

1 **Slip rate of trench-parallel normal faulting along the Mejillones Fault (Atacama Fault**
2 **System): Relationships with the northern Chile subduction and implications for seismic**
3 **hazards**

4
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32

33 **ABSTRACT**

34 The recent tectonics of the arid northern Chile Andean western forearc is characterized by
35 trench-parallel normal faults within the Atacama Fault System (AFS). Since the 1995-Mw8.1
36 Antofagasta earthquake, the mechanism driving this recent and localized extension is considered
37 to be associated with the seismic cycle within the subduction zone. Analyzing morphotectonic
38 patterns along these faults allows examining the seismic potential associated with the subduction
39 zone. Using field Digital Elevation Models and in situ-produced cosmogenic ^{10}Be , we determined
40 a 0.2 mm.yr^{-1} long-term vertical slip rate along the Mejillones fault, one of the most prominent
41 structures within the AFS. This result suggests that the AFS corresponds to slow slip rate faults
42 despite the rapid subduction context. However, the size of coseismic slips observed along the
43 AFS faults suggests that larger subduction earthquakes ($M_w > 8.1$) may occur episodically in the
44 area.

45

46 1. INTRODUCTION

47 The northern part of the arid Chile Andean forearc is known to have undergone major interplate
48 underthrusting seismic events over the past 150 years (e.g. Comte and Suárez, 1995; Madariaga
49 et al., 2010) (Figure 1A). It is described as a strongly coupled plate boundary between oceanic
50 and continental crusts, with a strong interseismic coupling (i.e. Madariaga et al., 2010; Chlieh et
51 al., 2011). Since the 1877-M8.6 mega earthquake, only four events of lesser magnitude occurred,
52 in 1995 (Mw 8.1 Antofagasta), in 2007 (Mw 7.7, Tocopilla), and in 2014 (Mw 7.6 and Mw 8.1,
53 Iquique), and a large seismic moment deficit remains in this region (e.g. Ruegg et al., 1996;
54 Delouis et al., 1998; 2009; Peyrat et al., 2010; Béjar-Pizarro et al., 2010, Ruiz et al., 2014)
55 (Figure 1A).

56 The whole northern Chile region is also characterized by a set of N-S sub-parallel faults
57 that stand out clearly in the morphology (Figure 1B). Called the Atacama Fault System (AFS)
58 (Arabasz, 1971), these eastward dipping faults, inherited from the Mesozoic tectonics (e.g.,
59 Brown et al., 1993; Scheuber and González, 1999; Cembrano et al., 2005) show evidences of
60 recent activity (e.g., Okada, 1971; Armijo and Thiele, 1990; Gonzales and Carrizo, 2003;
61 Gonzales and Carrizo, 2003, 2006; Vargas et al., 2011; Cortés et al., 2012). Armijo and Thiele
62 (1990) started describing the Atacama fault as a left-lateral strike-slip fault, and invoked a local
63 consequence of an E-W extension affecting the coast line (i.e. Coastal Scarp).. But, the following
64 works showed that the Atacama fault system, including the Mejillones peninsula, corresponded to
65 a system of normal faults (Delouis et al., 1998; Gonzales and Carrizo, 2003; Gonzales et al., 2003;
66 2006; Marquardt, 2005; Vargas et al., 2011; Cortés et al., 2012; Cortés-Aranda et al., 2015).

67 Observing fresh normal displacement along the Paposo fault that did not exist before the
68 1995 Antofagasta earthquake (see location “Fig.8A” in Figure 1C), Delouis et al. (1998)
69 proposed that the Atacama fault system was strongly coupled with the seismic cycle along the

70 subduction zone, with normal faulting events contemporaneous with strong co-seismic
71 subduction events. The same kind of observations were made during the M_w 8.8, 2010 Maule
72 earthquake near Santiago, with the occurrence of large aftershocks nucleated on normal faults in
73 the upper plate (Aron et al., 2013).

74 This interpretation is consistent with the geodetic data recorded during the 1995
75 Antofagasta earthquake that evidence a rapid increase of the horizontal coseismic displacement
76 towards the trench when moving from points located inside the forearc to points located near the
77 coast (Ruegg *et al.* 1996; Klotz *et al.* 1999). On the contrary, the horizontal displacement of the
78 forearc during the interseismic period is characterized by vectors directed landward that decrease
79 progressively from west to east (Chlieh et al., 2004). Analysing InSAR data, Loveless and Prichard
80 (2008) obtained ambiguous results with some images showing offset consistent with coseismic
81 faulting on the Paposó segment, while others lack such signal.

82
83 For Gonzales and Carrizzo (2003) and Gonzales et al. (2003, 2006) the normal faulting process
84 along the AFS is associated with a long-term extensional strain due to the buckling of the margin
85 with no direct co-seismic coupling with the subduction events.

86 In Delouis et al. (1998)'s interpretation, large, shallow underthrusting events at the plate
87 interface induce an increment of \sim E-W extension within the overlying upper crust, triggering
88 ruptures along the Atacama normal fault system. According to Delouis et al. (1998), this
89 increment of extension is associated with an increase of the Coulomb stress for normal failure on
90 the AFS. A comparison of the distributions of coseismic and interseismic horizontal
91 displacements on the overriding plate (Ruegg et al., 1996; Klotz et al., 1999; Chlieh, 2004)
92 indicates that coseismic extension is concentrated on the coastal area whereas interseismic
93 contraction is more broadly distributed in the whole forearc area and beyond towards the East.

94 This may explain why extension across the AFS is not cancelled at the end of a seismic cycle,
95 and may slowly increase from one cycle to the other.

96 Delouis et al. (1998)'s interpretation has three implications: 1) each complete seismic
97 cycle at the subduction interface produces a net increase of extension in the coastal area,
98 explaining the long-lasting activity and persistent morphological signature of the normal faults;
99 2) during the interseismic period, extension in the coastal area is only partially reduced, which
100 implies a partial reduction of the Coulomb Failure Stress (CFS) on the AFS. This is supported by
101 the absence of present-day seismicity along the AFS; and 3) movements along the AFS are
102 expected to coincide with the abrupt increase of the CFS on the normal faults, i.e. when large
103 subduction events occur. As confirmed by Aron et al. (2013), the static deformation field
104 produced by a great subduction earthquake is an effective mechanism for generating permanent
105 extension above the seismogenic zone, reactivating suitably oriented, long-lived normal faults. In
106 the case of the Maule earthquake, Aron et al. (2013) calculated that the earthquake produced a
107 radial pattern of static extension and deviatoric tension, with a CFS increment larger than 1.5
108 MPa, along the Coastal Cordillera enclosing the rupture area.

109 Cortés-Aranda et al. (2015) explored further the idea of a CFS link between the
110 subduction, the Mejillones fault system (MFS) within the Mejillones Peninsula, and the rest of
111 the Atacama fault system further east, with a more thorough modelling of the CFS associated to
112 both the co-seismic and inter-seismic periods. In particular, they show that the lower seismic
113 coupling below the Mejillones peninsula (Métois et al., 2013) may even be associated with a
114 positive increase of CFS on the MFS in the interseismic stage. They hence confirm that the MFS
115 activation can be explained in terms of CFS changes generated by slip on the plate interface.

116 In the present study, we do not discuss the role and origin of the west-dipping coastal
117 scarp, which could act differently than the AFS (Armijo & Thiele, 1990; Contreras-Reyes et al.,

118 2012). We consider, as in Delouis et al. (1998) and Cortés et al. (2015), that the MFS is part of
119 the AFS and respond to the same kind of CFS mechanical interaction with the subduction plate
120 interface.

121 If the mechanism driving this trench parallel normal faulting is associated with the
122 seismic cycle within the northern Chile subduction zone, analyzing geomorphology along these
123 faults may provide information on the seismic potential associated with the subduction zone. In
124 this paper, we present a study carried out in February 1999 and April 2002 (Marquardt, 2005 PhD
125 thesis), aiming at quantifying precisely the long-term average slip rate along the Mejillones fault,
126 one of the most prominent features within the AFS (Figures 1B and 2). The precise quantification
127 of the offsets affecting the alluvial surfaces at three sites along the fault, and the determination of
128 their ages taking into account the recent developments and knowledge enhancements concerning
129 ^{10}Be cosmic-ray exposure dating, allow us to determine the slip rate along the fault for the past 60
130 ka. Combining our results with new observations about the size of potential coseismic
131 displacements along the AFS, we bring new insights about the seismic hazard in the region.

132

133 **2. TECTONIC SETTING**

134 The AFS extends along the northern Chilean coast over more than 1,000 km. It corresponds to an
135 eastward dipping fault system with a mean N-S direction. Between 23° and 23.5°S , at the latitude
136 of the Mejillones peninsula, the AFS distributes over a 40 km wide fault zone (Fig. 1B) with fault
137 directions evolving from NNE-SSW (i.e. Salar del Carmen, Cerro Gordo, Cerro Moreno and
138 northern Cerro Fortuna faults) to NNW-SSE (i.e. Caleta Herradura fault) with N-S trending faults
139 or fault sections in between (i.e. Mejillones and southern Cerro Fortuna faults). Bathymetry and
140 seismic imagery show that this set of normal faults extends offshore and forms a seaward verging
141 domino-type structural domain (von Huene and Ranero, 2003).

142 On land Neogene to Quaternary reactivated segments of the fault system are recognized
143 north of 27°S, where fault scarps and open cracks are found within the youngest deposits (Armijo
144 and Thiele, 1990; Riquelme et al., 2003; González et al., 2003, Vargas et al., 2011; Cortés et al.,
145 2012). Age constraints on neotectonic features are scarce because of the hyperarid climate
146 (Vargas et al., 2000), which precludes the use of radiocarbon dating. K-Ar dating of biotites from
147 reworked ashes interbedded with alluvial deposits cut by the AFS leads a maximum age for the
148 onset of neotectonic activity between 2.9 to 6.1 Ma (González et al., 2003b). Morphological
149 modelling based on a diffusional model of erosion allowed González and Carrizo (2003)
150 estimating a post-400 ka activity along the Salar del Carmen fault, the easternmost fault of the
151 AFS (Fig. 1B). During the past 500 years, the reactivation of the AFS has been observed only
152 once, during the 1995, Mw8.0 Antofagasta earthquake. Surface breaks were found along several
153 segments of the southern Salar del Carmen fault, sometimes associated with a slight normal
154 offset (Delouis et al., 1998; González and Carrizo, 2003).

155 Within the Atacama Fault System, the Mejillones Peninsula appears as a prominent
156 geomorphic feature. Interrupting the N-S linear trend of the coastline, the peninsula displays
157 uplifted marine terraces, beach ridges and alluvial fans, which are crosscut by several normal
158 faults and large-scale tension gashes (Figures 1B, 1C). One of the main fault scarps is observed at
159 the north-eastern termination of the peninsula where the Mejillones fault cuts through Quaternary
160 alluvial fans. There, cumulative offsets of several tens of meters are observed (Figure 2). The low
161 denudation rates due to the regional pervasive hyper-aridity, and the lithology makes this site
162 suitable to apply the cosmic-ray exposure age dating method using in situ-produced ^{10}Be (e.g.
163 Ritz et al., 1995; Bierman et al., 1995; Marquardt, 2005).

164 Using both ^{10}Be and OSL dating techniques and detailed analysis of paleoseismic
165 trenches, Cortés et al. (2012) determined a steady slip rate of 0.61 ± 0.26 mm/yr along the

166 Mejillones fault between 35 and 3 ka. They also estimated that the fault was capable of producing
167 $M_w \sim 7$ earthquakes with mean recurrence intervals of 5 ± 3.5 ka. This is consistent with Vargas
168 et al., (2011)'s paleoseismic investigations, which used the U-series disequilibrium method to
169 constrain the timing of earthquake-induced fissures in gypsum precipitated in extensional fissures
170 along the fault. Both studies agree to date the last significant earthquake around ~ 3 , and
171 concluded that the cycle of the large earthquakes along the Mejillones Fault was not in phase
172 with the cycle of large ($M_w > 8.5$) subduction earthquakes.

173

174 **3. MORPHOTECTONIC STUDY**

175 The 20 km long Mejillones normal fault bounds to the east an uplifted faulted block (Morro
176 Mejillones) which displays a series of Pliocene and Pleistocene wave-cut platforms abraded in
177 basement rocks (Figure 3). Eastwards the Mejillones normal fault, in the hanging-wall, the Pampa
178 Mejillones is a sedimentary basin filled by Neogene marine deposits and covered by E-W
179 trending Quaternary sequences of beach-ridges formed over the last 400 ka (Ortlieb et al., 1996)
180 (Figure 3). Near the fault, the marine deposits are interbedded with thick alluvial sequences
181 deposited at the outlets of the drainage basins incising the eastern flanks of the Morro Mejillones.
182 The obvious cumulative fault scarp cutting through the latest alluvial deposits demonstrates the
183 recent activity on the Mejillones fault (Figure 2).

184 We carried out a morphological study at three sites located at the outlet of three different
185 drainage basins (Figure 3). At all three sites (O1, O2 and O3), two main alluvial surfaces are
186 displaced vertically above the youngest fan surface S0 incised by intermittent stream channels
187 (Figure 4A). These features reflect the interplay between climate-forced alluviation cycles and
188 co-seismic movements along the fault leading to the abandonment of the surfaces. S2, the most
189 obvious is the landscape, corresponds to the oldest preserved fan surface. S1, inset in S2,

190 corresponds to a younger surface. The youngest surface S0 does not show any offset. South of
191 site O3, in the direct extension of the easternmost normal fault strand (see dashed line in Figure
192 3B), we noticed that S0 was affected by a centimetric width fissure (Figure 4B). This feature
193 could be related to the 1995 Antofagasta earthquake, like the surfaces ruptures observed by
194 Delouis et al (1998) along the Salar del Carmen fault after this event.

195 The surfaces S1 and S2 are present on both sides of the Mejillones fault. In the hanging-
196 wall, eastwards of the Mejillones fault, S1 is slightly inset in S2 (Figure 4C) showing that there is
197 a general uplift over the entire area as also shown by the uplifted beach-ridges within the Pampa
198 Mejillones (Ortlieb, 1995). Fan material contains both clast-supported stream flow deposits and
199 poorly consolidated coarse-grained debris flow deposits found at the top of alluvial sequences.
200 These debris flows contain angular boulders embedded in a sand-gravel matrix with dimensions
201 scaling with the size of the feeding mountain streams. The largest boulders have diameters up to
202 1 m, with the exposed faces of the boulders displaying significant desert varnish.

203

204 **3.1 Cumulative offsets and cosmic-ray exposure ages**

205 Surfaces S2 and S1 found on both sides of the fault scarp enable accurate estimates of the
206 tectonic cumulative offsets. A differential GPS survey, with 2 Trimble 5700 GPS stations was
207 carried out within three outlets O1, O2 and O3. Digital Elevation Models (DEM) were generated
208 from the interpolation of the topographic data using Surfer software and a Kriging gridding
209 method, from which fault offsets were measured (Figure 5, see also Figures Sup1, Sup2 and Sup3
210 in Supplementary Material). The vertical offsets affecting S1 surface at the three studied outlets
211 O1, O2 and O3, are 5.2 ± 0.4 m, 4.2 ± 0.4 m and 5.8 ± 0.2 m, respectively (these values
212 correspond to the mean value between the maximum and minimum vertical offsets affecting the
213 surface (see Figure 5); the uncertainties correspond to the differences between the mean value

214 and the extreme ones). The vertical offsets affecting S2 surface at the three studied outlets are
215 12.7 ± 0.7 m (Site O1), 11.4 ± 0.8 m and 14.7 ± 0.4 m (Site O2), 11.8 ± 0.5 m and 13.2 ± 0.8 m
216 (Site O3). These measurements yield average cumulative displacements of 5.1 ± 0.8 m for
217 surface S1 and 12.8 ± 1.3 m for surface S2.

218 The minimum cosmic ray exposure ages of boulders sampled from the surfaces of offset
219 S1 and S2 alluvial fans were determined measuring in situ-produced cosmogenic ^{10}Be
220 concentrations in quartz (e.g., Brown et al., 1991; Gosse et al. 2001). Our sampling strategy
221 aimed to minimize the effects of exposure prior to deposition and of denudation following
222 deposition. Debris flows related to surfaces S1 and S2 correspond to alluvial pulses reworking
223 massive quantities of slope material, which suggests rapid transport from the upstream drainage
224 basins to their outlets. However, it has been shown that pre-exposure due to complex
225 exhumation-transport history may occur even in the case of major alluviation (e.g. Ritz et al.,
226 2003; 2006; Vassallo et al., 2005 ; 2011; Rizza et al., 2011). Since ^{10}Be production decreases
227 exponentially with depth, any deviation from this theoretical distribution indicates pre-deposition
228 inheritance or post-deposition displacement. When possible, two or three samples (top-bottom or
229 top-middle-bottom) have been collected from granite boulders embedded in S1 and S2 surfaces to
230 identify any such previous complex exposure histories (Figure 6).

231 Within both S1 and S2 surfaces, five boulders distributed on both sides of the fault scarp
232 were sampled (Table 1). To minimize the effects of denudation after deposition, the biggest
233 boulders showing desert varnished and standing above the surface, but still encased in it, were
234 sampled (Figure 6). Surface S2 was sampled at Site O3 (Figures 3B, 5). Five boulders were
235 sampled; four in the footwall and one in the hanging wall. Surface S1 was sampled at Site O2

236 (Figures 3B, 5). Three samples were we also collected along a 2 m deep soil pit excavated in
237 surface S1 to determine the evolution of the ^{10}Be concentration as a function of depth.

238 Quartz was isolated from crushed and sieved samples by dissolving all other minerals in a
239 mixture of 1/3 HCl – 2/3 H_2SiF_6 acidic solution. Atmospheric ^{10}Be was then eliminated by
240 successive HF sequential dissolutions (e.g., Brown et al., 1991). After complete dissolution of the
241 purified quartz in Suprapur HF, a reference 300 μg ^9Be spike (Merck Certipur) was added (e.g.,
242 Bourlès, 1988). Beryllium was separated from these solutions by successive solvent extractions
243 and alkaline precipitations. All ^{10}Be measurements were performed in June 2002 by accelerator
244 mass spectrometry at the Tandétron facility, Gif-sur-Yvette, France (Raisbeck et al., 1994)
245 against the NIST SRM4325 standard of certified $^{10}\text{Be}/^9\text{Be}$ value of $2.68 \cdot 10^{-11}$. However, together
246 with the absolute measurement of this ratio by Nishiizumi et al. (2007), the half live of ^{10}Be has
247 also been more precisely measured in 2010 by two independent teams (Korschineck et al; 2010,
248 Chmeleff et al. 2010). Therefore, to incorporate these two recent advances, the original ratios
249 have been recalculated, and the measured ^{10}Be concentrations presented in Table 1 are now
250 consistent with an assigned NIST value of $2.79 \cdot 10^{-11}$ and a ^{10}Be half-life of 1.387 Ma.

251 - Minimum exposure ages have been calculated from: the measured concentrations
252 extrapolated to surface assuming neutron attenuation only (see Table 2), the CREp
253 (Cosmic Ray Exposure program) (Martin et al., 2017) with a production rate at sea level
254 high altitude of $4.02 \pm 0.16 \text{at/g/a}$ (Borchers, 2016), and using the polynomials of Stone
255 (2000), a standard atmosphere and a custom geomagnetic database based on authigenic
256 $^{10}\text{Be}/^9\text{Be}$ reconstruction from Simon et al. (2016). The muon scheme is based on Braucher
257 et al (2011), which uses an altitudinal scaling only and sea level production rates of 0.012
258 at/g/a and 0.039 at/g/a for slow and fast muons, respectively. Maximum denudation rates

259 have been similarly calculated by setting the time to infinite. All necessary equations
260 (eq.1 and eq. 2) are presented in the Supplementary Material.

261
262 To combine and compare the different sample concentrations, the method proposed by Ward and
263 Wilson (1978) was then applied (Table 2, Figure 7, and Figure Sup4 in Supplementary Material):

264 - This method is based on Chi-square analysis. To obtain the most adequate number of
265 samples (n) per surface (i.e. samples that are statistically undistinguishable), the 0.05
266 critical value for a Chi-square with (n-1) degrees of freedom is calculated and compared
267 with the theoretical value given by the chi-square table. If the calculated value for these n
268 samples, is lower than the theoretical one, then all samples are used to calculate a
269 weighted mean concentration; otherwise outliers are rejected until the distribution passes
270 the test and the weighted mean concentration is calculated with the remaining samples
271 (i.e. the n samples less the outliers) (Figure 7).

272
273 We applied the method described above, on the oldest alluvial surface S2, considering the 12
274 samples, and found a chi2 value of 98.2. This is much higher than the theoretical one of 19.68. It
275 appears that for boulder B, the concentration measured in the lowest sample B3 is larger when
276 extrapolated to the surface than that measured in the sample B1 collected at the surface.
277 Moreover, sample B2, collected in between leads to a depleted concentration once extrapolated to
278 the surface. The decrease of the in situ-produced ^{10}Be concentration at the core of the boulder and
279 its increase at the bottom suggests that the boulder has been flipped and exposed from one face to
280 another at least once before its present exposition (e.g. Carretier and Regard, 2011). We therefore
281 consider boulder B as an outlier. Similarly, the wide dispersion of the in situ-produced ^{10}Be
282 concentrations measured in the samples from boulder E most likely indicates a complex exposure

283 history. Boulder E is therefore considered as an outlier. Thus, considering these concentrations
284 extrapolated to the surface for the samples A, C and D, a chi2 value of 6.11 is calculated, yielding
285 to a minimum surface cosmic ray exposure age of 61.0 ± 3.1 ka for surface S2 assuming
286 negligible denudation rate (Table 2; note that because ^{10}Be surface concentrations of samples A,
287 C and D are similar, inheritance can be neglected).

288 The same approach was applied to the concentrations obtained from the younger surface
289 S1. The surface concentrations (F1, G1, I1) and the concentrations extrapolated to the surface of
290 the samples collected at depth (F2, G2, and I2) are internally consistent considering all the
291 samples (F, G, H, I) excluding here also inherited ^{10}Be from previous history. A mean surface
292 concentration of 121.7 ± 7.4 kat/g is thus calculated, yielding to a minimum surface exposure age
293 of 35.3 ± 2.0 ka. (Table 2).

294 However, one can argue that using a negligible denudation rate may not be right in such
295 environment. Therefore, taking benefit of the multiple sampling within the same boulder, a
296 maximum denudation rate can be determined for each sample from its measured ^{10}Be
297 concentration. This is presented in Table Sup1 in Supplementary Material. One can observe that
298 top and bottom denudations rates are similar within uncertainties. This may suggest that boulders
299 are closed to steady state. To estimate their minimum exposure age, a so called integration times
300 (Lal, 1991; Rizza et al., 2011) can be determined from each sample concentration, as well as
301 mean maximum denudation rates and their related integration time for surfaces S1 and S2, from
302 the ^{10}Be mean surface concentrations of S1 and S2 (Table 2). This yields denudation rates of
303 19.59 ± 1.18 m/Myr and 10.18 ± 0.78 m/Myr, and integration times of 37.1 ± 2.2 ka and $67.4 \pm$
304 5.1 ka for S1 and S2, respectively.

305

306 We also modeled the depth profile ^{10}Be concentrations obtained in the soil pit dug in S1 surface
307 following Braucher et al (2009) and using the approach of Hidy et al (2010). The best fit
308 (associated to the lower chi2) yields a minimum age of 38.49 (+ 14.43 / -3.45) ka (Figure Sup5
309 and Table Sup2 in Supplementary Material). This is consistent with the integration time
310 calculated for S1. Note that the inheritance is totally negligible within the soil pit material
311 suggesting that S1 emplaced rapidly with short transport time. This is consistent with Vargas et
312 al. (2006)'s observation indicating that the period 35–15 ka corresponded to a period of strong
313 alluviations.

314

315 **3.2 Slip rates**

316 To determine the slip rate along the Meji llones fault, we used the minimum surface exposure
317 ages calculated above without denudation, which are very close to the integration ages. Dividing
318 S2 and S1 mean offsets (12.8 ± 1.3 and 5.1 ± 0.8 m, respectively) by their minimum surface
319 cosmic ray exposure ages (61.0 ± 3.1 ka and 35.3 ± 2.0 ka, respectively) yield maximum vertical
320 slip rates of 0.21 ± 0.03 mm/yr and 0.15 ± 0.03 mm/yr, respectively. These results suggest that
321 the vertical slip rate along the Meji llones fault has slightly decreased through time during the past
322 60 ka. However, given the uncertainties, these two estimates are not significantly different, and a
323 mean 0.18 mm/yr vertical slip rate can be proposed for the Meji llones fault since the mid-upper
324 Pleistocene. Where the fault cross cut the present shoreline, in the northern part of the peninsula,
325 we could observe the main fault plane outcropping (see Figure 3B for location). There, we
326 measured a fault azimuth of N008°E, a dip of 63°E, and a pitch for the slickensides of 84°N.
327 This allows calculating an actual mean slip rate along the Meji llones fault of 0.20 mm/yr.

328

329 **3.3 Coseismic displacements along the AFS**

330 Along the Mejillones fault, we observed few subtle cracks within S0 surface that can be
331 interpreted as surface breaks associated with the Mw 8.1, 1995 Antofagasta earthquake (See
332 section 3.1; Figure 4B). However, the cumulative displacements observed along the fault within
333 S1 and S2 surfaces are clearly the result of larger dislocations. After the Antofagasta earthquake,
334 Delouis et al. (1998) observed surface ruptures with decimetric vertical displacements along the
335 Paposo fault (the southern extension of the Salar Del Carmen fault) 100 km southwards the
336 Mejillones peninsula (Figures 8A, B), within a site situated in the middle of the epicentral area
337 of the 1995 Mw8.1 Antofagasta earthquake (see Figure 1A for location). Looking for the
338 extension of this surface rupture further north along the Paposo fault, and the northern Salar del
339 Carmen fault, we observed an outstanding fault scarp predating the 1995 earthquake (Figure 8C).
340 Given the absence of beveled surfaces above and beneath the scarp, and given also the fact that
341 we did not observed stepped surfaces, we interpreted this fault scarp as the result of a single co-
342 seismic surface dislocation, and not the result of cumulative offsets. A detailed topographic
343 survey of the fault scarp allowed us determining a co-seismic vertical offset of 2.5 ± 0.5 m
344 (Figure 9), a value which is of the same order than the offsets estimated along the Mejillones
345 fault by Cortés et al. (2012) from paleoseismic investigations. Therefore, the question of the
346 magnitude of the event during which such a vertical displacement occurred along the Salar del
347 Carmen fault arises if we assume that it is contemporary with a subduction earthquake, as is the
348 1995 fault rupture described by Delouis et al. (1998) farther south along the Paposo fault.

349

350 **4. DISCUSSION**

351 Our study allows determining a mean slip rate along the Mejillones fault of 0.20 mm/yr during
352 the past 60 ka. This is lower than the mean ~ 0.6 mm/yr slip rate estimated by Cortés et al. (2012)

353 for the time period comprised between 35 and 3 ka. However, it should be noted that Cortés et al.
354 (2012) detailed estimates for the time intervals 35-14ka and 3ka-Present (respectively $0.46 \pm$
355 0.41 mm/yr and $0.22 + 0.06$ mm/yr, respectively) are not significantly different with our result.
356 Therefore, we cannot confirm whether or not, the Mejillones fault has known an acceleration
357 between 14 and 3 ka (i.e. 0.6 mm/yr) as proposed by Cortés et al., (2012). On the other hand, the
358 0.2 mm/yr differential uplift estimated by Ortlieb (1995) over the past 400 ka, between the
359 Pampa Mejillones (uplift rate of 0.2 mm/yr) and the Morro Mejillones faulted block (uplift rate of
360 0.4 mm/yr) tends to confirm that the long term slip rate is closer to 0.2 mm/yr than 0.6 mm/yr.

361 If we consider that the faults within the AFS are moving contemporaneously with the
362 strong subduction earthquakes, as it has been observed for instance during the 1995, Mw8.1
363 Antofagasta earthquake (Delouis et al., 1998), we may assume that the recurrence of surface
364 ruptures along the AFS should be the same as the recurrence of subduction earthquakes. With a
365 convergence rate between the Nazca plate and Chile of about ~ 8 cm/yr (DeMets et al., 1994), an
366 average 150 ± 100 years recurrence interval for strong ($M_w \geq 8$) subduction earthquakes can be
367 calculated, assuming a regular recurrence intervals and a coseismic displacement between 4 and
368 23 m (range of displacement estimated from Wells and Coppersmith (1994)'s general statistical
369 function, linking $M_w \geq 8$ and the average coseismic displacement). Combined with our long term
370 slip rate estimated along the Mejillones fault (0.2 mm/yr), a 150 ± 100 yrs return period would
371 yield a mean vertical offset of 3 ± 2 cm per event along the fault if this one moves during each
372 subduction earthquake occurring bellow the peninsula. This is not consistent with the cumulative
373 fault scarps observed along the fault, which are clearly the result of larger dislocations in surface,
374 as also shown by the paleoseismic studies carried out by Cortés et al., (2012) suggesting that the
375 cumulative fault scarps along the Mejillones fault are the results of surface rupturing events with
376 vertical displacements comprised between 0.6 and 3.0 m. These last displacements are of the

377 same order of the co-seismic slip that we measured along the Paposo or the Salar del Carmen
378 faults (Figure 8). This suggests that while the mean recurrence time of fault activation in the AFS
379 region would be the same than those of the strong subduction earthquakes (ca 150 yrs), the
380 recurrence time of the AFS faults, taken separately is longer. This also suggests that the different
381 AFS faults can ruptured individually after several Mw8 earthquakes and/or that subduction
382 earthquakes larger than the Mw 8.1 1995 Antofagasta event can occur in the area. In
383 consequence, we can wonder whether the 2.5 m high coseismic fault scarp observed along the
384 southern part of the Salar del Carmen fault, eastwards of Antofagasta, could not be related to the
385 last major historical subduction earthquake in 1877 (M 8.6).

386

387 **5. CONCLUSIONS**

388 Our morphotectonic study along the Mejillones fault - mapping the topography and the geology
389 of the offset alluvial fans, and dating them with in situ-produced cosmogenic ^{10}Be – allows
390 determining that this main normal fault of the Atacama fault system (AFS) has a 0.2 mm/yr long-
391 term slip rate. This result suggests that the AFS is composed of slow slip rates faults despite the
392 fast rate (8 cm/yr) of subduction. However, the potential co-seismic slip observed along the
393 different faults of the AFS, combined with the fact they seem to be moving contemporaneously
394 with subduction earthquakes – as during the Mw 8.1, 1995, Antofagasta earthquake – suggest
395 that a larger earthquake could occur episodically in the area.

396 In this line of reasoning, the 2007 Tocopilla (M 7.7) earthquake having released only 2.5
397 per cent of the moment deficit accumulated on the interface during the 130 years period
398 following the 1877 historical event (Béjar-Pizarro et al., 2010) could be regarded as a possible
399 precursor of a larger subduction earthquake rupturing partially or completely the 500-km-long

400 North Chile seismic gap. Keeping on quantifying morphotectonic and paleoseismic features
401 along the AFS may help better asses this issue.

402

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413

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- 597

598 **FIGURES CAPTIONS**

599

600 **Figure 1. A:** Simplified map of the northern Chilean subduction zone with large historical
 601 (dashed grey ellipses) and instrumental (dashed red rectangles) earthquakes with their date and
 602 magnitude. AFS: Atacama Fault System. **B:** Sketch map of the quaternary faults affecting the
 603 Mejillones Peninsula, in the central part of the Atacama Fault System. **C:** Landsat Image (Google
 604 Earth) of the Mejillones peninsula with examples of uplifted marine terrace (white triangles),
 605 beach ridges (small black triangles), fault scarps (black arrows) and tension gashes (white
 606 arrows).

607

608 **Figure 2:** Field views of the Mejillones normal fault scarp within its northern part (A) and its
 609 central part (B). White arrows point out the base of the scarp (coordinates of picture shots; A: 23°
 610 8.865'S, 70° 29.715'W; B: 23° 6.963'S, 70° 29.705'W; see also Figure 3 for location).

611

612 **Figure 3.** Aerial photograph (A) and corresponding geological interpretation (B) of the
 613 Mejillones fault within its northern termination. The three studied sites (O1, O2, and O3) are
 614 indicated by the black frames. The black star indicated the site where were measured fault slip
 615 data along the main fault plane.

616

617 **Figure 4.** A: Field picture showing S2 and S1 alluvial surfaces displaced by the Mejillones fault
 618 in Site O1 (white arrows point out the main fault scarp; grey arrows a secondary scarp; see also
 619 the corresponding DEM in Figure 5). B: Fissure affecting surface S0 immediately to the south of
 620 studied Site O3. C: Downslope view of Site O3, showing the occurrence of S1 and S2 surfaces on
 621 both sides of the fault, with S1 inset in S2, both in the footwall and the hanging-wall.

622 **Figure 5.** Oblique view of the Digital Elevation Models of the three studied sites O1, O2 and O3,
623 built up from DGPS survey (see Supplementary Material). Topographic profiles across the fault
624 scarps allow estimating maximum and minimum offsets values for S1 (yellow) and S2 (orange)
625 fan surfaces. Capital letters A to I indicate samples locations.

626

627 **Figure 6:** A: Picture showing surface S2 sampled (boulder A) at site O2 within the footwall. B:
628 Picture showing surface S2 sampled (boulder E) at site O3 within the hanging wall compartment.

629

630 **Figure 7.** ^{10}Be concentrations from boulders sampled on S2 (boulders A to E) and S1 (boulders F
631 to I) fan surfaces. Grey lines represent the mean weighted minimum concentrations.

632

633 **Figure 8.** A and B: Pictures showing the surface rupture observed along the Paposo Fault within
634 the AFS (~100 km southwards of the Mejillones Peninsula), which was contemporaneous with
635 the 1995, Mw8.1 Antofagasta earthquake (see Figure 1A) (field pictures after Delouis et al.,
636 1998). C: Picture showing the fault scarp observed along the southern part of the Salar del
637 Carmen fault (see Figure 1B for location). The free faces (in shadow) attest of the latest, metric
638 vertical coseismic displacement.

639

640 **Figure 9.** A: Contours map of the scarp observed along the southern part of the Salar del Carmen
641 fault (see Figure 8C) generated from topographic survey with kinematics DGPS. B: Topographic
642 profiles across the scarp with corresponding coseismic vertical displacements.

643

644

645

646 **TABLES**

647 **Table 1.** ^{10}Be data: the ^{10}Be concentrations have been calculated relatively to the NIST Standard
648 Reference Material 4325, using an assigned ratio of ^{10}Be to ^9Be of $(2.79 \pm 1.5) \times 10^{-11}$. ^{10}Be
649 uncertainties (1σ) include a 3% contribution conservatively estimated from observed variations in
650 the standard during the runs, a 1σ statistical error in the number of ^{10}Be events counted,
651 uncertainty in the blank correction. Associated $^{10}\text{Be}/^9\text{Be}$ blank ratio was $(4.37 \pm 1.32) \times 10^{-15}$.

652

653 **Table 2.** ^{10}Be concentrations extrapolated to surface assuming a neutron attenuation length of
654 150 g/cm^2 and samples depths and densities as presented in Table 1. Minimum exposure ages
655 have been calculated using the CREp online calculator (see text). To combine and compare the
656 different samples concentrations, the method proposed by Ward and Wilson (1978) was applied
657 (see text).