
NEOGENE
Chattian
Rupelian
Oligocene
PALEOGENE
?

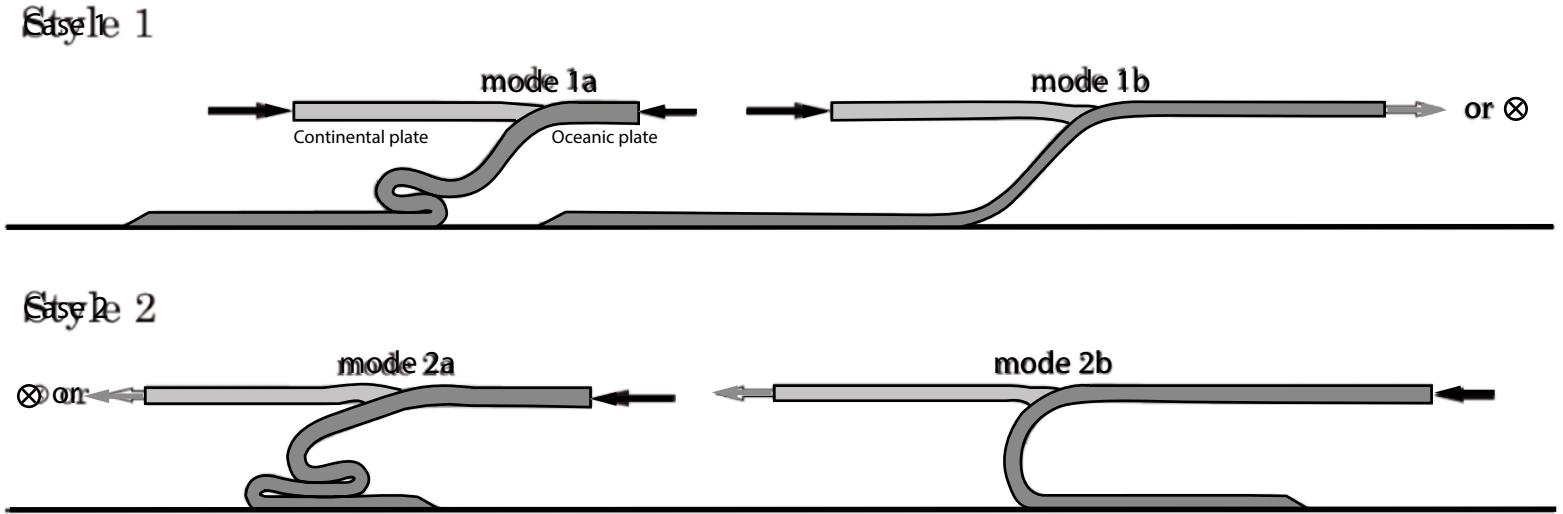

Oceanic Spreading rate (mm/yr)
140
120
100
80
60
40
20
0
Pacific ocean
40°S
37°S
Atlantic ocean
40°S
37°S
Plio.
Miocene
Oligocene
0
5
10
15
20
25
30
Time (My)




40°S
Convergence rates between the Nazca and South America plates (mm/yr)
Angle of convergence (°)
Time (My)
Plio.
Miocene
Oligocene
0
20
40
60
80
100
120
140
230
235
240
245
250
255
260
5
10
15
20
25
30


W
E
C
I
1152
0
300
600
900
1200
1500
Lat = 37
0°
3°
6°
9°
W
E
D
1121
Lat = 39
2
0
−2
A
B


A
Trench
650 km
B
650 km
C
650 km


Case 1
mode 1a
Continental plate
Oceanic plate
mode 1b
or ⊗
Case 2
mode 2a
⊗ or
mode 2b