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

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ABSTRACT

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

1. INTRODUCTION

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

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

Observing fresh normal displacement along the Paposo fault that did not exist before the 1995 Antofagasta earthquake (see location “Fig.8A” in Figure 1C), Delouis et al. (1998) proposed that the Atacama fault system was strongly coupled with the seismic cycle along the subduction zone, with normal faulting events contemporaneous with strong co-seismic subduction events. The same kind of observations were made during the M_w 8.8, 2010 Maule earthquake near Santiago, with the occurrence of large aftershocks nucleated on normal faults in the upper plate (Aron et al., 2013).

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

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

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

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

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

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

In the present study, we do not discuss the role and origin of the west-dipping coastal scarp, which could act differently than the AFS (Armijo & Thiele, 1990; Contreras-Reyes et al., 2012). We consider, as in Delouis et al. (1998) and Cortés et al. (2015), that the MFS is part of the AFS and respond to the same kind of CFS mechanical interaction with the subduction plate interface.

If the mechanism driving this trench parallel normal faulting is associated with the seismic cycle within the northern Chile subduction zone, analyzing geomorphology along these faults may provide information on the seismic potential associated with the subduction zone. In this paper, we present a study carried out in February 1999 and April 2002 (Marquardt, 2005 PhD thesis), aiming at quantifying precisely the long-term average slip rate along the Mejillones fault, one of the most prominent features within the AFS (Figures1B and 2). The precise quantification of the offsets affecting the alluvial surfaces at three sites along the fault, and the determination of their ages taking into account the recent developments and knowledge enhancements concerning ^{10}Be cosmic-ray exposure dating, allow us to determine the slip rate along the fault for the past 60 ka. Combining our results with new observations about the size of potential coseismic displacements along the AFS, we bring new insights about the seismic hazard in the region.

2. TECTONIC SETTING

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

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

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

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

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

3. MORPHOTECTONIC STUDY

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

We carried out a morphological study at three sites located at the outlet of three different drainage basins (Figure 3). At all three sites (O1, O2 and O3), two main alluvial surfaces are displaced vertically above the youngest fan surface S0 incised by intermittent stream channels (Figure 4A). These features reflect the interplay between climate-forced alluviation cycles and co-seismic movements along the fault leading to the abandonment of the surfaces. S2, the most obvious is the landscape, corresponds to the oldest preserved fan surface. S1, inset in S2, corresponds to a younger surface. The youngest surface S0 does not show any offset. South of site O3, in the direct extension of the easternmost normal fault strand (see dashed line in Figure 3B), we noticed that S0 was affected by a centimetric width fissure (Figure 4B). This feature could be related to the 1995 Antofagasta earthquake, like the surfaces ruptures observed by Delouis et al (1998) along the Salar del Carmen fault after this event.

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

3.1 Cumulative offsets and cosmic-ray exposure ages

Surfaces S2 and S1 found on both sides of the fault scarp enable accurate estimates of the tectonic cumulative offsets. A differential GPS survey, with 2 Trimble 5700 GPS stations was carried out within three outlets O1, O2 and O3. Digital Elevation Models (DEM) were generated from the interpolation of the topographic data using Surfer software and a Kriging gridding method, from which fault offsets were measured (Figure 5, see also Figures Sup1, Sup2 and Sup3 in Supplementary Material). The vertical offsets affecting S1 surface at the three studied outlets O1, O2 and O3, are 5.2 ± 0.4 m, 4.2 ± 0.4 m and 5.8 ± 0.2 m, respectively (these values correspond to the mean value between the maximum and minimum vertical offsets affecting the surface (see Figure 5); the uncertainties correspond to the differences between the mean value and the extreme ones). The vertical offsets affecting S2 surface at the three studied outlets are 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 (Site O3). These measurements yield average cumulative displacements of 5.1 ± 0.8 m for surface S1 and 12.8 ± 1.3 m for surface S2.

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

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

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

Quartz was isolated from crushed and sieved samples by dissolving all other minerals in a mixture of 1/3 HCl – 2/3 H_2SiF_6 acidic solution. Atmospheric ^{10}Be was then eliminated by successive HF sequential dissolutions (e.g., Brown et al., 1991). After complete dissolution of the purified quartz in Suprapur HF, a reference 300 µg ^{9}Be spike (Merck Certipur) was added (e.g., Bourlès, 1988). Beryllium was separated from these solutions by successive solvent extractions and alkaline precipitations. All ^{10}Be measurements were performed in June 2002 by accelerator mass spectrometry at the Tandétron facility, Gif-sur-Yvette, France (Raisbeck et al., 1994) against the NIST SRM4325 standard of certified $^{10}Be/^{9}Be$ value of $2.68.10^{-11}$. However, together with the absolute measurement of this ratio by Nishiizumi et al. (2007), the half live of ^{10}Be has also been more precisely measured in 2010 by two independent teams (Korschineck et al; 2010, Chmeleff et al. 2010). Therefore, to incorporate these two recent advances, the original ratios have been recalculated, and the measured ^{10}Be concentrations presented in Table 1 are now consistent with an assigned NIST value of $2.79.10^{-11}$ and a ^{10}Be half-life of 1.387 Ma.

- Minimum exposure ages have been calculated from: the measured concentrations extrapolated to surface assuming neutron attenuation only (see Table 2), the CREp (Cosmic Ray Exposure program) (Martin et al., 2017) with a production rate at sea level high altitude of 4.02 ± 0.16at/g/a (Borchers, 2016), and using the polynomials of Stone (2000), a standard atmosphere and a custom geomagnetic database based on authigenic $^{10}Be/^{9}Be$ reconstruction from Simon et al. (2016). The muon scheme is based on Braucher et al (2011), which uses an altitudinal scaling only and sea level production rates of 0.012 at/g/a and 0.039 at/g/a for slow and fast muons, respectively. Maximum denudation rates have been similarly calculated by setting the time to infinite. All necessary equations (eq.1 and eq. 2) are presented in the Supplementary Material.

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

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

We applied the method described above, on the oldest alluvial surface S2, considering the 12 samples, and found a chi2 value of 98.2. This is much higher than the theoretical one of 19.68. It appears that for boulder B, the concentration measured in the lowest sample B3 is larger when extrapolated to the surface than that measured in the sample B1 collected at the surface. Moreover, sample B2, collected in between leads to a depleted concentration once extrapolated to the surface. The decrease of the in situ-produced ^{10}Be concentration at the core of the boulder and its increase at the bottom suggests that the boulder has been flipped and exposed from one face to another at least once before its present exposition (e.g. Carretier and Regard, 2011). We therefore consider boulder B as an outlier. Similarly, the wide dispersion of the in situ-produced ^{10}Be concentrations measured in the samples from boulder E most likely indicates a complex exposure history. Boulder E is therefore considered as an outlier. Thus, considering these concentrations extrapolated to the surface for the samples A, C and D, a chi2 value of 6.11 is calculated, yielding to a minimum surface cosmic ray exposure age of 61.0 ± 3.1 ka for surface S2 assuming negligible denudation rate (Table 2; note that because ^{10}Be surface concentrations of samples A, C and D are similar, inheritance can be neglected).

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

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

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

3.2 Slip rates

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

3.3 Coseismic displacements along the AFS

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

4. DISCUSSION

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

for the time period comprised between 35 and 3 ka. However, it should be noted that Cortés et al. (2012) detailed estimates for the time intervals 35-14ka and 3ka-Present (respectively 0.46 ± 0.41mm/yr and 0.22 + 0.06 mm/yr, respectively) are not significantly different with our result. Therefore, we cannot confirm whether or not, the Mejillones fault has known an acceleration between 14 and 3 ka (i.e. 0.6 mm/yr) as proposed by Cortés et al., (2012). On the other hand, the 0.2 mm/yr differential uplift estimated by Ortlieb (1995) over the past 400 ka, between the Pampa Mejillones (uplift rate of 0.2 mm/yr) and the Morro Mejillones faulted block (uplift rate of 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.

If we consider that the faults within the AFS are moving contemporaneously with the strong subduction earthquakes, as it has been observed for instance during the 1995, Mw8.1 Antofagasta earthquake (Delouis et al., 1998), we may assume that the recurrence of surface ruptures along the AFS should be the same as the recurrence of subduction earthquakes. With a convergence rate between the Nazca plate and Chile of about ~8 cm/yr (DeMets et al., 1994), an average 150 ± 100 years recurrence interval for strong (Mw≥8) subduction earthquakes can be calculated, assuming a regular recurrence intervals and a coseismic displacement between 4 and 23 m (range of displacement estimated from Wells and Coppersmith (1994)'s general statistical function, linking Mw≥8 and the average coseismic displacement). Combined with our long term slip rate estimated along the Mejillones fault (0.2 mm/yr), a 150 ± 100 yrs return period would yield a mean vertical offset of 3±2 cm per event along the fault if this one moves during each subduction earthquake occurring bellow the peninsula. This is not consistent with the cumulative fault scarps observed along the fault, which are clearly the result of larger dislocations in surface, as also shown by the paleoseismic studies carried out by Cortés et al., (2012) suggesting that the cumulative fault scarps along the Mejillones fault are the results of surface rupturing events with vertical displacements comprised between 0.6 and 3.0 m. These last displacements are of the same order of the co-seismic slip that we measured along the Paposo or the Salar del Carmen faults (Figure 8). This suggests that while the mean recurrence time of fault activation in the AFS region would be the same than those of the strong subduction earthquakes (ca 150 yrs), the recurrence time of the AFS faults, taken separately is longer. This also suggests that the different AFS faults can ruptured individually after several Mw8 earthquakes and/or that subduction earthquakes larger than the Mw 8.1 1995 Antofagasta event can occur in the area. In consequence, we can wonder whether the 2.5 m high coseismic fault scarp observed along the southern part of the Salar del Carmen fault, eastwards of Antofagasta, could not be related to the last major historical subduction earthquake in 1877 (M 8.6).

5. CONCLUSIONS

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

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

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

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FIGURES CAPTIONS

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

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

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

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

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

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

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

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

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

TABLES

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

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