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Key Points:

- U-Pb and ^{10}Be SED from N Patagonia constrain patterns of deformation, and their temporal variations during a major climatic transition
- Data record a progressive decrease of upper-plate shortening after a first period of widespread deformation and uplift (13–7 Ma)
- Acceleration in erosion rates during the past 7 Ma may have influenced the regional geometry and kinematic history of the orogenic belt

Supporting Information:

- Supporting Information S1
- Table S1

Correspondence to:

E. G. Morabito,
ezequiel.morabito@gmx.ch

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The Influence of Climate on the Dynamics of Mountain Building Within the Northern Patagonian Andes

Ezequiel García Morabito^{1,2,3} , Alejandro Beltrán-Triviño^{3,4} , Carla M. Terrizzano¹, Florencia Bechis², Jeremías Likerman⁵, Albrecht Von Quadt³ , and Víctor A. Ramos⁵

¹Institute of Geography and Oeschger Centre for Climate Change Research, University of Bern, Bern, Switzerland,

²IIDyPCA, CONICET, Universidad Nacional de Río Negro, San Carlos de Bariloche, Argentina, ³Swiss Federal Institute of Technology in Zurich (ETHZ), Zürich, Switzerland, ⁴Department of Geology, Universidad EAFIT, Medellín, Colombia, ⁵IDEAN, University of Buenos Aires - CONICET, Viamonte, Argentina

Abstract Theoretical studies support the idea that the internal dynamics of actively deforming mountain ranges are influenced by spatial and temporal variations in climate. The identification of plausible correlations between orogen behavior and external climatic processes requires, among other factors, that the initiation and duration of any deformation event be precisely constrained. Here, we add new detrital zircon U-Pb ages and ^{10}Be surface exposure dating to the already extensive data set for the low-temperature cooling history, to analyze the spatial patterns of deformation, and their temporal variations across the foreland of North Patagonia, with the aim of evaluating potential feedbacks between climate and deformation. Stratal relationships, together with geomorphic evidences of deformation document the precise timing of individual structures within the fold-thrust belt and the broken foreland. Our data record a progressive decrease of upper-plate shortening rates and subsidence after a first period of widespread deformation and uplift (ca. 13–7 Ma). This “transitional” foreland phase coeval to the onset of glacial conditions at 7.4–5 Ma is followed by a structural reorganization after ca. 3 Ma, marked by the abandonment of the foreland and enhanced slip on interior faults, with the intensification of glacial erosion at this time. We propose that acceleration in erosion rates during the past 7 Ma may have influenced the regional geometry and kinematic history of the orogenic belt. Our empirical results match theoretical predictions and provide compelling evidence at the scale of individual thrust faults for the significant impact of climate change on orogenic behavior.

1. Introduction

Spatial and temporal variations in climate are hypothesized to have a significant influence on orogen behavior (e.g., Berger et al., 2008; Whipple, 2009). Empirical and theoretical studies, both postulate that a climatically modulated increase in erosion should force structural reorganization within an orogen, with the leading edge of deformation retracting during intervals of enhanced erosion (e.g., Davis et al., 1983; S. D. Willett, 1999; Whipple, 2009; Yagupsky et al., 2014). Extensive glaciation, in particular, is expected to have a significant impact on orogenesis (e.g., Tomkin & Roe, 2007), as glaciers are efficient enough to reduce summits within a few million years. Obtaining empirical evidence of a significant impact of climate in mountain building is, however, a challenging task (Whipple, 2009).

The North Patagonian Andes of South America (39°–47°S) is an optimal system for studying the interaction between climate and tectonics from this perspective (Figure 1). First, the orogen has experienced extensive mountain glaciation for ~7 Ma (J. H. Mercer & Sutter, 1982). Second, low-temperature apatite fission track and (U-Th)/He thermochronometry constrains the exhumation pattern in the orogen and its foreland (Figure 2) (Herman & Brandon, 2015; Savignano et al., 2016; Thomson et al., 2010). Third, Neogene deposits hosted in intermontane basins are widespread through the eastern piedmont (Figure 2), and growth strata within these basins provide insight into the nature and timing of thrust-related deformation (e.g., Butler et al., 2019; García Morabito et al., 2011; M. E. Ramos et al., 2015). Moreover, plateaus (“mesetas”) consisting of late Miocene and early Pliocene basalt flows, and tablelands (“pampas”) covered by gravel mantles are locally tilted, and may provide insight into spatio-temporal migration of deformation.

The Northern Patagonian Andes have previously been referred to as an example of a critical taper orogeny, where a climatically triggered increase in erosion forced a structural reorganization causing a reduction

Figure 1. General geological setting of southern South America showing sedimentary basins (NB, Neuquén Basin; SJGB, San Jorge Gulf Basin; MAB, Magallanes - Austral Basin; CB, Colorado Basin; VB, Valdés Basin; RB, Rawson Basin; SJB, San Julian Basin; MB, Malvinas Basin; NMB, Northern Malvinas Basin) and key orogenic and structural elements (North Patagonian Andes, South Patagonian Andes, Patagonian broken foreland, CC, Coastal Cordillera; CD, Central Depression; LOFZ, Liquiñe-Ofqui fault zone; FTB, Fold-thrust belt; NPM, North Patagonian Massif; DM, Deseado Massif). Black box shows location of the study region (Figure 2).

in the cross-sectional area of the orogen (Thomson et al., 2010). However, attempts to generalize orogen behavior based on modern snapshots of cooling ages in the absence of further constraints have to be taken cautiously (Burbank & Anderson, 2012). The identification of plausible feedbacks between climate change, erosion, and orogen behavior, requires that the time of initiation and duration of any deformation event be precisely constrained. So far, timing, rates, and spatial patterns of deformation in the North Patagonian Andes have not been properly constrained, since numeric age control in the foreland of Northern Patagonia has largely been missing. Here we add new detrital zircon U-Pb and ^{10}Be surface exposure data to the already extensive data set for the low-temperature cooling history, to constrain the spatial and temporal structural evolution of the North Patagonian Andes, and to evaluate potential feedbacks between climate and deformation.

2. The North Patagonian Andes

The North Patagonian Andes form a relatively narrow and low-elevation orogenic belt along the western margin of South America (Figure 1). It is the product of the ongoing oblique convergence between Nazca Plate oceanic lithosphere beneath the South American continent. Deformation is currently focused along the axis of the orogen within a right-lateral strike-slip to transpressional fault system (Liquiñe-Ofqui fault zone, LOFZ) that spans almost the entire length of the North Patagonian Andes controlling the emplacement of the volcanic arc (Cembrano et al., 1996). Transpression at the southern end of the fault since the Middle Miocene is generally attributed to the collision of the Chile Rise (Forsythe & Nelson, 1985;

Figure 2. Regional setting of the Andean orogen between 38°S and 44°S showing major Neogene depocenters (Lo, Lonquimay; Loe, Loncopué; Al, Aluminé; CL, Catán Lil; PA, Piedra del Aguila; CC, Collón Cura; Ni, Ñirihuau; Bo, Bolsón; Ga, Gastre; PS, Paso del Sapo; Ag, Agnia), location of growth strata localities, cosmogenic nuclide samples (blue dots), and U-Pb sample locations (black dots). Apatite (U-Th)/He data are shown as yellow squares, apatite fission-track data locations are shown as blue squares (Thomson et al., 2010), and instrumental surface earthquakes (<33 km depth) (www.earthquake.usg.gov) are represented by small pink dots.

Georgieva et al., 2016). The orogen intersects the Westerlies Wind Belt (WWB) concentrating heavy orographic precipitation of $5\text{--}10\text{ m yr}^{-1}$ on its windward side, which rapidly decreases across the leeward flank.

The Patagonian Andes were repeatedly covered by ice since the latest Miocene (7.4–5 Ma) (J. H. Mercer & Sutter, 1982; Rabassa et al., 2011; Ton That et al., 1999). At least during the Pleistocene, ice coverage was nearly complete, and many glaciers reached the lowlands through numerous glacial valleys (Davies et al., 2020; Rabassa et al., 2011). These glacial events of varied duration and intensity promoted important landscape transformations as a result of high rates of erosion and rapid sediment transfer to the adjacent basins (Christeleit et al., 2017; Melnick & Echtler, 2006; Thomson et al., 2010; Villaseñor et al., 2019; C. D. Willett et al., 2020). Erosion rates began to accelerate specifically in the main Cordillera along the LOFZ region around 5–7 Ma (Thomson et al., 2010), coincident with late Cenozoic global cooling (Zachos et al., 2001), increasing particularly rapidly during the past 2 m.y. (Herman & Brandon, 2015). The expansion of ice sheets occurred on a previously uplifted topography. In the North Patagonian Andes, the process of mountain building has been governed by contrasting tectonic regimes regulated by plate-scale changes in convergence and shifts in the geometry of the subducting oceanic slab (e.g., Folguera & Ramos, 2011; Horton, 2018). Pre- and synorogenic extension took place prior to the period investigated in the study. Widespread pre-Andean extension occurred in the backarc with the opening of a series of Mesozoic basins characterized by synrift and postrift phases of evolution (Figure 1) (e.g., Uliana et al., 1989). In turn, Cenozoic extension rather focused on restricted forearc and retroarc settings (e.g., Bechis et al., 2014; Burns et al., 2006; Charrier et al., 2002; García Morabito & Ramos, 2012; Jordan et al., 2001; D. Orts et al., 2012). Neogene reactivation of extensional faults resulted in a disorganized retroarc fold-and-thrust belt that includes a wide broken foreland province, referred to as the Patagonian broken foreland (PBF). This broad and discontinuous region is characterized by a series of uplifted basement blocks that grew separately from the main Andean axis and constitute the highest summits of the extra-Andean region. The PBF hosts a series of synorogenic sedimentary basins filled with volcanoclastic, and continental deposits, including documented growth strata along associated thrust faults (Figure 3) (Butler et al., 2019; Echaurren et al., 2016; García Morabito & Ramos, 2012; García Morabito et al., 2011; M. E. Ramos et al., 2015). Sediments within these basins provide a record of the regional geometry and kinematic history of the North Patagonian Andes since the Miocene. Furthermore, basaltic lava flows and fluvioglacial deposits covering the synorogenic successions are locally tilted, and provide insight into the manner in which deformation has migrated in space and time. Low-temperature apatite fission-track and (U–Th)/He thermochronometry constrains the exhumation pattern in the orogen (Adriasola et al., 2006; Glodny et al., 2007; Thomson, 2002; Thomson et al., 2010), and the syn-orogenic foreland deposits have been the focus of several works in recent years (Bucher et al., 2019; Butler et al., 2019; Echaurren et al., 2016; García Morabito et al., 2011; D. Orts et al., 2012; M. E. Ramos et al., 2015). The combination of these distinct data sets could provide a good general view on how the Patagonian orogen has evolved kinematically in time.

Figure 3. Representative stratigraphic columns of the main Neogene depocenters including compiled (K–Ar, Ar–Ar [in italics] and U–Pb [gray]) and new ages (U–Pb [red], ^{10}Be [blue]).

3. Regional Stratigraphy

The basement of northern Patagonia consists of Paleozoic-Mesozoic metamorphic complexes and deformed igneous suites (Colohuincul Complex; Dalla Salda et al., 1991) well exposed on the eastern flank of the North Patagonian Andes. These units were intruded by calc-alkaline subduction related plutonic rocks, which form part of the Early Jurassic Subcordillerean Batholith and the North Patagonian Batholith (González Díaz, 1982; Toubes & Spikerman, 1973). The ages of the North Patagonian Batholith and its volcanic equivalents range from middle Jurassic to Neogene, although they have clear predominance of Early-Late Cretaceous magmatic components (Castro et al., 2011; Pankhurst et al., 1999).

Paleozoic-Triassic metamorphic and igneous basement units are also exposed within the North Patagonian Massif (Figure 2). This distal basement domain bounds to the north with the Neuquén basin, a major petroleum province in Argentina (Figure 1). The basin contains thousands of meters of Mesozoic sedimentary successions accumulated on the eastern side of the Andean volcanic arc under variable sedimentary and tectonic conditions (e.g., Vergani et al., 1995). South of the North Patagonian Massif, exhumed depocenters of the Middle Jurassic to Early Cretaceous Cañadon Asfalto Basin are present (Figari, 2005). The Jurassic sequences are unconformably overlain by continental and volcanoclastic deposits of the late Early-Late Cretaceous Chubut Group. A depositional hiatus separates the Chubut Group from the overlying marine Upper Cretaceous-lower Paleogene Paso del Sapo and Lefipán Formations (Scasso et al., 2012; Spalletti, 1996).

Cenozoic volcanic rocks of the Paleocene-Eocene Pilcaniyeu Belt (~60–55 Ma, Huitrera Formation) and Eocene-lowermost Miocene of the El Maitén Belt (37–23 Ma, Ventana Formation) (Rapela et al. 1988) cover vast areas of the Patagonian foreland. The Ventana Formation forms part of the initial infill of the Ñirihuau, and probably of the Collón Cura basin (Figure 2). The late Cenozoic sedimentary infill is concentrated in a series of isolated depocenters well represented across the whole foreland region (Figure 2). The Neogene basin fill of the Patagonian foreland unconformably overlies Mesozoic sedimentary succession and Huitrera-Ventana volcanic deposits and consists of Miocene sedimentary and pyroclastic successions of the Ñirihuau, La Pava, Collón Curá, Caleufu and Chimehuin Formations (Figure 3).

4. Methods

We use U-Pb geochronology and in situ cosmogenic ^{10}Be surface exposure dating along with field and topographic surveys and structural relationships, to analyze the spatial patterns of deformation and their temporal variations in the North Patagonian Andes, and to evaluate a potential feedbacks between climate and deformation.

4.1. Field and Topographic Survey

To quantify deformation we surveyed terrace surfaces and lava plateaus or “mesetas.” Topographic data were obtained using a Stonex differential GPS. Additionally, we measured topographic profiles from digital elevation models (ALOS – PALSAR DEM, Alaska Satellite Facility [ASF]) to enlarge our survey coverage. Folded surfaces were restored to the initial horizontal state. Vertical offsets were calculated by assigning the origin of the profile ($x = 0$) to the foot of the fault scarp and by fitting straight lines to the survey points along the hanging wall and the footwall. Local variations attributed to erosion or folding, as well as erratic points were omitted from the best fit calculations.

4.2. Detrital Zircon U-Pb Geochronology

Eight samples were collected from volcanoclastic horizons and tuff layers from different Neogene growth stratal successions at key sites in the PBF. All eight samples were analyzed for detrital zircon U-Pb geochronology. Zircons were separated from samples following conventional physical and chemical mineral density separation techniques, and dated with laser ablation-inductively coupled key plasma-mass spectrometry (LA-ICP-MS) at ETH Zürich, following the method described by Beltrán-Triviño et al. (2013). We carried out a total of 1,060 age determinations, and we considered for analysis only concordant or sub-concordant results ($n:875$). Detrital zircon U-Pb isotopic data are provided in the Supporting information along with

Figure 4. Plots showing the youngest detrital zircon U-Pb age populations for individual samples. Maximum depositional ages were constrained using the youngest ages of two of more individual zircon grains that overlap at the 2σ level. MSWD, mean square weighted deviation; WMA, weighted mean age.

concordia plots of U-Pb ages and supplementary discussion. Plots showing the youngest detrital zircon U-Pb age populations for individual samples are represented in Figure 4. Frequency histograms and relative probability plots of U-Pb ages are shown in Figure 5. Analytical results are provided in Table S1.

4.3. Cosmogenic Dating

We sampled quartz-rich boulders and cobbles from the extensive Caleufu outwash plain located at the foothills of the cordillera (Figures 2 and 6), in order to determine the exposure age of this deformed surface. The surface is associated with a piedmont-style moraine lobe that may record a major expansion of the Patagonian Ice Sheet throughout the Quaternary. All samples ($n = 3$) were taken from flat and planar locations far from topographic highs and human disturbance. Samples were crushed and sieved. Quartz was purified chemically by following standard procedures according to Kohl and Nishiizumi (1992). $^{10}\text{Be}/^9\text{Be}$ ratios were measured at the ETH Zurich Accelerator Mass Spectrometry System TANDY (Christl et al., 2013; Müller et al., 2010). The ratios were normalized to the ETH in-house secondary standard S2007 N with a nominal value of $^{10}\text{Be}/^9\text{Be} = 28.1 \times 10^{-12}$. We used a ^{10}Be half-life of 1.387 ± 0.012 Ma (Chmeleff et al., 2010;

Figure 5. Detrital zircon U-Pb age data for eight samples of Neogene stratigraphic units and a composite age distribution, shown as probability density plots (curves) and age histograms (rectangles). For each age distribution, maximum depositional ages (MDAs) that may approximate true depositional ages are indicated.

Table 1
Sample Details, Cosmogenic ^{10}Be Concentrations, and ^{10}Be Surface Exposure Ages

Sample I.D.	Sample type	Context	Latitude S (dd)	Longitude W (dd)	Altitude (m)	Thickness (cm)	Shielding factor ^a	Quartz mass (g)	$^{10}\text{Be}/^9\text{Be}$ concentration ($\times 10^{-12}$)	^{10}Be concentration ($\times 10^4$ at.g ⁻¹) ^b	Uncertainty in ^{10}Be concentration ($\times 10^4$)	^{10}Be age (ka) ^c	$\pm 1\sigma$ (int.)	$\pm 1\sigma$ (ext.)
CA12	Granitic boulder	Caleufu outwash	-40.42	-70.79	794	5	0.99	15.74	5.183	658.75	15.41	1,084.7	33.7	113.1
CA13	Granitic boulder	Caleufu outwash	-40.43	-70.96	1,023	5	0.99	21.47	3.250	296.71	7.00	343.1	8.8	29.4
CA14	Pebbles quartz	Caleufu outwash	-40.43	-70.88	874	3	0.99	30.00	14.407	812.43	16.20	1,312.8	37.0	144.0

^aCalculated from the Geometric shielding calculator, version 1.1. (http://hess.ess.washington.edu/math/general/skyline_input.php). ^bBlank-corrected ^{10}Be concentrations. Normalization was done using the ETHZ Zurich standard S2007 N (nominal value of $^{10}\text{Be}/^9\text{Be} = 28.1 \times 10^{-12}$), which is a secondary standard normalized to ICN 01-5-1 (Nishiizumi et al., 2007). ^c ^{10}Be exposure ages calculated under the assumption of no erosion with the CRONUS-Earth ^{10}Be - ^{26}Al calculator, version 3.0 (Balco et al., 2008) using a ^{10}Be half-life of 1.387 Ma and a rock density of 2.65 g/cm³. We applied the time-dependent scaling scheme of Lal (1991) and Stone (2000), which takes temporal variations in the magnetic field intensity into account.

Korschinek et al., 2010). Exposure ages were calculated using the CRONUS-Earth online calculator (version 3.0) and the time-varying production model of Lal (1991) or Nishiizumi et al. (1989) (Lm) with a SLHL reference ^{10}Be production rate of 4.00 ± 0.32 atoms g⁻¹ yr⁻¹ (Best-fitting production rate of ^{10}Be by neutron spallation in atoms per gram of quartz per year from Borchers et al. (2016), uncertainty from Phillips et al. (2016)). We choose to report erosion uncorrected ages, so that exposure ages represent minimum estimates. Analytical results are provided in Table 1.

5. Records of Deformation

5.1. Field Evidence of Synorogenic Strata

Growth stratal successions are common features in the North Patagonian forelands. Since growth strata are stratigraphic intervals that were deposited during deformation, their ages can be used to constrain the timing of deformations. Below, we describe key areas where the most significant geometries of the growth sequences were mapped in detail. While some of these geometries have been described previously, others are new.

Growth strata evidence has been reported in the Catan Lil basin (Figure 6; García Morabito et al., 2011). At its eastern edge, the Neogene infill shows a series of progressive unconformities developed over the western limb of a west-verging anticline. Here, dips of strata decrease from steeply overturned at the bottom to subhorizontal at the top (Figure 7, Site 1). Throughout the section, the strata show wedge geometries, confirming that deposition was simultaneous with motion along an east-dipping thrust structure responsible for the growth of the Las Coloradas anticline (Figure 6). Similar relationships are observed in seismic reflection profiles across the Catan Lil basin (García Morabito et al., 2011).

The Sañico Massif is a basement high bounded by steeply dipping reverse faults (Figure 6). Two Cenozoic depocenters are located on either side of this crustal block. Strata with progressive unconformities are found associated with contractional structures at the western and eastern margins. Close to Piedra Pintada (Site 2), growth strata within the Neogene formations are associated with motion along an east-vergent thrust structure responsible for the uplift of this basement high (Figure 7, Site 2). To the west, the boundary between the Collón Cura basin and the Sañico Massif is defined by a west-verging fault system that uplifts basement rocks over the sedimentary units of the Caleufu Formation (Figure 6). Evidence for growth strata has been found at the footwall of this structure (Figure 7, Site 3a). Within the Caleufu Formation (Figure 3), that record can be linked to motion along this east-dipping fault. A few kilometers to the north, progressive unconformities within the La Pava and the Collón Cura Formations have also been described in the hanging wall of a NNE-trending splay related to the main fault (López et al., 2019). These have been interpreted as syncontractional deposits coeval to the uplift of the Sañico Massif. At this site, the Caleufu Formation shows no evidence of deformation (López et al., 2019). Further evidence is seen within the Collón Cura basin, where contractional structures affecting the upper Miocene sections in the Junín de los Andes area are locally associated with progressive unconformities.

Farther south, growth strata within the Collón Curá Formation (M. E. Ramos et al., 2011), are preserved at the eastern slope of the Cordón del Maitén Range (Figure 7, Site 4), a north-trending range constructed during Neogene inversion of late Eocene-Oligocene normal faults. This section presents progressive unconformities that range from 80°E to 10°E at the bottom and top, respectively (Figure 7, Site 4). Growth stratal packages in the upper levels of the Ñirihuau Formation have also been described in the Esquel Range of the North Patagonian foothills (Figure 2, Site 5) (Butler et al., 2019). Here, progressive unconformities have been found in association with a basement cored anticline (Echaurren et al., 2016). Surface growth stratal relationships are also reported within the Gastre and Paso del Sapo Basin (Bilmes et al., 2013; Butler et al., 2019). Thinning of the La Pava and Collón Cura Formations can be tied to middle Miocene displacement along the trailing basement structures (Sierra de Taquetren, Figure 2, Site 7).

Figure 6. Geomorphological and structural features of the North Patagonian foreland between 39° and 40°30'S showing evidence of deformation in the widespread lavas identified as Basalto II and locations of the topographic profiles in Figure 9. Whole rock K-Ar ages from Valencio et al. (1970), González Díaz et al. (1990), Linares et al. (1991), Vattuone and Latorre (1998), and Escosteguy and Franchi (2010).

Figure 7. Field photographs and line drawings of growth stratal packages. Stratal dip values and maximum depositional ages (MDAs) for samples are shown for pre-growth and growth strata. (a) Site 1: upper Miocene levels of the Chimehuin Formation exposed associated to a west-verging anticline at the eastern edge of the Catán Lil depocenter (after García Morabito et al., 2011). (b) Site 2: Miocene packages of the La Pava, Collón Cura, and Caleufu Formations in the broken foreland adjacent to the eastern margin of the Sanico Massif. (c) Site 3: Alicura fault at the eastern edge of the Collón Cura Basin and associated syngrowth strata in the upper Miocene levels of the Caleufu Formation. (d) Site 3a: Alicura fault. (e) Site 4: late Miocene upper levels of the Collón Cura Formation in the broken foreland region adjacent to the Cordón del Maitén (after M. E. Ramos et al., 2015).

5.2. Geomorphic Signature of Deformation

The extra-Andean lava fields of Patagonia form extensive flat-topped plateaus or mesetas, which cover large areas of the foreland, resting upon Mesozoic-Cenozoic volcanoclastic and sedimentary rocks. Although deformation in Patagonia is frequently assumed to predate the eruption of the late Miocene and early Pliocene basalts, folds and faults also occur in the volcanic cover (García Morabito & Ramos, 2012; Huyghe et al., 2015). Striking compressive deformation is seen in the widespread lavas identified as Basalto II (ca. 7.6–3 Ma), a large part of which is folded into anticlines and synclines associated with Andean thrusts (Figure 8). This basalt has been dissected into several relicts that rest unconformably on the Neogene syngrowth strata.

Lava flows are not ideal markers of deformation. This is because it is difficult to assess whether the flow ran over a pre-existing disrupted surface or whether it was disrupted by activation of an underlying fold or fault. In the studied area, however, multiple observations may indicate post-extrusion faulting. (i) At certain localities (Figure 9, P1 and P2), for instance, surface rupture of lava flows related to faulting produce offsets

Figure 8. Field photographs and line drawings of deformed basalts (Bii: Basalto II) and Quaternary deposits. (a–c) Morphological signature of the La Esperanza thrust. (d) Morphological signature of the La Pintada thrust. N-S-trending scarp with a high of ca. 250 m associated with the La Pintada thrust. It is difficult to discriminate between the pre-existing component of the topographic break and its reactivated part. (e) East facing scarp in outwash deposits, western Collón Cura Basin.

between 170 and 115 m (Figure 8). (ii) Topographic surveys reveal that tilting over certain structures has locally reversed the original slope of the lava flows (Figure 9, P2 and P4), indicating active uplift after its emplacement. (iii) Several basaltic lava remnants are located at higher elevations than their source areas, suggesting post-extrusion uplift (Figure 9, P3) (Huyghe et al., 2015).

The most striking evidence of deformation is seen in the Catan Lil depocenter, where basaltic flows are deformed by fault-related folds. A thin basaltic lava flow unconformably covers a westward verging anticline that folds Cretaceous sedimentary rocks (Mendoza Group). At its northern termination, the overlying lavas are folded and faulted, producing an offset of nearly 170 m (Figures 8c and 9). The lavas at the footwall dip between 25°W next to the fault trace to 5°W immediately to the west. An asymmetric anticline is developed in the hanging wall. A few kilometers to the south, the forelimb of the same anticline coincides with a scarp that cuts the plateau lavas, producing a vertical offset of 115 ± 4 m (Figure 8b). Further evidences of deformation include a series of short wavelength synclines, which can be interpreted as monocline structures associated with motion along N-S trending thrusts (Figure 8a). Similar evidence has been observed at the eastern margin of the main basaltic relict, where two regional topographic escarpments (Fria de los Guanacos and Piedra Pintada) coincide in position and trend with two westward dipping faults associated with uplift of the Sañico Massif (Figure 6). The southern scarp produces a well-defined topographic break of 247 m (Figure 9d). At the hanging wall, the Santo Tomas basalt gently dips to the west. Next to the fault trace the plateau lavas dip ca. 25°W. Farther to the north, a 30–70 m height N-S-trending scarp has been interpreted as the surface expression of the east-verging Fria de los Guanacos thrust (Huyghe et al., 2015). A further topographic escarpment is seen on a plateau located at the eastern margin of the Collón Cura river. Here, a NW-trending scarp produces a topographic break of 50 ± 1 m, and coincides in position and trend with the west-verging Badurial fault (Figure 6).

Figure 9. Topographic profiles from digital elevation models (ALOS—PALSAR DEM, Alaska Satellite Facility [ASF], spatial resolution of 12.5 m) and DGPS survey profiles. See location on Figure 6.

A further line of evidence that may record post-Miocene deformation comes from the occurrence of tilted fluvio-glacial deposits covering the vast extra-Andean plains or pampas. In the Collón Cura basin, several meters thick gravel deposits rest above the Miocene Calefú Formation. These consist of unconsolidated cobbles and boulders of Andean origin, and can be interpreted as a former outwash plain because of the high quantity of boulders and the lack of sorting and stratification. This surface is now dissected into different relicts by the modern drainage network. Remnants are particularly well preserved in the southern portion of the basin, where they can be directly linked to a degraded piedmont-style moraine lobe that marks former ice limits (Figure 10). At the western edge of the Collón Cura basin, between the Calefú and Limay rivers, a former outwash plain has been tilted, defining a N-S-striking fold (Figure 10). Our topographic surveys south of the Calefú river reveal that westward tilting over a former fault has locally reversed the original eastward slope of the outwash surface (Figure 10). The geometry of this structure is unknown, but could be explained by a westward dipping structure with reverse motion. This blind fault has displaced the outwash surface and forms a 20-km long east facing scarp with vertical displacements between ca. 100 and 50 m (Figure 8e). Restoration of this surface to its initial horizontal state allows us to estimate a shortening value of 30 ± 3 m (Table 2).

6. Geochronological Constraints

6.1. Detrital Zircon U-Pb Geochronological Results and Interpretations

6.1.1. Age of Growth Strata

The youngest U-Pb ages of zircon grains in populations of detrital zircons are commonly used to constrain maximum depositional ages (MDA) of stratigraphic units (e.g., Dickinson & Gehrels, 2009). Previously pub-

Figure 10. Geomorphic records of deformation. (a) DEM (ALOS—PALSAR DEM, Alaska Satellite Facility [ASF], spatial resolution of 12.5 m) and superimposed geomorphological and structural features of the southwestern Collón Cura Basin. Black dots indicate cosmogenic nuclide sampling sites. Yellow lines define traces of topographic surveys. (b) Topographic survey profiles across the tilted fluvio-glacial deposits that conform the Caleufu outwash surface. Survey points were projected onto W-E vertical planes.

lished and eight new estimates of MDAs are presented from samples from growth stratal packages that include the La Pava, Ñirihuau, Collón Cura, Chimehuin, and Caleufu formations (Figures 3 and 4). The presence of a nearby active volcanic arc means that youngest zircons are likely to be close to the true stratigraphy age.

Samples from a complete growth sequence associated with the steeply dipping frontal forelimb of the Las Coloradas anticline (Site 1) yield MDAs of 10.21 ± 0.18 Ma (LC01), 9.81 ± 0.85 Ma (LC06), and 7.55 ± 0.13 Ma (LC02). Two of these samples (LC01 and LC06) were collected from the basal Chimehuin Fm., the third one (LC02) was collected next to the top. These ages indicate continuous deposition during late Miocene motion along basement faults that propagate into the cover strata. In the Collón Cura Basin, a sample from a tuff layer intercalated within the intermediate levels of the Caleufu Formation (PO02) yields an age of 10.30 ± 0.11 Ma (PO02). At this site, the stratal packages are nearly horizontal, but growth strata within the same sequences are seen a few kilometers to the north, where they are associated with medium-scale reverse faults. At the eastern border of the Sañico Massif (Site 2), samples from two growth sequences linked to motion along the east-directed Piedras Paradas fault yield MDAs of 16.93 ± 0.27 Ma (SA01) and 9.66 ± 0.21 Ma (SA02). Farther south, at the eastern slope of the Cordón del Maitén Range (Site 4), samples from pre-growth and overlying growth strata within the Collón Cura Formation yield MDAs of 13.17 ± 0.18 Ma (MA02) and 10.57 ± 0.16 Ma (MA03), respectively. These U-Pb results are slightly older than a recently reported MDA of 10.1 ± 0.2 Ma and a single grain age of 9.3 ± 0.3 Ma (Butler et al., 2019; samples 17CUS04 and 17CUS05). MDAs from pre-growth and overlying growth strata within the Ñirihuau Fm. have also been reported for the Esquel Range, located in the frontal segment of the fold-thrust belt (Site 5). These indicate MDAs of 13.1 ± 0.2 Ma (17ESQ02) and 12.5 ± 0.3 Ma (17ESQ03) for the pre and syn-growth strata respectively (Butler et al., 2019). Farther east, Bucher et al. (2019) described growth strata in the La Pava and Collón Cura formations in the footwall of the southwest-verging Taquetren thrust (Site 6), from which they obtained MDAs of 14.45 ± 0.13 Ma (PBG) and 11.6 ± 0.4 Ma (PLG).

6.1.2. Sediment Provenance Results and Interpretation

U-Pb detrital zircon ages also provide insight into the development and relative importance of the sediment source areas. Previous provenance studies have delineated two main sediment source regions (western and

eastern) that could have delivered clastic sediment to the Patagonian foreland (Butler et al., 2019). These include four major western sources in the scattered outcrops of the Colohuincul Complex (420–345 Ma), Subcordilleran Batholith (~185–180 Ma), North Patagonian Batholith (~170–80 Ma; <20 Ma), and El Maitén Belt (~37–20 Ma), and eastern sources in the pre-Andean basement units (~400–364 Ma; ~370–310 Ma (Franzese, 1995; V. A. Ramos et al., 2010), subordinate exposures of magmatic arc units (~310–280 Ma; ~75–61 Ma), and intraplate volcanic units (Pilcaniyeu Belt [~60–42 Ma]).

Eight samples from the Neogene basin fill yield varied U-Pb results with multiple age distribution (Figure 5). Although there are variations among samples, many of these signatures can be linked to discrete sources. These include a pre-Andean bedrock, the Early Jurassic Subcordilleran Batholith, the Cretaceous North Patagonian Batholith, Paleogene intraforeland volcanic belts, and the Neogene Andean magmatic arc.

Miocene zircon grains, including syndepositional grains, dominate the detrital zircon components in most of the samples, which we interpret as the result of continued input from the contemporaneous Andean

Table 2
Vertical Offsets Derived From Topographic Surveys (Figures 9 and 10), Shortening Values, Ages, and Deformation Rates

Fault/Fold	Profile	Mean age $\pm$ uncertainty (Ma)	Source	Shortening (m)	Horizontal uncertainty (m)	Vertical offset (m)	Vertical uncertainty (m)	Fault dip ($^{\circ}$)	Dip uncertainty ($^{\circ}$)	Vertical slip rate (mm/yr)	Uncertainty (mm/yr)	Shortening rate (mm/yr)	Uncertainty (mm/yr)
Caleufu	P0i	1.31 $\pm$ 0.04	10 Be	30	3	97	1	NA	NA	0.07	0.00	0.02	0.00
Caleufu	P0i	0.51 $\pm$ 0.13 ^a	NA	30	3	97	1	NA	NA	0.19	0.05	0.06	0.02
La Esperanza N	P1	3.0 $\pm$ 0.5	K-Ar	64	6	NA	NA	NA	NA	NA	NA	0.02	0.00
La Esperanza N	P1	4.3 $\pm$ 0.1	K-Ar	64	6	NA	NA	NA	NA	NA	NA	0.02	0.00
La Esperanza N	P1	7.6 $\pm$ 0.2	K-Ar	64	6	NA	NA	NA	NA	NA	NA	0.01	0.00
La Esperanza N	P1	4.6 $\pm$ 0.5 ^b	NA	64	6	NA	NA	NA	NA	NA	NA	0.02	0.01
La Esperanza S	P2	3.0 $\pm$ 0.5	K-Ar	NA	NA	115	4	NA	NA	0.04	0.01	NA	NA
La Esperanza S	P2	4.3 $\pm$ 0.1	K-Ar	NA	NA	115	4	NA	NA	0.03	0.01	NA	NA
La Esperanza S	P2	7.6 $\pm$ 0.2	K-Ar	NA	NA	115	4	NA	NA	0.02	0.00	NA	NA
La Esperanza N	P2	4.6 $\pm$ 0.5 ^b	NA	NA	NA	115	4	NA	NA	0.03	0.00	NA	NA
Fria de los Guanacos	P3	3.0 $\pm$ 0.5	K-Ar	NA	NA	69	1	NA	NA	0.02	0.00	NA	NA
Fria de los Guanacos	P3	4.3 $\pm$ 0.1	K-Ar	NA	NA	69	1	NA	NA	0.02	0.00	NA	NA
Fria de los Guanacos	P3	7.6 $\pm$ 0.2	K-Ar	NA	NA	69	1	NA	NA	0.01	0.00	NA	NA
La Esperanza N	P3	4.6 $\pm$ 0.5 ^b	NA	NA	NA	69	1	NA	NA	0.02	0.00	NA	NA
Piedra Pintada	P4	3.0 $\pm$ 0.5	K-Ar	NA	NA	247	1	NA	NA	0.08	0.01	NA	NA
Piedra Pintada	P4	4.3 $\pm$ 0.1	K-Ar	NA	NA	247	1	NA	NA	0.06	0.01	NA	NA
Piedra Pintada	P4	7.6 $\pm$ 0.2	K-Ar	NA	NA	247	1	NA	NA	0.03	0.01	NA	NA
Piedra Pintada	P4	4.6 $\pm$ 0.5 ^b	NA	NA	NA	247	1	NA	NA	0.05	0.01	NA	NA
Bandurial	P5	5.6 $\pm$ 0.2	K-Ar	NA	NA	50	1	NA	NA	0.01	0.00	NA	NA
Bandurial	P5	6.7 $\pm$ 0.3	K-Ar	NA	NA	50	1	NA	NA	0.01	0.00	NA	NA
Bandurial	P5	4.6 $\pm$ 0.5 ^b	NA	NA	NA	50	1	NA	NA	0.01	0.00	NA	NA
Las Coloradas anticline	NA	2.7 $\pm$ 0.2	U-Pb	1,055	105	NA	NA	40	5	0.33	0.07	0.39	0.05
El Mañen Range	NA	2.6 $\pm$ 0.2	U-Pb	3,000	300	NA	NA	NA	NA	NA	NA	1.15	0.14

Note. The basalts flows have been dated to an extrusive cycle lasting from 7.6 to 3 Ma. We therefore consider three possible scenarios when estimating deformation rates. For this, we combine the offsets with a preferred (4.3 $\pm$ 0.1 Ma), a maximum (7.6 $\pm$ 0.2 Ma), and a minimum age (3.0 $\pm$ 0.5 Ma). Displacements are also averaged over a time interval of 4.6 $\pm$ 0.5 Ma (extrusive cycle lasting from 7.6 to 3 Ma and sealing of deformation structures after 3 Ma). NA, Not applicable. K-Ar, Whole rock K-Ar ages from Valencio et al. (1970), González Díaz et al. (1990) and Linares et al. (1991).

^aTime elapsed between the deposition of the Cauñu outwash (1.3 Ma) and the undeformed lava flow covering this surface (0.8 Ma). ^bTime interval of 4.6 $\pm$ 0.5 Ma (extrusive cycle lasting from 7.6 to 3 Ma).

magmatic arc. The Cretaceous detrital component is also present in most of the samples and has a western provenance from the North Patagonian Batholith, with clusters of ~170–80 Ma (Aragón et al., 2011; Castro et al., 2011; González Díaz, 1982; Pankhurst et al., 1999; Rolando et al., 2004), although some Cretaceous grains from the northernmost samples (LC01, LC06, and LC02) may reflect detrital contribution from rising structures at the southwestern edge of the Neuquén Basin, which involve sedimentary units of the Mendoza Group. There is also an almost continuous presence of Early Jurassic grains (200–170 Ma), likely derived from the North Patagonian and/or from the Sub-cordilleran Batholith (Pankhurst et al., 1999), which is involved in the North Patagonian fold-thrust belt at these latitudes. Alternatively, the Jurassic ages from samples LC01, LC06, and LC02 could be derived from the reworking of sedimentary units of the Neuquén Basin (e.g., pre-Cuyo and/or Cuyo groups), which outcrop along the eastern edge of the Catan Lil basin, where they constitute the core of several anticlines. The Late Jurassic detrital component is subordinate and may have a mixed provenance.

A broad range of roughly 400–270 Ma ages can be linked to pre-Andean components. However, the sources for these age populations are ambiguous, as they may have originated from late Paleozoic rocks in the west or from basement heights in the east. Two samples from the Catan Lil basin (LC06 and LC02) are particularly notable for the large proportion of zircon grains from Late Paleozoic sources (50%) with a main peak at ~303 Ma, and secondary Late Permian and Late Devonian peaks. This population is likely derived from Permian (250–300) intrusive rocks and associated host rocks of Devonian age (~400–364 Ma; ~370–310 Ma) exposed at the present-day southwestern margin of the Neuquén Basin (Franzese, 1995; Lucassen et al., 2004; V. A. Ramos et al., 2010; Schiuma & Llambías, 2008). Both samples are further distinguished by a unique Paleocene group (70–60 Ma) that is not observed in any other samples in this study. The most likely source could be attributed to scattered outcrops of subordinate magmatic arc units (~75–61 Ma) present north and south of the Catan Lil basin (García Morabito & Ramos, 2012). The U-Pb results from the two southernmost samples of the Collón Cura Formation (MA02 and MA03) show similar basement signatures as the northernmost ones but in different proportions (Figure 5). The continuous presence of Late Paleozoic grains attests to input from pre-Andean sources. A progressive upsection increase of pre-Andean bedrock derived ages attest to incremental input from basement sources, likely related to the uplift and exhumation of basement blocks in the foreland and its consequent fragmentation. This trend is particularly notable in samples LC01, LC06, and LC02, which record an inversion of the detrital pattern that switches from western Cordillera to the eastern foreland at ca. 9.8 Ma.

In summary, the age spectra obtained from our 1060 U-Pb detrital zircon ages record a provenance from both western and eastern sources during the compressional stage, with a progressive increase of basement sourced clastics coming from the exhumation of rising basement blocks during the late Miocene. In the light of these new data, many of the structural highs present across the Andean foreland can be identified as late Miocene growth structures that feed several intermontane basins.

6.2. Age of the Basalto II Formation

Within the North Patagonian foreland, plateaus consisting of Cenozoic lava flows are locally tilted, and may therefore provide insight into the manner in which deformation migrates in space and time. It is accepted that Cenozoic volcanism in the extra-Andean region of North Patagonia occurred in cycles (Valencio et al., 1970). Within the region of interest, the Late Miocene—Early Pliocene volcanic products that rest unconformably on the Miocene sediments were commonly included in the Basalto II Formation (Groeber, 1929). In the literature, these basalts are also locally referred to as Santa Isabel, Coyochó, Chenqueniyeu, or Tipilihuque basalts (Leanza & Hugo, 1997; Turner, 1965). Within the Aluminé depocenter, the basaltic lava flows that cover the Late Miocene sedimentary succession have been dated between 6.2 and 3.4 Ma (Re et al., 2000; Vattuone & Latorre, 1998). In the area located between the Collón Cura and Catan Lil depocenters, the Santa Isabel basalt flows constrain an extrusive cycle lasting from 7.6 to 4.3 Ma, with some older and younger outcrops of 8.5 and 3 Ma (Franzese et al., 2018; González Díaz et al., 1990; Linares et al., 1991; Valencio et al., 1970; Figure 6). These are slightly older than three ages varying between 4.2 and 3.7 Ma from similar basaltic flows located at the western edge of the Collón Cura Basin (Escosteguy & Franchi, 2010).

6.3. Cosmogenic Surface Exposure Results

The ^{10}Be exposure ages are presented in Table 1. Exposure ages assume no erosion. Uncertainties correspond to the internal 1σ . Two boulders from the Caleufu outwash terrace in the Collón Cura Basin yield ages of 343 ± 9 ka (CA13) and $1,085 \pm 34$ ka (CA12). A third sample consisting of amalgamated quartz pebbles ($n = 90$) from the same surface yielded an exposure age of $1,313 \pm 37$ ka (CA14). The boulder giving the youngest age was probably exhumed from depth any time after the deposition of the fluvioglacial deposits. The two older samples give ages consistent with a K-Ar age that bracket the timing of the Caleufu outwash (Linares & González, 1990). Specifically, their ages ($1,085 \pm 34$ ka and $1,313 \pm 37$ ka) are consistent with the age of a basaltic flow (Sample AK-1567: 0.8 ± 0.4 ka) that overlies an equivalent surface ~ 20 km to the south (Linares & González, 1990).

With the exception of prior exposure to cosmic rays, which results in an overestimate of the true age of the sampled landform, geological factors like erosion generally reduce cosmogenic nuclide concentrations in rock surfaces, resulting in apparent surface exposure ages that underestimate the true depositional age of the sampled landform. In this regard, the oldest sample (Sample CA14: $1,313 \pm 37$ ka [± 144 ka external]) is considered closest to the age of deposition, because nuclide inheritance is expected to be low in outwash sediment of Patagonia (Cogez et al., 2018; Hein et al., 2009), and because of field evidence for deflation, which could potentially reduce the exposure age recorded by the sample. If the arithmetic mean of the two oldest samples is considered as the best approximation to the timing of deposition, then the formation of the Caleufu outwash dates to $1,199 \pm 161$ ka. Our results are consistent with ^{10}Be and ^{26}Al surface exposure ages from outwash surfaces of southern Patagonia, which approximate the timing of J. Mercer (1976) Greatest Patagonian glaciation Glaciations (GPG) as ca. 1.2 Ma (Hein et al., 2011).

7. Rates of Deformation Among Individual Structures

Geochronologic data, in combination with balanced cross sections and topographic surveys, constrain the rate of motion on specific structures in the foreland of North Patagonia. All results are compiled in Table 2.

Late Miocene shortening magnitudes were obtained from the reconstruction of two balanced cross sections at sites 1 and 4 (Figures 2 and 11). These are sections previously described in García Morabito et al. (2011) and D. L. Orts et al. (2015), which are based on seismic-profile information, together with field and surface data. Geologic cross section A crosses the Las Coloradas anticline, a basement-cored anticline developed at the eastern margin of the Catan Lil basin (Figure 11a). The restoration of this section indicates 1,055 m of shortening. Section B crosses the Cordón del Maitén, a north-trending frontal antiform related to a basement-involved wedge structure (Figure 11b). Restoration of this cross section yielded a total shortening of 3,000 m. This value is slightly higher than a previously reported magnitude of 2,500 m (M. E. Ramos et al., 2011). Taking into account these magnitudes, the U-Pb age analyses of horizons from lower and upper levels of the associated growth stratal successions, we estimated shortening rates at both sites. These rates should be interpreted as maxima, because some shortening could have occurred before the deposition of the growth strata packages, and because the values encompass several faults averaged in space. Combining a shortening of $1,055 \pm 25$ m across section A, with a time interval of 2.7 ± 0.2 Ma, yields a maximum shortening rate of 0.4 ± 0.3 mm yr $^{-1}$. If the 3,000 m of shortening obtained from section B corresponds to the horizontal component absorbed by the Cordón del Maitén anticline, and assuming that this antiform acquired its present configuration between 13.17 and 10.57 Ma, then the maximum shortening rate for this feature is 1.15 ± 0.2 mm yr $^{-1}$.

Further constraints are derived from reconstructed geomorphic offsets recorded by deformed lava fields and tilted fluvioglacial deposits. In this case, the ages are interpreted to be maximum ages with regard to calculated deformation rates, because offset must have occurred at some time after deposition or volcanic extrusion. As such, the slip rates should be interpreted as minima. At particular sites, however, it is difficult to discriminate between the pre-existing component of the escarpment and its reactivated part. In Bandurial and Piedra Pintada (profiles 4 and 5, Figure 9), for instance, where well-defined topographic breaks are related to thrusts that partly pre-date the basaltic extrusion, it is highly probable that the lava flows ran over a pre-existing scarp. Hence, the observed offsets at these particular sites would represent an upper limit for the latest Miocene—early Pliocene increment in deformation. In sum, the deformation

Figure 11. Balanced cross-sections across the west-verging Las Coloradas anticline (a) and the Cordón del Maitén (b). These are sections previously described in García Morabito et al. (2011) and D. L. Orts et al. (2015), which are based on seismic-profile information, together with field and surface data.

rates estimated on structures responsible for the tilting and offset of different lava flows range between 0.01 and 0.08 mm yr^{-1} .

If the displacements are averaged over a time interval of $4.6 \pm 0.5 \text{ Ma}$ (extrusive cycle lasting from 7.6 Ma to 3 Ma and sealing of deformation structures after 3 Ma), the vertical rates range between 0.01 and 0.05 mm yr^{-1} (Figure 12). These values are one order of magnitude smaller than the deformation rates calculated for late Miocene active structures

We also constrained the deformation rate for a blind thrust that folded an outwash surface at the mountain front by combining topographic profiling and ^{10}Be exposure dating (Figure 10). Topographic profiles yield a maximum vertical offset of $97 \pm 1 \text{ m}$, and restoration of this surface to its initial state allows us to estimate a shortening value of $30 \pm 3 \text{ m}$ (Table 2). If the oldest sample (Sample CA14: $1,313 \pm 37 \text{ ka}$ ($\pm 144 \text{ ka}$ external)) is taken as closest to the age of deposition, then the vertical deformation rate is $0.074 \pm 0.002 \text{ mm yr}^{-1}$, with a horizontal slip rate of $0.023 \pm 0.002 \text{ mm yr}^{-1}$ (Figure 12). However, if the measured displacements are averaged over $0.51 \pm 0.13 \text{ Ma}$, the time elapsed between the deposition of the Caleufu outwash (1.3 Ma) and the undeformed lava flow covering this surface (0.8 Ma) (Figure 6), then the vertical and horizontal rates increase to $0.19 \pm 0.05 \text{ mm yr}^{-1}$ and $0.06 \pm 0.02 \text{ mm yr}^{-1}$, respectively. Although relatively low, these rates are one order of magnitude higher than those determined on latest Miocene—Pliocene active structures (Figure 12).

8. Discussion

8.1. Spatial-Temporal Evolution of Growth Structures

Depositional age constraints for different growth stratal sequences, together with geomorphic evidences of deformation along the foreland region of Northern Patagonia, provide a refined view on the spatial-temporal evolution of growth structures along the eastern flank of the North Patagonian orogen (Figure 13). Our data record several deformational phases since the early Miocene characterized by a progressive decrease of upper-plate contraction after an initial period of widespread deformation (Figure 12). Syntectonic sediments in the eastern foreland record a main and relatively sustained pulse of shortening in a narrow fold-thrust system governed by reactivation of earlier extensional structures during the middle to late Miocene ($\sim 17\text{--}7 \text{ Ma}$). Widespread shortening commenced no later than 13 Ma , although early deformation is locally recorded in distal regions of the broken foreland by ca. 17 Ma (Figure 13). Between 13 and 10 Ma , deformation recorded by growth stratal geometries seems to have been distributed simultaneously over the entire

Figure 12. Late Cenozoic variations in deformation rates. Cosmogenic ^{10}Be and U-Pb ages plotted against vertical offsets (gray) and shortening (blue). Slowdown in deformation rates coincides with the onset of widespread glaciation.

retroarc segment (Figures 13 and 14), with no clear signal of progressive thrust propagation. Shortening in the frontal thrust belt foothills started at 13–12 Ma (Esquel Range), coeval with early and ongoing uplift along structures in the broken foreland farther east (Cordón del Maitén, Sanico Massif, Taquetrén Range), and continued through the late Miocene. Late Miocene basement reactivation and associated thrusting is recorded by growth stratal successions expressed in the Lonquimay, Catan Lil, Colón Cura, and Piedra del Aguila basin fill. This evidence indicates that deformation continued during the late Miocene, with important slip focused along several reverse faults that uplifted basement rocks at least during the ca. 10–7.5 Ma time window (Figures 13 and 14). Final retroarc basement partitioning is, therefore, a result of widespread late Miocene growing structures, with individual fault activity and related block uplift varying irregularly in space and time. This is in accordance with sediment provenance patterns, which indicate clastic systems progressively fed by Paleozoic crystalline rocks of the pre-Andean basement, associated with incremental uplift and exhumation of basement blocks. The growth stratal relations are also in accordance with available low-temperature thermochronological data, which demonstrate exhumation-related late Miocene – Pliocene cooling along the main Cordillera and the frontal fold-thrust belt (Savignano et al., 2016), that is also supported by high sedimentation rates registered in the broken foreland (Bucher et al., 2019).

The late Miocene to early Pliocene transition represents a time interval of progressive change in Patagonia. From the latest Miocene onward, continental sedimentation nearly stalled. The Miocene synorogenic units are covered by a thin carpet of piedmont and glaciofluvial gravel deposits and widespread basaltic lava flows. Since deformation is not recorded stratigraphically, there has been an assumption that tectonic growth in the foreland of northern Patagonia ended by this time (Thomson et al., 2010). There are, however, multiple lines of geomorphic evidence that suggest that deformation in the broken foreland continued into the Pliocene. This younger phase of reactivation of basement structures is recorded by faulted and folded basalts, which have been affected by deformation at least during a 7.6–3 Ma window. During this time interval, low-magnitude shortening in the broken foreland was synchronous with tectonic activity in the frontal thrust belt foothills, as inferred from low-temperature thermochronological data (Savignano et al., 2016), but may have been inhibited in distal foreland positions south of 41°S, where robust evidence of shortening is lacking. Our data indicate that during this stage, motion on individual structures along the broken foreland occurred at least 10 times slower compared to the late Miocene (Figure 12).

We propose that low-magnitude deformation in the broken foreland continued until the end of the extrusive cycle represented by the Basalto II flows. The timing of this later stage of deformation is bracketed by the deformed Basalto II flows (7.6–3 Ma) and the widespread sealing of deformation structures by a younger extrusive cycle dated between 3 and 0.8 Ma, with four ages around 2.3 Ma (Linares & González, 1990; Linares et al., 1991; Rabassa, 1975). By the end of this phase, we observe abandonment of distal faults and renewed activity at the mountain front reflected by deformation of an outwash surface after ca 1.3 Ma and coeval uplift and denudation at the core of the orogen.

Collectively, previous and new chronological constraints are indicative of a progressive decrease of upper-plate shortening in the foreland of Patagonia. The initial period of intense deformation and uplift (ca. 13–7 Ma) (Figure 13) was followed by a structural reorganization characterized by reduced recent activity on thrust faults in the foreland, and renewed activity in the hinterland close to the orogen's core. This leads to the conclusion that upper-plate shortening in the foreland of North Patagonia has been extremely slow or absent throughout the last few million years. The cessation of fault activity is consistent with a lack of shallow crustal seismicity, and represents an orogen-wide shift in deformation away from the external foreland region. This indicates that the conditions for upper-plate deformation varied across the convergent margin during the Neogene.

Figure 13. Spatial patterns of deformation in the North Patagonian Andes. Cross sections through the North Patagonian foreland (for location see Figure 2) showing detrital zircon U-Pb ages, ^{10}Be surface exposure data and projected low-temperature thermochronometric data from Savignano et al. (2016). Cross section a is based on García Morabito et al. (2011). Cross sections b and c are based on D. L. Orts et al. (2015). Cross section d is based on Echaurren et al. (2016).

8.2. Role of Climate in Orogenic Deformation

Theoretical analysis predicts that enhanced erosion related to spatial and temporal variations in climate can influence the internal dynamics of mountain building, forcing a structural reorganization within a critical-taper orogenic wedge, and leading to a reduction in orogen width and height (e.g., Tomkin & Roe, 2007; Whipple, 2009; Whipple & Meade, 2006). Although both, analogue and numerical models can be effectively used as a guide to the tectonic changes expected to occur within a mountain range in response to climate change, compelling field evidence has proved challenging (Whipple, 2009). So far, direct evidence of a coordinated response of an orogen to climate change has been demonstrated in the European Alps and the St. Elias range of Alaska (Berger et al., 2008; S. D. Willett et al., 2006). More recently, Thomson et al. (2010), used an extensive low-temperature thermochronologic data set from the Patagonian Andes to argue that the internal dynamics of mountain building and the reduced height and width of the orogen north from 49°S were largely influenced by late Cenozoic glaciations. However, the timing and distribution of Andean deformation in the back-arc region, crucial for diagnosing a tectonic response to climate change, is only loosely constrained. Our results complement snapshots of cooling ages, and provide new empirical evidence of an active tectonic response to climate cooling in Northern Patagonia. The pattern of Neogene “stepwise” evolution of the Patagonian foreland region closely matches the onset of Patagonian glaciation at 7 Ma and the successive acceleration of erosion rate (Herman & Brandon, 2015; Thomson et al., 2010; C. D. Willett

Figure 14. Composite diagram showing the correlation of motion along individual faults (gray bars) with rock exhumation in North Patagonia, global climate evolution, plate kinematic parameters, and sedimentation rates at the Argentine margin. U-Pb ages (this study; Bucher et al., 2019; Butler et al., 2019; Niviere et al., 2019) and surface exposure ages (this study) are shown as black and blue dots respectively. Apatite fission-track ages (blue squares) from Thomson et al. (2010), Apatite (U-Th)/He cooling ages (yellow dots) from Thomson et al. (2010) and Savignano et al. (2016). Instrumental surface earthquakes (<33 km depth) are represented by small gray dots. Global $\delta^{18}\text{O}$ marine isotope record redrawn after Zachos et al. (2001). Plate kinematic data from Maloney et al. (2013). Rates of sedimentation at the Argentine margin from Gruetzner et al. (2016). Al, Alicura; Ba, Bandurial; Ca, Caleufu; Esq, Esquel; FG, Fria de los Guanacos; Ga, Gastre; LC, Las Coloradas anticline; LE, La Esperanza; Ma, Cordón del Maitén; PP, Piedra Pintada; PS, Pino Solo; Tr, Taquetrén.

et al., 2020). Our interpretation is that acceleration in erosion rates during the past 7 Ma coeval with the onset of major Patagonian glaciation may have influenced not only mean elevation and relief, but also the regional geometry and kinematic history of the orogenic belt.

The Neogene structural evolution of Northern Patagonia suggests a coordinated response of the orogen to climate change. However, additional controls on spatial and temporal variations in upper-plate contraction may have intervened as well. The construction of the Andes, in general, has been governed largely by plate-scale changes in convergence and shifts in the geometry of the subducting oceanic slab (e.g., Folguera & Ramos, 2011; Horton, 2018). Although there seems to be no general straightforward correlation between deformational history and plate kinematic parameters, there is a further complication in Northern Patagonia, where the convergence rate has declined from ca. 15 Ma onward (Maloney et al., 2013). This decrease could explain progressive abandonment of the foreland front even without a tectonic response to climate change. However, this seems unlikely, given that convergence was already decreasing during late Miocene thrusting (Figure 14). Moreover, progressive decrease of convergence rate does not explain the Quaternary renewed slip recorded at the Caleufu outwash plain. In addition, the trench-normal absolute overriding plate velocity, which explains several key components of the Mesozoic-Cenozoic tectonic history in the Central and Southern Andes (Horton, 2018; Maloney et al., 2013; Vietor & Echtler, 2006), does not show any correlation with the deformational history across this segment, and thus appears to have had no significant

effect on foreland deformation. Another driving mechanism of variable mechanical coupling along the plate boundary is the variability of slab dip (Horton, 2018). Fluctuations in the geometry of the subducting slab, have commonly been invoked as a cause of variations in deformation and magmatic patterns in several regions east of the Main Andes (e.g., Folguera & Ramos, 2011). In North Patagonia, however, the observed Neogene foreland reactivation at 13–7 Ma occurred in a period where no significant arc shifting is recorded (Gianni et al., 2015). This rules out the possibility that shifts in the geometry of the subducting oceanic plate during this time span controlled spatial and temporal variations in upper-plate contraction. Further driving mechanisms involve the role of trench sediments in modulating stress transmission across the plate interface. South of the Juan Fernandez Ridge at 33°S, more than 1.5 km of terrigenous sediments fill the trench (Bangs & Cande, 1997). Most of these sediments are supplied from the southern region and re-distributed northward along the plate boundary (Völker et al., 2006). Thick layers of subducted sediment lubricate and thus promote weaker coupling along the plate interface (Lamb & Davis, 2003). It is highly probable that the sediment supply resulting from substantial exhumation in Northern Patagonia after onset of glacial conditions contributed to a decrease in compressive deformation, explaining the observed slowing in upper-plate contraction from ca. 7 Ma onward. Nonetheless, if the observed decline in shortening was driven by an increase in sediment flux to the trench, as is plausible, a similar response should be expected in the Southern Central Andes backarc region, where the thickness of trench fill exceeds 2 km (Hoffmann-Rothe et al., 2006). There is, however, multiple lines of evidence that attest to active tectonics at the front of the fold-thrust belt and in the broken foreland and ongoing deformation with deformation rates ranging between from 2.5 and 0.5 mm/yr (Branellec et al., 2016; Messenger et al., 2010). This suggests that even a thick trench fill along the Southern Central Andes margin was unable to sufficiently lubricate and weaken the plate interface to slow upper-plate shortening. Sediment supply to the trench may represent a second-order mechanism linked to climate change that partially regulated the several-phase Neogene history of North Patagonia. In the absence of further drivers, however, this fails to explain the recent evolution of neighboring Andean segments.

Conditions required for the onset of glacial erosion and denudation were achieved in the North Patagonian Andes by the latest Miocene–earliest Pliocene. The elevations required for the onset of glacial denudation were reached via mountain building processes by the end of the Miocene (Blisniuk et al., 2005), and cooling since ca. 6 Ma, as indicated by different proxies (Lear et al., 2000), is consistent with the onset of major Antarctic ice-sheet expansion at 7–5 Ma (Zachos et al., 2001), and the oldest glacial deposits in Patagonia (7.4–4.6 Ma) (J. H. Mercer & Sutter, 1982; Ton That et al., 1999; Figure 14). Before late Miocene glacial expansion, denudation and erosion rates in the Andean interior were relatively low (Savignano et al., 2016; C. D. Willett et al. 2020), deformation was distributed over the entire foreland, and sedimentation rates in the associated intermontane basins were high (Bucher et al., 2019; Figure 15). Hundreds of meters of Neogene sediment fill were deposited. This record suggests major subsidence in agreement with evidence for increasing uplift of basement blocks. After the onset of glacial conditions, however, erosion rates accelerated, coeval with an increase in sediment flux to the trench (Kilian & Behrmann, 2003; Melnick & Echtler, 2006; Thomson et al., 2010; C. D. Willett et al. 2020). These changes in the core of the northern Patagonian Andes and its forearc were accompanied by a marked slowing of foreland upper-plate shortening and a reduction in subsidence. From the latest Miocene onward, continental sedimentation nearly stalled. The Miocene syn-orogenic units are covered by a thin carpet of piedmont and glaciofluvial gravel deposits and by widespread basaltic lava flows. In contrast, the upper Miocene—lower Pliocene constitutes an aggradational period that is particularly well developed in the Argentine continental margin (Figure 14; Ercilla et al., 2019; Ghiglione et al., 2016; Gruetzner et al., 2016). The sediment eroded from the interior of the range likely bypassed Patagonia and reached the offshore basins, where they are represented by thick sequences with high sedimentation rates (Gruetzner et al., 2016). This regional reorganization of the sediment routing pattern includes evidence of diminished foreland subsidence and can be interpreted as the result of progressive foreland abandonment and sediment bypass during reduced activity on thrust faults (Figure 15). During this time interval, significant shortening seems to have been confined to more internal parts of the orogen, recorded locally by transpressional uplift along the LOFZ zone since 10–7 Ma (Rosenau et al., 2006), and exhumation ages younger than 7 Ma throughout the Andean interior (Savignano et al., 2016). This decrease of shortening across the foreland is unusual, given the expectation that a contractional mountain belt should grow continuously, accommodating the greatest fraction of convergence on the frontal thrusts (Davis et al., 1983).

In contrast to the slowing in upper-plate shortening the transpressional motion along the LOFZ seems to have accelerated since the Late Miocene as a consequence of oblique ridge collision (Adriasola et al., 2006). In this regard, it is highly probable that accelerated uplift efficiently balanced by glacial denudation contributed to sustain the observed high erosion rates. This combination of active block uplift efficiently balanced by glacial erosion might provide an additional mechanism for the shift of deformation.

The cessation of shortening in the broken foreland and the final narrowing of the deformation zone was not coeval with the onset of glacial conditions at 7.4–5 Ma, but rather with the intensification of glacial erosion since 2–3 Ma, perhaps with a shift toward more erosionally effective ice streams (Herman & Brandon, 2015). A modeling study identified an erosion hotspot in northern Patagonia around 2–3 Ma that may have resulted from the northward migration of the WWB during Pleistocene glaciations (Herman & Brandon, 2015). This increase is in line with a proposed global increase in late-Cenozoic erosion rates in response to a cooling climate (Herman et al., 2013). However, there remains much debate about the effect of the Cenozoic cooling of Earth's climate on the efficiency of erosion in orogenic systems (Schildgen et al., 2018). Alternative perspectives suggest that late-Cenozoic increase in erosion rates throughout mountainous landscapes derived from thermochronological data (Herman et al., 2013) appear to be an artifact related to combining data with disparate exhumation histories (Schildgen et al., 2018). This could be valid for the Northern Patagonian Andes, where steep gradients in thermochronological ages and exhumation rates are found. There, many thermochronological systems shows young ages close to the Liquine-Ofqui fault.

Alternatively, the Plio-Quaternary slowing or termination of the subduction orogeny could be reflecting a certain time lag between the climate forcing at 7.4–5 Ma and the orogen's response to the onset of glaciation. Once glaciation has started, the system may need a certain time to evacuate material and respond. Recent studies, for instance, place landscape response to glaciation between a few tens of thousands and a few millions of years (Herman et al., 2018; C. D. Willett et al., 2020). These response times are, however, difficult to address because the wide range of processes and parameters involved.

The retreat of the deformation front on the eastern foreland around 3 Ma, as indicated by the widespread sealing of deformation structures by Pliocene volcanics, seems to be followed by enhanced slip on interior faults after 1.3 Ma (Figure 15) as inferred from preliminary evidence for fault activity close to the interior of the orogen. Evidence of renewed slip in the Andean interior is implied by active folding and tilting of a Pliocene outwash surface at the eastern mountain front, with deformation rates close to the order of magnitude of those recorded on late Miocene thrust faults (Figure 12). This response is precisely what theoretical analysis predicts for convergent mountain ranges subjected to enhanced erosion (Dahlen & Suppe, 1988; Whipple, 2009; Whipple & Meade, 2006). Although persistent field evidence of enhanced deformation close to the core of the range remains elusive, given the absence of Neogene cover sequences in the Main Cordillera, the observed behavior seems to favor a case for a tectonic response to climate change in the North Patagonian Andes.

9. Conclusions

New U-Pb and ^{10}Be surface exposure data from North Patagonia provide a refined view on the spatial patterns of deformation, and their temporal variations during a major climatic transition. Our results complement snapshots of cooling ages and provide new empirical evidence at the scale of individual thrust faults that is relevant for diagnosing an active tectonic response to climate cooling. Based on the combination of these distinct datasets we draw the following conclusions:

1. Syntectonic sediments record a first period of widespread deformation and uplift during the ca 13–7 Ma time window, with local evidences of early deformation by ca. 17 Ma. During this time interval, deformation, as recorded by growth stratal geometries and low-temperature thermochronological data, was distributed simultaneously over the entire retroarc segment, with no clear signal of progressive thrust propagation
2. Final retroarc basement partitioning is recorded by a series of widespread late Miocene growth structures, with individual fault activity and related block uplift varying irregularly in space and time. This is in accordance with sediment provenance patterns, which indicate clastic systems progressively fed by pre-Andean basement sources and with high sedimentation rates registered in the broken foreland

Figure 15. Schematic description of Neogene evolution of Northern Patagonia showing tectonic structures and morphotectonic units during (a) Late Miocene deformation of the thrust belt and the broken foreland, (b) unroofing of the orogen's core coeval with onset of glacial conditions and decrease of upper-plate shortening; and (c) final narrowing of the deformation zone after ca. 3 Ma.

3. From the latest Miocene onward, our data record a progressive decrease of upper-plate shortening and slowing of the rate of subsidence in the foreland of Patagonia. Geomorphic evidence indicates that deformation continued into the Pliocene, but upper-plate contraction drastically slows after 7 Ma, and even terminates in the Pliocene. The final abandonment of the contractional deformation in the foreland after ca. 3 Ma seems to be followed by renewed fault activity close to the interior of the orogen at ca. 1.3 Ma
4. The pattern of Neogene episodic evolution of the Patagonian foreland region closely matches the onset of Patagonian glaciation at ca. 7 Ma and the successive acceleration of erosion rate values within the orogen interior. The observed correlation between timing of onset of glaciation (7.4–5 Ma), acceleration of erosion rates (7–5 Ma) and slowing in upper-plate shortening and subsidence (7–3 Ma) is followed by the abandonment of distal faults superseded by probable enhanced slip on interior faults, likely related with the intensification of glacial erosion since 2–3 Ma
5. Our interpretation is that acceleration in erosion rates during the past 7 Ma, after the onset of major Patagonian glaciation influenced not only mean elevation and relief, but also the regional geometry and kinematic history of the eastern flank of the orogenic belt, with deformation fluctuations responding to changes in climatic conditions

Data Availability Statement

All of the data have been archived at the Geochron data base (https://www.geochron.org/dataset/html/geochron_dataset_2020_12_21_p6eaB)

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