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Northern Chile Intermediate-Depth Earthquakes Controlled by Plate Hydration

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Abstract

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

Keywords: Intralab Intermediate Depth Earthquakes, Aftershock, Chile, Subduction, Slab.

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

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

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

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

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

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

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

3 Aftershock productivity

3.1 Detection of aftershocks using template matching

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

TM has been widely used to detect events in aftershocks sequences, where seismicity is known to repeat (Shelly et al., 2007; Peng & Zhao, 2009; Frank et al., 2017; Li et al., 2018). In a similar way, we here apply TM to improve the detection of aftershocks following the analyzed events (Figure 1). We base our analysis on continuous three-component velocity seismological data from the CSN, IPOC, and CAUGHT seismic networks, using the five nearest stations to the source region of each event (Figure S15). Data were continuously recorded at a sampling rate of 100 Hz for IPOC and CSN stations, and 40 Hz for CAUGHT stations. To detect non-reported aftershocks, we searched for events with similar waveforms to those reported as mainshock and aftershocks for the CSN (template candidates, see Supplementary Information). A space-time 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 $\text{SNR} > 5$ 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

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

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

3.2 Aftershock Activity

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

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

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

4 Thermal structure

4.1 Thermal-Fluid Numerical Model

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

4.1.1 Numerical Model

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

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

in the region of interest Ω , where T is the temperature and the first term corresponds to heat diffusion where κ is the thermal diffusivity depending on space (ratio of the thermal conductivity k and the thermal capacity ρc_p of the media), the second term corresponds to the heat transport due to lithospheric speed and asthenospheric fluxes $\vec{u}$, and the last term stands for the external heat sources such as radiogenic or frictional heat. No friction heat effects are considered in this study 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:

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

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:

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

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

supplementary information for more details.

4.1.2 Thermal Model Description (Parameter Selection)

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

4.2 Thermal Conditions of the Slab

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

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

5 Seismic Source properties

5.1 Seismic Source Kinematic Inversions

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

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

5.2 Kinematic Parameters

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

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

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

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

6 Discussion

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

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

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

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

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

bend-faulting occur and lead to partial hydration of the crust and upper mantle. However, fluids may not be able to penetrate any deeper than neutral plane (approximately 450 °C isotherm-depth). Thus, the neutral plane separates a high-hydrated from a dry or poorly hydrated zones.

(2) As the oceanic plate subducts, a deepening of the neutral plane occurs by the increase of the slab-pull forces, separating it from the 450 °C isotherm, and increasing the region subject to extensional failure. (3) Normal (tensional) events occur at different distances from the top of the slab, but their behavior is mostly controlled by variable physical background conditions. In particular, in the high-hydrated region a greater number of aftershocks is observed, differently from what is observed in the dry deeper zone.

7 Conclusions

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

413 **Declaration of competing interest**

414 The authors declare that they have no known competing financial interests or personal
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416

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665

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_{max}</i> (m)	<i>V_r</i> (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.

670

Figure 1. Seismotectonic setting of the northern Chile region. **a** Orange circles indicate the mainshocks (labeled as E1, E2 and so on according to their distance from the top of the slab) and aftershocks epicenters (CSN catalogue), scaled and colored by magnitude and depth. Focal mechanisms are obtained from USGS, and also date and magnitude according to CSN are indicated. Black inverted triangles and cyan squares are the locations of the stations used for the kinematic inversions (KI) and template matching (TM), respectively. Yellow triangles represent active volcanoes. Red lines indicate faulting located on the outer-rise 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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Figure 2. Example of new events detected using template matching. **a** Map of stations used for E2 (cyan inverted triangles) and E5 (green dots). Stars represent epicenter for each event (CSN), and their focal mechanism according to USGS are also plotted. (Inset) Global map with the specified study area. **b** Example of a new detection in the continuous data for E2. Cyan traces represent templates aligned considering their respective moveouts. The average correlation coefficient (Mean CC) is 0.10 and the inferred magnitude (M_w) 1.8. **c** Example of a new detection in the continuous data for E5. Green traces represent templates aligned considering their respective moveouts. The average correlation coefficient (Mean CC) is 0.12 and the inferred magnitude (M_w) for the new event is 2.7.

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

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

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

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.

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

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