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The western Andes at ~20–22°S: A contribution to the quantification of crustal shortening and kinematics of deformation

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1 **The western Andes at ~20–22°S: A contribution to the quantification of crustal**
2 **shortening and kinematics of deformation.**

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9 **Key Points:**

- 10 • New data on the kinematics of shortening of the West Andean Fold-and-Thrust-Belt and
11 Andean Basement Thrust at ~20–22°S
- 12 • Multi-kilometric shortening across the western Andes after ~68 Ma, implying significant
13 contribution in the early stages of Andean orogeny
- 14 • Significant slowing-down of deformation rates after ~29 Ma, starting possibly by ~44 Ma

15 **Abstract**

16 The Andes are an emblematic active Cordilleran orogen. It is admitted that mountain-building in
17 the Central Andes at $\sim 20^\circ\text{S}$ started by Late-Cretaceous to Early-Cenozoic along the subduction-
18 margin, and propagated eastward. In general, the structures sustaining the uplift of the West
19 Andean flank are dismissed, and their contribution to mountain-building remains poorly solved.
20 Here, we focus on two sites along the western Andes at $\sim 20\text{--}22^\circ\text{S}$, where structures are well
21 exposed. We combine mapping from high-resolution satellite-images with field-observations and
22 numerical trishear-forward-modeling to provide quantitative constraints on the kinematic
23 evolution of the western Andes. Our results confirm the existence of two main structures: (1) the
24 Andean Basement Thrust, a west-vergent thrust system placing Andean Paleozoic basement over
25 Mesozoic strata; and (2) a series of west-vergent folds pertaining to the West Andean Fold-and-
26 Thrust-Belt, deforming primarily Mesozoic units. Once restored, we estimate that both structures
27 accommodate together at least $\sim 6\text{--}9$ km of shortening across the sole $\sim 7\text{--}17$ km-wide
28 outcropping fold-and-thrust-belt. This multi-kilometric shortening represents only a fraction of
29 the total shortening accommodated along the whole western Andes. The timing of the main
30 deformation recorded in the fold-and-thrust-belt can be bracketed sometime between ~ 68 and
31 ~ 29 Ma – and possibly between ~ 68 and ~ 44 Ma – from dated folded geological layers, with a
32 subsequent significant slowing-down of shortening-rates. Even though negligible when
33 compared to total shortening across the whole orogen, the contribution of the structures of the
34 West Andes has been likely significant at the earliest stages of Andean mountain-building before
35 deformation was transferred eastward.

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.

46 Andean mountain-building initiated by Late Cretaceous-Early Cenozoic along the
47 western Andes of the Bolivian Orocline (between 16–22°S), and proceeded since then with the
48 progressive eastward propagation of deformation onto the South American continent (e.g.
49 Anderson et al., 2017; Armijo et al., 2015; Barnes et al., 2008; Barnes & Ehlers, 2009; Charrier
50 et al., 2007; DeCelles et al., 2015; Eichelberger et al., 2013; Elger et al., 2005; Faccenna et al.,
51 2017; Kley & Monaldi, 1998; McQuarrie et al., 2005; Oncken et al., 2006; Sheffels, 1990; and
52 references therein). Most local and mountain-wide previous studies have essentially focused on
53 the structures of the Altiplano-Puna plateau and on those of the various cordilleras to the east,
54 but those located along the western flank of the orogen have remained up to now under-studied
55 and their contribution to the significant topographic relief (Figure 1) and crustal thickness (e.g.
56 Allmendinger et al., 1990; Introcaso et al., 1992; Isacks, 1988) of this part of the orogen poorly
57 understood.

58 In most classical models of Andean mountain-building, the western flank is described as
59 a passive monoclinical-like crustal-scale flexure (e.g., Isacks, 1988; Lamb, 2011, 2016;
60 McQuarrie, 2002). However, in the late 1980's, Mpodozis and Ramos (1989) described west-
61 vergent thrusting on the western Andean margin as potential major tectonic structures. Later,
62 other authors described various thrusts, mostly west-vergent, at several localities along the
63 western Andean flank (e.g. Charrier et al., 2007; Fariás et al., 2005; Fuentes et al., 2018; Garcia
64 & Hérail, 2005; Martínez et al., 2021; Muñoz & Charrier, 1996; Victor et al., 2004), but they
65 generally gave these thrusts a minor role in the building of the western flank of the orogen. Only
66 further south, at the latitude of Santiago de Chile (~33°30'S), a clear west-verging fold-and-
67 thrust-belt (fold-and-thrust-belt hereafter simplified as FTB) has been documented along the
68 western Andes (Armijo et al., 2010; Riesner et al., 2017, 2018). This FTB emerges at the active
69 San Ramon Fault in front of the capital city of Santiago de Chile, and has absorbed a significant
70 amount of shortening (Riesner et al., 2017). It has been proposed to link this western FTB to a
71 crustal-scale west-verging thrust (the West Andean Thrust or WAT) thought to have played a
72 major role in the initiation of orogenic building (Armijo et al. 2010, Riesner et al. 2018, 2019),
73 although this interpretation is still debated (e.g. Astini & Dávila, 2010; Barrionuevo et al., 2021;
74 Lossada et al., 2020).

75 We note that at 33°30'S, the orogen is relatively younger and narrower than in the
76 Bolivian Orocline further north. In contrast, at ~20–22°S, where the Andes-Altiplano system is
77 much wider and structurally more complex, the contribution of similar west-vergent structures
78 along the western Andes is probably small compared to the >300 km total shortening (e.g.
79 Anderson et al., 2017; Barnes & Ehlers, 2009; Eichelberger et al., 2013; Elger et al., 2005;
80 Faccenna et al., 2017; Kley & Monaldi, 1998; McQuarrie et al., 2005; Oncken et al., 2006;
81 Sheffels, 1990) across the entire >650 km wide orogen, but their role at the start of orogenic
82 building may have been significant (Armijo et al., 2015). At this latitude, Victor et al. (2004)
83 showed the existence of west-vergent thrusts rooting on a deep decollement dipping eastward
84 beneath the western Andes. They also estimated that these structures absorbed ~3 km of
85 shortening. However, this relatively minor shortening only characterizes the deformation
86 affecting the post ~29 Ma Altos de Pica Formation deposited above the Choja erosional surface
87 (or Choja Pediplain). The Mesozoic series beneath this surface appear much more deformed (e.g.
88 Armijo et al., 2015; Blanco & Tomlinson, 2013) but this deformation remains to be precisely
89 described and quantified. One of the difficulties in such quantification is that a very large part of
90 the deformation is hidden under blanketing mid-upper Cenozoic deposits and volcanics (Armijo
91 et al., 2015; Fariás et al., 2005; SERNAGEOMIN, 2003; Victor et al., 2004). A quantitative
92 analysis of this deformation and its kinematics is only possible at the few sites along the western
93 flank where deformed Mesozoic series crop out and which are accessible despite the hostile
94 desert conditions in North Chile.

95 In this study, we provide quantitative data to better constrain the geometry of structures,
96 the shortening they accommodated and their kinematics of deformation over time in two of the
97 few areas along the west Andean flank where erosion of the Cenozoic units allows for exposures
98 of the underlying deformed Mesozoic layers (Figure 1). The Pinchal area, at ~21°30'S, exhibits a
99 major west-vergent thrust that brings the Paleozoic basement of the Cordillera Domeyko over a
100 FTB of Mesozoic units. These structures have never been described in detail. In the Quebrada
101 Blanca zone, ~80 km further north, the excellent exposure of the FTB affecting the Mesozoic
102 series allows for a more quantitative estimate of the shortening and of the timing of the main
103 deformation episodes. Despite our detailed and quantitative approach, these two study areas only
104 give a limited minimal vision of total deformation of this region, as their spatial extent remains
105 minor at the scale of the whole western Andean flank (Figure 1). We find that the shortening of

106 these structures is multi-kilometric, revealing that the contribution of the west Andean flank to
107 Andean mountain building is not negligible. Additionally, we show that the main deformation
108 recorded by folded Mesozoic units occurred sometime between ~68 and ~29 Ma (and possibly
109 between ~68 Ma and ~44 Ma), further emphasizing that these structures mostly participated to
110 the early stages of mountain-building.

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

112 **2.1 General geological framework**

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

117 At ~20–22°S, the mountain-belt is characterized by its largest width (>650 km), highest
118 average elevation (~4–4.5 km above sea level, hereafter a.s.l., Figure 1), thickest crust (70–80
119 km, e.g. Tassara et al., 2006; Wölbern et al., 2009; Yuan et al., 2000) and greatest total
120 shortening (>300 km, e.g. Anderson et al., 2017; Eichelberger et al., 2013; Elger et al., 2005;
121 McQuarrie et al., 2005; Sheffels, 1990). Here, the Andean margin along the western border of
122 the continent is described by three major morpho-tectonic ensembles, which are, from west to
123 east: (1) the subduction margin (including the Peru–Chile Trench, the oceanward forearc, and the
124 Coastal Cordillera that reaches altitudes >1 km and that corresponds to the former Mesozoic
125 volcanic arc); (2) the Atacama Bench or Central Depression (at an altitude of ~1 km,
126 corresponding to a modern continental forearc basin, particularly well expressed in the
127 morphology and topography of North Chile); and (3) the strictly speaking Andean orogen,
128 including the current volcanic arc and the Altiplano plateau reaching elevations over 4000 m
129 a.s.l. at ~20°S (e.g. Charrier et al., 2007; McQuarrie et al., 2005; Oncken et al., 2006). Following
130 the terminology of Armijo et al. (2010, 2015), the morpho-tectonic units located west of the
131 Andean orogen constitute the Marginal Block (i.e. the oceanward forearc, the Coastal Cordillera
132 and the Atacama Bench) (Figure 1).

133 At ~20–22°S latitude, the Andean orogen itself is composed of several major tectono-
134 stratigraphic ensembles, which are, from west to east: (1) the Western Cordillera (Figure 1),

135 including the Cordillera Domeyko and the modern volcanic arc (following here the terminology
136 of e.g. Armijo et al., 2015; McQuarrie, 2002; Eichelberger et al., 2013; Garziona et al., 2017;
137 Oncken et al., 2006); (2) the Altiplano Plateau, a high-elevation internally drained low-relief
138 basin; (3) the Eastern Cordillera, a bi-vergent portion of the East Andean FTB; (4) the
139 Interandean zone (or Cordillera Oriental); and (5) the Subandean ranges, east of which the South
140 American craton underthrusts the Andes (e.g. Armijo et al., 2015; Isacks, 1988; McQuarrie et al.,
141 2005; Oncken et al., 2012). The building of the Andean mountain-belt *stricto sensu* proceeded
142 since the Late Cretaceous - Early Cenozoic at $\sim 20\text{--}22^\circ\text{S}$ and was associated with crustal
143 shortening and thickening (e.g. Amilibia et al., 2008; Andriessen & Reutter, 1994; Armijo et al.,
144 2015; Arriagada et al. 2006; Barnes et al., 2008; Bascuñan et al., 2016; Charrier et al., 2007;
145 DeCelles et al., 2015; Faccenna et al., 2017; Henriquez et al., 2019; McQuarrie et al., 2005;
146 Mpodozis et al., 2005; Oncken et al., 2006). Based on the regional syntheses and reviews by
147 McQuarrie et al. (2005), Oncken et al. (2006), Charrier et al. (2007), Armijo et al. (2015),
148 Garziona et al. (2017) and Horton (2018), the across-strike growth of the orogen may be
149 summarized as follows: (1) by Late Cretaceous, the Mesozoic arc and backarc basin (formed
150 during the early Andean cycle) is located at the position of the present-day forearc, and most of
151 the current Andes shows mainly flat topography; (2) by Late Cretaceous - Early Cenozoic,
152 orogenic growth initiates and deformation primarily affects the western margin of the present-
153 day Altiplano; (3) by $\sim 45\text{--}30$ Ma, shortening vanishes along the western flank of the Andes, and
154 is transferred to the Eastern Cordillera; (4) by ~ 25 Ma, deformation ends in the Eastern
155 Cordillera and migrates into the Interandean Belt; (5) from ~ 10 Ma until present, deformation
156 within the Subandean Belt proceeds with the underthrusting of the Brazilian Craton beneath the
157 Andes. It is therefore clear that the Andean shortening started along the western Andes and
158 subsequently propagated eastward, progressively enlarging the orogen to form the different
159 cordilleras and the Altiplano plateau in between.

160 Different authors investigated crustal shortening and thickening at $\sim 20\text{--}22^\circ\text{S}$ at the scale
161 of the whole Andean mountain-belt. From these earlier studies, total crustal shortening is
162 estimated to ~ 360 km (e.g. Anderson et al., 2017; Barnes & Ehlers, 2009; Eichelberger et al.,
163 2013; Elger et al., 2005; Kley & Monaldi, 1998; McQuarrie et al., 2005; Sheffels, 1990). This
164 crustal shortening contributed to crustal thickening. With a crustal thickness of $\sim 70\text{--}80$ km (e.g.
165 Heit et al., 2007; Tassara et al., 2006; Wölbern et al., 2009; Yuan et al., 2000; Zandt et al., 1994)

166 beneath the Western Cordillera, the Altiplano and the Eastern Cordillera at these latitudes, the
167 crust is over-thickened compared to the ~45 km thick crust of the South America craton (e.g.
168 Wölbern et al., 2009).

169 **2.2 Geological setting of the Western flank of the Andes at ~20–22°S**

170 The Andean western flank is formed of three tectono-stratigraphic units at ~20–22°S,
171 aside from the present-day volcanic arc. Starting from the East (oldest and deepest units, exposed
172 at high altitudes) to the West (youngest units, lower altitudes), these are (Figure 1): (1) Andean
173 basement consisting of metamorphic rocks of Precambrian and Paleozoic ages; (2) volcano-
174 sedimentary deposits of Mesozoic age (Triassic–Cretaceous), folded and deformed in a FTB, and
175 (3) unconformably overlain by less-deformed mid-upper Cenozoic (Oligocene – Quaternary)
176 volcanics and sedimentary cover. Magmatic intrusions locally alter these different units, and are
177 mostly Cenozoic (SERNAGEOMIN, 2003). This along-strike structuration of the western
178 Andean flank at these latitudes is here only given to the first-order as Mesozoic strata may be
179 locally trapped in between two basement units, and Cenozoic layers may be unconformably
180 overlying older strata even to the east (Figure 1). Laterally, and in particular further south (i.e.
181 south of the city of Calama, ~22°27'), it should be noted that the structural organization of the
182 western flank of the Andes is more complex, and the description proposed here does not directly
183 apply.

184 The pre-Andean basement rocks formed during the Late Proterozoic and Paleozoic, when
185 the Amazonian craton was progressively assembled from various terranes (e.g. Charrier et al.,
186 2007; Lucassen et al., 2000; Ramos, 1988; Rapela et al., 1998). At the end of this period of
187 subduction and continental accretion, intensive magmatic activity (volcanism and major granite
188 intrusions) welded together the basement during the Late Carboniferous to Early Permian
189 (Charrier et al., 2007; Ramos, 2008; Vergara & Thomas, 1984).

190 The Mesozoic deposits (Triassic to Cretaceous), found today along the west Andean
191 flank, formed in a proto-Andean arc and backarc basin system during the early period of the
192 Andean cycle (e.g. Charrier et al., 2007; Mpodozis & Ramos, 1989). Marine and continental
193 sediments are interbedded with volcano-magmatic rocks (Aguilef et al., 2019; SERNAGEOMIN,
194 2003). These Mesozoic units attain locally thicknesses up to ≥ 10 km (e.g. Buchelt & Tellez,
195 1988; Charrier et al., 2007; Mpodozis & Ramos, 1989).

196 A regional erosional surface called the Choja Pediplain (Galli-Olivier, 1967) developed
197 during the Eocene to Early Oligocene (~50–30 Ma) (e.g. Armijo et al., 2015; Victor et al., 2004).
198 Above this angular unconformity, the up to ~1600 m thick (Labbé et al., 2019) Cenozoic
199 deposits of the Altos de Pica Formation (Galli & Dingman, 1962) are composed of continental
200 clastic sediments, interbedded with volcanic layers (Victor et al., 2004). The oldest documented
201 age within the Altos de Pica Formation is of ~24–26 Ma from dated ignimbrites (Fariás et al.,
202 2005; Victor et al., 2004). From there, an age of ~27–29 Ma for the base of the Altos de Pica
203 Formation is inferred regionally when extrapolated to the basal erosional surface using an
204 average sedimentation rate. The youngest ignimbrites within the Altos de Pica Formation are
205 dated at ~14–17 Ma (Middle Miocene) (Vergara & Thomas, 1984; Victor et al., 2004). Based
206 thereon and in addition to other younger dated ignimbrites (Baker, 1977; Vergara & Thomas,
207 1984), Victor et al. (2004) deduced from stratigraphic correlations that the development of the
208 Altos de Pica Formation finished by ~5–7 Ma (Late Miocene) at ~20–22°S. |

209 The Paleozoic basement of the Western Cordillera is disrupted at places in the form of
210 various basement highs boarded by reverse faults (Figure 1), such as the Sierra del Medio to the
211 east and the Sierra de Moreno to the the west at ~22°S (e.g. Haschke & Günther, 2003;
212 Henriquez et al., 2019; Puigdomenech et al., 2020; Tomlinson et al., 2001) – not to be confused
213 with the north–south trending strike-slip Domeyko Fault System, also called West Fissure
214 System (or Falla Oeste) (e.g. Charrier et al., 2007; Reutter et al., 1996; Tomlinson & Blanco,
215 1997a, 1997b) along the Late Cretaceous - Early Cenozoic magmatic arc, east and out of our
216 field study area. At ~21°30'S, the geological map of Skarmenta and Marinovic (1981) indicates a
217 west-vergent thrust bringing the Paleozoic basement over folded Mesozoic units. Such thrust
218 contact is in structural continuity with other similar basement thrusts locally described further
219 north and south by other authors (Aguilef et al., 2019; Haschke & Günther, 2003;
220 SERNAGEOMIN, 2003; Skarmenta & Marinovic, 1981). These basement thrusts, if pertaining
221 to a common thrust system, would imply significant crustal shortening across the western
222 Andean margin, yet to be further documented in the field.

223 Using apatite fission track dating, Makshev and Zentilli (1999) proposed significant
224 exhumation of the basement units between 50 Ma and 30 Ma, possibly related to basement
225 overthrusting. Older exhumation ages (Late Cretaceous to Early Cenozoic (U-Th)/He zircon and
226 apatite ages) are however provided by Reiners et al. (2015) for the western Andean basement at

227 ~21°42'S, but from only one sample and without modeling. Together, these ages indicate that
228 data remain missing to better quantify the exhumation, uplift and timing of deformation of the
229 basement thrusts reported along this flank of the Andes.

230 In the folded sedimentary series further west, Victor et al. (2004) determined ~3 km of
231 shortening recorded by the Cenozoic deposits of the Altos de Pica Formation, i.e. accumulated
232 between ~29 Ma and ~5–10 Ma. However, these authors did not take into account the
233 deformation of the underlying more deformed Mesozoic units. They interpret the underlying
234 structures and folding as part of a system of west-vergent thrusts, re-analyzed recently by other
235 authors (Fuentes et al., 2018; Martinez et al., 2021), but the poor quality of the seismic profiles at
236 these depths renders these interpretations quite tenuous and disputable. Haschke and Günther
237 (2003) estimated that >9 km of shortening across the western flank in the outcropping Sierra de
238 Moreno area (~21°45'S) occurred since the Late Cretaceous to Eocene on a west- and east-
239 verging thrust system. It follows that even if published data hint at the existence of a west-
240 verging fault system along the western Andean front at ~20–22°S, its geometry, kinematics and
241 total amount of shortening have not yet been satisfactorily evaluated.

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

252 **3. Data and Methods**

253 **3.1 Available Data**

254 The most detailed existing geological map for the Pinchal area is the Quillagua map
255 (1:250,000 scale, Skarmenta and Marinovic, 1981), which only provides very large-scale

256 information but hints for the existence of a major west-vergent basement thrust. For the
257 Quebrada Blanca area, the recent Guatacondo map (1:100,000 scale, Blanco & Tomlinson, 2013)
258 provides detailed and updated information on the stratigraphy and structure. There, the folded
259 Mesozoic rocks are well exposed on a relatively wide area (~15 km east–west extent) and their
260 structure has been preliminarily mapped and qualitatively described by Blanco and Tomlinson
261 (2013), Armijo et al. (2015) and Fuentes et al (2018).

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

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

276 **3.2 Establishing structural maps**

277 We establish structural maps for the two investigated sites. We use an approach based on
278 the 3D-mapping of stratigraphic layers on satellite imagery (Armijo et al., 2010; Riesner et al.,
279 2017). More precisely, layers are traced and correlated on Google Earth satellite images. The so-
280 obtained georeferenced traces are projected on the DEM-derived topographic map, and
281 compared with geological maps, mainly for stratigraphic and age references. Field observations
282 allow ground verifications and provide supplementary details, such as the existence of minor
283 thrusts and folds, the observation of polarity criteria or the local measurement of dip angles.

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

294 **3.3 Building structural cross-sections**

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

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

301 In contrast, in the Quebrada Blanca area, we mostly build our subsurface cross-section
302 from mapping on satellite imagery. Here, we follow the approach already proposed in Armijo et
303 al. (2010) and described in detail in Riesner et al. (2017). The mapped georeferenced horizons
304 are projected on the high-resolution Pléiades DEMs. Using a 3D-modeler, the horizons can be
305 visualized interactively. In order to precisely assess the local average dip and strike angles of
306 deformed Mesozoic layers, we project these layers along swath profiles chosen where Mesozoic
307 strata crop out the best, where folds are mostly cylindrical and where incision (and therefore
308 topographic relief) is most significant. It should be noticed that river incision is here significantly
309 lower (a few hundred meters at most) than at the latitude of Santiago de Chile (~33°30'S) where
310 this approach has been previously employed (Riesner et al., 2017, 2018). In any case, we
311 successfully obtain the overall sectional geometry of layers, and by comparing with the structural
312 map, we determine the approximate locations of the major synclinal and anticlinal fold axes. By
313 respecting the classical structural rule of constant layer thickness, we derive fold geometries.

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

319 **3.4 Crustal shortening and kinematic modeling**

320 We use our subsurface cross-sections to estimate the minimum shortening across the
321 investigated sites. We employ a simple line-length-balancing approach to determine shortening
322 related to folding. However, we have no precise indication on the structure and geometry of
323 layers at depth, in particular within the footwall of thrusts or within the hanging wall nearby the
324 interpreted propagating faults. Given this, thrust offsets can not be precisely documented from
325 field observations only. To solve this and get a more complete view on shortening estimates, we
326 model anticlinal geometries interpreted to be related to fault-propagation using a numerical
327 trishear approach (e.g. Allmendinger, 1998; Erslev, 1991). We use the code FaultFold Forward
328 (version 6) (Allmendinger, 1998) in order to jointly model thrust displacement and anticlinal
329 folding. Trishear models the deformation distributed within a triangular zone located at the tip of
330 a propagating fault (Erslev, 1991). This forward modeling relies on a set of parameters that are
331 here adjusted by trial and error to fit structural geometries. By adding sedimentary layers at
332 various steps during ongoing deformation, we model syntectonic deposition and subsequent
333 deformation, in order to reproduce deformation of Cenozoic layers. Additional information on
334 trishear modeling, together with the range of tested parameters, are provided in supplementary
335 material. We recognize that our best-fit model parameters may not be unique. This is not
336 expected to impact much estimated total shortening as this result depends mostly on final cross-
337 sections. This point will be further discussed in section 7.3 and in the supporting information.

338 Deformation is expressed in terms of shortening (in km) but also in terms of relative
339 shortening (in %). Relative shortening is hereafter defined as the ratio of the estimated
340 shortening by the initial length of the undeformed section.

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

342 **4.1 Field observations**

343 Our observations confirm the existence of a major basement thrust in the Pinchal area.
344 Because our observations may seem in contradiction with previous stratigraphic and structural
345 interpretations of the folded Mesozoic series along the Western Andean flank, we hereafter
346 describe in detail our field observations. We subsequently discuss and compare them to previous
347 interpretations, and propose a solution reconciling these observations with a priori regional
348 stratigraphic knowledge.

349 **4.1.1 Stratigraphic observations**

350 We propose a first-order stratigraphic column from our structural, stratigraphic and
351 sedimentary field observations. In the landscape, the three main tectono-stratigraphic units are
352 clearly distinguishable (Figure 2): (1) the metamorphic basement, (2) the continuous Mesozoic
353 sedimentary series (with a continuum from continental upward to marine facies) and (3) the
354 continental Cenozoic cover. The first-order stratigraphic column (Figure 3) is hereafter described
355 from the oldest to the youngest units. Detailed field pictures of identified and individualized
356 sedimentary formations are provided in supplementary material to complement the forthcoming
357 stratigraphic descriptions. We acknowledge not to have any constraint on the absolute ages of
358 these series, but the relative stratigraphic ages are deduced from the kilometer-scale structural
359 geometry and from clear sedimentary or structural polarity criteria observed in the field (further
360 details below). We also indicate that thicknesses are inferred only locally, and that we cannot
361 exclude thickness variations within the sedimentary series over our study area.

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

366 The older part of the outcropping Mesozoic series consists of continental deposits, with a
367 high content of Paleozoic lithics and volcano-clastic and tuffitic low-rounded conglomerates, of
368 greenish, beige and brownish colors, and clast-sizes varying from a few millimeters to few
369 decimeters (Figure S2). At places, these rocks bear sedimentary polarity criteria such as grain-

370 grading, grain-sorting, cross-bedding and tangential beds (Figure S3). In the eastern part of the
371 Pinchal area, we locally observed below this series some dark green detrital pelites (lutites)
372 (Figure S4). On the basis of petrographic and sedimentological correlations, these detrital
373 Mesozoic sediments resemble to units mapped as Triassic north of the Pinchal zone (between
374 21°–21°30'S) in the Quehuita area (Aguilef et al., 2019).

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

383 The calcareous crest is overlain by thin-bedded (cm–dm) limestone layers of rose-beige
384 color (Figure S8), over a thickness of ~50–100 m. Going upsection, the marine series becomes
385 progressively more marly, limestone layers become more rare and the color more beige,
386 evidencing a deeper marine paleo-environment bearing fossiliferous marl layers. Belemnite
387 fossils were encountered in the lower part of this limestone-to-marl sequence. Characteristic
388 calcareous oval concretions of variable diameter (cm to m) (Figure S9), are pervasive at the
389 transition from marly limestones to marls. The marls bear ammonite fossils, which we have not
390 identified. These Ammonite species could be *Perisphinctes*, *Euaspidoceras*, *Mirosphinctes* and
391 *Gregoryceras*, according to the notice of the Quillagua geological map (Skarmenta & Marinovic,
392 1981) if applicable here. In this case they would be associated with a Middle Jurassic age
393 (Bajocian to Callovian). The series from the thin-bedded limestones to the top of the beige marls
394 is ~200 m thick, along one of the canyons and sections investigated in the field (Quebrada
395 Tania).

396 Upsection, the beige marls become progressively more calcareous again, with the
397 presence of thin limestone layers (Figure S10). Finally, this marine sequence ends with black
398 marls containing layers of beige sandstones (mm to few cm – rarely dm – thick) (Figure S11),
399 indicative of a detrital component in a probable deep seated basin, comparable to the "flysch"

400 series in the Alpine basins (Homewood & Lateltin, 1988). This unit is hereafter called "black
401 flysch", and has a minimum thickness of ~50 m.

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

410 ***4.1.2 Structural Observations***

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

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

426 West of the Pinchal Thrust, a series of folds involving folded Mesozoic units is
427 observable (Figures 2 and 4). From east to west, an asymmetric and overturned syncline is first
428 found (Figure 8a), followed by a relatively symmetric anticline (Figure 8b). The eastern limb of

429 the syncline, right beneath the Pinchal Thrust, is inverted and locally highly faulted and folded
430 (Figures 5 and 7b-c). Within this inverted limb, the series goes westward (and upsection) from
431 sheared lutites beneath the Pinchal Thrust followed by Mesozoic detrital series with
432 conglomerates, to the Mesozoic marine series from the calcareous crest upsection to the marly
433 limestones. The overturned strata are steeply dipping ($50\text{--}70^\circ\text{E}$). Penetrative small-scale
434 deformation can be observed pervasively within the marine Mesozoic series, in the form of
435 numerous local small folds, kinematically indicative of an inverted fold limb (used here as a
436 structural polarity criterium) (Figure 7b), and local secondary shear zones and thrusts (Figure
437 7c).

438 Going westward, as observed in detail along Quebrada Tania (Figure 5a), the eastern part
439 of the black flysch bears small-scale folds characteristic of the inverted fold limb, whereas
440 normal limb folds (used here also as structural polarity criteria) are observed slightly further
441 west: the axis of the overturned west-vergent syncline therefore passes through the black flysch.
442 Part of the Mesozoic series is missing, as overthrusting within the flysch and (marly) limestones
443 is observed frequently along Quebrada Tania (Figure 5a). The overturned syncline is therefore
444 found to be broken by a secondary thrust fault striking approximately parallel to the Pinchal
445 Thrust and roughly coinciding with the synclinal fold axis (Figures 4-5). Westward, the normal
446 western limb of the syncline encompasses the whole Mesozoic series from the black flysch
447 down-section to the Mesozoic volcano-detrital series, with more gentle dip angles ($20\text{--}40^\circ\text{E}$)
448 (Figures 2 and 5). Penetrative deformation is observed to be limited here.

449 The continental Mesozoic layers of the normal limb of the syncline flatten toward the
450 west. The section along Quebrada Tambillo (Figure 5c) shows a broad, overall symmetrical,
451 anticlinal fold (Figure 8b). Its fold axial plane is steep, dipping $\sim 80^\circ\text{E}$. The western flank of this
452 large anticline is marked by smaller, secondary folds with westward decreasing wavelength and
453 amplitude. Field logistics did not permit further detailed structural observations within this
454 anticline.

455 The Mesozoic sediments immediately west of the basement are unconformably covered
456 by sheet-like, river-incised Cenozoic fluvial deposits, forming aggradational terraces deposited
457 above erosional surfaces at different elevations, of varying spatial extent and of probably
458 different ages (Figure 2b). The majority of these erosional surfaces shows a westward tilt (Figure

459 5c). Further west, the Cenozoic deposits become thicker and bury the westward extent of the
460 folded Mesozoic units. Westward thickening of the Cenozoic layers is clearly observable in
461 Quebrada Tambillo and indicates the presence of growth strata at the front of the anticline
462 deforming the Mesozoic series (Figures 5c and 8b).

463 ***4.1.3 Comparison to previous stratigraphic and structural interpretations***

464 In the Pinchal area, a major basement thrust was reported in the Quillagua 1:250,000
465 geological map (Skarmenta & Marinovic, 1981). In this map, the Mesozoic units are interpreted
466 as pertaining to the Jurassic Quinchamale formation, deposited in a backarc basin context and
467 composed of an Oxfordian (~157–163 Ma) and a younger Kimmeridgian (~152–157 Ma) sub-
468 unit. Based on this age interpretation and relying on a regionally established Mesozoic
469 stratigraphy where marine sequences are followed upward by younger clastic deposits,
470 Skarmenta & Marinovic (1981) interpreted the main structure of the Pinchal zone as an anticline.

471 Our field investigations confirm the existence of a basement thrust, but contradict the
472 earlier interpretation of the structure of the folded Mesozoic series and of the local Mesozoic
473 stratigraphy. Even though we do not know the absolute ages of the folded sedimentary series, our
474 structural and sedimentary field observations allow for clearly constraining the relative
475 stratigraphic ages of Mesozoic units, from either structural or sedimentary polarity criteria, and
476 unambiguously indicate that detrital continental units are here geometrically and stratigraphically
477 below a marine sequence (Figure 3). In the case that the marine strata are Jurassic in age from
478 their likely fossiliferous content, the older continental clastic units could be Triassic, by
479 comparison to recent observations not far from the Pinchal area (Aguilef et al., 2019). These ages
480 need to be confirmed by future chronological and stratigraphic analyses.

481 Given this, even though the stratigraphic sequence we observe in Pinchal may look in
482 contradiction with the regionally known stratigraphy, it may rather be viewed as complementary:
483 the detrital component observed here below marine series may be older than the marine-to-
484 continental upward succession that has been well described regionally. In this sense, the Pinchal
485 area may provide a key outcrop to refine our knowledge of older series.

486 In any case, we recall that relative ages are only needed here for the scope of this study to
487 decipher the general structure and deformation pattern.

488 **4.2 Structural Interpretation**

489 The cross-section of Quebrada Tambillo (Figure 5c) summarizes our interpretation of the
490 sub-surface structural geometry of the Pinchal area. Tectonic shortening in the Pinchal area is
491 evidenced by the presence of the Pinchal Thrust and by the folded and faulted Mesozoic strata.

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

501 Considering that the folds west of the Pinchal Thrust develop above underlying thrusts
502 that connect onto a common detachment is a reasonable and classical assumption for fold-and-
503 thrust-belts (hereafter simplified as FTB). This detachment is expected to root at least at the base
504 of the outcropping Mesozoic series, or deeper (Figure 5c). Assuming that the layer thickness is
505 constant over our study area, it can be extrapolated that such detachment is located at least 2 km
506 beneath the topographic surface (i.e. at ~ 0.2 km a.s.l.), or deeper. To the West of our field area,
507 at the front of the anticline, the small-scale folds with westward decreasing wavelength and
508 amplitude (Figure 8b) are interpreted as the possible expression of disharmonic folding within
509 the forelimb of the anticline and/or of a thrust ramping-up toward the sub-surface at the front of
510 the anticline (Figure 5c).

511 Because of the internal shortening of the thrust sheet, with the pervasive presence of
512 small-scale folding and thrusting, in particular within the inverted limb of the overthrust
513 syncline, our shortening estimate only represents a minimum value. Using the simplified cross-
514 section along Quebrada Tambillo (Figure 5c), line-length balancing results in a minimum of ~ 1
515 km of shortening absorbed by folding only across the two folds documented here, from the
516 Pinchal Thrust to the front of the anticline. A significant – but unconstrained – amount of
517 shortening related to the pervasive deformation observed in the field (Figures 5a-b and 7b-c) is to

518 be added, as well as the thrust offsets on the faults of the FTB and on the Pinchal Thrust. An
519 estimate of the contribution of thrusting within the FTB will be provided below by modeling
520 (section 6.2).

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

526 **5. Structure of the folded Mesozoic series within the Quebrada Blanca area ($\sim 20^\circ 45'S$)**

527 **5.1 Stratigraphy of the Quebrada Blanca area**

528 The stratigraphy of the western Andean flank at $\sim 20^\circ 45'S$ is well described in the
529 Guatacondo geological map (Blanco & Tomlinson, 2013). Unlike in the Pinchal area, basement
530 rocks do not crop out in the investigated zone nearby the Quebrada Blanca (Figure 9), but larger
531 scale maps (e.g. SERNAGEOMIN, 2003) show Paleozoic basement units further east and higher
532 in the topography (Figure 1).

533 The Mesozoic units of the Quebrada Blanca are of Jurassic to Cretaceous age (Blanco &
534 Tomlinson, 2013), have been deposited in a back-arc basin context in successive transgression–
535 regression sequences (Charrier et al., 2007), and are subdivided into three formations: (1) The
536 Late Oxfordian Majala Formation, a clastic unit of sandstones, shales and subordinately
537 stromatolitic limestones of transitional marine origin (Blanco et al., 2012; Blanco & Tomlinson,
538 2013; Galli-Olivier, 1967); (2) the Late Jurassic / Early Cretaceous Chacarilla Formation, a
539 continental (fluvial) clastic sequence (Blanco & Tomlinson, 2013; Dingman & Galli, 1965); and
540 (3) the Late Cretaceous Cerro Empexa Formation, an andesitic volcanic and continental
541 sedimentary unit (Blanco et al., 2000; Blanco & Tomlinson, 2013; Dingman & Galli, 1965). The
542 Majala and Chacarilla Formations are both of reddish and beige colors and predominantly bear
543 detritic sediments. The Cerro Empexa Formation appears greyish and massive in the field. In the
544 Quebrada Blanca area, uranium-lead (U/Pb) dated zircons from this formation bear ages between
545 ~ 75 and ~ 68 Ma (Blanco et al., 2012; Blanco & Tomlinson, 2013; Tomlinson et al., 2015)
546 (Figure 9).

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

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

554 **5.2 Structural observations**

555 The structural map of the Quebrada Blanca area (Figure 9) highlights the main
556 stratigraphic and structural elements observed in the field and by mapping from satellite imagery.
557 Although the cartography of the folds is complicated by the persistent Cenozoic cover (notably
558 in the west and south), and by magmatic intrusions and hydrothermalism (particularly to the
559 east), three large-scale folds are clearly observable: a wide syncline in the center (Higueritas
560 syncline), bounded by two anticlines to the west (Chacarilla anticline) and east (fold names from
561 Blanco & Tomlinson, 2013; Fuentes et al. 2018). The scale of these major folds is multi-
562 kilometeric (Figure 9). The cross-section of Figure 10 illustrates the asymmetry of the folds. Both
563 anticlinal folds have steeper western limbs (dip angles vary mostly between ~50–80°W), whereas
564 their eastern limbs have more gentle dip angles (varying mostly between ~20–50°W) (Figure
565 10a). Despite the fact that the eastern flank of the eastern anticline is widely hidden by magmatic
566 intrusions and hydrothermalism, its southern part is well observed in the field and mapped
567 (Figure 9). The central Higueritas syncline is wider and more symmetric, with dip angles of ~40–
568 50° on both limbs. The anticlines involve the Majala and Chacarilla Formations, while the core
569 of the syncline bears the Cerro Empexa Formation. Overall, the folded series – and in particular
570 the anticlines – document a clear west-vergence of the folds (Figure 10c). From our projection of
571 the strata mapped on satellite imagery, the Mesozoic series are observed to be all concordant
572 (Figures 10a-b). The cross-section of the Guatacondo map (Blanco & Tomlinson, 2013) proposes
573 an angular unconformity of <10° between the Jurassic and Cretaceous units, at the base of the
574 Cerro Empexa formation, however not observed here from our large-scale high-resolution
575 mapping. As this unconformity does not produce any evident change in the geometry of layers

576 from Jurassic to Cretaceous, we consider it to be minor for our analysis, in particular with respect
577 to the main large-scale folding documented here.

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

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

591 **5.3 Structural interpretations**

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

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

607 **6. Kinematics of shortening of the folds and thrusts at the Pinchal and Quebrada** 608 **Blanca sites**

609 **6.1 Timing of deformation**

610 The time frame for the tectonic deformation observed within the two investigated sites
611 can be bounded from our data, and more specifically from our results at the Quebrada Blanca site
612 (Figure 10c). Indeed, our field observations and 3D-mapping do not reveal any relevant angular
613 unconformity within the folded Mesozoic series in both study areas. The main deformation and
614 folding of the investigated FTB therefore post-dates the deposition of these series. In the
615 Quebrada Blanca area, the youngest folded Mesozoic layers that form the core of the mapped
616 syncline belong to the Cerro Empexa Formation and bear maximum U-Pb ages of 68.9 ± 0.6 Ma
617 and 68 ± 0.4 Ma (Blanco & Tomlinson, 2013) (Figures 9 and 10c). Here, we can therefore
618 conclude that the main deformation of the documented folds post-dates ~68 Ma. This does not
619 preclude that minor deformation happened locally earlier, as suggested by the minor
620 unconformities reported from previous field works (Blanco & Tomlinson, 2013; Martínez et al.,
621 2021).

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

628 Even though we suspect that the deformed series of the Pinchal zone are Triassic to
629 Jurassic (section 4.1), we do not have any absolute ages of the folded units. Therefore, we
630 postulate that the main deformation here also post-dates ~68 Ma by analogy to our observations
631 at the Quebrada Blanca.

632 The FTB is unconformably covered by the Cenozoic deposits of the Altos de Pica
633 Formation at both investigated sites. This is also the case for the Pinchal Thrust and secondary
634 thrusts at few places in the Pinchal zone (Figure 4). The presence of growth strata at the front of
635 the westernmost anticlines in both study areas, over the erosional Choja Pediplain, suggests that
636 some deformation proceeded after ~ 29 Ma, during deposition of the Altos de Pica Formation.
637 However, the deformation recorded by folded Mesozoic layers appears of greater intensity than
638 that of the Cenozoic growth layers (Figures 5c and 10c).

639 Given this, we propose that the timing of main folding of the Mesozoic layers forming
640 the documented sections of the FTB at ~ 20 – 22° S can be loosely bracketed to a maximum time
641 span of ~ 40 Myr, sometime between ~ 68 Ma and ~ 29 Ma, with additional relatively minor
642 deformation after ~ 29 Ma. Possibly, the main deformation period could be even shorter (~ 24
643 Myr), sometime between ~ 68 Ma and ~ 44 Ma, with minor shortening after the Eocene
644 intrusions. In the case of the Pinchal Thrust, we can only propose from our observations that
645 thrusting took place prior to ~ 29 Ma.

646 **6.2 Further constraints on total shortening deduced from trishear modeling**

647 Line-length-balancing only reveals a fraction of the shortening related to the folding of
648 the Mesozoic series. Because the deduced underlying faults of the FTBs have not reached the
649 surface (Figures 5c and 10c), we assume fault-propagation-folding to be the dominant mode of
650 deformation in the studied FTBs. To further explore and quantify the associated deformation, we
651 use kinematic trishear modeling (e.g. Allmendinger, 1998; Erslev, 1991) of the westernmost
652 anticlines documented at the Quebrada Tambillo (Pinchal area) and Quebrada Blanca. This
653 approach accounts for slip on propagating thrust-faults and models the deformation distributed at
654 the tip of these evolving faults. The trishear formalism relies on a set of parameters that are
655 adjusted here by trial and error so as to fit the deduced structural geometries of the modeled
656 anticlines. The values of these parameters are within the range considered in previous studies
657 (e.g. Allmendinger, 1998; Allmendinger & Shaw, 2000; Cristallini & Allmendinger, 2002;
658 Hardy & Ford, 1997; Zehnder & Allmendinger, 2000). Here we present our best-fitting model,
659 which allows for reproducing satisfactorily our structural results, acknowledging that it is most
660 probably not unique. From there, we further discuss the kinematics of the investigated sites.

661 Further details are provided in supplementary material. Tables S1 to S3 provide the set of
662 parameters used for our modeling.

663 The structural geometries of the westernmost anticlines of the two investigated sites are
664 reproduced, and the evolution of deformation is modeled over time taking into account the
665 Cenozoic growth strata. The final geometries of our best-fitting models are reported over our
666 cross-sections, represented in Figure 13. The various stages of deformation are shown in Figures
667 S13 and S14 in the supplementary material. We find that the geometries of the western anticlines
668 can be reproduced with a cumulative shortening of 3.1 km for Quebrada Tambillo (Pinchal area),
669 and of 6.6 km for Quebrada Blanca (Figure 13). These values account for both thrusting and
670 folding across the western anticlines. They are however minimum shortening values as (1) the
671 depth of the detachment considered for modeling is minimal and could be deeper than the base of
672 the outcropping units, and (2) the model formalism does not account for small-scale internal
673 deformation, especially within the forelimb of the anticlines where the ramps approach the
674 surface.

675 The above shortening values deduced from trishear modeling only account for the
676 deformation (folding and thrusting) absorbed across the westernmost anticlines of our two
677 investigated sites. The synclinal folding accounts for a shortening of ~ 0.4 km as deduced by line-
678 length-balancing in the Pinchal area, leading to a minimum amount of shortening of ~ 3.5 km
679 across the whole Quebrada Tambillo section. This includes folding of the outcropping FTB, as
680 well as slip on the detachment and western thrust ramp. When adding the minimum ~ 2.6 km of
681 thrusting deduced on the Pinchal Thrust, we get a minimum shortening of ~ 6.1 km across the
682 whole Pinchal area. Similarly, in the Quebrada Blanca area, the easternmost anticline and
683 syncline take up ~ 2 km of shortening deduced by line-length-balancing, leading to a minimum
684 amount of shortening of ~ 8.6 km across the whole Quebrada Blanca section, including folding of
685 the outcropping FTB in addition to slip on the underlying detachment and western ramp.

686 The two investigated FTBs take up differing amounts of minimum shortening. These
687 variations may relate to the disparate extents of outcropping structures, in particular because the
688 scale of the two sections are significantly different, in terms of across-strike width (~ 7 km long
689 section for the Quebrada Tambillo vs. ~ 17 km long section for the Quebrada Blanca). Indeed, the
690 calculated shortenings similarly represent $\sim 47\%$ and $\sim 34\%$ of minimum shortening when scaled

691 to the extent of the Quebradas Tambillo and Blanca sections, respectively. Differences between
692 these sections may also relate to the depth of the underlying detachment (altitude of ~ 0.2 km for
693 Quebrada Tambillo vs. depth of ~ 2 km for Quebrada Blanca, relative to sea level) (Figure 13).
694 Lateral variations in deformation can also not be excluded.

695 **6.3 Kinematics of shortening**

696 Trishear modeling allows for simulating the evolution of thrust slip and folding in the
697 case of the westernmost anticlines of the two investigated sites. By adding syntectonic layers
698 while deformation proceeds, we also reproduce the overall geometry of the base of the Cenozoic
699 Altos de Pica Formation deposits and of the subsequent growth strata (Figures S13 and S14).
700 Syntectonic surfaces and layers are prescribed an initial $3\text{--}6^\circ$ W dipping angle, similar to the
701 present-day overall regional topographic slope (Figure 1). From there, we find that ~ 0.5 km and
702 ~ 0.4 km of shortening are needed to reproduce to the first order the geometry of the base of the
703 Altos de Pica Formation deposits at the front of the Quebrada Tambillo (Pinchal area) and
704 Quebrada Blanca sections, respectively, using the trishear models adjusted to the final cross-
705 sections. When compared to the minimum 3.1 km and 6.6 km of total shortening cumulated since
706 ~ 68 Ma across the westernmost anticlines of these two sections, this indicates that the ~ 29 Ma
707 old basal Cenozoic layers above the Choja surface only record at most 16% and 6% of this total
708 shortening, respectively. We have tested the possibility of initial horizontal Cenozoic syntectonic
709 layers. In this case, a post- ~ 29 Ma shortening of 0.8 km at most is needed to best adjust the
710 observed geometry of the basal Altos de Pica Formation layers, even though a good fit to both
711 the geometry of the growth strata and of the finite fold structure cannot be satisfactorily found.

712 These results are then used to quantitatively describe the evolution of shortening over
713 time across the westernmost anticlines of the two investigated sections, with account on the
714 timing of deformation discussed in section 6.1 (Figure 13d). We find that shortening rates were
715 on average of $\sim 0.07\text{--}0.16$ km/Myr over the time span $\sim 68\text{--}29$ Ma. They could have been even as
716 high as $\sim 0.11\text{--}0.26$ km/Myr if considering that the main deformation phase is confined to $\sim 68\text{--}$
717 44 Ma. Subsequently, deformation rates decreased to an average value of ~ 0.015 km/Myr after
718 ~ 29 Ma, starting possibly earlier.

719 It should be noted that these average values are most probably minimum values. Indeed,
720 thrusting and folding are here only modeled for the westernmost anticlines of our study sites, and

721 do not account for the shortening cumulated neither across the other structures of the FTB nor on
722 the Pinchal Thrust. Also, the main phase of deformation prior to ~29 Ma could have lasted less
723 than the ~68–29 Ma or ~68–44 Ma time intervals, respectively (Figure 13d).

724 Our results therefore quantitatively emphasize our former qualitative conclusion that the
725 major phase of deformation recorded at the two investigated sites occurred sometime between
726 ~68 and ~29 Ma, with a significant subsequent slowing down of deformation rates afterwards,
727 possibly as soon as ~44 Ma or earlier (Figure 13d).

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729 **7. Discussion**

730 **7.1 The Andean Basement Thrust**

731 *7.1.1 Evidencing a large basement thrust system along the West Andean flank* 732 *(~20–22°S)*

733 We have further documented in the Pinchal zone the existence of a west-vergent
734 basement thrust – the Pinchal Thrust along the western flank of the Andes, after its initial
735 pointing out on earlier geological maps. Here, this thrust brings basement units of the Sierra de
736 Moreno westward over folded Mesozoic units. Our study in the Pinchal area suggests that this
737 thrust bears local complexities with several strands and minor splays, most probably related to
738 the reactivation of structures in the initial pre-Andean back-arc basins. Laterally, the geological
739 map of Skarmenta and Marinovic (1981), on which we based our investigations of the Pinchal
740 Thrust, clearly documents this structure from ~21°15'S to 21°35'S, and possibly down to ~22°S
741 with some structural complexities by ~21°35'S with the junction of two possible strands of this
742 basement thrust.

743 Structurally similar basement thrust segments have been described along the Cordillera
744 Domeyko between ~20°S and ~22°S. North of the map by Skarmenta and Marinovic (1981), the
745 Quehuita (up to ~21°11'S) and Choja (between ~21°08'S–21°01'S) Faults are west-vergent
746 thrusts similarly bringing basement units over folded Mesozoic sediments (Aguilef et al., 2019).
747 North of ~21°S, intrusions, hydrothermalism and surface volcanics hamper any clear observation
748 of similar basement thrusts. Such basement thrust, if existent, would however provide a

749 reasonable mechanism for the exhumation and exposure of basement rocks east of the folded
750 Mesozoic units and at higher elevations, at the latitude of Quebrada Blanca (~20°45'S) (Figure
751 1). For these reasons, we cannot tell with any certainty whether a thrust contact similar to that
752 described in Pinchal (this study) and further north (Aguilef et al., 2019) exists at this latitude, but
753 such structure is to be suspected.

754 South of the map by Skarmenta and Marinovic (1981), in the Sierra de Moreno at
755 ~21°45'S, Haschke and Günther (2003)'s section report a basement thrust over folded Mesozoic
756 units, in agreement with the style of deformation documented here, but with a relatively minor
757 displacement on this thrust compared to our results in the Pinchal area. This thrust is here called
758 the Sierra de Moreno Thrust. Together with the (1:1,000,000) Geological map of Chile
759 (SERNAGEOMIN, 2003), Haschke and Günther (2003)'s map suggests that this basement thrust
760 is cartographically continuous southward to the southern end of the Sierra de Moreno, at
761 ~22°05'S. This possibly documents the lateral termination of this basement thrust.

762 As a conclusion, there exists a thrust system formed of various basement thrusts all along
763 the Western Andean flank, bringing the basement of the Cordillera Domeyko westward over
764 folded Mesozoic units – and therefore contributing to the uplift of the western margin of the
765 Altiplano. This thrust system is segmented, with various strands mapped as local basement
766 faults, as in our study (Figure 4) or in other maps (Aguilef et al., 2019; Haschke & Günther,
767 2003; SERNAGEOMIN, 2003; Skarmenta & Marinovic, 1981; Tomlinson et al., 2001).
768 Altogether these various thrust segments document a much larger thrust system extending
769 laterally over at least ~120 km (Figure 1), that we propose to name here the Andean Basement
770 Thrust (hereafter ABT) system.

771 We interpret the ABT and the other thrusts west of it to root onto a low-angle, eastward
772 dipping décollement, situated >2 km (Pinchal area) or >4 km (Quebrada Blanca area) beneath the
773 present-day topographic surface. Deeper and eastward, this décollement probably steepens and
774 forms a crustal-scale ramp, needed to sustain the uplift and topographic rise of the Western
775 Andes, as proposed by Victor et al. (2004) and Armijo et al. (2015). Such crustal-scale structure
776 has been termed the West Andean Thrust (or WAT) by Armijo et al. (2015).

7.1.2 Shortening and timing of deformation of the Andean Basement Thrust

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From the likely minimum offset of the basement in the Pinchal area we estimated that the Pinchal Thrust (as part of the ABT system) alone accommodated locally a strict minimum of ~2.6 km of shortening on a horizontal distance of ~1 km. This multi-kilometric shortening would be associated with multi-kilometric basement exhumation, but only limited thermochronological data actually permit to evaluate the actual amount of exhumation. These data are presently absent locally in Pinchal, but sparsely exist at a regional scale when considering the ABT system over its whole extent. From apatite fission track dating in basement samples taken ~20 km east and south-east of our study sites of the Pinchal and Quebrada Blanca areas, Makshev and Zentilli (1999) inferred at least 4–5 km of basement exhumation occurring between ~50–30 Ma. This is in good agreement with our results in terms of amount of uplift that would result from basement overthrusting on the ABT and above the WAT. Older thermochronological ages – (U-Th)/He zircon and apatite ages of ~91 Ma and ~57 Ma, respectively – were however found by Reiners et al. (2015) from the basement of the Quebrada Arcas, ~30 km south of our study site, in a structural setting equivalent to the basement of the Pinchal zone documented here. These ages do not contradict the previous estimates on total exhumation by Makshev and Zentilli (1999), even though modeling would be needed here to precisely test this. However, they question the exact timing of basement exhumation, and, from there, of thrusting over the ABT. In the absence of properly analyzed and modeled samples closer to the ABT, it is difficult to assess more precisely its timing or amount of exhumation, uplift and shortening.

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At a few places, the Pinchal segment of the ABT is covered by Cenozoic deposits. Given this observation and with existing thermochronological ages, we postulate that the ABT was most probably active sometime by Early Cenozoic to possibly Late Cretaceous – and that its activity had ceased by Early Miocene. This suggests it may have been coeval with deformation of the FTB documented immediately further west – or starting slightly before.

802 **7.2 The West Andean Fold-and-Thrust-Belt at ~20–22°S**

803 *7.2.1 Evidencing a west-vergent fold-and-thrust belt along the West Andean* 804 *flank (~20–22°S)*

805 The series of west-vergent folds described in this study at ~20–22°S as deforming
806 Mesozoic units are interpreted to form above thrusts that root at depth onto a common east-
807 dipping decollement, as proposed classically in FTBs. A similar system of folds and faults
808 affecting Mesozoic units is expected to extend laterally further north and south than just the two
809 sites described here, most probably over our entire study zone of ~20–22°S (Figure 1), even
810 though a large part north of the Quebrada Blanca area is covered by Cenozoic strata. This is
811 deduced from existing maps and previous works (e.g. Aguilef et al., 2019; Haschke & Günther,
812 2003; SERNAGEOMIN, 2003; Skarmenta & Marinovic, 1981). It therefore probably spreads out
813 over a north-south distance of at least ~200 km – and possibly more as folded Mesozoic
814 sediments are mapped on the (1:1,000,000) Geological map of Chile (SERNAGEOMIN, 2003)
815 in the north- and south-ward continuation of the two zones investigated here.

816 Further west, structures at depth are covered by Cenozoic deposits. Seismic profiles from
817 the Chilean Empresa Nacional del Petroleo (ENAP), as re-interpreted by Victor et al. (2004),
818 Jordan et al. (2010), Fuentes et al. (2018), Labbé et al. (2019) or Martínez et al. (2021), show a
819 series of several blind mostly west-verging thrust-faults affecting both the Cenozoic and the
820 underlying Mesozoic units. Nonetheless, deformation is mostly well-imaged for post ~29 Ma
821 growth strata within the Cenozoic series deposited above the Choja erosional surface, and
822 remains less well-resolved for underlying Mesozoic units. These observations may reflect the
823 fact that Mesozoic units are much more deformed than Cenozoic layers, a deduction in line with
824 our own field observations in the Pinchal and Quebrada Blanca areas (Figures 4 and 9). As
825 proposed by Victor et al. (2004), these blind west-vergent thrust-faults at ~20–22°S can
826 reasonably be interpreted as connecting onto an east-dipping detachment, deepening towards the
827 mountain range, again in line with our interpretation of the structures in our field study areas
828 (Figures 5c and 10c).

829 Altogether, these data suggest that all these thrust faults, either blind or deduced from
830 outcropping folds, pertain to the same FTB system. By analogy to what has been proposed at the
831 latitude of Santiago de Chile (~33.5°S) (Armijo et al., 2010; Rault, 2011; Riesner et al., 2017;

832 Riesner et al., 2018), we propose to name this FTB hereafter as the West Andean Fold-and-
833 Thrust-Belt (WAFTB). The WAFTB at $\sim 20\text{--}22^\circ\text{S}$ therefore extends laterally over at least ~ 200
834 km, and across-strike over a much wider region (~ 50 km, maybe locally more) than the two $\sim 7\text{--}$
835 17 km wide sites investigated in this study (Figure 1), as most of the FTB is hidden beneath the
836 less deformed Cenozoic cover (Figure 1).

837 ***7.2.2 Shortening across the Western Andean Fold-and-Thrust Belt ($\sim 20\text{--}22^\circ\text{S}$)***

838 The WAFTB of northern Chile ($\sim 20\text{--}22^\circ\text{S}$) accommodates a minimum shortening of $\sim 3\text{--}$
839 9 km, as quantified from the $\sim 7\text{--}17$ km wide cross-sections representative of the two investigated
840 areas (not including the contribution of the ABT in the case of the Pinchal area). Few authors
841 attempted to quantify the shortening in this part of the Andes. At $20^\circ 30'\text{S}$, Victor et al. (2004)
842 only evaluated the deformation affecting the post ~ 29 Ma deposits and not the total shortening as
843 could be derived from folded Mesozoic series. At $\sim 21^\circ 45'\text{S}$, ~ 30 km south of the Pinchal area,
844 Haschke and Günther (2003) reported a minimum shortening of >9 km from a ~ 50 km wide
845 cross-section, but without providing nor discussing the data used to make this estimate. Their
846 section encompasses an equivalent of the WAFTB and ABT investigated in this study in the
847 Sierra de Moreno area, but also extends further east. Within the $\sim 8\text{--}10$ km wide Sierra de
848 Moreno area itself, they estimate a minimum shortening of ~ 4 km (i.e. a minimum of $\sim 30\%$ of
849 shortening), a value consistent with our results. This study of Haschke and Günther (2003) is to
850 our knowledge the only other work attempting to estimate the minimum total shortening
851 absorbed by the WAFTB at $20\text{--}22^\circ\text{S}$. It becomes obvious that the various structures of the
852 WAFTB in northern Chile, wherever they are (Quebrada Blanca, Pinchal or Sierra Moreno
853 areas), all absorb multi-kilometric shortening, at the scale of only one to three major folds and
854 thrusts.

855 This conclusion further emphasizes that the minimum $\sim 3\text{--}9$ km of shortening proposed
856 here from the folds of the Quebrada Blanca and Pinchal areas (when excluding the contribution
857 of the ABT in the Pinchal area) are under-estimates of the total shortening across the whole
858 WAFTB at this latitude. When applying the minimum $\sim 34\text{--}47\%$ shortening estimated across our
859 two investigated sites to the ~ 50 km across-strike extent of the whole WAFTB, we find a
860 minimum crustal shortening of $\sim 26\text{--}44$ km, a value consistent – even though in the high range –
861 with the $\sim 20\text{--}30$ km qualitatively estimated by Armijo et al. (2015) by scaling with structural

862 relief and crustal thickness. A precise quantification of the deformation recorded by buried
863 folded Mesozoic units west of our study sites is at the moment not possible from available
864 seismic profiles (Victor et al., 2004; Jordan et al., 2010; Labbé et al., 2019).

865 **7.3 Temporal evolution of deformation along the western Andes (~20–22°S)**

866 Our investigations underline that the deformation of the Quebrada Blanca and Pinchal
867 areas is not linearly distributed over time, and can be assigned to two main periods: (1) a period
868 of major deformation sometime between ~68–29 Ma (possibly ~68–44 Ma) at a minimum
869 average shortening rate of ~0.1–0.3 km/Myr; and (2) a subsequent period of moderate
870 deformation from ~29–0 Ma (starting possibly earlier) at an average rate of <0.1 km/Myr (Figure
871 13d). These deductions and rates hold for the westernmost anticline of the study sites, but the
872 reduction in deformation rates is expected at the scale of both whole investigated sites. Indeed,
873 the difference in the deformation cumulated by Mesozoic units and by post ~29 Ma Cenozoic
874 layers can be qualitatively – but clearly – intuited from our cross-sections (Figures 5, 10 and 13).
875 Westward, it may also be inferred but with less certainty from the ENAP seismic profiles (see
876 discussion above). This deformation slow-down, starting by ~29 Ma at latest and possibly earlier
877 by ~44 Ma, could therefore be regional across the entire WAFTB.

878 This reasoning essentially applies to the WAFTB but may also hold for the ABT. If the
879 age of basement thrusting is not precisely known, it most probably occurred by Early Cenozoic
880 (Maksaev & Zentilli, 1999) or even Late Cretaceous - Early Cenozoic (deduced after Reiners et
881 al., 2015), and had ceased by ~29 Ma (see discussion in section 7.1).

882 This proposed time window for major folding and possibly for thrusting over the ABT is
883 generally consistent with the main Incaic phase of deformation inferred by various authors as the
884 main period of Andean mountain-building *stricto sensu* (e.g. Charrier et al., 2007; Cornejo et al.,
885 2003; Pardo-Casas & Molnar, 1987; Steinmann, 1929).

886 Based on the ENAP seismic profiles in the westward prolongation of our study areas,
887 Victor et al. (2004) investigated the folding and thrusting recorded by the growth strata of the
888 Cenozoic Altos de Pica Formation. They determined a post ~29 Ma shortening of ~3 km,
889 accommodated by several west-vergent thrusts within the ~40 km wide Atacama Bench. In both
890 our study areas, we were able to reproduce with trishear modeling the first-order pattern of the

891 slightly deformed Cenozoic growth strata over and in front of the western anticlines. We found
892 ~0.4–0.5 km of post ~29 Ma shortening on one single most frontal fault and fold in the case of
893 our two investigated sections (Figures S12, S13), that is over a distance of ~5–8 km. These
894 values are in overall good agreement with the results of Victor et al. (2004) when setting them to
895 the same spatial scale, as they together consistently represent ~6–8% of shortening. Compared to
896 the minimum ~3–6 km of ante- ~29 Ma shortening (or ~34–47% of shortening) quantified on
897 one single structure from each study section (Figures S12, S13), the post ~29 Ma shortening is
898 clearly of limited importance.

899 The simplest interpretation would be that this post ~29 Ma decline of the shortening rate
900 results from the slow-down of the same protracted regional compressional event which caused
901 the formation of the west-vergent WAFTB and ABT. With the presently available data at 20–
902 22°S, we cannot exclude that this slow-down may have started before ~29 Ma – possibly as soon
903 as ~44 Ma, or even before (section 6.3) – but definitely not afterwards. Anyhow, in the absence
904 of sedimentary markers that would provide further quantitative details on the incremental
905 deformation between ~68 Ma and ~29 Ma, the evolution of shortening cannot be quantified more
906 precisely over time, and only average rates can be proposed over the large time spans of the
907 phases of major and moderate deformation.

908 **7.4 Regional implications**

909 Even though multi-kilometric, the shortening accommodated by the west-vergent
910 structures of the western Andes outlined in this study represents a modest contribution to the
911 total crustal shortening of >300 km across the entire Central Andes at ~20°S (e.g. Anderson et
912 al., 2017; Barnes & Ehlers, 2009; Eichelberger et al., 2013; Elger et al., 2005; Faccenna et al.,
913 2017; Kley & Monaldi, 1998; McQuarrie et al., 2005; Oncken et al., 2012; Sheffels, 1990). It
914 should however be recalled that the deformation absorbed across the western Andes took place
915 mostly in the early stages of the Andean orogeny, sometime between ~68–29 Ma (possibly ~68–
916 44 Ma) in the case of the WAFTB, starting possibly earlier for the ABT – in any case during the
917 Incaic phase. In fact, when replaced within the temporal evolution of Andean mountain-building
918 at these latitudes (e.g. Armijo et al., 2015; Charrier et al., 2007; McQuarrie et al., 2005; Oncken
919 et al., 2006), the early multi-kilometric shortening evidenced here represents in fact a major
920 contribution to initial Andean deformation, which has been most often neglected in orogen-wide

921 studies. The slowing down of deformation across the western Andean flank by ~29 Ma – and
922 possibly starting after ~44 Ma – may have accompanied the jumping and transfer of deformation
923 towards the East (i.e. towards the eastern Altiplano and further east, e.g. Isacks et al., 1988;
924 McQuarrie et al., 2005; Oncken et al., 2006).

925 **Conclusion**

926 In this study, we investigate and explore two major structural features within the western
927 flank of the Andes at ~20–22°S: (1) the Andean Basement Thrust (ABT), which stands as a
928 west-vergent, >120 km long system of ~north-south trending thrusts bringing Paleozoic
929 basement over folded Mesozoic series; (2) the West Andean Fold-and-Thrust-Belt (WAFTB),
930 which is a west-vergent FTB deforming Mesozoic and Cenozoic sediments, mostly covered by
931 the Cenozoic Altos de Pica Formation, but cropping out in few (up to ~10–20 km wide) places
932 along the mountain flank. The WAFTB extends over at least ~200 km north-south and ~50 km
933 across-strike. Even though our investigations only rely on two limited outcropping sites, our
934 deductions have regional implications when compared and up-scaled with previous results.

935 Using field and satellite observations, we build structural cross-sections and quantify the
936 recorded shortening at two key sites along the western mountain flank. We find a minimum
937 shortening of ≥ 2.6 km on the ABT and of ≥ 3 –9 km on the few exposed structures of the
938 WAFTB. This strict minimum shortening – derived from outcrop areas of limited extent –
939 corresponds only to a fraction of the entire deformation at the scale of the whole Western
940 Cordillera at ~20–22°S. When set on scale with the extent of the investigated structures, it clearly
941 implies the possibility of multi-kilometric shortening across the western flank of the Andes,
942 possibly up to 26–44 km or more.

943 We further exploit the differential deformation recorded by folded Mesozoic layers and
944 Cenozoic growth strata of the post ~29 Ma Altos de Pica Formation. We show that the
945 outcropping WAFTB was mainly active between ~68–29 Ma (possibly ~68–44 Ma), and that its
946 deformation rates significantly decreased after ~29 Ma (a decrease that may have started
947 sometime earlier, e.g. by ~44 Ma). By comparison to previous studies of the blind portions of the
948 WAFTB west of our study sites, we propose that such slowing-down of deformation rates was
949 regional rather than local. In addition, field observations and published thermochronological
950 results of basement exhumation suggest that this temporal evolution of deformation rates may

951 also hold for the ABT. We therefore propose that the post ~29 Ma (or post ~44Ma) decline in
952 shortening rates resulted from the regional slowing-down of the same protracted compressional
953 event that caused the formation of the west-vergent WAFTB and ABT, most probably
954 accompanying the transfer of Andean deformation towards the Altiplano Plateau, Eastern
955 Cordillera, and further eastward.

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973 **Open Research: Data Availability Statement**

974 Pleiades satellite imagery was obtained through the ISIS program of the CNES under an
975 academic license and is not for open distribution. Numerical computations for the DEMs were
976 performed using the free and open source MicMac software suite (Rosu et al., 2014; Rupnik et
977 al., 2016) freely available at <https://micmac.engg.eu/index.php>. For cartographic mapping, we
978 also used Google Earth imagery (Landsat 7, DigitalGlobe) freely accessible at

979 <https://earth.google.com>. All geological maps used in this work are cited in the main text and in
 980 the reference section. All field measurements have been collected by us on our field missions
 981 (march 2018, january 2019) and are provided in the main text, in the figures and in the
 982 supporting information. These field measurements will be uploaded to [Zenodo.org](https://zenodo.org) upon
 983 acceptance.

984 The kinematic modeling was made using FoldFault Forward version 6, freely available from
 985 <http://www.geo.cornell.edu/geology/faculty/RWA/programs/faultfoldforward.html>
 986 (Allmendinger, 1998).

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1266 Zehnder, A. T., & Allmendinger, R. W. (2000). Velocity field for the trishear model. *Journal of*
1267 *Structural Geology*, 22, 1009–1014.

1268 **Figures**

1269 **Figure 1.** Simplified geological and structural map of the western Central Andes at ~20–22°S,
 1270 Northern Chile (modified from Armijo et al., 2015), and average topographic profile (top; ve:
 1271 vertical exaggeration). The two main structural ensembles are here the Marginal Block and the
 1272 Western Cordillera. The Marginal Block encompasses the Coastal Cordillera and the longitudinal
 1273 valley of the Atacama Bench (or Central Depression). The Western Cordillera includes the West
 1274 Andean Fold-and-Thrust-Belt (WAFTB), a basement high (Cordillera Domeyko), and the
 1275 modern volcanic arc. A large part of the WAFTB is hidden beneath blanketing Cenozoic
 1276 deposits and only outcrops in few places. The Andean Basement Thrust (ABT) separates the
 1277 WAFTB and the basement high of the Western Cordillera. The location of Figures 4 (Pinchal
 1278 area) and 9 (Quebrada Blanca area) is given by black boxes. Inset: Location of the map (red box)
 1279 within the Central Andes along the South American Continent. WAT: West Andean Thrust (after
 1280 Armijo et al., 2015); FTB: Fold-and-Thrust-Belt; Cz: Cenozoic; Mz: Mesozoic; Pz-Pc:
 1281 Paleozoic and Precambrian.

1282

1283 **Figure 2.** Landscape field overviews of the Pinchal area depicting the main tectono-stratigraphic
 1284 units. The Paleozoic (Pz) basement stands clearly out in the background, characterized by its
 1285 darker color and higher altitudes. The Mesozoic (Mz) series in the central part and in the
 1286 foreground bear a marine part and a volcano-detrital part, delimited by an outstanding calcareous
 1287 (Calc.) crest. Unconformable Cenozoic erosional surfaces, with limited fluvial deposits can also
 1288 be observed. View points of both pictures are located on Figure 4.

1289

1290 **Figure 3.** First-order stratigraphic column of the Pinchal area derived from field observations
 1291 obtained mainly along Quebrada Tania (Figures 4 and 5a) where the Mesozoic series seems to be
 1292 most complete. Thicknesses of the stratigraphic units are not at scale on the figure, but are given
 1293 in the main text (section 4.1.1). By analogy to regional descriptions, these layers are suspected to
 1294 be Triassic at the base, and Jurassic in the case of the marine fossiliferous levels (see section
 1295 4.1.3 for additional details). The description of Cenozoic units is here completed based on the
 1296 work of Victor et al. (2004). Color-code in line with maps (Figures 1, 4 and S13) and cross-
 1297 sections (Figure 5). In the Pinchal area, Paleozoic basement overthrusts folded Mesozoic series

1298 along the Pinchal Thrust, so that part of the deeper and older Mesozoic series may be missing
1299 here (as depicted by “?”). See Figures S1–S12 and corresponding captions (in supplementary
1300 material) for detailed sedimentologic descriptions.

1301

1302 **Figure 4.** Structural map of the Pinchal area (at $\sim 21^{\circ}30'S$) derived from mapping in the field and
1303 on satellite imagery (location on Figure 1). White thin lines highlight Mesozoic layers mappable
1304 on satellite images. Thick blue line depicts the calcareous crest, which is used as a marker bed
1305 (Figure 2). A–A' and B–B' sections locate the topographic profiles used for the surface cross-
1306 sections of Quebrada Tania and Quebrada Martine, respectively (Figures 5a-b). In the case of the
1307 Quebrada Tambillo cross-section, a topographic swath profile was used along C–C'. The fold
1308 axes are relatively well defined for the synclinal fold, but less well constrained for the anticlinal
1309 fold because only observable along Quebrada Tambillo. Black dots refer to the location of field
1310 photographs, and are numbered according to the figures where these pictures are reported. PT:
1311 Pinchal Thrust; Q: Quebrada.

1312

1313 **Figure 5.** Cross-sections along **(a)** the Quebrada Tania (A–A' on Figure 4), **(b)** the Quebrada
1314 Martine (B–B' on Figure 4), and **(c)** the Quebrada Tambillo (C–C' on Figure 4). Reported dip
1315 angles have been measured in the field. Faults are outlined in black, and dashed when they are
1316 only observable at a local spatial scale. Only larger faults (continuous lines) are mapped on
1317 Figure 4. Fold axes are depicted above their surface trace, based on our field observations, and
1318 their orientation illustrates the deduced orientation of the corresponding axial planes. Grey
1319 numbers with arrows point out to field pictures and indicate the associated figure. In the case of
1320 the Quebrada Tania section **(a)**, the sedimentary polarity criterion (β) indicated to the west of the
1321 section has been observed ~ 1 km further downstream than reported here. For the Quebrada
1322 Martine section **(b)**, note the stripe of continental Mesozoic rocks trapped in between two strands
1323 of the Pinchal Thrust. Sub-surface interpretation from surface observations is reported with
1324 transparent colors in the case of the Quebrada Tambillo section **(c)**. Note the different spatial
1325 scales of the three sections. PT: Pinchal Thrust.

1326

1327 **Figure 6.** Field view of the Pinchal Thrust (PT), thrusting the dark-grayish Paleozoic basement
1328 over the greenish folded Mesozoic units. Reddish rocks on the hanging wall to the east-northeast
1329 correspond to the thrust shear zone (hatched area in picture). Location on Figure 4. Non-
1330 interpreted photograph can be found in the supporting information (Figure S15).

1331

1332 **Figure 7.** Field pictures of small-scale structural features characteristic of the deformation within
1333 the Pinchal zone (Location on Figures 4 and 5). Non-interpreted photographs for (b) and (c) can
1334 be found in the supporting information (Figures S16).

1335 **(a)** Shear band with characteristic C/S-fabric (for "Cisaillement/Schistosité") indicative of top-to-
1336 the-west thrusting. Observation within the metamorphic basement in the hanging wall of the
1337 Pinchal Thrust.

1338 **(b)** Example of a small-scale fold within the marine Mesozoic units (blue line) in Quebrada
1339 Tania, within the inverted limb of the mapped syncline, nearby the fold axis. Note also the
1340 erosional surface (yellow) forming the unconformable contact between the Cenozoic deposits
1341 over the deformed Mesozoic.

1342 **(c)** Small-scale thrusts (steep red line to the right) and décollements (flat red line to the left)
1343 observed within the marine Mesozoic strata (blue) of the inverted synclinal limb along Quebrada
1344 Tania. The limestone-dominated cm–dm beds are characteristic of the lower part of the marine
1345 Mesozoic units (Figure 3).

1346 **Figure 8.** Field pictures of the two major folds within the Pinchal area (location on Figure 4).
1347 Non-interpreted photos can be found in the supporting information (Figures S17).

1348 **(a)** Panoramic view over the north-eastern part of the Pinchal area. The Paleozoic basement (red
1349 crosses) overthrusts the Mesozoic units (blue and violet horizons) along the Pinchal Thrust (red
1350 line with triangles). The topographic low locates the synclinal axis. The calcareous crest on both
1351 sides is highlighted by the thick blue lines. For better visibility, Cenozoic erosional surfaces
1352 covered by thin deposits are not highlighted.

1353 **(b)** Panoramic view along Quebrada Tambillo, in the southern part of the Pinchal area. The ~200
1354 m deep incised canyon reveals the geometry of the large western anticline affecting Mesozoic
1355 layers (violet) underneath the unconformable Cenozoic strata (yellow). The fold axis (black line)
1356 probably coincides with an approximately vertical fault, also well observable on satellite

1357 imagery. Note also the repetition of smaller folds with westward decreasing amplitude and
1358 wavelength discernable beneath the westward thickening Cenozoic growth strata to the right of
1359 the picture. The Mesozoic calcareous crest (blue) and the Paleozoic basement (red crosses) over
1360 the Pinchal Thrust (red) appear in the far eastern background.

1361

1362 **Figure 9.** Structural map of the Quebrada Blanca zone (at $\sim 20^{\circ}45'S$), refined from Armijo et al.
1363 (2015) (location on Figure 1). Colored lines report mappable layers. For visibility, only major,
1364 well-correlated layer traces are represented here. Black boxes locate the swath profiles from
1365 which layers were projected for the construction of the structural east–west cross-section (Figure
1366 10). The A–B section corresponds to the topographic profile used for this same cross-section.
1367 Strike and dip measurements are extracted from 3D-mapping (see section 3.3) or observed in the
1368 field. Strike symbols without dip magnitude are derived from satellite imagery. Thick black lines
1369 correspond to major fold axes. Field pictures are located (with view direction), and numbered
1370 according to the associated figure. Ages from uranium-lead (U/Pb) radioisotope dating on zircon
1371 are taken from the Guatacondo geological map (Blanco & Tomlinson, 2013). Letters C, D, E, F,
1372 G and H to the north-east (within the folded Chacarilla and Majala Formations) report the layers
1373 illustrated on Figure 12. Cz: Cenozoic; K: Cretaceous; Jr: Jurassic; Q: Quebrada.

1374

1375 **Figure 10.** East–west cross-section of the Quebrada Blanca site, established from the projection
1376 of selected, well-expressed layers mapped on satellite imagery. APF: Altos de Pica Formation.

1377 **(a)** Observations, reporting the geometry of projected layers and associated dip angles, together
1378 with their stratigraphic ages (color-code).

1379 **(b)** Sub-surface interpretation and extrapolation of observations.

1380 **(c)** East–west cross-section based on (a) and (b). Interpretation at depth is indicated with
1381 transparent colors, in contrast with sub-surface observations. Extrapolation above the
1382 topographic surface is drawn with dashed lines. Ages from uranium-lead (U/Pb) radioisotope
1383 dating on zircon are taken from the Guatacondo geological map (Blanco & Tomlinson, 2012).
1384 The ~ 27 – 29 Ma age of the basal deposits of the Altos de Pica formation is derived from regional
1385 considerations (Victor et al., 2004).

1386

1387 **Figure 11.** Field picture of the western limb of the western anticline in the Quebrada Blanca
1388 area. Non-interpreted photographs are provided in supplementary material (Figure S18).
1389 Location on Figures 9 and 10.

1390 **(a)** Series of folds with westward decreasing amplitude and wavelength (hundreds to tens of
1391 meters) observed at the front of the western anticline.

1392 **(b)** Detailed view of the westernmost outcropping small-scale anticlines, located on (a) by the
1393 black box.

1394 **Figure 12.** Landscape view on the western limb of the eastern large-scale anticline in the
1395 Quebrada Blanca area (Location on Figures 9 and 10). Here, steeply inclined Mesozoic horizons
1396 are very well discernible in the landscape. Bedding traces C, D, E, F, G and H underlined here
1397 are also georeferenced on the structural map (Figure 9) from mapping on satellite imagery. Note
1398 the thrust-affected small-scale fold (red dashed line) emphasizing the west-vergence of tectonic
1399 structures. The non-interpreted picture is provided in supplementary material (Figure S19).

1400 **Figure 13.** Kinematics of folding of the western anticlines of the Quebrada Blanca and Pinchal
1401 zones as deduced from field observations and trishear modeling. Modeling was performed with
1402 FaultFold Forward v.6 (Allmendinger, 1998).

1403 **(a-c)** Final stages of the best-fit models in the case of **(a)** the Quebrada Blanca area; **(b)** the
1404 Quebrada Tambillo (Pinchal area), shown here at the same scale as (a). **(c)** Detailed and enlarged
1405 view of our results for the Quebrada Tambillo (Pinchal area). Note the large scale-difference
1406 between the sections of the two investigated sites (a,b). Thicker lines outline model results, while
1407 transparent lines and colors refer to the cross-sections of Figures 10c and 5c. These lines are
1408 color-coded according to the stratigraphic level they represent, as in the original cross-sections.
1409 Black lines report the modeled thrusts and horizontal arrows report the model total shortening.
1410 PT: Pinchal Thrust.

1411 **(d)** Shortening vs. time, as deduced from trishear modeling of the western anticlines of the
1412 Quebrada Blanca and Pinchal areas, and the ages of deformed layers. The three temporal
1413 benchmarks correspond to the age of the youngest folded Cretaceous (Kr) unit (~68 Ma), to the
1414 age of magmatic intrusions (~44 Ma) that are cartographically discordant, both derived from the
1415 Guatacondo geological map (Blanco & Tomlinson, 2013 – see also Figures 9 and 10c), and to
1416 the ~29 Ma age of the oldest Cenozoic layer of the Altos de Pica Formation (APF) (Victor et al.,
1417 2004) above the Choja erosional surface. It is possible that most deformation occurred prior to

1418 ~44 Ma, as deduced from the age of the intrusions cartographically seemingly post-dating
1419 folding (Figure 9), even though this argument is to be taken with caution. Our results underline
1420 two phases of deformation, with a slowing down of deformation since ~29 Ma at least, possibly
1421 even before. Intermediate stages of the trishear modeling are reported on Figures S20 and
1422 S21 (supplementary material) for the cross-sections of Quebrada Tambillo and Quebrada
1423 Blanca, respectively. Model parameters are indicated in Tables S1–S3 in supplementary material.

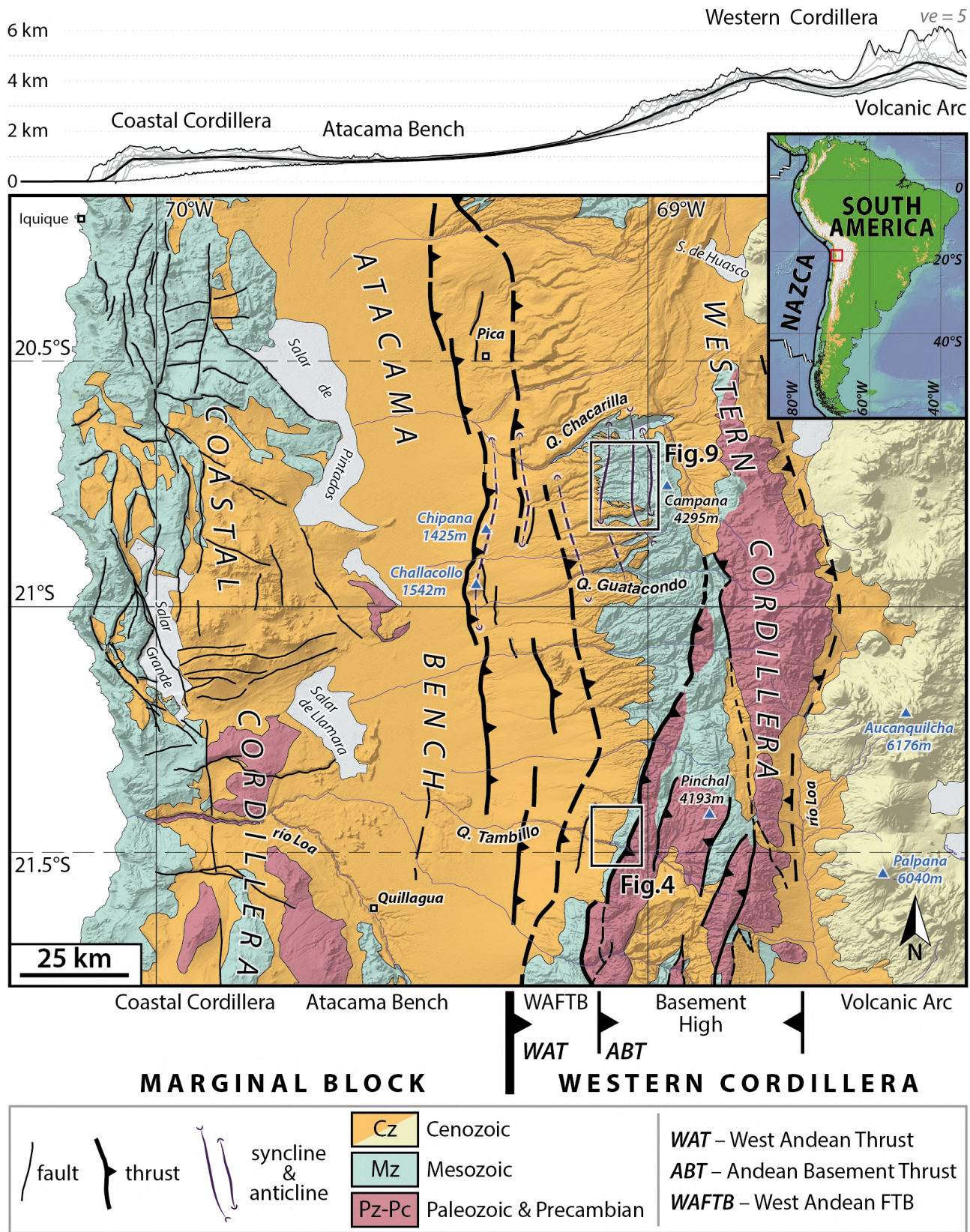


Figure 1.

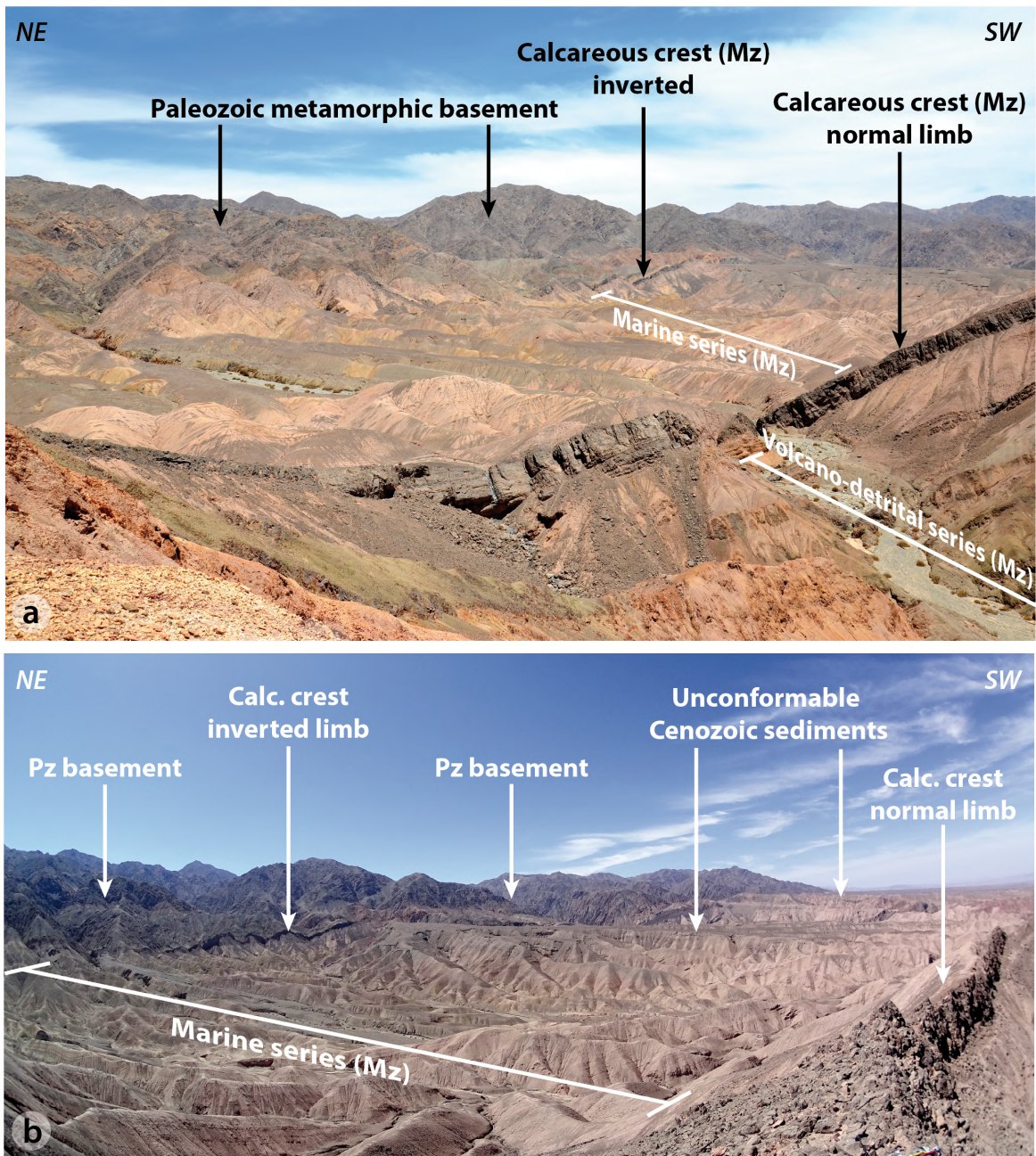


Figure 2.

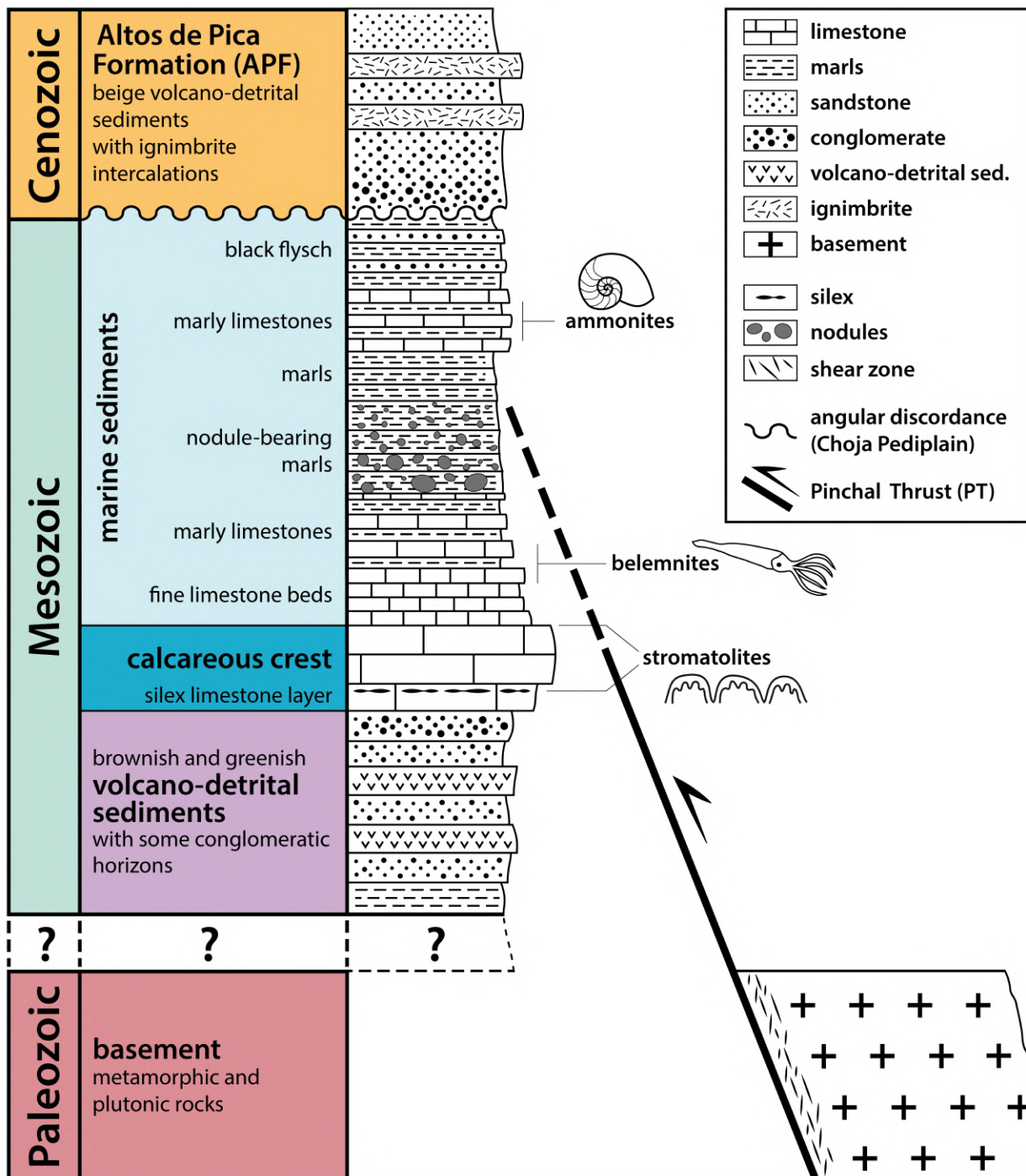


Figure 3.

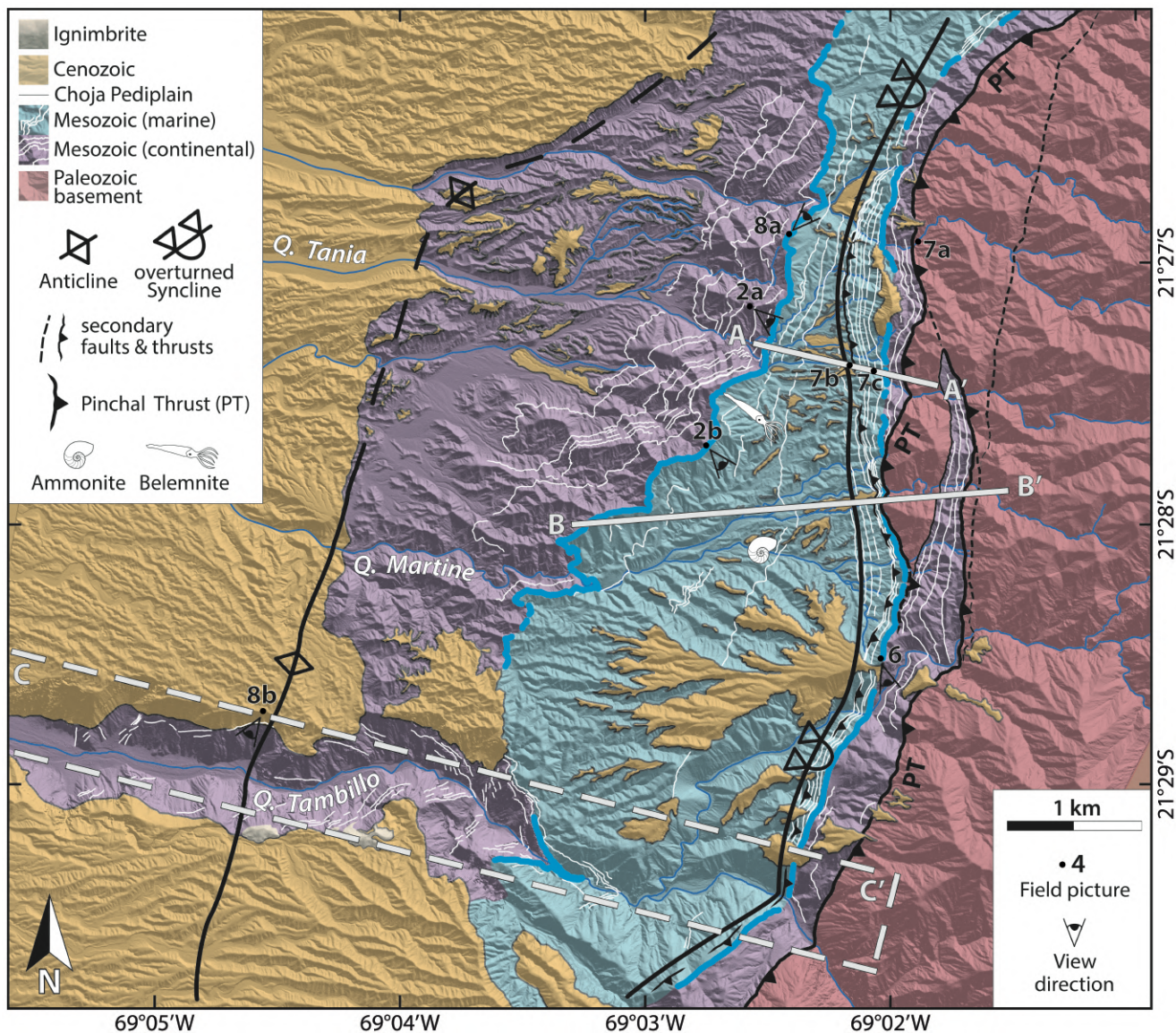


Figure 4.

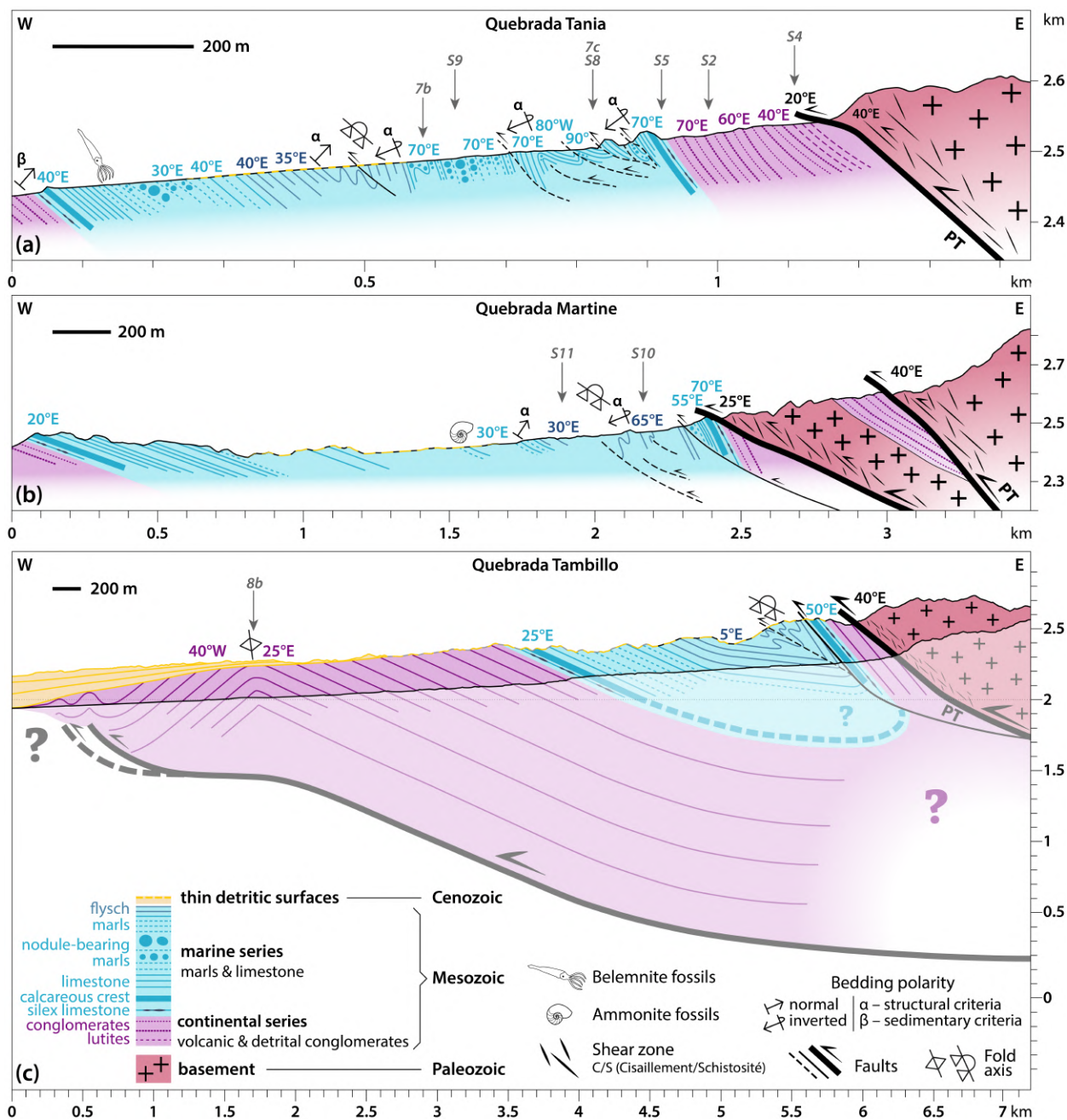


Figure 5.

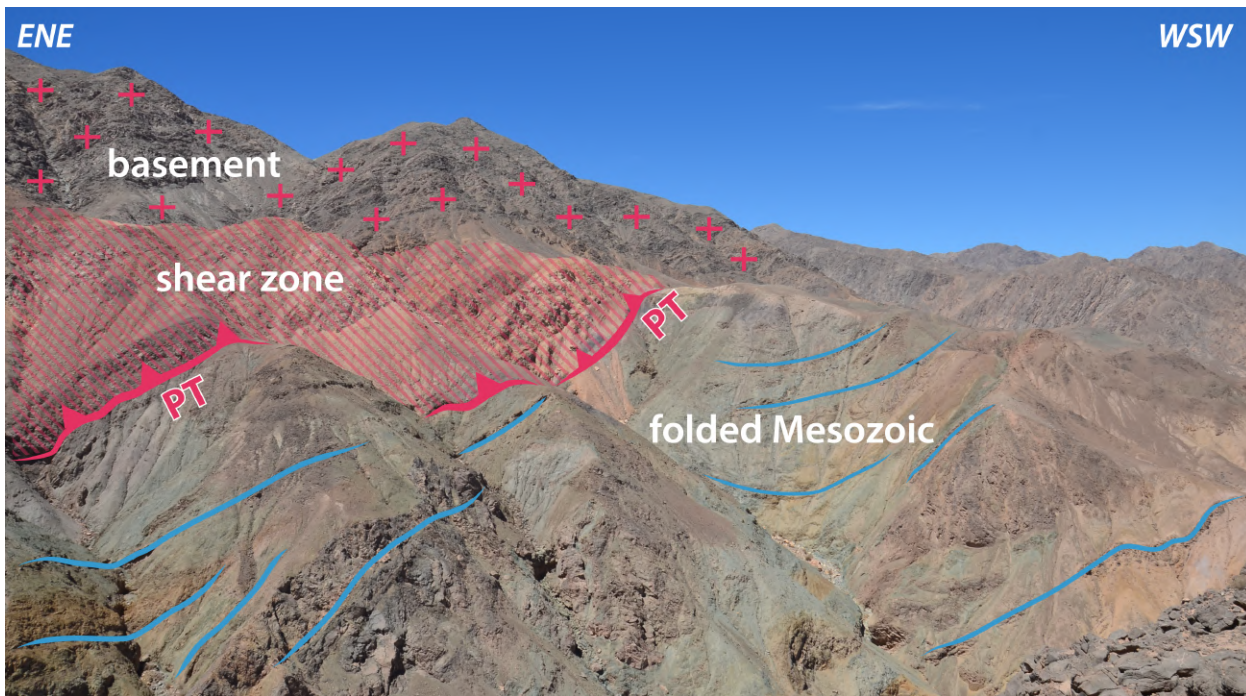


Figure 6.

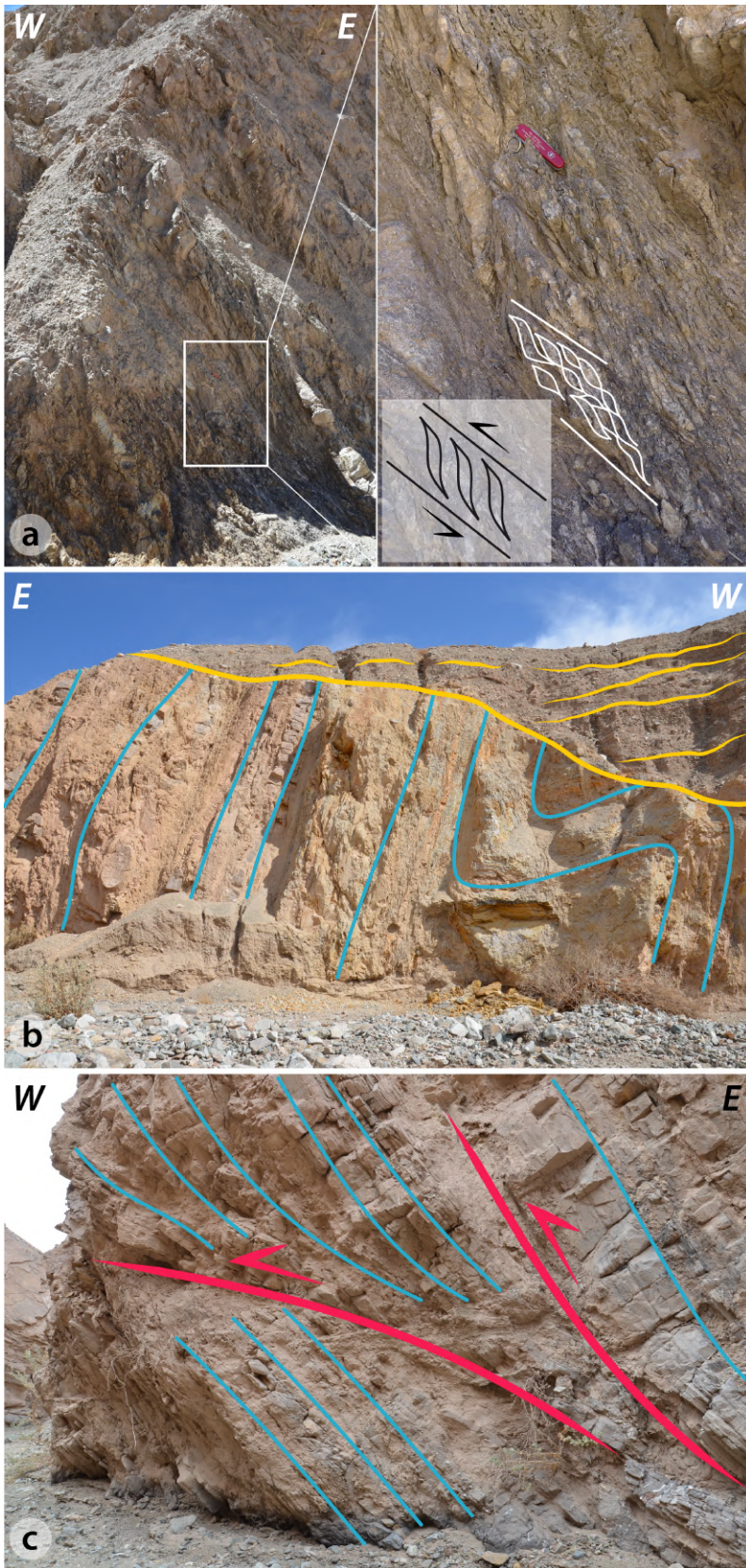


Figure 7.

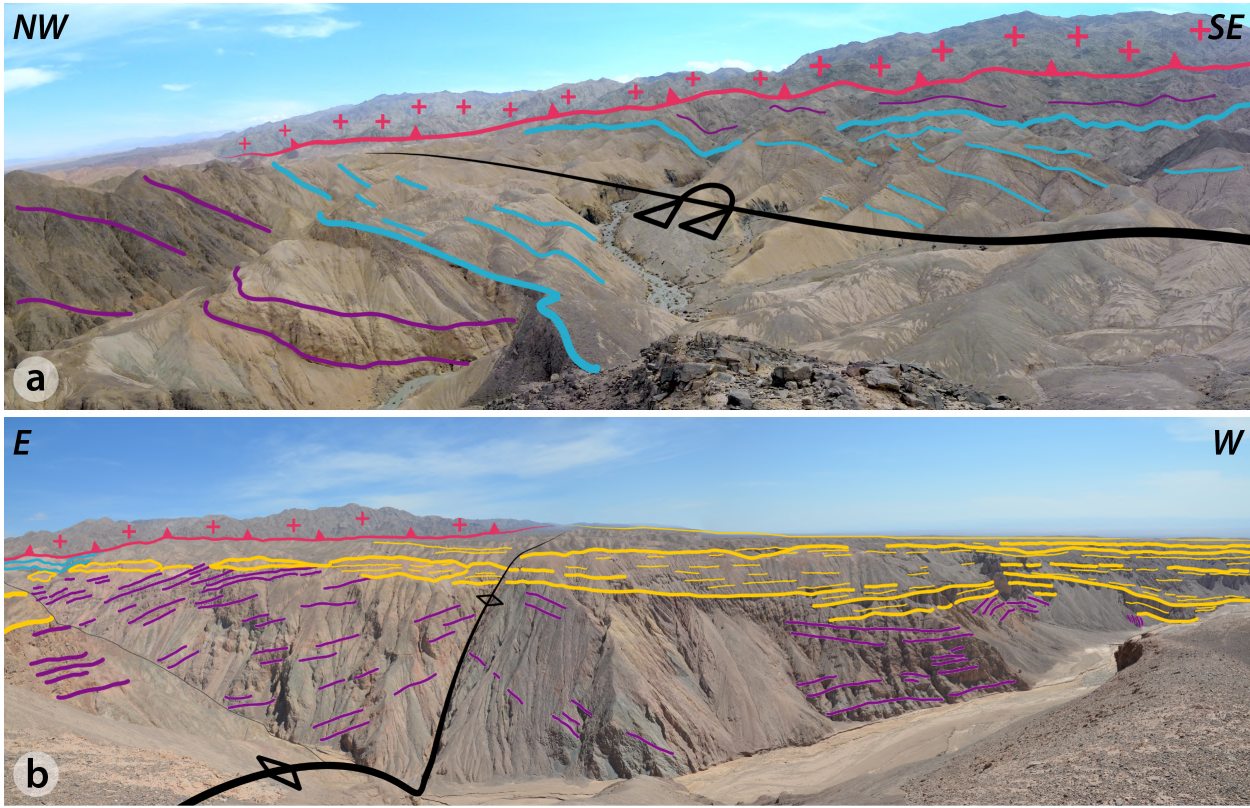


Figure 8.

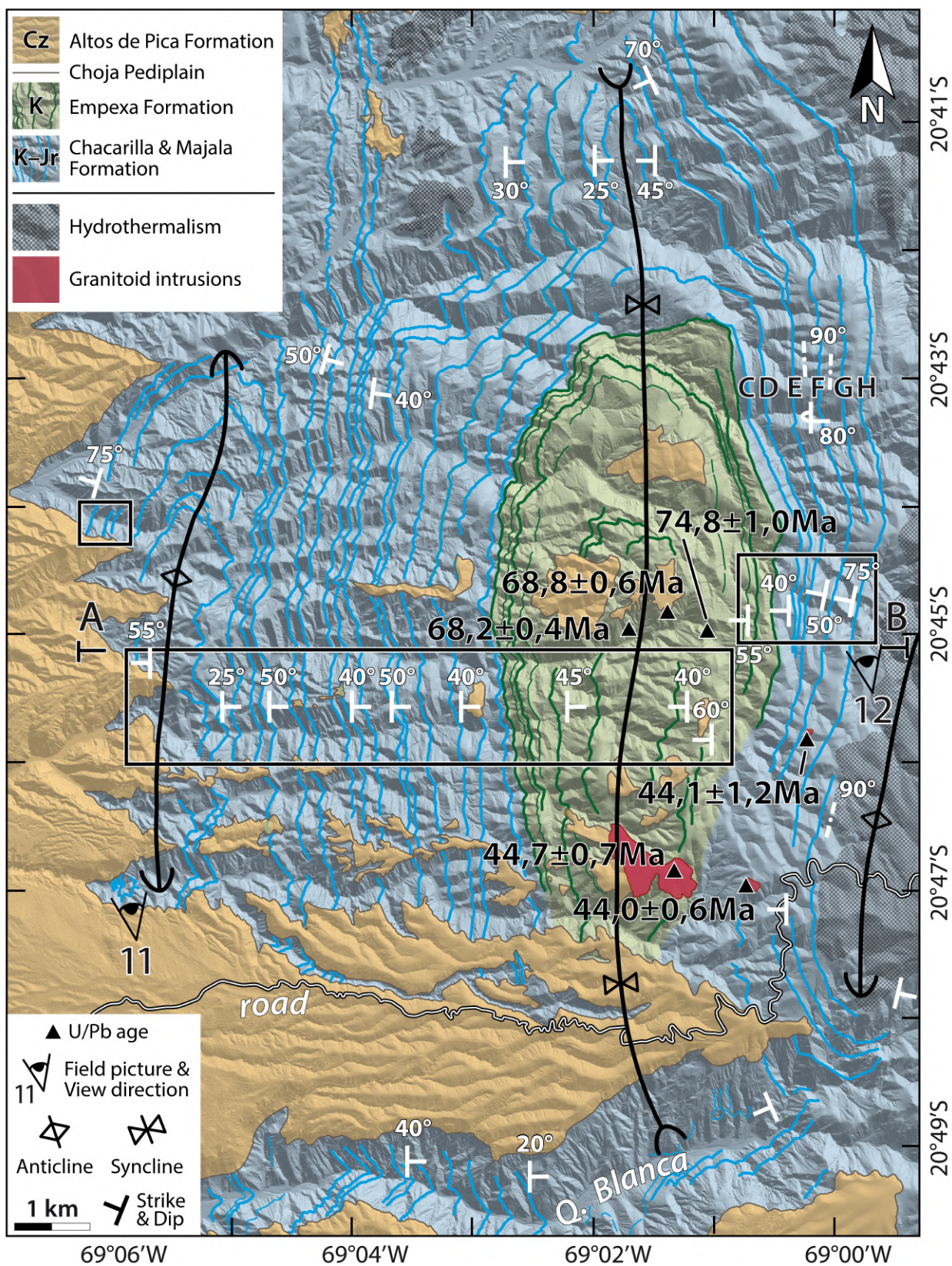


Figure 9.

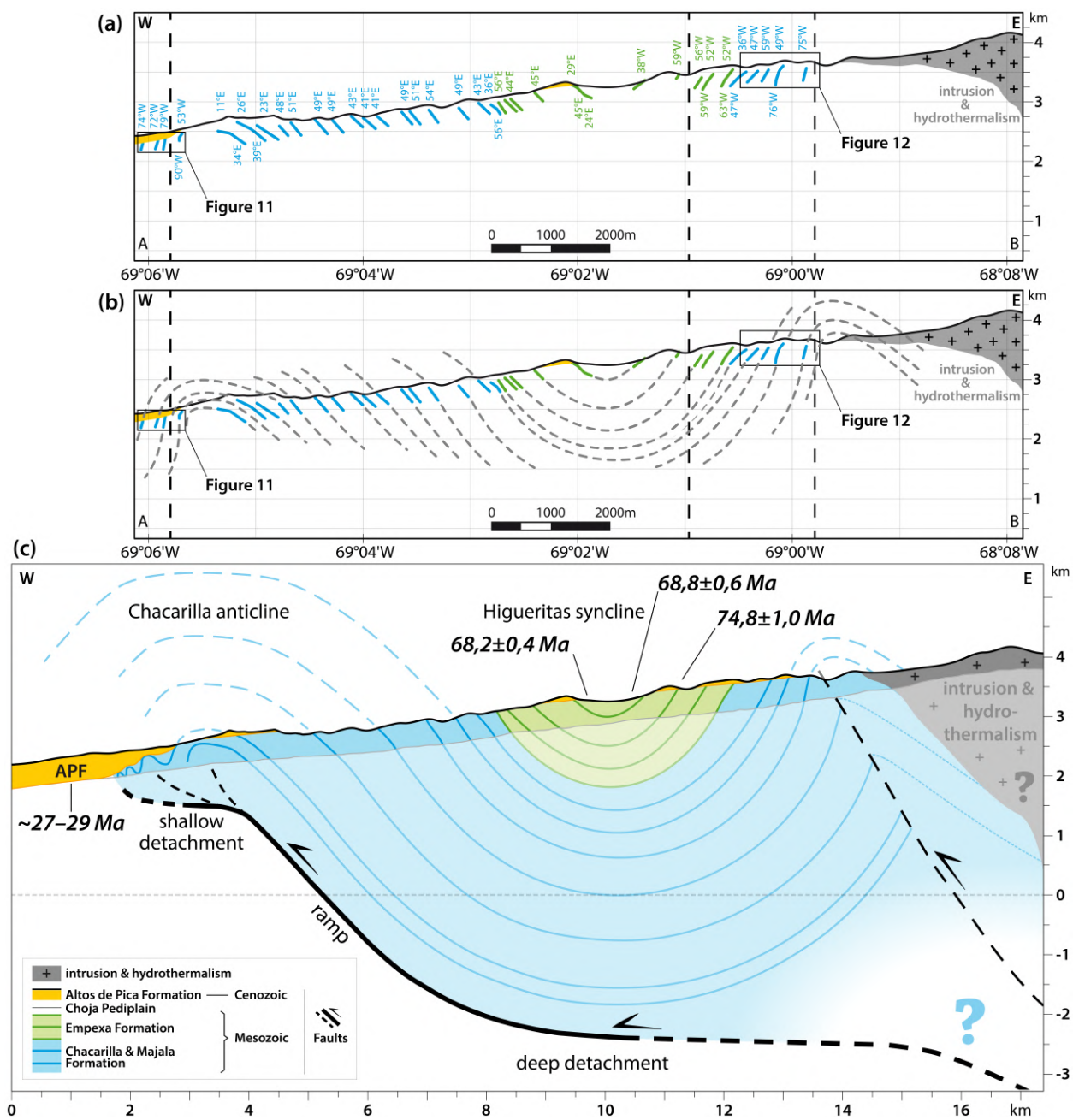


Figure 10.

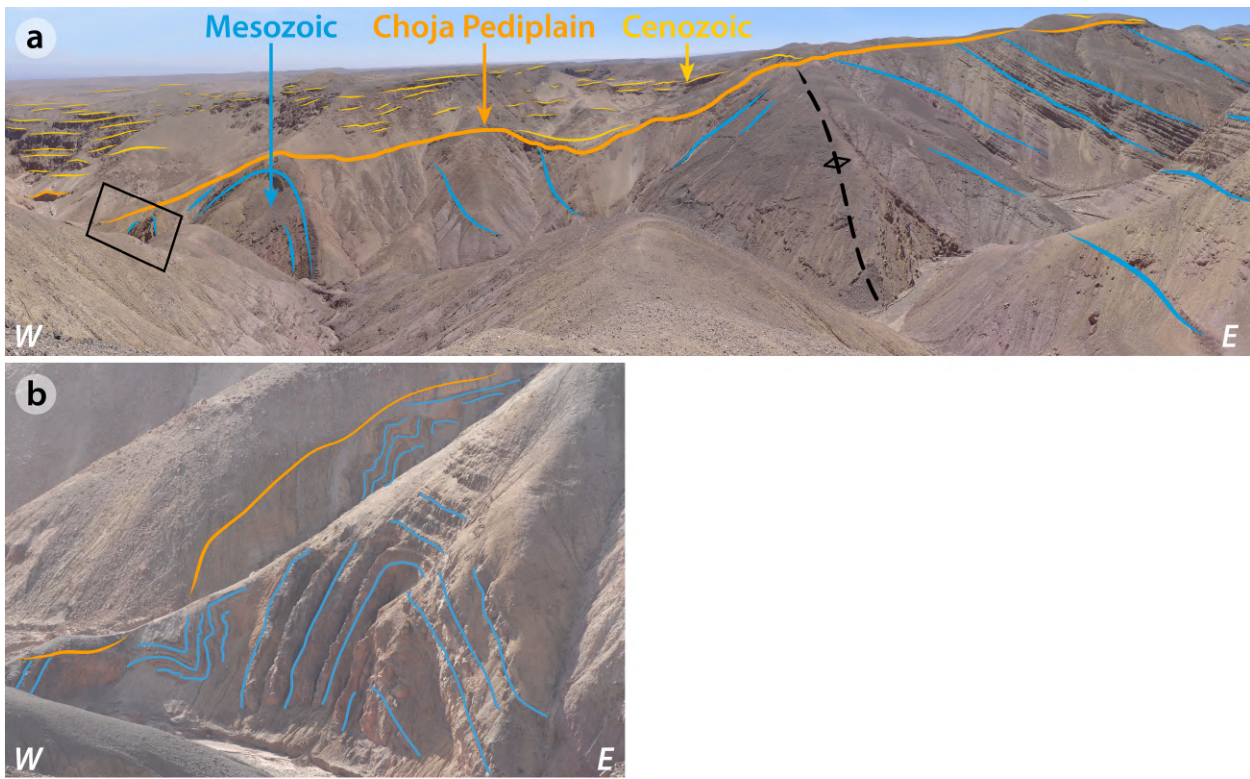


Figure 11.



Figure 12.

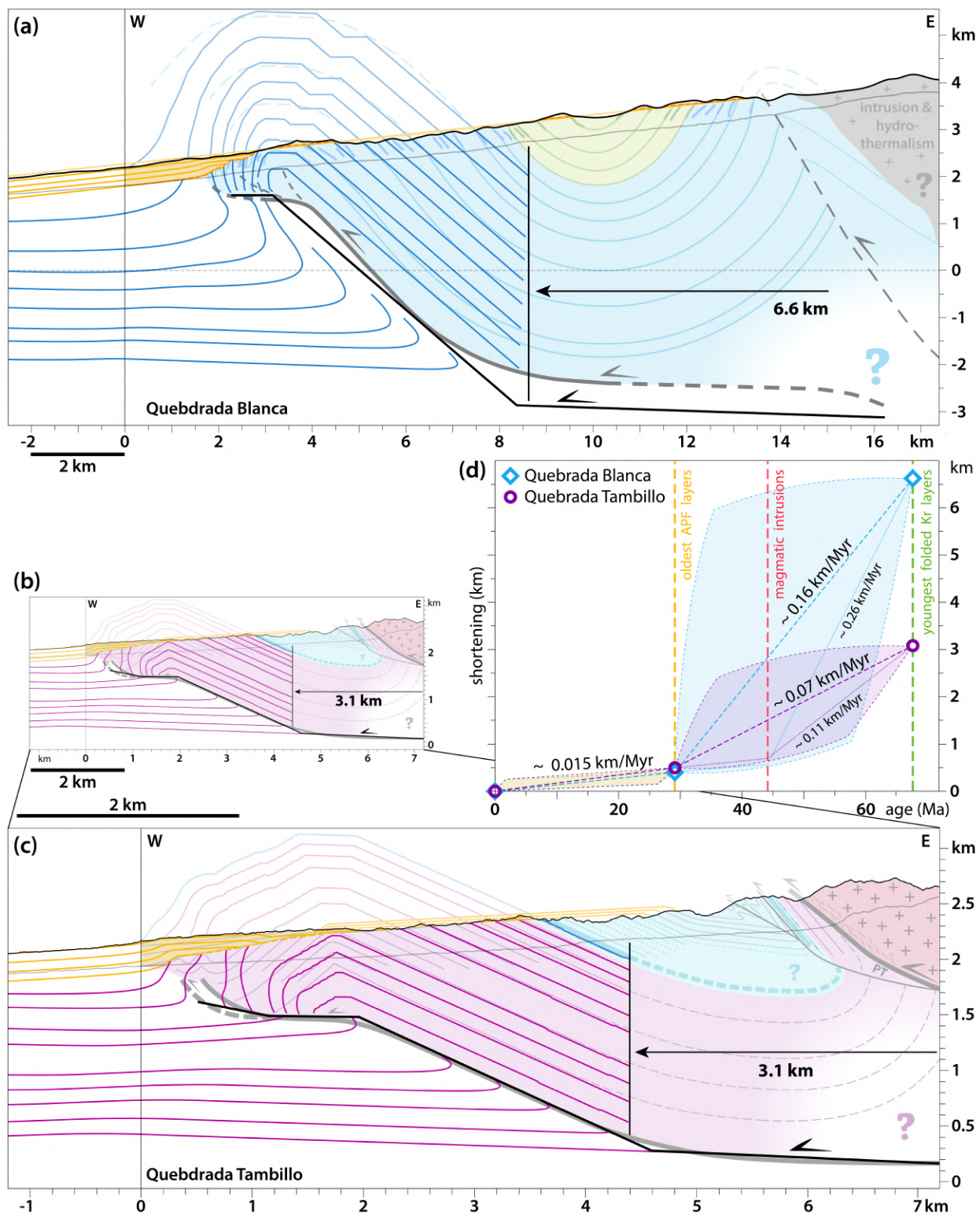


Figure 13.

The western Andes at ~20–22°S: A contribution to the quantification of crustal shortening and kinematics of deformation.

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Text S1

Figures S1 to S21

Tables S1 to S3

Additional Supporting Information (Files uploaded separately)

Caption for Dataset S1

Introduction

This supporting information document is subdivided into three sections:

(1) A sedimentary description of the units encountered at the Pinchal zone, and further illustrating section 4.1.1. Field photographs are provided (Figures S1–S12) and their position is pointed out on the corresponding structural map (Figure S13) and within the stratigraphic column (Figure S14).

(2) Non-interpreted and interpreted field photographs (Figures S15–S19), to be compared to their interpreted version in the main text (Figures 6, 7b-c, 8a-b, 11a-b, 12).

(3) Additional information on trishear modeling. Here, we present further details on the trishear modeling method and on our results (Text S1), complementary to section 3.4 and 6.2 in the main text. This text is accompanied by figures illustrating six key stages of our best-fitting trishear model for Quebrada Tambillo (Figure S20) and Quebrada Blanca (Figure S21); by a table showing the range of tested parameters during modeling (Table S1); as well as tables with the parameters for each of the six stages of our preferred model for Quebrada Tambillo (Table S2) and Quebrada Blanca (Table S3).

All the photographs (presented in the supporting information and in the main text) were taken by us during our field missions in March 2018 and January 2019 in the Western Cordillera of Northern Chile.

In addition to this document, we provide our georeferenced field-logistics dataset (field-logistics.kmz). Therewith one can visualize (e.g. on Google Earth) the off-road track we used to reach the extremely remote Pinchal area, the localisation of our Pinchal base-camp, and the GPS positions of field photographs for both study sites. In hyper-arid environments such as here, tracks may be preserved for several years, in between two rare rain episodes. We therefore recall that the off-road track and base-camp indicated here were those of March 2018 and January 2019, and cannot guarantee their state and usability after that period.

(1) Sedimentary description (Pinchal area)



Figure S1. Migmatitic gneisses, a common metamorphic facies found in the Paleozoic basement rocks to the east of the Pinchal area. Hiking stick given for scale. Location #S1 on Figures S13 and S14.

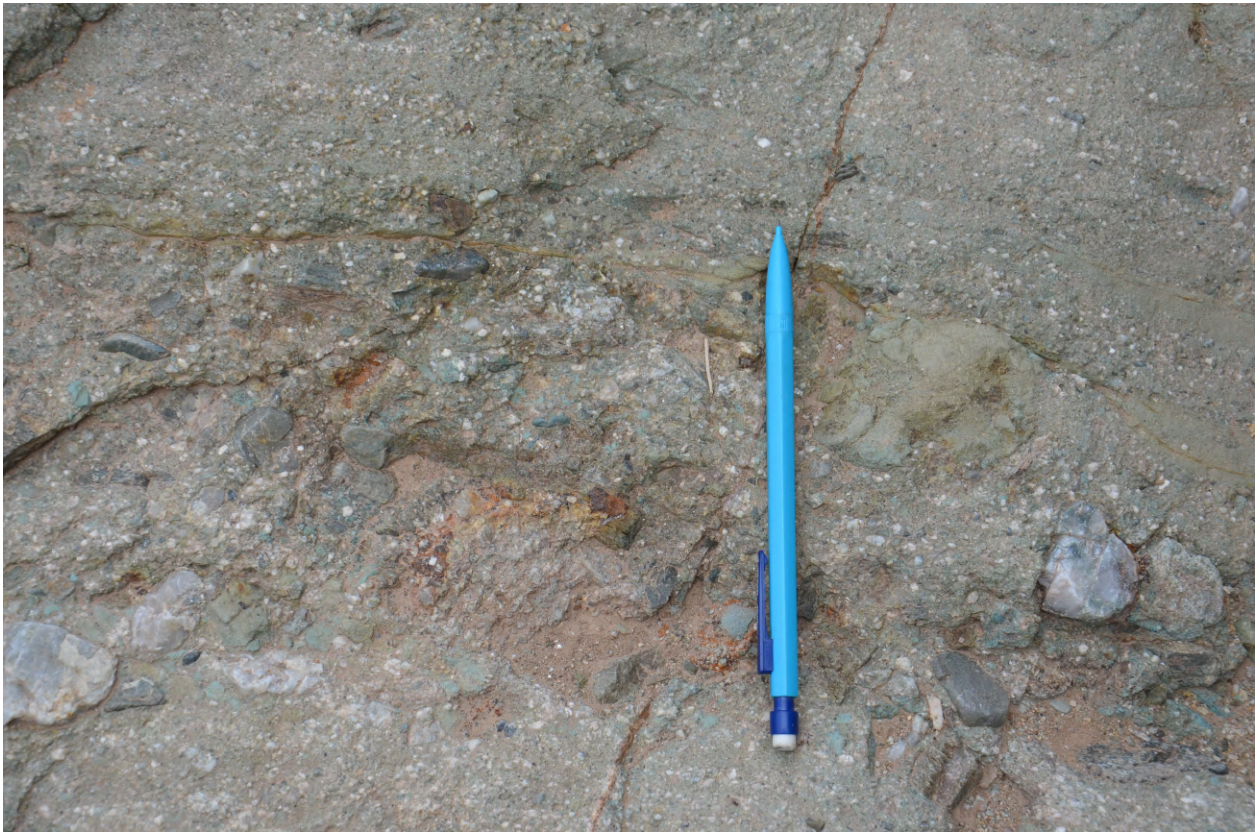


Figure S2. Volcano-detrital conglomerates, here of greenish color, belonging to the continental Mesozoic series, with millimetric to centimetric clasts. Location #S2 on Figures 5a, S13 and S14.

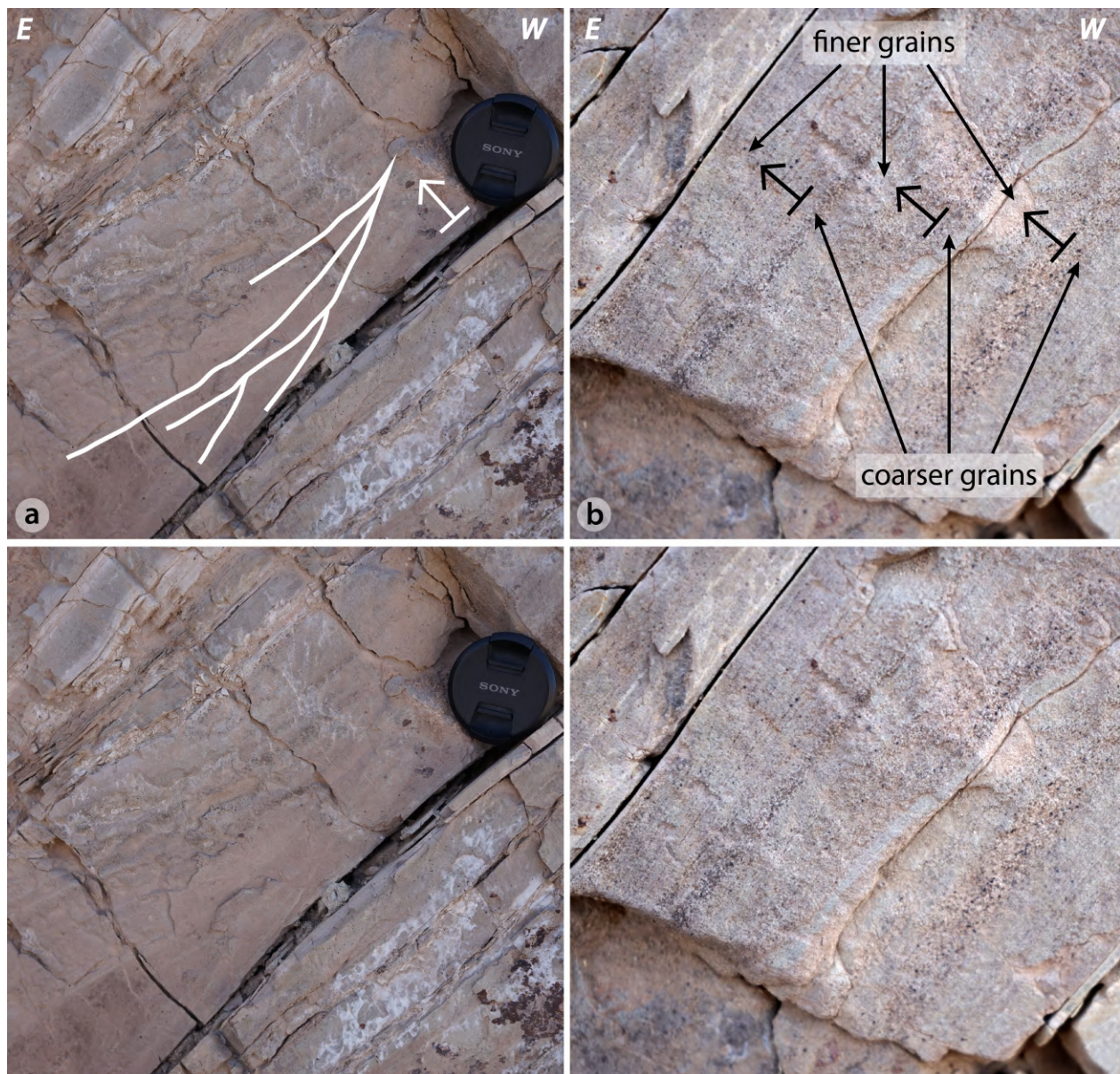


Figure S3. Detrital layers within the Mesozoic units, characterized by **(a)** tangential beds, indicative of normal polarity with top-to-the-east; **(b)** grain-size grading (finer at the top, coarser at the base of the layer) indicating top-to-the-east bedding polarity. Note also the erosive base contrasting with the sharp top of the layer. Top: interpreted field pictures; bottom: non-interpreted pictures. Location #S3 on Figures S13 and S14.



Figure S4. Dark green detrital pelites (lutites). This unit resembles sediments given a Triassic age (Aguilef et al., 2019) immediately north of the Pinchal area. Note the strong deformation as these lutites are located next to the Pinchal Thrust. Location #S4 on Figures 5a, S13 and S14.



Figure S5. Silex nodules at the base of the calcareous crest (blue pencil for scale). Location #S5 on Figures 5a, S13 and S14.

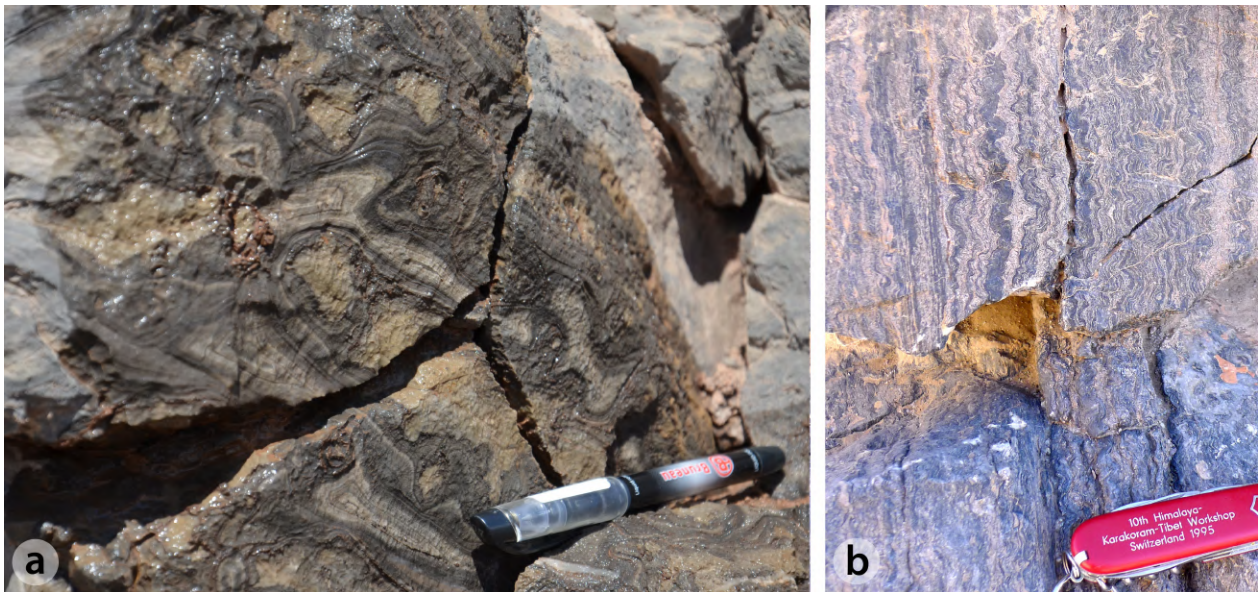


Figure S6. Stromatolite fossils from the Mesozoic (marine) series, located within (a) the western normal synclinal fold limb; and (b) the eastern inverted synclinal fold limb. Location #S6a and S6b, respectively, on Figures S13 and S14.



Figure S7. Bivalve fossils from the Mesozoic (marine) series, located within the western normal synclinal fold limb. Location #S7 on Figures S13 and S14.



Figure S8. Fine limestone layers within the Mesozoic (marine) series, characterized by a rose-beige color and an alternance of thin-bedded (cm–dm), regular beds. Location #S8 on Figures 5a, S13 and S14.



Figure S9. Nodule bearing marls from the Mesozoic (marine) series. The nodules vary in size from centimeters to few meters, as in the case of the large ones illustrated in this field photo. Location #S9 on Figures 5a, S13 and S14.

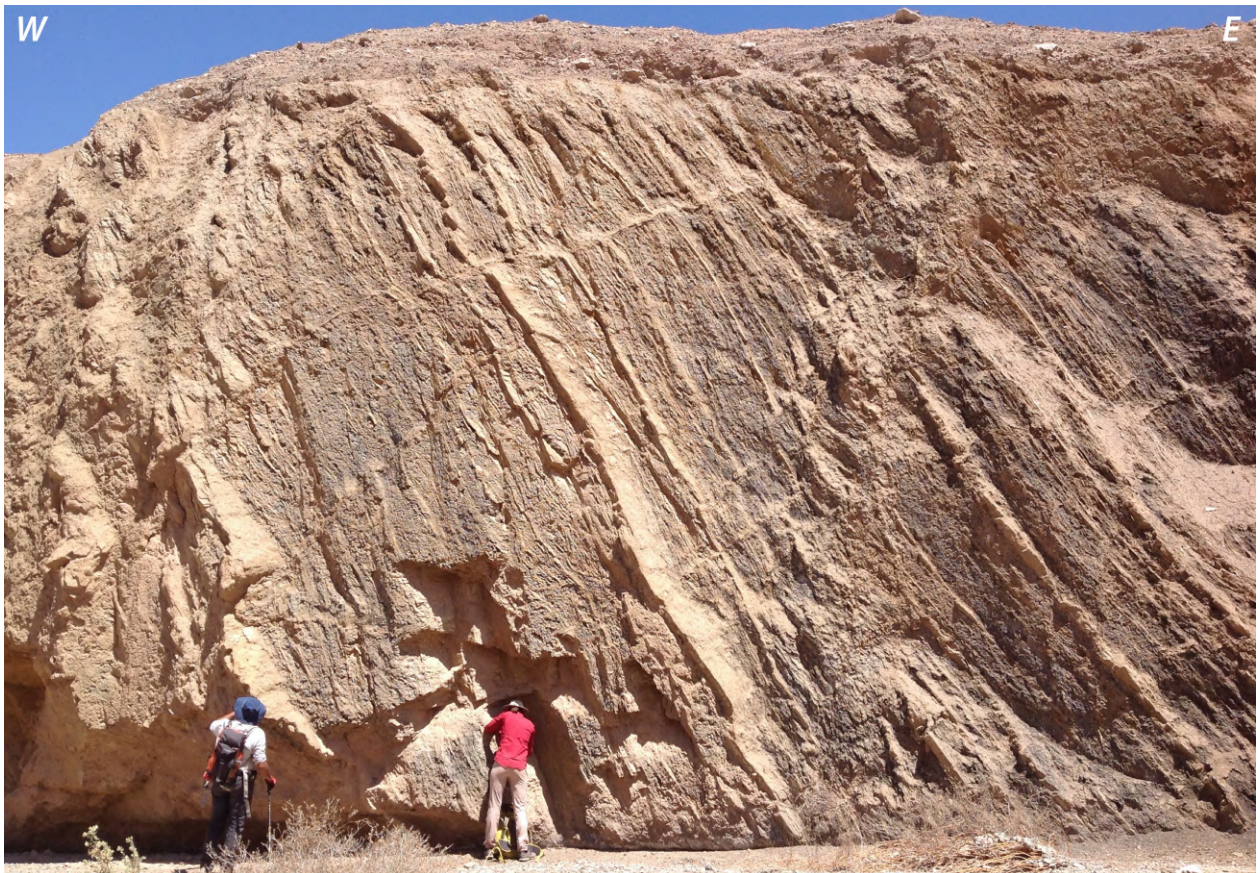


Figure S10. Thin-layered limestone series within marls. Location #S10 on Figures 5b, S13 and S14.

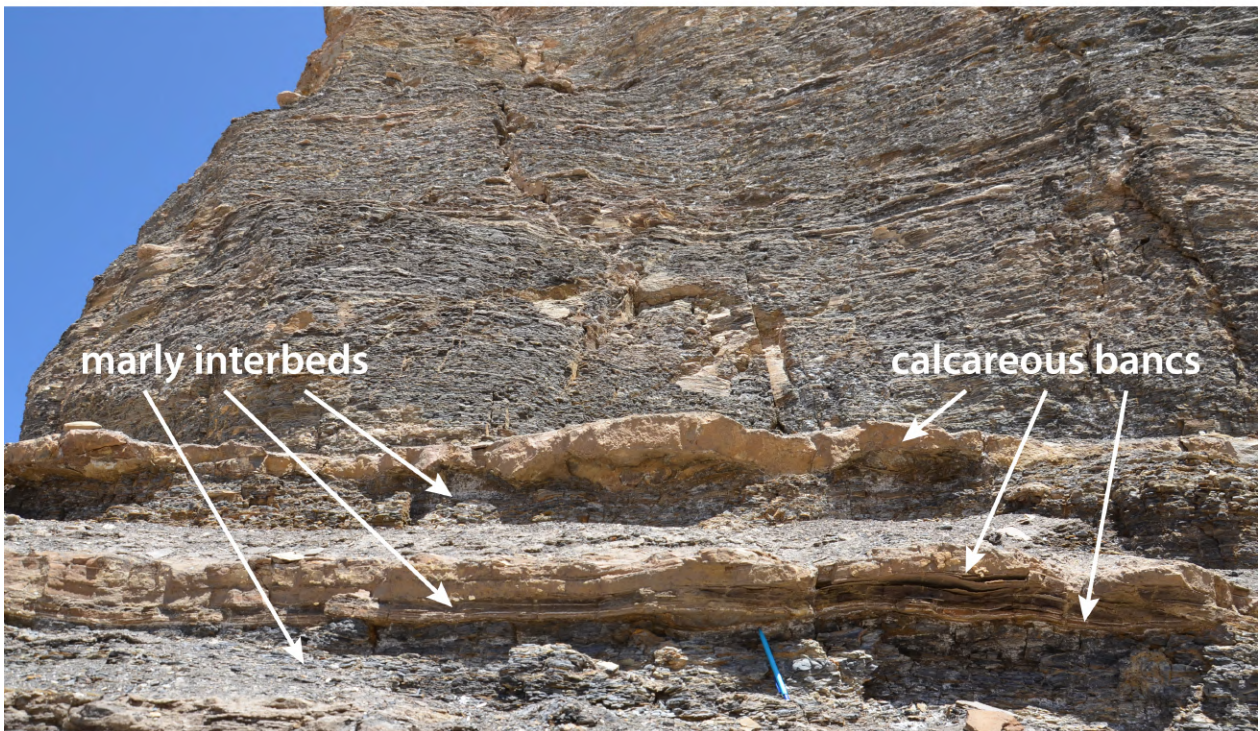


Figure S11. Flysch formation characterized by beige, resistant calcareous beds of millimetric to decimetric thickness, within dark-grey, more friable, marls. Top: landscape view; bottom: detailed outcrop view (blue pencil for scale). Location #S11 on Figures 5b, S13 and S14.



Figure S12. Red arenites at the base of the Cenozoic series bearing detrital and volcanic clasts of millimetric to pluri-centimetric size. Location #S12 on Figures S13 and S14.

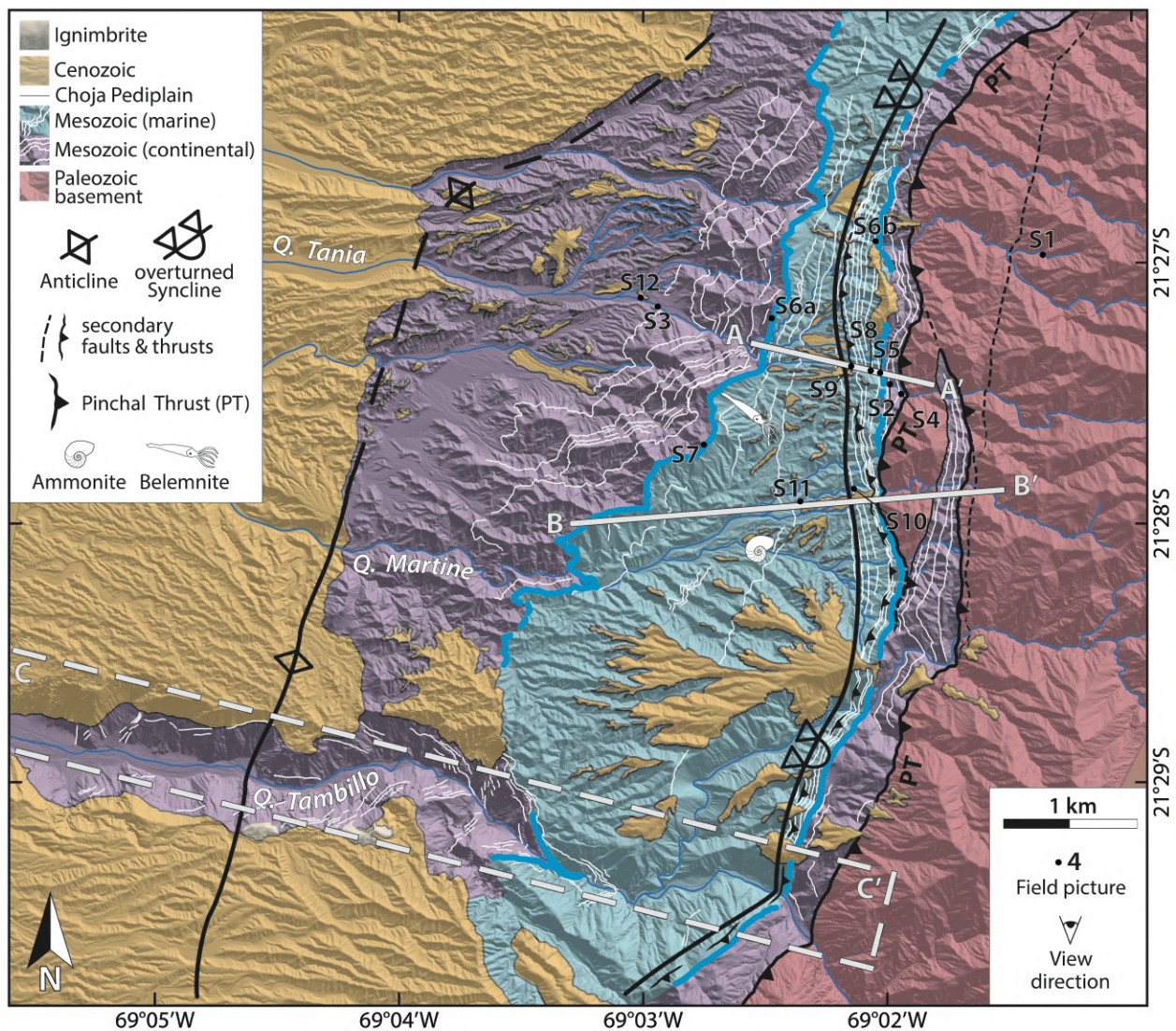


Figure S13. Structural map of the Pinchal area (~21°30'S), as Figure 4 in main text, but here with locations of supplementary field references (S1–S12). Black dots refer to the location of field photographs, and are numbered according to the figures where these pictures are reported. Location of this map is the same as that of Figure 4, reported on Figure 1. White thin lines highlight Mesozoic layers mappable on satellite images. Thick blue line depicts the calcareous crest, which is used as a marker bed (Figure 2). A–A' and B–B' sections locate the topographic profiles used for the 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. PT: Pinchal Thrust; Q: Quebrada.

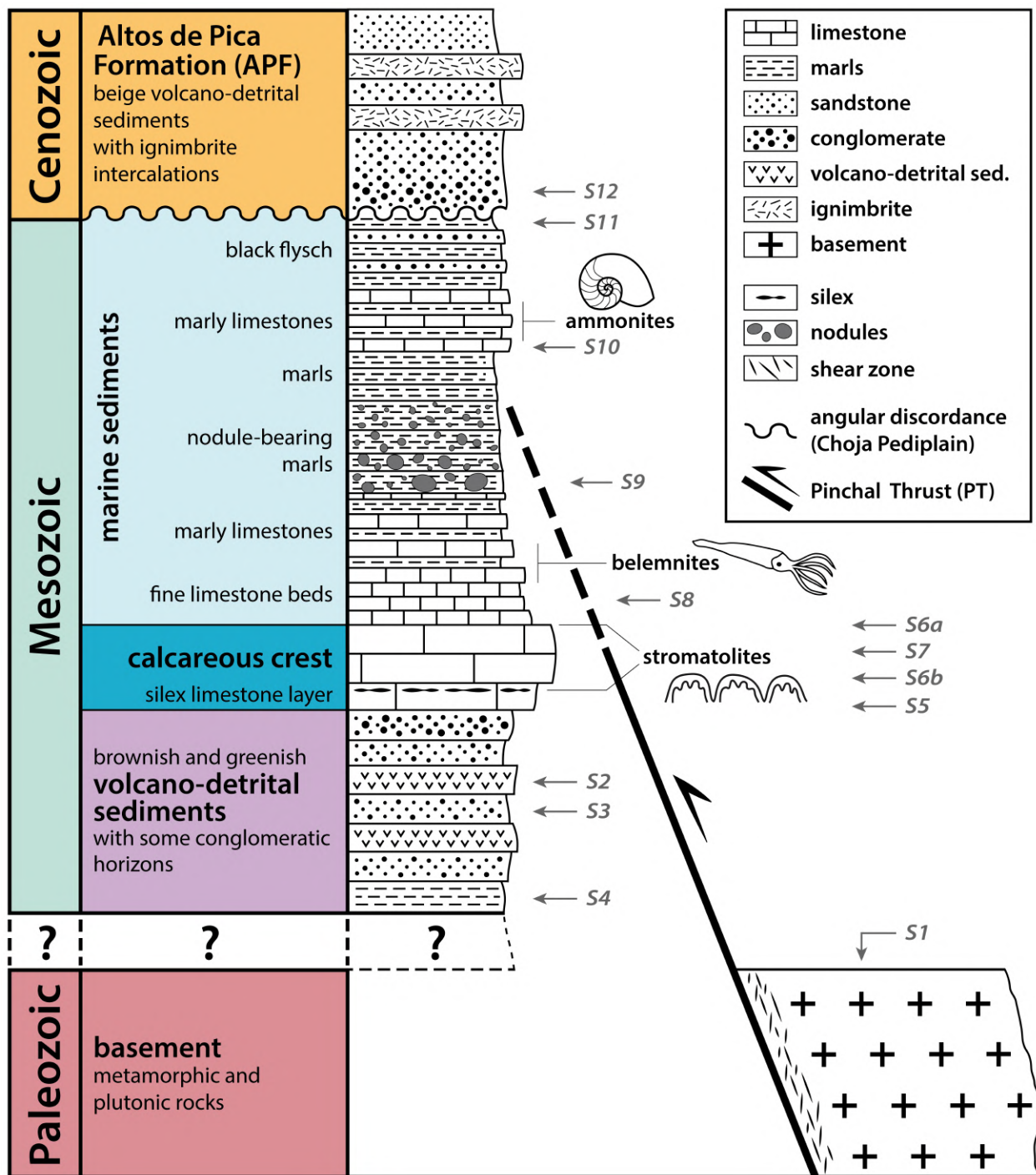


Figure 14. First-order stratigraphic column of the Pinchal area derived from field observations obtained mainly along Quebrada Tania (Figures 4 and S13) where the Mesozoic series seems to be most complete. Thicknesses of the stratigraphic units are not at scale on the figure, but are given in the main text (section 4.1.1). This column is also reported on Figure 3 (main text), but here with the addition of the stratigraphic locations of field observations illustrated on Figures S1 to S12. In the Pinchal area, Paleozoic basement overthrusts folded Mesozoic series along the Pinchal Thrust (PT), 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.

(2) Non-interpreted field photographs

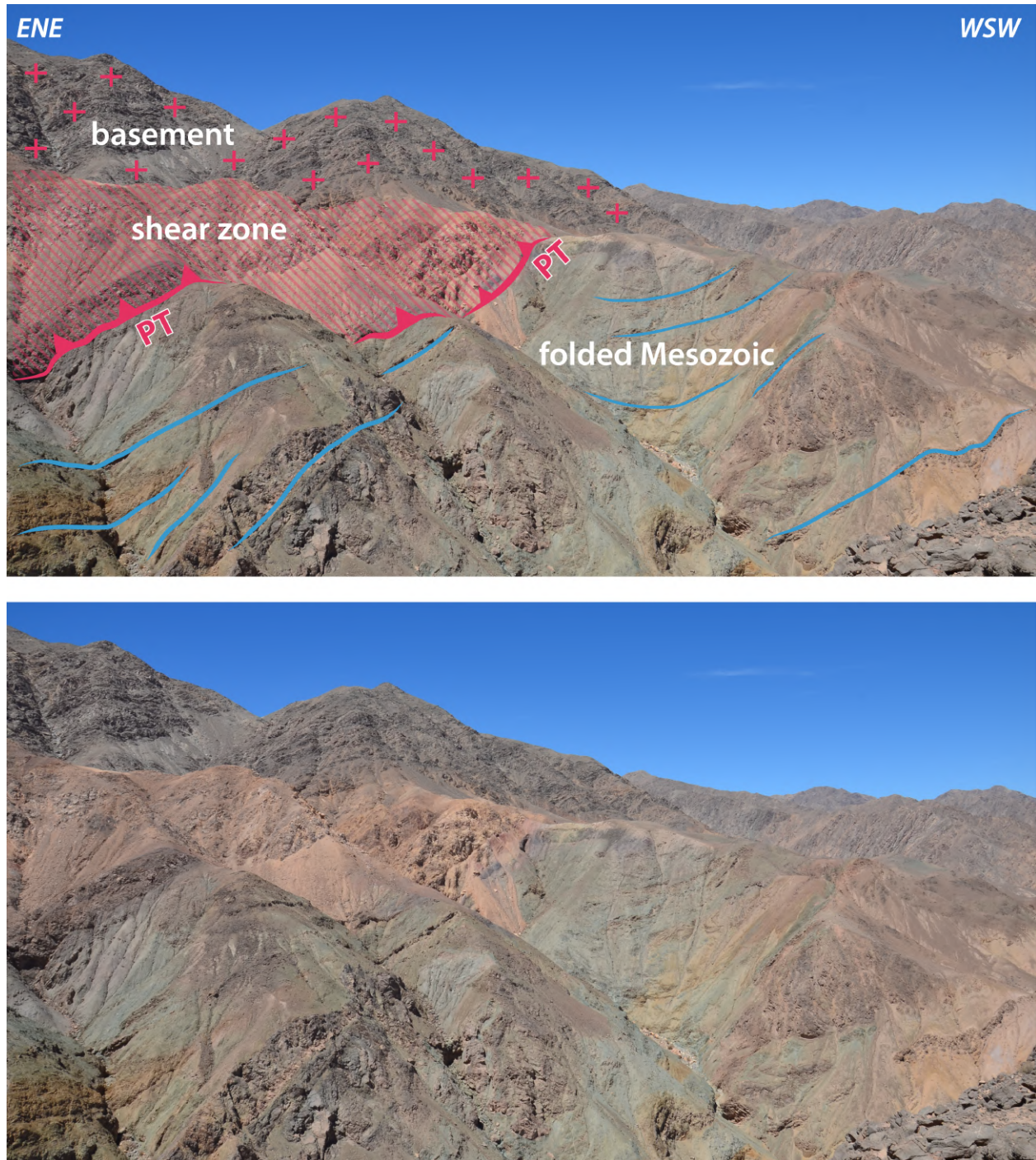


Figure 15. Field view of the Pinchal Thrust (PT), overthrusting the dark-grayish Paleozoic basement over the greenish folded Mesozoic units. Reddish rocks on the hanging wall to the East-Northeast correspond to the thrust shear zone (hatched area in picture). Same as Figure 6 (Top), but with non-interpreted field picture (bottom). Location #6 on Figure 4.

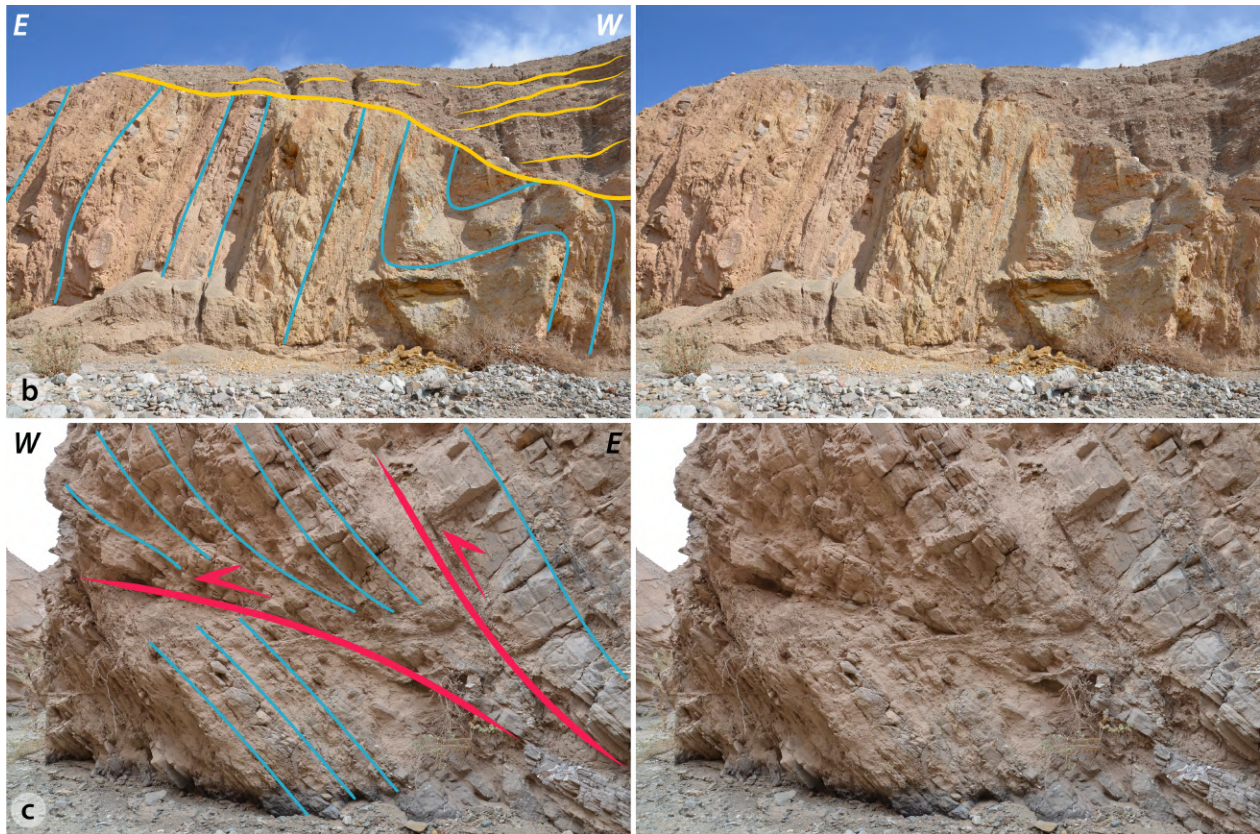


Figure S16. Field pictures of small-scale structural features characteristic of the deformation within the Pinchal zone, as in Figures 7b-c in main text. Locations #7b and 7c in Figure 4, respectively. Left: interpreted picture as in main text; right: non-interpreted picture.

(b) Example of a small-scale fold within the marine Mesozoic units (blue lines) 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.

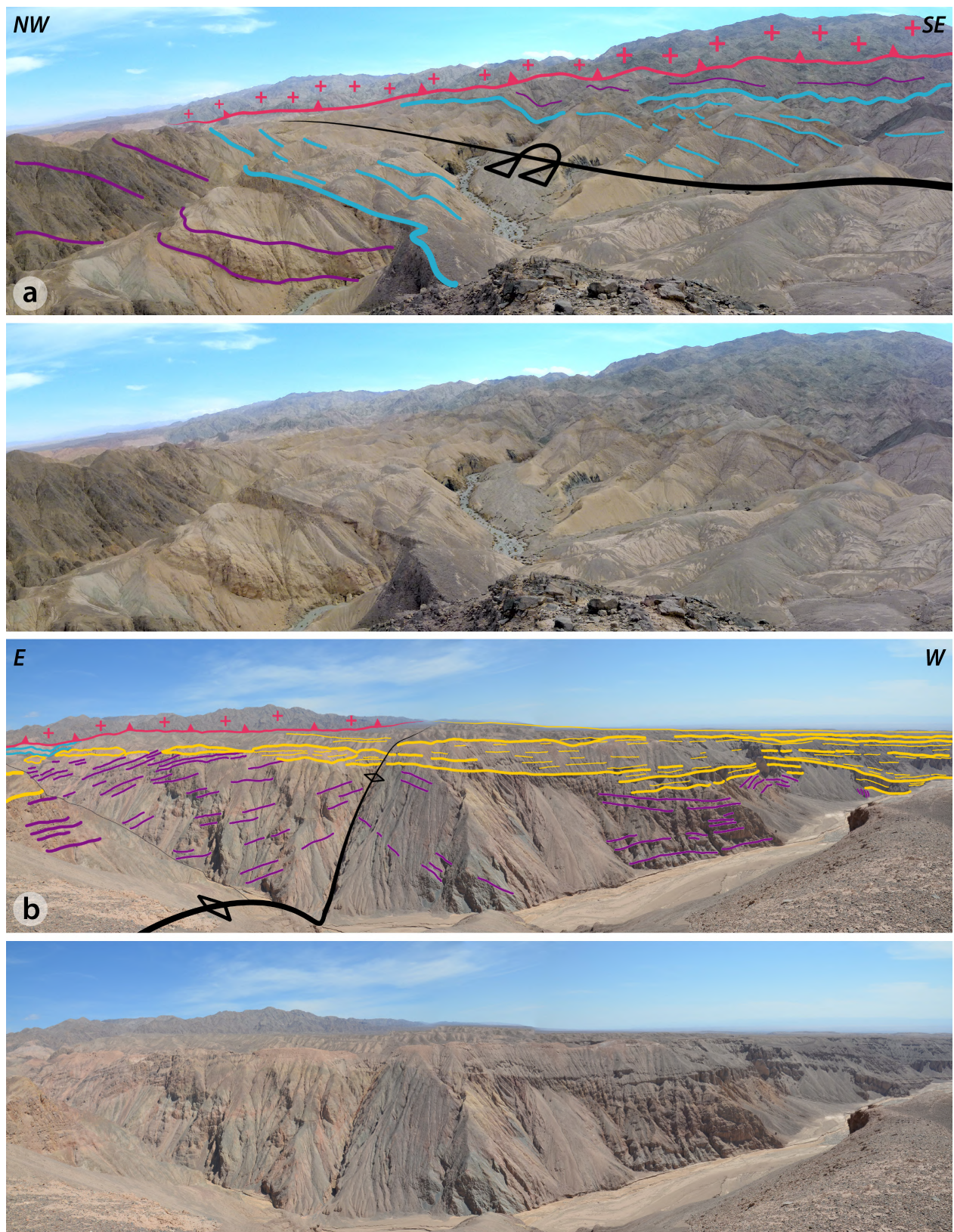


Figure S17. Field pictures of the two major folds within the Pinchal zone, as in Figure 8 in main text. Location #8a and 8b in Figure 4. **(a)** Panoramic view over the north-eastern part of the Pinchal area. **(b)** Panoramic view along Quebrada Tambillo, in the southern part of the Pinchal area. Top: interpreted picture; bottom: non-interpreted picture, for (a) and (b) respectively. For complete figure descriptions see main text.

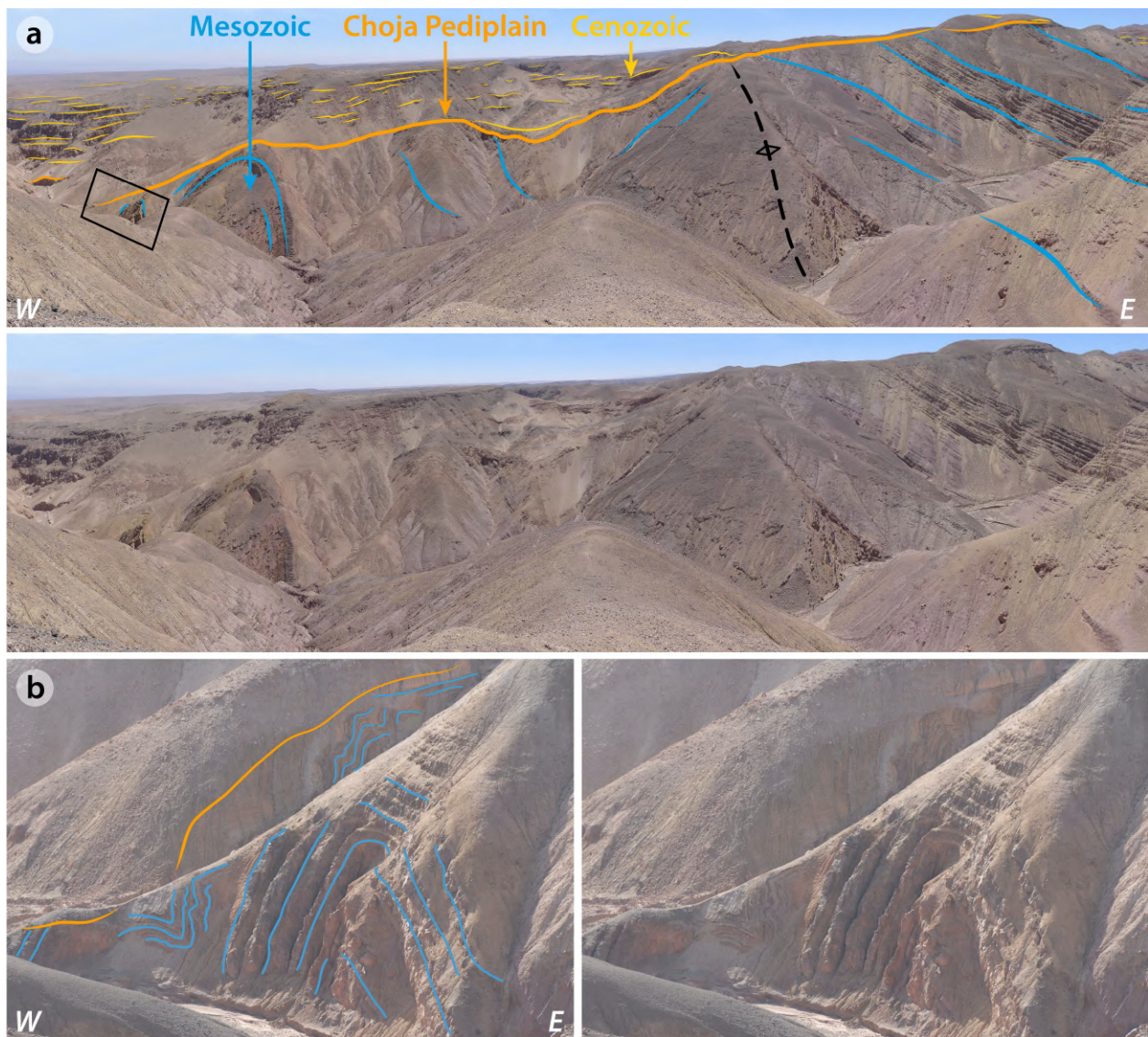


Figure S18. Field picture of the western limb of the western anticline in the Quebrada Blanca area. Same as Figure 11 in main text. Location #11 in Figure 9.

(a) Series of folds with westward decreasing amplitude and wavelength (hundreds to tens of meters) observed at the front of the western anticline. Top: interpreted picture; bottom: non-interpreted picture.

(b) Detailed view of the westernmost outcropping small-scale anticlines, located on (a) by the black box. Left: interpreted picture; right: non-interpreted picture.

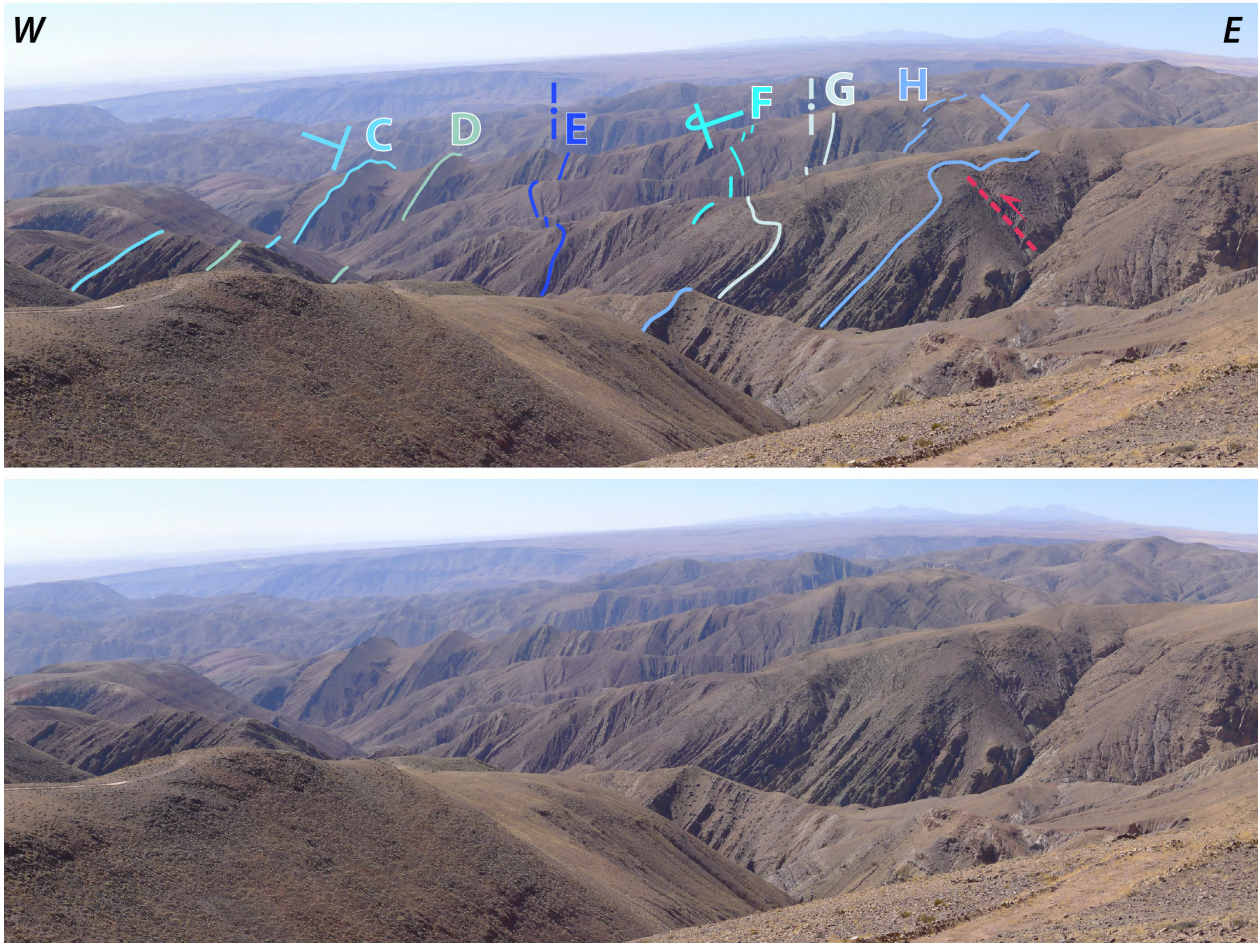


Figure S19. Landscape view on the western limb of the eastern large-scale anticline in the Quebrada Blanca area. Same as Figure 12 in main text; location #12 in Figure 9. Here, steeply inclined Mesozoic horizons are very well discernible in the landscape. Bedding traces C, D, E, F, G and H underlined here are also georeferenced on the structural map (Figure 9) by mapping on satellite imagery. Note the thrust-affected small-scale fold (red dashed line) emphasizing the west-vergence of tectonic structures. Top: interpreted picture; bottom: non-interpreted picture.

(3) Trishear modeling

Text S1. Additional information on trishear modeling (method and results)

As briefly resumed in sections 3.4 and 6.2, we used the trishear folding approach (e.g. Allmendinger, 1998; Erslev, 1991) to better constrain the amount of shortening across our study areas. We further detail the trishear method and results here.

Method

We assume fault-propagation folding to be the dominant mode of deformation in the studied fold-and-thrust-belt, and in particular in the case of the western anticlines of the cross-sections along Quebrada Tambillo (Pinchal area) and Quebrada Blanca.

We use the code FaultFold Forward version 6 (freely available from <http://www.geo.cornell.edu/geology/faculty/RWA/programs/faultfoldforward.html>) (Allmendinger, 1998) that models the distributed deformation in triangular zones at the tip of propagating faults. The formalism relies on the following parameters: the coordinates of the fault tip, the angle of the propagating fault ramp, the slip on the fault, the propagation-to-slip-ratio (P/S) of the fault, the trishear angle (i.e. the angle of the triangular zone at the tip of the fault where distributed deformation occurs), and the inclined shear angle (either parallel or similar folding) controlling the backlimb kinematics. We assume here the case of linear symmetric trishear to keep models as simple as possible, meaning that folding of the backlimb occurs parallel to the fault. We tested non-linear trishear but these trials lead to unsatisfying results when compared to our cross-sections.

For the initial conditions of the models, we assume slightly sub-horizontal layers, with a slight eastward tilt (3°E at Quebrada Tambillo, 2°E at Quebrada Blanca) as expected in the initial Andean basin. The final geometry of the fold and sedimentary cover, the fault ramp angles and bends are constrained to fit our geological cross-sections.

By adding sedimentary layers step by step during progressing deformation, we model syntectonic deposition of the Cenozoic series and reproduce the angular unconformity of the Cenozoic over Mesozoic units. The syntectonic deposits are also assumed to be initially slightly sub-horizontal, here with a slight westward tilt (3°W at Quebrada Tambillo, 6°W at Quebrada Blanca), following the approximate angle of the present-day average topography in the corresponding study areas. In fact, the basal Choja Pediplain may be comparable to the first order to the present-day average rising topography; subsequent deposition along the western mountain flank had most certainly a slight westward tilt as observed today within the eastern Atacama Bench (Figure 1). We tested the addition of horizontal syn-tectonic Cenozoic layers, but the outcoming modeled fold-forms were much less consistent with our field and map observations. We tested different dipping angles (between 1–7°W) and chose the values that allowed us to best fit our data.

The trishear modeling confirms the necessity of a fault ramp propagating from a deep décollement towards the surface to fit the observed folds, as classically observed in fold-and-thrust-belts. We simplify the fault geometry into a few fault segments: 4 segments for Quebrada Tambillo, and 3 segments for Quebrada Blanca (Figures 13, S20 and S21). Segment 1 corresponds to the flat deep detachment which is parallel to the initial Mesozoic bedding. The second fault segment ramps-up from this regional décollement with an eastward dipping angle of 24°E for Quebrada Tambillo, 40.6°E for Quebrada Blanca. These geometries are needed to fit the dip angles observed in the hanging wall of the faults (i.e. within the backlimb of the modeled folds). A shallow flat detachment is needed at both study sites to reproduce the large-scale tilt of the Cenozoic cover and the geometry of fold forelimbs. For Quebrada Tambillo, an 11.3° eastward-dipping fourth segment is necessary to best fit the surface observations in the western part of the cross-section.

Neither the trishear angle, nor the P/S ratio are deductible from geological observations (Allmendinger & Shaw, 2000), while their effect on folding is crucial as pointed out by various studies

(e.g. Hardy & Ford, 1997; Allmendinger, 1998; Zehnder & Allmendinger, 2000). With the aim to satisfactorily reproduce the geometry of the folded Mesozoic units and the tilt of the Cenozoic strata cover (Figures 5 and 10), we tested numerous combinations of parameters, in the range of values considered as reasonable in the cited studies, and regarding our geological constraints. By trial and error, we thus establish a set of best-fitting parameters for Quebrada Tambillo and Quebrada Blanca respectively, indicated in Tables S1 to S3. We recognize here that these may not be unique but only represent possible geologically viable solutions.

Results

The final stages (i.e. present-day deformation pattern) of our best fitting models are represented in Figure 13 (main text), together with the corresponding structural cross-section. The cumulative shortening, as constrained by the trishear modeling, in agreement with our geological cross-sections, is of 3.1 km for Quebrada Tambillo, and of 6.6 km for Quebrada Blanca (Figure 13). We recall here that these values account for folding and fault slip, but only for the westernmost anticlines and the two study sites.

Figures S20 and S21 illustrate the various stages of folding and fault propagation of our models, and complement the findings and discussion of section 6.3 in the main text. Some chronological constraints can be added to these various stages using geological observations, from the initial conditions prior to folding (~68 Ma), the first Cenozoic syn-tectonic deposits (~29 Ma) and to the present-day situation (0 Ma). Considering this, we find that most folding occurred prior to the first Cenozoic deposits of the Altos de Pica Formation at ~29 Ma (stage 3 on Figures S20-S21): before ~29 Ma, 2.6 km (out of the total 3.1 km) and 6.2 km (out of the total 6.6 km) of shortening had been completed for Quebrada Tambillo and for Quebrada Blanca, respectively. After ~29 Ma, the amount of additional shortening is only of 0.5 km and 0.4 km for Quebrada Tambillo and Quebrada Blanca, respectively, and corresponds to less than 20% (16% and 6%, respectively) of the total shortening, even though the duration of both time spans (~68–29 Ma and ~29–0 Ma periods) is of the same order. These findings are further represented on the graph of Figure 13d in the main text).

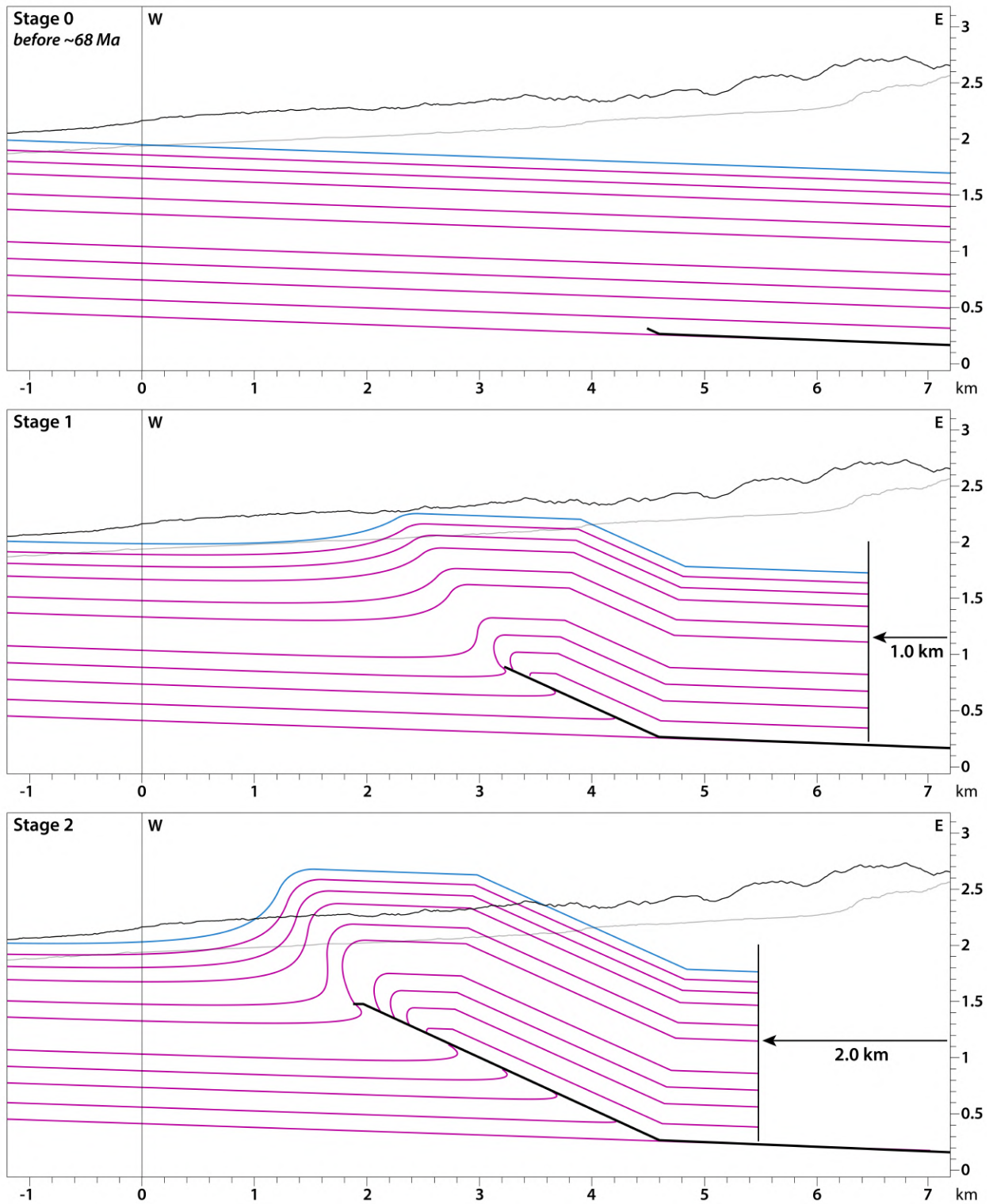


Figure S20. See second part of the figure and description next page.

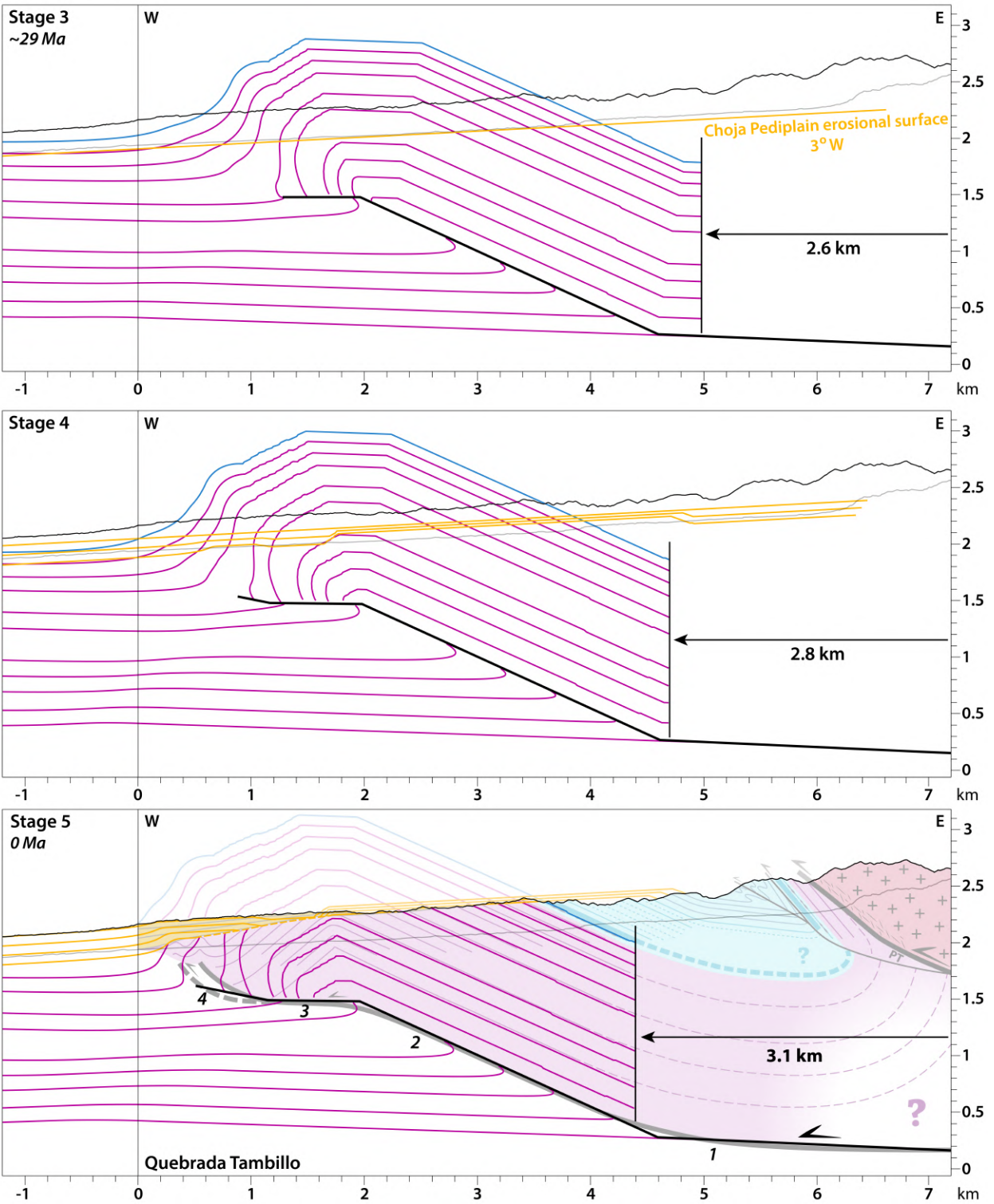


Figure S20. Outcomes from trishear modeling performed with FaultFold Forward v.6 (Allmendinger, 1998) in the case of the Quebrada Tambillo section, with chronological constraints provided from geological observations and data (see section 6.1 in main text). The black horizontal arrows underline the cumulated shortening at each stage. The final stage (present-day situation) is overlapped with our structural cross-section. Model parameters are provided in Table S2. In final stage, fault segments are numbered as in Table S1.

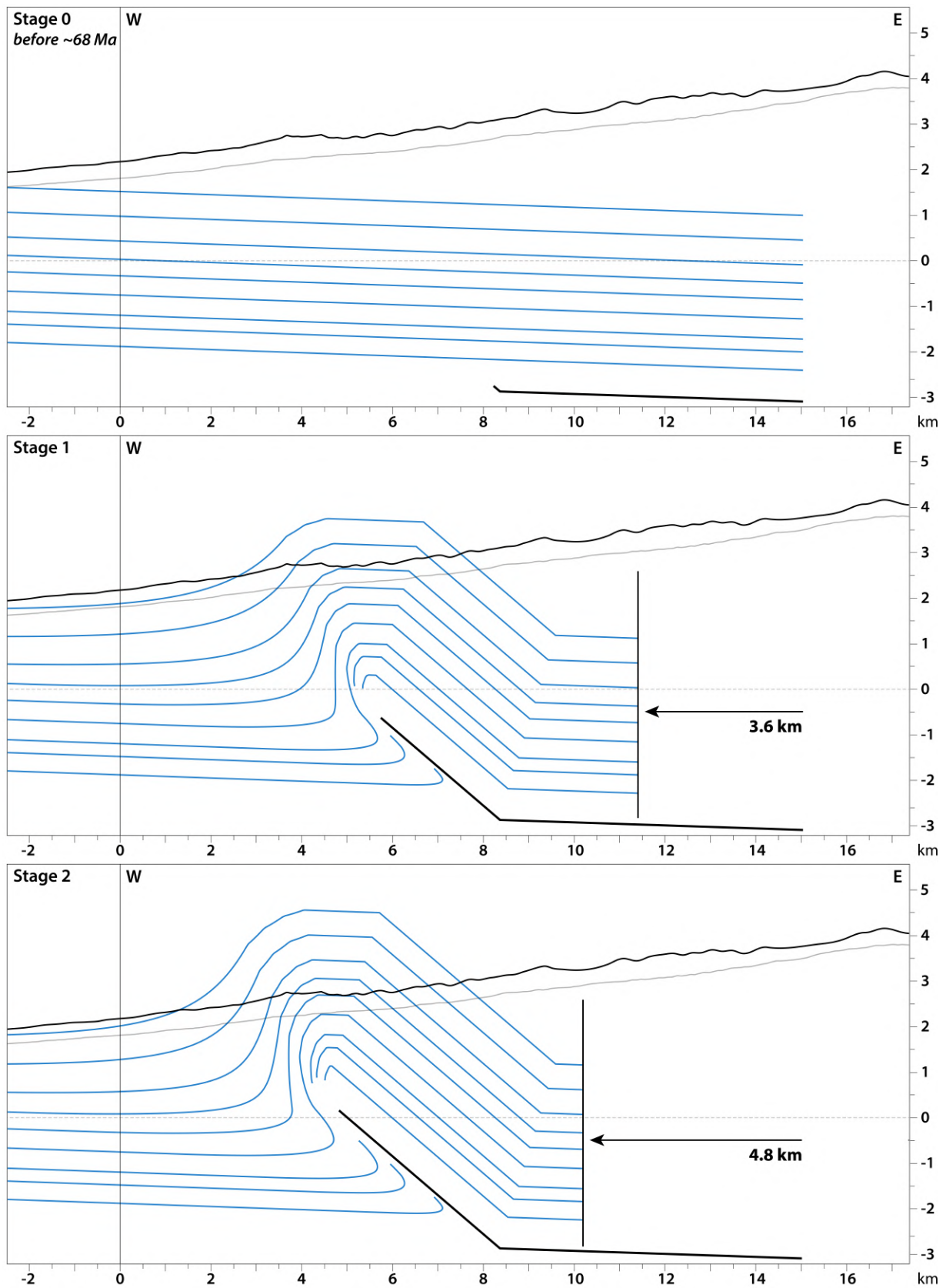


Figure S21. See second part of the figure and description next page.

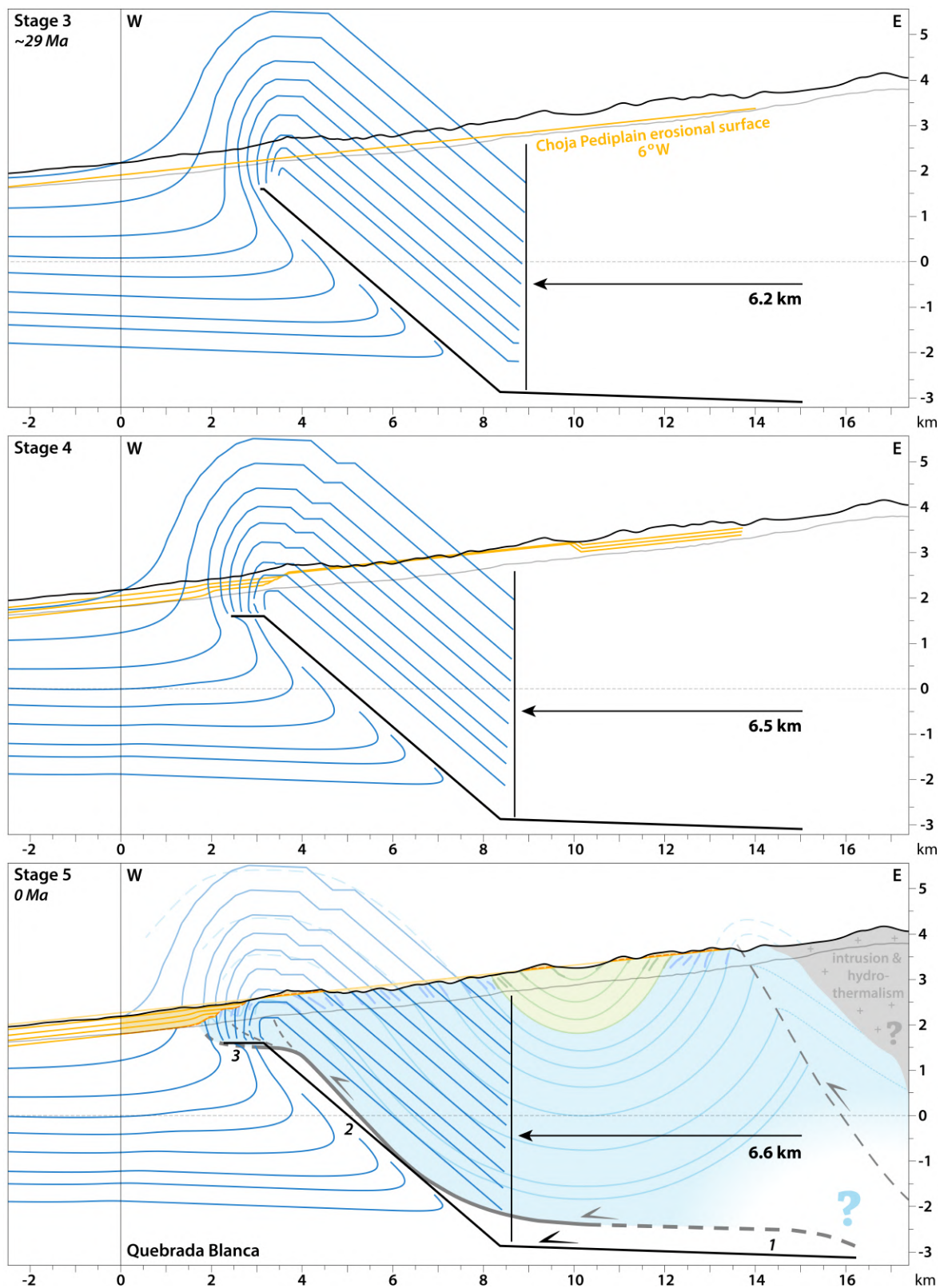


Figure S21. Outcomes from trishear modeling performed with FaultFold Forward v.6 (Allmendinger, 1998) in the case of the Quebrada Blanca section, with chronological constraints provided from geological observations and data (see section 6.1 in main text). The black horizontal arrows underline the cumulated shortening at each stage. The final stage (present-day situation) is overlapped with our structural cross-section. Model parameters are provided in Table S3. In final stage, fault segments are numbered as in Table S1.

	Fault angle	Trishear angle	P/S	Inclined shear angle	Initial bedding dip
Quebrada Tambillo	Segment 1: 0–4°E Segment 2: 20–30°E Segment 3: 0–10°E Segment 4: 5–20°E	50–110°	0.8–3.0	Parallel and similar folding tested	Mesozoic: 0–5°E Cenozoic: 0–4°W
Quebrada Blanca	Segment 1: 0–4°E Segment 2: 30–40°E Segment 3: 0–10°E	50–120°	0.7–3.0	Parallel and similar folding tested	Mesozoic: 0–4°E Cenozoic: 0–7°W

Table S1. Range of tested parameters for the trishear modeling performed with FaultFold Forward v.6 (Allmendinger, 1998). Trial and error forward modeling lead to ~65 tested models for Quebrada Tambillo and ~100 models for Quebrada Blanca. Best results came out with a fault-ramp composed of 4 segments for Quebrada Tambillo, and 3 segments for Quebrada Blanca; segments are here numbered from the deepest to the shallowest. Fault position (tips and bends) and slip on the fault are derived from our geological cross-sections. Initial layer dip angles are chosen in view of the current topography in a range of reasonable initial geometries, which allow to correctly reproduce the sections. The trishear angle controls the size of the deformed area, the P/S (propagation/slip) ratio controls the degree of folding accommodated in the trishear zone. Values for both parameters were tested based on values described as common in the literature. Concerning the inclined shear angle, best fit is obtained with parallel folding for all stages. The inclined shear angle controls the shape of the fold backlimb (Cristallini & Allmendinger, 2002).

Stage	Fault angle	Trishear angle	P/S	Slip (km)	Initial bedding dip
0	3°E	60°	1.4	0	Mesozoic: 3°E
1	24.0°E	60°	1.4	1.0	
2	0°	60°	1.2	2.0	
3	0°	60°	1.2	2.5	
4	11.3°E	90°	1.2	2.6	Cenozoic: 3°W
5	11.3°E	90°	1.2	3.1	

Table S2. Best-fit parameters for the trishear modeling performed with FaultFold Forward v.6 (Allmendinger, 1998) at Quebrada Tambillo. Model results are illustrated on Figure S20.

Stage	Fault angle	Trishear angle	P/S	Slip (km)	Initial bedding dip
0	2°E	50°	0.9	0	Mesozoic: 2°E
1	40.6°E	70°	0.9	3.6	
2	40.6°E	80°	1.0	4.8	
3	0°	90°	0.9	6.2	Cenozoic: 6°W
4	0°	90°	0.9	6.5	
5	0°	90°	0.9	6.6	

Table S3. Best-fit parameters for the trishear modeling performed with FaultFold Forward v.6 (Allmendinger, 1998) at Quebrada Blanca. Model results are illustrated on Figure S21.

Additional Supporting Information (Files uploaded separately)

Data Set S1. Georeferenced dataset (field-logistics.kmz) for visualization (e.g. on Google Earth) of strategic points and paths for the realization of our field missions (in 2018 and 2019). Data points are organized in self-explaining folders:

The folders “major roads” and “dirt tracks” contain lines showing the main paths we followed. Pink, red and orange lines are practicable by (4W drive) cars. White lines are practicable by foot only. Pay attention that we followed the “dirt tracks” for the last time in January 2019. In the case of subsequent rain, even moderate, part of the tracks may have become impracticable by car since then.

The folder “guiding points” comprises strategic (turning-) points on the road, towns and the position of our base-camp in the Pinchal area. Please leave the base-camp always clean and tidy, in the same way as you wish to find it. There, we enjoyed the moon and the stars while cooking excellent French-German-Chilean dishes.

The folder “GPS positions photos” includes two sub-folders for the two investigated areas with the localisations of the field photographs equivalent to those depicted on the structural schemes (Figures 4, 9 and S13). Color and symbol of the point-markers give additional information: Paddle symbol stands for view points, pushpin symbol for pictures illustrating stratigraphic and sedimentary observations. Red for Paleozoic basement, blue for Mesozoic units, yellow for Cenozoic deposits.

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