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1 **The Fuegian thrust-fold belt: from arc-continent collision to thrust-related**  
2 **deformation in the southernmost Andes**

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25 **Abstract**

26 New detailed structural data from the Fuegian Andes including new ages and cross-  
27 cutting relationships with intrusive rocks, as well as an appraisal of published structural  
28 data, support that this orogen evolved as a basement-involved thrust-fold belt after  
29 initial formation in an arc-continent collision scenario. New structural data from a  
30 deformed 84 Ma intrusive indicate that structures from the collisional event in the  
31 Argentine Fuegian Andes are of Campanian age, comprising only the youngest and less  
32 intense deformation of the orogenic wedge. In the internal thrust-fold belt, these  
33 structures are cut by intrusives with new ages of 74 Ma (Ar/Ar on hornblende). The  
34 superposition of thrusts on these early structures indicates a subsequent event in  
35 which a thrust-fold belt formed since the Maastrichtian-Danian. Additional new data  
36 confirm brittle-ductile thrusting in the central belt, with thrusts joining a common  
37 upper detachment in the base of the Lower Cretaceous rocks. These thrusts formed a  
38 first-order duplex system that transferred the shortening accommodated in the  
39 foreland until the Miocene.

40

41 **Keywords:** arc-continent collision, thrust-fold belt, Ar/Ar dating, deformation

42 chronology, Fuegian Andes

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**49 1. Introduction**

50 In the last 10 years a systematic research has been carried out in the Fuegian Andes  
51 attempting to unveil the structural relationships between a poorly known central belt  
52 (between Lago Fagnano and Canal Beagle, Fig. 1a), and the foreland thrust-fold belt  
53 located northwards. The connection between both domains was obscured not only by  
54 limited structural knowledge of the central belt, but also because the geometry of the  
55 foreland structures was only constrained by studies at its external portion, especially at  
56 the best exposed Atlantic coast (Álvarez-Marrón et al., 1993; Ghiglione, 2002;  
57 Ghiglione et al., 2010; Torres Carbonell et al., 2011, 2013a; Zanella et al., 2014). In  
58 addition, the internal part of the thrust-fold belt (Fig. 1a), involving mostly Cretaceous  
59 mudstones and slates with scarce stratigraphic control, was practically not addressed  
60 from a structural geology perspective, with the exception of a limited area in Chile  
61 (Klepeis, 1994; Rojas and Mpodozis, 2006).

62 Despite a greater amount of structural data from early works in the central belt of  
63 the Fuegian Andes, especially from Cordillera Darwin and surrounding areas (Fig. 1a),  
64 our analysis of that research shows us that the deformation of the foreland has been  
65 implicitly circumvented. Indeed, early investigations focused their attention on the  
66 spectacular deformation of high-grade metamorphic rocks exposed in Cordillera  
67 Darwin. These studies, as well as others along the central belt up to Isla de los Estados  
68 (Fig. 1a), determined that most of the shortening registered in the Fuegian Andes was  
69 of “mid” Cretaceous age (the “main phase” of early works), related to multiple folding  
70 and faulting generations and associated regional metamorphism (Bruhn, 1979; Dalziel  
71 and Palmer, 1979; Nelson et al., 1980; Cunningham, 1994, 1995; Kohn et al., 1995). An  
72 interesting aspect of these studies is that they led to a distinction of the structure of

73 the Fuegian Andes with respect to the rest of the Andean Cordillera. Accordingly, the  
74 style of deformation was found to be more similar to the Alpine chain (Nelson et al.,  
75 1980), and it was clear that a collisional process closing a prior back-arc basin was  
76 responsible for this structural style, in contrast with the geodynamic context of the  
77 rest of the Andes of South America (Dalziel et al., 1974; Dalziel, 1986).

78 Following, and building on this early research, more recent works revealed that  
79 after the “main phase” of deformation, thrust-related uplift of the central belt  
80 progressed during the Late Cretaceous and the Paleogene (Klepeis, 1994; Kohn et al.,  
81 1995; Gombosi et al., 2009; Klepeis et al., 2010; Maloney et al., 2011). It was  
82 recognized that at least part of the exhumation in Cordillera Darwin and its  
83 surroundings was coeval with thrusting in the foreland (e.g. Bruhn, 1979; Dalziel and  
84 Palmer, 1979; Álvarez-Marrón et al., 1993; Barbeau et al., 2009a; Gombosi et al.,  
85 2009). However, an implicit separation between deformation in both domains came  
86 out from these works, since most of the central belt thrusting was considered out-of-  
87 sequence with the structures in the foreland thrust-fold belt (e.g. Klepeis, 1994). With  
88 the incorporation of new data from the foreland, which led to Eocene-Miocene  
89 shortening estimates of tens of kilometers (Torres Carbonell et al., 2011, 2017a), it  
90 became clear that models explaining the connection between both ends of the  
91 orogenic system were increasingly needed.

92 One early approach to explain Cenozoic shortening in the foreland with coeval  
93 development of structures in the central belt was proposed by Torres Carbonell and  
94 Dimieri (2013). As a working hypothesis, these authors presented a schematic cross-  
95 section of the Fuegian Andes, combining the central belt and the foreland into a  
96 thrust-fold belt with basement involvement, in which a first-order duplex in the central

97 belt transferred all the shortening recorded in the foreland (Fig. 1b). In this sense, even  
98 if punctuated by successive deformation phases (or superposed generations of  
99 structures) as previously suggested (Bruhn, 1979; Dalziel and Palmer, 1979; Nelson et  
100 al., 1980; Cunningham, 1995; Klepeis et al., 2010), deformation in the Fuegian Andes  
101 comprised a protracted history from the Late Cretaceous to the Miocene.

102 In this work we analyze previously published structural information from the  
103 Fuegian Andes, and integrate it with new data that fill gaps in knowledge regarding the  
104 age and style of deformation of the Argentine part of the orogen. Our new data  
105 include detailed descriptions and cross-sections of two previously uncharted areas  
106 where the relationships and superposition of the main phases of deformation in the  
107 Fuegian Andes are well-revealed. We also include a detailed description of cross-  
108 cutting relationships between two intrusive suites and deformation of their host rocks,  
109 together with three new geochronological analyses of the youngest of these suites (of  
110 previously unknown age), providing key temporal constraints on the timing of these  
111 deformation phases in the Argentine Fuegian Andes. The critical appraisal of the  
112 published information together with our new data, support the proposed scheme of  
113 structural evolution of the orogen (Fig. 1b), with a first event comprising pre-  
114 Maastrichtian collision, ductile deformation and metamorphism, and a second event  
115 involving Maastrichtian-Danian to early Neogene faulting and uplift. The present study  
116 validates the interpretation of the Fuegian Andes as a basement-involved regional  
117 thrust-fold belt with a linked evolution between the central belt and the foreland.

118

119

120

## 121 **2. Regional geologic setting**

122 The origin of the Fuegian Andes involved the closure of a prior back-arc basin called  
123 Rocas Verdes Basin (Fig. 1b). This back-arc basin formed after Middle to Late Jurassic  
124 rifting of the southwestern margin of Gondwana, which preceded Early Cretaceous  
125 generation of oceanic floor between a volcanic arc and the continental margin. The  
126 basin was filled with marine successions during the Early Cretaceous (Dalziel et al.,  
127 1974; Dalziel, 1981; Stern and de Wit, 2003). Fragments of the oceanic floor are  
128 exposed as incomplete ophiolitic strips in the Tortuga, Sarmiento, and Capitán Aracena  
129 Complexes (Fig 1a), as well as intensely deformed and metamorphosed amphibolites  
130 at Cordillera Darwin (Cunningham, 1994; Stern and de Wit, 2003; Klepeis et al., 2010;  
131 Calderón et al., 2016). These ophiolitic rocks represent remnants of the Late  
132 Cretaceous collision between the magmatic arc and the continent during closure of the  
133 Rocas Verdes Basin (Nelson et al., 1980; Dalziel, 1986; Cunningham, 1994, 1995;  
134 Klepeis et al., 2010). Near Cordillera Darwin, the ophiolitic rocks are highly strained,  
135 involved in mylonitic zones that have been interpreted either as related to convergent  
136 strike-slip deformation (Cunningham, 1995), or as thrust sheets reflecting north-  
137 northeastward obduction of part of the oceanic floor during basin closure (Klepeis et  
138 al., 2010). The Canal Beagle (Fig. 1a) manifests itself as a conspicuous structural  
139 boundary separating strongly deformed and metamorphosed rocks in Cordillera  
140 Darwin from less deformed and mainly magmatic rocks towards the south.

141 Closure of the Rocas Verdes Basin and further development of the Fuegian Andes  
142 also led to formation of a Late Cretaceous-Neogene foreland basin system in front of  
143 the orogen, called Austral basin (Magallanes in Chile, Biddle et al., 1986, Fig. 1a, b).  
144 Protracted deformation and development of the thrust-fold belt, therefore, affected



145 syntectonic successions deposited in this basin. The stratigraphy of these successions  
146 as well as their overprinting relationships with the structures of the thrust-fold belt has  
147 been helpful to constrain several stages of uplift, erosion, and orogenic expansion  
148 (Torres Carbonell and Olivero, 2019). A more detailed description of the stratigraphy  
149 and a summary of the tectonic stages of the Fuegian Andes are given in the following  
150 sections.

151

## 152 **2.1. Stratigraphic framework**

153 The stratigraphic framework of the Fuegian thrust-fold belt is synthetically  
154 described, focusing on lithologies and thicknesses involved in deformation, and  
155 chronologic information useful to constrain structural stages (Fig. 2). For the sake of  
156 brevity, we omit further detailed information that can be found in the cited references.  
157 Throughout the text, we use formal names defined in Argentina for units that are also  
158 recognized in Chile, as shown in Fig. 2. For simplicity, we omit in our framework the  
159 back-arc mafic floor, which does not crop out in Argentina, as well as some intrusives  
160 of unknown age.

161

### 162 **2.1.1. Paleozoic basement**

163 The basement is included in the Cordillera Darwin Metamorphic Complex, which  
164 portrays metamorphic grades from greenschist to upper amphibolite facies (Hervé et  
165 al., 2010a, 2010b). Within these metamorphic rocks, a distinction is made amongst the  
166 Mesozoic cover (Jurassic synrift deposits and Cretaceous sedimentites) and the “true  
167 basement” of the Rocas Verdes Basin (Figs. 2 and 3). The latter, composed mainly of  
168 phyllites, schists and metabasites, is characterized by pre-Permian detrital zircon ages

169 (Hervé et al., 2010b). A single zircon U/Pb crystallization age of  $153.12 \pm 0.93$  Ma from  
170 a granite dike intruding a penetrative foliation in a schist at southern Cordillera Darwin  
171 (Klepeis et al., 2010, see also Hervé et al., 2010b), constrains the pre-Jurassic  
172 deformation of the basement (called  $D_B$  by Nelson et al., 1980).

173

174 2.1.2. Jurassic synrift and Lower Cretaceous back-arc basin volcanic and  
175 sedimentary rocks

176 2.1.2.1. *Lapataia Formation*

177 The Lapataia Formation (Figs. 2 and 3) includes greenschists facies  
178 metasedimentites and metabasites that are thrust over the Upper Jurassic Lemaire  
179 Formation (Bruhn, 1979; Olivero et al., 1997; Cao et al., 2018). The total thickness of  
180 the Lapataia Formation is undetermined due to its intense deformation, unknown  
181 base, and the tectonic nature of its contact with the Lemaire Formation. Recently  
182 published structural and petrographic data suggest that this unit represents a highly  
183 deformed section of the synrift Jurassic succession, stratigraphically between the  
184 Paleozoic basement and the Lemaire Formation (Cao et al., 2018; see also Olivero et  
185 al., 1997; Acevedo, 2019).

186

187 2.1.2.2. *Lemaire Formation*

188 The Lemaire Formation (Figs. 2 and 3), widely exposed in the central belt domain, is  
189 composed of very-low to low grade metasedimentary and metavolcanic-volcaniclastic  
190 rocks deposited during the synrift stage (Bruhn et al., 1978; Hanson and Wilson, 1991;  
191 Olivero and Martinioni, 1996a; González Guillot et al., 2016; González Guillot, 2017;  
192 Cao et al., 2018). This unit is in tectonic contact with the Lapataia Formation, and

193 covered by the Lower Cretaceous Yahgan and Beauvoir formations, although a  
194 detachment surface usually overprints this contact (Torres Carbonell and Dimieri,  
195 2013; Cao et al., 2018). The total thickness of the Lemaire Formation is unconstrained.

196 In Argentine Tierra del Fuego, a U/Pb zircon age in rhyolite yielded an age of 164  
197  $\pm 3.6$  Ma (Palotti et al., 2012). The equivalent Tobífera Formation (Fig. 2), by correlation  
198 with the Darwin suite (see below), has geochronological ages between 178-152 Ma  
199 (Pankhurst et al., 2000; Barbeau et al., 2009a; Hervé et al., 2010b; Klepeis et al., 2010).

200

#### 201 *2.1.2.3. Yahgan Formation*

202 The Yahgan Formation (Figs. 2 and 3) comprises very low-grade metasedimentary  
203 rocks, originated from marine epiclastic and volcanoclastic successions of the Rocas  
204 Verdes Basin (Suárez and Pettigrew, 1976; Winn, 1978; Dalziel, 1981; Olivero and  
205 Martinioni, 1996a). The unit rests on the Lemaire Formation, the contact usually  
206 overprinted by a detachment surface (Torres Carbonell and Dimieri, 2013; Cao, 2019).  
207 The top of the formation is unknown.

208 Estimations of the original thickness of the Yahgan Formation average from 1400 m  
209 near Ushuaia and increasing southwards (Winn, 1978; Caminos et al., 1981; Olivero  
210 and Malumián, 2008). The formation is not younger than the late Albian, according to  
211 its fossils and scarce geochronological data (Olivero and Martinioni, 1996b; Barbeau et  
212 al., 2009b).

213

#### 214 *2.1.2.4. Beauvoir Formation*

215 The Beauvoir Formation (Figs. 2 and 3) comprises slates with a marine protolith,  
216 that may interfinger southwards with the Yahgan Formation (Olivero and Malumián,

217 2008; Martinioni et al., 2013). South of Lago Fagnano, the contact with the underlying  
218 Lemaire Formation is a few meters below a north-northeast-dipping detachment  
219 surface (Torres Carbonell and Dimieri, 2013; González Guillot et al., 2016; Cao, 2019;  
220 this work), whilst at Península Mitre this contact is a south-dipping reverse fault  
221 (Torres Carbonell et al., 2017b). Its top is not well defined due to lack of stratigraphic  
222 definition north of Lago Fagnano. A minimum thickness of 450 m has been estimated  
223 for the formation, which contains Aptian-Albian marine invertebrates, mostly  
224 inoceramids (Olivero et al., 2009; Martinioni et al., 2013).

225

226 2.1.3. Upper Cretaceous-lower Neogene foreland basin sedimentary rocks

227 2.1.3.1. *Upper Cretaceous-Danian succession*

228 An Upper Cretaceous-Danian sedimentary package forms the oldest sedimentary  
229 succession of the Austral foreland basin (Figs. 2 and 3). The older rocks in this  
230 succession include several mudstone-dominated units with Turonian-Campanian  
231 fossils (Fig. 2, Olivero et al., 2009; Martinioni et al., 2013). The middle part of the  
232 succession is formed by coarse-grained deposits interbedded with mudstones, of late  
233 Campanian-?early Maastrichtian age according to ammonites and foraminifera (Bahía  
234 Thetis Formation, Olivero et al., 2003). The youngest rocks in this package are  
235 bioturbated fine sandstone-mudstone intercalations with Maastrichtian-Danian fossils  
236 and detrital zircon ages (Policarpo Formation, Olivero et al., 2002, 2003; Barbeau et al.,  
237 2009a; Martinioni et al., 2013).

238 The base of the Upper Cretaceous-Danian succession is not well-defined, whilst the  
239 top is marked by an unconformity between the Policarpo Formation and coarse

240 deposits of the Río Claro Group (Fig. 2, Olivero et al., 2003; Martinioni et al., 2013). The  
241 accumulated minimum thickness of the succession exceeds 1000 m.

242

#### 243 *2.1.3.2. Paleocene-lower Miocene succession*

244 The Paleocene to lower Miocene sedimentary rocks of the Austral basin comprise a  
245 heterogeneous succession including several units that crop out, or are drilled in  
246 subsurface, from the mountain front to northern Tierra del Fuego (Figs. 2 and 3).  
247 Minimum estimated thicknesses of each unit are given in Fig. 2, from composite  
248 sections. The base of this succession rests on an unconformity with the Policarpo  
249 Formation (Olivero et al., 2003; Martinioni et al., 2013).

250 The upper part of the succession is formed by uppermost Oligocene-Miocene beds  
251 coeval with the end of contraction in the thrust-fold belt (Torres Carbonell and Olivero,  
252 2019). They include growth strata formed above the youngest folds (Ghiglione, 2002;  
253 Malumián and Olivero, 2006; Ponce et al., 2008; Torres Carbonell et al., 2017a).

254

#### 255 *2.1.4. Fuegian batholith and peripheral intrusives north of Canal Beagle*

##### 256 *2.1.4.1. Darwin suite*

257 The oldest magmatic rocks north of Canal Beagle comprise orthogneisses mostly  
258 derived from granites intruded in the basement at Cordillera Darwin (Fig. 3, Nelson et  
259 al., 1980; Hervé et al., 1981, 2010b; Klepeis et al., 2010). The rocks are ductilely  
260 deformed and intruded by less deformed granitoids of the Beagle suite (Nelson et al.,  
261 1980). The Darwin granites are Late Jurassic, with U/Pb zircon ages of 164-153 Ma  
262 (Mukasa and Dalziel, 1996; Klepeis et al., 2010).

263

264 *2.1.4.2. Beagle suite*

265 The Beagle suite is the main unit of the Fuegian batholith (Fig. 3). North of Canal  
266 Beagle, the granitoids that compose this suite have U/Pb zircon ages ranging between  
267 86-74 Ma (Hervé et al., 1984; Klepeis et al., 2010; McAtamney et al., 2011). They  
268 postdate most of the ductile structures related to the arc-continent collision, although  
269 some intrusives overlap with the late stages of that deformation and the associated  
270 peak metamorphism (Maloney et al., 2011).

271

272 *2.1.4.3. Rear-arc suites*273 *2.1.4.3.1. Ushuaia Peninsula Andesites*

274 The Ushuaia Peninsula Andesites (Fig. 3) include small ultramafic to silicic plugs  
275 cross-cut by a set of andesitic (the main lithology) and lamprophyre dikes (González  
276 Guillot et al., 2011). This suite intrudes the Yahgan Formation. One andesite dike  
277 yielded a zircon U/Pb age of  $84.1 \pm 1.6$  Ma (González Guillot et al., 2018). The  
278 overprinting relationships between the Ushuaia Peninsula Andesites and the  
279 deformation of the Yahgan Formation are reported in this work as a constraint on  
280 deformation timing (see section 4.1.1).

281

282 *2.1.4.3.2. Fuegian Potassic Magmatism*

283 The Fuegian Potassic Magmatism comprises small ( $< 25$  km<sup>2</sup>), isolated plutons  
284 emplaced in the Yahgan and Beauvoir formations (Fig. 3, González Guillot et al., 2009).  
285 The plutons are composite, ranging from ultramafic to felsic facies, and with a  
286 characteristic mildly alkaline chemistry (González Guillot, 2016). The intrusives are the  
287 Ushuaia, Kranck, Moat and Jeu-Jepén plutons (Fig. 3).

288 The Kranck pluton has especial interest for this work since it is associated with two  
289 sets of dikes and sills (Cerro Rodríguez dikes) that intrude the Beauvoir Formation  
290 towards the north of the pluton, within the internal thrust-fold belt domain (Torres  
291 Carbonell et al. 2017c). We provide below (section 4.2), and later discuss, the  
292 overprinting relationships of these dikes with the deformation of the host rock, as well  
293 as new geochronological data of the Kranck pluton and Cerro Rodríguez dikes (section  
294 5). Previously reported U/Pb zircon ages are  $75 \pm 1.0$  Ma to  $70.9 \pm 1.7$  Ma for the  
295 Ushuaia pluton (Barbeau et al., 2009b; González Guillot et al., 2018), and  $72.01 \pm 0.75$   
296 Ma for the Jeu-Jepén pluton (Cerrodo et al., 2011).

297

## 298 **2.2. Tectonic framework**

299 Contractual deformation in the Fuegian Andes started with the northward  
300 obduction of the Rocas Verdes Basin oceanic floor in Cordillera Darwin, which was also  
301 associated with southward underthrusting of the continental margin of the basin, both  
302 processes comprising a first deformation phase in the region associated with arc-  
303 continent collision (Fig. 1b, Klepeis et al. 2010). At mid-crustal depths, underthrusting  
304 and obduction caused high shear strain and peak metamorphism under upper-  
305 amphibolite facies conditions, reaching 12 kbar and  $\sim 600$  °C, and later decompressing  
306 to 9 kbar during peak temperatures of more than 600 °C. This is recorded in Bahía Pía  
307 (all localities shown in Fig. 3), where a high-grade shear zone has been described (Kohn  
308 et al., 1993, 1995; Klepeis et al., 2010; Maloney et al., 2011).

309 Modelled Ar/Ar ages of hornblende, muscovite, biotite and K-feldspar from  
310 metamorphic rocks of Cordillera Darwin indicate that first-phase deformation was  
311 ongoing by 90 Ma, with a stage of rapid cooling between 90-70 Ma (Fig. 4, Kohn et al.,

1995; see also Nelson, 1982). Other evidence of this rapid cooling trend, and associated uplift and erosion of the orogenic core, is recorded in the Campanian-?lower Maastrichtian Bahía Thetis Formation and equivalent Cerro Matrero Formation in Chile (Fig. 2). These units contain clasts of eroded metavolcanic and metasedimentary rocks, and detrital zircons in the Cerro Matrero Formation with the younger ages in the range of 82-145 Ma, derived from the back-arc basin fill, basaltic floor and older synrift sequence (Fig. 4, Olivero et al., 2003; McAtamney et al., 2011).

The first-phase structures are intruded by Beagle suite granites without ductile deformation, which suggests an age older than 86-74 Ma for this phase (Klepeis et al., 2010). However, some contemporaneity between Beagle granite intrusions and first-phase structures is indicated by an age of  $72.6 \pm 2.2$  Ma from U-Th-Pb in-situ dating of a late first-phase metamorphic monazite from Bahía Pía (Fig. 4, Maloney et al., 2011). In Argentina, first-phase structures are intruded by the Ushuaia pluton, with a well-developed contact-metamorphism aureole that overprints a previous foliation (Fig. 4, González Guillot et al., 2018). In Península Mitre, the Bahía Thetis Formation is the youngest unit affected by first-phase deformation (Fig. 4, Torres Carbonell et al., 2017b). This broadly suggests a late Campanian-?early Maastrichtian age for the end of this deformation phase.

The Beagle suite granites are affected by a second, more brittle deformation phase associated with at least three major thrust sheets in the Cordillera Darwin region (Klepeis et al., 2010). Similar structures from Argentina, and their relationship with the second phase, are reported in this work. The thrusts in Cordillera Darwin are responsible for uplift of the high-grade metamorphic rocks simultaneously with retrograde metamorphism (Kohn et al., 1993) since at least  $\sim 70$  Ma (Fig. 4, Klepeis et



336 al., 2010; Maloney et al., 2011). Nonetheless, there are structures in Bahía Pía related  
337 to retrogression of metamorphic assemblages (thus second-phase), which are cut by  
338 the Beagle suite (F3/S3 in Klepeis et al., 2010). Therefore, we notice a possibly  
339 contemporaneous development between early second-phase thrusting and intrusion  
340 of some Beagle granites in that region. This probable contemporaneity between the  
341 Beagle suite and the transition from the first collision-related phase and the second  
342 thrust-related phase highlights the need for additional independent age constraints  
343 from different portions of the Fuegian Andes. This is especially important to document  
344 a possible diachronic progression of deformation in the orogen.

345 The second thrusting phase has been compared with the 60-40 Ma rapid cooling  
346 trend modelled from Ar/Ar closure temperature ages on K-feldspar, hornblende,  
347 muscovite, and biotite, Rb/Sr closure temperature ages on biotite, and fission track  
348 ages of titanite, zircon and apatite, from metamorphic rocks of Cordillera Darwin (Fig.  
349 4, Kohn et al., 1995; see also Nelson, 1982). More recent zircon and apatite fission  
350 track and (U-Th-Sm)/He ages of samples from southern Cordillera Darwin, Isla Gordon  
351 and the Argentine central belt, suggest the onset of a rapid cooling stage since ~48 Ma,  
352 ending by ~34 Ma (Fig. 4, Gombosi et al., 2009). These cooling trends are consistent  
353 with further uplift, erosion, and a more prominent supply of detritus eroded from the  
354 basement and especially Jurassic components of the orogen to the foreland basin, as  
355 recorded from 150-180 Ma detrital zircons and from the petrography of Paleocene-  
356 Eocene sedimentary rocks (Barbeau et al., 2009a; Torres Carbonell and Olivero, 2019).  
357 Additional evidence of thrust-related uplift in the central belt comes from the  
358 documented Oligocene thrusting affecting the Lemaire Formation at Bahía Sloggett  
359 (Fig. 4, Olivero et al., 1998).

360 Development of the foreland thrust-fold belt is coeval with both phases of  
361 deformation in the central belt of the orogen (Fig. 4). The thrust-fold belt is subdivided  
362 into an internal and an external portion (Figs. 1a and 3), given their different structural  
363 styles (cf. Torres Carbonell et al., 2017a). Both portions and the central belt, however,  
364 are structurally linked, as we will discuss in this work.

365 The internal thrust-fold belt comprises structures between the central belt and a  
366 major structure called Apen-Malvinera thrust system (Fig. 3) that roughly defines the  
367 mountain front in Tierra del Fuego. Previous structural work in the internal thrust-fold  
368 belt has been published by Klepeis (1994), Zanella et al. (2014) and Torres Carbonell et  
369 al. (2013b, 2017b). The structure of the internal thrust-fold belt will be addressed in  
370 the following sections together with new structural and stratigraphic data from this  
371 region.

372 The external thrust-fold belt comprises the shallowest structures of the Fuegian  
373 Andes. These structures, as observed in seismic sections, are rooted in two main  
374 detachments localized below Paleocene strata and below the Cretaceous succession,  
375 respectively (Álvarez-Marrón et al., 1993; Zanella et al., 2014; Torres Carbonell et al.,  
376 2017a). Below these detachments, some deeper structures appear to be splays  
377 associated with the deep thrusts emplaced in the central belt (Torres Carbonell et al.,  
378 2017a).

379 The structures in the external thrust-fold belt have been interpreted as fault-  
380 propagation and fault-bend folds, with both foreland and hinterland vergence. Most of  
381 the thrusts are rarely exposed, except at the Apen-Malvinera thrust system and at the  
382 Atlantic coast, where few of them display cataclastic fabrics (Torres Carbonell et al.,

383 2011). No penetrative fabrics (foliations) have been recognized in the external thrust-  
384 fold belt.

385 Several fracture sets, analyzed for paleostress directions, indicate compression  
386 perpendicular to most of the structures during the Paleogene (Maestro et al., 2019). In  
387 Península Mitre, on the other hand, the curvature of the thrust-fold belt forms a recess  
388 (Península Mitre recess). Based on fracture paleostress directions and strain  
389 measurements, this recess has been interpreted as a progressive arc with  
390 development of superposed deformations related to tangential longitudinal strain  
391 (Torres Carbonell et al., 2019).

392 In the Atlantic coast, previous work defined several contractional stages (Df2-Df6 in  
393 Fig. 4) based on the overprinting relationships of structures with well-exposed  
394 unconformity-bounded units (Torres Carbonell et al., 2011). The latter units, in turn,  
395 have well-constrained ages based on foraminifers, nanoplankton, dinocysts, and in  
396 some cases detrital zircon ages (Malumián and Olivero, 2006; Olivero and Malumián,  
397 2008; Barbeau et al., 2009a; Torres Carbonell and Olivero, 2019).

398 These contractional stages, further grouped as D2 phase in Torres Carbonell et al.  
399 (2017a), developed from the early Eocene to the latest Oligocene-earliest Miocene.  
400 During these times, contractional deformation ceased simultaneously with deposition  
401 of growth strata in the deformation front (Df6, Fig. 4, Ghiglione, 2002; Torres Carbonell  
402 et al., 2011, 2013a).

403 According to balanced cross-sections, accumulated shortening in the Paleogene  
404 sequence vary between 11 and 28 km (16-45%, Torres Carbonell et al. 2017a). Torres  
405 Carbonell and Dimieri (2013), and later Torres Carbonell et al. (2017a), argued that this  
406 shortening must necessarily be generated by transference of major structures in the

407 hinterland, carried on ramps above the main detachments below the Cretaceous and  
408 Paleocene sequences. According to this model, the second-phase structures of the  
409 central belt are associated with uplift and transference of shortening to the foreland  
410 through formation of an antiformal stack of basement and Upper Jurassic rocks (Fig.  
411 1b, Torres Carbonell and Dimieri, 2013). The upper detachment of this first-order  
412 duplex is located near the contact between the Upper Jurassic synrift sequences and  
413 the Lower Cretaceous fill of the Rocas Verdes Basin (Klepeis, 1994; Kley et al., 1999;  
414 Rojas and Mpodozis, 2006; Torres Carbonell and Dimieri, 2013; this work).

415

### 416 **3. Methods and new data**

417 New data used in this paper come from fieldwork in the Argentine Fuegian Andes.

418 Field data consist of attitudes and cross-cut relationships between different  
419 generations of structures, and between structures and several igneous bodies.

420 Oriented samples have been studied under the microscope in order to define  
421 metamorphic assemblages and/or microstructural relationships. In particular we  
422 present here detailed descriptions of two previously unmapped areas of the Fuegian  
423 Andes (located in Fig. 3): Cañadón Bianchi, at Montes Martial, and a cross-section  
424 along a creek in northern Sierra Alvear (Arroyo Velazquito). In these areas we obtained  
425 new supporting evidence for the superposition of younger, brittle-ductile thrust-  
426 related structures on older ductile structures. All spherical projections shown were  
427 constructed with the software Stereonet 10.1.6 from R. W. Allmendinger.

428 In order to constrain the ages of both structural generations, we studied the cross-  
429 cutting relationships of selected intrusives that due to their position in the Fuegian  
430 Andes, allow a spatiotemporal control on the progression of deformation. On one

431 hand, we studied the Ushuaia Peninsula Andesites, which constitute the southernmost  
432 intrusives north of Canal Beagle in Argentine Tierra del Fuego, with an available  
433 crystallization age. On the other hand, we studied the northernmost intrusives in the  
434 thrust-fold belt: the Kranck pluton and related dikes and sills between the pluton and  
435 Cerro Rodríguez (Cerro Rodríguez dikes). We present new structural descriptions, and  
436 three new radiometric ages that constitute key constraints on the chronology of  
437 deformation in the Fuegian thrust-fold belt.

438 A U/Pb zircon age was obtained from a quartz monzonite of the Kranck pluton at  
439 the Geochronology Laboratory of the University of Brasilia (Brazil), by LA-ICP-MS.  
440 Other two hornblende Ar/Ar ages were obtained from a hornblendite of the Kranck  
441 pluton, and from an undeformed hornblende lamprophyre from the Cerro Rodríguez  
442 dikes, described in this work. Both Ar/Ar ages were obtained at the Earth Sciences  
443 Institute at Orléans (ISTO), INSU-CNRS, University of Orléans (France), by step heating.  
444 The methodological details and the tabulated results are given in the Supplementary  
445 File 1.

446 A balanced cross-section is presented in section 7.2. This cross-section was  
447 constructed with Move 2018.2 software by sequential restoration of individual  
448 structures, with conservation of shortening transferred from deeper structures to  
449 shallower detachments as a premise. After initial construction, the cross-section was  
450 tested by forward modelling using Move-on-Fault tools in Move. The method selected  
451 was fault-bend folding, except from a few frontal structures modelled with trishear.

452

453

454

#### 455 **4. New structural data**

##### 456 4.1. New structural data of the central belt

457 Our new structural data from the central belt includes detailed structural  
458 descriptions of, on one hand, the cross-cutting relationships between the Ushuaia  
459 Peninsula Andesites and deformation in their host rock, and on the other hand the  
460 superposition of brittle-ductile thrusts on prior ductile deformation, as well as the  
461 connection between these thrusts and a main upper detachment below the  
462 Cretaceous rocks. These latter brittle-ductile structures are described in two different  
463 areas of the central belt (Cañadón Bianchi and Arroyo Velazquito). Even though some  
464 thrusts and the detachment have been reported without much detail in other areas of  
465 the Fuegian Andes (e.g. Torres Carbonell and Dimieri, 2013; Torres Carbonell et al.,  
466 2017b), we provide here the first detailed description of these structures, and their  
467 superposed nature on previous ductile structures, in the area between Lago Fagnano  
468 and Canal Beagle.

469

##### 470 4.1.1. Deformation of the Ushuaia Peninsula Andesites

471 The Ushuaia Peninsula Andesites were previously thought to postdate ductile  
472 deformation, based on unreported deformation or metamorphism in the intrusive  
473 rocks (González Guillot et al., 2011). Indeed, the host rock (Yahgan Formation),  
474 affected by contact metamorphism, preserves its original well-stratified character,  
475 without recognizable centimeter-scale folds (Fig. 5a-b) such as it is observed elsewhere  
476 surrounding Ushuaia, and along the Canal Beagle only a few kilometers away from  
477 Península Ushuaia (e.g. surrounding the Ushuaia pluton). Only occasionally the Yahgan  
478 Formation shows a spaced rough cleavage away from the main body of the intrusive

479 suite (Fig. 5c). Farther away ( $> 1$  km), however, at the southeastern tip of the  
480 peninsula, the Yahgan Formation reveals a more intense deformation, with  
481 centimeter-scale folds and a closely spaced cleavage formed by pressure-solution  
482 seams. At this location, an isolated, 1.5 m thick andesite dike, shows spaced pressure-  
483 solution cleavage parallel to the foliation in the Yahgan Formation (Fig. 5d-f).

484 A comparison between the poorly deformed host rock, with only a spaced,  
485 pressure-solution cleavage of stylolitic morphology (Fig. 5e-f), and the stronger  
486 deformation in the Yahgan Formation away from the main intrusive, suggests that the  
487 host rock has been protected from ductile deformation due to the existence of a  
488 coherent, very competent igneous rock. The absence of penetrative structures in the  
489 stronger rheologies, with selective development in the weaker rocks, is typical of this  
490 part of the central belt, and consistent with the temperature conditions suggested by  
491 the low-grade metamorphism (see section 6.1).

492

#### 493 4.1.2. Brittle-ductile thrusts and detachment superposed on ductile structures

494 At Montes Martial, in the area known as Cañadón Bianchi, we have mapped three  
495 brittle-ductile fault zones that constitute thrusts below a common upper detachment  
496 (Fig. 6a-b). The thrusts cut up-section metavolcaniclastic and metasedimentary rocks  
497 of the Lemaire Formation. These fault zones are composed of cohesive cataclastic  
498 rocks (Fig. 6c-d), with folded and deflected older foliations, forming characteristic s-c  
499 type fabrics with top-to-north shear sense (Fig. 6e). A clear superposition of these  
500 brittle-ductile structures on the prior ductile fabrics is observed in outcrop as well as  
501 under the microscope (Fig. 6f-g). The older foliations are continuous cleavages formed  
502 mostly by white micas and chlorite, as well as deformed quartz and plagioclase. In the

503 metasedimentary facies this cleavage transposes bedding forming a characteristic  
504 banded fabric (Fig. 6f).

505 The southernmost thrust joins a detachment located near the base of the Yahgan  
506 Formation, just below a thick basic sill known as Puente Quemado Gabbro (described  
507 in González Guillot et al., 2016, Fig. 6b). The detachment comprises a brittle-ductile  
508 fault zone about 5 meters thick, characterized by cataclastic fabrics similar to the ones  
509 described for the thrusts, but affecting in this case slates of the Yahgan Formation. A  
510 notorious feature is the presence of faulted folds (classes 2 or 1C) in the detachment  
511 zone (Fig. 6h), as well as crenulation of the prior slaty cleavage (Fig. 6i). Kinematic  
512 indicators (deflected foliation in s-c type fabrics) indicate top-to-north movement (Fig.  
513 6j). Both the thrusts and the detachment are folded by first-order structures, and in a  
514 SW-NE cross section they are involved in a large synform that plunges to the SE (Fig.  
515 6a, k).

516 At the northern face of Sierra Alvear, along Arroyo Velazquito, the same  
517 detachment horizon is observed at the base of the Beauvoir Formation (Fig. 7). At this  
518 location, the detachment surface is folded and exposed in the vertical, frontal limb of  
519 an antiform affecting metavolcaniclastic facies of the Lemaire Formation, covered by  
520 lapillitic mudstones of the Beauvoir Formation (Fig. 7a). Towards the north, the  
521 Beauvoir Formation attains a gentle northward dip, thus the detachment dips below  
522 Lago Fagnano in that direction (see Fig. 3). The detachment itself is characterized by  
523 cataclastic fabrics, faulted folds, and deflected older foliations with s-c type fabrics  
524 showing top-to-north shear sense (Fig. 7b-d, see also Fig. 8b).

525

526



## 527 4.2. New structural data of the internal thrust-fold belt

528 We report here new detailed descriptions of the structures affecting Cretaceous  
529 rocks in the internal thrust-fold belt (Sierra Beauvoir), and their overprinting  
530 relationships with dikes intruding the Beauvoir Formation nearby Cerro Rodríguez  
531 (Cerro Rodríguez dikes, Fig. 9). The general deformation in this area has been  
532 described for the Upper Cretaceous rocks by Torres Carbonell et al. (2013b). Detailed  
533 descriptions of the deformation in older rocks have not been published so far, except  
534 from brief mentions in Torres Carbonell et al. (2017c). Further descriptions are  
535 included within the new data presented below.

536 Towards the lower stratigraphic horizons of Sierra Beauvoir (e.g. Beauvoir  
537 Formation), the rocks reveal a continuous cleavage formed mainly by very fine sericite.  
538 Folds across Sierra Beauvoir are involved in at least five orders, and in general they  
539 present NE vergence across the range. Wavelengths are of a few kilometers in the first  
540 order folds, and a few meters in the highest order (Fig. 8a). Competent beds form  
541 parallel folds, however multilayers can be classified as 1C or 2 Ramsay folds, since less-  
542 competent layers accommodate deformation forming class 3 folds between class 1B  
543 (parallel) competent folded beds. A wide spectrum in fold style is observed, however,  
544 with both angular and rounded hinges, and tightness varying between tight and open.

545 At least three thrusts have been recognized at Sierra Beauvoir, involved in the  
546 stacking of the Cretaceous sequence. The thrusts have NE vergence, and occasionally  
547 cut the frontal limb of the first order folds. Thrusts are characterized by shear zones  
548 several tens of meters thick, where a deflection of prior cleavage planes or bedding  
549 (forming folds and s-c fabrics), disrupted stratigraphy, and widespread cataclastic  
550 fabrics are observed (Fig. 8b). Shear sense is always top-to-NE. These characteristics

551 are similar to those described for thrusts affecting the same sequence in Chile (Klepeis,  
552 1994), and at Península Mitre (e.g. Bahía Thetis thrust in Torres Carbonell et al.,  
553 2017b).

554 Between Cerro Rodríguez and the Kranck pluton (Fig. 9) two distinct sets of dikes  
555 and sills crop out (Cerro Rodríguez dikes, first reported by Martinioni et al., 1999). The  
556 dikes intrude the Beauvoir Formation and have thicknesses of 0.2-2.5 m. The  
557 abundance of intrusions increases towards the Kranck pluton (Fig. 9). As briefly  
558 reported by Torres Carbonell et al. (2017c), we identified two dike generations. One of  
559 these generations is composed of clinopyroxene-biotite or clinopyroxene  
560 lamprophyres, affected by spaced pressure-solution cleavage parallel to the  
561 continuous cleavage in the host rock. These dikes are here called “deformed set” (Figs.  
562 9 and 10). The second generation has no cleavage and is composed of hornblende-  
563 clinopyroxene lamprophyres and felsic dikes of andesite and trachy-andesite  
564 (“undeformed set”). The deformed set is concentrated surrounding Cerro Rodríguez,  
565 whereas the undeformed set has a more uniform distribution, but is more abundant  
566 close to the Kranck pluton (Fig. 9). Martinioni et al. (1999) obtained a whole rock K/Ar  
567 age of  $104 \pm 4$  Ma for one undeformed, hornblende-phyric dike, which we discuss  
568 below. On the basis of hornblende chemical composition, Acevedo et al. (2007)  
569 correlated the hornblende undeformed dikes with hornblende gabbros from the  
570 Kranck pluton.

571

## 572 **5. Geochronological results**

573 Sample BR47 (Fig. 9) is a quartz monzonite of the Kranck pluton, which lacks  
574 penetrative deformation. Cross-cutting relationships indicate that it represents one of

575 the youngest facies of the pluton (a thorough description of the facies was presented  
576 by González Guillot et al., 2012). A total of 47 zircon grains were analyzed. The data set  
577 shows a constant  $^{206}\text{Pb}/^{238}\text{U}$  age between 64 and 72 Ma, with a weighted average at  
578  $68.01 \pm 0.52$  Ma ( $2\sigma$ , MSWD 11.8, Fig. 11a, Supplementary File 1). We interpret this as  
579 the age of emplacement and crystallization of the rock.

580 Sample BR29 (Fig. 9) is an undeformed hornblendite from the Kranck pluton. It is  
581 intruded by other facies, thus represents one of the earliest pulses of crystallization  
582 (see more details in González Guillot et al., 2012). The sample has a weighted mean  
583 age, integrated over the steps between arrows in Fig. 11b, of  $73.4 \pm 1.6$  Ma ( $2\sigma$ , MSWD  
584 0.81, Supplementary File 1). Sample 104, an undeformed hornblende lamprophyre of  
585 the Cerro Rodríguez dikes, has a weighted mean age, integrated over the steps  
586 between arrows in Fig. 11c, of  $74.0 \pm 0.8$  Ma ( $2\sigma$ , MSWD 0.79, Supplementary File 1).  
587 Both  $^{40}\text{Ar}/^{39}\text{Ar}$  ages are indistinguishable; therefore the undeformed hornblende  
588 lamprophyres of Cerro Rodríguez were emplaced at the same time than hornblendites  
589 in the Kranck pluton, and before the quartz monzonite.

590

## 591 **6. Integration of previous and new structural data from the Fuegian thrust-fold belt**

592 In the following section we integrate our new structural and geochronological data  
593 with previous work, which allows us to define structural generations and correlate  
594 them across the Fuegian Andes. Based on previous work (Torres Carbonell and Dimieri,  
595 2013), we refer to the Fuegian Andes as a thrust-fold belt that involves basement and  
596 thickens toward the south (hinterland). For convenience, we address separately the  
597 central belt from the internal thrust-fold belt.

598

599 **6.1. Central belt**

600 The central belt of the Fuegian Andes reveal multiple generations of structures with  
601 complex correlations between different regions. This is especially so in the highly  
602 metamorphosed and deformed area of Cordillera Darwin, where glacier cover adds  
603 difficulty in connecting structures from different coastal outcrops. However, two major  
604 deformations can be roughly defined in Cordillera Darwin, called D1<sub>CD</sub> and D2<sub>CD</sub>, each  
605 of them comprising one or more generations of structures. This distinction follows the  
606 criteria and data of Nelson et al. (1980), Cunningham (1995), Kohn et al. (1995), Klepeis  
607 et al. (2010), and Maloney et al. (2011), and correlates with the first and second  
608 deformation phases mentioned in section 2.2.

609 D1<sub>CD</sub> comprises foliations and folds formed during the obduction and  
610 underthrusting of the oceanic floor and continental margin, respectively, with  
611 evidence of a southward-dipping mid-crustal shear zone nearby Bahía Pía (Nelson et  
612 al., 1980; Cunningham, 1995; Klepeis et al., 2010). This process formed an initial S1  
613 metamorphic fabric and superposed S2 crenulation cleavage and F2 folds (Nelson et  
614 al., 1980; Kohn et al., 1993; Klepeis et al., 2010; Maloney et al., 2011), during what we  
615 interpret as a phase of progressive deformation with two generations of structures (S1  
616 + S2/F2). Away from the zone of higher metamorphic grade at Bahía Pía, D1<sub>CD</sub>  
617 structures are limited to a single generation of S1 structures associated to top-to-  
618 northeast obduction and ductile shear (Nelson et al., 1980; Klepeis et al., 2010).

619 The second deformation, D2<sub>CD</sub>, includes crenulation, kink bands, thrusts and folds  
620 that deform the prior D1<sub>CD</sub> structures. At Bahía Pía and in northern Cordillera Darwin  
621 D2<sub>CD</sub> structures have NE vergence and are coeval with retrograde metamorphism  
622 (Nelson et al., 1980; Kohn et al., 1993). West of Bahía Pía (at SW Cordillera Darwin)

623 D2<sub>CD</sub> structures are associated with backfolds and backthrust shear zones (Klepeis et  
624 al., 2010). D2<sub>CD</sub> developed during uplift of Cordillera Darwin, caused by at least three  
625 identified major thrusts (Garibaldi -blind-, Parry and Marinelli thrusts, Fig. 3), which  
626 together with the backfolds and backthrusts recognized in SW Cordillera Darwin form a  
627 doubly-vergent wedge (Nelson et al., 1980; Klepeis et al., 2010).

628 Formation of D1<sub>CD</sub> structures, as detailed in section 2.2, started before ~90 Ma  
629 (Kohn et al., 1995) and acted at least until ~73 Ma, coeval with the last intrusions of  
630 the Beagle suite (Fig. 4, Klepeis et al., 2010; Maloney et al., 2011; McAtamney et al.,  
631 2011). Thus, mid-crustal shearing during D1<sub>CD</sub> may be transitional with the beginning of  
632 thrusting during the following deformation (D2<sub>CD</sub>), which has ambiguous overprinting  
633 relationships with the Beagle suite (Klepeis et al., 2010).

634 The thrusting stage related to D2<sub>CD</sub> progressed through the Paleogene, as evidenced  
635 by thermochronology results detailed in section 2.2 (Fig. 4, Kohn et al., 1995; Gombosi  
636 et al., 2009). An appraisal of previous work, however, shows no clear connection with  
637 the structures towards the foreland, which have been interpreted as structures formed  
638 prior to out-of-sequence uplift of Cordillera Darwin (Klepeis, 1994; Kley et al., 1999;  
639 Klepeis et al., 2010).

640 In Argentine Tierra del Fuego, overprinting structural relationships from the central  
641 belt also allowed to define two main structural styles. An older, ductile deformation,  
642 which we will call D1<sub>CB</sub> (first deformation -D1- recorded in the central belt -CB-  
643 excluding Cordillera Darwin), and a younger, brittle-ductile deformation that we call  
644 D2<sub>CB</sub>. D1<sub>CB</sub> is characterized by ductile NE-vergent folds and axial-plane foliations  
645 associated with low-grade regional metamorphism, and with the development of shear  
646 zones with mylonitic foliations in sectors of localized higher strains (Fig. 8a, Kranck,

647 1932; Bruhn, 1979; Dalziel and Palmer, 1979; Torres Carbonell and Dimieri, 2013;  
648 Torres Carbonell et al., 2017b; Cao et al., 2018; Cao, 2019). Between Canal Beagle and  
649 Lago Fagnano, D1<sub>CB</sub> folds range from tight or isoclinal in the Lapataia and Lemaire  
650 formations, to tight or very tight folds with NE vergence, usually classes 1C to 2, in the  
651 Yahgan and Beauvoir formations. The axial-plane foliation is defined by low-grade  
652 metamorphic minerals, and occasionally transposes bedding (Fig. 8a). An exception  
653 occurs in the more competent coherent volcanic-subvolcanic bodies of all the  
654 formations of this region of the central belt, in which the folds and cleavage are not  
655 uniformly developed; some of these bodies show poor cleavage or none at all. The  
656 metamorphic grade in this part of the central belt is within prehnite-pumpellyite to  
657 greenschist facies (Bruhn, 1979; Cao et al., 2018; Cao, 2019).

658 In the zones of higher strain, especially localized in quartz-rich lithologies, the main  
659 structures are mylonitic fabrics with top-to-north or northeast shear sense (Fig. 8a).  
660 These mylonites, first recognized by Kranck (1932) and Bruhn (1979), led the latter  
661 author to interpret formation of the central belt between Canal Beagle and Lago  
662 Fagnano as a result of progressive simple shear. An alternative interpretation based on  
663 scattered and undetailed data suggested that these mylonites formed part of the D2<sub>CD</sub>  
664 thrusting of Cordillera Darwin (Torres Carbonell and Dimieri, 2013). However, as will be  
665 addressed below, the correlation of the D2<sub>CD</sub> brittle-ductile structures with the brittle-  
666 ductile thrusts reported in section 4.1.2, suggests that these mylonites are older.  
667 Therefore, they can be interpreted as zones of concentrated D1<sub>CB</sub> deformation in the  
668 simple shear scheme proposed by Bruhn (1979, cf. Torres Carbonell et al., 2017b; Cao  
669 et al., 2018).

670 In southern Península Mitre, the D1<sub>CB</sub> structures and metamorphic grades in the  
671 Lemaire and Beauvoir formations are similar to those between Lago Fagnano and  
672 Canal Beagle, with a NW-SE structural trend (Torres Carbonell et al., 2017b). Moreover,  
673 in this region it is clear how the intensity of the D1<sub>CB</sub> structures decrease from  
674 mylonitic foliations and transposition cleavage to spaced pressure-solution cleavage  
675 towards upper structural levels, affecting rocks as young as late Campanian-?early  
676 Maastrichtian (Bahía Thetis Formation, Figs. 2 and 8a, Torres Carbonell et al., 2013b,  
677 2017b). These NW-SE structures are included in the first deformation phase of Torres  
678 Carbonell et al. (2017b).

679 Previous constraints on the age of the D1<sub>CB</sub> structures were limited to their  
680 inclusion in the “main phase of deformation” (“mid” Cretaceous in Bruhn, 1979; Dalziel  
681 and Palmer, 1979), and the statement that they affected rocks not younger than the  
682 Campanian-?Maastrichtian Bahía Thetis Formation (Torres Carbonell et al., 2017b).  
683 González Guillot et al. (2018) established that these structures are older than the  
684 Ushuaia pluton (75-71 Ma), giving a stronger constraint for the southern portion of the  
685 central belt (Fig. 4). The same overprinting relationship is observed in other pluton of  
686 the Fuegian Potassic Magmatism, the Jeu-Jepén pluton, which has a contact  
687 metamorphism aureole that overprints cleavage in the host rocks (Fig. 5g).

688 The deformed dike depicted in Fig. 5d-f confirms that ductile structures (D1<sub>CB</sub>)  
689 affected the Ushuaia Peninsula Andesites. From correlation of that dike with the dated  
690 andesite at the main body of the intrusion ( $84.1 \pm 1.6$  Ma, González Guillot et al.,  
691 2018), we can establish an oldest time constraint on the age of these structures (Fig.  
692 4).

693 The younger deformation,  $D2_{CB}$ , includes large folds (up to tens of kilometers of  
694 wavelength) and crenulation cleavages that grade from a pervasive development in  
695 the lower structural levels (Lapataia Formation), to spaced crenulation cleavages  
696 localized in high-strain zones in the upper structural levels (Bruhn, 1979; Cao et al.,  
697 2018). These high-strain zones comprise thrusts and detachments (Fig. 8b), the former  
698 include the thrust zones of Cañadón Bianchi (section 4.1.2), as well as thrusts  
699 previously reported at the contact between the Lapataia and Lemaire formations  
700 (Beatriz thrust, Fig. 3, Cao et al., 2018; Cao, 2019), within the Lemaire Formation at  
701 Sierra Sorondo and Sierra Alvear (Cao, 2019), and between the Lemaire and Beauvoir  
702 formations at Montes Negros (Torres Carbonell et al., 2017b). The detachments are  
703 consistently located in the surroundings of the contact between the Lemaire and  
704 Yahgan or Beauvoir formations, as stated in section 4.1.2 (cf. Torres Carbonell and  
705 Dimieri, 2013; Cao, 2019). In Chile, the same detachment level was identified between  
706 the Tobífera and Río Jackson formations and has been called Río Jackson detachment  
707 by Klepeis (1994).

708 The  $D2_{CB}$  thrusts form characteristic zones of protracted deformation during  
709 decreasing temperature conditions. Accordingly, while crenulation cleavages and  
710 deflected foliations formed in the lower structural levels, these structures were  
711 overprinted by cataclastic fabrics in shallower levels. For example, as shown in Fig. 6f,  
712 ductile  $D2_{CB}$  folds affecting the older  $D1_{CB}$  transposition foliation are faulted and  
713 incorporated in a brittle  $D2_{CB}$  cataclasite. In the more brittle settings and especially  
714 near detachment horizons, the crenulations affecting previous  $D1_{CB}$  foliations are  
715 localized within a few meters from the detachment surfaces or shear zones, and  
716 dissipate away from the fault surfaces (Fig. 8b). Our data from Cañadón Bianchi and



717 Arroyo Velazquito (Figs. 6 and 7) confirm the superposition of these thrusts and  
718 detachment on the prior ductile D1<sub>CB</sub> deformation, thus constraining D2<sub>CB</sub> to a post-  
719 Campanian age (Fig. 4).

720 A third generation of structures in the Argentine central belt has been described by  
721 Cao (2019) and Cao et al. (2018), characterized by small (cm-scale) folds, kink bands  
722 and shear zones of deflected D1<sub>CB</sub> cleavage. These structures consistently have a N-S  
723 orientation, which is almost perpendicular to the regional structural trend. They have  
724 been interpreted as a distinct phase by the cited authors, but it is also possible that  
725 they are cross-folds or similar structures (see for example Butler, 1982) formed during  
726 the same D2<sub>CB</sub> deformation.

727

## 728 **6.2. Internal thrust-fold belt**

729 The connection between the central belt and the internal thrust-fold belt has been  
730 described north of Seno Almirantazgo and Lago Fagnano, in Chile (Klepeis, 1994; Kley  
731 et al., 1999; Rojas and Mpodozis, 2006). This connection, as mentioned in the previous  
732 section, involves a detachment at the base of the Lower Cretaceous shaly sequence,  
733 mapped as the Río Jackson detachment by Klepeis (1994) and reported in Argentina in  
734 this work (Fig. 7). Previous work by Klepeis (1994) and Torres Carbonell and Dimieri  
735 (2013), defined this detachment as the base of the cover sequence deformed in the  
736 foreland thrust-fold belt (Fig. 8b).

737 Between Lago Blanco and Lago Fagnano, the deformation sequence comprises  
738 folding and axial plane, pressure-solution cleavage development (S1 in Klepeis, 1994)  
739 and refolding of these structures by open folds (F2). Both generations developed,  
740 according to Klepeis (1994), during progressive shortening associated with thrust

741 emplacement in the Cretaceous succession. Thrusts in this region are recognized as  
742 brittle-ductile structures such as s-c type, and brittle cataclastic fabrics. Klepeis (1994)  
743 also interpreted two backthrusts, associated with deformation just above the Río  
744 Jackson detachment.

745 At Sierra Beauvoir, the Upper Cretaceous rocks are deformed by folds and an axial-  
746 plane foliation, both with variable degrees of development (Torres Carbonell et al.,  
747 2013b). The foliation varies from smooth disjunctive cleavage in the lower structural  
748 levels or high-strain zones, to a more spaced cleavage ( $> 0,05$  mm) formed by  
749 pressure-solution in the uppermost part of the succession. In the Policarpo Formation,  
750 for example, the cleavage is absent or very roughly developed and usually it is only a  
751 pencil structure (Torres Carbonell et al., 2013b). Our new data from the internal thrust-  
752 fold belt (section 4.2) complements these previous descriptions from Sierra Beauvoir,  
753 and show the change to more pronounced cleavages and very low-grade  
754 metamorphism in the older units. We recognized no backfolds or backthrusts,  
755 however, which could have been correlated with similar structures described in Chile,  
756 just tens of kilometers west of our study area (cf. Klepeis, 1994).

757 At Península Mitre (Bahía Thetis), structures similar to those described at Sierra  
758 Beauvoir are included in the second deformation phase described by Torres Carbonell  
759 et al. (2017b). These structures comprise NE-SW oriented folds, brittle-ductile shear  
760 zones (e.g. Bahía Thetis thrust and Buen Suceso backthrust) and SE-dipping pressure-  
761 solution cleavages in the Policarpo Formation and older Upper Cretaceous rocks, which  
762 grade to crenulation cleavages in the Beauvoir and Lemaire formations further south.  
763 No significant growth of metamorphic minerals is associated with the formation of this  
764 crenulation. The thrusts are characterized by zones several hundred meters thick,

765 integrated by shear fabrics superposed on the prior D1<sub>CB</sub> foliation. These include s-c  
766 type fabrics and disrupted strata, evidencing deformation in brittle-ductile conditions  
767 (Torres Carbonell et al., 2017b).

768 Both at Península Mitre and Sierra Beauvoir, the youngest spaced cleavages affect  
769 the Policarpo Formation. The Paleocene Río Claro Group, resting on an unconformity  
770 above the Policarpo Formation, lacks penetrative deformation (Torres Carbonell et al.,  
771 2013b). This led to define a deformation phase of approximately Danian age (D1 in  
772 Torres Carbonell et al., 2017a), which in Península Mitre cross-cuts the prior D1<sub>CB</sub>  
773 structures with a highly oblique trend (NE-SW against NW-SE). Here we call this  
774 deformation D1<sub>FB</sub>, namely the first deformation distinct to the foreland thrust-fold belt  
775 (FB, Fig. 4).

776 At Sierra Beauvoir, where both D1<sub>CB</sub> and D1<sub>FB</sub> are coaxial (NW-SE trend), their  
777 distinction is aided with our new geochronological data (section 5). The age of sample  
778 104 is significantly younger than the age obtained for a similar undeformed dike by  
779 Martinioni et al. (1999) using K/Ar in whole rock. We consider our age better  
780 constrained, whereas previous work in the Fuegian Andes has shown the significant  
781 errors associated with K/Ar dating (González Guillot et al., 2018). Therefore, our new  
782 age allows us to place the formation of foliations in the deformed set of the Cerro  
783 Rodriguez dikes and host rock before ~74 Ma, placing these structures in the D1<sub>CB</sub>  
784 deformation. In this manner, younger spaced cleavages affecting rocks as young as the  
785 Policarpo Formation, mentioned before, can be attributed to D1<sub>FB</sub> consistently with  
786 the assignment made in Península Mitre (cf. Torres Carbonell et al., 2017b). Younger  
787 deformations in the foreland (Df2-Df6, mentioned in section 2.2), which are not  
788 discussed in this work, are grouped within a single deformation D2<sub>FB</sub> (Fig. 4).

## 789 **7. Protracted deformation history of the Fuegian thrust-fold belt**

790 The combination of published and new data just presented allows us to integrate  
791 the central belt and the foreland thrust-fold belt of the Fuegian Andes in a protracted  
792 deformation history, from the Late Cretaceous to the early Neogene. The evidence for  
793 structural connections between the central belt and the internal thrust-fold belt, in  
794 turn linked with the external thrust-fold belt, permits to analyze these portions of the  
795 orogenic front in a coherent manner, using as an example the restored cross-section  
796 presented below.

797 We propose a simplified, regional deformation sequence in which distinct structural  
798 styles develop simultaneously at different portions of the orogenic belt. Accordingly,  
799 the phases and stages of deformation mentioned throughout this text, which group  
800 generations of structures in different sectors of the thrust-belt, are here included in  
801 two major events forming this protracted history. The first event (red double-arrows in  
802 Fig. 4) involved the closure of the back-arc basin leading to obduction, underthrusting,  
803 and arc-continent collision (cf. Nelson et al., 1980; Dalziel, 1986; Cunningham, 1995;  
804 Klepeis et al., 2010), during which the orogeny was initiated. The second event (black  
805 double-arrows in Fig. 4) involved major uplift of the orogenic core (central belt) leading  
806 to formation of a first-order thrust system (antiformal stack) that transferred  
807 shortening towards the foreland, expanding the thrust-fold belt into the coeval  
808 foreland basin system (cf. Torres Carbonell and Dimieri, 2013).

809

### 810 **7.1. First event: Arc-continent collision**

811 Our new data on the overprinting relationships between the Ushuaia Peninsula  
812 Andesites and the D1<sub>CB</sub> deformation (Fig. 5d-f) indicates an age not older than ~84 Ma

813 for the first event in the southern central belt in Argentina (Fig. 4). The youngest age of  
814  $D1_{CB}$  in the same area has already been established by the intrusion of the Ushuaia  
815 pluton by 75-71 Ma (González Guillot et al., 2018), which is consistent with the  
816 Campanian age determined in Península Mitre (first phase in Torres Carbonell et al.,  
817 2017b). At the northern border of the central belt, the ~72 Ma Jeu-Jepén pluton  
818 (Cerredo et al., 2011) also overprints the  $D1_{CB}$  foliation (Fig. 5g).

819 Even if these temporal constraints are consistent with the correlation of  $D1_{CB}$  and  
820  $D1_{CD}$ , the beginning of ductile deformation in the Argentine part seems to have been  
821 delayed in comparison with the development of deep-seated structures in Cordillera  
822 Darwin, which started before 90 Ma (Fig. 4). In addition, during the end of the  
823 collisional event (by ~73 Ma), deformation in Cordillera Darwin occurred under high-  
824 temperature conditions, coincident with decompression from 12 to 9 kbar and  
825 initiation of thrust-related uplift (Maloney et al., 2011). Conversely, the coeval  
826 deformation between Canal Beagle and Lago Fagnano was milder, especially in the  
827 upper stratigraphic levels (Yahgan-Beauvoir formations). This is notorious within and  
828 around some competent rock bodies that developed poor cleavage or were almost  
829 undeformed (Fig. 5a-d). The more deformed rocks in this region, indeed, are  
830 represented by low-grade mylonitic zones in the Lapataia and Lemaire formations (Fig.  
831 8a). This is consistent with a change from high- to low-grade deformation across the  
832 orogen, i.e. from the internal and lower structural levels now exposed at Cordillera  
833 Darwin to the external, upper structural levels exposed between Lago Fagnano and  
834 Canal Beagle. Farther towards the Campanian deformation front, in the internal thrust-  
835 fold belt, only very low-grade deformation affects the Beauvoir Formation.

836 The data presented here in section 4.2 also contribute to constrain the age of the  
837 collisional event in the internal thrust-fold belt. Accordingly, the age of the  
838 undeformed set of the Cerro Rodríguez dikes constrains the age of collisional  
839 deformation in that region to before  $\sim 74$  Ma (Fig. 11), which is consistent with the ages  
840 in the central belt.

841 In summary, the progression of deformation during the arc-continent collision event  
842 can be described as foreland directed, ductile simple shear deformation above the  
843 underthrusting continental plate, concurring with Bruhn (1979, see also Tanner and  
844 Macdonald, 1982; Storey, 1983, Fig. 8a). Whilst high-grade deformation started prior  
845 to 90 Ma in Cordillera Darwin ( $D1_{CD}$ ), in the Argentine Fuegian Andes only the last,  
846 Campanian part of this deformation is recorded in the rock structures, comprising low-  
847 grade deformation ( $D1_{CB}$ ).

848

## 849 **7.2. Second event: Major thrusting and expansion of the orogenic wedge**

850 Our new data from the central belt contributes to the definition of the second event  
851 during the orogenic history, which involved thrust-related deformation throughout the  
852 orogen. We presented evidence from Cañadón Bianchi (Fig. 6) which confirms part of  
853 the thrust faults forming the duplex proposed by Torres Carbonell and Dimieri (2013).  
854 This new evidence adds to previous data reported by Torres Carbonell et al. (2017b)  
855 and Cao et al. (2018). These thrusts, ramping from an unexposed lower detachment  
856 (possibly located in the basement unit), cut through Jurassic syn-rift deposits and join  
857 an upper detachment in the vicinity of the contact between the Lemaire and Yahgan or  
858 Beauvoir formations (Figs. 6 and 7). The brittle-ductile thrusts and detachment are

859 superposed on previous ductile D1<sub>CB</sub> structures; therefore they post-date the  
860 Campanian (Fig. 4).

861 In a similar way, the D1<sub>CD</sub> structures are overprinted by D2<sub>CD</sub> structures in Cordillera  
862 Darwin, which are related to ductile-brittle thrusts emplaced in that region. The  
863 northernmost of these thrusts, called Marinelli thrust (Klepeis, 1994), can be laterally  
864 traced to join the Beatriz thrust in our map (Fig. 3). Klepeis (1994) suggested that the  
865 Marinelli thrust exhumed and placed basement rocks on top of back-arc basin cover  
866 rocks, proposing that the thrust timing was out-of-sequence with deformation in the  
867 internal thrust-fold belt. Building on that interpretation, it has been proposed that all  
868 the D2<sub>CD</sub> thrusts (Garibaldi, Parry and Marinelli) formed out-of-sequence with the  
869 internal thrust-fold belt (Kraemer, 2003; Rojas and Mpodozis, 2006; Klepeis et al.,  
870 2010). This has important implications for the transference of shortening (or lack of it)  
871 that built the foreland portion of the thrust-fold belt.

872 An alternative view has been proposed by Torres Carbonell and Dimieri (2013),  
873 interpreting the Marinelli thrust as a ramp which, after being actively involved in the  
874 duplex, broke up-section as a breaching thrust (cf. Butler et al., 2007), cutting the  
875 duplex roof detachment. This interpretation is more adequate to allow additional  
876 structures in the hinterland to transfer shortening to the foreland, before being  
877 effectively uplifted above the roof thrust (see also Torres Carbonell et al., 2017a).

878 Our interpretation for the progression of D2<sub>CB</sub> structures is shown in a balanced  
879 cross-section constrained by field and seismic data (Fig. 12). This cross-section uses a  
880 depth to the basal detachment controlled by depth-migrated seismic data and wells in  
881 the foreland, and with an arbitrary uniform dip towards the hinterland. Notice that the

882 basement-Upper Jurassic contact in the cross-section is speculative. Further details on  
883 the balancing method are given in section 3.

884 The cross-section assumes that all the deformation during  $D2_{CB}$  was accommodated  
885 by movement on single thrusts. This clearly is an oversimplification, but allows creating  
886 a geologically reasonable and viable cross-section that explains the first-order central  
887 belt structure and coeval foreland deformation. It also implies amounts of shortening  
888 transferred to the foreland that are consistent with balanced cross-sections in the  
889 external thrust-fold belt (Torres Carbonell et al., 2017a).

890 In summary, this regional, balanced cross-section shows that emplacement of  
891 brittle-ductile thrusts in the central belt during  $D2_{CD}$  and  $D2_{CB}$  progressed  
892 simultaneously with accommodation of transferred shortening in shallower structures  
893 in the thrust-fold belt ( $D1_{FB}$  and  $D2_{FB}$ , Fig. 8b). The latter argument is consistent with  
894 the cooling history of the central belt, as reported in section 2.2, which indicates  
895 coeval uplift of the central belt and deformation in the foreland at least during the Late  
896 Cretaceous and Paleogene (Fig. 4, Kohn et al., 1995; Gombosi et al., 2009).

897 The interpretation shown in Fig. 12, with a crustal duplex responsible for uplift of  
898 the central belt after the initial closure of the Rocas Verdes Basin, also implies the  
899 existence of basement thrust wedges below the central belt as a requisite to explain  
900 the structural level at which the Lemaire Formation is uplifted, in comparison with the  
901 same horizons in the foreland (Fig. 12). This interpretation has the advantage of  
902 explaining the occurrence of rocks uplifted from depths of more than 20 km (12 kbar  
903 and  $\sim 600$  °C according to Kohn et al. 1995) at the current surface (at Bahía Pía), using a  
904 reasonable geometric model for this amount of uplift.



905 Conditions of deformation during the thrusting event varied also with structural  
906 position. In the central belt the metamorphic grade was low in Cordillera Darwin (Kohn  
907 et al., 1993), to very low or even absent in Argentina, and penetrative deformation was  
908 restricted to crenulation cleavages with minor growth of retrograde metamorphic  
909 minerals associated with the formation of major brittle-ductile thrusts (Fig. 8b).  
910 Towards the foreland, shortening transferred from the duplex to the Cretaceous cover  
911 was accommodated by pressure-solution and spaced cleavage development, followed  
912 by folding and brittle-ductile thrusting (Fig. 8b). This deformation comprises, in the  
913 scheme presented here, the  $D1_{FB}$  structures now exposed in the internal thrust-fold  
914 belt.

915 Simultaneous sedimentation and deformation in the Austral basin characterized the  
916 following deformation in the external part of the thrust-fold belt, called  $D2_{FB}$  in this  
917 work. This deformation is mostly related to the development of major thrust systems  
918 in the foreland until the early Miocene (Fig. 12, see Torres Carbonell et al., 2017a).

919

## 920 **8. Conclusions**

921 An integration of new structural and geochronological data with an analysis of  
922 previously published data from the Fuegian Andes, allows us to document the  
923 sequence of deformation within a tectonic model of the orogen formed in the  
924 southernmost Andes. Accordingly, we constrain the timing of the two main events  
925 involved in deformation in the Argentine portion of the orogen. The first event was  
926 caused by closure of the Rocas Verdes Basin, with obduction, underthrusting, and arc-  
927 continent collision. As previously documented from Cordillera Darwin, this event  
928 spanned from ca. 100-90 Ma to ca. 73 Ma ago, causing high-grade deformation (e.g.

929 Kohn et al., 1995; Maloney et al., 2011). However, in the central belt in Argentina, our  
930 new data reveal that only Campanian (post-84 Ma), low-grade structures were  
931 developed in the uppermost structural levels of the collision-related deformation. This  
932 deformation is also recorded in the internal thrust-fold belt, and we provide a new age  
933 constraint for the structures in that region, which are post-dated by intrusives with  
934 Ar/Ar hornblende ages of 74 Ma.

935 The second event was characterized by the formation of a first-order thrust system  
936 since the Maastrichtian-Danian. This led to the definitive formation and expansion of  
937 the thrust-fold belt until the early Miocene, with major uplift episodes recorded in the  
938 hinterland associated with the emplacement of major basement-involved thrust  
939 sheets. Our new structural data confirms the occurrence of these thrusts, and their  
940 connection with a common upper detachment placed at the base of the Lower  
941 Cretaceous rocks, which forms the roof thrust of a major duplex in the central belt.

942 This scheme explains the different styles of deformation across the Fuegian Andes  
943 in the context of a progressive evolution from collisional-style orogenesis to thrust-  
944 related deformation. This model also allows explaining the shortening in the deformed  
945 Austral-Magallanes foreland basin as a consequence of major thrust emplacement in  
946 the central belt. Our new structural and geochronological data, in addition, give key  
947 constraints to correlate structure generations across and along the strike of the  
948 orogen.

949

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962

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1242

1243 **Figure captions**

1244 **Figure 1: a.** Geologic map of southernmost Patagonia and Tierra del Fuego with  
1245 regional structural elements discussed in this work. S: Sarmiento Complex, CA: Capitán  
1246 Aracena Complex. T: Tortuga Complex, LF: Lago Fagnano, CB: Canal Beagle. Based on  
1247 works cited in the text. **b.** Cartoon showing closure of the back-arc basin and  
1248 development of the Fuegian Andes during two main tectonic events, involving collision  
1249 and later thrust-fold belt (TFB) expansion. Known intrusive pulses during both events  
1250 are schematically depicted (modified from Torres Carbonell et al., 2014).

1251

1252 **Figure 2:** Stratigraphic framework of the Fuegian thrust-fold belt north of Canal Beagle,  
1253 with nomenclature used in Argentina and Chile. For a more complete regional  
1254 correlation see Torres Carbonell and Olivero (2019). References are 1: Biddle et al.  
1255 (1986), Cañón (2000), Sánchez et al. (2010), McAtamney et al. (2011), Malumián et al.

1256 (2013), 2: Malumián and Olivero (2006), Olivero and Malumián (2008), Martinioni et al.  
1257 (2013), Cao et al. (2018), 3: Klepeis et al. (2010), McAtamney et al. (2011), Cerredo et  
1258 al. (2011), González Guillot et al. (2018). An idealized lithologic column is shown,  
1259 although lateral facies variations exist. Minimum thicknesses reported for reference  
1260 are from the thicker composite sections and in some cases highly variable.

1261

1262 **Figure 3:** Geologic map of the Fuegian thrust-fold belt with main structural features  
1263 and stratigraphic units mentioned in the text. See location in Fig. 1a, and text for  
1264 details of stratigraphic units. Compiled from Klepeis (1994), SERNAGEOMIN (2003),  
1265 Olivero and Malumián (2008), Hervé et al. (2010b), Klepeis et al. (2010), González  
1266 Guillot (2016), Cao (2019), Torres Carbonell and Olivero (2019), and authors' data.  
1267 Cross-sections x-y and y-z indicate location of Fig. 12a. Pt: Parry thrust, MM: Montes  
1268 Martial, UP: Ushuaia pluton, UPA: Ushuaia Peninsula Andesites, JP: Jeu-Jepén pluton,  
1269 MP: Moat pluton, BP: Bahía Pía, BT: Bahía Thetis.

1270

1271 **Figure 4:** Time constraints for the deformations recorded in the Fuegian thrust-fold  
1272 belt, north of Canal Beagle, as addressed in this work. Modified from Torres Carbonell  
1273 and Dimieri (2013). Bold bars indicate  $2\sigma$  uncertainties. UPA: Ushuaia Peninsula  
1274 Andesites, FPM: Fuegian Potassic Magmatism. References: 1: Nelson (1982), 2: Kohn et  
1275 al. (1995), 3: Gombosi et al. (2009), 4: Klepeis et al. (2010), 5: Olivero and Martinioni  
1276 (1996b), 6: Nelson et al. (1980), 7: Olivero et al. (2003), 8: McAtamney et al. (2011), 9:  
1277 Torres Carbonell et al. (2011), 10: Maloney et al. (2011), 11: Cerredo et al. (2011), 12:  
1278 Barbeau et al. (2009b), 13: González Guillot et al. (2018), 14: Torres Carbonell et al.



1279 (2017b), 15: Olivero et al. (1998), 16: this work. Red and black double arrows labeled  
1280 "D" are discussed in sections 6 and 7.

1281

1282 **Figure 5: a-d.** Outcrop photographs of the Ushuaia Peninsula Andesites (UPA) and host  
1283 rock (Yahgan Formation, YG). Notice the lack of intense deformation in pictures **a**  
1284 (person for scale) and **b**, from the main body of the intrusion. In picture **c** the Yahgan  
1285 Formation shows slight folding and cleavage away from the main body of the intrusion  
1286 (scale is 6 cm long). Picture **d** shows a dike and its host rock affected by pressure-  
1287 solution cleavage (parallel to compass), farther away from the main intrusion. **e-f.**  
1288 Photomicrographs of cleavage in the dike of picture **d**, plane-polarized light. Note  
1289 horizontal pressure-solution seams with wiggly morphology defined by opaque  
1290 insoluble residue. Pl: plagioclase, Hbl: hornblende (chloritized), Aln: allanite. g.  
1291 Hornfels from the Jeu-Jepén pluton (Beauvoir Formation) with scarce pressure-solution  
1292 seams (from upper-right to lower-left) reflecting the prior cleavage, which is  
1293 obliterated by recrystallization of the plagioclase + quartz (Qz) groundmass,  
1294 dissemination of insoluble residue inherited from the pressure-solution seams, and  
1295 growth of biotite (Bt) parallel to the preexisting cleavage planes.

1296

1297 **Figure 6: a.** Geologic map of the Cañadón Bianchi area (location in Fig. 3). Equal area,  
1298 lower hemisphere projection shows orientation of crenulation folds in the area. Brittle-  
1299 ductile shear zones (BDZS) are numbered for reference to picture **b**. Location of  
1300 photographs **c-j** is shown. **b.** Panoramic photograph with depiction of major structures  
1301 as shown in **a**. Acronyms are for stratigraphic units in the map. **c-d.** Outcrop (backpack  
1302 for scale) and close-up view of cataclastic fabrics in thrust zones. **e.** S-C type fabric

1303 showing top-to-north component of movement. **f.** Cataclasite block revealing faulted  
1304 folds (horizontal trace) affecting an older transposition foliation (alternating lighter  
1305 and darker bands). Encircled 23-mm coin for scale. **g.** Photomicrograph of a cataclasite  
1306 from a thrust zone affecting older quartz-mica schist, cross-polarized light. **h.** Faulted  
1307 folds in the detachment at the base of the Yahgan Formation. Notice the style of  
1308 folding with thickened hinge and thinned limbs in the lower left of the photograph. **i.**  
1309 Crenulation of a previous foliation in the upper flat of the BDSZ 1. Scale is 6 cm long. **j.**  
1310 S-C type fabric showing northeastward component of displacement in the detachment  
1311 zone. **k.** Cross-section (located in **a**) showing the first-order synform affecting the  
1312 detachment.

1313

1314 **Figure 7: a.** Panoramic photograph of the structure at Arroyo Velazquito. The  
1315 detachment near the base of the Beauvoir Formation is folded by a large antiform.  
1316 Towards the north, the beds dip northwards with progressively shallower angles  
1317 (values given in dip direction/dip). **b-d.** Details of brittle-ductile fault fabrics in the  
1318 detachment zone (located in **a**), superposed on older foliations and folds. In **b**, s-c type  
1319 fabrics show a component of northward displacement (notice that north points to the  
1320 lower left). 6-cm scale highlighted. Brecciated folds (arrows) and cataclastic fabric are  
1321 shown in **c** and **d**.

1322

1323 **Figure 8:** Schematic description and examples of  $D1_{CB}$  structures (**a**) and  $D2_{CB}$ - $D1_{FB}$   
1324 structures (**b**), further explained in section 6. S0: bedding, J: Jurassic and older, Kl:  
1325 Lower Cretaceous, Ku: Upper Cretaceous. Cross-sections are idealized and not to scale.

1326

1327 **Figure 9:** Map of the Kranck pluton and Cerro Rodríguez dikes, with distribution of  
1328 deformed and undeformed dike sets mentioned in the text. Location of dated samples  
1329 is shown. Equal area projection shows poles to cleavage in the Beauvoir Formation  
1330 (black) and the deformed dike set (red), with n number of measurements.

1331

1332 **Figure 10: a-c.** Outcrop examples of deformed Cerro Rodríguez dikes. Notice the good  
1333 development of pressure-solution cleavages parallel to the cleavage in the host rock.  
1334 **d-f.** Photomicrographs (plane-polarized light) showing the stylolitic or wiggly pressure-  
1335 solution seams defining the spaced cleavage. Phenocrysts are very altered  
1336 pseudomorphs of plagioclase, clinopyroxene or hornblende. In picture **f** a detail of  
1337 bended biotite is shown, with the axial planes of the microfolds subparallel to cleavage  
1338 (subvertical).

1339

1340 **Figure 11: a.** Wetherill concordia plot for zircon age determination from a quartz  
1341 monzonite of the Kranck pluton (sample BR47). **b-c.** Hornblende (Hbl) Ar/Ar plateau  
1342 ages for **(b)** sample BR29 (hornblendite, Kranck pluton) and **(c)** sample 104 (Hbl  
1343 lamprophyre, Cerro Rodríguez dikes). The weighted mean age was integrated over the  
1344 steps between the arrows.

1345

1346 **Figure 12: a.** Balanced cross-section of the Fuegian thrust-fold belt. The cross-section  
1347 was constructed along an assumed continuous trace x-z (see location in Fig. 3),  
1348 considering that the segment y-z was located behind x-y before the c. 50 km left-  
1349 lateral strike offset produced by the Neogene Fagnano transform fault system (cf.  
1350 Torres Carbonell et al., 2014). AMTS: Apen-Malvinera thrust system. **b.** Sequential

- 1351 forward modelling of the interpreted structures shown in **a**, using Move-on-Fault  
1352 algorithms in MOVE software. Eroded spaces are left blank for simplicity.

Journal Pre-proof

Arc-continent collision related orogenesis caused high-grade to mild deformation

New structural data and Ar/Ar - U/Pb ages of intrusives is reported

First deformation in the central belt of Argentina occurred from 84 to 74 Ma

Subsequent deformation involved a brittle-ductile to brittle thrust system

These thrusts join a common upper detachment transferring shortening to foreland

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**Declaration of interests**

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

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