Crustal faults in the Chilean Andes:
geological constraints and seismic potential

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ABSTRACT. The Chilean Andes, as a characteristic tectonic and geomorphological region, is a perfect location to unravel the geologic nature of seismic hazards. The Chilean segment of the Nazca-South American subduction zone has experienced mega-earthquakes with Moment Magnitudes (Mw) >8.5 (*e.g.,* Mw 9.5 Valdivia, 1960; Mw 8.8 Maule, 2010) and many large earthquakes with Mw >7.5, both with recurrence times of tens to hundreds of years. By contrast, crustal faults within the overriding South American plate commonly have longer recurrence times (thousands of years) and are known to produce earthquakes with maximum Mw of 7.0 to 7.5. Subduction-type earthquakes are considered the principal seismic hazard in Chile, with the potential to cause significant damage to its population and economy. However crustal (non-subduction) earthquakes can also cause great destruction at a local scale, because of their shallower hypocentral depth. Nevertheless, the nature, timing and slip rates of crustal seismic sources in the Chilean Andes remain poorly constrained. This work aims to address the seismic potential of the crustal faults in Chile, contributing to the estimation of key fault parameters for the seismic hazard assessment. We have examined the main parameters involved in the magnitude of an earthquake, including length, width and mean displacement of some case studies crustal faults and their morphotectonic settings, exposing the parametrical similarities in longitudinal domains (N-S stripes) and disparity from W to E, across latitudinal domains. The maximum hypocentral depths for crustal earthquakes vary across margin parallel tectonic domains aligned parallel, from 25-30 km in the outer forearc to 8-12 km in the volcanic arc, thus allowing for a first-order approach for seismic potential assessment. Current structural, paleoseismological and geodetic data, although sparse and limited, suggest that slip rates of Chilean crustal faults range from 0.2 mm/yr (in the forearc region) to up to 7.0 mm/yr (in the intra-arc region). The different tectonic modes for crustal fault reactivation and their wide range of slip rates complicates the estimation of seismic hazard. A rigorous seismic hazard assessment must therefore consider the different tectonic settings, timing and slip rates of Andean crustal faults. Understanding the nature of these faults will allow a better evaluation of the associated seismic hazard, and better constraints to be placed on their relationship with the subduction seismic cycle.

Keywords: Crustal Faults, Seismic Hazard, Neotectonics, Crustal earthquakes, Paleoseismology, Chilean Andes.

RESUMEN. Fallas corticales en los Andes chilenos: limitaciones geológicas y potencial sísmico. Los Andes chilenos, entendidos como una región tectónica y geomorfológica característica, son un lugar perfecto para revelar la naturaleza geológica del peligro sísmico. El segmento chileno de la zona de subducción Nazca-Sudamericana ha experimentado mega-terremotos de Magnitud de Momento (Mw) >8,5 (*e.g.,* Mw 9,5 Valdivia, 1960; Mw 8,8 Maule, 2010) y numerosos terremotos de Mw >7,5, ambos tipos con tasas de recurrencia de decenas a cientos de años. En contraste, las fallas corticales en la placa superior sudamericana comúnmente tienen tasas de recurrencia mayores (miles de años) y son conocidas por producir terremotos con magnitud Mw máximas 7,0 a 7,5. Los terremotos del tipo subducción han sido considerados como el principal peligro sísmico en Chile, con el potencial de causar daño importante a la población y economía, pero los terremotos corticales también tienen la capacidad de causar gran destrucción a escala local, debido a la menor profundidad de sus hipocentros. Sin embargo, la naturaleza, las tasas de recurrencia y el deslizamiento de las fallas corticales en Chile están pobremente delimitadas. Este trabajo tiene como objetivo investigar el potencial sismogénico de las fallas corticales en Chile, y contribuir en la estimación de los principales parámetros de las fallas para la evaluación del peligro sísmico. Se examinaron los parámetros principales involucrados en la magnitud de un terremoto, incluyendo largo, ancho y desplazamiento medio de las fallas corticales y su disposición morfotectónica, revelando las similitudes en los parámetros en cada dominio longitudinal (franjas N-S) y una mayor disparidad de W a E (dominios latitudinales). Se muestra que las máximas profundidades de hipocentros corticales cambian en los dominios tectónicos paralelos al margen continental desde 25-30 km en el antearco externo a 8-12 km en el arco volcánico, lo que permite una primera aproximación a la evaluación del potencial sísmico. Aunque dispersos y limitados, los datos estructurales, paleosismológicos y geodésicos actuales sugieren tasas de deslizamiento de las fallas corticales en Chile entre 0,2 mm/a (dominio antearco) y 7,0 mm/a (dominio intraarco). Las diferentes condiciones tectónicas de reactivación de las fallas corticales junto al amplio rango de tasas de deslizamiento complican la estimación del peligro sísmico. La investigación rigurosa del peligro sísmico debe considerar los diferentes escenarios tectónicos, las tasas de recurrencia y de deslizamiento de las fallas corticales andinas. Entender la naturaleza de estas fallas nos ayudará no solo a entender mejor el peligro sísmico asociado, sino que también a precisar la conexión con el ciclo sísmico de subducción.

Palabras clave: Fallas corticales, Peligro sísmico, Neotectónica, Terremotos corticales, Paleosismología, Andes chilenos.

1. Introduction

Chile is located in an outstanding physical laboratory to investigate the nature of earthquakes and related seismic hazard: The Chilean Andes is a distinctive tectonic and geomorphological orogenic region formed, in the north, by the ocean-continent convergence between the South American and Nazca plates, and in the south by the convergence between the Antarctic and Scotia plates. Its primary features include the Coastal Cordillera, Central Depression and Main and Patagonian Cordillera (*e.g.,* Pankhurst and Hervé, 2007; Ramos, 2009b).

Continental Chile is affected by three main types of earthquakes: subduction, or thrust-type events, due to the Nazca-South America interplate motion, which typically have a hypocentral depth up to 60 km (*e.g.,* Tichelaar and Ruff, 1993; Suárez and Comte, 1993; Allmendinger and González, 2010; Scholz and Campos, 2012); intraplate earthquakes, which occur within the subducting Nazca plate, and typically have hypocentral depths of 60-200 km (*e.g.,* Barrientos, 1980; Campos and Kausel, 1990); and crustal intraplate earthquakes, which occur within the overriding South American plate, and generally have hypocentral depths of less than 30 km (*e.g.,* Barrientos *et al*., 2004; Leyton *et al*., 2010).

The Nazca-South American Chilean subduction zone is capable of producing large magnitude earthquakes with Moment Magnitudes (Mw) greater than 8.5 (*e.g.,* Mw 9.5 Valdivia, 1960; Mw 8.8 Maule, 2010). Earthquake recurrence times in these areas have been estimated at between 80 to 120 years for earthquakes with Mw >7.5, and between 250 to 500 years for earthquakes with Mw >8.5 (Comte *et al*., 1986; Barrientos and Ward, 1990; Lomnitz, 2004; Cisternas *et al*., 2005; Barrientos, 2007). Such, "megathrust" subduction type earthquakes are considered the principal seismic hazard in Chile, with the potential to cause significant damage to its population and economy.

Crustal faults within the overriding South American plate (intraplate crustal faults/earthquakes are hereafter named simply "crustal" faults/earthquakes), by contrast, commonly have recurrence times of thousands of years (*e.g.,* Barrientos *et al*., 2004; Silva, 2008; Leyton *et al*., 2010) and according to the seismic records are known to produce earthquakes of only moderate magnitudes Mw < 7.5, without inducing documented primary surface deformation. However,

the lack of more detailed seismic and neotectonic studies may have biased our current understanding of crustal faults. Such faults are controlled by contemporary tectonics, or "Neotectonics" (*e.g.,* Mw 6.2, Aysén, 2007, Vargas *et al.*, 2013) or are produced by the reactivation of inherited mechanical anisotropies under a different setting (Paleotectonics) with respect to the present-day stress-field (*e.g.,* Mw 6.9 and Mw 7.0, Pichilemu, 2010, Aron *et al.*, 2013). The term Neotectonics, which has been used variably by previous authors (Mercier, 1976.; Fairbridge, 1981; Hancock and Williams, 1986; Vita-Finzi, 1987; Hancock, 1988; Pavlides, 1989; Mörner, 1994; Costa *et al.*, 2006a), here refers to the study of the processes and structures acting since the contemporary stress field of a given region was established (Stewart y Hancock, 1994). In Chile, this refers to the tectonic system which has been active since at least the start of the Pliocene (González *et al.*, 2003; Cembrano *et al.*, 2007; Ramos 2009a).

The available structural, paleoseismological and geodetic data, though limited, suggest that current slip rates of Chilean crustal faults range from 0.2 mm/yr in the fore-arc to up to 6.5 mm/yr in the intra-arc (*e.g.,* Wang *et al.*, 2007; Armijo *et al.*, 2010; Cortés *et al.*, 2012; Stanton-Yonge *et al.*, 2016). This implies recurrence times of Mw>6 crustal fault earthquakes in the range of hundreds-to-tens of thousands of years (Sibson, 2002). The various modes of tectonic fault reactivation, and the wide range of slip rates that these faults exhibit, greatly complicate crustal seismic hazard assessment in Chile. Many faults which are considered active under the traditional classification (*i.e.,* over the last 10,000 years, USGS, 2018b) may not be capable of generating significant earthquakes in thousands of years according to their recurrence time (Machette, 2000; Costa, 2004; Costa *et al*, 2006a), while faults of quaternary age or older, with no instrumentally recorded seismicity, may have the potential to trigger socially consequential earthquakes with Mw>7 (*e.g.,* 2011 Mw 7.1 Darfield earthquake, New Zealand; Li *et al.*, 2014).

In this work we attempt to characterize and estimate key crustal fault parameters and general conditions related to their tectonic and geomorphological setting, to introduce a first approximation for the crustal fault seismic hazard assessment (SHA) in Chile. Despite the large epistemic uncertainties involved in most fault data, basic fault conditions parameters were populated in order to better understand the input of geological structures in SHA and also to provide guidance for further investigations in this subject matter.

2. Geological framework of Chilean Andes crustal faults

The oceanic Nazca plate subducts beneath the continental South American plate along almost all of the latter's western margin, in a compressional setting defined by Uyeda and Kanamori (1979) as "Chilean type". In Chile, the contact between the two plates runs offshore in a roughly north-south orientation for the majority of the country's length of more than 4,000 km, from the Arica elbow in the north to the Chile ridge subduction site at the Nazca-South America-Antarctica Triple Junction (latitude 46° S, Fig. 1). The subducted plate dips ~25-30° E beneath South America, except in the central "flat slab" segment, between 28° and 33° S, where it dips ~5° E with a convergence oblique to the margin (N78° E) at a rate of 66-79 mm/yr (Cande and Leslie, 1986; DeMets *et al.*, 1994; Angermann *et al.*, 1999) in the strongest compressional-transpressional regime along the Andean margin (Ramos, 2009a).

South of 46° S until ~52° S the subducting oceanic plate beneath South America is the Antarctic plate, having a convergence rate of ~20 mm/yr and a perpendicular to the trench convergence direction from the early Eocene (Cande and Leslie, 1986).

The southernmost tectonic margin of Chile curves to the east and is characterized by a transform boundary between the South American and Scotia plates, whose continental segment is known as the Magallanes-Fagnano fault. This portion of the plate boundary has a left-lateral strike-slip motion estimated between 6.6±1.3 mm/yr to 9.6±1.4 mm/yr (Smalley *et al.*, 2003; DeMets *et al.*, 2010). To the west and southwest of this margin, the Antarctic plate subducts beneath the South American plate at a rate of 20-24 mm/yr (Cande and Leslie, 1986; DeMets *et al.*, 1990), and beneath the Scotia plate at a rate of 11-13 mm/yr (Pelayo and Wiens, 1989; Thomas *et al.*, 2003).

The convergence between the Nazca and South American plates has caused crustal shortening and thickening of the overriding (SA) plate, resulting in the formation of the modern Andes mountain range over the past 25 Ma (Mpodozis and Ramos, 1989; Tassara and Yáñez, 1996; Yáñez and Cembrano,
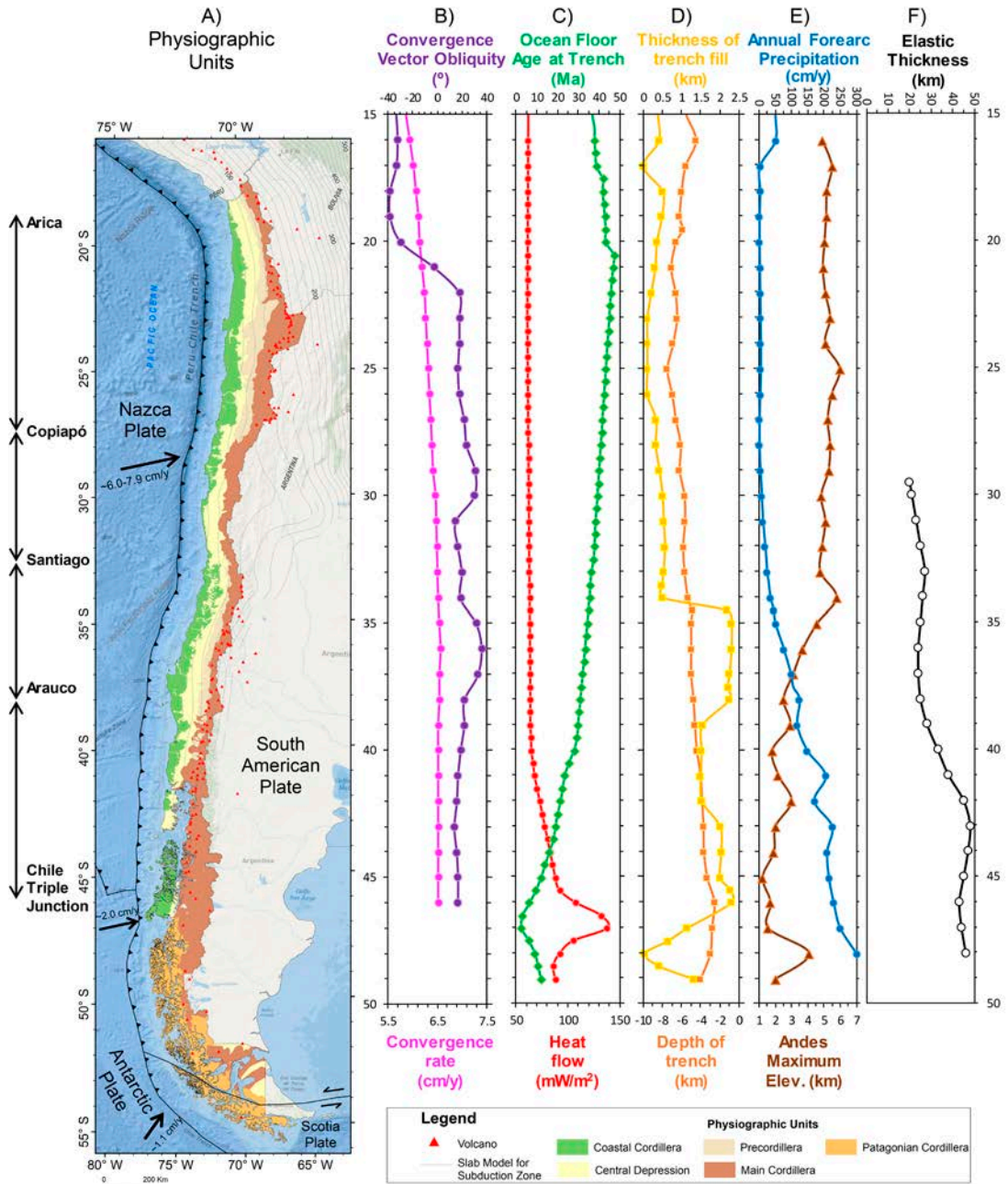
FIG. 1. **A.** Morphotectonic context and segmentation of the Chilean Andes. Main physiographic units of the Chilean Andes: Coastal Cordillera, Central Depression, Precordillera, Main Cordillera and Patagonian Cordillera (compiled and modified from Mpodozis and Ramos, 1989; Cembrano *et al*., 2007; Riquelme *et al*., 2007; Mpodozis and Cornejo, 2012; Armijo *et al*., 2015). Red triangles symbolize the present-day active volcanoes. Grey lines show depth in km of the SLAB-1 model (Hayes *et al*., 2012) representing the converging plates interface from the trench (black thick line); The graphs show **B.** Convergence vector obliquity and convergence rate (Angermann *et al*., 1999); **C.** The reducing age of the subducting plate from north to south (Müller *et al*., 1997) and heat flow from Stein (2003); **D.** Trench maximum depth in brown diminishing N to S (Lindquist *et al*., 2004) and its sediment filling depth in yellow (Bangs and Cande, 1997; Hampel *et al*., 2004) that increases from 33° S to south; **E.** Average annual precipitation along the Chilean fore-arc (New *et al*., 2002) and topographic elevations of the Andes (Lindquist *et al*., 2004); and **F.** The variation of elastic thickness (Tassara and Yáñez, 2003).

2004; Mpodozis and Ramos, 2008; Ramos, 2009a). The study of the present and past tectonics of the Andes exposes the marked changes that occur along both north-south and east-west transects. First-order tectonic segments have been identified in both the Nazca and the South American plates, ranging from continental scale (*e.g.,* Isacks and Barazangi, 1977; Jordan *et al*, 1983a; Jordan *et al*, 1983b; Jarrard, 1986; Yáñez and Cembrano, 2004; Hoffmann-Rothe *et al*., 2006; Ramos, 2009a) to local scale (*e.g.,* Rosenau, 2004; Rosenau *et al*., 2006; Rehak *et al.*, 2008), and trending both along- and across-strike. Several factors control this tectonic segmentation, most significantly the absolute motion of the overriding plate and slab retreat, the length of the Benioff zone, the relative plate convergence rate, the direction of mantle flow, climate, the crustal thickness of the South American plate, the age and subduction angle of the Nazca plate, and the subduction of passive and/or active ridges, among others.

The causes of this first-order segmentation are still a subject of debate. Tassara and Yáñez (2003) propose a first-order segmentation caused by a change in the composition of the continental crust from north to south (more felsic in the central Andes (15-33.5° S) against a more mafic crust in the southern Andes (33.5-47° S), and a second-order segmentation associated with variations in the thermo-mechanical regime of the plate convergence process. Yáñez and Cembrano (2004) postulate that the arc-foreland deformation is controlled by the absolute velocity of the continental plate and the degree of plate coupling between the oceanic and continental plates, coupling subordinate to the age of the subducting plate and on the convergence rate. Ramos (2009a) indicates that the absolute motion of the upper plate relative to the hotspot frame and the consequent trench rollback velocity are the first order parameters that control the arc-foreland deformation.

Along with fundamental tectonic causes, marked long-term latitudinal climatic variations have been invoked as controlling factors accounting for the geological evolution and segmentation of the continental plate (*e.g.,* Jordan *et al*., 1983a, b, 2001, 2014; Jarrard, 1986; Mpodozis and Ramos, 1989; Kley *et al*., 1999; Gutscher *et al*., 2000; Yáñez *et al*., 2001; Lamb and Davis, 2003; Tassara and Yáñez, 2003; Yáñez and Cembrano, 2004; Sobolev and Babeyko, 2005; Armijo *et al*., 2015). Climatic characteristics vary both along and across the Chilean

Andes. Precipitation increases from a hyperarid north to a cold and rainy south, exerting a critical control on the sediment contribution to the trench (Dunai *et al*., 2005; Hoke, 2006; Kober *et al*., 2007; Matmon *et al*., 2009), which is almost empty of sediment fill between 10° to 33.5° S. Hyperaridity has likely persisted since at least *ca*. 12±1 Ma (Jordan *et al*., 2014) and has been recognized as a key factor on the formation of the Andes (Lamb and Davis, 2003; Jordan *et al*., 2014). In addition, the Perú-Chile oceanic current and the associated upwelling, induces colder oceanic waters and thus limits precipitation in its vicinity (Lamb and Davis, 2003).

A latitudinal segmentation (across-strike relative to the continental margin/Chilean trench) of the Chilean continental lithosphere into at least five different tectonic domains has been recognized by various authors: Arica-Copiapó, Copiapó-Santiago, Santiago-Arauco, Arauco-Chile Triple Junction, and South of Chile Triple Junction (*e.g.,* Jordan *et al*., 1983a, b; Tassara and Yáñez, 2003; Yáñez and Cembrano, 2004; Ramos, 2009b; Gerbault *et al*., 2009). The main tectonic characteristics of these domains are shown in table 1 and figure 1.

There is also a longitudinal segmentation into three main tectonic domains aligned parallel to the Chilean coastal margin: namely, the Outer-Forearc (OF), Inner-Forearc (IF) and Volcanic Arc (VA). These domains are associated with the main physiographic units of the Chilean Andes: The Coastal Cordillera, Central Depression and Main and Patagonian Cordillera, respectively (Pankhurst and Hervé, 2007). These units correspond broadly to the seismic regionalization proposed by Barrientos (1980) along Chile, which is used by the Chilean Earthquake-resistant design of industrial structures and facilities standard (NCh 433) that only accounts for the subduction earthquake hazard.

The aforementioned tectonic segmentation results in diverse deformation styles and subduction convergence rates along the length of Chile, as exemplified by the different fault systems (compressional, transpressional, normal, strike-slip) and the strain partitioning across the Chilean Andes, which have been active within the same converging regime since the late Cretaceous, and none of which has covered the entire 3,500 km length of the range (Hoffmann-Rothe *et al*., 2006). Inherited structural anisotropy and current geodynamical conditions further complicate the understanding of the seismogenic

TABLE 1. LATITUDINAL TECTONIC SEGMENTATION OF CHILEAN ANDES.

Segment	Continental margin trend	Subduction Angle (°)	Crustal Thickness	Elastic Thickness	Coupling	Volcanic arc (Actual)	Cenozoic Andean shortening	Trench depth (km)	Trench sediment fill (km)
Arica-Copiapó	N-S	Normal ~25-30°E	thick up to 70 km	low <10 km in arc	high	active	300-200 km	>7.0	-
Copiapó-Santiago	N-S	Flat ~5°E	up to 60 km	very low 0-5 km in arc	-	non active	200-100 km	6.0-6.5	<0.5
Santiago-Arauco	NNE	Normal ~25-30°E	high to normal < 60 km	medium to high <10 km	medium	active	100-50 km	5.0-6.0	0.5-1.0
Arauco-Chile Triple Junction	N-S	Normal ~25-30°E	normal to low <40 km	high >35 km	medium to low	active	<50 km	<5.0	>1.0

Data compiled from: Jordan *et al*., 1983a, b; Tassara and Yáñez, 2003; Yáñez and Cembrano, 2004; Asch *et al*., 2006; Tassara *et al*., 2007; Ramos, 2009b; Gerbault *et al*., 2009.

potential of crustal faults in Chile, considering the subduction as a primal driving source.

The complicated interaction between the Nazca and South American plates has also placed scientific challenges in other Andean countries such as Peru and Ecuador, where the relationship between subduction and crustal seismicity is only just beginning to be understood (*e.g.,* Machare and Ortlieb, 1992; Paris *et al*., 2000; Machare *et al*., 2003; Allmendinger and González, 2010; Costa *et al*., 2010; Alvarado *et al*., 2014; Yepes *et al*., 2016; Alvarado *et al*., 2016).

3. Use of geological and paleoseismological data in seismic-hazard analysis: issues and challenges

The amount, repeatability and age of last movement on a fault are often included under the concept of fault "activity" and its related hazard. However, the concept of fault activity has proved to be highly controversial and misleading since it has no agreed definition despite being widely used (Costa *et al.*, 2006a). For example, an "active fault" can refer to one which has undergone slip in either Historic, Holocene or Quaternary times depending on the author (*e.g.,* Wallace, 1986; Yeats *et al*., 1997; Aki and Lee, 2003; McCalpin, 2009; USGS, 2018b among others). Still other authors define an active

fault as one "that may have movement within a future period of concern to humans" (Wallace, 1981), or else fail to define the term at all.

In order to assess the seismic hazard posed by a given fault, it is first necessary to identify whether that fault is capable of generating an earthquake. The seismogenic potential of a fault -that is, the size and recurrence rate of the earthquakes it produces- may be clearer and more tectonically significant when that fault is viewed in the geodynamic context of the subduction earthquake cycle. For example, inactive crustal faults which are optimally oriented with respect to inter- or co-seismic stress fields may be reactivated and produce shallow earthquakes with magnitudes of up to Mw 7 (*e.g.,* Farías *et al*., 2011; Aron *et al*., 2013; González *et al*., 2015). This scenario is particularly relevant for optimally-oriented basement (inherited) crustal faults undergoing sudden high strain rates, a condition which can potentially be achieved co-seismically with large 'megathrust' subduction earthquakes (Mw>8, King *et al*., 1994; Stein *et al*., 1994; Stein, 1999; Kilb *et al*., 2000; Lin and Stein, 2004; Loveless *et al*., 2010; Seeber and Armbruster, 2000). In contrast, during the inter-seismic period, when the crustal deformation rate is relatively low, the reactivation of optimally-oriented faults may be less likely (*e.g.,* Aron *et al*., 2014).

The estimation of earthquake magnitudes from geological data is typically performed using empirical relationships and assumptions of earthquake parameters based on the premise that the rupture parameters of an earthquake are proportional to its magnitude (Slemmons, 1977; Hanks and Kanamori, 1979; Bonilla *et al*., 1984; Slemmons and de Polo, 1986; Wells and Coppersmith, 1994; Hanks and Bakun, 2008; Wesnousky, 2008; and many others summarized in Stirling *et al*., 2013). Accordingly, the rupture length and surface displacement which are preserved in the paleoseismological record can be used in empirical regression relationships to estimate earthquake magnitude. The uncertainties of these parameters must be considered, and the assumption that a fault will always rupture with a constant displacement and slip rate thus creating a "characteristic earthquake" must be used with caution (Kagan, 1993; Burbank and Anderson, 2001; McCalpin, 2009). Accordingly, the assessment of the key parameters in the field, and the subsequent estimation of other parameters using empirical relationships, are subject to significant uncertainties (see McCalpin, 2009 for a review).

Models of fault behavior acknowledge that faults may be segmented into discrete sections that act distinctly over their seismic cycle (*e.g.,* Schwartz and Coppersmith, 1986; Slemmons, 1995). Such segmentation can control the onset and termination of fault rupture, and accordingly also the size of the rupture area and earthquake magnitude (Grant, 2002). Commonly, the length of a single fault segment (which may be identified through field observations and/or satellite imagery) indicates the maximum possible surface rupture length during an earthquake, because field features are formed by several ruptures over time. Furthermore, if an inappropriate work scale is chosen, several consecutive segments could be mistaken for a single one. Fault segments which have generated earthquakes independently of each other may additionally be capable of combining into a single segment rupture and thereby generate an earthquake of higher magnitude (*e.g.,* Estay *et al*., 2016).

Co-seismic displacement of faults is another difficult parameter to assess. Slip from colluvial wedge thickness formed coseismically in dip-slip faults (Nelson, 1992; McCalpin, 2009) assume that the colluvial wedge has not experienced erosion, represents a single event and is representative of the entire rupture zone slip. If all these assumptions are

met, the colluvial wedge height is considered to be half of the total slip displacement (McCalpin, 2009). Accordingly, paleoseismological studies should be conducted along the entire length of a fault in order to better represent its behavior. Even when all of this is taken into account, the estimation of earthquake size via paleoseismological or terrain analysis techniques introduces uncertainties which may be difficult to constrain.

4. Historical and instrumentally recorded crustal earthquakes

Seismic data is an essential tool with which to constrain and characterize plate tectonics. Major seismic catalogues, such as those maintained by the International Seismological Centre (ISC), National Earthquake Information Centre (NEIC), the ANSS Comprehensive Earthquake Catalog from USGS (2018a), Integrated Plate Boundary Observatory Chie (IPOC) and CMT-Harvard, include data from crustal earthquakes all over the Chilean Andes. Nevertheless, in several cases the hypocentral locations of these events are not sufficiently well constrained to distinguish shallow intra-crustal earthquakes from subduction earthquakes. This is particularly important for events recorded in the forearc, where seismicity in the overriding plate detected from teleseismic observations is often masked by the stronger interplate seismicity. However, catalogue data show a clear concentration of seismicity in the Chilean fore-arc region, in contrast with scarce intraplate crustal seismicity in the Main Cordillera of the Andes (*e.g.,* Pardo *et al*., 2002; Asch *et al*., 2006). Along the length of the Central and Southern Andes, intraplate seismicity tends to be localized within the weak lithosphere, where the brittle-ductile transition is located at less than 15 km depth (Tassara *et al*., 2007).

There are relatively few published studies documenting crustal earthquakes in Chile (*e.g.,* Pardo *et al*., 2002; Barrientos *et al*., 2004). In recent years, the Pichilemu Mw 7.0 earthquake (which struck 12 days after the 2010 Maule interplate mega-earthquake) and the Pisagua Mw 6.7 earthquake (16 days before the 2014 Iquique interplate earthquake) have attracted the attention of geologists and seismologists. These events motivated several studies regarding the nature of these events and their tectonic significance in the subduction cycle, due to their spatial proximity to

the trench and the fact that they were temporally associated with Mw 8.8 and Mw 8.1 interplate earthquakes (*e.g.,* Farías *et al.*, 2011; Aron *et al.*, 2014; González *et al.*, 2015).

In order to assess the modern crustal seismicity of the Chilean Andes, we used the Harvard CMT catalogue (http://www.globalcmt.org/ (last visit 01/06/2015)), one of the most accurate global seismic event catalogues (Kagan, 2002). When hypocentral depth is undetermined, which is the case on more than 30% of historical-period earthquakes in the different catalogues, an hypocentral depth based on tectonic setting or nearby seismicity is assigned (Bondar *et al.*, 2015). The CMT catalogue assigns events a "fixed depth solution" with an assumed hypocenter located at 10, 15 or 33 km depth.

A search of the CMT catalogue for earthquakes located in the Chilean Andes, having hypocentral depths shallower than 50 km and magnitudes above Mw 4.9, and occurring in a 39-year period between January 1976 and April 2015, yielded 6.601 events. Information for each event included date, latitude, longitude, depth, half duration, moment tensor, magnitude metrics (Mw, Ms, mb, etc.) and fault plane solution. From this population we selected 372 earthquakes that were considered to be of crustal origin, by eliminating events that were determined to be subduction-related via analysis of their focal mechanisms, while also eliminating events with poorly determined hypocentral locations. These 372 events are spatially distributed along two bands, one in the coastal outer forearc, and one located inland in the volcanic arc (Fig. 2).

Several earthquakes from this population have been of scientific and/or public interest due to their large magnitude (Mw>6.0) and shallow hypocentral depth (<16 km). Epicentral locations of these events are distributed across several different latitudinal and longitudinal domains, and are good models of
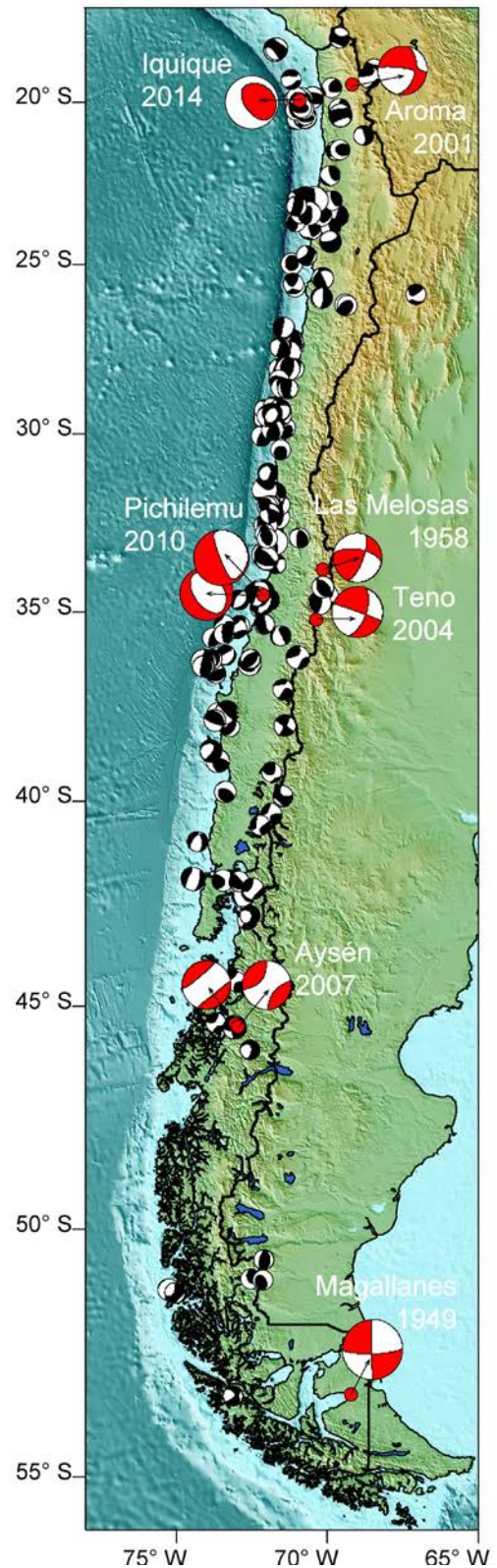

FIG 2. Crustal earthquakes recorded in the last 40 years. Location map of 372 crustal earthquakes along the Chilean Andes, with circle sizes corresponding to their magnitude. Focal mechanism of crustal earthquakes with Mw >=6 are displayed. Based on data from Harvard CMT catalogue between January 1976 and April 2015 (The Harvard Centroid Moment Tensor Catalog, http://www.globalcmt. org/ (last visit 01/06/2015)). Also, larger historical crustal events are highlighted: Aroma 2001, Iquique 2015, Las Melosas 1958, Pichilemu 2010, Teno 2004, Aysén 2007 and Magallanes 1949.

crustal earthquake types as explained in this section (summarized characteristics in table 2). Moreover, the hypocenters and focal mechanisms of some of the earthquakes show strong correspondence with the identified traces and slip characteristics of crustal faults in the region.

In this section, we detail some prominent crustal earthquakes (Mw>6.0), included in the crustal earthquakes selected from the CMT catalog (Fig. 2).

4.1. Aroma earthquake (2001), Inner Forearc Domain

This Mw 6.3 earthquake occurred on July 24, 2001 (Table 2, Fig. 2), and is one of the few shallow earthquakes to have occurred in northern Chile in recent years (Comte *et al.*, 2001). Legrand *et al.* (2007) used different seismic data to locate the event and its aftershocks. They additionally constrained the focal depth of the event using the SP phase and

its focal mechanism. Their analysis produced a fault plane solution of the main event with strike 14°±10°, dip 53°±15° and rake 163°±15°, indicating dextral movement on an oblique-slip fault. This is consistent with the distribution of aftershocks, which also indicated a fault direction of N14° E and dip 50° E. Additionally, Legrand *et al.* (2007) determined the earthquake's hypocentral depth to be 5±1 km, although neither surface deformation nor related morphotectonic features were reported.

4.2. Las Melosas earthquake (1958), Volcanic Arc Domain

On September 4, 1958 a seismic event was recorded in the Cordillera Principal of central Chile (Fig. 2), located near the intersection of the Maipo and Volcán rivers (Table 2, Fig. 2), with a hypocentral depth of 15 km. No field evidence of surface rupture associated with this event was observed (Sepúlveda *et al.*, 2008). The

TABLE 2. INSTRUMENTALLY RECORDED CRUSTAL EARTHQUAKES IN CHILE.

Location	Year	Magnitude Mw	Epicentral Location			Reference	Longitudinal Tectonic Domain
			Latitude °S	Longitude °W	Depth km		
Aroma	2001	6.3	19.5	69.2	~5.0	Legrand *et al.*, 2007	Inner Forearc
Las Melosas	1958	6.3	33.9	70.2	5.0-9.0	Alvarado *et al.*, 2009	Arc
		6.9-6.7-6.8				Sepúlveda *et al.*, 2008	
Pichilemu	2010	6.9	34.5	72.1	~12.9	Farías *et al.*, 2011	Outer Forearc
		7.0			16.3	Aron *et al.*, 2013	
Teno	2004	6.5	34.9	70.6	4.7	González, 2008	Arc
Aysén	2007 Apr 21	6.2	45.4	73.0	<8.0	Mora *et al.*, 2010	Arc
					4.0	Legrand *et al.*, 2011	
Aysén	2007 Apr 2	6.1	45.4	73.1	5.3	Legrand *et al.*, 2011	Arc

earthquake involved a sequence of three separate events, with surface wave magnitudes of between Ms 6.7 and 6.9 (Sepúlveda *et al*., 2008) and moment-magnitudes of Mw 6.3 (Alvarado *et al*., 2009). For the main event, the focal mechanism determined is consistent either with sinistral slip on a NS-oriented fault plane or dextral slip on a nearly E-W-striking fault plane (Lomnitz, 1960; Sepúlveda *et al*., 2008; Alvarado *et al*., 2009; Alfaro, 2011). Sepúlveda *et al*. (2008) analyzed the hazard associated with the Las Melosas earthquake, estimating intensities and Peak Ground Acceleration (PGA) values from the reinterpretation of historical documents and the study of landslides caused by similar earthquakes. The maximum intensity felt in the epicentral area was 9 (MSK scale), and 6 at a distance of 40 km. Furthermore, according to analysis using the Newmark method of two landslides triggered by this event, the authors estimated a PGA of between 0.58 and 1.30 g. These features highlight the large seismic hazard existing in the vicinity of this blind fault, close to the El Fierro Fault System (Farías *et al*., 2010).

4.3. Teno earthquake (2004), Volcanic Arc Domain

A magnitude Mw 6.7 earthquake occurred on August 28, 2004, in the Cordillera Principal of the Maule region near the headwaters of the Teno river, north of Peteroa volcano (Comte *et al*., 2008; González, 2008; Alfaro, 2011). According to the National Seismological Service of Chile, the hypocenter was located at 35.2° S, 70.5° W and 5 km depth, with a local magnitude of M_L 6.2. The Harvard CMT catalog gives a focal mechanism for this event indicating either dextral movement on the fault plane N21º E/61º E and rake -178º, or sinistral movement on the fault plane N70º W/88º N and rake -29º. González (2008) relocated the hypocenter at 34.9° S, 70.5° W and 4.7 km depth. Using aftershock earthquakes placed nearby this structure, she was able to delimit a N20º E/W-dipping, 18 km long and 11 km deep structure. The same author also suggests a northward spreading of the rupture, from the area of Termas del Flaco to the Maipo river valley, where a Mw 5.6 magnitude earthquake with reverse focal mechanism occurred on September 12, 2004. If we consider the focal mechanism reflecting a NNE-oriented structure and associated dextral fault movement, this earthquake would relate coherently

with east-vergent fault systems that reflect the tectonic control proposed by various authors in the south-central Principal Cordillera area (*e.g.,* El Diablo Fault in Fock, 2005; El Fierro fault in Farías, 2007).

The intensities recorded for the Teno earthquake reveal that only moderate to weak damage was caused in the area: a maximum intensity of VI on the Modified Mercalli scale, while in Talca and Curicó, intensities of V to VI (National Seismological Service) were recorded.

4.4. Aysén earthquake (2007), Volcanic Arc Domain

The biggest two earthquakes of a swarm of 7,200 events which occurred within a small area in the Aysen Fjord region in 2007, were the Mw 6.1 and Mw 6.2 earthquakes of April 2007. These events triggered a landslide which induced a tsunami that killed 10 people (Mora *et al*., 2010; Legrand *et al*., 2011). The hypocentral depths of the two events have been estimated at less than 12 km depth, by Legrand *et al*. (2011), and at 8 km depth by Mora *et al*. (2010). Fault plane solutions indicate an almost NS-striking, steeply-dipping nodal plane, consistent with long- and short-term Liquiñe-Ofqui Fault System (LOFS) tectonics, similar to the other events in the swarm. Such earthquakes have also been of interest due to its combined tectonic and volcanic origin (Legrand *et al*., 2011).

The locations of these earthquakes coincide with the Punta Cola fault (PCF) and with the location of the submarine rupture observed in subsurface profiles. (Vargas *et al*., 2013; Villalobos-Claramunt *et al*., 2015). The Punta Cola Fault, a main branch of the LOFS, is a 15-20 km long, N-S-striking, structure that crosses the Aysén Fjord about 5 km to the east of the main Liquiñe-Ofqui fault (Legrand *et al*., 2011),

4.5. Pichilemu earthquakes (2010), Outer Forearc Domain

A series of crustal earthquakes were felt in the area of Pichilemu for more than a year following the Maule (Mw 8.8) subduction earthquake that affected central Chile on 27 February 2010 (Arriagada *et al*., 2011; Alfaro, 2011; Farías *et al*., 2011).

The magnitudes of the main events in this series, which occurred in March 11 2010, were Mw 6.9 and 7.0 (Farías *et al*., 2011). However, significant

aftershocks of magnitudes Mw 6.4, 6.1, 6.0 and 5.9 were also recorded. Both the aftershocks and the largest earthquake had normal fault plane solutions (Harvard CMT catalog).

The Mw 6.9 and 7.0 earthquakes were located at ~34.3° S, 71.9° W, had hypocentral depths of 12 and 20 km respectively, and had normal type focal mechanism solutions with fault planes N144º E/ 55º SE and N155º E/74º S (Ruiz *et al*., 2014). The Mercalli intensities of the earthquakes recorded by the ONEMI were VII in Rancagua and VI in Talca while the USGS records intensities of IX near Pichilemu town (Alfaro, 2011).

The earthquake series and its aftershocks correlate spatially with a NW-oriented normal fault, dipping to the SW from the surface to the interplate boundary. Equally oriented extensional fault structures in the region are shown in the 1:1,000,000 scale Chilean geologic map (Sernageomin, 2003).

The 2010-2011 earthquake series has been proposed to have been caused by the large change in Coulomb stress caused by the Maule earthquake (Farías *et al*., 2011; Aron *et al*., 2013). These authors link the Pichilemu earthquake with other neotectonic activity, in agreement with the short-term deformation, and suggest that this event might be closely related to the long-term subduction seismic cycle: extensional faulting during coseismic subduction earthquakes and thrusting events in the interseismic period.

5. Neotectonic crustal faults

The territory of Chile hosts several prominent crustal faults which have undergone neotectonic activity, and which are predominantly spatially and/or temporally associated with strike-slip fault systems formed by the strain partitioning since at least the middle Jurassic (Hoffmann-Rothe *et al*., 2006). In this work, we reassess and compile over 1,600 faults from existing publications (Fig. 3). The basis of our work included the compilations of Lavenu *et al.* (2000), Lavenu (2005), PMA (2009) and SARA (2016).

The main neotectonic crustal faults and fault systems in Chile are spatially distributed across the three main longitudinal tectonic domains that are placed parallel to the Chilean coastline: Outer-forearc (OF), Inner-forearc (IF) and Volcanic arc (VA). These domains are additionally crossed by a series of long-lived transverse faults (T). Previous research has primarily

focused on geomorphological and tectonic analysis with relatively very few paleoseismological studies. More than 1,000 fault structures regarded as having neotectonic activity but without a rigorous assessment of basic fault parameters are currently under study. This lack of data is largely due to the fact that neotectonic faults are spatially associated with older tectonic structures, making it difficult to separate ancient from recent activity. In the current work we characterize 141 faults using various fundamental parameters for seismic hazard assessment, by compiling information from well-studied faults (Fig. 3).

5.1. Outer Forearc Domain

The outer forearc master faults strike predominantly N-S and exhibit concave-to-the-west morphologies (*e.g.,* the three segments of the Atacama Fault System, Arabasz, 1971). Regional-scale fault segments are steeply-dipping and have lengths of up to several hundreds of km. These structures commonly display evidence of early strike-slip to late dip-slip displacement. NW-striking subvertical secondary faults splay off the master faults and show similar kinematics. The most relevant faults of the outer forearc domain are the following:

5.1.1. *Atacama Fault System*

The Atacama Fault System (AFS) is a *ca.* 1,000 km -long fault zone running along the Coastal Cordillera of northern Chile from Iquique (20° S) to La Serena (30° S). The AFS has been the focus of several studies since the late 1960s due to the prominent geomorphological characteristics that it displays, reflecting a displacement history dating to the Jurassic. Three main segments of the AFS have been identified, all of them curved and concave to the west: Salar del Carmen, Paposo and El Salado (Okada, 1971; Arabasz, 1971; Naranjo, 1987; Hervé, 1987).

The earliest AFS activity is evidenced by mylonitic and cataclastic rocks formed under sinistral displacements in the volcanic arc as result of oblique convergence in the Jurassic-early Cretaceous (Scheuber and Andriessen, 1990; Brown *et al*., 1993).

Progressive cooling of the magmatic arc in the mid-Cretaceous may have coincided with the AFS deformation changing from ductile to brittle sinistral strike-slip (Brown *et al*., 1993). The structures generated in this brittle phase of the AFS are duplex-type, limited by NS- and NW-striking
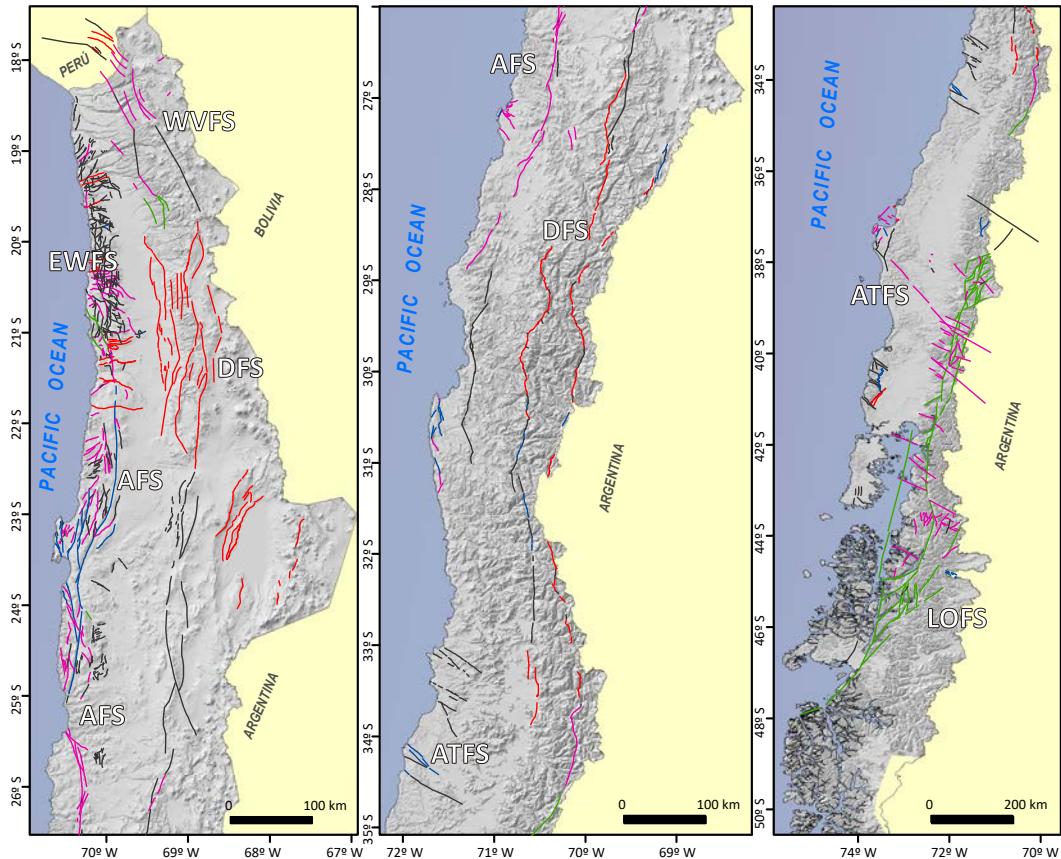
FIG 3. Compilation of Crustal Faults in the Chilean Andes with evidence of neotectonic activity. Normal (blue), Reverse (red), Strike slip dextral (green), Strike-slip Sinistral (magenta) and undetermined (black). Many of them belong to fault systems such as Atacama (AFS), East-West Transversal (EWFS), West Vergent Thrust (WVTS), Domeyko (DFS), Liquiñe-Ofqui (LOFS) faults systems and AndeanTransverse faults (ATF). Only 141 faults have specific studies that allow for a correct categorization as neotectonic faults.

strike-slip faults, segmented into numerous sections each in excess of 60 km in length (*e.g.,* Cembrano *et al.*, 2005).

Several authors have documented regional-scale sinistral and locally dextral strike-slip displacements along various segments of the AFS. However, the main Neogene-to-recent displacements are normal-style, consistent with the extensional regime of the present day forearc location (Arabasz, 1971; Okada, 1971; Hervé, 1987; Naranjo, 1987; DeLouis *et al.*, 1998; Scheuber and González, 1999; González *et al.*, 2003; Riquelme *et al.*, 2003; Cembrano *et al.*, 2005; González *et al.*, 2008; Vargas *et al.*, 2011; Allmendinger and González, 2010; Cortés *et al.*, 2012). Quaternary deformation of the AFS is preserved in the horst-and-graben and hemi-graben

coastal topography, developing fault scarps from a few meters to over 100 m in height.

Seismic activity had not been definitively associated with any AFS fault segments until the study of Metcalf and Kapp (2015). These authors compiled crustal seismicity data from existing catalogues, and found extensional focal mechanisms in the forearc region, consistent with the distribution of mapped active fault scarps. This interpretation, however, should be viewed with caution considering that the seismic data used was not obtained from local networks.

One of the best-known AFS fault segments is the Salar del Carmen fault (SCF) that develops near Antofagasta (23° S) as a N15° E-striking, 75° SE dipping normal fault. Several displaced ~400 ka old alluvial fan sediments, forming 3 to 5 m-high east

dipping scarps, reflect the most recent deformation along the AFS (González *et al*., 2006; Cortés, 2012). These authors constrain the age of the youngest displacement to the late Pleistocene. More recent work, including paleoseismological trenching and Optically Stimulated Luminescence (OSL) dating on colluvial sediments, indicate that SCF has generated at least two earthquakes in the last *ca*. 5 ka, with ~2.2 m of fault slip for each event. By using Wells and Coppersmith relationships (1994) it has been estimated that each earthquake had a Mw ~7 (Cortés, 2012).

5.1.2. Mejillones Fault

The Mejillones Fault (MF) is a 40 km long (30 km onshore, 10 km offshore to the north) N10° E-striking, 75° SE- dipping normal fault located in the northern part of the Mejillones Peninsula, 60 km NW from Antofagasta city. Activity on this fault has been established since at least 3.3 My, forming a conspicuous scarp, ~500 m in height at its northern termination and progressively lowering southward to ~65 m height (Marquardt, 2005). Its most recent activity is expressed in the deformation of the piedmont area, which formed an escarpment in late Pleistocene alluvial gravels (Marquardt *et al*., 2004; Cortés *et al*., 2012). This escarpment has an approximate length of 12 km and reaches a height of 3 to 10 m.

The MF surface rupture pattern indicates that the fault consists of at least two segments, connected at depth (Cortés *et al*., 2012). By combining morphological observations, mapping of displaced alluvial surfaces, ^{10}Be dating, OSL dating of colluvial deposits at the foot of the escarpment, and paleoseismological trench mapping, it has been determined that this fault has generated three earthquakes of Mw ~7 between *ca*. 35 ka and *ca*. 14 ka, as well as possibly an earthquake of Mw ~ 6.6 more recently (*ca*. 3 ka). These results also indicate a slip rate of 0.61±0.26 mm/yr, during the first three intervals of the fault evolution, and 0.22±0.06 mm/yr during the fourth interval. The displacement involved in each of these events has been estimated to range from 0.7 to 2.2 m, distributed along rupture surfaces ~40 km in length and 23 km in width. Recurrence times for Mw ~7 sized earthquakes on the MF have been estimated at 5.0±3.5 ka (Cortés *et al*., 2012).

5.1.3. Caleta Herradura Fault

The Caleta Herradura fault (CHF) is located in the southwestern half of the Mejillones Peninsula

(Arabasz, 1971; Okada, 1971). Its morphological expression, an escarpment in Quaternary alluvial gravels, suggests that this is a normal fault with conspicuous tectonic activity. The CHF has a length of ~40 km, and five distinct fault sections can be identified. These have accumulated hectometric displacement (Espinoza, 2013; Marquardt, 2005). The state of preservation of the Quaternary scarps is similar to that of the Mejillones Fault, so a late activity of similar age from Pleistocene to Holocene times is inferred (Cortés-Aranda *et al*., 2015). Ongoing research in neotectonic activity on the fault suggests a transfer of deformation between the Mejillones and Caleta Herradura faults, likely through a connection between the two at around 15 km depth.

5.1.4. E-W Fault System

One of the most distinctive features in the Coastal Cordillera of northern Chile is a series of EW-oriented topographic highs, located between 19º to 22º S latitude, that are spatially associated with eroded fault- and fold-scarps forming the E-W Fault System (EWFS). These scarps are between 20 and 300 m in height, contain open fractures in 20 to 1600 m-wide zones, and extend from the coastline to the eastern side of the Coastal Cordillera, ending in fault-propagation folds affecting the Central Depression Tertiary infill (Allmendinger *et al*., 2005; González *et al*., 2008). The scarps are associated with reverse faults oriented sub-perpendicular to the N-S coastal line, with dip-slip displacement accommodating parallel to the trench shortening (Allmendinger *et al*., 2005; González *et al*., 2008; Allmendinger and González, 2010). Some of the scarps are related to emergent reverse faults while others result from surface folding by blind reverse faults.

Morphotectonic and geochronologic studies conducted at a local scale, as well as regional-scale mapping, indicate that the E-W-oriented scarps formed in the late Miocene to early Pliocene, and that activity persisted into the Quaternary (Allmendinger *et al*., 2005). The scarps blocked the ancient drainage system, leading to the formation of evaporite deposits. The authors suggest that climatic conditions in the Coastal Cordillera were probably wetter during the late Miocene-Pliocene than during the present day. Their results further indicate that the Coastal Cordillera uplifte is younger than mid-Miocene and/or that the Coastal Escarpment has retreated substantively to the east (Allmendinger *et al*., 2005) and also suggest

that the origin of the EWFS is linked to the curved continental margin of the Bolivian Orocline, due to its limited latitudinal extent. Concurrently, the EWFS may be associated with north-trending shortening resulting from the oblique convergence between the Nazca and South American plates (*e.g.,* Allmendinger *et al.*, 2005; Carrizo *et al.*, 2008).

The Bajo Molle Fault (BMF), located to the south of Iquique, is one of the nearly EW-striking transversal faults exposed in the coastal active cliff area with a N80° E/30° NW orientation. The most recent activity on this fault is evidenced by a fault propagation fold gentle scarp, placing Jurassic volcanic rocks above late Pleistocene marine deposits (Allmendinger and González, 2010). OSL dating on deformed marine deposits suggests that at least one Mw 7.0 earthquake has occurred in the last 10 ka (González *et al.*, 2015, Supplemental Information). These authors estimated an uplift rate of 0.38±0.02 mm/yr, a horizontal shortening rate of 0.135±0.015 mm/yr and a net slip rate of 0.4± 0.1 mm/yr. The rupture displacement calculated for that event is 0.3 to 2.0 m distributed between the main fault and its branches. Depth of the earthquake source obtained by two methods are ~14 km (González *et al.*, 2015, Supplemental Information).

5.1.5. Chomache Fault

The Chomache Fault (ChF) is a strike-slip fault, aligned N42° W/78° NE, and located at 21°10' S in the vicinity of Salar Grande. The main evidence of recent activity on this fault is the dextral offset of channels which are developed in Quaternary alluvial surfaces. This deformation pattern occurs along the three main segments of the ChF with a total length of 20 km (González *et al.*, 2003; Masana *et al.*, 2005; Carrizo *et al.*, 2008; Allmendinger and González, 2010). OSL dating on alluvial/colluvial deposits and paleoseismological trench mapping suggest that at least two seismic events of Mw ~7 have occurred along this fault during the last *ca.* 4 ka. The fault displacement for these events has been estimated at between 3 and 6 m, over a rupture area with a length of around 20 km, and a width of no less than 20 km. The time between both events is ~2.1 ka, which gives some indication of the recurrence time of Mw ~7 earthquakes in ChF.

5.1.6. Pichilemu Fault

Two earthquakes, one of Mw 6.9 and one of Mw 7.0, occurred 12 days after the 2010 Mw 8.8

Maule megathrust earthquake. These nucleated on the Pichilemu Fault (PF), an upper plate normal fault (Aron *et al.*, 2013). These events showed that the rupture planes have an orientation of N145° E/55° E and N155º E/74º E respectively, and extend down to the plate interface, which lies approximately at 33 km depth in this region (Farías *et al.*, 2011; Ryder *et al.*, 2012; Aron *et al.*, 2013; Ruiz *et al.*, 2014). Ryder *et al.* (2012) and Ruiz *et al.* (2014) suggest that both earthquakes were related to a part of a typical normal fault diverging splay, where the secondary fault connects to the main fault at depth. InSAR and GPS measurements show that ~0.6 m of vertical offset occurred during the main Mw 7.0 event, and finite slip inversions suggest maximum normal-fault displacements of ~3 m (2.4 m along the vertical component) at 10-11 km depth (Ryder *et al.*, 2012).

The main Pichilemu earthquake fault plane has been spatially associated by Aron *et al.* (2014) with a 141°-striking, SW-dipping normal fault trace mapped in the 1:1.000.000 Chilean geologic map (Sernageomin, 2003). Several studies have concluded that there was no surface rupture associated with the 2010 Pichilemu earthquakes (Arriagada *et al.*, 2011; Farías *et al.*, 2011; Aron *et al.*, 2014). Nevertheless, Aron *et al.* (2014) state that the geological relations found in the field, such as a record of Pleistocene-Holocene (?) uplift of paleo-beach sequences in the footwall, indicate that the Pichilemu fault has repeatedly ruptured the surface in the past and has moved as a normal fault.

5.2. Inner Forearc Domain

The inner forearc fault systems exhibit a general NS strike, a sinuous shape and splay faults to both the west and east (*e.g.,* Domeyko fault system, Reutter *et al.*, 1991, 1996). The main fault segments reach tens of kilometers in length, have subvertical dips, and exhibit both dextral and sinistral strike slip movements. There are also reverse faults, such as the West-Vergent Thrust System in northern Chile (Muñoz and Charrier, 1996) and the San Ramón Fault in Central Chile (Vargas *et al.*, 2014).

5.2.1. Domeyko Fault System

The Domeyko Fault System (DFS) is an anastomosed system of mostly NS-striking, steeply dipping faults comprising a complex array of strike-slip, normal, and reverse faults, together with

thin- and thick-skinned folds and thrusts. The DFS is traceable parallel to the Chilean trench for more than 1,000 km and has a width of between 40 and 60 km (*e.g.,* Reutter *et al*., 1991, 1996; Cornejo *et al*., 1997; Lindsay *et al*., 1995; Mpodozis and Cornejo, 2012). The DFS consists of several regional-scale segments, each of which has undergone distinct deformation events at various times.

The DFS has been active since the upper Eocene-lower Oligocene (Reutter *et al*., 1996), and has contributed to the emplacement and mineralization processes of some of the largest porphyry copper deposits on Earth. Many authors have documented an initial dextral strike-slip movement in the Incaic tectonic phase, followed by sinistral deformation associated with lower convergence rates (*e.g.,* Reutter *et al*., 1996, Lindsay *et al*., 1995). The youngest deformation is evidenced by dextral strike-slip fault zones, the same kinematic regime that prevailed during the earliest Eocene-Oligocene phase. Tectonic inversion of this fault system is evidenced by dissimilar displacement amounts: dextral displacement of 0.5-2.0 km, and sinistral displacement of 35-37 km (Reutter *et al*., 1991; Tomlinson and Blanco, 1997a, b).

Seismic records from the Domeyko Cordillera at 21º S show that the present-day activity of the DFS includes, at least in part, both dextral- and reverse-faulting kinematics (Salazar, 2011; Bloch *et al*., 2014). Field observations in the Chuquicamata mine pit document active "en echelon" tension cracks, further supporting ongoing dextral displacement along traces of the DFS (D. Faulkner, T. Mitchell, E. Jensen, personal communication, 2011).

5.2.2. San Ramón Fault

Studies carried out along the San Ramón fault (SRF) associate this fault with the structural system that controls the morphotectonic boundary between the Central Depression and the Main Range in central Chile, at a latitude of approximately 33º S, near the eastern fringe of the Santiago de Chile metropolitan region (Farías, 2007; Armijo *et al*., 2010; Rauld, 2011). This west-vergent reverse structure is part of the larger West Andean Thrust system, where regional-scale deformation has migrated eastward (Farías, 2007). Surface observations show folded and faulted Eocene-Miocene Abanico Formation volcanoclastic rocks thrusting over Quaternary deposits (Rauld, 2002).

The surface morphology in the region of the SRF is a semi-continuous escarpment that extends north-south for at least 35 km, from the Mapocho river to the Maipo river (Armijo *et al*., 2010), decreasing its expression southward. Rauld (2011) applied a kinematic deformation model to the mountain front and piedmont, where a series of west verging kilometric folds affect rocks of the Abanico and Farellones (Miocene) formations. These structures were modeled by the fold-thrust-belt propagation to the west by four reverse faults, with 45-65º dip, attached to a 4-5º E dip detachment plane at 10-12 km depth.

Rauld (2011) quantitative study of deformation associated with fault scarps in the Andean piedmont (lower Pleistocene-Holocene deposits) and its morpho-stratigraphic relationship with deformed fluvial terraces of the Maipo and Mapocho rivers, shows fault scarps heights between 3 and 200 m along 3 segments. The terraces of these rivers, dated using OSL, are deformed by the SRF, suggesting Pleistocene incision rates of 0.5 to 1.0 mm/yr.

A more recent study using fault trenching and cosmogenic isotope data, documents two large earthquake ruptures (Mw 7.5) on the SRF within the past 8-19 ka; with fault displacements of ~5 m in each event (Vargas *et al*., 2014). Other research shows scattered present day seismic activity clustered at around 10 km depth 20-25 km east of Santiago (Pérez *et al*., 2014, crustal seismicity recorded between 2000 and 2011). According to Pérez *et al*. (2014), these data provide the first evidence of seismic activity consistent with the crustal-scale structural model proposed for the SRF system in the area (Armijo *et al*., 2010). Also, Pérez *et al*. (2014) model different rupture models for a Mw 6.9 earthquake in the San Ramón Fault, obtaining PGA maximum values in the order of 0.7-0.8 g.

Recently, Estay *et al*. (2016) defined four segments along the trace of the SRF, with a mean length of 10 km. They used time-integration methodologies to estimate the most likely scenarios of rupture and seismic event distribution along the fault, as well as Transient Electromagnetic (TEM) geophysics to constrain the fault dip angle (65°) and its capacity for surface rupture. By subsequently applying the empirical equation of Chiou and Youngs (2014) for crustal faults and considering the characteristic seismic event (high-angle thrust fault, 10 km length, Mw 6.2-6.7), the authors also provided estimates

for the PGA distribution in Santiago that would be caused by such an event, and thereby identified "hazard zones" which are capable of undergoing PGA in excess of 0.5 g.

5.2.3. West-Vergent Thrust System

The West-Vergent Thrust System (WVTS) is the largest tectonic feature in the precordillera of northern Chile. The system, of Miocene- Pliocene age, flanks the western reach Altiplano (Muñoz and Charrier, 1996), and has emplaced Precambrian rocks over Neogene rocks. Faults can be observed at the surface at 18° S, while further south they turn blind, and form surface flexures. The WVTS may be responsible for part of the Altiplano uplift (David, 2007, Farías *et al.*, 2005).

Seismic activity at the WVTS has been documented by David *et al.* (2002) and Legrand *et al.* (2007). This activity includes the Mw 6.3 Aroma earthquake of 2001, which had a dextral focal mechanism, similar to WVTS earthquakes reported further south and east by Salazar (2011). Although the Aroma earthquake did not develop a surface rupture, the position of its hypocenter has been correlated with a high angle reverse fault that underlies a regional-scale flexure at the western boundary of the Domeyko Cordillera (Farías *et al.*, 2005).

5.3. Volcanic Arc Domain

The volcanic arc faults exhibit roughly north-south strikes with horsetail-type terminations, subvertical dips and strike slip kinematics (*e.g.,* Liquiñe-Ofqui fault system; Cembrano *et al.*, 1996). These develop in the present-day volcanic arc region, where deformation partitioning has taken place from at least the late Miocene.

5.3.1. Liquiñe-Ofqui Fault System

The Liquiñe-Ofqui fault system (LOFS) is the most prominent seismogenic crustal fault in the Chilean Andes. It runs for more than 1,200 km from Copahue Volcano in the north (37° S) to Golfo de Penas in the south (47° S), approximately 150 km SE of the Chile Triple Junction. Early studies by Moreno and Parada (1976), Hervé (1976), and Hervé *et al.* (1979) recognized the LOFS only along its northern stretch. Later, Forsythe and Nelson (1985) identified and studied the southernmost segments of the LOFS, and postulated a causal link between the

Chile Ridge subduction and Cenozoic activity of the fault system. Field, structural and geochronological studies conducted in the nineties by Pankhurst *et al.* (1992), Cembrano and Hervé (1993), Cembrano *et al.* (1996), and Lavenu and Cembrano (1999) provided regional scale constraints on the geometry, nature and timing of ductile to brittle deformation along the fault system. The LOFS is currently regarded as a major dextral strike-slip duplex, associated with both Mio-Pliocene deformation and present day volcanism. In the last decade, more detailed structural and geochronological studies have documented Pliocene ductile-to-brittle transpressional deformation within a regional scale transpressional double-vergent system associated with high exhumation rates during the last 4 Ma (Adriasola *et al.*, 2006; Cembrano *et al.*, 2002, Thomson, 2002; Rosenau *et al.*, 2006).

Brittle fault segments, tens of kilometers in length, commonly overprint pre-existing ductile deformation zones along the LOFS, forming typical strike-slip fault geometries such as duplexes, Riedel-type splay faults, pull apart basins and tension cracks spatially associated with Holocene volcanoes. The master faults have an orientation N10° E, and subsidiary branches are mainly NE to ENE striking faults.

Teleseismic data and data from local seismic networks show that the LOFS is currently active as a dextral strike-slip fault (*e.g.,* Cifuentes, 1989; Lange *et al.*, 2008; Legrand *et al.*, 2011; Mora *et al.*, 2010; Agurto *et al.*, 2012; Agurto *et al.*, 2014). Recent seismic activity (Aysén Fjord) includes the April 2007 Mw 6.2 dextral strike-slip earthquake, with a hypocentral depth of 4 km, which was the last of a swarm (from January to June 2007) of thousands of lower magnitude quakes, including one normal dip-slip Mw 6.1 earthquake, with a hypocentral depth of 5.3 km, on a NE-striking splay fault (Legrand *et al.*, 2011). In this paper they re-located 10 events, obtaining hypocentral depths between 1 and 8.5 km. Maximum depths for tectonic seismic activity of 12 km are reported in Sielfeld *et al.* (2017).

However, the lack of paleoseismological studies, coupled with the scarcity of exposures of related Quaternary deformation, prevents a more detailed assessment of the Holocene displacement history, and thus the seismogenic potential, of the LOFS. Slip rates can only be roughly estimated from long-term geological data (Adriasola *et al.*, 2006) and short-term, regional-scale, GPS surveys (Wang *et al.*, 2007). Estimates fall in the range of 1-9 mm/yr,

decreasing northward (*e.g.,* Rosenau *et al*., 2006). The boundary element numerical model by Stanton-Yonge *et al*. (2016) suggests slip rates in the range of 2-7 mm/yr for the LOFS master faults. Iturrieta *et al*. (2017) finite element model results show slip rates of 5 to 10 mm/yr for the eastern master fault, whereas the western and subsidiary faults display rates of 1 to 5 mm/yr.

5.4. Other crustal faults

5.4.1. Andean Transverse Faults

The Andean Transverse Faults (ATF) are deep-seated lithospheric-scale structural elements that cross all margin-parallel tectonic domains. These structures display various alignments; the most conspicuous ones are the arc-oblique WNW-striking fault systems as defined by Pérez-Flores *et al*. (2016), based on the previous work of several authors (*e.g.,* Cembrano *et al*., 1996; Yáñez *et al*., 1998; Rosenau *et al*., 2006; Cembrano and Lara, 2009; Sánchez *et al*., 2013). ATF are well developed from the Andes main cordillera to the coastal cordillera between 33° and 46°S. The ATF comprises a series of discrete faults, with a predominantly NW strike, that are spatially or genetically related to either of the following features and/or processes: **1)** Regional basin boundaries (*e.g.,* Radic, 2010); **2)** Location of several ore deposits (*e.g.,* Chernicoff *et al*., 2002); **3)** Subduction earthquakes rupture zones (Melnick *et al*., 2009); **4)** Quaternary volcanism in the Central and Southern Andes (Cembrano and Lara, 2009; Cembrano and Moreno, 1994; Lara *et al*., 2004, 2006; Sánchez *et al*., 2013; Tibaldi *et al*., 2005). The ATF exhibits evidence of long-term fluid flow pathways, volcanic activity and shallow seismic activity (Haberland *et al*., 2006; Lara *et al*., 2006, 2004; Sánchez *et al*., 2013; Tardani *et al*., 2016). Segments of these faults are included in the longitudinal classification of this paper, as is the Pichilemu fault in the outer arc domain, because fault segments act as independent faults; there is no record of one entire fault within the ATF system with neotectonic displacement. Consequently, although ATF have prominent, long traces in surface by their behavior respond well to the domain characteristics.

Other prominent ATF are the Lanalhue Fault, which interseismic transpressional deformation in the forearc (38° S) include a M_L 5.2 earthquake at 12 km depth with a left-lateral focal mechanism (Haberland *et al*., 2006) and the Biobío-Alumine Fault System (Melnick *et al*., 2006; Muñoz and Stern, 1988).

The seismic activity observed with a temporal network in the Haberland *et al*. (2006) work in the Arauco Peninsula shows seismicity in the continental crust situated between 10 and 30 km depth. They associated those earthquakes, in most cases, with mapped faults, revealing that these faults are almost all forearc crust deep.

5.4.2. Magallanes-Fagnano Fault System

The 600 km-long Magallanes-Fagnano Fault (MFS) is a transform-type margin, and one of the major segments of the continental South America-oceanic Scotia plate boundary. The MFS runs from the western arm of the Magallanes Strait, where the South American-Scotia-Antarctic plates triple junction is located, to the offshore Atlantic. It transects the Isla Grande of Tierra del Fuego into two continental blocks (Lodolo *et al*., 2002, 2003; Cisternas and Vera, 2008; Civile *et al*., 2012).

Lodolo *et al*. (2003) describe the MFS as consisting of distinct tectonic lineaments that are segments of the transform system and predominantly consists of near-vertical faults. In their study of the asymmetric basin sediments they suggest a close relation between strike-slip motion and transform-normal extensions, a common feature found in other continental trans-tensional environments.

South of the South American-Nazca-Antarctic plates triple junction, the South American and Antarctic plates converge at a rate of *ca*. 1.9 cm/y, in a margin-perpendicular way in the north varying to strongly oblique in the south. To the south, the convergence rate is reduced to <1.2 cm/y since part of the convergence is accommodated by the MFS at a <0.5 to 0.7 cm/y rate (Lodolo *et al*., 2003; Cisternas and Vera, 2008). This convergence is almost perpendicular at 52° S, but rapidly becomes oblique to the south reaching 60° at 57° S.

In Magallanes, two important historical earthquakes of M_L=7.5; M_S=7.8 occurred, in 1879 and 1949 respectively (Lomnitz, 1970; Cisternas and Vera, 2008; Costa *et al*., 2006b). Recent seismicity, recorded in 1997 and 1998, of lower magnitude M_L=4.3, occurred within the continental crust. Furthermore, some very shallow concentrated earthquakes have occurred around the active Reclus and Burney volcanoes (Cisternas and Vera, 2008).

The main geomorphological expression associated with the MFS is an 11m-high scarp developed in late Pleistocene-Holocene (?) deposits. This suggests that there have been other earthquakes responsible for this feature because the vertical slip of the 1949 event was less than 1m. Paleoseismic analysis of this fault indicates two, and possibly three pre-1949 paleoearthquakes preserved in the stratigraphy (Costa *et al.,* 2006b).

More recent work by González-Bonorino *et al* (2012) found normal faults buried beneath gravelly strand plains on the Atlantic coast of Patagonia with displacement ages between 0.9 and 6.4 ka and a recurrence rate of about 1 ka. Their most likely coseismic source is the MFS, more than 300 km distant, which suggests high-magnitude earthquake activity on this fault system throughout the Holocene.

6. Discussion

One of the main aims of this work is to address how the seismogenic potential of neotectonic faults in Chile can be estimated. We have examined the main parameters involved in determining the magnitude of an earthquake, including length, width and mean displacement of crustal faults. We compiled the characteristics of the most significant crustal faults regarding their morphotectonic imprint and characteristic rupture/displacement segments revealed by their surface trace. This study highlights the similarity of neotectonic fault parameters along and across different tectonic domains, with longitudinal segmentation being dominant.

Hence, we collected crustal fault parameters and recorded crustal earthquakes in the corresponding longitudinal domain, in order to gather all the necessary parameters to estimate Moment Magnitude and check the consistency of those estimations. Length of fault rupture and slip during an earthquake are the general parameters applied for earthquake magnitude estimations (Bonilla *et al*., 1984; Wells and Coppersmith, 1994; McCalpin, 2009). We applied one of the most commonly-used formulae developed after Hanks and Kanamori (1979):

$Mw=2/3 (\log Mo-9.1)$
with $Mo=\mu A D$, where μ=shear modulus (=30 GPa in crust)
A=Length * Width=area
D=average displacement during rupture

Where the rupture plane is a km^2 area with 1 order of magnitude range (km to tens of km) and the average displacement ranges from centimeters to meters in scale, with, the latter parameter being the most sensitive to changes in the formulae.

This initial exercise constrains very well the order of magnitude of fault parameters, indicating which ones are key for the seismogenic hazard assessment.

6.1. Fault parameters for longitudinal domains

The seismogenic zone is where deformation processes are predominantly frictional and earthquakes can be nucleated and propagate as slip on preexisting fault surfaces or intact rock failure (Lockner and Beeler, 2002). The progressive transition from predominantly frictional deformation to a mainly quasi-plastic one, is commonly referred to as the base of this seismogenic zone or the brittle-plastic transition (*e.g.,* Scholz, 1988; Sibson, 2002), and bounds the maximum possible depth for an earthquake in continental crust.

Crustal heterogeneities, coupled with variations in regional heat flow, account for much of the observed variation in the depth of the seismogenic zone (Sibson, 1984; Smith and Bruhn, 1984). The main parameters controlling crustal heterogeneity are plate rheology, thermal gradient, and elastic and crustal thickness. Fault zone rheology is strongly affected by the quartz/feldspar ratio: the temperature of the transitional zone can change from T~350 °C for quartz rich material to T~450 °C where feldspar is the dominant load-bearing constituent within the rock mass. Complexity at the base of the seismogenic zone, however, may be further compounded by fluid pressure cycling near the transition zone (Sibson, 1992, 1994).

Numerous studies evaluating and modelling subducting plate structures and geometries in the Andes have been performed since the general recognition of the subduction processes in the 1960s. Most models postulate a large, predominantly east-verging ramp detachment thrusting structure connecting the subduction zone with the cordillera throughout the Andes (*e.g.,* Isacks, 1988; Farías *et al*., 2010; McQuarrie *et al*., 2005; Lamb, 2011, 2016; Armijo *et al*., 2015). These models depict a brittle and rigid forearc crust (high elastic thickness) and a weak ductile lower crust under the arc region (low elastic thickness), thus implying variable thicknesses

of the brittle-ductile transition from the subduction trench to the Andes Cordillera.

By analyzing seismicity and surface geology in Central Chile, Farías *et al.* (2010) predict a brittle-ductile transition of less than 10 km under the Main Cordillera, at ~25 km depth below the Central Depression and with the ramp detachment structure connecting with the subduction zone at ~60 km depth. Their results also suggest the presence of similar structures for northern and south-central Chile. It is important to highlight that the lower limit of crustal seismicity correlates well with the 400 °C isotherm (Fig. 4).

Arndt *et al.* (1997) suggest that the brittle-ductile transition occurs at depths of less than 20 km based on analysis of the thermal and elastic behavior of three granodioritic rocks from the Northern Chilean Coastal Cordillera. Ongoing thermal lithospheric modelling suggests that almost all of the crustal seismicity between 1996 and 2006 occurring at latitudes

of 33°-34° S is constrained in depth by the 350 °C isotherm (Yáñez *et al.*, personal communication, 2016). It follows that in the external forearc domain, earthquakes can be generated at any depth above the Benioff zone; from there to the arc, the isotherm shallows progressively from 30 to 15 km providing a rough estimate of the maximum depth for earthquake nucleation (Fig. 4). Accordingly, fault depth will increase away from the high heat flow domain of the volcanic arc and reach a maximum at the colder outer forearc region.

Our fault and earthquake data compilation reveals that the maximum known depths for crustal hypocenters are 25-30 km for the Outer Forearc domain, 15-20 km for Inner Forearc domain and 8-12 km for the Arc domain. Accordingly, the width of fault planes will depend on plane dip; subhorizontal faults could have bigger rupture areas, as it is on listric faults. Strike-slip faults, having almost vertical fault planes, will also share this limit.
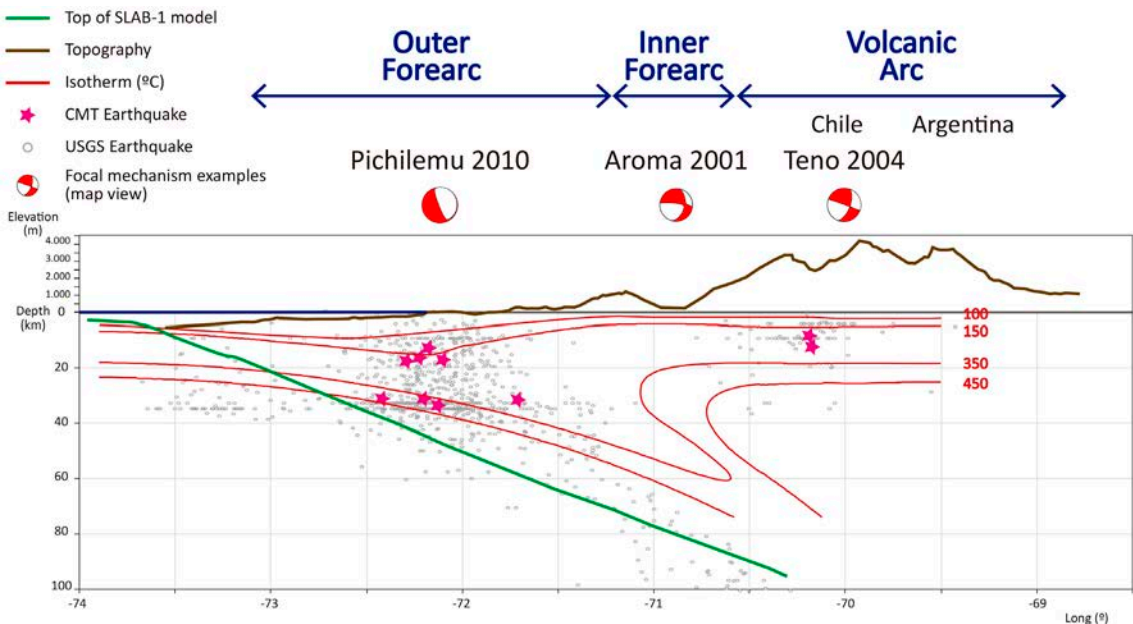

FIG. 4. Schematic cross section at central Chile latitude (~34º S) with studied earthquakes. Shown are the west to east domains and examples of crustal earthquakes with focal mechanisms for each domain. The Nazca-South American subduction zone is represented by the SLAB-1 model (Hayes *et al.*, 2012) and the variable thicknesses, from the subduction trench to the Andes Cordillera, of the brittle-ductile transition on the South American plate is correlated with the generalized isothermal line ~350 ºC. The compiled crustal earthquakes located between 33.5º and 34.5º S from CMT catalogue are highlighted (pink stars) from the totality of USGS (2018b) earthquakes in this range (grey circles in the background, 2018). Note crustal earthquake distribution across west to east domains, with greater crustal depths for outer forearc crustal events compared to inner forearc and the arc itself, augmenting the potential fault rupture area thus generating greater magnitude events.

The length of fault planes was estimated from the geomorphic features of the numerous crustal faults along the Andean region. Fault segments of up to 40 km in length are easily identifiable in satellite images and Digital Elevation Models (DEM) for different fault systems or single fault traces. These fault traces are the result of several seismic events, superimposing their effects over thousands of years of slip in same and different segments of the fault. This repetition will most likely result in recognizable maximum-length segments for the involved period.

Two of the most difficult parameters to estimate are the fault slip rate and the accumulated net displacement controlling seismic momentum. Moreover, because fault slip is the most sensitive parameter in the earthquake magnitude equation, researchers should be cautious when assigning magnitudes to events with poorly estimated slip rates. From the few paleoseismological works which indicate discrete measurable event slips, and the dating of these events, slip rates and recurrence times were obtained (*e.g.,* Cortés, 2012; Cortés *et al.,* 2012; Vargas *et al.,* 2014; González *et al.,* 2015). These works suggest that maximum average displacements during earthquakes can generally reach up to 2 m, or over 2 m in extraordinary cases.

By integrating studied faults, recorded earthquakes, morphotectonic and rheological features, we present a conceptualization of the main characteristics of Chilean crustal faults in table 3. There we show an outline of where, how and by what distance a crustal fault could slip during a hazardous seismic event.

We conclude that the along-strike tectonic domains (longitudinal domains) are the most relevant for the seismogenic potential assessment of crustal faults in Chile, since it is more significant for the faults magnitude parameters.

6.2. Estimated earthquakes

Following the previous analysis, it is possible to estimate the maximum magnitude earthquake for each longitudinal domain, using scaling relationships (Stirling *et al.,* 2013) and the assessed parameters seismogenic depth, length and slip, and assuming a constant shear modulus of 32 GPa in the crust. These magnitude estimations are in complete agreement with the recorded seismic events for each domain (Table 3).

The earthquakes with the largest moment magnitudes (Mw 7.5 for normal and strike-slip faults to Mw 7.0 reverse faults) occur in the Outer Forearc

Domain which experiences 0.2 to 0.6 mm/yr of slip rate, with a recurrence time of a few thousand years. Events with magnitudes between 6.9 and 7.4 could occur in the Inner Forearc Domain, which experiences slip rates of 0.1 to 0.4 mm/yr and recurrence times of a few thousand years. Events smaller than Mw 6.7 may occur in the Arc Domain, where the upper brittle crust thickness is lowest due to the presence of active magmatism. However, because slip rates of some arc domain faults are one order of magnitude higher than those of the outer and inner forearc domains, recurrence times could be as short as a few hundred years in this domain.

6.3. Parameter uncertainties assessment

The estimation, calculation or assessment of fault parameters is susceptible to errors, with the largest source being field data collection. The identification of recurrent rupture planes in the field (tectonic morphologies and/or deposits) becomes very difficult with the superposition of multiple seismic events, as previously mentioned. Consequently, the estimation of fault length for a characteristic earthquake or maximum credible earthquake is challenging and paleoseismological studies should be conducted along the entire length of a fault in order to be more fully representative and precise.

Fault displacement is also a difficult parameter to assess. For example, slip fault appraisal from colluvial wedge thickness formed coseismically in dip-slip faults: it must be identified a single seismic event colluvial wedge representative of an entire rupture zone slip. If all these conditions are found and the colluvial wedge had been under "normal" erosion rates, its height is thought to be half of the total slip as "rule of thumb" (McCalpin, 2009). But this relation must be used with caution, there are many problems in measuring 1 event-only colluvial wedges, scarp free-faces or displacement (*e.g.,* Ostenaa, 1984; Nelson, 1992; McCalpin, 2009). However, even when all of this is considered, earthquake size is still commonly overestimated. The bigger range for average displacement with the common error percentage make this a key parameter.

6.4. Activation mechanisms and hazard assessment

Crustal faults are widely scattered across the entire length and breadth of the Andes. After analyzing fault

TABLE 3. CRUSTAL FAULTS SEISMOGENIC PARAMETERS.

Longitudinal Tectonic Domain		Maximum Depth (km)	Maximum Mw Earthquake			Fault types	Examples	Slip rate (mm/y)	Recurrence (y)	References
			Estimated	Recorded	Paleo-seismological					
Outer Forearc	Coastal Cordillera	25-30	7.5-7.0	-	7.0	Normal-strike slip	Mejillones, Mititus	0.2-0.6	10^{-3}	Arriagada et al., 2011; Farías et al., 2011; Cortés et al., 2012; Aron et al., 2013.
				7.0			Pichilemu			Ruiz et al., 2014,
			7.0	6.7	7.0	Reverse	Pisagua, Bajo Molle			Allmendinger and González, 2010; Cortés et al., 2015
Inner Forearc	Central Depression, Main Cordillera	10-15	7.2	6.9	-	Strike slip	Aroma, Las Melosas, Domeyko, Teno	0.1-0.4	10^{+3}-10^{-4}	Comte et al., 2002; Legrand et al., 2007
			6.9-7.4	-	7.5	Reverse	San Ramón			Armijo et al., 2010; Pérez et al., 2014
Volcanic Arc	Main Cordillera	08-12	6.7-6.2	6.2	-	Strike slip	LOFS, Aysén	7.0	10^{+2}-10^{-3}	Mora et al., 2010; Legrand et al., 2011; Vargas et al., 2013
Transform fault (plate boundary)	Magallanes-Fagnano fault	30	7.8	7.5	-	Strike slip	Magallanes-Fagnano fault	7.0	10^{+2}-10^{-3}	Lommitz, 1970; Cisternas and Vera, 2008.

characteristics in their morphotectonic environment, it is possible to predict some seismogenic parameters for crustal faults located in particular tectonic domains. A summary of these conditions are shown in table 3.

There is a strong documented correlation between crustal seismicity and large interplate earthquakes, acting in earthquake segmentation (*e.g.,* Ecuador-Colombia, Collot *et al*., 2004; Alvarado *et al*., 2016; and Perú, Audin *et al*., 2008), slip partitioning (Melnick *et al*., 2006) or acting as mega-earthquake triggers (González *et al*., 2015) or coseismic/postseismic pseudo-aftershocks (Farías *et al*., 2011; Aron *et al*., 2013).

Outstanding examples in recent years in Chile are the Pichilemu and Iquique crustal earthquakes related to Maule 2010 and Iquique 2014 megathrust earthquakes respectively. The Pichilemu earthquakes Mw 7.0 and 6.9 (see section 4.5), 2 weeks after 2010 Maule earthquake, have been allegedly caused by the large change in Coulomb stress caused by the Maule earthquake (*e.g.,* Farías *et al*., 2011; Aron *et al*., 2013). The other case is a Mw 6.7 reverse crustal earthquake which was apparently the trigger for the 2014 Iquique-Pisagua Mw 8.1 megathrust earthquake, also by creating a change in the stress conditions, as suggested by González *et al*. (2015). That work indicates that several other crustal earthquakes, with reverse and normal slip, took place before and after the intraplate earthquake.

Other crustal earthquake clusters were also temporally associated with the 1985 Valparaíso, 2007 Tocopilla and 2005 Antofagasta thrust earthquakes, with the amount of crustal earthquakes generally increasing in the period following these large earthquakes.

Normal faulting crustal earthquakes were also observed following the Tohoku 2011 (Mw 9.1) and Sumatra-Andaman 2004 (Mw 9.1) mainshocks (*e.g.,* Dewey *et al*., 2007; Asano *et al*., 2011; Nettles *et al*., 2011; Hardebeck, 2012; Miyakawa and Otsubo, 2015). Toda *et al*. (2011a and 2011b) suggests that those faults were activated by static stress changes complementing local normal faulting regimes and stress heterogeneity. Hardebeck (2012) studied the rotations of the principal stress axes in those events, finding that they exhibit similar coseismic rotations; shallow plunge of σ_1 (maximum compressive stress axis) pre-event changing abruptly to 45º plunge σ_1 and σ_3 (maximum and minimum compressive stress axes) immediately after. In that context, crustal faults can be oriented for either reverse or normal faulting in this post-mainshock stress field, depending on their dip, explaining the observed normal-faulting aftershocks. Hardebeck (2012) also states that the stress rotation outcome following these 3 earthquakes does not occur in all subduction earthquakes. Deeper earthquakes exhibit smaller coseismic stress rotations, likely due to increasing deviatoric stress with depth.

Farías *et al*. (2011) recommend that crustal faults, especially those in the outer forearc, should be considered for seismic hazard analysis despite the absence of historic seismic activity, because events such as the Pichilemu sequence after the Maule earthquake prove their seismogenic potential.

The seismogenic potential of crustal faults in Chile is fundamentally linked to the subduction earthquake cycle. For instance, optimally-oriented faults, regarding the deformation field and stress changes, may be reactivated at virtually any time (King *et al*., 1994; Stein *et al*., 1994; Stein, 1999; Kilb *et al*., 2000; Lin and Stein, 2004; Loveless *et al*., 2010; Seeber and Armbruster, 2000). This is particularly true for optimally-oriented ancient basement faults which are subjected to high strain rates, a condition which can be reached co-seismically with large subduction earthquakes, and the slip being partitioned with these faults (*e.g.,* Melnick *et al*., 2006; Aron *et al*., 2013). In contrast, during the interseismic period, strain rates are kept relatively low making the reactivation of optimally oriented faults less likely. However, the precise nature of the link between crustal faults and the subduction zone earthquake cycle remains poorly understood.

We hypothesize that some crustal faults, especially those in the outer forearc, have the potential to reactivate co-seismically when optimally oriented with respect to the instantaneous extension direction arising from elastic rebound of mega-earthquakes (*e.g.,* Pichilemu fault; Aron *et al*., 2013). Other faults may activate during the subduction interseismic period, including regional strike-slip faults and thrust faults in the main cordillera (*e.g.,* Liquiñe-Ofqui fault, LOFS; Legrand *et al*., 2011; Mora *et al*., 2010).

7. Conclusions

The continental margin of the Chilean territory is segmented longitudinally into three tectonic domains -External Forearc, Inner Forearc and Volcanic Arc-along its length, which incorporate similar earthquake

parameters as fault size, slip rate and earthquake magnitude for crustal faults, thus allowing for a first-order approach for seismic potential assessment.

Although sparse and limited, current structural, paleo-seismological and geodetic data suggests that slip rates in Chilean crustal faults range from 0.2 mm/yr in the forearc to up to 7.0 mm/yr for the intra-arc. This implies recurrence times in the range of 200 to 50,000 years for *ca*. Mw 7 earthquakes.

Outer Forearc domain faults have the largest seismogenic potential, with earthquake magnitudes estimated at between Mw 7.0 and Mw 7.5, in agreement with recorded crustal earthquakes of *ca*. Mw 7.0 and similar magnitudes obtained from paleoseismological evidence. Slip rates range between 0.2 to 0.6 mm/yr with recurrence times of several thousands of years for *ca*. Mw 7 earthquakes (Table 3).

Volcanic arc domain faults, by contrast, have the lowest estimated seismic potential with earthquake magnitudes of Mw 6.2 to 6.7. However, these faults have slip rates which are one order of magnitude greater than those of the Outer Forearc (7.0 mm/yr) and have recurrence times of hundreds to thousands of years (Table 3).

The various different tectonic modes for reactivation of crustal faults in Chile -related to large interplate earthquakes and the stress field within the seismic cycle (section 6.4)- and their wide range of slip rates, greatly complicates the estimation of seismic hazard. Many structures considered active using traditional classification (see Machette, 2000 for a review) may not generate significant earthquakes in thousands of years according to their recurrence times, whereas other less-well-known faults, that have no instrumentally recorded seismicity, could trigger up to Mw 7 earthquakes. Furthermore, fault segments that have generated earthquakes independently from each other may be capable of merging into a single rupture zone and generate an earthquake of greater magnitude.

A rigorous seismic hazard assessment must then consider the different tectonic settings across the Chilean Andes, timing with interplate seismic cycle and slip rates of Andean crustal faults. Understanding the nature of these faults will help us not only to better assess the associated seismic hazard, but also to better understand and constrain their underlying connection with the subduction zone seismic cycle, as triggers, slip partitioning or result of subduction earthquakes.

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