



HAL
open science

Deformation of the western Andes at 20–22°S: a contribution to the quantification of crustal shortening

Tania Habel, Robin Lacassin, Martine Simoes, Daniel Carrizo, Germán Aguilar

► To cite this version:

Tania Habel, Robin Lacassin, Martine Simoes, Daniel Carrizo, Germán Aguilar. Deformation of the western Andes at 20–22°S: a contribution to the quantification of crustal shortening. 2021. insu-03373164

HAL Id: insu-03373164

<https://insu.hal.science/insu-03373164>

Preprint submitted on 22 Nov 2021

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

Manuscript submitted to Tectonics :

Deformation of the western Andes at ~20–22°S: a contribution to the quantification of crustal shortening

Tania Habel – Université de Paris, Institut de physique du globe de Paris, CNRS, F-75005 Paris, France – habel@ipgp.fr

Robin Lacassin – Université de Paris, Institut de physique du globe de Paris, CNRS, F-75005 Paris, France – lacassin@ipgp.fr

Martine Simoes – Université de Paris, Institut de physique du globe de Paris, CNRS, F-75005 Paris, France – simoes@ipgp.fr

Daniel Carrizo – GeoEkun SpA, Santiago 7500593, Chile – carrizo@geoekun.com

Germán Aguilar – Advanced Mining Technology Center, Facultad de Ciencias Físicas y Matemáticas, Universidad de Chile, Avenida Tupper 2007, Santiago, Chile – german.aguilar@amtc.cl

This a non peer-reviewed paper.
This manuscript has been submitted for publication to Tectonics on
2020-12-17

Any comments can be sent to the corresponding
authors: habel@ipgp.fr

16 **Abstract**

17 The Andes are an emblematic active Cordilleran orogen. It is admitted that mountain-building in
18 the Central Andes, at $\sim 20^\circ\text{S}$, started $\sim 50\text{--}60$ Myr ago, along the subduction margin, and
19 propagated eastward. In general, the structures sustaining the uplift of the western flank of the
20 Andes are dismissed, and their contribution to mountain-building remains poorly solved. Here,
21 we focus on two sites along the western Andes at $20\text{--}22^\circ\text{S}$, in the Atacama Desert, where
22 structures are well exposed. We combine mapping from high-resolution satellite images with
23 field observations and numerical trishear forward modeling to provide quantitative constraints on
24 the kinematic evolution of the western Andes. Our results confirm the existence of two main
25 structures: (1) the Andean Basement Thrust, a west-vergent thrust placing Andean Paleozoic
26 basement over Mesozoic strata; and (2) the west-vergent West Andean Fold-and-Thrust-Belt,
27 deforming primarily Mesozoic units. Once restored, we estimate that both structures
28 accommodate together at least $\sim 6\text{--}9$ km of shortening across the sole $\sim 7\text{--}17$ km-wide
29 outcropping fold-and-thrust-belt. This multi-kilometric shortening represents only a fraction of
30 the total shortening accommodated along the whole western Andes. The timing of the recorded
31 main deformation can be bracketed sometime between ~ 68 and ~ 29 Ma – and possibly between
32 ~ 68 and ~ 44 Ma – from dated deformed geological layers, with a subsequent significant slowing-
33 down of shortening rates. Our results therefore reveal that the contribution of the structures of the
34 West Andes can no longer be neglected, in particular at the earliest stages of Andean mountain-
35 building.

36 **1 Introduction**

37 One of the most active convergent plate boundaries is located along the western margin
38 of South America (Figure 1). There, the oceanic Nazca plate plunges beneath the South
39 American continent, with a convergence rate currently of ~ 8 cm/yr at $\sim 20^\circ\text{S}$, according to the
40 NUVEL-1A model (DeMets et al., 1994). The major part of this convergence is absorbed by the
41 subduction megathrust in the form of large earthquakes (magnitude $M_w \geq 8$). A small fraction of
42 this convergence – presently about 1 cm/yr at 20°S (e.g., Brooks et al., 2011; Norabuena et al.,
43 1998) – contributes to the deformation of the upper plate over millions of years and to the
44 formation of one of the largest reliefs at the Earth's surface: the Andean Cordilleras and the
45 Altiplano-Puna plateau in between (Figure 1).

46 Andean mountain-building processes are still a matter of debate. A widely accepted
47 model explains the building of the current topography and the crustal thickness of the whole
48 orogen as the result of thrusting propagating eastward over time and concentrated presently on
49 the eastern Andean front, whereas the western flank is described as a passive monoclinical-like
50 crustal-scale flexure (e.g., Isacks, 1988; Lamb, 2011, 2016). However, in the late 1980's,
51 Mpodozis and Ramos (1989) suggested that the Andes-Altiplano was a bivergent orogen with
52 east-vergent structures on the eastern side (along the Sierras Subandinas) and west-vergent
53 thrusting on the western margin (along the Cordillera Domeyko). Later, other authors described
54 west-vergent thrusts at several localities along the western Andean flank (e.g., Charrier et al.,
55 2007; Farías et al., 2005; Garcia & Hérail, 2005; Muñoz & Charrier, 1996; Victor et al., 2004),
56 but they generally gave to these thrusts a minor role in the building of the whole orogen by
57 considering them as secondary backthrusts. A recent and still debated model actually assigns a
58 more important role to these structures, with a west-verging fold-and-thrust-belt (fold-and-thrust-
59 belt hereafter simplified as FTB) along the west Andean flank, thus depicting again the Andes-
60 Altiplano as a bivergent orogen (Armijo et al., 2010, 2015; Riesner et al., 2018). The
61 contribution of this western west-verging FTB to crustal Andean shortening and thickening has
62 been evaluated and quantified at the latitude of Santiago de Chile, $\sim 33^\circ 30'\text{S}$ (Armijo et al., 2010;
63 Riesner et al., 2017, 2018, 2019). There, the orogen is relatively younger and structurally more
64 simple, with a narrower width (Figure 1) and less structural units than in the Central Andes
65 further north. It has been proposed that the FTB emerging along the west Andean front roots on a
66 deeper thrust ramp – the West Andean Thrust (or WAT) – that would represent the primary

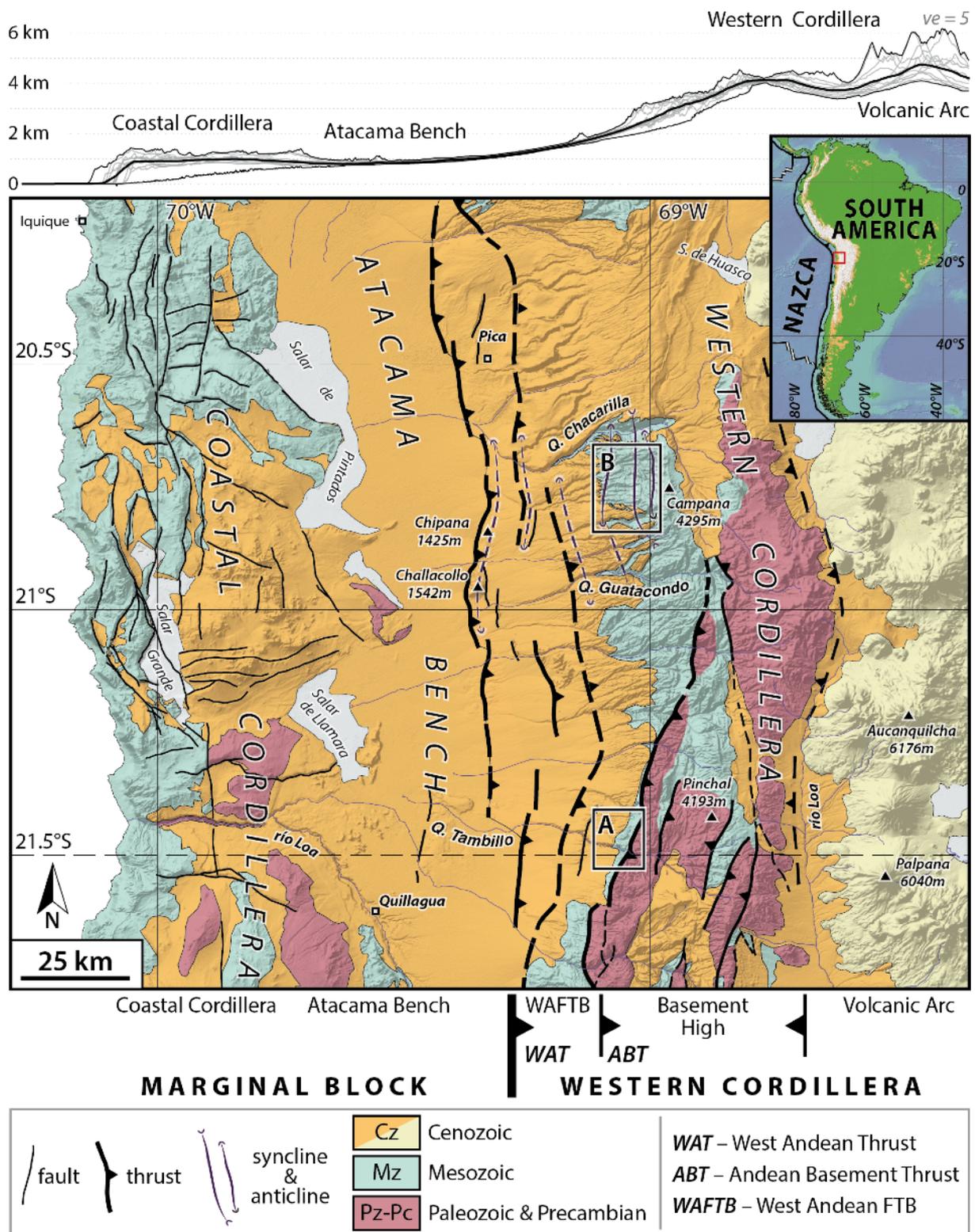


Figure 1. Simplified geological and structural map of the western Central Andes at ~20–22°S, Northern Chile (modified from Armijo et al., 2015), and average topographic profile (top; ve : vertical exaggeration). The two main structural ensembles are here the Marginal Block and the Western Cordillera. The Marginal Block encompasses the Coastal Cordillera and the longitudinal valley of the Atacama Bench (or Central Depression). The Western Cordillera includes the West Andean Fold-and-Thrust-Belt (WAFTB) and Basement High (Cordillera Domeyko), and the modern volcanic arc. A large part of the WAFTB is hidden beneath blanketing Cenozoic deposits and only outcrops in few places. The Andean Basement Thrust (ABT) separates the WAFTB and the Basement High of the Western Cordillera. The location of Figures 4 and 9 is given by black boxes (A and B, respectively). Inset: Location of the map (red box) within the Central Andes along the South American Continent. WAT: West Andean Thrust (after Armijo et al., 2015); FTB: Fold-and-Thrust-Belt; Cz: Cenozoic; Mz: Mesozoic; Pz-Pc: Paleozoic and Precambrian.

67 crustal-scale structure on which the orogeny initiated at $\sim 33^{\circ}30'S$ (Armijo et al., 2010; Riesner et
68 al., 2018, 2019).

69 In contrast, the contribution of similar west-vergent structures further north, at $\sim 20\text{--}22^{\circ}S$
70 (in the so-called Bolivian Orocline), is probably small compared to the total shortening across
71 the entire Andes-Altiplano system, but their role at the start of orogenic building, ~ 50 Myr ago,
72 may have been similarly important (Armijo et al., 2015). At this latitude, Victor et al. (2004)
73 have shown the existence of west-vergent thrusts rooting on a deep decollement dipping
74 eastward beneath the western Andes. They also estimated that these structures absorbed a
75 shortening of ~ 3 km. However, this relatively minor shortening only characterizes the
76 deformation affecting the post ~ 29 Ma Altos de Pica Formation deposited above the Choja
77 erosional surface (or Choja Pediplain). The deformation of the Mesozoic series beneath this
78 surface appears much stronger (Armijo et al., 2015; Blanco & Tomlinson, 2013) but remains to
79 be precisely described and quantified. One of the difficulties in such quantification is that a very
80 large part of the deformation is hidden under blanketing mid-upper Cenozoic deposits and
81 volcanics (Armijo et al., 2015; Fariás et al., 2005; SERNAGEOMIN, 2003; Victor et al., 2004).
82 A quantitative analysis of this deformation and its kinematics is only possible at the few sites
83 along the western flank where deformed Mesozoic series crop out and which are accessible
84 despite the hostile desert conditions in North Chile.

85 In this study, we provide quantitative data to better constrain the geometry of structures,
86 the shortening they accommodated and their kinematics of deformation over time in two of the
87 few areas along the west Andean front where erosion of the Cenozoic units allows for exposures
88 of the underlying deformed Mesozoic layers (Figure 1). The Pinchal area, at $\sim 21^{\circ}30'S$, exhibits a
89 major west-vergent thrust that brings the Paleozoic basement of the Cordillera Domeyko over a
90 FTB of Mesozoic units. These structures have never been described in detail. In the Quebrada
91 Blanca zone, ~ 80 km further north, the excellent exposure of the FTB affecting the Mesozoic
92 series allows for a more quantitative estimate of the shortening and of the timing of the main
93 deformation episodes. Despite our detailed and quantitative approach, these two study areas only
94 give a limited minimal vision of total deformation of this region, as their spatial extent remains
95 minor at the scale of the whole western Andean flank (Figure 1). We find that the shortening of
96 these structures is multi-kilometric, revealing that the contribution of the west Andean flank to
97 Andean mountain building is not negligible. Additionally, we show that the main recorded

98 deformation occurred sometime between ~68 and ~29 Ma (and possibly between ~68 Ma and
99 ~44 Ma), further emphasizing that these structures mostly participated to the early stages of
100 mountain-building.

101 **2 Geological Context of the Andes (~20–22°S)**

102 2.1 General geological framework

103 The Central Andean mountain-belt extends parallel to the Peru–Chile trench (Figure 1),
104 where the Nazca oceanic plate subducts slightly obliquely beneath the South American
105 continent. Spreading out north–south over several thousands of kilometers, the morpho-tectonic
106 structure of the belt varies not only across the range, but also along its ~north–south axis.

107 At ~20–22°S, the mountain-belt is characterized by its largest width (>650 km), highest
108 average elevation (~4–4.5 km above sea level, hereafter a.s.l., Figure 1), thickest crust (70–80
109 km, e.g., Tassara et al., 2006; Wölbern et al., 2009; Yuan et al., 2000) and greatest total
110 shortening (>300 km, e.g., Anderson et al., 2017; Elger et al., 2005; McQuarrie et al., 2005;
111 Sheffels, 1990). Here, the Andean margin along the western border of the continent is described
112 by three major morpho-tectonic ensembles, which are, from west to east (Figure 1): (1) the
113 subduction margin (including the Peru–Chile Trench, the oceanward forearc, and the Coastal
114 Cordillera that reaches altitudes >1 km and that corresponds to the former Mesozoic volcanic
115 arc); (2) the Atacama Bench or Central Depression (at an altitude of ~1 km, corresponding to a
116 modern continental forearc basin, particularly well expressed in the morphology and topography
117 in North Chile) and (3) the strictly speaking Andean orogen, including the current volcanic arc
118 and the Altiplano plateau reaching elevations over 4000 m a.s.l. at ~20°S (e.g., Charrier et al.,
119 2007; McQuarrie et al., 2005; Oncken et al., 2006). Following the terminology of Armijo et al.
120 (2010, 2015), the morpho-tectonic units located west of the Andean orogen constitute the
121 Marginal Block (i.e. the oceanward forearc, the Coastal Cordillera and the Atacama Bench).

122 At ~20–22°S latitude, the Andean orogen itself is composed of several major tectono-
123 stratigraphic ensembles, which are, from west to east: (1) the Western Cordillera (Figure 1),
124 including the Cordillera Domeyko and the modern volcanic arc; (2) the Altiplano Plateau, a
125 high-elevation internally drained low-relief basin; (3) the Eastern Cordillera, a bi-vergent portion
126 of the East Andean FTB; (4) the Interandean zone (or Cordillera Oriental); and (5) the

127 Subandean ranges, east of which the South American craton underthrusts the Andes (e.g., Armijo
128 et al., 2015; Isacks, 1988; McQuarrie et al., 2005; Oncken et al., 2012). The Andean orogen is
129 here tectonically and structurally delimited by two main bivergent thrust systems: the west-
130 vergent West Andean Thrust (WAT) along the western flank, and the east-vergent East Andean
131 Thrust (EAT) along the eastern flank (according to the terminology proposed by Armijo et al.,
132 2015).

133 Although there exist indications of local minor deformation during the Cretaceous, the
134 building of the Andean mountain-belt *stricto sensu* proceeded during the past ~50 Myr at ~20–
135 22°S and was associated with crustal shortening and thickening (e.g., Armijo et al., 2015; Barnes
136 et al., 2008; Charrier et al., 2007; DeCelles et al., 2015; Faccenna et al., 2017; McQuarrie et al.,
137 2005; Oncken et al., 2006). Based on the regional syntheses and reviews by McQuarrie et al.
138 (2005), Oncken et al. (2006), Charrier et al. (2007) and Armijo et al. (2015), the across-strike
139 growth of the orogen may be summarized as follows: (1) at ~70–50 Ma, the Mesozoic arc and
140 backarc basin (formed during the early Andean cycle) is located at the position of the present-
141 day forearc, and most of the current Andes shows mainly flat topography; (2) by ~50 Ma,
142 orogenic growth initiates and deformation primarily affects the Western Cordillera and the
143 western margin of the present-day Altiplano; (3) by ~30 Ma, main shortening vanishes in the
144 Western Cordillera, and is transferred to the Eastern Cordillera, and subsequently to the
145 Interandean Belt; (4) by ~20 Ma, deformation in the Eastern Cordillera and Interandean Belt
146 ends as suggested by the development of erosional surfaces; (5) from ~10 Ma until present,
147 deformation within the Subandean Belt proceeds while the Brazilian Craton underthrusts the
148 Andes. It is therefore clear that the Andean shortening started along the western Andes and
149 subsequently propagated eastward, progressively enlarging the orogen to form the different
150 cordilleras and the Altiplano plateau in between.

151 Different authors investigated crustal shortening and thickening at ~20–22°S at the scale
152 of the whole Andean mountain-belt. From these earlier studies, total crustal shortening is
153 estimated to ~360 km (e.g., Anderson et al., 2017; Barnes & Ehlers, 2009; Elger et al., 2005;
154 Kley & Monaldi, 1998; McQuarrie et al., 2005; Sheffels, 1990). This crustal shortening
155 contributed to crustal thickening. With a crustal thickness of ~70–80 km (e.g., Heit et al., 2007;
156 Tassara et al., 2006; Wölbern et al., 2009; Yuan et al., 2000; Zandt et al., 1994) beneath the
157 Western Cordillera, the Altiplano and the Eastern Cordillera at these latitudes, the crust is over-

158 thickened compared to the ~45 km thick crust of the South America craton (e.g., Wölbern et al.,
159 2009).

160 2.2 Geological setting of the Western flank of the Andes at ~20–22°S

161 The Andean western flank is formed of three tectono-stratigraphic units at ~20–22°S,
162 aside from the present-day volcanic arc. Starting from the East (oldest and deepest units, exposed
163 at high altitudes) to the West (youngest units, lower altitudes), these are (Figure 1): (1) Andean
164 basement consisting of metamorphic rocks of Precambrian and Paleozoic ages; (2) volcano-
165 sedimentary deposits of Mesozoic age (Triassic–Cretaceous), folded and deformed in a FTB, and
166 (3) unconformably overlain by less-deformed mid-upper Cenozoic (Oligocene – Quaternary)
167 volcanics and sedimentary cover. Magmatic intrusions locally alter these different units, and are
168 mostly Cenozoic (SERNAGEOMIN, 2003).

169 The pre-Andean basement rocks formed during the Late Proterozoic and Palaeozoic,
170 when the Amazonian craton was progressively assembled from various terranes (e.g., Charrier et
171 al., 2007; Lucassen et al., 2000; Ramos, 1988; Rapela et al., 1998). At the end of this period of
172 subduction and continental accretion, intensive magmatic activity (volcanism and major granite
173 intrusions) welded together the basement during the Late Carboniferous to Early Permian
174 (Charrier et al., 2007; Ramos, 2008; Vergara & Thomas, 1984).

175 The Mesozoic deposits (Triassic to Cretaceous), found today along the west Andean
176 flank, formed in a proto-Andean arc and backarc basin system during the early period of the
177 Andean cycle (e.g., Charrier et al., 2007; Mpodozis & Ramos, 1989). Marine and continental
178 sediments are interbedded with volcano-magmatic rocks (Aguilef et al., 2019; SERNAGEOMIN,
179 2003). These Mesozoic units attain locally thicknesses up to ≥ 10 km (e.g., Buchelt & Tellez,
180 1988; Charrier et al., 2007; Mpodozis & Ramos, 1989).

181 A regional erosional surface called the Choja Pediplain (Galli-Olivier, 1967) developed
182 during the Eocene to Early Oligocene (~50–30 Ma) (e.g., Armijo et al., 2015; Victor et al.,
183 2004). Above this angular unconformity, the up to ~1600 m thick (Labbe, 2019) Cenozoic
184 deposits of the Altos de Pica Formation (Galli & Dingman, 1962) are composed of continental
185 clastic sediments, interbedded with volcanic layers (Victor et al., 2004). The oldest documented
186 age within the Altos de Pica Formation is of ~24–26 Ma from dated ignimbrites (Fariás et al.,
187 2005; Victor et al., 2004). From there, an age of ~27–29 Ma for the base of the Altos de Pica

188 Formation is inferred regionally when extrapolated to the basal erosional surface using an
189 average sedimentation rate. The youngest ignimbrites within the Altos de Pica Formation are
190 dated at ~14–17 Ma (Middle Miocene) (Vergara & Thomas, 1984; Victor et al., 2004). Based
191 thereon and in addition to other younger dated ignimbrites (Baker, 1977; Vergara & Thomas,
192 1984), Victor et al. (2004) deduced from stratigraphic correlations that the development of the
193 Altos de Pica Formation finished by ~5–7 Ma (Late Miocene) at ~20–22°S.

194 Using apatite fission track dating, Makshev and Zentilli (1999) proposed significant
195 basement exhumation between 50 Ma and 30 Ma, possibly related to basement overthrusting. At
196 ~21°30'S, the geological map of Skarmenta and Marinovic (1981) indicates a west-vergent
197 thrust bringing Paleozoic basement over folded Mesozoic units. Such a thrust contact would
198 imply significant crustal shortening across the western Andean margin, yet to be further
199 documented in the field and quantified. In the folded sedimentary series further west, Victor et
200 al. (2004) determined ~3 km of shortening recorded by the Cenozoic deposits of the Altos de
201 Pica Formation, i.e. accumulated between ~29 Ma and ~5–10 Ma. However, these authors did
202 not take into account the deformation of the underlying, more deformed Mesozoic units.
203 Haschke and Günther (2003) estimated that >9 km of shortening across the western flank in the
204 Sierra Moreno area (~21°45'S) occurred since the Late Cretaceous to Eocene on a west- and
205 east-verging thrust system. It follows that even if published data hint at the existence of a west-
206 verging fault system along the western Andean front at ~20–22°S, its geometry, kinematics and
207 total amount of shortening have not yet been satisfactorily evaluated.

208 Unconformable mid-upper Cenozoic clastic sediments and ignimbrites hide most often
209 the folded Mesozoic layers and their contact with the basement. Investigation is thus limited to
210 sparse areas of few tens of km of extent, only where the interplay of erosion, canyon incision and
211 exhumation has removed this Cenozoic cover and allows for structural observations (Aguilef et
212 al., 2019; SERNAGEOMIN, 2003) (Figure 1). In this study, we focus on two relatively
213 accessible outcrop sites: (1) At ~21°30'S, where the Paleozoic basement thrusts over the
214 Mesozoic according to Skarmenta and Marinovic (1981). This zone will be referred to as the
215 Pinchal area (next to Cerro Pinchal, 4193 m a.s.l.) (box A on Figure 1). (2) At ~20°45'S, where
216 the FTB composed of deformed Mesozoic units has been significantly eroded and allows
217 observations. This zone is hereafter named Quebrada Blanca area, after its largest canyon (box B
218 on Figure 1).

219 **3. Data and Methods**

220 3.1 Available Data

221 The most detailed existing geological map for the Pinchal area is the Quillagua map
222 (1:250,000 scale, Skarmenta & Marinovic, 1981), which only provides very large-scale
223 information but hints for the existence of a major basement thrust. For the Quebrada Blanca area,
224 the recent Guatacondo map (1:100,000 scale, Blanco & Tomlinson, 2013) provides detailed and
225 updated information on the stratigraphy and structure. There, the folded Mesozoic rocks are well
226 exposed on a relatively wide area (~15 km east–west extent) and their structure has been
227 preliminarily mapped and qualitatively described by Blanco and Tomlinson (2013) and Armijo et
228 al. (2015).

229 Enhanced cartographic details can be deduced from high-resolution satellite imagery. We
230 use Google Earth imagery (Landsat 7, DigitalGlobe) whose resolution varies from a few meters
231 to a few tens of meters depending on the zones. In addition, this work benefits from very high-
232 resolution imagery from the European Pléiades satellites. Using the MicMac software suite
233 (Rosu et al., 2014; Rupnik et al., 2016), we calculate high-resolution DEMs from tri-stereo
234 Pléiades imagery, with a 0.5 m resolution. These DEMs are down-sampled to a resolution of 2 m
235 to enhance data treatment and calculations (e.g., stratigraphic projection and image processing).
236 Relative vertical accuracy may reach ~1 m, depending on local slope.

237 Field observations acquired during two field surveys in March 2018 and January 2019
238 complete the dataset and permit the verification of the large-scale data acquired from maps and
239 satellite imagery. Difficult accessibility and field logistics in the remote and desert Pinchal area
240 only allow detailed field observations on a relatively limited area. Observation points and the
241 off-road track followed to reach our field site in the Pinchal area are provided as supplementary
242 material.

243 3.2 Establishing structural maps

244 We establish structural maps for the two investigated sites. We use an approach based on
245 the 3D-mapping of stratigraphic layers on satellite imagery (Armijo et al., 2010; Riesner et al.,
246 2017). More precisely, layers are traced and correlated on Google Earth satellite images. The so-
247 obtained georeferenced traces are projected on the DEM-derived topographic map, and

248 compared with geological maps, mainly for stratigraphic and age references. Field observations
249 allow ground verifications and provide supplementary details, such as the existence of minor
250 thrusts and folds, the observation of polarity criteria or the local measurement of dip angles.

251 The approach used here is mainly limited by local geological complications. Continuous
252 mapping of Mesozoic strata is indeed locally complicated where incision of Cenozoic strata is
253 limited, where magmatic intrusions and associated hydrothermalism alter the surrounding
254 structural geometries, where soft layers with no well-expressed bedding such as marls are present
255 (ex: Pinchal area), or where small landslides or recent sediment deposits hide the underlying
256 deformation pattern. Therefore, geometrical observations and detailed mapping of the structures
257 may be locally difficult, in some zones impossible. These difficulties cause uncertainties in
258 precisely correlating mapped layers and may result in metric to decametric errors (if correlating a
259 layer with its neighbor by error) but do not modify our large-scale (km) results and
260 interpretations.

261 3.3 Building structural cross-sections

262 We use structural measurements, field observations and the obtained structural map to
263 build cross-sections of the two investigated areas.

264 In the Pinchal area – because of limited canyon incision, marls, and frequent blanketing
265 of the structures by Cenozoic cover – we build our structural cross-sections mainly from field
266 observations (strike and dip angles, polarity criteria, first-order stratigraphic column), with
267 additional information taken from satellite imagery.

268 In contrast, in the Quebrada Blanca area, we mostly build our subsurface cross-section
269 from mapping on satellite imagery. Here, we follow the approach already proposed in Armijo et
270 al. (2010) and described in detail in Riesner et al. (2017). The mapped georeferenced horizons
271 are projected on the high-resolution Pléiades DEMs. Using a 3D-modeler, the horizons can be
272 visualized interactively. In order to precisely assess the local average dip and strike angles of
273 deformed Mesozoic layers, we project these layers along swath profiles chosen where Mesozoic
274 strata crop out the best, where folds are mostly cylindrical and where incision (and therefore
275 topographic relief) is most significant. It should be noticed that river incision is here significantly
276 lower (a few hundred meters at most) than at the latitude of Santiago de Chile (~33°30'S) where
277 this approach has been previously employed (Riesner et al., 2018, 2017). In any case, we

278 successfully obtain the overall sectional geometry of layers, and by comparing with the structural
279 map, we determine the approximate locations of the major synclinal and anticlinal fold axes. By
280 respecting the classical structural rule of constant layer thickness, we derive fold geometries.

281 The limits of our interpretations mostly relate to the difficulty of unambiguously
282 correlating stratigraphic layers, and to the fact that, in reality, layers may not always keep
283 constant thicknesses. As incision and local topographic relief are reduced to a few hundred
284 meters at most, the construction of cross-sections is mostly restricted to extrapolating surface dip
285 angles at depth.

286 3.4 Crustal shortening and kinematic modeling

287 We use the obtained subsurface cross-sections to estimate the minimum fold-related
288 shortening across the investigated sites, employing a simple line-length-balancing approach.
289 However, as this approach does not allow for quantifying the shortening absorbed by thrusting,
290 we additionally model deduced anticlinal geometries using a numerical trishear approach (e.g.,
291 Allmendinger, 1998; Erslev, 1991). We use the code FaultFold Forward (version 6)
292 (Allmendinger, 1998) in order to jointly model thrust displacement and anticlinal folding.
293 Trishear models the deformation distributed within a triangular zone located at the tip of a
294 propagating fault. This forward modeling relies on a set of parameters that are here adjusted by
295 trial and error to fit structural geometries. By adding sedimentary layers at various steps during
296 ongoing deformation, we model syntectonic deposition and subsequent deformation, in order to
297 reproduce deformation of Cenozoic layers. Additional information on trishear modeling, together
298 with the range of tested parameters, are provided in supplementary material. We recognize that
299 our best-fit model parameters may not be unique. This is not expected to impact much estimated
300 total shortening as this result depends mostly on final cross-sections. This point will be further
301 discussed in section 7.3 and in the supporting information.

302 **4. Basement thrust and deformed Mesozoic series within the Pinchal area (~21°30'S)**

303 4.1 Synopsis of the previous structural interpretation

304 In the Pinchal area, the Quillagua 1:250,000 geological map (Skarmenta & Marinovic,
305 1981) reports a thrust contact between the Paleozoic basement and the folded Mesozoic series of

306 the Quinchamale formation to the west. These Mesozoic deposits, formed supposedly in a
307 backarc basin context, would be of Jurassic age and would contain two sub-units. Skarmeta and
308 Marinovic (1981) propose that the core of the folded Mesozoic series is Oxfordian (~157–163
309 Ma) with younger Kimmeridgian (~152–157 Ma) layers east and west of it. The mapped
310 structure is therefore an anticline, with an east-dipping axial plane.

311 Our observations confirm the existence of a major basement thrust in the Pinchal area.
312 However, our field investigations disagree with the structural interpretation of the folded
313 Mesozoic series proposed by Skarmeta and Marinovic (1981) just west of the basement thrust.
314 Even though we do not know the absolute ages of the sedimentary series, our structural
315 observations and the relative stratigraphic ages of Mesozoic units, as deduced from either
316 structural or sedimentary polarity criteria, suggest a synclinal structure rather than an anticlinal
317 one.

318 Hereafter, we describe our own stratigraphic and structural observations and
319 subsequently present our interpretation of the Pinchal area.

320 4.2 Field observations

321 4.2.1 Stratigraphic observations

322 In the absence of an existing detailed local stratigraphic documentation, we propose a
323 first-order stratigraphic column from our structural, stratigraphic and sedimentary field
324 observations. In the landscape, the three main tectono-stratigraphic units are clearly
325 distinguishable (Figure 2): (1) the metamorphic basement, (2) the continuous Mesozoic
326 sedimentary series (with a continuum from continental to marine facies) and (3) the continental
327 Cenozoic cover. The first-order stratigraphic column (Figure 3) is hereafter described from the
328 oldest to the youngest units. Detailed field pictures of identified and individualized sedimentary
329 formations are provided in supplementary material to complement the forthcoming stratigraphic
330 descriptions.

331 The Paleozoic basement (Figure S1) dominates the eastern part of the Pinchal area, and is
332 composed of mainly coarse-grain granodiorites and diorites, as well as metamorphic rocks
333 comprising gneisses, migmatites and mica-schist, consistent with documented characteristics of
334 the basement in the area (Skarmeta & Marinovic, 1981).

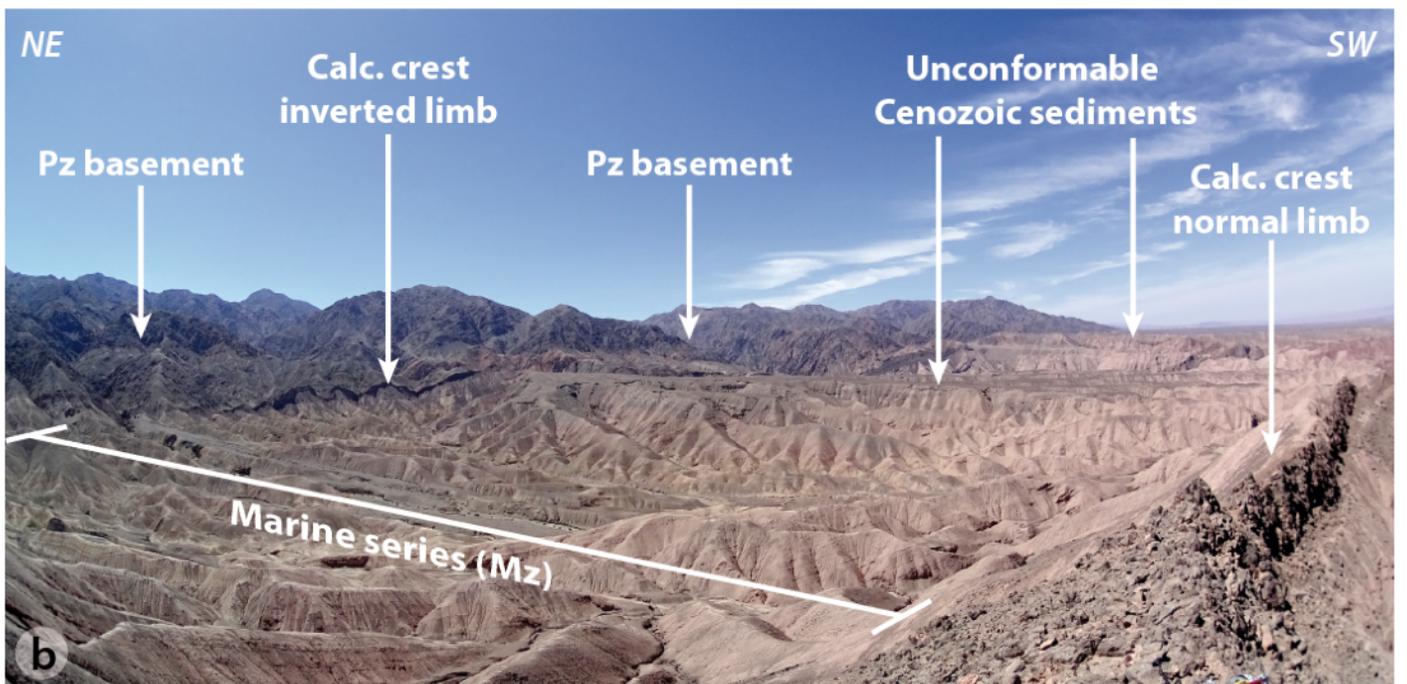
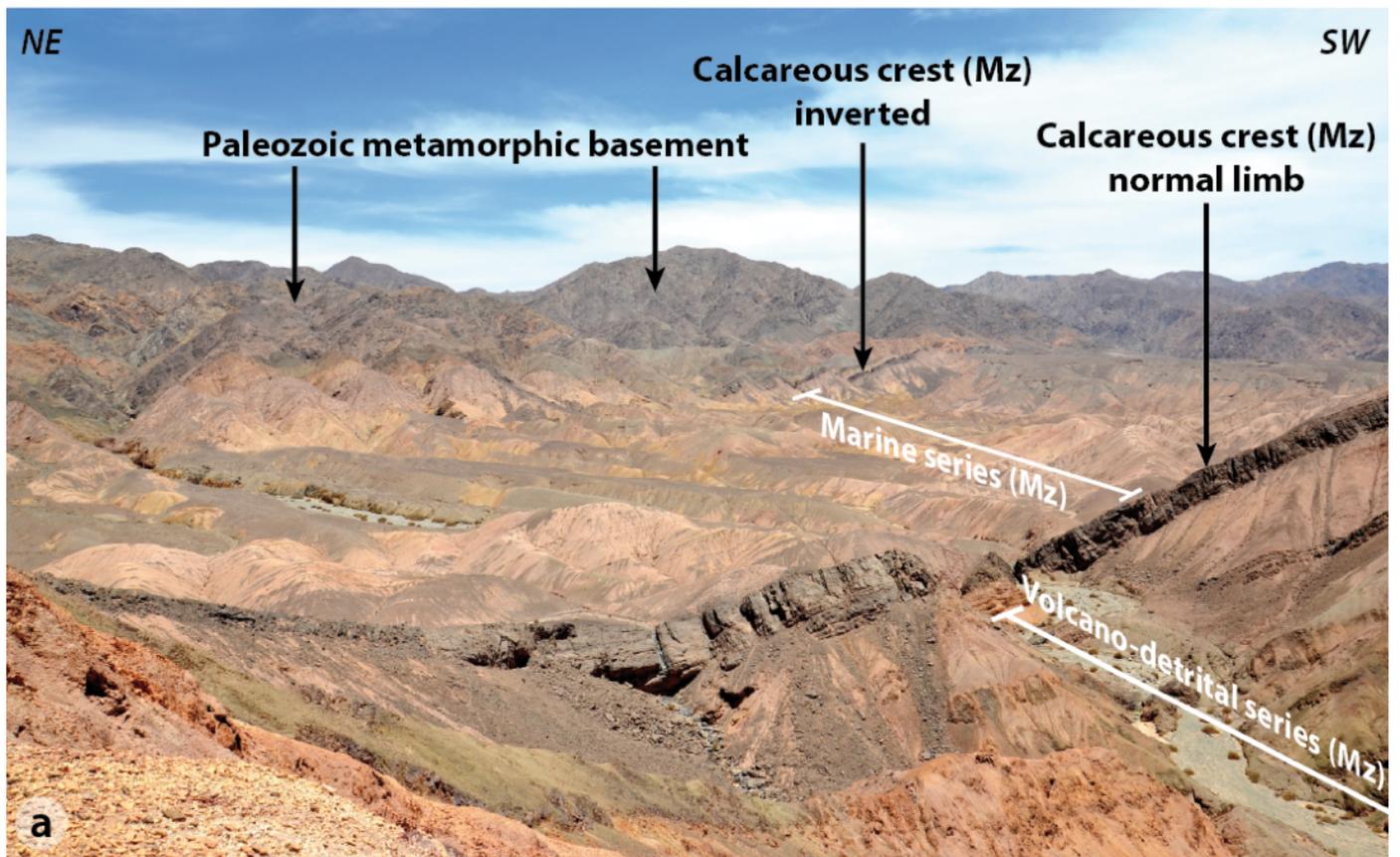


Figure 2. Landscape field overviews of the Pinchal area depicting the main tectono-stratigraphic units. The Paleozoic (Pz) basement stands clearly out in the background, characterized by its darker color and higher altitudes. The Mesozoic (Mz) series in the central part and in the foreground bear a marine part and a volcano-detrital part, delimited by an outstanding calcareous (Calc.) crest. Unconformable Cenozoic erosional surfaces, with limited fluvial deposits can also be observed. View points of both pictures are located on Figure 4.

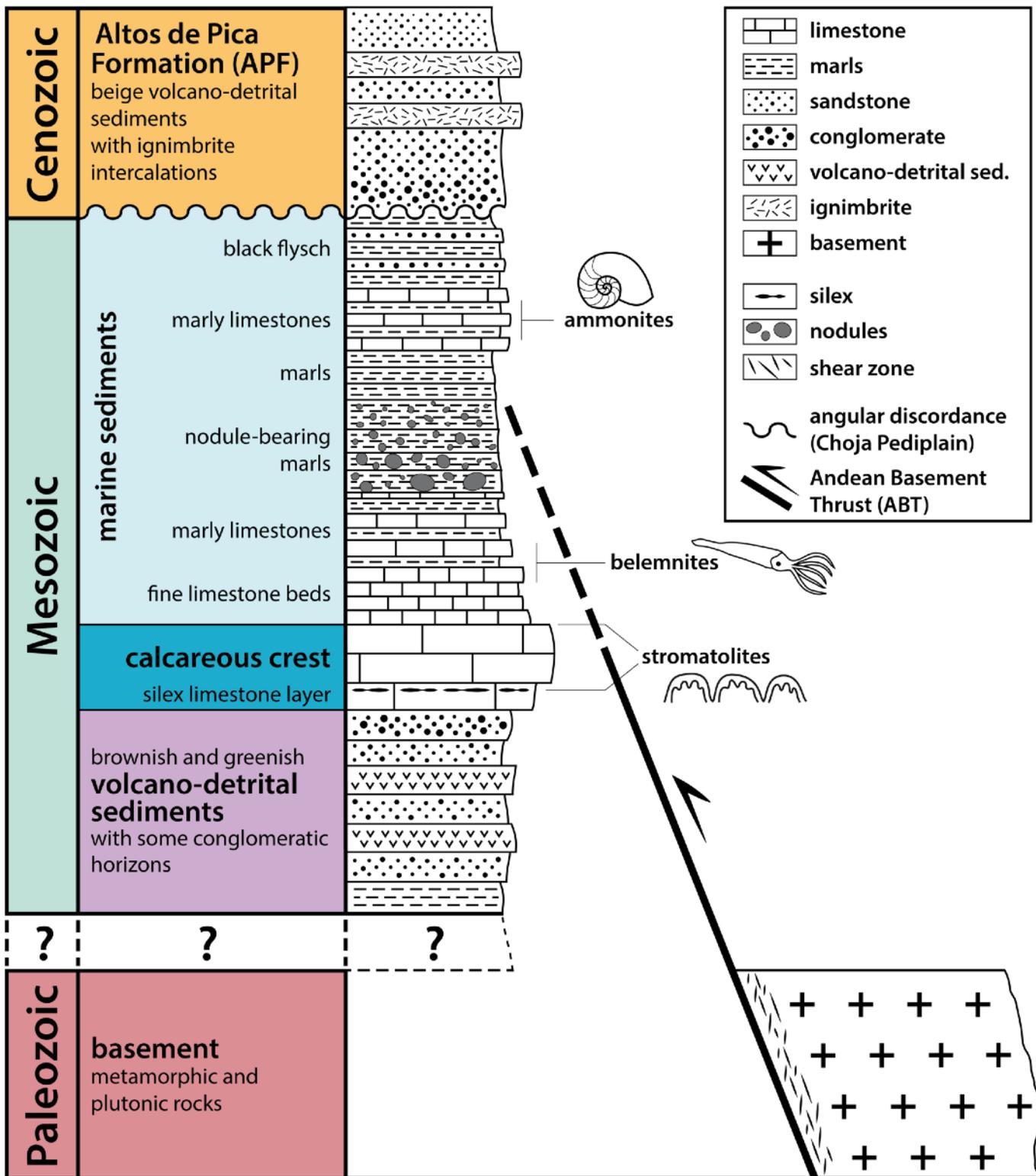


Figure 3. First-order stratigraphic column of the Pinchal area derived from field observations obtained mainly along Quebrada Tania (Figures 4 and 5a) where the Mesozoic series seems to be most complete. By analogy to regional descriptions, these layers are suspected to be Triassic at the base, and Jurassic in the case of the marine fossiliferous levels (see section 4.2.1 for additional details). The description of Cenozoic units is here completed based on the work of Victor et al. (2004). Color-code in line with maps (Figures 1, 4 and S13) and cross-sections (Figure 5). In the Pinchal area, Paleozoic basement overthrusts folded Mesozoic series along the Andean Basement Thrust (ABT), so that part of the deeper and older Mesozoic series may be missing here (as depicted by “?”). Abbreviation “sed.” for sediments. See Figures S1–S12 and corresponding captions (in supplementary material) for detailed sedimentologic descriptions.

335 The older part of the outcropping Mesozoic series consists of continental deposits, with a
336 high content of Paleozoic lithics and volcano-clastic and tuffitic low-rounded conglomerates, of
337 greenish, beige and brownish colors, and clast-sizes varying from a few millimeters to few
338 decimeters (Figure S2). At places, these rocks bear sedimentary polarity criteria such as grain-
339 grading, grain-sorting, cross-bedding and tangential beds (Figure S3). In the eastern part of the
340 Pinchal area, we locally observed below this series some dark green detrital pelites (lutites)
341 (Figure S4). On the basis of petrographic and sedimentological correlations, these detrital
342 Mesozoic sediments resemble to units mapped as Triassic north of the Pinchal zone (between
343 21°–21°30'S) in the Quehuita area (Aguilef et al., 2019).

344 In paraconformity, a characteristic limestone layer marks the beginning of a marine
345 sequence within the Mesozoic series, evidencing a marine transgression process. We hereafter
346 refer to this layer as the "calcareous crest" as it is prominent in the landscape (Figure 2) and as
347 such can be easily used as a reference marker in the field or in satellite images. The base of the
348 calcareous crest is characterized by the presence of silex layers or nodules (Figure S5).
349 Upsection, numerous stromatolites (Figure S6) and bivalves (Figure S7) are found within the
350 unit. Its thickness varies between a few meters (less than 10 m) in the eastern part, to ~10–20 m
351 to the west.

352 The calcareous crest is overlain by thin-bedded (cm–dm) limestone layers of rose-beige
353 color (Figure S8), over a thickness of ~50–100 m. Going upsection, the marine series becomes
354 progressively more marly, limestone layers become more rare and the color more beige,
355 evidencing a deeper marine paleo-environment bearing fossiliferous marl layers. Belemnite
356 fossils were encountered in the lower part of this limestone-to-marl sequence. Characteristic
357 calcareous oval concretions of variable diameter (cm to m) (Figure S9), are pervasive at the
358 transition from marly limestones to marls. The marls bear ammonite fossils, which we have not
359 identified. These Ammonite species could be *Perisphinctes*, *Euaspidoceras*, *Mirosphinctes* and
360 *Gregoryceras*, according to the notice of the Quillagua geological map (Skarmenta & Marinovic,
361 1981) if applicable here. In this case they would be associated with a Middle Jurassic age
362 (Bajocian to Callovian). The series from the thin-bedded limestones to the top of the beige marls
363 is ~200 m thick, along one of the canyons and sections investigated in the field (Quebrada
364 Tania).

365 Upsection, the beige marls become progressively more calcareous again, with the
366 presence of thin limestone layers (Figure S10). Finally, this marine sequence ends with black
367 marls containing layers of beige sandstones (mm to few cm – rarely dm – thick) (Figure S11),
368 indicative of a detrital component in a probable deep seated basin, comparable to the "flysch"
369 series in the Alpine basins (Homewood & Lateltin, 1988). This unit is hereafter called "black
370 flysch", and has a minimum thickness of ~50 m.

371 Continental-clastic Cenozoic deposits (Altos de Pica Formation), unconformably overlies
372 the folded Mesozoic series over the Choja erosional surface (Galli & Dingman, 1962; Galli-
373 Olivier, 1967; Victor et al., 2004), evidencing a regression process (Figures 2 and 3). They are
374 mainly composed of alluvial fan facies that were sourced from the mountain front immediately to
375 the east, with different aggradational terraces. Locally, ignimbrites are observed to cover these
376 clastic series. We encountered red arenites at the base of the Cenozoic series in the western part
377 of the Pinchal area (Figure S12). The age of the oldest sedimentary deposits above this erosional
378 surface is regionally inferred to be ~27–29 Ma (Victor et al., 2004; see also section 2.2).

379 4.2.2 Structural Observations

380 The structural map of Figure 4 illustrates the main stratigraphic and structural features
381 observed in the field and by mapping on satellite imagery. Two ~east–west cross-sections show
382 detailed surface observations along two accessible representative canyons: Quebrada Tania and
383 Quebrada Martine (Figure 5a,b). The Quebrada Tambillo incises deeper into folded units, and as
384 such surface structural observations can be further extrapolated at depth (Figure 5c).

385 The easternmost part of our study area is marked by a major west-vergent thrust bringing
386 the metamorphic basement over the Mesozoic units. This basement thrust is hereafter named the
387 Andean Basement Thrust (ABT). The west-vergent thrust-nature of the shear zone between the
388 Paleozoic basement and the Mesozoic units is observable in the field (Figures 2 and 6). The
389 characteristic C/S-fabric ("Cisaillement/Schistosité") – underlines the penetrative shearing of the
390 basement rocks within and nearby the thrust shear zone, and indicates a top-to-the-west thrusting
391 direction (Figure 7a). The ABT roughly follows a north–south direction (Figure 4). This major
392 contact often resumes to a single basement thrust (Figure 5a,c), but may also show local
393 geometrical complexities, with secondary thrusts and branches, eventually involving basement
394 with stripes of trapped Mesozoic units, as for example along Quebrada Martine (Figure 5b).

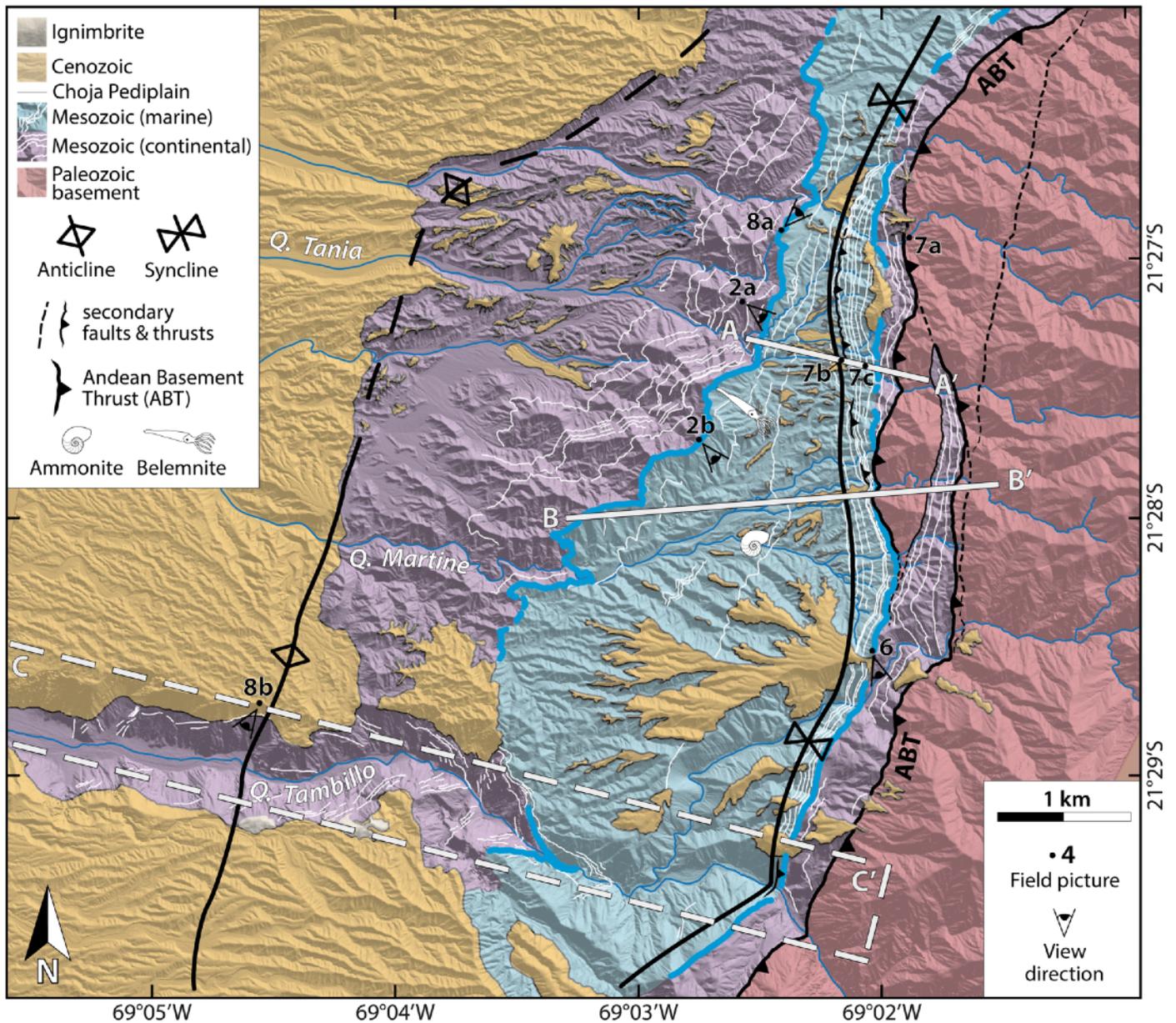


Figure 4. Structural map of the Pinchal area (at ~21°30'S) derived from mapping in the field and on satellite imagery (location on Figure 1, box A). White thin lines highlight Mesozoic layers mappable on satellite images. Thick blue line depicts a robust, out-sticking Mesozoic limestone layer forming a prominent calcareous crest in the landscape (Figure 2). A–A' and B–B' sections locate the topographic profiles used for the surface cross-sections of Quebrada Tania and Quebrada Martine, respectively (Figures 5a-b). In the case of the Quebrada Tambillo cross-section, a topographic swath profile was used along C–C'. The fold axes are relatively well defined for the synclinal fold, but less well constrained for the anticlinal fold because only observable along Quebrada Tambillo. Black dots refer to the location of field photographs, and are numbered according to the figures where these pictures are reported.

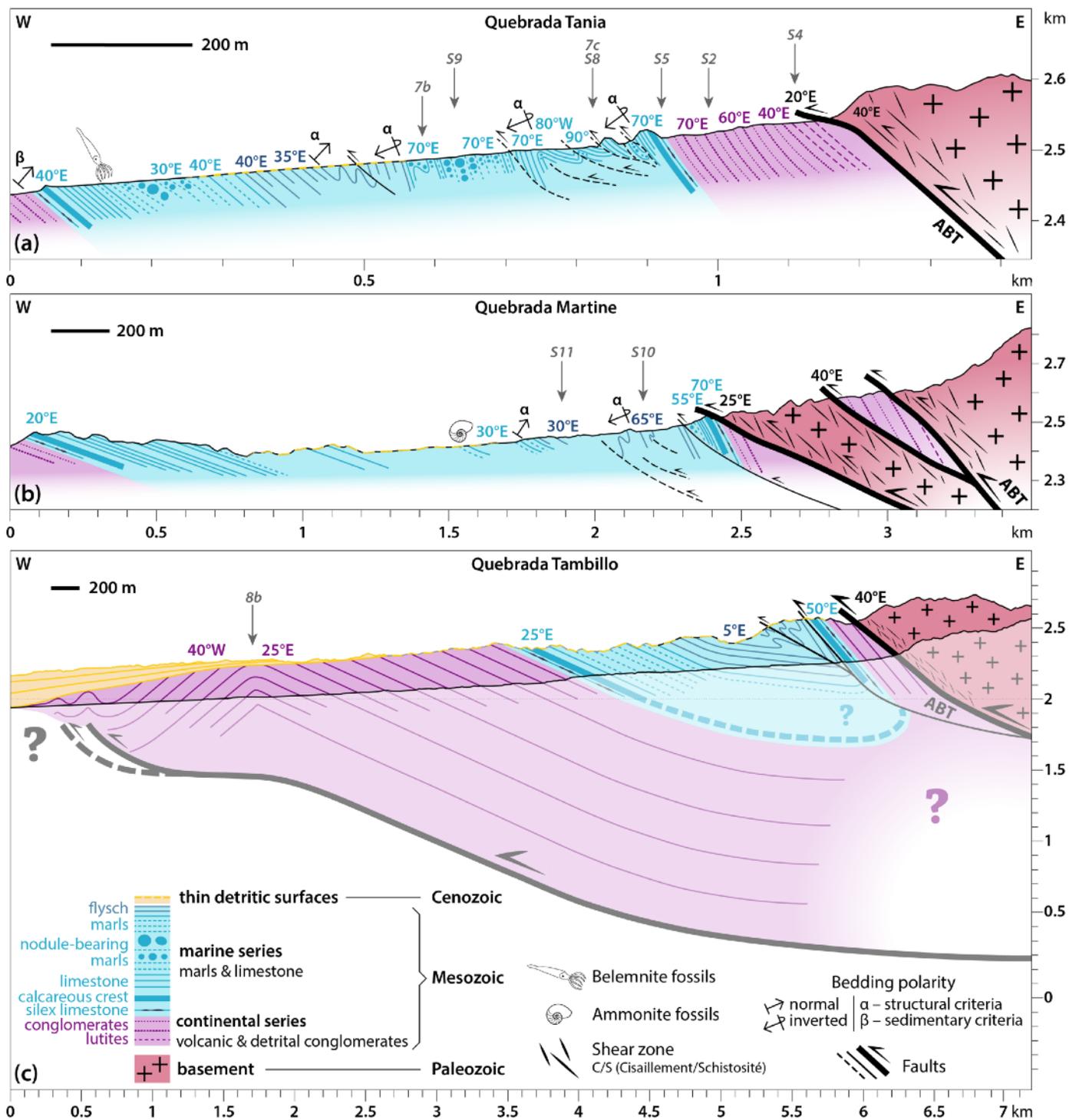


Figure 5. Surface field observations and cross-sections along (a) the Quebrada Tania (A–A' on Figure 4), (b) the Quebrada Martine (B–B' on Figure 4), and (c) the Quebrada Tambillo (C–C' on Figure 4). Reported dip angles have been measured in the field. Faults are outlined in black, and dashed when they are only observable at a local spatial scale. Only larger faults (continuous lines) are mapped on Figure 4. Grey numbers with arrows point out to field pictures and indicate the associated figure. In the case of the Quebrada Tania section (a), the sedimentary polarity criterion (β) indicated to the West of the section has been observed ~ 1 km further downstream than reported here. For the Quebrada Martine section (b), note the stripe of continental Mesozoic rocks trapped in between two strands of the Andean Basement Thrust (ABT). Sub-surface interpretation from surface observations is reported with transparent colors in the case of the Quebrada Tambillo section (c). Note the different spatial scales of the three sections.

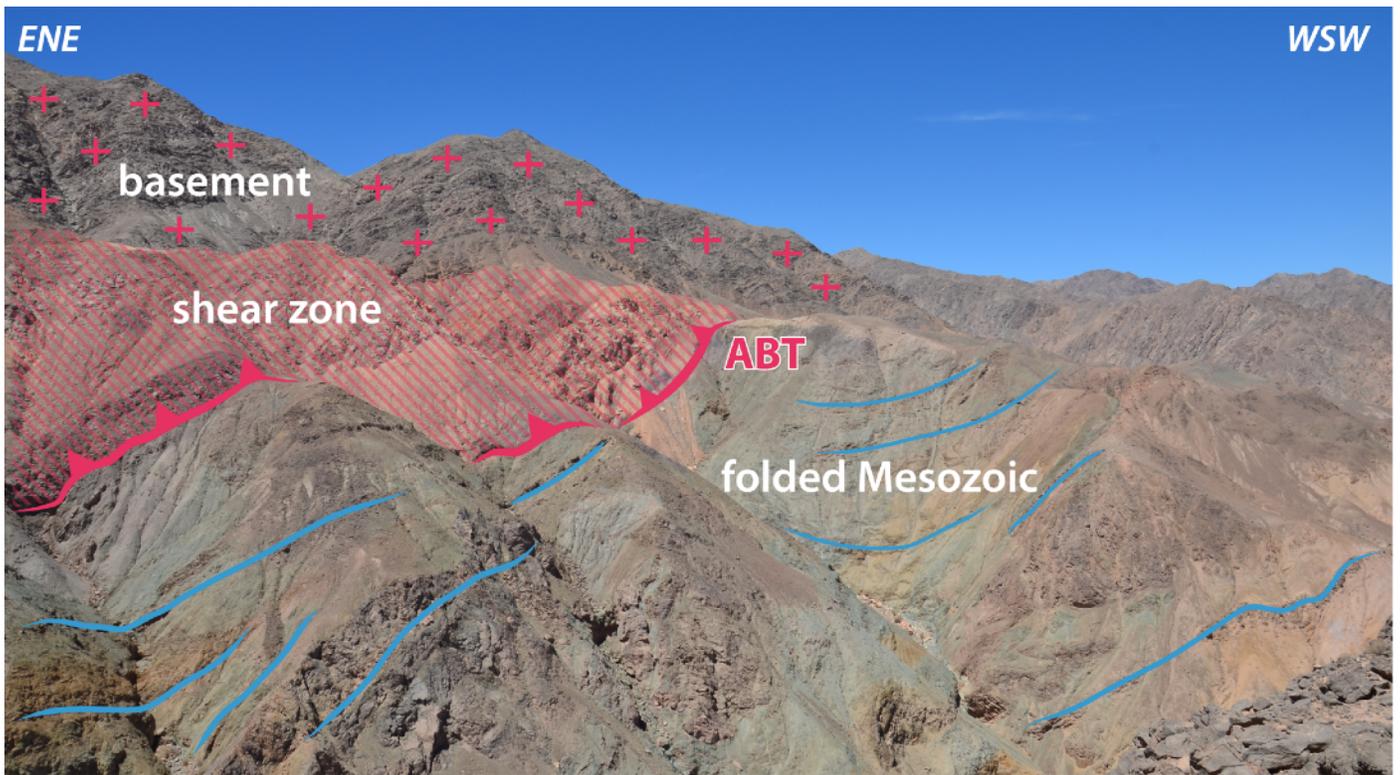


Figure 6. Field view of the Andean Basement Thrust (ABT), overthrusting the dark-grayish Paleozoic basement over the greenish folded Mesozoic units. Reddish rocks on the hanging wall to the East-Northeast correspond to the thrust shear zone (hatched area in picture). Location on Figure 4. Non-interpreted photograph can be found in the supporting information (Figure S15).

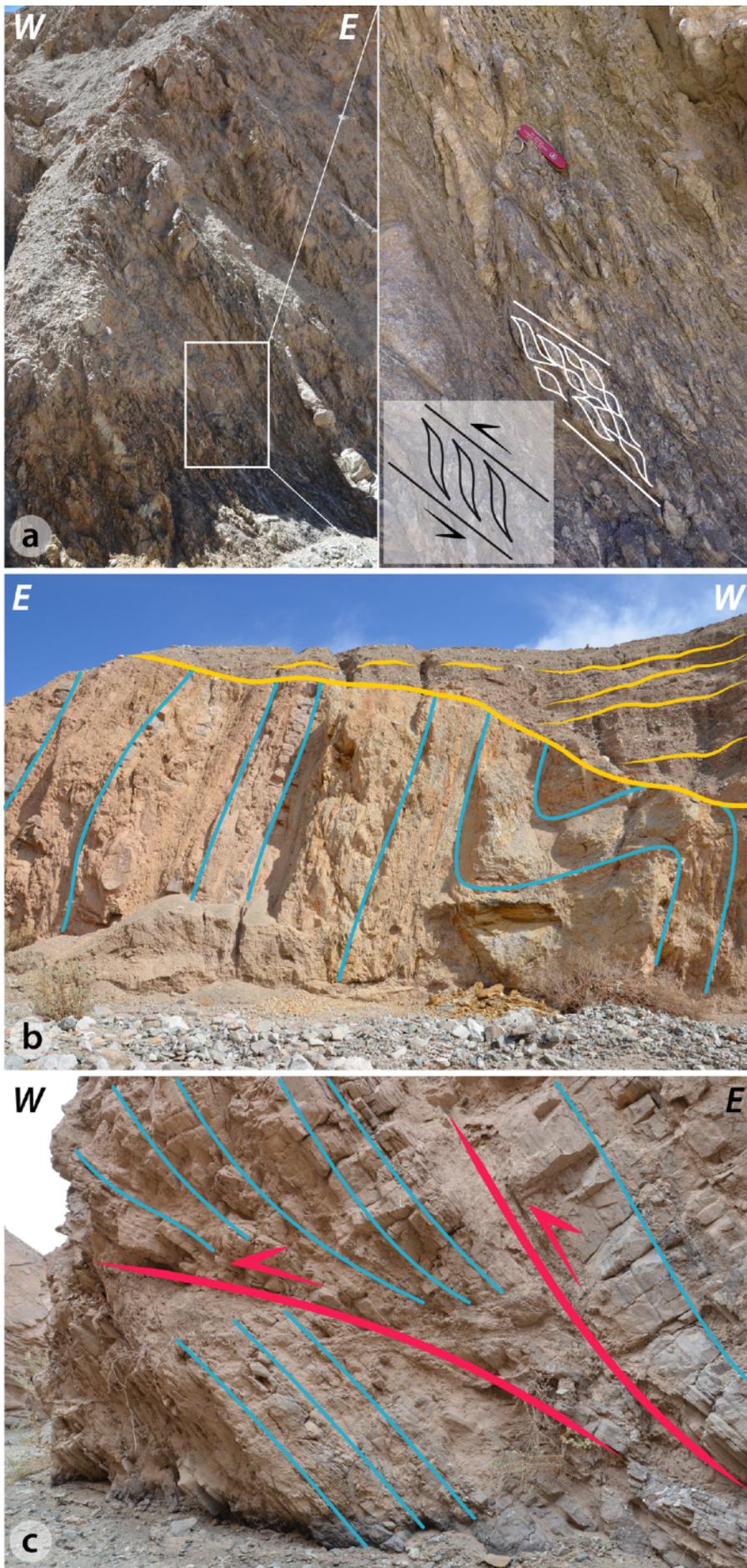


Figure 7. Field pictures of small-scale structural features characteristic of the deformation within the Pinchal area (Location on Figures 4 and 5). Non-interpreted photographs for (b) and (c) can be found in the supporting information (Figures S16).

(a) Shear band with characteristic C/S-fabric (for "Cisaillement/Schistosité") indicative of top-to-the-west thrusting. Observation within the metamorphic basement in the hanging wall of the Andean Basement Thrust.

(b) Example of a small-scale fold within the marine Mesozoic units (blue line) in Quebrada Tania, within the inverted limb of the mapped syncline, nearby the fold axis. Note also the erosional surface (yellow) forming the unconformable contact between the Cenozoic deposits over the deformed Mesozoic.

(c) Small-scale thrusts (steep red line to the right) and décollements (flat red line to the left) observed within the marine Mesozoic strata (blue) of the inverted synclinal limb along Quebrada Tania. The limestone-dominated cm–dm beds are characteristic of the upper part of the marine Mesozoic units (Figure 3).

395 West of the ABT, a FTB consisting of folded Mesozoic units is observable (Figures 2 and
396 4). From east to west, this west-vergent FTB is first formed of an asymmetric and overturned
397 syncline (Figure 8a), followed by a relatively symmetric anticline (Figure 8b). The eastern limb
398 of the syncline, right beneath the ABT, is inverted and locally highly faulted and folded (Figures
399 5 and 7b-c). Within this inverted limb, the series goes westward (and upsection) from sheared
400 lutites beneath the ABT followed by Mesozoic detrital series with conglomerates, to the
401 Mesozoic marine series from the calcareous crest upsection to the marly limestones. The
402 overturned strata are steeply dipping (50–70°E). Penetrative small-scale deformation can be
403 observed pervasively within the marine Mesozoic series, in the form of numerous local small
404 folds, kinematically indicative of an inverted fold limb (Figure 7b), and local secondary shear
405 zones and thrusts (Figure 7c).

406 Going westward, as observed in detail along Quebrada Tania (Figure 5a), the eastern part
407 of the black flysch bears small-scale folds characteristic of the inverted fold limb, whereas
408 normal limb folds are observed slightly further west: the axis of the overturned west-vergent
409 syncline therefore passes through the black flysch. Part of the Mesozoic series is missing, as
410 overthrusting within the flysch and (marly) limestones is observed frequently along Quebrada
411 Tania (Figure 5a). The overturned syncline is therefore found to be broken by a secondary thrust
412 fault striking approximately parallel to the ABT and roughly coinciding with the synclinal fold
413 axis (Figures 4-5). Westward, the normal western limb of the syncline encompasses the whole
414 Mesozoic series from the black flysch down-section to the Mesozoic volcano-detrital series, with
415 more gentle dip angles (20–40°E) (Figures 2 and 5). Penetrative deformation is observed to be
416 limited here.

417 The continental Mesozoic layers of the normal limb of the syncline flatten toward the
418 west. The section along Quebrada Tambillo (Figure 5c) shows a broad, overall symmetrical,
419 anticlinal fold (Figure 8b). Its fold axial plane is steep, dipping ~80°E. The western flank of this
420 large anticline is marked by smaller, secondary folds with westward decreasing wavelength and
421 amplitude. Field logistics did not permit further detailed structural observations within this
422 anticline.

423 The Mesozoic sediments immediately west of the basement are unconformably covered
424 by sheet-like, river-incised Cenozoic fluvial deposits, forming aggradational terraces
425 unconformably deposited above erosional surfaces at different elevations, of varying extension

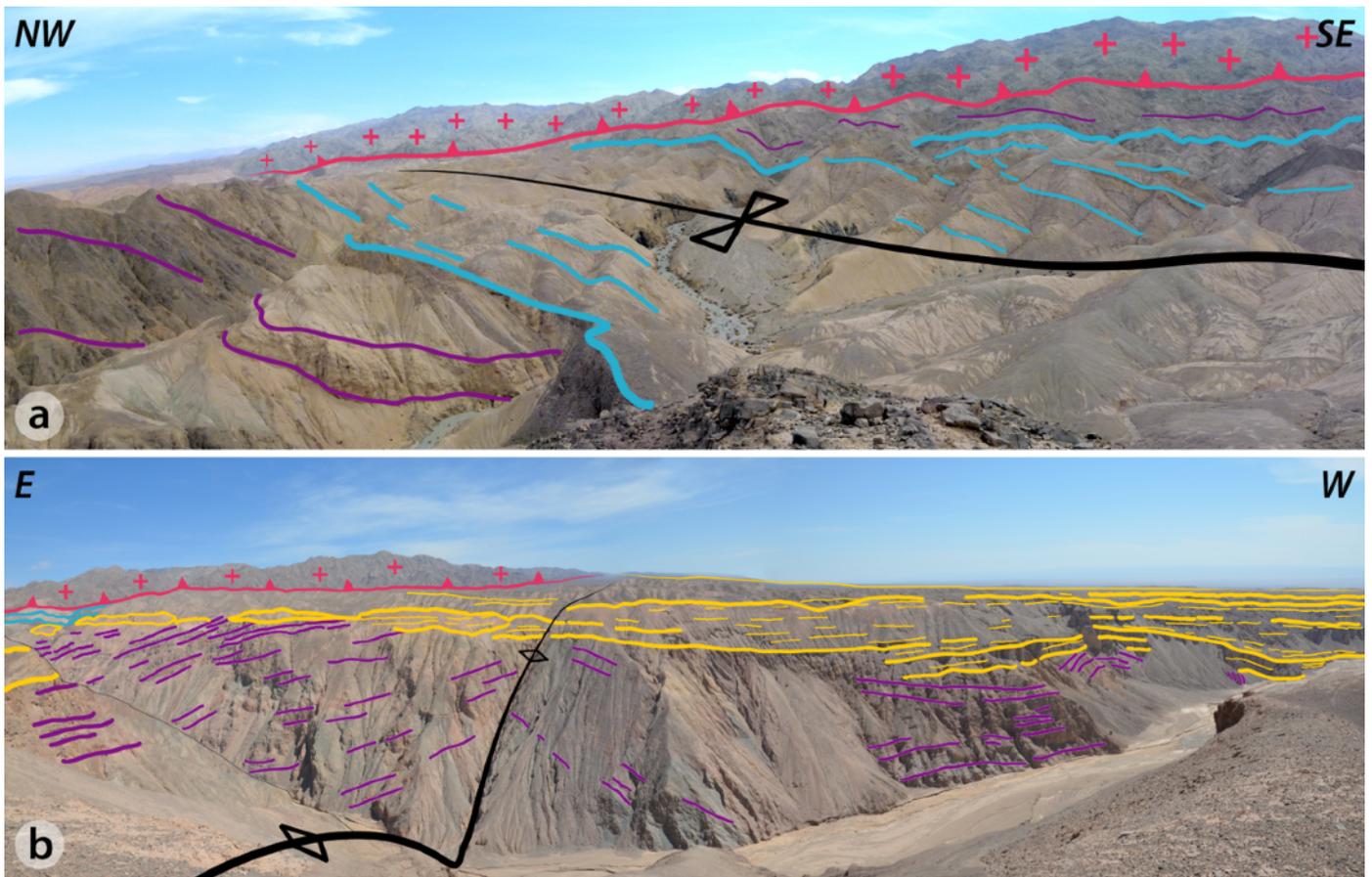


Figure 8. Field pictures of the two major folds within the Pinchal area (location on Figure 4). Non-interpreted photos can be found in the supporting information (Figures S17).

(a) Panoramic view on the north-eastern part of the Pinchal area. The Paleozoic basement (red crosses) overthrusts the Mesozoic units (blue and violet horizons) along Andean Basement Thrust (ABT - red line with triangles). The topographic low locates the synclinal fold axis. The calcareous crest on both sides is highlighted by the thick blue lines. For better visibility, Cenozoic erosional surfaces covered by thin deposits are not highlighted.

(b) Panoramic view along Quebrada Tambillo, in the southern part of the Pinchal area. The ~200 m deep incised canyon reveals the geometry of the large western anticline affecting Mesozoic layers (violet) underneath the unconformable Cenozoic strata (yellow). The fold axis (black line) probably coincides with an approximately vertical fault, also well observable on satellite imagery. Note also the repetition of smaller folds with westward decreasing amplitude and wavelength discernable beneath the westward thickening Cenozoic growth strata to the right of the picture. The Mesozoic calcareous crest (blue) and the Paleozoic basement (red crosses) over the ABT (red) appear in the far eastern background.

426 and of probably different ages (Figure 2b). The majority of these erosional surfaces shows a
427 westward tilt (Figure 5c). Further west, the Cenozoic deposits become thicker and bury the
428 westward extent of the fold-and-thrust-belt. Westward thickening of the Cenozoic layers is
429 clearly observable in Quebrada Tambillo and indicates the presence of growth strata at the front
430 of the anticline deforming the Mesozoic series (Figures 5c and 8b).

431 4.3 Structural Interpretation

432 The synthetic cross-section of Quebrada Tambillo (Figure 5c) summarizes our
433 interpretation of the sub-surface structural geometry of the Pinchal area. Tectonic shortening in
434 the Pinchal area is thus materialized by the presence of the ABT and by the folded and faulted
435 Mesozoic strata.

436 Based on the dip angle of the C/S-fabric in the ABT shear-zone (Figure 7a) and on the
437 mapping of the ABT on satellite imagery, we estimate that the ABT has a subsurface dip angle of
438 $\sim 40^\circ\text{E}$, even though locally flatter such as along Quebradas Tania and Martine (Figures 5a-b).
439 All secondary strands of the ABT are expected to root at depth onto the main shear-zone. The
440 secondary thrust breaking the core of the syncline is roughly parallel to the ABT (Figure 4) and
441 is probably a frontal splay fault of the basement thrust. It is therefore expected to also connect
442 onto it at depth. A similar reasoning is proposed to all small-scale thrusts and décollements
443 observed within the inverted synclinal limb, in particular along Quebrada Tania.

444 Considering that the folds west of the ABT develop above underlying thrusts that connect
445 onto a common detachment is a reasonable and classical assumption for FTBs. This detachment
446 is expected to root at least at the base of the outcropping Mesozoic series, or deeper (Figure 5c).
447 Assuming that the layer thickness is constant over our study area, it can be extrapolated that such
448 detachment is located at least 2 km beneath the topographic surface (i.e. at ~ 0.2 km a.s.l.), or
449 deeper. To the West of our field area, at the front of the anticline, the small-scale folds with
450 westward decreasing wavelength and amplitude (Figure 8b) are interpreted as the possible
451 expression of disharmonic folding within the forelimb of the anticline and/or of a thrust ramping-
452 up toward the sub-surface at the front of the anticline (Figure 5c).

453 Because of the pervasive presence of small-scale folding and thrusting, in particular
454 within the inverted limb of the overthrust syncline, our shortening estimate represents a
455 minimum value and is indicative only of the first-order magnitude of the deformation. Using the

456 simplified first-order cross-section along Quebrada Tambillo (Figure 5c), line-length balancing
457 results in a minimum of ~1 km of shortening across the two folds documented here, from the
458 ABT to the front of the anticline. A significant – but unconstrained – amount of shortening
459 related to the pervasive deformation observed in the field (Figures 5a-b and 7b-c) is to be added.
460 Further, this minimal shortening of 1 km only encompasses folding of Mesozoic units and does
461 not account for thrusting neither on the thrusts of the FTB nor on the ABT. An estimate of the
462 contribution of thrusting within the FTB will be provided below by modeling (section 6.2).

463 The minimum thickness of the Mesozoic series is ~2.2 km, as estimated from the normal
464 limb of the syncline along the Quebrada Tambillo section. Thus, it can be considered that the
465 strict minimum exhumation of the basement is equally of ~2.2 km. Assuming a 40°E dip angle
466 for the ABT, this yields a strict minimum displacement of ~2.6 km on this thrust, which has to be
467 added to the minimum shortening estimated from the folding of the Mesozoic series.

468 **5. Structure of the folded Mesozoic series within the Quebrada Blanca area (~20°45'S)**

469 5.1 Stratigraphy of the Quebrada Blanca area

470 The stratigraphy of the western Andean flank at ~20°45'S is well described in the
471 Guatacondo geological map (Blanco & Tomlinson, 2013). Unlike in the Pinchal area, basement
472 rocks do not crop out in the investigated zone nearby the Quebrada Blanca (Figure 9), but larger
473 scale maps (e.g., SERNAGEOMIN, 2003) show Paleozoic basement units further east and higher
474 in the topography (Figure 1).

475 The Mesozoic units of the Quebrada Blanca are of Jurassic to Cretaceous age (Blanco &
476 Tomlinson, 2013), have been deposited in a back-arc basin context in successive transgression–
477 regression sequences (Charrier et al., 2007), and are subdivided into three formations: (1) The
478 Late Oxfordian Majala Formation, a clastic unit of sandstones, shales and subordinately
479 stromatolitic limestones of transitional marine origin (Blanco et al., 2012; Blanco & Tomlinson,
480 2013; Galli-Olivier, 1967); (2) the Late Jurassic / Early Cretaceous Chacarilla Formation, a
481 continental (fluvial) clastic sequence (Blanco & Tomlinson, 2013; Dingman & Galli, 1965); and
482 (3) the Late Cretaceous Cerro Empexa Formation, an andesitic volcanic and continental
483 sedimentary unit (Blanco et al., 2000; Blanco & Tomlinson, 2013; Dingman & Galli, 1965). The
484 Majala and Chacarilla Formations are both of reddish and beige colors and predominantly bear

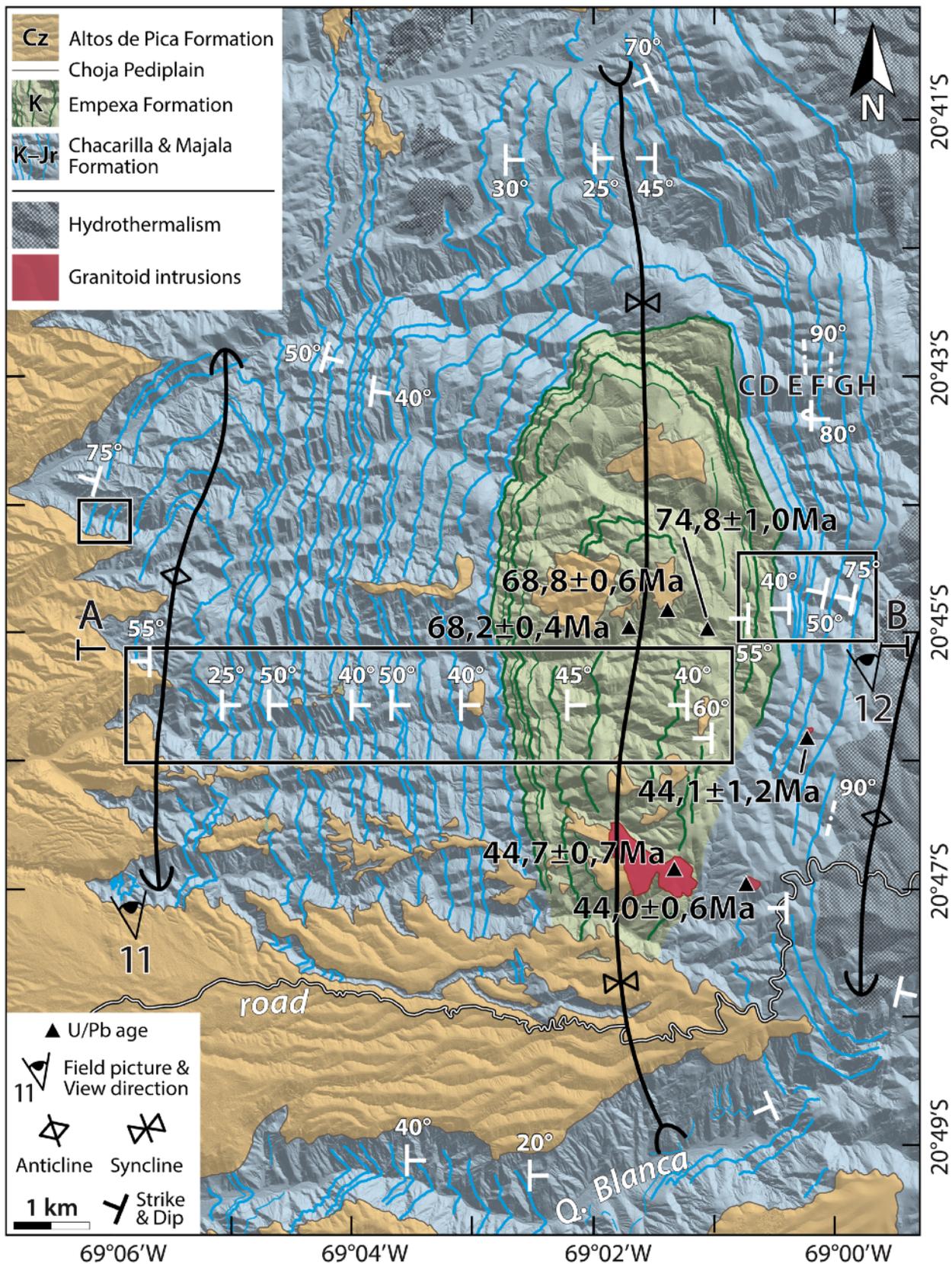


Figure 9. Structural map of the Quebrada Blanca zone (at ~20°45'S), refined from Armijo et al. (2015) (location on Figure 1, box B). Colored lines report mappable layers. For visibility, only major, well-correlated layer traces are represented here. Black boxes locate the swath profiles from which layers were projected for the construction of the structural east–west cross-section (Figure 10). The A–B section corresponds to the topographic profile used for this same cross-section. Strike and dip measurements are extracted from 3D-mapping (see section 3.3) or observed in the field. Strike without dip derives from satellite imagery. Thick black lines correspond to major fold axes. Field pictures are located (with view direction), and numbered according to the associated figure. Ages from uranium-lead (U/Pb) radioisotope dating on zircon are taken from the Guatacondo geological map (Blanco and Tomlinson, 2012). Letters C, D, E, F, G and H to the North-East (within the folded Cretaceous Empexa Formation) report the layers illustrated on Figure 12. Cz: Cenozoic; K: Cretaceous; Jr: Jurassic; Q: Quebrada.

485 detritic sediments. The Cerro Empexa Formation appears greyish and massive in the field, and its
486 age has been determined from uranium-lead (U/Pb) dating on zircon minerals to ~68–75 Ma
487 (Blanco et al., 2012; Blanco & Tomlinson, 2013; Tomlinson et al., 2015) (Figure 9).

488 The Cenozoic deposits of the Altos de Pica Formation here also unconformably overlie
489 the Mesozoic series, over the Choja Pediplain angular unconformity (Galli-Olivier, 1967; see
490 also section 2.2). The age of the basal deposits of the Altos de Pica Formation is estimated
491 regionally to ~27–29 Ma (Blanco & Tomlinson, 2013; Victor et al., 2004).

492 Magmatic intrusions and hydrothermalism occur locally, and hide the eastern
493 continuation of the folded Mesozoic series. Some of these intrusions are dated by uranium-lead
494 (U/Pb) on zircons at ~44 Ma (Blanco & Tomlinson, 2013) (Figure 9).

495 5.2 Structural observations

496 The structural map of the Quebrada Blanca area (Figure 9) highlights the main
497 stratigraphic and structural elements observed in the field and by mapping from satellite imagery.
498 Although the cartography of the FTB is complicated by the persistent Cenozoic cover (notably in
499 the west and south), and by magmatic intrusions and hydrothermalism (particularly to the east),
500 three large-scale folds are clearly observable: a wide syncline in the center, bounded by two
501 anticlines to the east and west. The scale of these major folds is multi-kilometric (Figure 9). The
502 cross-section of Figure 10 illustrates the asymmetry of the folds. Both anticlinal folds have
503 steeper western limbs (dip angles vary mostly between ~50–80°W), whereas their eastern limbs
504 have more gentle dip angles (varying mostly between ~20–50°W) (Figure 10a). It should be
505 noted that the eastern flank of the eastern anticline is widely hidden by magmatic intrusions and
506 hydrothermalism. The central syncline is wider and more symmetric, with dip angles of ~40–50°
507 on both limbs. The anticlines involve the Majala and Chacarilla Formations, while the core of the
508 syncline bears the Cerro Empexa Formation. Overall, the folded series – and in particular the
509 anticlines – document a clear west-vergence of the FTB (Figure 10c). From our projection of the
510 strata mapped on satellite imagery, the Mesozoic series are observed to be concordant (Figures
511 10a-b). The cross-section of the Guatacondo map (Blanco & Tomlinson, 2013) proposes an
512 angular unconformity between the Jurassic and Cretaceous units, however not observed here
513 from our large-scale mapping. As this unconformity does not produce any evident change in the

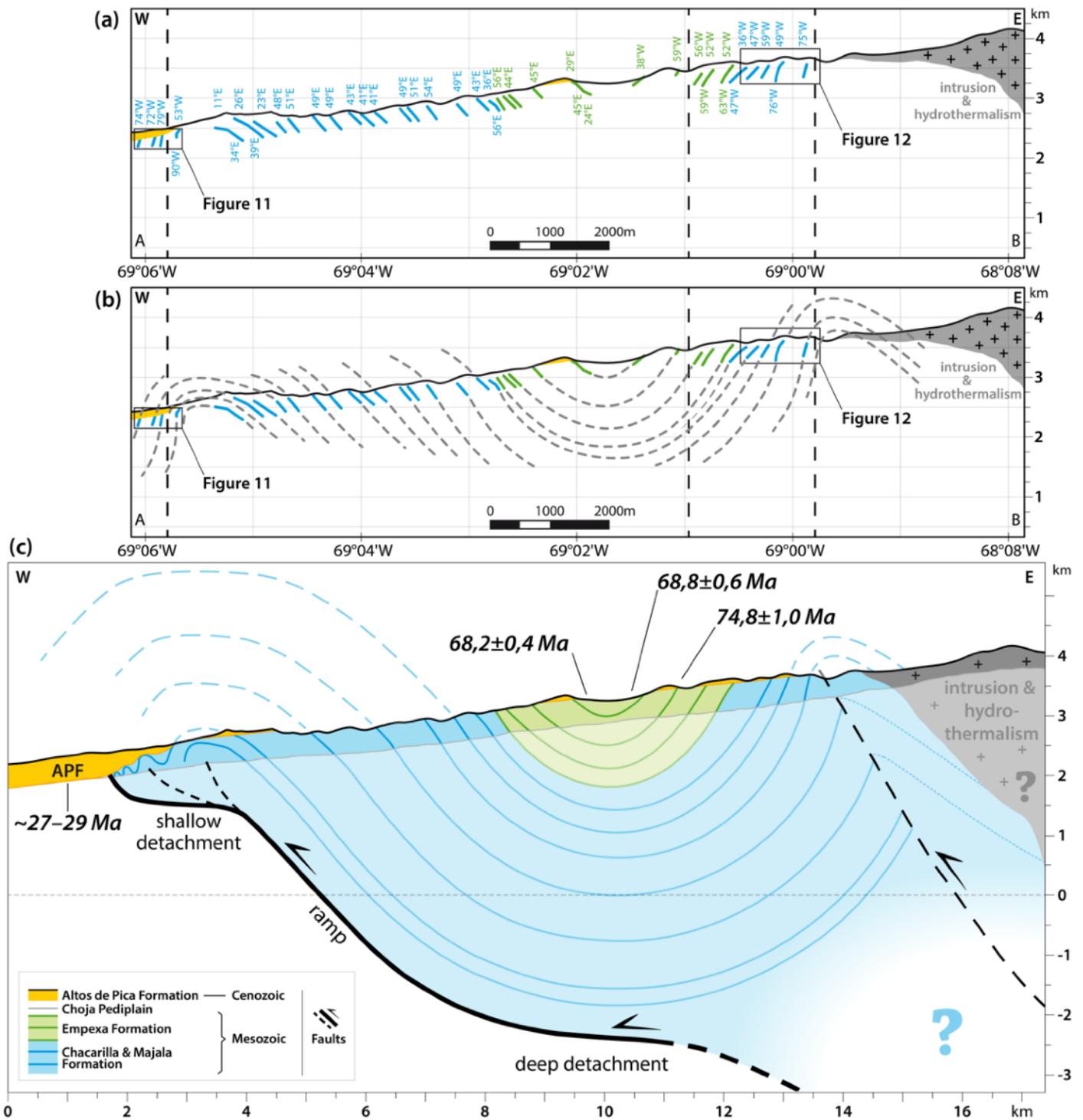


Figure 10. East–west cross-section of the Quebrada Blanca area, established from the projection of selected, well-expressed layers mapped on satellite imagery. APF: Altos de Pica Formation.

(a) Observations, reporting the geometry of projected layers and associated dip angles, together with their stratigraphic ages (color-code).

(b) Sub-surface interpretation and extrapolation of observations.

(c) Synthetic east–west cross-section based on (a) and (b). Interpretation at depth is indicated with transparent colors, in contrast with sub-surface observations. Extrapolation above the topographic surface is drawn with dashed lines. Ages from uranium-lead (U/Pb) radioisotope dating on zircon are taken from the Guatacondo geological map (Blanco and Tomlinson, 2012). The ~27–29 Ma age of the basal deposits of the Altos de Pica formation is derived from regional considerations (Victor et al., 2004).

514 geometry of layers from Jurassic to Cretaceous, we consider it minor for our analysis of the
515 folding here.

516 In the field, we observe small-scale deformation within both anticlines (Figure 10). A
517 series of anticlines with westward decreasing amplitude and wavelength (of a few tens to a few
518 hundreds of meters – to be compared to the ~4 km wavelength of the main anticline) are
519 observable on the western edge of the western anticline (Figures 10c and 11). In the field, at least
520 one of these small-scale folds seems affected by a minor thrust. Additionally, within the eastern
521 large-scale anticline, a thrust-affected small-scale fold is observed (Figures 10c and 12), and
522 confirms the west-vergence at this smaller scale.

523 The Cenozoic detrital units are unconformably deposited above the folded Mesozoic
524 series. Thin sheet-like river-incised Cenozoic surfaces remain in the central part, becoming more
525 dominant to the South and West (Figure 9). These superficial erosional surfaces show an overall
526 westward tilt (Figure 11). Westward thickening of the Cenozoic layers deposited above the
527 erosional Choja surface is clearly observed at the front of the western anticline (Figure 11) and
528 reveals the presence of growth strata.

529 5.3 Structural interpretations

530 As for the Pinchal area and by analogy with other FTBs, we interpret the Quebrada
531 Blanca area folds as related to ramp thrusts rooting onto a deep detachment (Figure 10c). The
532 detachment probably roots at least at the base of the observed Late Jurassic series, or possibly
533 deeper. Assuming constant layer thicknesses over the study area, it can be extrapolated that the
534 detachment locates at least 4 km beneath the current topographic surface (i.e. at least at –2 km
535 a.s.l.). To the East of our investigated area and in order to balance the proposed cross-section, the
536 detachment is interpreted to deepen. An alternative interpretation would be that the detachment
537 keeps a shallow eastward dip angle with some local thickening beneath the eastern anticline. The
538 secondary frontal folds with westward decreasing wavelength (Figures 10c and 11) can be
539 explained as disharmonic folds within the forelimb of the large western anticline and/or be
540 interpreted as reflecting the existence of a shallow thrust (Figure 10c). Such a feature is also in
541 good agreement with secondary (steeper) thrusts affecting the center of anticlines (Figure 10).

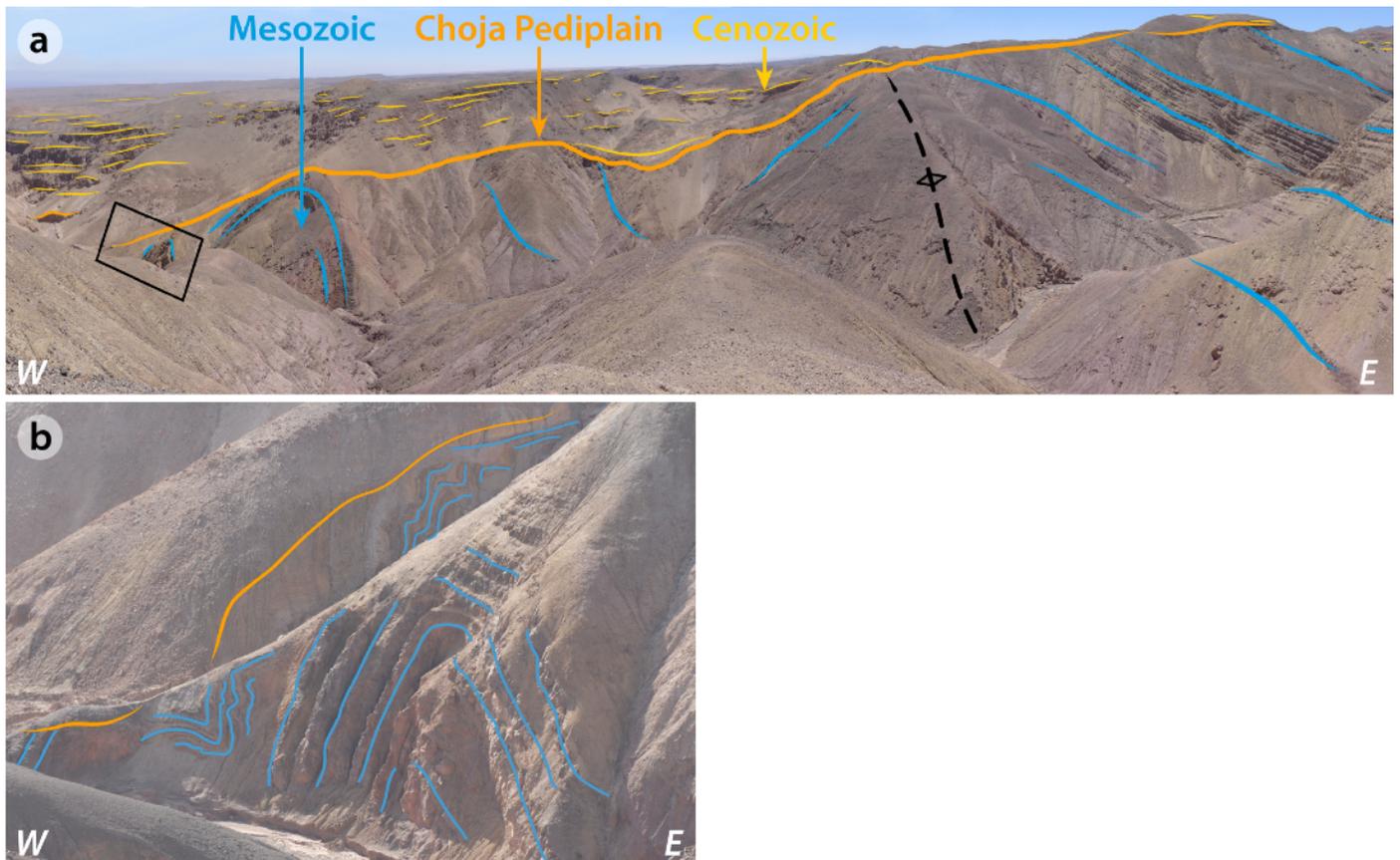


Figure 11. Field picture of the western limb of the western anticline in the Quebrada Blanca area. Non-interpreted photographs are provided in supporting material (Figure S18). Location on Figures 9 and 10. (a) Series of folds with westward decreasing amplitude and wavelength (hundreds to tens of meters) observed at the front of the western anticline. (b) Detailed view of the westernmost outcropping small-scale anticlines, located on (a) by the black box.



Figure 12. Landscape view on the western limb of the eastern large-scale anticline in the Quebrada Blanca area (Location on Figures 9 and 10). Here, steeply inclined Mesozoic horizons are very well discernible in the landscape. Bedding traces C, D, E, F, G and H underlined here are also georeferenced on the structural map (Figure 9) from mapping on satellite imagery. Note the thrust-affected small-scale fold (red dashed line) emphasizing the west-vergence of tectonic structures. The non-interpreted picture is provided in supporting material (Figure S19).

542 Line-length-balancing of the cross-section of Figure 10c results in ~3.8 km of shortening
543 solely related to folding. This value is only a minimum as it does not account neither for slip on
544 the interpreted thrusts nor for the observed small-scale deformation.

545 **6. Kinematics of shortening of the fold-and-thrust-belt at the Pinchal and Quebrada Blanca** 546 **sites**

547 6.1 Timing of deformation

548 The time frame for the tectonic deformation observed within the two investigated sites
549 can be bounded from our data, and more specifically from our results at the Quebrada Blanca site
550 (Figure 10c). Indeed, our field observations and 3D-mapping do not reveal any relevant angular
551 unconformity within the folded Mesozoic series in both study areas. The deformation of the
552 investigated FTB therefore post-dates the deposition of these series. In the Quebrada Blanca
553 area, the youngest folded Mesozoic layers that form the core of the mapped syncline belong to
554 the Cerro Empexa Formation and bear U-Pb ages of 68.9 ± 0.6 Ma and 68 ± 0.4 Ma (Blanco &
555 Tomlinson, 2013) (Figures 9 and 10c). Here, we can therefore conclude that the deformation of
556 the documented folds post-dates ~68 Ma.

557 Magmatic intrusions dated at ~44 Ma intruded the folded Mesozoic units, and appear
558 cartographically not affected by folding (Blanco & Tomlinson, 2013) (Figure 9). This possibly
559 suggests that the major part of the folding occurred during the ~68–44 Ma time interval.
560 However, without additional observations of the deformation – or not – of these intrusions
561 (geometry of the contact with surrounding host units, mineral deformation...), we cannot
562 unequivocally conclude here from this simple cartographic observation.

563 Even though we suspect that the deformed series of the Pinchal zone are Triassic to
564 Jurassic (section 4.2), we do not have any absolute ages of the folded units. Therefore, we
565 postulate that deformation here also post-dates ~68 Ma by analogy to our observations at the
566 Quebrada Blanca.

567 The FTB is unconformably covered by the Cenozoic deposits of the Altos de Pica
568 Formation at both investigated sites. This is also the case for the ABT and secondary thrusts at
569 few places in the Pinchal zone (Figure 4). The presence of growth strata at the front of the
570 westernmost anticlines in both study areas, over the erosional Choja Pediplain, suggests that

571 some deformation proceeded after ~ 29 Ma, during deposition of the Altos de Pica Formation.
572 However, the deformation recorded by folded Mesozoic layers appears of greater intensity than
573 that of the Cenozoic growth layers (Figures 5c and 10c). Given this, we propose that the timing
574 of the main tectonic deformation at $\sim 20\text{--}22^\circ\text{S}$ can be loosely bracketed to a maximum time span
575 of ~ 40 Myr, sometime between ~ 68 Ma and ~ 29 Ma, with additional relatively minor
576 deformation after ~ 29 Ma. Possibly, the main deformation period could be shorter (~ 24 Myr),
577 sometime between ~ 68 Ma and ~ 44 Ma, with minor shortening after the Eocene intrusions.

578 6.2 Further constraints on total shortening deduced from trishear modeling

579 Line-length-balancing only reveals a fraction of the shortening related to the folding of
580 the Mesozoic series. Because the deduced underlying faults of the FTBs have not reached the
581 surface (Figures 5c and 10c), we assume fault-propagation-folding to be the dominant mode of
582 deformation in the studied FTBs. To further explore and quantify the associated deformation, we
583 use kinematic trishear modeling (e.g., Allmendinger, 1998; Erslev, 1991) of the westernmost
584 anticlines documented at the Quebrada Tambillo (Pinchal area) and Quebrada Blanca. This
585 approach accounts for slip on propagating thrust-faults and models the deformation distributed at
586 the tip of these evolving faults. The trishear formalism relies on a set of parameters that are
587 adjusted here by trial and error so as to fit the deduced structural geometries of the modeled
588 anticlines. The values of these parameters are within the range considered in previous studies
589 (e.g., Allmendinger, 1998; Allmendinger & Shaw, 2000; Cristallini & Allmendinger, 2002;
590 Hardy & Ford, 1997; Zehnder & Allmendinger, 2000). Here we present our best-fitting model,
591 which allows for reproducing satisfactorily our structural results, acknowledging that it is most
592 probably not unique. From there, we further discuss the kinematics of the investigated sites.
593 Further details are provided in supplementary material. Tables S1 to S3 provide the set of
594 parameters used for our modeling.

595 The structural geometries of the westernmost anticlines of the two investigated sites are
596 reproduced, and the evolution of deformation is modeled over time taking into account the
597 Cenozoic growth strata. The final geometries of our best-fitting models are reported over our
598 cross-sections, represented in Figure 13. The various stages of deformation are shown in Figures
599 S20 and S21 in the supplementary material. We find that the geometries of the western anticlines
600 can be reproduced with a cumulative shortening of 3.1 km for Quebrada Tambillo (Pinchal area),

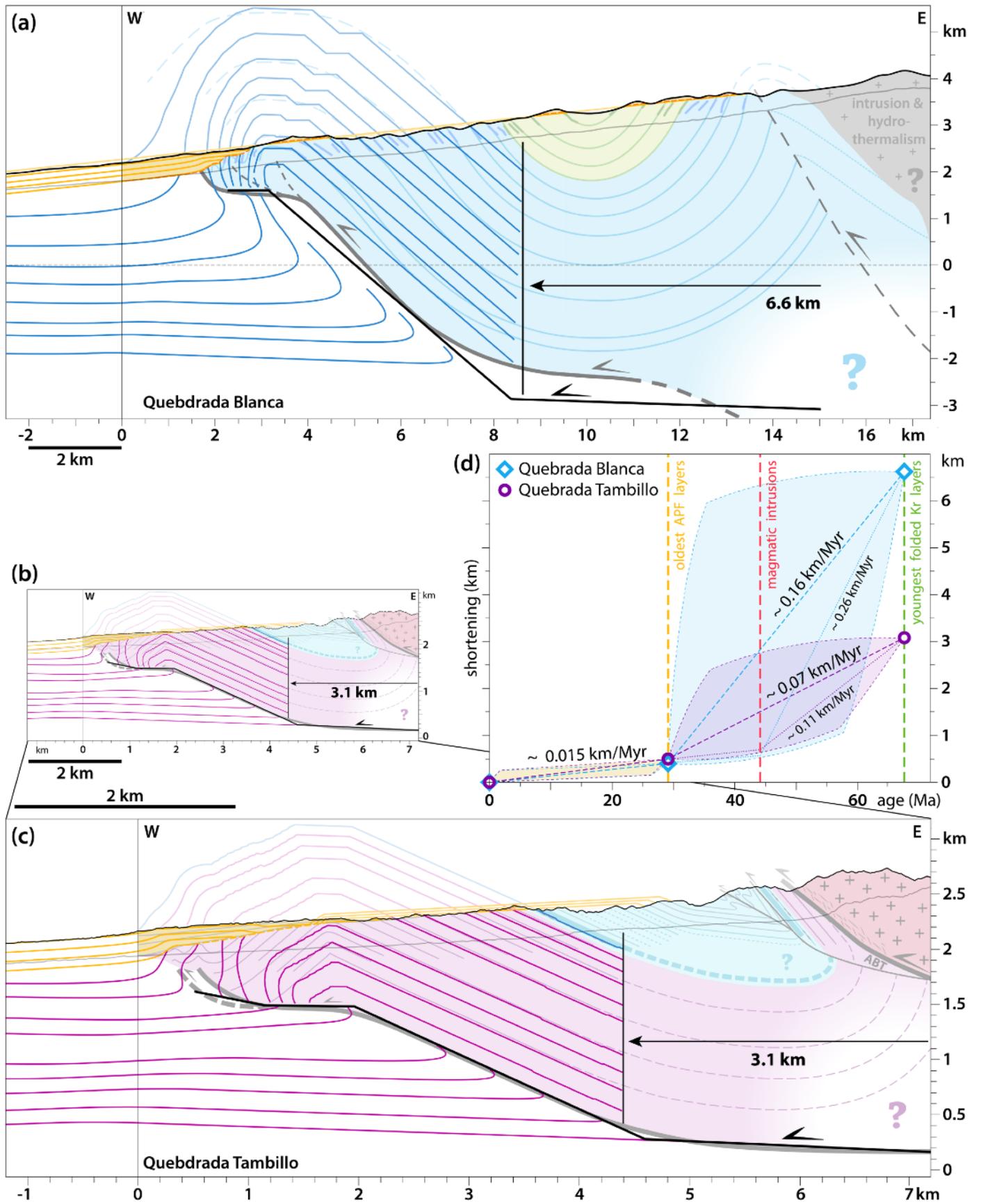


Figure 13. Caption on next page.

Figure 13. Kinematics of folding of the western anticlines of the Quebrada Blanca and Pinchal areas as deduced from field observations and trishear modeling. Modeling was performed with FaultFold Forward v.6 (Allmendinger, 1998).

(a-c) Final stages of the best-fit models in the case of **(a)** Quebrada Blanca area; **(b)** Quebrada Tambillo (Pinchal area), shown here at the same scale as (a). **(c)** Detailed and enlarged view of our results for Quebrada Tambillo (Pinchal area). Note the large scale-difference between the sections of the two investigated sites (a-b). Thicker lines outline model results, while transparent lines and colors refer to the cross-sections of Figures 10c and 5c. These lines are color-coded according to the stratigraphic level they represent, as in the original cross-sections. Black lines report the modeled thrusts and horizontal arrows report the model total shortening.

(d) Shortening vs. time, as deduced from trishear modeling of the western anticlines of the Quebrada Blanca and Pinchal areas, and the ages of deformed layers. The three temporal benchmarks correspond to the age of the youngest folded Cretaceous (Kr) unit (~68 Ma), to the age of magmatic intrusions (~44 Ma) that are cartographically discordant, both derived from the Guatacondo geological map (Blanco and Tomlinson, 2012 – see also Figures 9 and 10c), and to the ~29 Ma age of the oldest Cenozoic layer of the Altos de Pica Formation (APF) (Victor et al., 2004) above the Choja erosional surface. It is possible that most deformation occurred prior to ~44 Ma, as deduced from the age of the intrusions cartographically seemingly post-dating folding (Figure 9), even though this argument is to be taken with caution. Our results underline two phases of deformation, with a slowing down of deformation since ~29 Ma at least, possibly even before. Intermediate stages of the trishear modeling are reported on Figures S20 and S21 (supplementary material) for the cross-sections of Quebrada Tambillo and Quebrada Blanca, respectively. Model parameters are indicated in Tables S1–S3 in supplementary material.

601 and of 6.6 km for Quebrada Blanca (Figure 13). These values account for both thrusting and
602 folding across the western anticlines. They are however minimum shortening values as (1) the
603 depth of the detachment considered for modeling is minimal and could be deeper than the base of
604 the outcropping units, and (2) the model formalism does not account for small-scale deformation,
605 especially within the forelimb of the anticlines where the ramps approach the surface.

606 The above shortening values deduced from trishear modeling only account for the
607 deformation (folding and thrusting) absorbed across the westernmost anticlines of our two
608 investigated sites. If synclinal folding accounts for half of the shortening deduced previously by
609 line-length-balancing in the Pinchal area (i.e. ~ 0.5 km of shortening), a minimum amount of
610 shortening of ~ 3.6 km can be proposed across the whole Quebrada Tambillo section, and
611 includes folding of the outcropping FTB, as well as slip on the detachment and western thrust
612 ramp. When adding the minimum ~ 2.6 km of thrusting deduced on the ABT, we get a minimum
613 shortening of ~ 6.2 km across the whole Pinchal area. Similarly, in the Quebrada Blanca area, if
614 the easternmost anticline and syncline take up half of the folding deduced previously by line-
615 length-balancing (i.e. ~ 2 km of shortening), a minimum amount of shortening of ~ 8.6 km is
616 deduced across the whole Quebrada Blanca section, including folding of the outcropping FTB in
617 addition to slip on the underlying detachment and western ramp.

618 The two investigated FTBs take up differing amounts of minimum shortening. These
619 variations may relate to the disparate extents of outcropping structures, in particular because the
620 scale of the two sections are significantly different, in terms of length (~ 7 km long section for the
621 Quebrada Tambillo vs. ~ 17 km long section for the Quebrada Blanca) but also in terms of depth
622 of the underlying detachment (altitude of ~ 0.2 km for Quebrada Tambillo vs. depth ~ 2 km for
623 Quebrada Blanca, relative to sea level) (Figure 13). Lateral variations in deformation can also not
624 be excluded.

625 6.3 Kinematics of shortening

626 Trishear modeling allows for simulating the evolution of thrust slip and folding in the
627 case of the westernmost anticlines of the two investigated sites. By adding syntectonic layers
628 while deformation proceeds, we also reproduce the overall geometry of the base of the Cenozoic
629 Altos de Pica Formation deposits and of the subsequent growth strata (Figures S20 and S21).
630 Syntectonic surfaces and layers are prescribed an initial $3\text{--}6^\circ$ W dipping angle, similar to the

631 present-day overall regional topographic slope (Figure 1). From there, we find that ~0.5 km and
632 ~0.4 km of shortening are needed to reproduce to the first order the geometry of the base of the
633 Altos de Pica Formation deposits at the front of the Quebrada Tambillo (Pinchal area) and
634 Quebrada Blanca sections, respectively, using the trishear models adjusted to the final cross-
635 sections. When compared to the 3.1 km and 6.6 km of total shortening cumulated since ~68 Ma
636 across the westernmost anticlines of these two sections, this indicates that the ~29 Ma old basal
637 Cenozoic layers above the Choja surface only record at most 16% and 6% of this total
638 shortening, respectively. We have tested the possibility of initial horizontal Cenozoic syntectonic
639 layers. In this case, a post ~29 Ma shortening of 0.8 km at most is needed to best adjust the
640 observed geometry of the basal Altos de Pica Formation layers, even though a good fit to both
641 the geometry of the growth strata and of the finite fold structure cannot be satisfactorily found.

642 These results are then used to quantitatively describe the evolution of shortening over
643 time across the westernmost anticlines of the two investigated sections, with account on the
644 timing of deformation discussed in section 6.1 (Figure 13d). We find that shortening rates were
645 on average of ~0.07–0.16 km/Myr over the time span ~68–29 Ma. They could have been even as
646 high as ~0.11–0.26 km/Myr if considering that the main deformation phase is confined to ~68–
647 44 Ma. Subsequently, deformation rates decreased to an average value of ~0.015 km/Myr after
648 ~29 Ma, starting possibly earlier.

649 It should be noted that these average values are most probably minimum values. Indeed,
650 thrusting and folding are here only modeled for the westernmost anticlines of our study sites, and
651 do not account for the shortening cumulated neither across the other structures of the FTB nor on
652 the ABT. Also, the main phase of deformation prior to ~29 Ma could have lasted less than the
653 ~68–29 Ma or ~68–44 Ma time intervals, respectively (Figure 13d).

654 Our results therefore quantitatively emphasize our former qualitative conclusion that the
655 major phase of deformation recorded at the two investigated sites occurred sometime between
656 ~68 and ~29 Ma, with a significant subsequent slowing down of deformation rates afterwards,
657 possibly as soon as ~44 Ma or earlier (Figure 13d).

658 **7. Discussion**

659 7.1 The Andean Basement Thrust

660 We have further documented the existence of a major west-vergent basement thrust – the
661 ABT, for Andean Basement Thrust – along the western flank of the Andes, after its initial
662 pointing out on earlier geological maps.

663 Our study in the Pinchal area suggests that the ABT bears local complexities with several
664 strands and minor splays, most probably related to the reactivation of structures in the initial pre-
665 Andean back-arc basins. Laterally and at a larger scale, these complexities may also exist so that
666 the ABT is possibly segmented all along the west Andean flank at $\sim 20\text{--}22^\circ\text{S}$. The geological
667 map of Skarmenta and Marinovic (1981), on which we based our investigations of the ABT,
668 clearly documents this structure from $\sim 21^\circ 15'\text{S}$ to $21^\circ 35'\text{S}$, and possibly down to $\sim 22^\circ\text{S}$ with
669 some structural complexities by $\sim 21^\circ 35'\text{S}$ with the junction of two possible strands of this same
670 basement thrust system.

671 Immediately north, the ABT is mapped locally as the Quehuita Fault up to $\sim 21^\circ 11'\text{S}$
672 (Aguilef et al., 2019). The Choja Fault between $\sim 21^\circ 08'\text{S}$ – $21^\circ 01'\text{S}$ (Aguilef et al., 2019) is also
673 mapped as a thrust bringing basement over folded Mesozoic sediments. Even though the
674 cartographic continuity between both basement thrusts of Aguilef et al. (2019)'s map is not
675 straightforward, they can be reasonably considered as strands of the same major basement thrust
676 system, similar to that documented for the ABT in this study. North of $\sim 21^\circ\text{S}$, intrusions,
677 hydrothermalism and surface volcanics hamper any clear observation of a similar basement
678 thrust. Such basement thrust, if existent, would however provide a reasonable mechanism for the
679 exhumation and exposure of basement rocks east of the folded Mesozoic units and at higher
680 elevations, at the latitude of Quebrada Blanca ($\sim 20^\circ 45'\text{S}$) (Figure 1). For these reasons, we
681 cannot tell with any certainty whether a thrust contact similar to the ABT (bringing basement
682 westward over folded Mesozoic) exists at this latitude, but such structure is to be suspected.

683 South of the map by Skarmenta and Marinovic (1981), in the Sierra Moreno at $\sim 21^\circ 45'\text{S}$,
684 Haschke and Günther (2003)'s section report a basement thrust over folded Mesozoic units, in
685 agreement with the style of deformation documented here, but with a relatively minor
686 displacement on this thrust compared to our results in the Pinchal area. Together with the
687 (1:1,000,000) Geological map of Chile (SERNAGEOMIN, 2003), Haschke and Günther

688 (2003)'s map suggests that the ABT is cartographically continuous southward to the southern
689 end of the Sierra Moreno, at $\sim 22^{\circ}05'S$. This possibly documents the lateral termination of this
690 section of the ABT.

691 As a conclusion, the ABT stands as a major crustal structure, bringing basement units
692 westward over folded Mesozoic units – and therefore uplifting the western margin of the
693 Altiplano – possibly segmented but extending over at least ~ 120 km (Figure 1). Even though its
694 various strands are mapped as local basement faults in some maps, as in our study (Figure 4) or
695 in other maps (Aguilef et al., 2019; McElderry et al., 1996; Skarmenta & Marinovic, 1981;
696 Tomlinson et al., 2001), altogether these thrust segments document a much larger thrust system
697 (Figure 1). We interpret the ABT and the other thrusts west of it to root onto a low-angle,
698 eastward dipping décollement, situated >2 km (Pinchal area) or >4 km (Quebrada Blanca area)
699 beneath the present-day topographic surface. Deeper and eastward, this décollement probably
700 steepens and forms a crustal-scale ramp, needed to sustain the uplift and topographic rise of the
701 Western Andes, as proposed by Victor et al. (2004) and Armijo et al. (2015). Such crustal-scale
702 structure is termed the West Andean Thrust (or WAT) by Armijo et al. (2015).

703 From the likely minimum offset of the basement in the Pinchal area we estimated that the
704 ABT alone accommodated a strict minimum of ~ 2.6 km of shortening on a horizontal distance of
705 ~ 1 km. This multi-kilometric shortening would be associated with multi-kilometric basement
706 exhumation, but only limited thermochronological data actually permit to evaluate the amount of
707 exhumation. These data presently exist at a regional scale but are absent locally. From apatite
708 fission track dating in basement samples taken ~ 20 km east and south-east of our study sites of
709 the Pinchal and Quebrada Blanca areas, Makshev and Zentilli (1999) inferred at least 4–5 km of
710 basement exhumation occurring between ~ 50 – 30 Ma. This is in good agreement with our results
711 in terms of amount of uplift that would result from basement overthrusting on the ABT and
712 above the WAT. In the absence of dated samples closer to the ABT, it is difficult to assess more
713 precisely its timing. At a few places in the Pinchal area, the ABT is covered by Cenozoic
714 deposits, so that it was probably mostly active before ~ 29 Ma. Given this observation and by
715 comparison and extrapolation with published thermochronological results (Makshev & Zentilli,
716 1999), we postulate that the ABT was most probably active between ~ 50 – 30 Ma, suggesting that
717 it was therefore overall coeval with deformation of the FTB documented immediately further
718 west.

719 7.2 The West Andean Fold-and-Thrust-Belt at ~20–22°S

720 The series of west-vergent folds and faults at ~20–22°S deforming Mesozoic units is
721 interpreted as a west-vergent FTB, along the western flank of the Andes. By analogy to what has
722 been proposed at the latitude of Santiago de Chile (~33.5°S) (Armijo et al., 2015; Rauld, 2011;
723 Riesner et al., 2018, 2017), we propose to name this FTB hereafter as the West Andean Fold-
724 and-Thrust-Belt (WAFTB).

725 The WAFTB extends over our entire study zone from ~20–22°S, even though a large part
726 north of the Quebrada Blanca area is covered by Cenozoic strata. It therefore spreads out over a
727 north–south-distance of at least ~200 km – and possibly more as folded Mesozoic sediments are
728 mapped on the (1:1,000,000) Geological map of Chile (SERNAGEOMIN, 2003) in the north-
729 and south-ward continuation of the zone investigated here. The WAFTB of northern Chile
730 accommodates a minimum shortening of ~3–9 km, as quantified from the ~7–17 km large cross-
731 sections representative of the two investigated areas (not including the contribution of the ABT
732 in the case of the Pinchal area).

733 Few authors attempted to quantify the shortening in this part of the Andes. At 20°30'S,
734 Victor et al. (2004) only evaluated the deformation affecting the post ~29 Ma deposits and not
735 the total shortening as could be derived from folded Mesozoic series. We further discuss their
736 results hereafter (below and section 7.3). At ~21°45'S, ~30 km south of the Pinchal area,
737 Haschke and Günther (2003) reported a minimum shortening of >9 km from a ~50 km wide
738 cross-section, but without providing nor discussing the data used to make this estimate. Their
739 section encompasses an equivalent of the WAFTB and ABT investigated in this study in the
740 Sierra Moreno area, but also extends further east. Within the ~8–10 km wide Sierra Moreno area
741 itself, they estimate a minimum shortening of ~4 km, a value consistent with our results. This
742 study of Haschke and Günther (2003) is to our knowledge the only other work attempting to
743 estimate the minimum total shortening absorbed by the WAFTB at 20–22°S. It becomes obvious
744 that the various structures of the WAFTB in northern Chile, wherever they are (Quebrada
745 Blanca, Pinchal or Sierra Moreno areas), all absorb multi-kilometric shortening, at the scale of
746 only one to three major folds and thrusts.

747 To the West and laterally, the WAFTB is covered by Cenozoic deposits. Seismic profiles
748 from the Chilean Empresa Nacional del Petróleo (ENAP), as re-interpreted by Victor et al.
749 (2004), Jordan et al. (2010) or Labbé et al. (2019), show a series of several blind west-verging

750 thrust-faults affecting both the Cenozoic and the underlying Mesozoic units. Nonetheless,
751 deformation is mostly well-imaged for post ~29 Ma growth strata within the Cenozoic series
752 deposited above the Choja erosional surface, and remains less well-resolved for underlying
753 Mesozoic units. These observations may reflect the fact that Mesozoic units are much more
754 deformed than Cenozoic layers, a deduction in line with our own field observations in the
755 Pinchal and Quebrada Blanca areas (Figures 4 and 9). As proposed by Victor et al. (2004) and
756 Armijo et al. (2015), the west-vergent thrust-faults beneath the western Andean flank at ~20–
757 22°S, can reasonably be interpreted as connecting onto an east-dipping detachment, deepening
758 towards the mountain range, again in line with our interpretation of the structure in our study
759 areas (Figures 5c and 10c). Altogether, these data suggest that all these thrust faults, either blind
760 or deduced from outcropping folds, pertain to the same WAFTB system. The WAFTB at ~20–
761 22°S therefore extends over a much wider region (~50 km, maybe locally more) than the two
762 ~7–17 km wide sites investigated in this study (Figure 1), as most of the FTB is hidden beneath
763 the less deformed Cenozoic cover (Figure 1). This conclusion further emphasizes that the
764 minimum ~3–9 km of shortening proposed here from the folds of the Quebrada Blanca and
765 Pinchal areas (when excluding the contribution of the ABT in the Pinchal area) are clearly under-
766 estimates of the total shortening across the whole WAFTB at this latitude. A precise
767 quantification of the deformation recorded by buried folded Mesozoic units west of our study
768 sites is not possible from available seismic profiles (Victor et al., 2004; Jordan et al., 2010;
769 Labbé et al., 2019). The amount of ~20–30 km of total shortening along the western flank of the
770 Andes qualitatively estimated by Armijo et al. (2015) would be a reasonable value at scale with
771 the structural relief of the western flank and the crustal thickness beneath.

772 7.3 Temporal evolution of deformation

773 Our investigations underline that the deformation of the Quebrada Blanca and Pinchal
774 areas is not linearly distributed over time, and can be assigned to two main periods: (1) a period
775 of major deformation sometime between ~68–29 Ma (possibly ~68–44 Ma) at a minimum
776 average shortening rate of ~0.1–0.3 km/Myr; and (2) a subsequent period of moderate
777 deformation from ~29–0 Ma (starting possibly earlier) at an average rate of <0.1 km/Myr (Figure
778 13d). These deductions and rates hold for the westernmost anticline of the study sites, but the
779 reduction in deformation rates is expected at the scale of both whole investigated sites. Indeed,

780 the difference in the deformation cumulated by Mesozoic units and by post ~29 Ma Cenozoic
781 layers can be qualitatively – but clearly – intuited from our cross-sections (Figures 5, 10 and 13).
782 Westward, it may also be inferred but with less certainty from the ENAP seismic profiles (see
783 discussion above). This deformation slow-down, starting by ~29 Ma and possibly earlier, could
784 therefore be regional across the entire WAFTB.

785 The period of major deformation deduced at the scale of the WAFTB sections
786 investigated here, sometime from ~68 Ma to ~29 Ma (even possibly to ~44 Ma), is coeval to the
787 first order with the timing of basement exhumation deduced from thermochronology by Makshev
788 and Zentilli (1999), and with the timing postulated for thrusting on the ABT (see section 7.1).
789 The proposed time window is also consistent with the main Incaic phase of deformation inferred
790 by various authors as the main period of Andean mountain-building *stricto sensu* (e.g., Charrier
791 et al., 2007; Cornejo et al., 2003; Pardo–Casas & Molnar, 1987; Steinmann, 1929).

792 Based on the ENAP seismic profiles in the westward prolongation of our study areas,
793 Victor et al. (2004) investigated the folding and thrusting recorded by the growth strata of the
794 Cenozoic Altos de Pica Formation. They determined a post ~29 Ma shortening of ~3 km,
795 accommodated by several west-vergent thrusts within the ~40 km wide Atacama Bench. In both
796 our study areas, we were able to reproduce with trishear modeling the first-order pattern of the
797 slightly deformed Cenozoic growth strata over and in front of the western anticlines. We found
798 ~0.4–0.5 km of post ~29 Ma shortening on one single most frontal fault and fold for our two (~7
799 and ~17 km long) investigated sections (Figures S20, S21). These values are in overall good
800 agreement with the results of Victor et al. (2004) when setting them to the same spatial scale.
801 Compared to the minimum ~3–6 km of ante ~29 Ma shortening quantified on one single
802 structure from each study section (Figures S20, S21), the post ~29 Ma shortening is clearly of
803 limited importance.

804 The simplest interpretation would be that this post ~29 Ma decline of the shortening rate
805 results from the slow-down of the same protracted compressional event which caused the
806 formation of the west-vergent WAFTB and ABT. With the presently available data at 20–22°S,
807 we cannot exclude that this slow-down may have started even before ~29 Ma – possibly as soon
808 as ~44 Ma, or even before (section 6.3) – but definitely not afterwards. Anyhow, in the absence
809 of sedimentary markers that would provide further quantitative details on the incremental
810 deformation between ~68 Ma and ~29 Ma, the evolution of shortening cannot be quantified more

811 precisely over time, and only average rates can be proposed over the large time spans of the
812 phases of major and moderate deformation.

813 **8 Conclusion: Contribution of the western Andes to mountain-building at ~20–22°S**

814 In this study, we investigate and explore two major structural features within the western
815 flank of the Andes at ~20–22°S: (1) the Andean Basement Thrust (ABT), which stands as a
816 west-vergent, >120 km long system of ~north–south trending thrusts bringing Paleozoic
817 basement over folded Mesozoic series; (2) the West Andean Fold-and-Thrust-Belt (WAFTB),
818 which is a west-vergent FTB deforming Mesozoic and Cenozoic sediments, mostly covered by
819 the Cenozoic Altos de Pica Formation, but cropping out in few (up to ~10–20 km wide) places
820 along the mountain flank. The WAFTB extends over at least ~200 km north–south. Even though
821 our investigations only rely on two limited outcropping sites, our deductions have regional
822 implications when compared and set on scale with previous results.

823 Using field and satellite observations, we build structural cross-sections and quantify the
824 recorded shortening at two key sites along the western mountain flank. We find a minimum
825 shortening of ≥ 2.6 km on the ABT and of ≥ 3 –9 km on the few exposed structures of the
826 WAFTB. This strict minimum shortening – derived from outcrop areas of limited extent –
827 corresponds only to a fraction of the entire deformation at the scale of the whole Western
828 Cordillera at ~20–22°S. When set on scale with the extent of the investigated structures, it clearly
829 implies the possibility of multi-kilometric shortening across the western flank of the Andes.

830 We further exploit the differential deformation recorded by folded Mesozoic layers and
831 Cenozoic growth strata of the post ~29 Ma Altos de Pica Formation. We show that the
832 outcropping WAFTB was mainly active between ~68–29 Ma (possibly ~68–44 Ma), and that its
833 deformation rates significantly decreased after ~29 Ma (or starting sometime before). By
834 comparison to previous studies of the blind portions of the WAFTB west of our study sites, we
835 propose that such slow-down of deformation rates was regional rather than local. In addition,
836 field observations and published thermochronological results of basement exhumation suggest
837 that this temporal evolution of deformation rates also holds for the ABT. We therefore propose
838 that the post ~29 Ma decline in shortening rates resulted from the regional slow-down of the
839 same protracted compressional event that caused the formation of the west-vergent WAFTB and
840 ABT.

841 Even though multi-kilometric, the shortening accommodated by the west-vergent
842 structures of the western Andes outlined in this study represents a minor contribution to the total
843 crustal shortening of >300 km across the entire Central Andes at ~20°S (e.g., Anderson et al.,
844 2017; Barnes & Ehlers, 2009; Elger et al., 2005; Kley & Monaldi, 1998; McQuarrie et al., 2005;
845 Oncken et al., 2012; Sheffels, 1990). It should however be recalled that this deformation took
846 place mostly in the early stages of the Andean orogeny, sometime between ~68–29 Ma (possibly
847 ~68–44 Ma), during the so-called Incaic phase. In fact, when replaced within the temporal
848 evolution of Andean mountain-building at these latitudes (e.g., Armijo et al., 2015; Charrier et
849 al., 2007; McQuarrie et al., 2005; Oncken et al., 2006), the early multi-kilometric shortening
850 evidenced here represents a major contribution to initial Andean deformation, which has been
851 most often neglected. The slowing down of deformation across the western Andean flank by ~29
852 Ma – and possibly starting before – may have accompanied the jumping and transfer of
853 deformation towards the East on east-vergent structures (i.e. towards the eastern Altiplano and
854 further east, e.g., Isacks et al., 1988; McQuarrie et al., 2005; Oncken et al., 2006), building from
855 there a bi-vergent orogen (e.g., Armijo et al., 2015; Faccenna et al., 2013; Martinod et al., 2020).
856 As such, the early stages of Andean deformation described here at ~20–22°S recall the first-order
857 kinematics of crustal shortening deduced for the ~20–25 Myr old growing Andes at ~33.5°S
858 (Riesner et al., 2018, 2019). This analogy suggests the continuity of the WAT all along the
859 western flank of the Central Andes, re-sketching the relevance of this structure in the earlier
860 stages of the Andean orogeny.

861 **Acknowledgements**

862 This study was supported by grants from CNRS-INSU (program TELLUS-SYSTER) and from
863 the Institut de physique du globe de Paris (IPGP). Field work was also funded by the Andean
864 Tectonics Laboratory of the Advanced Mining Technology Center, University of Chile. Earlier
865 work on this zone by RL and DC was supported by ANR project MegaChile (grant ANR-12-
866 BS06-0004-02) and LABEX UnivEarthS project. TH benefitted from a PhD grant attributed by
867 the French Ministry of Higher Education and Research. Pleiades satellite imagery was obtained
868 through the ISIS program of the CNES under an academic license and is not for open
869 distribution. The authors thank A. Delorme for his technical assistance in producing the DEMs.

870 Numerical computations for the DEMs were performed on the S-CAPAD platform, Institut de
 871 physique du globe de Paris (IPGP). The kinematic modeling was made using FoldFault Forward
 872 version 6, freely available from
 873 <http://www.geo.cornell.edu/geology/faculty/RWA/programs/faultfoldforward.html>
 874 (Allmendinger, 1998). R. Armijo and the late R. Thiele are warmly thanked for the fruitful
 875 discussions that led over the years to this work and manuscript. We also benefited from
 876 discussions with C. Creixell, N. Blanco, A. Tomlinson and F. Sepulveda (SERNAGEOMIN),
 877 from the valuable help of M. Riesner for the 3D mapping, and that of L. Barrier for facies and
 878 polarity identifications. L. Barrier and N. Bellhasen are also thanked for discussions that inspired
 879 and led to trishear modeling. This study was partly supported by IdEx Université de Paris ANR-
 880 18-IDEX-0001. This is IPGP contribution number XXX.

881 **References**

- 882 Aguilef, S., Franco, C., Tomlinson, A., Blanco, N., Alvarez, J., Montecino, D., et al. (2019).
 883 *Geología del área Quehuíta-Chela, regiones de Tarapacá y Antofagasta*. Servicio
 884 Nacional de Geología y Minería (Carta Geológica de Chile, Serie Geología Básica 207:
 885 293 p., 1 mapa escala 1:100,000). Santiago, Chile.
- 886 Allmendinger, R. W. (1998). Inverse and forward numerical modeling of trishear fault-
 887 propagation folds. *Tectonics*, 17(4), 640–656. <https://doi.org/10.1029/98TC01907>
- 888 Allmendinger, R. W., & Shaw, J. H. (2000). Estimation of fault propagation distance from fold
 889 shape: Implications for earthquake hazard assessment. *Geology*, 28(12), 1099–1102.
- 890 Anderson, R. B., Long, S. P., Horton, B. K., Calle, A. Z., & Ramirez, V. (2017). Shortening and
 891 structural architecture of the Andean fold-thrust belt of southern Bolivia (21°S):
 892 Implications for kinematic development and crustal thickening of the central Andes.
 893 *Geosphere*, 13(2), 538–558. <https://doi.org/10.1130/GES01433.1>
- 894 Armijo, R., Rauld, R., Thiele, R., Vargas, G., Campos, J., Lacassin, R., & Kausel, E. (2010). The
 895 West Andean Thrust, the San Ramón Fault, and the seismic hazard for Santiago, Chile.
 896 *Tectonics*, 29(TC2007). <https://doi.org/10.1029/2008TC002427>

- 897 Armijo, R., Lacassin, R., Coudurier-Curveur, A., & Carrizo, D. (2015). Coupled tectonic
898 evolution of Andean orogeny and global climate. *Earth-Science Reviews*, *143*, 1–35.
899 <https://doi.org/10.1016/j.earscirev.2015.01.005>
- 900 Baker, M. C. W. (1977). Geochronology of upper Tertiary volcanic activity in the Andes of north
901 Chile. *Geologische Rundschau*, *66*(1), 455–465. <https://doi.org/10.1007/BF01989588>
- 902 Barnes, J., Ehlers, T., McQuarrie, N., O’Sullivan, P., & Tawackoli, S. (2008).
903 Thermochronometer record of Central Andean Plateau growth, Bolivia (19.5°S).
904 *Tectonics*, *27*(TC3003). <https://doi.org/10.1029/2007TC002174>
- 905 Barnes, J. B., & Ehlers, T. A. (2009). End member models for Andean Plateau uplift. *Earth-*
906 *Science Reviews*, *97*(1–4), 105–132. <https://doi.org/10.1016/j.earscirev.2009.08.003>
- 907 Blanco, N., & Tomlinson, A. J. (2013). *Carta Guatacondo, Region de Tarapaca*. (Carta
908 Geologica de Chile, Serie Geologia Basica). Santiago, Chile: Servicio Nacional de
909 Geología y Minería.
- 910 Blanco, N., Tomlinson, A. J., Moreno, K., & Rubilar, D. (2000). *Importancia estratigráfica de*
911 *icnitas de dinosaurios en la Fm. Chacarilla (Jurásico Superior - Cretácico Inferior), I*
912 *Región, Chile*. Presented at the IX Congreso Geológico Chileno. Servicio Nacional de
913 Geología y Minería, Puerto Varas, Chile.
- 914 Blanco, N., Vásquez, P., Sepúlveda, F., Tomlinson, A. J., Quezada, A., & Ladino, M. (2012).
915 *Levantamiento Geológico para el Fomento de la Exploración de Recursos Minerales e*
916 *Hídricos de la Cordillera de la Costa, Depresión Central y Precordillera de la Región de*
917 *Tarapacá (20°–21°S)*. (Informe Registrado IR-12-50, 246 p., 7 maps, scale 1:100,000).
918 Santiago, Chile: Servicio Nacional de Geología y Minería.
- 919 Brooks, B. A., Bevis, M., Whipple, K., Ramon Arrowsmith, J., Foster, J., Zapata, T., et al.
920 (2011). Orogenic-wedge deformation and potential for great earthquakes in the central
921 Andean backarc. *Nature Geoscience*, *4*(6), 380–383. <https://doi.org/10.1038/ngeo1143>
- 922 Buchelt, M., & Tellez, C. (1988). The Jurassic La Negra Formation in the area of Antofagasta,
923 northern Chile (lithology, petrography, geochemistry). In *The Southern Central Andes*
924 (Springer-Verlag Berlin Heidelberg, Vol. 17, pp. 169–182).
925 <https://doi.org/10.1007/BFb0045181>

- 926 Charrier, R., Pinto, L., & Rodríguez, M. P. (2007). Tectonostratigraphic evolution of the Andean
927 Orogen in Chile. In T. Moreno & W. Gibbons (Eds.), *The Geology of Chile* (First, pp.
928 21–114). The Geological Society of London. <https://doi.org/10.1144/GOCH.3>
- 929 Cornejo, P., Matthews, S., & Pérez de Arce, C. (2003). The “K-T” compressive deformation
930 event in northern Chile (24°–27°S). In E. Campos (Eds), *X Congreso Geológico Chileno*
931 *Actas*, Concepción, Chile, (Thematic Session ST1, p. 1–13,CD-ROM).
- 932 Cristallini, E. O., & Allmendinger, R. W. (2002). Backlimb trishear: a kinematic model for
933 curved folds developed over angular fault bends. *Journal of Structural Geology*, 24(2),
934 289–295. [https://doi.org/10.1016/S0191-8141\(01\)00063-3](https://doi.org/10.1016/S0191-8141(01)00063-3)
- 935 DeCelles, P. G., Zandt, G., Beck, S. L., Currie, C. A., Ducea, M. N., Kapp, P., et al. (2015).
936 Cyclical orogenic processes in the Cenozoic central Andes. In P. G. DeCelles, M. N.
937 Ducea, B. Carrapa, & P. A. Kapp, *Geodynamics of a Cordilleran Orogenic System: The*
938 *Central Andes of Argentina and Northern Chile*. Geological Society of America.
939 [https://doi.org/10.1130/2015.1212\(22\)](https://doi.org/10.1130/2015.1212(22))
- 940 DeMets, C., Gordon, R., Argus, D., & Stein, S. (1994). Effect of recent revisions to the
941 geomagnetic reversal time scale on estimates of current plate motions. *Geophysical*
942 *Research Letters*, 21. <https://doi.org/10.1029/94GL02118>
- 943 Dingman, R. J., & Galli, C. O. (1965). *Geology and Ground-Water Resources of the Pica Area,*
944 *Tarapaca Province, Chile*. Geological Survey Bulletin 1189. US Dept of Interior.
- 945 Elger, K., Oncken, O., & Glodny, J. (2005). Plateau-style accumulation of deformation: Southern
946 Altiplano. *Tectonics*, 24(4). <https://doi.org/10.1029/2004TC001675>
- 947 Erslev, E. A. (1991). Trishear fault-propagation folding. *Geology*, 19, 617–620.
948 [https://doi.org/10.1130/0091-7613\(1991\)019<0617:TFPF>2.3.CO;2](https://doi.org/10.1130/0091-7613(1991)019<0617:TFPF>2.3.CO;2)
- 949 Faccenna, C., Becker, T. W., Conrad, C. P., & Husson, L. (2013). Mountain building and mantle
950 dynamics. *Tectonics*, 32(1), 80–93. <https://doi.org/10.1029/2012TC003176>
- 951 Faccenna, C., Oncken, O., Holt, A. F., & Becker, T. W. (2017). Initiation of the Andean orogeny
952 by lower mantle subduction. *Earth and Planetary Science Letters*, 463, 189–201.
953 <https://doi.org/10.1016/j.epsl.2017.01.041>

- 954 Farías, M., Charrier, R., Comte, D., Martinod, J., & Hérail, G. (2005). Late Cenozoic
955 deformation and uplift of the western flank of the Altiplano: Evidence from the
956 depositional, tectonic, and geomorphologic evolution and shallow seismic activity
957 (northern Chile at 19°30'S): Western Altiplano Uplift. *Tectonics*, 24(4), TC4001.
958 <https://doi.org/10.1029/2004TC001667>
- 959 Galli, C., & Dingman, R. J. (1962). *Cuadrángulos Pica, Alca, Matilla y Chacarilla: Con un*
960 *estudio sobre los recursos de agua subterránea, Provincia de Tarapacá*. Instituto de
961 Investigaciones Geológicas (Carta Geológica de Chile 7–10, 125 p., 4 maps, scale
962 1:50,000).
- 963 Galli-Olivier, C. (1967). Pediplain in Northern Chile and the Andean Uplift. *Science*, 158(3801),
964 653–655. <https://doi.org/10.1126/science.158.3801.653>
- 965 Garcia, M., & Hérail, G. (2005). Fault-related folding, drainage network evolution and valley
966 incision during the Neogene in the Andean Precordillera of Northern Chile.
967 *Geomorphology*, 65(3–4), 279–300. <https://doi.org/10.1016/j.geomorph.2004.09.007>
- 968 Hardy, S., & Ford, M. (1997). Numerical modeling of trishear fault propagation folding.
969 *Tectonics*, 16(5), 841–854. <https://doi.org/10.1029/97TC01171>
- 970 Haschke, M., & Günther, A. (2003). Balancing crustal thickening in arcs by tectonic vs.
971 magmatic means. *Geology*, 31(11), 933. <https://doi.org/10.1130/G19945.1>
- 972 Heit, B., Sodoudi, F., Yuan, X., Bianchi, M., & Kind, R. (2007). An S receiver function analysis
973 of the lithospheric structure in South America. *Geophysical Research Letters*, 34(14).
974 <https://doi.org/10.1029/2007GL030317>
- 975 Homewood, P., & Lateltin, O. (1988). Classic Swiss Clastics (Flysch and Molasse) - The Alpine
976 connection. *Geodinamica Acta*, 2(1), 1–11.
977 <https://doi.org/10.1080/09853111.1988.11105150>
- 978 Isacks, B. L. (1988). Uplift of the Central Andean Plateau and bending of the Bolivian Orocline.
979 *Journal of Geophysical Research*, 93(B4), 3211–3231.
980 <https://doi.org/10.1029/JB093iB04p03211>

- 981 Kley, J., & Monaldi, C. R. (1998). Tectonic shortening and crustal thickness in the Central
982 Andes: How good is the correlation? *Geology*, *26*(8), 723–726.
983 [https://doi.org/10.1130/0091-7613\(1998\)026<0723:TSACTI>2.3.CO;2](https://doi.org/10.1130/0091-7613(1998)026<0723:TSACTI>2.3.CO;2)
- 984 Lamb, S. (2011). Did shortening in thick crust cause rapid Late Cenozoic uplift in the northern
985 Bolivian Andes? *Journal of the Geological Society*, *168*(5), 1079–1092.
986 <https://doi.org/10.1144/0016-76492011-008>
- 987 Lamb, S. (2016). Cenozoic uplift of the Central Andes in northern Chile and Bolivia –
988 reconciling paleoaltimetry with the geological evolution. *Canadian Journal of Earth
989 Sciences*, *53*(11), 1227–1245. <https://doi.org/10.1139/cjes-2015-0071>
- 990 Lucassen, F., Becchio, R., Wilke, H. G., Franz, G., Thirlwall, M. F., Viramonte, J., & Wemmer,
991 K. (2000). Proterozoic–Paleozoic development of the basement of the Central Andes
992 (18–26°S) – a mobile belt of the South American craton. *Journal of South American
993 Earth Sciences*, *13*(8), 697–715. [https://doi.org/10.1016/S0895-9811\(00\)00057-2](https://doi.org/10.1016/S0895-9811(00)00057-2)
- 994 Martinod, J., Gérard, M., Husson, L., & Regard, V. (2020). Widening of the Andes: An
995 interplay between subduction dynamics and crustal wedge tectonics. *Earth-Science
996 Reviews*, *204*, 103170. <https://doi.org/10.1016/j.earscirev.2020.103170>
- 997 McElderry, S., Chong, G., Prior, D., & Flint, S. S. (1996). *Structural styles in the Domeyko
998 range, northern Chile*. In Géodynamique andine : résumés étendus, Paris : ORSTOM,
999 439-442. Third International Symposium on Andean Geodynamics (ISAG), Saint-Malo,
1000 France, 1996/09/17-19. ISBN 2-7099-1332-1. Retrieved from
1001 https://horizon.documentation.ird.fr/exl-doc/pleins_textes/divers4/010008618.pdf
- 1002 McQuarrie, N., Horton, B. K., Zandt, G., Beck, S., & DeCelles, P. G. (2005). Lithospheric
1003 evolution of the Andean fold–thrust belt, Bolivia, and the origin of the central Andean
1004 plateau. *Tectonophysics*, *399*(1–4), 15–37. <https://doi.org/10.1016/j.tecto.2004.12.013>
- 1005 Mpodozis, C., & Ramos, V. A. (1989). The Andes of Chile and Argentina. In G. E. Ericksen, M.
1006 T. Canas Pinochet, & J. A. Reinemund (Eds.), *Geology of the Andes and its Relation to
1007 Hydrocarbon and Mineral Resources* (pp. 59–90). Circum-Pacific Council for Energy
1008 and Mineral Resources Earth Sciences Series, Houston, Texas.

- 1009 Muñoz, N., & Charrier, R. (1996). Uplift of the western border of the Altiplano on a west-
1010 vergent thrust system, Northern Chile. *Journal of South American Earth Sciences*, 9(3–
1011 4), 171–181. [https://doi.org/10.1016/0895-9811\(96\)00004-1](https://doi.org/10.1016/0895-9811(96)00004-1)
- 1012 Norabuena, E., Leffler-Griffin, L., Mao, A., Dixon, T., Stein, S., Sacks, I. S., et al. (1998). Space
1013 geodetic observations of Nazca–South America convergence across the Central Andes.
1014 *Science*, 279(5349), 358–362. <https://doi.org/10.1126/science.279.5349.358>
- 1015 Oncken, O., Boutelier, D., Dresen, G., & Schemmann, K. (2012). Strain accumulation controls
1016 failure of a plate boundary zone: Linking deformation of the Central Andes and
1017 lithosphere mechanics. *Geochemistry, Geophysics, Geosystems*, 13(Q12007).
1018 <https://doi.org/10.1029/2012GC004280>
- 1019 Oncken, Onno, Chong, G., Franz, G., Giese, P., Götze, H.-J., Ramos, V. A., et al. (Eds.). (2006).
1020 *The Andes: Active Subduction Orogeny*. Berlin, Heidelberg: Springer. Retrieved from
1021 <https://doi.org/10.1007/978-3-540-48684-8>
- 1022 Pardo-Casas, F., & Molnar, P. (1987). Relative motion of the Nazca (Farallon) and South
1023 American Plates since Late Cretaceous time. *Tectonics*, 6(3), 233–248.
1024 <https://doi.org/10.1029/TC006i003p00233>
- 1025 Ramos, V. A. (1988). Late Proterozoic - Early Paleozoic of South America – a Collisional
1026 History. *Episodes*, 11(3), 7.
- 1027 Ramos, V. A. (2008). The Basement of the Central Andes: The Arequipa and Related Terranes.
1028 *Annual Review of Earth and Planetary Sciences*, 36(1), 289–324.
1029 <https://doi.org/10.1146/annurev.earth.36.031207.124304>
- 1030 Rapela, C. W., Pankhurst, R. J., Casquet, C., Baldo, E., Saavedra, J., & Galindo, C. (1998). Early
1031 evolution of the Proto-Andean margin of South America. *Geology*, 26(8), 707–710.
- 1032 Rauld, R. A. (2011). *Deformación cortical y peligro sísmico asociado a la falla San Ramón en el*
1033 *frente cordillerano de Santiago, Chile Central (33°S)*, (Doctoral dissertation). Retrieved
1034 from http://www.tesis.uchile.cl/tesis/uchile/2011/cf-rauld_rp/html/index.html. Dep. de
1035 Geol., Univ. de Chile, Santiago.

- 1036 Riesner, M., Lacassin, R., Simoes, M., Armijo, R., Rauld, R., & Vargas, G. (2017). Kinematics
1037 of the active West Andean fold-and-thrust belt (central Chile): Structure and long-term
1038 shortening rate. *Tectonics*, 36(2), 287–303. <https://doi.org/10.1002/2016TC004269>
- 1039 Riesner, Magali, Lacassin, R., Simoes, M., Carrizo, D., & Armijo, R. (2018). Revisiting the
1040 Crustal Structure and Kinematics of the Central Andes at 33.5°S: Implications for the
1041 Mechanics of Andean Mountain Building. *Tectonics*, 37(5), 1347–1375.
1042 <https://doi.org/10.1002/2017TC004513>
- 1043 Riesner, Magali, Simoes, M., Carrizo, D., & Lacassin, R. (2019). Early exhumation of the
1044 Frontal Cordillera (Southern Central Andes) and implications for Andean mountain-
1045 building at ~33.5°S. *Scientific Reports*, 9(7972). [https://doi.org/10.1038/s41598-019-](https://doi.org/10.1038/s41598-019-44320-1)
1046 [44320-1](https://doi.org/10.1038/s41598-019-44320-1)
- 1047 Rosu, A.-M., Deseilligny, M., Delorme, A., Binet, R., & Klinger, Y. (2014). Measurement of
1048 ground displacement from optical satellite image correlation using the free open-source
1049 software MicMac. *ISPRS Journal of Photogrammetry and Remote Sensing*, 100.
1050 <https://doi.org/10.1016/j.isprsjprs.2014.03.002>
- 1051 Rupnik, E., Deseilligny, M., Delorme, A., & Klinger, Y. (2016). Refined satellite image
1052 orientation in the free open-source photogrammetric tools Apero/Micmac. *ISPRS Annals*
1053 *of Photogrammetry, Remote Sensing and Spatial Information Sciences*, III–1, 83–90.
1054 <https://doi.org/10.5194/isprs-annals-III-1-83-2016>
- 1055 SERNAGEOMIN. (2003). Mapa Geológico de Chile: versión digital. (No. 4, CD-ROM, versión
1056 1.0, 2003). Santiago, Chile: Servicio Nacional de Geología y Minería, Publicación
1057 Geológica Digital.
- 1058 Sheffels, B. M. (1990). Lower bound on the amount of crustal shortening, in the central Bolivian
1059 Andes. *Geology*, 18(9), 812–815. [https://doi.org/10.1130/0091-](https://doi.org/10.1130/0091-7613(1990)018<0812:LBOTAO>2.3.CO;2)
1060 [7613\(1990\)018<0812:LBOTAO>2.3.CO;2](https://doi.org/10.1130/0091-7613(1990)018<0812:LBOTAO>2.3.CO;2)
- 1061 Skarmenta, J., & Marinovic, N. (1981). *Hoja Quillagua*. Inst. Invest. Geol. (Carta Geol. de Chile
1062 No.51, 1:250,000)
- 1063 Steinmann, G. (1929). Geologie von Peru. *The Journal of Geology*, Vol 38, No 2, University of
1064 Chicago. <https://doi.org/10.1086/623704>

- 1065 Tassara, A., Götze, H.-J., Schmidt, S., & Hackney, R. (2006). Three-dimensional density model
1066 of the Nazca plate and the Andean continental margin. *Journal of Geophysical Research:*
1067 *Solid Earth*, 111(B09404). <https://doi.org/10.1029/2005JB003976>
- 1068 Tomlinson, A., Blanco, N., Makshev, V., Dilles, J., Grunder, A., & Ladino, M. (2001). *Geología*
1069 *de la Precordillera Andina de Quebrada Blanca-Chuquicamata, Regiones I y II (20°30'-*
1070 *22°30'S)*. Informe Registrado IR-01-20.
- 1071 Tomlinson, A. J., Blanco, N., & Ladino, M. (2015). *Carta Mamina, Región de Tarapacá*. (Carta
1072 Geológica de Chile, Serie Geología Básica, scale 1:100,000). Santiago, Chile: Servicio
1073 Nacional de Geología y Minería.
- 1074 Vergara, H., & Thomas, A. (1984). *Hoja Collacagua, Región de Tarapacá*. (scale 1:250.000).
1075 Santiago, Chile: Servicio Nacional de Geología y Minería.
- 1076 Victor, P., Oncken, O., & Glodny, J. (2004). Uplift of the western Altiplano plateau: Evidence
1077 from the Precordillera between 20° and 21°S (northern Chile): Altiplano West Flank.
1078 *Tectonics*, 23(4). <https://doi.org/10.1029/2003TC001519>
- 1079 Wölbern, I., Heit, B., Yuan, X., Asch, G., Kind, R., Viramonte, J., et al. (2009). Receiver
1080 function images from the Moho and the slab beneath the Altiplano and Puna plateaus in
1081 the Central Andes. *Geophysical Journal International*, 177(1), 296–308.
1082 <https://doi.org/10.1111/j.1365-246X.2008.04075.x>
- 1083 Yuan, X., Sobolev, S. V., Kind, R., Oncken, O., Bock, G., Asch, G., et al. (2000). Subduction
1084 and collision processes in the Central Andes constrained by converted seismic phases.
1085 *Nature*, 408(6815), 958–961. <https://doi.org/10.1038/35050073>
- 1086 Zandt, G., Velasco, A. A., & Beck, S. L. (1994). Composition and thickness of the southern
1087 Altiplano crust, Bolivia. *Geology*, 22(11), 1003–1006. [https://doi.org/10.1130/0091-](https://doi.org/10.1130/0091-7613(1994)022<1003:CATOTS>2.3.CO;2)
1088 [7613\(1994\)022<1003:CATOTS>2.3.CO;2](https://doi.org/10.1130/0091-7613(1994)022<1003:CATOTS>2.3.CO;2)
- 1089 Zehnder, A. T., & Allmendinger, R. W. (2000). Velocity field for the trishear model. *Journal of*
1090 *Structural Geology*, 22, 1009–1014.