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► **To cite this version:**

Leoncio Cabrera, Sergio Ruiz, Piero Poli, Eduardo Contreras-Reyes, Axel Osses, et al.. Northern Chile intermediate-depth earthquakes controlled by plate hydration. *Geophysical Journal International*, 2021, 226 (1), pp.78-90. 10.1093/gji/ggaa565 . hal-03197367

HAL Id: hal-03197367

<https://hal.univ-grenoble-alpes.fr/hal-03197367>

Submitted on 13 Apr 2021

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1 **Northern Chile Intermediate-Depth Earthquakes Controlled by Plate Hydration**
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14

15

16 **Abstract**

17 We investigate the variations of the seismic source properties and aftershock activity using
18 kinematic inversions and template-matching, for six large magnitude intermediate-depth
19 earthquakes occurred in northern Chile. Results show similar rupture geometry and stress drop
20 values between 7-30 MPa. Conversely, aftershocks productivity systematically decreases for the
21 deeper events within the slab. Particularly there is a dramatic decrease in aftershock activity
22 below the 400-450 °C isotherm-depth, which separates high and low-hydrated zones. The events
23 exhibit tensional focal mechanisms at unexpected depths within the slab, suggesting a deepening
24 of the neutral plane, where the extensional regimen reaches the 700-800°C isotherm-depth. We
25 interpret the reduction of aftershocks in the lower part of the extensional regime as the absence
26 of a hydrated-slab at those depths. Our finding highlights the role of the thermal-structure and
27 fluids in the subducting plate, in controlling the intermediated-depth seismic activity and shed
28 new light in their causative mechanism.

29

30 **Keywords:** Intraslab Intermediate Depth Earthquakes, Aftershock, Chile, Subduction, Slab.

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32

33

34 **1 Introduction**

35 Rupture mechanism of intermediate-depth earthquakes (IDEs) is one of the biggest open
36 questions in seismology (Frohlich, 2006; Houston, 2015). They occur mostly at subduction zones
37 in the so called double seismic zones (DSZ), where two parallel seismicity planes are observed.
38 The upper seismicity plane (USP) is inferred to occur within the subducting oceanic crust and/or
39 the upper mantle due to dehydration reactions (Hacker et al., 2003; Kirby, 1995). In contrast, the
40 lower seismicity plane (LSP) occurs in the lithospheric mantle, and its mechanism is still under
41 debate (Ferrand et al., 2017; Ohuchi et al., 2017; Peacock, 2001; Reynard, 2010). Hypothesis for
42 this process include dehydration-embrittlement of antigorite (Peacock, 2001), reactivation of pre-
43 existing shear zones (Reynard, 2010) and quasi-adiabatic shear-heating instabilities (John et al.,
44 2009). Furthermore, laboratory experiments also suggest that faulting at intermediate depths may
45 occur under dry conditions (Ohuchi et al., 2017) or by dehydration-driven stress transfer under
46 partially hydrated mantle conditions (Ferrand et al., 2017).

47 The DSZ is usually characterized by compressional deep events within the oceanic lithospheric
48 mantle in trench-outer rise regions prior to subduction, while the upper part of the oceanic
49 lithosphere is affected by a tensional regime. This stress field in the trench-outer rise region is
50 controlled by the plate bending caused by the interplay of slab pull forces and the rheology of the
51 oceanic lithosphere (Contreras-Reyes & Osses, 2010). Furthermore, based on a worldwide
52 compilation of trench outer rise earthquakes, the position of the neutral plane of the stress field
53 within the oceanic lithosphere have been estimated to be approximately coincident with the 400°-
54 450°C isotherm-depth (Seno & Yamanaka, 1996). Likewise, Ruiz & Contreras-Reyes (2015) and
55 Carrasco et al., (2019) found that tensional outer rise events are confined to the upper oceanic
56 lithosphere to isotherm-depths shallower than 400°C offshore central Chile. Off Japan, Kita et

57 al., (2010) reported that the transition from upper tensional to compressional intraplate
58 earthquakes occur at 22 km beneath the base of the Eurasian plate beneath Tohoku. However,
59 this transition occurs only ~10 km beneath the upper plate beneath Hokkaido. The authors
60 attribute this difference to the complexity of the oblique subduction beneath Hokkaido, and they
61 further pointed out that the stress regime is the combination of two stress fields: (i) the
62 bending/unbending force, and (ii) the thermomechanical and petrological forces.

63 Although IDEs are characterized by shear slip on faults similar as inland crustal or interplate
64 earthquakes, anomalous behavior is often observed for the first ones such as thermal-dependence
65 radiated seismic energies (Wiens, 2001), significant non-double-couple mechanisms (Richardson
66 & Jordan, 2002), and high stress drops (Frohlich, 2006; Poli et al., 2014; Houston, 2015) among
67 others. A remarkable characteristic of intermediate and deep earthquakes is their general lower
68 aftershock productivity as compared with shallower events (Frohlich, 1987; Wiens et al., 1994;
69 Zhan, 2014; Houston, 2015), although few cases with large activity exist (Wiens et al., 1994).
70 Aftershocks activity has been related to fault properties as state of stress (Shebalien & Narteau,
71 2017), rupture processes (Poli et al., 2016), seismic coupling (Hainzl et al., 2019) and the
72 thermal state of the slab, with colder slabs showing a greater number of aftershocks (Wiens &
73 Gilbert, 1996). A main issue is to understand if this general decrease in the number of
74 aftershocks is real, or it is due to the lack of near-field data (Li et al., 2018). It is thus
75 fundamental to tackle this issue, to better assess the aftershock activity (Dascher-Cousineau et
76 al., 2020) and get new insights about the mechanism(s) and physical conditions controlling the
77 occurrence of IDEs.

78 Since 2010, six intermediate-depth events with magnitudes (M_w) greater than 6 occurred in
79 northern Chile between 17°S and 21°S (Figure 1 and Table S1), where the Centro Sismológico

80 Nacional de la Universidad de Chile (CSN; Barrientos et al., 2018), Integrated Plate Boundary
81 Observatory Chile (IPOC; GFZ, & CNRS-INSU, 2006) and Central Andean Uplift and the
82 Geodynamics of the High Topography (CAUGHT; Beck et al., 2010) seismic networks have
83 been deployed (Figure 1a). These events are of main interest for several reasons. First, they are
84 registered by a dense network of broad-band and strong-motion sensors in the near-field
85 providing an opportunity to avoid problems with detection levels (Poli et al., 2016; Shebalien &
86 Narteau, 2017; Zhan et al., 2014). Second, the six events are located in a very limited volume,
87 reducing issues related with attenuation and other propagation effects, which are possibly
88 affecting the estimation and thus the comparison of source properties (Prieto et al., 2012).
89 Finally, their magnitudes ($M_w \sim 6.3$) and focal mechanisms are similar (Figure 1). We thus have
90 a unique opportunity to compare source properties and aftershock activity, with limited bias due
91 to propagation and rupture geometry. In addition, according to the CSN catalogue they exhibit
92 different aftershock productivity. For the shallowest event (E1 in Figure 1c), there are more than
93 40 aftershocks reported, including some of them with magnitude equal or greater than $M_w 4.0$
94 (see Supplementary Information). For the deeper ones there is a decrease in the number of
95 aftershocks with no aftershocks for the deepest event (E6 in Figure 1c).

96 To better characterize if differences in between these events exist, we study their aftershock
97 activity and rupture properties, using template matching (TM) and kinematic inversions (KI).
98 Furthermore, we developed a thermal model using the finite element method (FEM), with the
99 aim of analyze the impact of slab temperature on seismic behavior. In the next sections, the
100 tectonic setting and methodologies used in this study followed by their respective results are
101 described, and finally the discussions and conclusions are presented.

103 2 Tectonic Setting

104 The northern Chilean convergent margin is characterized by the subduction of the oceanic Nazca
105 plate beneath the continental South American plate (Figure 1) at a currently convergence rate of
106 ~67 mm/a (Khazaradze & Klotz, 2003), although the average convergence rate during the last 20
107 Ma is about 85 mm/a (DeMets et al. 2010). The Nazca plate in the study region (18° - 21°S) was
108 formed in the East Pacific Rise 45-52 Ma (Mueller et al., 2008), whereas its crustal thickness is
109 6-7 km thick inferred by seismic constraints (Patzwahl et al., 1999; Ranero and Sallarès, 2004;
110 Contreras-Reyes et al., 2012; Maksymowicz et al., 2018). The northern Chilean margin has a
111 poorly sedimented trench (< 200 m; Maksymowicz et al., 2018) due to the reduced sediment
112 supply from the Andes in the extremely arid region of the Atacama Desert. The continental slope
113 is steep as the frontal part of the margin lacks a well-developed frontal prism (Contreras-Reyes
114 2018; Maksymowicz et al., 2018).

115 The northern Chilean margin has undergone subduction erosion since at least the Jurassic, which
116 is inferred from the eastward migration of the volcanic arc (e.g. Rutland, 1971; Stern, 2011) as
117 well as the long-term arcward retreat of the trench, crustal thinning and subsidence of the outer
118 forearc (e.g., Kukowski & Oncken 2006; Contreras-Reyes 2018). Long-term subduction rates for
119 northern Chile have been estimated with values of 40-45 km³Myr⁻¹km⁻¹ (Kukowski & Oncken
120 2006). On the other hand, the oceanic Nazca plate presents well developed tensional faults seen
121 in the bathymetric data (Geersen et al., 2018), which were formed by bending of the oceanic
122 plate (Ranero et al., 2005). These bend faults provide the pathways for seawater infiltration

123 reaching the oceanic mantle according to seismic and geodynamic studies (see Contreras-Reyes
124 & Osses, 2010 and references therein).

125

126 **3 Aftershock productivity**

127 **3.1 Detection of aftershocks using template matching**

128 The six studied IDEs have tensional focal mechanism and span a range of depth from 90 to 130
129 km (Figure 1), in a region where a double seismic zone (DSZ) has been previously reported
130 (Comte et al., 1999; Dorbath et al., 2008; Sippl et al., 2018; Florez & Prieto, 2019) and the
131 seismic catalog has magnitude of completeness \sim Ml 3.5 (Barrientos et al., 2018). It is thus likely
132 that aftershocks are missing for some the studied events, which is improved using template
133 matching (TM).

134 TM has been widely used to detect events in aftershocks sequences, where seismicity is known
135 to repeat (Shelly et al., 2007; Peng & Zhao, 2009; Frank et al., 2017; Li et al., 2018). In a similar
136 way, we here apply TM to improve the detection of aftershocks following the analyzed events
137 (Figure 1). We base our analysis on continuous three-component velocity seismological data
138 from the CSN, IPOC, and CAUGHT seismic networks, using the five nearest stations to the
139 source region of each event (Figure S15). Data were continuously recorded at a sampling rate of
140 100 Hz for IPOC and CSN stations, and 40 Hz for CAUGHT stations. To detect non-reported
141 aftershocks, we searched for events with similar waveforms to those reported as mainshock and
142 aftershocks for the CSN (template candidates, see Supplementary Information). A space-time
143 window needs to select reference events and build templates. A large window implies including

144 more events, but also more background events as well, while a smaller window decreases this
145 effect, but is more likely to miss some aftershocks. Since the six earthquakes have similar
146 characteristics such as magnitude, focal mechanism, hypocenter and station coverage, we use the
147 same spatial and temporal scale for all of them in order to preserve a homogeneous criterion that
148 allows their comparison. After some tests with different windows size and considering some
149 previous works (e.g., Persh and Houston, 2004; Dascher-Cousineau et al., 2020) we define the
150 spatial limit using a 3D radius of 25 km from each hypocenter, and 25 days of data after each
151 mainshock, as this is the maximum number of days for which the five stations closest to each
152 event were operating continuously. Both continuous data and template waveforms were bandpass
153 filtered from 5 to 10 Hz because this frequency range exhibits better signal to noise ratios, and
154 decimated to 25 Hz. To select high-quality template events, we follow Frank et al. (2017) and we
155 estimate the signal-to-noise ratio (SNR) on the vertical component as the ratio between the RMS
156 velocity during the first 25 s of the P-wave and the RMS velocity during a 25 s of noise before
157 the P-wave arrival time. Only template candidates with $SNR > 5$ in at least 3 stations are used as
158 templates. We define the resulting template waveforms as 30 s time windows that start 5 s before
159 event's P-wave arrival at each station. Then we calculate the correlation coefficients between
160 template waveforms and continuous data in a sliding window that preserves the seismic
161 moveouts (Frank et al., 2017), to get a time series that represents the similarity of the continuous
162 data to the matched-filter template. We search sample-by-sample considering a detection
163 threshold that is 10 times the median absolute deviation (MAD) of the correlation sum to detect
164 event significantly similar to the template. Events detected with this criterion, are then
165 considered to occur at the same hypocenter (determined by the CSN) as their template, and we
166 estimate their magnitude by computing the median amplitude ratio between the template event

167 and the aftershock over the 5-station network, assuming that a tenfold increase in amplitude
168 corresponds to one unit increase in magnitude. Figures 2b-c show an example for E2 and E5,
169 where using the mainshocks waveforms as templates, two new detections with inferred
170 magnitudes (M_w) of 1.8 and 2.7 occurred four days (for E2) and some hours (for E5) after
171 mainshocks.

172 To avoid redundancy of events detected by multiple templates, we remove detections within 30 s
173 of another detections, by keeping the event with the highest network correlation coefficient.
174 Finally, we use the new catalogue (including CSN and new detected aftershocks) as templates
175 candidates, and the process is repeated once.

176 **3.2 Aftershock Activity**

177 Figure 3a shows all new detections and catalog events, for 25 days after each mainshock. As is
178 common for aftershock sequences, the number of events decreases over time, with a higher
179 concentration close to the mainshock. On the other hand, the number of aftershocks shows a
180 clear reduction as function of depth, down to E6 that is showing no aftershocks (Figures 3a-c).
181 The most productive event is E1, located just ~ 7 km from the top of slab (Figure 1). For this
182 earthquake, we used 45 events as templates, and we found 2044 aftershocks, nearly 47 times
183 more.

184 It is important to highlight that there are no aftershocks reported in the CSN catalog for E2, E5,
185 and E6. For these events, we use the mainshock waveforms for detection. This strategy permits
186 to find new events for E2 and E5, suggesting that the sole main event waveforms can improve
187 the detection. Figure 2 shows an example of detections for E2 and E5 using mainshock

188 waveforms as templates (see Table S2 for more details about detections). According to the
189 above, lack of detection seen for E6 is a robust feature of our analysis.

190

191 **4 Thermal structure**

192 **4.1 Thermal-Fluid Numerical Model**

193 To assess the thermal conditions at the depth of the analyzed events, we developed a 2D-thermal
194 model using Finite Element Method (FEM) for northern Chile, along the BB' profile (Figure 1).
195 The model is constrained by the plate geometry of SLAB2.0 (Hayes et al., 2018), hypocenter
196 data, and published thermal parameters in the area (see Table S5).

197 **4.1.1 Numerical Model**

198 The thermal subduction model consists in a two-dimensional finite element discretization of a
199 coupled system of a heat equation for the temperature and and the Stokes's system for the
200 subducting lithospheric plate speed and asthenospheric fluxes. The heat equation is given by:

$$-\kappa\nabla^2 T + \vec{u} \cdot \nabla T - Q = 0 \quad (1)$$

201 in the region of interest Ω , where T is the temperature and the first term corresponds to heat
202 diffusion where κ is the thermal diffusivity depending on space (ratio of the thermal conductivity
203 k and the thermal capacity ρc_p of the media), the second term corresponds to the heat transport
204 due to lithospheric speed and asthenospheric fluxes \vec{u} , and the last term stands for the external heat
205 sources such as radiogenic or frictional heat. No friction heat effects are considered in this study
206 since in the case young lithospheres the effects of friction can be neglected (Vöelker et al., 2011).

207 Thermal effects due to the subducted eroded material above the incoming/subducting lithosphere
 208 are also neglected for the sake of model simplicity.

209 Given a velocity field \vec{u} in the whole region Ω , we obtain the steady state temperature T as the
 210 limit of T_n solution of the following variational problem discretized with Lagrange finite elements
 211 of type P2 (space V_h) and iterating for $n \geq 1$ starting from $T_0 = 0$ until convergence, where Δt is
 212 some fictitious numerical time step:

$$\frac{1}{\Delta t} \int_{\Omega} (T^n - T^{n-1})S + \int_{\Omega} \kappa \nabla T^n \cdot \nabla S + \int_{\Omega} (\vec{u} \cdot \nabla T^n)S - \int_{\Omega} QS = 0 \quad \forall S \in V_h \quad (2)$$

213 Given the units we use, we observe that it is sufficient to make the order of 15 iterations with a
 214 time step of 10 myrs corresponding to 150 myrs in total in order to reach a steady state.

215 In order to compute the heat transport field \vec{u} , we consider zero velocity in the continental crust
 216 and we impose a rigid and constant movement along the parallel shape lines of the incoming plate.
 217 We model the fluxes produced in the mantle wedge using a divergence free Stokes flux,
 218 considering both momentum and mass conservation:

$$-\nu_A \Delta \vec{u} + \nabla p = 0 \quad (3)$$

$$\nabla \cdot \vec{u} = 0 \quad (4)$$

219 in the region Ω_A occupied by the mantle wedge, where $\vec{u} = (u_x, u_y)$ and p are the velocity and
 220 pressure of the fluid flow and ν_A is the kinematic viscosity of the asthenosphere in the wedge
 221 mantle. Thus, only for the wedge mantle, the velocity field $\vec{u} = (u_x, u_y)$ is computed from the

222 Stokes's variational formulation using finite elements of type P1b for the velocity (space U_h) and
 223 P1 for the pressure (space P_h) by solving:

$$\begin{aligned}
 v_A \int_{\Omega_A} \frac{\partial u_x}{\partial x} \frac{\partial v_x}{\partial x} + \frac{\partial u_x}{\partial y} \frac{\partial v_x}{\partial y} + \frac{\partial u_y}{\partial x} \frac{\partial v_y}{\partial x} + \frac{\partial u_y}{\partial y} \frac{\partial v_y}{\partial y} \\
 + \int_{\Omega_A} \frac{\partial p}{\partial x} v_x + \frac{\partial p}{\partial y} v_y + q \left(\frac{\partial u_x}{\partial x} + \frac{\partial u_y}{\partial y} \right) = 0 \quad \forall \vec{v} \in U_h, q \in P_h.
 \end{aligned}
 \tag{5}$$

224 4.1.2 Thermal Model Description (Boundary Conditions)

225 For the heat equation, we consider that the temperature on the edge condition of the
 226 incoming/subducting plate increases in deep y by:

$$T(y) = T_S + (T_m - T_S) \operatorname{erf} \left(\frac{y}{\sqrt{4\kappa t}} \right)
 \tag{6}$$

227 where $T_S=0$ and $T_m=1350$ °C are the surface and mantle temperatures, where erf is the Gaussian
 228 error function, t is the age of the plate (50 myrs) and κ is the thermal diffusivity of the incoming
 229 plate. This is a good approximation for deep temperature profiles for low aged plates that is the
 230 case for the entire Chilean trench (see Stein & Stein, 1992; Vöelker et al., 2011). For the other
 231 boundary conditions (Figure 4), we assume that T varies quadratically from 1350° to 1400 °C in
 232 interior of the right boundary of the mantle wedge, $T=0$ at the surface, $T=1350$ °C at the bottom
 233 part of the subducting lithosphere and zero heat flux in the rest of the boundaries.

234 For the Stokes's equation, we take $\vec{u} = 0$ in the continental crust and $\vec{u} = u_0 \vec{\tau}(y)$ in the subducting
 235 lithosphere where $u_0=67$ mm/yr is the imposed constant speed of the incoming plate with $\vec{\tau}(y)$ the
 236 unit tangent vector to the subducting lithosphere shape lines depending on depth y . The boundary
 237 conditions for the flow in the mantle wedge are $\vec{u} = 0$ on the upper part of the wedge, $\vec{u} = u_0 \vec{\tau}(y)$
 238 on the lower part of the wedge and free flow in the right edge of the wedge. See Text S1 in the

239 supplementary information for more details.

240 **4.1.2 Thermal Model Description (Parameter Selection)**

241 Topography and bathymetry are taken from published data (Ryan et al., 2009). The geometry of
242 the subducted lithosphere is taken from Slab2.0 (Hayes et al., 2018). In addition, to assess our
243 model, we consider three different upper continental mantle depths of 60, 80 and 100 km.
244 However, as they show roughly similar results for the location of the mainshocks, we consider the
245 more realistic model of 60 km (Wada and Wang, 2009). Figures S16 and S17 show models with
246 80 and 100 km, respectively). A detail list of the other parameters used to develop the two-
247 dimensional model through finite element are listed in Table S5 where values are taken from
248 several authors (Hyndman & Wang, 1993; Herzberg et al., 2007; Wang et al., 1995; Oleskevich et
249 al., 1999; Vöelker et al., 2011, Höink et al., 2011) for rheologic and geometric values. Please note
250 that heat diffusivity takes different values in three different regions: the subduction lithosphere,
251 the upper part of the continental crust and sediments and the rest of the domain. For the sources of
252 temperature, we also consider different radiative sources in the upper and lower continental crust
253 and sediments.

254 **4.2 Thermal Conditions of the Slab**

255 From our results of the slab thermal-conditions in northern Chile (Figure 3d), we observe that E1
256 and its aftershocks occurs near the 400 °C isotherm-depth and does not exceed the 450 °C
257 isotherm, while the rest of the events occurs around or below this isotherm. Figures 3b and 3c
258 illustrate the relationship between the temperature of the slab at which events occur, the distance
259 from the top of the slab and number of aftershocks, respectively. Our analysis clearly reveals an
260 inverse relationship between the number of aftershocks and temperature. It is important to

261 highlight that E6, the deepest within the slab, occurs at the highest temperatures range between
262 700 – 800 °C, and is the only event that does not exhibit aftershocks.

263

264 **5 Seismic Source properties**

265 **5.1 Seismic Source Kinematic Inversions**

266 Having observed a clear difference of aftershock productivity as function of depth, we now
267 attempt to resolve any difference in rupture properties. To obtain the slip distribution of each
268 mainshock, we perform kinematic inversions using near-field stations on hard rock (Leyton et
269 al., 2018) (Figure 1), including 19 strong-motion stations and one broad-band station (Figure
270 S17). Since it is not possible to distinguish the causative fault plane from aftershocks geometry,
271 we consider the two possible fault planes, according to moment tensors reported by the USGS
272 (Tables S3 and S4). We assume a finite-fault model with an a-priori elliptical-patch slip
273 distribution. This methodology has been successfully used by several authors (Madariaga and
274 Ruiz, 2016; Ruiz et al., 2019). We inverted seven parameters. Five of them, correspond to
275 geometrical characteristic: semi-axes a and b of the ellipse, the rotation angle of the ellipse α
276 and the location (x_0, y_0) of its center into the fault plane, and the other two parameters are
277 maximum slip D_{max} and rupture velocity V_r . From this inverted rupture model, we compute
278 dynamic parameters using circular crack approximation obtained averaging both axes of the
279 ellipse (Ruiz et al., 2019 and references there in). Strong motion and broad band records were
280 corrected for the instrumental response and linear trend, filtered using a causal Butterworth filter
281 with corner frequencies of 0.02 and 0.1 Hz, and then integrated to displacement. It is important
282 to mention that the high corner frequency is controlled by our simple elliptical model and the 1-

283 D velocity model (Husen et al., 1999) used to simulate the wave propagation from the source to
284 the receivers. The misfit between observed and synthetic records was computed using a L2 norm
285 starting from origin time up to 75 s later to avoid including surface waves, and considering both
286 possible rupture planes reported by the USGS (Tables S3 and S4).

287 **5.2 Kinematic Parameters**

288 Considering all six events, our results show similar geometries of rupture (Table 1), with an
289 approximately circular rupture ranging from ~3.5 to ~5 km.

290 Figure 5 shows an example of the kinematic inversion results, for event E1 considering NP1. The
291 observed and synthetic displacement waveforms show a good agreement, with a maximum value
292 about 0.4 cm for the East-West component in station A20P (waveforms for all events in
293 Supplementary Figures S1-S12). In addition, although there are differences in the misfit values
294 for each of the fault planes considered (Tables S3 and S4), these are small enough to give
295 preference to one or the other.

296 The estimated stress drop values vary between 7.5 MPa and 29.5 MPa (Table 1), which are high
297 compared with typical values for shallow earthquakes (Derode & Campos, 2019), but similar to
298 others observed for IDEs in Chile (Ruiz & Madariaga, 2011; Ruiz et al., 2019) and at global
299 scale (Poli & Prieto, 2014; Prieto et al., 2012). Despite a variability of rupture parameters exists
300 (Table 1), we do not see any clear correlation between them and the depth of the events. We thus

301 suggest that the analyzed events, while occurring under different thermal conditions, all have a
302 similar rupture physics.

303

304 **6 Discussion**

305 While co-seismic rupture properties do not vary with distance from the top of the slab, we
306 observe clear differences in the post-seismic activity pattern, with a decrease of the aftershocks
307 as the distance from the top of the slab and temperature increase (700-800 °C isotherm-depths).

308 Difference in the number of aftershocks between E1 and the rest of events could be related to the
309 hypocenter-depth within the slab. E1 is located ~7 km from top of slab, close to the oceanic
310 Moho (Figure 1b; Patzwahl et al., 1999; Ranero and Sallarès, 2004). In addition, E1 aftershocks
311 locations reported by the CSN delineate a potential pre-existing fault aligned and similar to those
312 extensional faults located in the outer-rise region (Figure S13). These tensional faults are likely
313 caused by plate bending, and could provide a pathway for fluid infiltration, producing hydration
314 into the crust and uppermost mantle (Boneh et al., 2019; Contreras-Reyes et al., 2008; Iyer et al.,
315 2012). However, how deeply water can infiltrate is still under debate, since the buoyancy of
316 water (or equivalently, confining pressure) makes it difficult to bring water down even when
317 faulting is deep (Korenaga, 2017). By contrast, numeric models suggest that stress variation
318 induced by plate bending can produce sub-hydrostatic or even negative pressure gradients along
319 normal faults, favoring thus downward pumping of fluids (Faccenda et al., 2009). On the other
320 hand, previous studies in the DSZ of this region (Dorbath et al., 2008) observed that the upper
321 seismicity plane (USP) corresponding to oceanic crust, is characterized by intermediate V_p (~7.7
322 km/s) and low V_p/V_s (1.67) values, concluding that the USP is related to fluid releases associated

323 with metamorphic reactions occurring within jadeite-lawsonite blueschists. Similarly, it has been
324 observed that the USP has significantly larger b-values than the LSP (Florez & Prieto, 2019),
325 suggesting an hydrated USP and a relatively dry lithospheric mantle, since previous observations
326 have shown that high b-values anomalies correlate well with regions where dehydration reactions
327 are expected (Wiemer & Benoit, 1996) or with hydrated fracture zones (Schlaphorst et al., 2016).
328 Thus, this hydrated zone could lead to a greater occurrence of aftershocks.

329 Furthermore, we observe a dominant tensional stress regime, where mainshocks with normal
330 focal mechanism occur up to ~40 km from the top of the slab (i.e., 700 – 800 °C isotherm). This
331 isotherm-depth interval is much deeper than the expected depth for this type of focal mechanism
332 compared to neutral plane of the stress field estimated by Seno & Yamanaka, (1996), who show
333 that the transition from tensional to compressional of outer rise events occur at the 400°-450 °C
334 isotherm-depth. If this is correct, our results show that the plate bending becomes stronger during
335 subduction and is likely more predominant than the cyclic stresses associated to the seismic
336 coupling at the thrust zone at depths where IDEs occur (Seno & Yamanaka, 1996). Furthermore,
337 seismic tomographic studies off south-central Chile (Contreras-Reyes et al., 2008) and off
338 Nicaragua (Lefeldt et al., 2009) show a reduction in the upper oceanic mantle velocities from the
339 oceanic Moho up to 400°-450 °C isotherm-depth. This lower bound has been interpreted as the
340 maximum potential depth for hydro-alteration within the upper part of the oceanic lithosphere,
341 where extensional stresses dominate, and fluids may not be able to penetrate any deeper than
342 neutral plane (Contreras-Reyes et al., 2008; Lefeldt et al., 2009; Faccenda et al., 2009). Another
343 proposed interpretation is that these seismic velocity anomalies could be related to small-crack
344 porosities, which can be produced by a mixture between thermal cracking and bending-related
345 faulting (Korenaga, 2017). As described above, our observations could suggest the existence of a

346 deepening of the neutral plane from 400-450 °C to 700-800 °C isotherm-depths regardless of the
347 presence of deep-water percolation.

348 A remaining question is, what could be controlling this deepening of the neutral plane along-dip?
349 Northern Chile corresponds to the region where the oceanic Nazca plate is relatively old (Figure
350 1) and therefore colder and with a deep fragile-ductile system (Figure 6), with an age of 54 Ma
351 near the trench axis (Müller et al., 2008) and 58 Ma estimated for the area where the mainshocks
352 occur. A plausible mechanism for a deepening of the neutral plane along-dip in northern Chile
353 would be the increase in bending stresses due to a greater slab-pull associated to the relative old,
354 cold and heavier already subducted oceanic Nazca plate. In particular, in this segment of the
355 Chilean margin, the largest known IDE has been registered for the slab-pull Tarapacá 2005 Mw
356 7.7 event (Peyrat et al., 2006; Legrand and Delouis, 2006; Kuge et al., 2010; Ruiz and
357 Madariaga, 2018). Moreover, the 1950 Ms 8.0 tensional event (Kausel & Campos, 1992)
358 occurred a little further south, reflecting the tensional character of the stress field within the
359 subducting slab. We also note how the moment tensors in northern Chile DSZ show a downdip
360 extensive stress field throughout the seismically active area (Sippl et al., 2019), which may imply
361 that plate bending, and unbending forces are small compared to slab-pull forces. Another
362 mechanism previously proposed by Kita et al. (2010) to explain the difference in the location of
363 the neutral plane beneath Tohoku and Hokkaido in the northeastern Japan is related to
364 differences in the thermal structure between the two regions. They concluded that the thermal
365 age of the subducted lithosphere and the oblique component of convergence rate affect the
366 density of the mantle wedge. This process triggers buoyancy force variations exerted in the slab
367 affecting finally the stress field of the subducting slab. However, this mechanism should be more

368 dominant along strike (thermal variations are more important in that direction) rather than in
369 along dip.

370 On the other hand, the occurrence of tensional events up to depths of 40 km as E6 is consistent
371 with the brittle region predicted by a yield strength envelope (YSE) for an oceanic lithosphere
372 under tensional stresses (Figure 6), which reaches depths of 40-42 km for olivine and dry olivine
373 mineralogy, and it decreases for wet conditions. Likewise, using high-precision relocations, a
374 DSZ south of Iquique (21.5 °S) is observed (Rietbrock & Waldhauser, 2004), with
375 predominantly extensional faulting in both seismic layers, and a clear down-dip tensional regime
376 with T-axes oriented in slab-parallel directions (Rietbrock & Waldhauser, 2004; Sippl et al.,
377 2019). The consistency between the strength envelope and the maximum depths of intraplate
378 tensional events suggests that these earthquakes have ruptured through the entire brittle part of
379 the oceanic lithosphere. Another interesting observation is that events E5 and E6 highlight for its
380 isotherm-depth occurrence, about 700-800 °C, which are higher than the stability limit of
381 antigorite at 600–650 °C commonly proposed as a lower limit for the LSP. As observed by Sippl
382 et al. (2018; 2019) these events are more common in the Northern Chile DSZ, and a runaway-
383 type process could be responsible, under specific conditions in the lower plate (e.g. composition,
384 hydration, stress state). Likewise, according to McKenzie et al. (2005) particularly large strain
385 rates could also lead to earthquakes in regions below the 600–650 °C isotherm-depth. In our
386 study region, this could be related to an increase in the slab pull. Finally, thermal cracking
387 hypothesis (Korenaga, 2017) could also play a role, since the deepest hydrated cracks could be
388 more widely spaced than the most superficial cracks

389 Considering previous observations and our own results, we propose a conceptual model shown in
390 Figure 7, where the processes that occur are as follows: (1) in the outer-rise region extensional

391 bend-faulting occur and lead to partial hydration of the crust and upper mantle. However, fluids
392 may not be able to penetrate any deeper than neutral plane (approximately 450 °C isotherm-
393 depth). Thus, the neutral plane separates a high-hydrated from a dry or poorly hydrated zones.
394 (2) As the oceanic plate subducts, a deepening of the neutral plane occurs by the increase of the
395 slab-pull forces, separating it from the 450 °C isotherm, and increasing the region subject to
396 extensional failure. (3) Normal (tensional) events occur at different distances from the top of the
397 slab, but their behavior is mostly controlled by variable physical background conditions. In
398 particular, in the high-hydrated region a greater number of aftershocks is observed, differently
399 from what is observed in the dry deeper zone.

400

401 **7 Conclusions**

402 We presented a detailed study of seismic source and aftershock productivity for the six largest
403 IDEs occurred in northern in Chile since 2010 and recorded by dense seismic arrays in the near-
404 field. Our results show that, although all of these events are located at different depths and under
405 different thermal conditions, all have similar rupture physics, with analogous geometries
406 considering an elliptical-patch approach and stress drop values between 7 MPa and 30 MPa. On
407 the other hand, a clear decrease of the number of aftershocks as the distance from the top of the
408 slab and temperature increase is observed. We propose that this behavior could be controlled by
409 the incoming plate hydration, where the 400 – 450 °C isotherm-depths and neutral plane of the
410 stress field acts as limits for hydration in the outer-rise region, which is deepened by the slab-pull
411 as the slab subducts.

412

413 **Declaration of competing interest**

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

416

417 **Acknowledgments**

418 We thank the Incorporated Research Institutions for Seismology Data Management Center and
419 Centro Sismológico Nacional who provided the raw data used in this work. LC thanks the
420 scholarship PFCHA/MagísterNacional/2018-22189566, granted by CONICYT. SR and ECR
421 thank the support of Chilean Fondo Nacional de Desarrollo Científico y Tecnológico
422 (FONDECYT), Grants 1170430 and 1170009, respectively. PP and LC were supported by the
423 European Union Horizon 2020 Research and Innovation Programme (grant agreements, 802777-
424 MONIFaults). AO acknowledges FONDECYT Grants 1191903 and 1201311, Basal CMM
425 ANIF PIA AFB170001 and Fondap CR2 Grant 15110009. We also acknowledge the support of
426 CONICYT PIA/Anillo de Investigación en Ciencia y Tecnología ACT172002. We also thank the
427 editor Eiichi Fukuyama, Jun Korenaga and another anonymous reviewer for their comments,
428 which improved the manuscript. We used Generic Mapping Tools 6 software for preparing some
429 plots.

430

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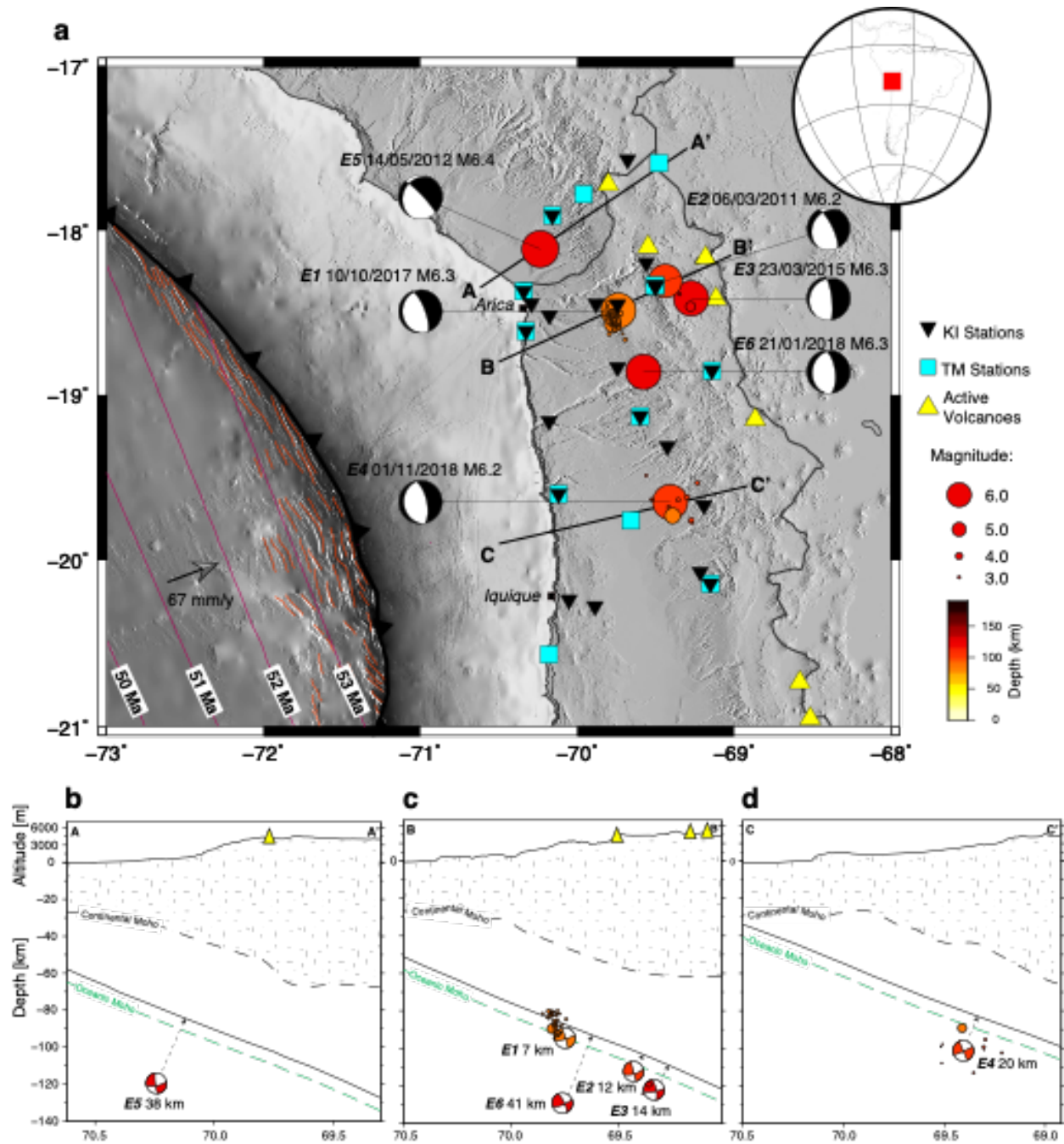
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666 **Table 1**667 *Resume of kinematic parameters and number of aftershocks obtained for each mainshock.*

Event	<i>a</i> (km)	<i>b</i> (km)	<i>D</i>_{max} (m)	<i>V</i>_r (km/s)	Stress drop (MPa)	Number of aftershocks	Distance from the top of the slab (km)
E1	3.49	5.13	1.08	1.07	18.2	2044	7
	3.67	5.68	0.94	1.19	14.7		
E2	4.38	6.99	0.65	1.59	8.5	123	12
	4.12	6.58	0.73	1.51	9.9		
E3	5.05	6.16	1.12	0.67	14.3	30	14
	4.14	6.44	1.24	0.82	16.7		
E4	4.70	3.90	0.73	1.22	12.8	46	20
	4.56	5.60	0.52	1.91	7.5		
E5	5.89	4.25	0.71	1.68	10.0	16	38
	4.88	6.11	0.60	2.20	8.1		
E6	3.12	4.42	1.61	0.68	29.5	0	41
	3.43	6.40	0.95	0.80	14.1		

668 Note: Green and yellow rows indicate parameters obtained using nodal plane 1 (NP1) and nodal plane 2 (NP2). For more details
669 see Tables S3 and S4.

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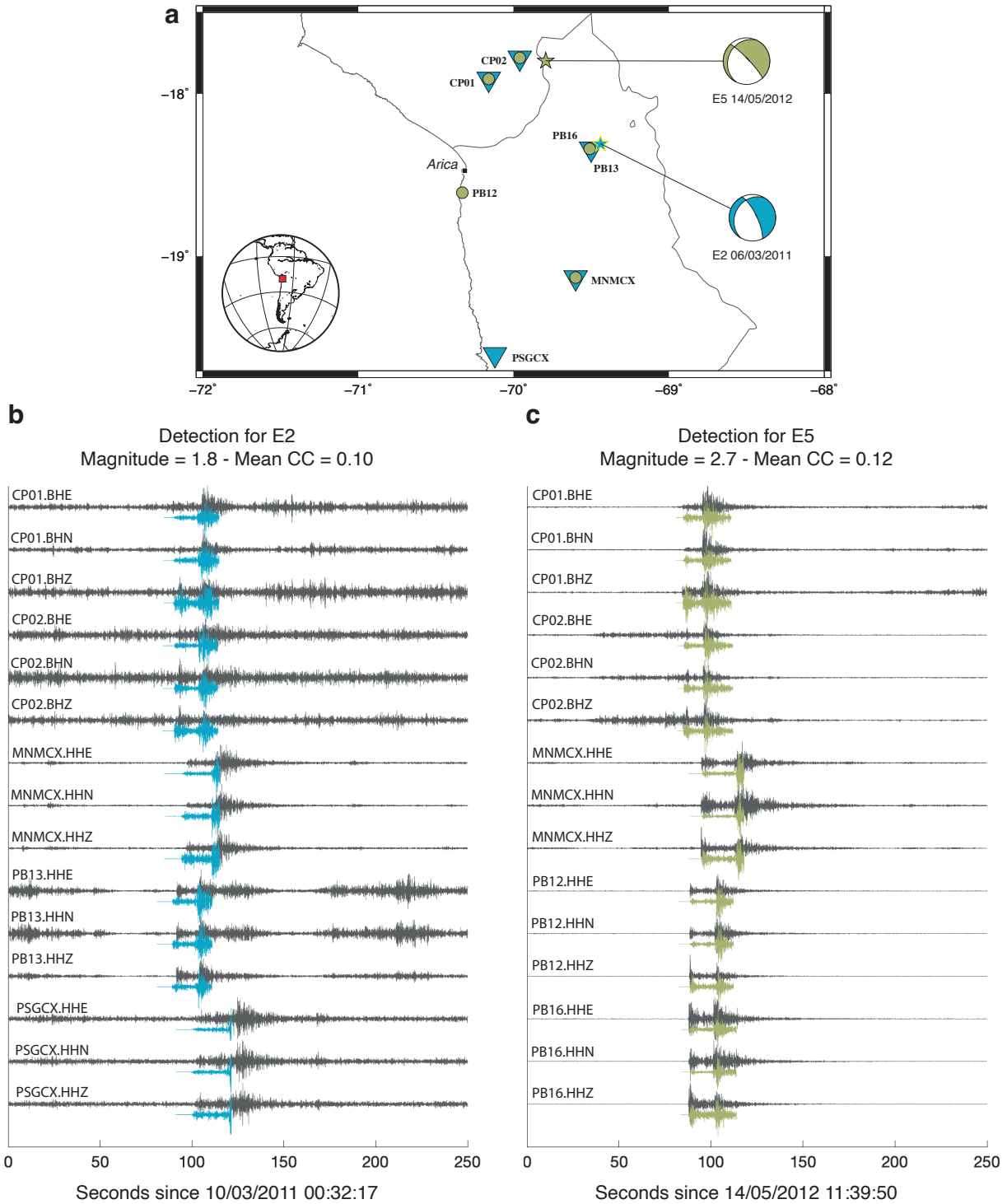


671

672 **Figure 1.** Seismotectonic setting of the northern Chile region. **a** Orange circles indicate the mainshocks (labeled as E1, E2 and so
 673 on according to their distance from the top of the slab) and aftershocks epicenters (CSN catalogue), scaled and colored by
 674 magnitude and depth. Focal mechanisms are obtained from USGS, and also date and magnitude according to CSN are indicated.
 675 Black inverted triangles and cyan squares are the locations of the stations used for the kinematic inversions (KI) and template
 676 matching (TM), respectively. Yellow triangles represent active volcanoes. Red lines indicate faulting located on the outer-rise
 677 region. Isochrones for the Nazca plate (Müller et al., 2008) are indicated with magenta lines. (Inset) Global map with the

678 specified study area. **b**, **c** and **d** cross sections along segments AA', BB' and CC' respectively in **a**. Distance from the top of the
679 slab is indicated for each mainshock. The slab (black solid line) corresponds to the projection along cross-sections of the Nazca
680 slab model SLAB2.0 (Hayes et al., 2018). Black and green dashed lines correspond to continental Moho (Tassara & Echaurren,
681 2012) and oceanic Moho (Contreras-Reyes et al., 2012), respectively.

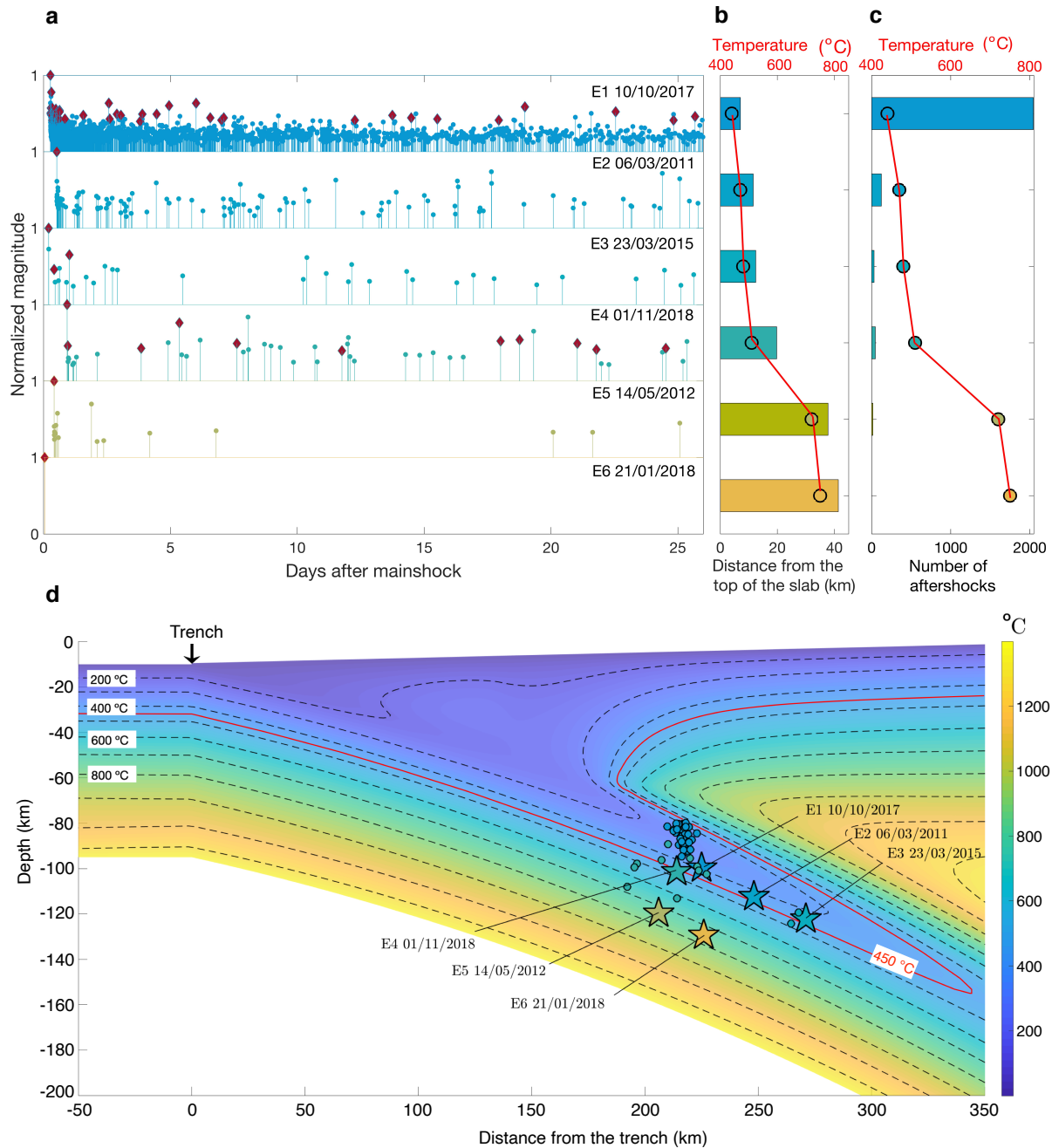
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684 **Figure 2.** Example of new events detected using template matching. **a** Map of stations used for E2 (cyan inverted triangles) and
 685 E5 (green dots). Stars represent epicenter for each event (CSN), and their focal mechanism according to USGS are also plotted.
 686 (Inset) Global map with the specified study area. **b** Example of a new detection in the continuous data for E2. Cyan traces
 687 represent templates aligned considering their respective moveouts. The average correlation coefficient (Mean CC) is 0.10 and the
 688 inferred magnitude (M_w) 1.8. **c** Example of a new detection in the continuous data for E5. Green traces represent templates
 689 aligned considering their respective moveouts. The average correlation coefficient (Mean CC) is 0.12 and the inferred magnitude
 690 (M_w) for the new event is 2.7.

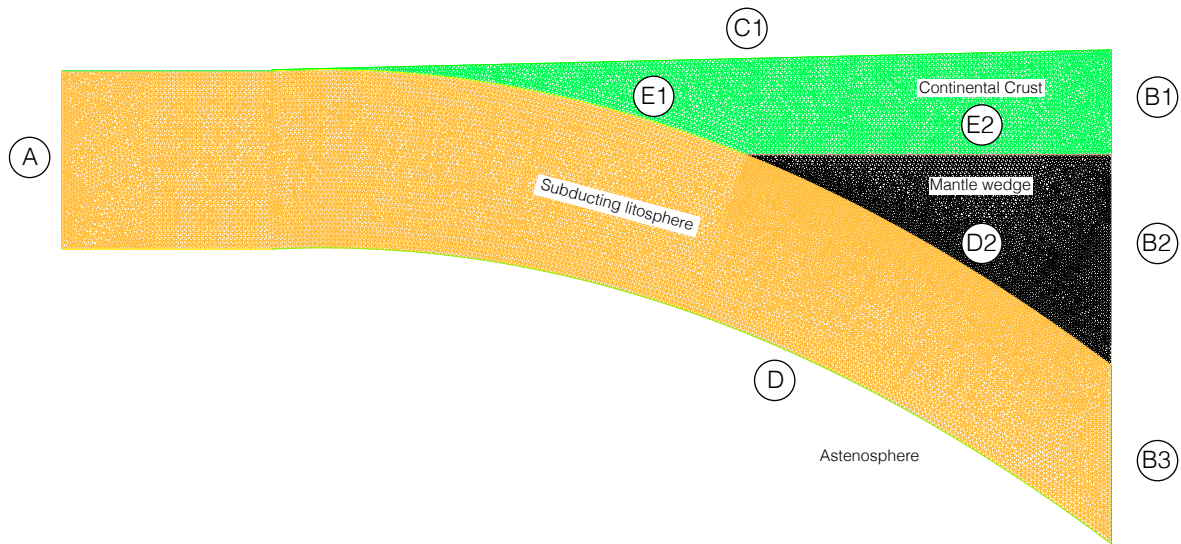
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693 **Figure 3.** Results for aftershocks efficiency and thermal model. **a** Detections using TM. For each row, stems show aftershocks
 694 detections until 25 days after mainshocks using TM as indicated in section 2.1. Stem height is normalized by the mainshock
 695 magnitude, and templates used are indicated as magenta diamonds. **b** Distance from the top of slab (Figure 1) is indicated with
 696 horizontal bars, and red line shows estimated temperature for hypocenter location according to the thermal model in **d**. **c**
 697 Horizontal bars indicate total number of aftershocks and red line the temperature. **d** Thermal model for northern Chile.

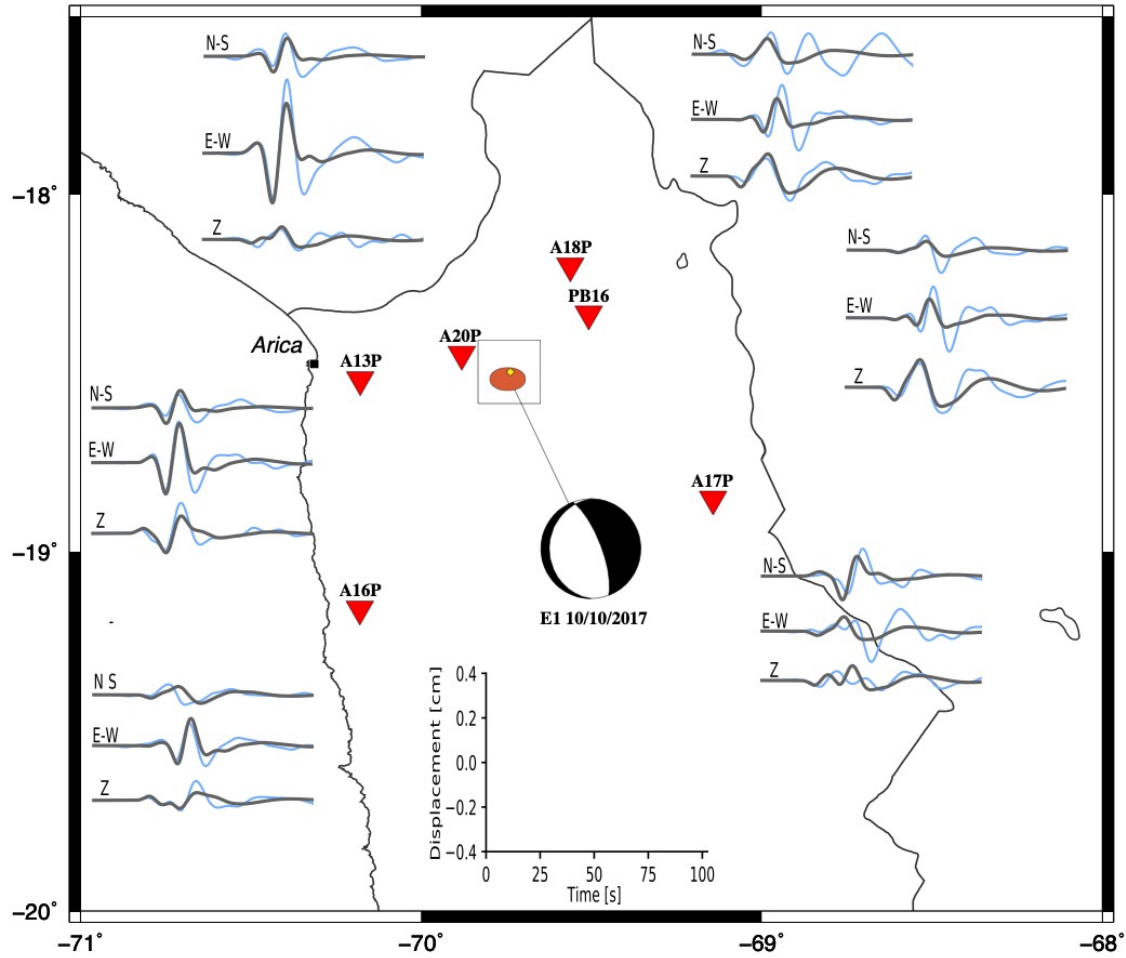
698 Mainshock hypocenters and its aftershocks reported by the CSN are indicated with stars and circles, respectively. In addition,
699 trench and dates of events are labeled.



700

701 **Figure 4.** FEM grid and boundary conditions used for thermal numerical modelling. Continental crust, mantle wedge and oceanic
702 plate are indicated with green, black, and light brown colors, respectively. Capital letters circled represent where border conditions
703 are imposed (see details in Text S1).

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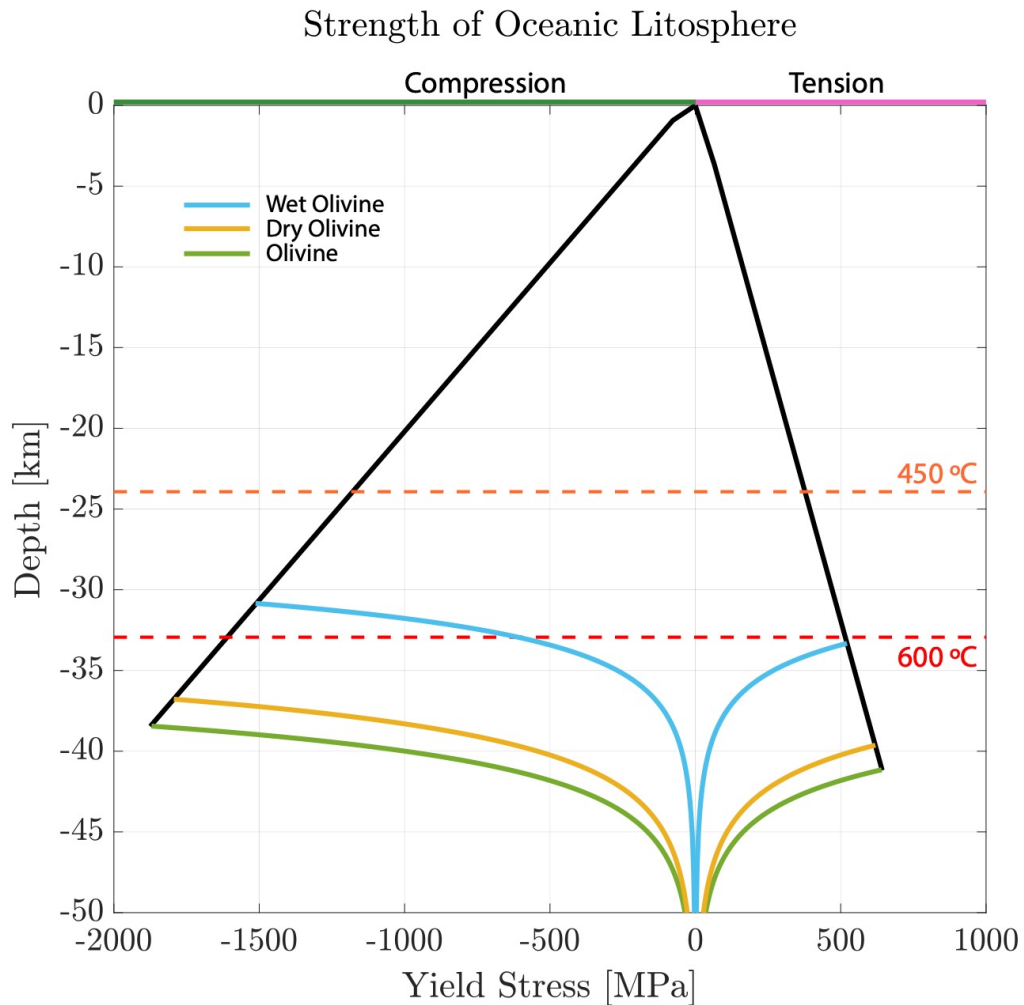
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Figure 5. Example of kinematic inversion for E1. Ellipse shows the rupture area of the mainshock, with at maximum slip amplitude of 1.08 m. Yellow diamond is the National Seismological Center (CSN) mainshock epicenter. Inverted red triangles are near-field stations considered for inversion. Gray and blue lines are simulated and observed displacement waveforms respectively. Focal mechanism from USGS.

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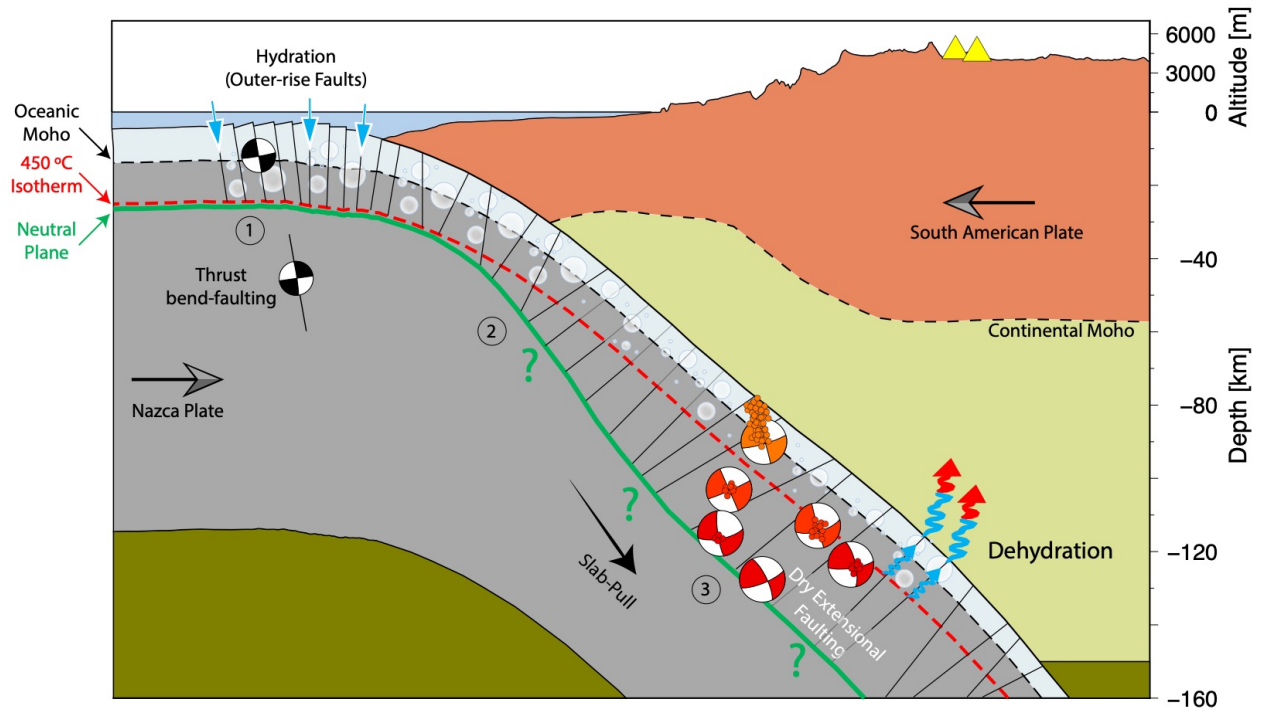
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713 **Figure 6.** Yield strength envelope for an oceanic lithosphere 58 Myr old. Stress differences or yield stress are limited at the top of
 714 the plate by frictional sliding rule according to Byerlee's frictional sliding rule (black curve). Yield stress is limited at the base of
 715 the plate by steady state creep, which depends on the cube of the stress and exponentially on temperature (Kohlstedt et al., 1995),
 716 and computed considering steady state flow properties for olivine (green line; Goetze, 1978), dry olivine (yellow line; Karato et
 717 al., 1986) and wet olivine (cyan line; Karato et al., 1986). Isotherms were computed based on the cooling of a semi-infinite half-
 718 space mode (Turcotte & Schubert, 2002).

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720

721 **Figure 7.** Cartoon showing a conceptual model for the evolution of subducting lithosphere in northern Chile. The topography
 722 of the plate in the outer-rise/trench region has been exaggerated to show better the deformation associated to plate bending. Scale
 723 is approximate everywhere else. Bubbles in oceanic lithosphere indicate hydration.

724