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# Interplay of seismic and a-seismic deformation during the 2020 sequence of Atacama, Chile.

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## Abstract

An earthquake sequence occurred in the Atacama region of Chile throughout September 2020. The sequence initiated by a mainshock of magnitude  $M_w = 6.9$ , followed 17 hours later by a  $M_w = 6.4$  aftershock. The sequence lasted several weeks, during which more than a thousand events larger than  $M_l = 1$  occurred, including several larger earthquakes of magnitudes between 5.5 and 6.4. Using a dense network that includes broad-band, strong motion and GPS sites, we study in details the seismic sources of the mainshock and its largest aftershock, the afterslip they generate and their aftershock, shedding light of the spatial temporal evolution of seismic and aseismic slip during the sequence. Dynamic inversions show that the two largest earthquakes are located on the subduction interface and have a stress drop and rupture times which are characteristics of subduction earthquakes. The mainshock and the aftershocks, localised in a 3D velocity model, occur in a narrow region of interseismic cou-

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28 pling (ranging 40%-80%), *i.e.* between two large highly coupled areas, North and South of  
29 the sequence, both ruptured by the great  $M_w \sim 8.5$  1922 megathrust earthquake. High rate  
30 GPS data (1 Hz) allow to determine instantaneous coseismic displacements and to infer co-  
31 seismic slip models, not contaminated by early afterslip. We find that the total slip over 24  
32 hours inferred from precise daily solutions is larger than the sum of the two instantaneous  
33 coseismic slip models. Differencing the two models indicates that rapid aseismic slip de-  
34 veloped up-dip the mainshock rupture area and down-dip of the largest aftershock. During  
35 the 17 hours separating the two earthquakes, micro-seismicity migrated from the mainshock  
36 rupture area up-dip towards the epicenter of the  $M_w$  6.4 aftershocks and continued to propa-  
37 gate upwards at  $\sim 0.7$  km/day. The bulk of the afterslip is located up-dip the mainshock and  
38 down-dip the largest aftershock, and is accompanied with the migration of seismicity, from  
39 the mainshock rupture to the aftershock area, suggesting that this aseismic slip triggered the  
40  $M_w = 6.4$  aftershock. Unusually large post-seismic slip, equivalent to  $M_w = 6.8$  developed  
41 during three weeks to the North, in low coupling areas located both up-dip and downdip  
42 the narrow strip of higher coupling, and possibly connecting to the area of the deep Slow  
43 Sleep Event detected in the Copiapo area in 2014. The sequence highlights how seismic and  
44 aseismic slip interacted and witness short scale lateral variations of friction properties at the  
45 megathrust.

**Key Words:**

Chilean subduction zone, Atacama, earthquake sequence, GPS, post seismic, seismology

**1 Introduction**

The Atacama region (26°S-30°S) is one of the long lasting seismic gaps of Chile [Lomnitz, 2004; Métois *et al.*, 2016; Ruiz and Madariaga, 2018]. In this region, the last megathrust earthquake occurred in 1922, a  $M_w$  8.6 event that stroke North-Central Chile and triggered a transpacific tsunami [Willis, 1929; Beck *et al.*, 1998; Soloviev and Go, 1976; Ruiz and Madariaga, 2018; Kanamori *et al.*, 2019]. After 1922, the largest earthquake that occurred in the area was in 1983 with a magnitude 7.7 [Pacheco and Sykes, 1992; Comte *et al.*, 1992]. More recently, in 2013, an event of magnitude 6.8 located around 50 km depth occurred, probably at the bottom-end of the seismogenic zone along the plate interface. A decade of survey GPS measurements conducted in this region revealed two large highly coupled zones, the Atacama and the Chañaral segments, separated by a relatively large intersegment of intermediate to low coupling, named the Baranquilla low coupling zone (LCZ) [Métois *et al.*, 2013; Métois *et al.*, 2016; Klein *et al.*, 2018a]. Additionally, a 1.5 year-long,  $M_w \sim 7$ , slow slip event (SSE) was also detected in the region in 2014, but occurred deeper (40-60 km) than usual seismogenic depths (10-40 km) [Klein *et al.*, 2018b]. A detailed analysis of the only continuous GPS site in the region at this time also revealed two episodes of transient deformation, prior to the 2014 event, in 2005 and 2009, suggesting a possible recurrence of about 5-years for deep slow slip events in the region [Klein *et al.*, 2018b].

Here, we study a large seismic sequence that occurred in the Atacama region throughout September 2020 (Fig. 1), South of an area where seismic swarms have occurred several times in the past, *i.e.* in 1973, 1976 and 2016, offshore the town of Caldera [27°S, Fig. 1, Comte *et al.*, 2002; Holtkamp *et al.*, 2011]. The 2020 sequence initiated on September, 1<sup>st</sup>, at 04:09 UTC, with an earthquake of magnitude 6.9. It was followed 20 minutes later by an event of magnitude 6.3, close to the mainshock epicenter and 17 hours later, at 21:09 UTC, by another event of magnitude 6.4, the largest aftershock of the entire sequence, 20 km updip the mainshock. Overall, the sequence lasted several weeks with more than a thousand events and includes several large earthquakes of magnitude larger than 5. We use a complete set of seismological sites deployed in the area prior to the sequence that includes broad-band, strong motion and GPS to monitor the spatio-temporal evolution of this sequence (Fig. 1).

78 Thanks to this dense network we greatly improve the threshold detection down to magnitude  
 79 1 (with a magnitude of completeness of 2.5) and the precision of the localisation through a  
 80 3-D refined velocity model. Focusing on the first day, we compare the high rate and the daily  
 81 GPS solutions to quantify the amount of seismic and aseismic deformation that took place  
 82 after the mainshock. Finally, we discuss how this sequence takes place in the earthquakes  
 83 history of this area and how it may alter the potential seismic hazard of the nearby highly  
 84 coupled zones.

## 85 **2 Seismic analysis**

86 The Atacama region is poorly covered by the national seismic network (CSN, *Centro*  
 87 *Sismológico Nacional*, University of Chile, Santiago) with only 2 broad-band stations at less  
 88 than 100 km from the sequence. Since 2013, less than 2000 earthquakes have been located in  
 89 Chile between latitudes 30° S and 26° S.

### 90 **2.1 Building the sequence catalogue**

91 We built a catalogue using data from 14 broad-band stations of the CSN in a 300 km  
 92 radius around the sequence, completed by data from three semi-permanent stations in the  
 93 Copiapo region (30-150 km North), 10 temporary stations between Vallenar and Ovalle  
 94 (100-300 km South), and 30 stations from the national strong-motion network of the CSN  
 95 [Barrientos, 2018; Leyton *et al.*, 2018] providing data only for the 16 largest events (Fig. 1).  
 96 Event detection was performed by STA/LTA method using the six closest stations, with two  
 97 constraints: firstly, one of the three closest stations had to be first in triggering a detection  
 98 and secondly, each event had to trigger detections on at least 5 stations to be considered.  
 99 These criteria geographically restricted the area of study and filtered out the smallest local  
 100 events and the hundreds of earthquakes happening everyday in Chile. Between the 25<sup>th</sup> of  
 101 August and the 25<sup>th</sup> of September included, 1354 events have been detected, out of which  
 102 50 % happened within the first four days of the sequence. No significant raise in seismic ac-  
 103 tivity was detected prior to the main event: 1 to 9 events/day occurred between the 25<sup>th</sup> and  
 104 the 31<sup>st</sup> of August (Fig. 1-B). Manual P- and S-wave arrival-times readings were performed,  
 105 leading to 916 earthquake locations out of which 843 events belong to the dense core of the  
 106 sequence and 74 correspond to surrounding activity that may or may not be related to the se-  
 107 quence. Specifically, half of these (35 events) occur up North in a 80 x 80 km<sup>2</sup> area, 11 are

108 located further inland, 15 are poorly located beyond the trench or very deep below the con-  
109 tact and the last 13 correspond to quarry blasts.

110 Earthquake locations were determined by a double-difference approach in a regional  
111 3D velocity model obtained by regional tomography [Potin *et al.*, 2019]. Figure 2 represents  
112 a trench-perpendicular vertical cross-section across the sequence, with P-wave velocities and  
113 P- over S-wave velocity ratios [based on earthquakes arrival times, Potin *et al.*, 2019]. The  
114 seismicity associated with the sequence is located at the interface, mainly between 15 km and  
115 40 km deep, with some events scattered within the first 15 km of the upper plate. The back-  
116 ground seismicity visible on Fig. 2, located within a 50 km range on both sides of the cross-  
117 section, appears to extend within the plunging oceanic plate and can be interpreted as the  
118 double seismic zone observed in several places along the Chilean coast [Bloch *et al.*, 2014,  
119 2018; Comte and Suarez, 1994; Sippl *et al.*, 2018], although these events are poorly located  
120 due to the lack of local observations. P-wave velocities and P- over S-wave velocity ratios for  
121 the upper plate, the interface and the upper oceanic mantle are consistent with others local  
122 tomographic models obtained in northern Chile [Pastén-Araya *et al.*, 2018, 2021].

123 Figure 3 shows the spatio-temporal evolution of the seismicity over the first 72 h fol-  
124 lowing the mainshock. Immediately after the mainshock, seismicity spread over a  $20 \times 20 \text{ km}^2$   
125 region, a size roughly consistent with the rupture area (Fig. 3-C and 3-D). This initial spatial  
126 extension shows the area of influence of the stress increase due to the mainshock. Through-  
127 out these first 72 hours, both the North-South and Eastern (downdip) boundaries of seismic-  
128 ity remain stable. On the contrary, seismicity slowly spreads updip (westward), with an av-  
129 erage velocity of approximately  $\sim 0.7 \text{ km/hour}$  (considering a dip of  $20^\circ$ ; red dashed line on  
130 Fig. 3-C), resulting in almost doubling the initial area of aftershock.

## 131 2.2 Moment magnitudes

132 To constrain the magnitude of the largest events of the sequence, we perform regional  
133 W-phase source inversions [Duputel *et al.*, 2012] combined with a bootstrap analysis [Efron  
134 and Tibshirani, 1993]. We use broad-band velocimetric data from the Federation of Digi-  
135 tal Seismic Networks (FDSN) (C, C1 (doi.org/10.7914/SN/C1), CX (doi:10.14470/  
136 PK615318), G (doi:10.18715/GEOSCOPE.G), GT (doi.org/10.7914/SN/GT) and II  
137 (doi.org/10.7914/SN/II) networks) within 26 degrees of epicentral distance. To im-  
138 prove the homogeneity of the data coverage, we select one station per cell in a  $100 \text{ km} \times$

139 100 km grid in the vicinity of the source. The used time window starts at the P-wave arrival  
 140 time. Its duration is 300 s for epicentral distances smaller than  $12^\circ$  and grows with distance  
 141 ( $15 \times \Delta s/\circ$ ) for farther stations. Waveforms are filtered using a frequency band-pass that  
 142 varies with the Global CMT magnitude. Here we filter between 50-80 s and 120-250 s. The  
 143 average  $M_w$  and  $\pm 2\sigma$  uncertainties are:  $6.87 \pm 0.07$ ,  $6.29 \pm 0.04$ ,  $6.42 \pm 0.07$  for the events  
 144 that occurred on 2020/09/01 at 04:09 UTC, 04:30 UTC and 21:09 UTC. The bootstrap his-  
 145 tograms are shown on Fig. 4 and estimated parameters are gathered in the supporting infor-  
 146 mation.

### 147 **2.3 Characterisation of the sequence: Mainshock-Aftershock sequence or Seismic** 148 **swarm ?**

149 To evaluate the difference of the 2020 Atacama seismicity compared to a standard  
 150 mainshock-aftershock sequence, we analyze earthquake sizes and temporal distribution in  
 151 the area. Considering seismic events in the epicentral area since 2017 in the CSN catalog, we  
 152 estimate a b-value of  $b = 0.8 \pm 0.2$  using the *Aki* [1965] approach (consistently, we estimate  
 153  $b = 0.7 \pm 0.1$  for the 2020 Atacama sequence using the catalog presented in section 2.1). The  
 154 time of aftershocks relative to the  $M_w = 6.9$  mainshock is consistent with the Omori-Utsu  
 155 law  $r(t) = K(t + c)^{-p}$  with  $p = 1.0$ ,  $c = 0.1$  days and  $K = 16.3$  [see Fig. S1 of the Sup-  
 156 porting Information; *Omori*, 1894; *Utsu*, 1957]. Looking independently into the magnitude  
 157 and temporal distribution of the earthquakes, the sequence does not seem different from a  
 158 classical mainshock-aftershock sequence. However, what seems anomalous is the occurrence  
 159 of two  $M_w > 6$  aftershocks within 24 hours after the mainshock. Using a simple approach  
 160 similar to *Reasenber and Jones* [1989], we forecast the number of aftershocks of magnitude  
 161  $M_w \geq 6.3$  within 24 hours after the mainshock using  $b = 0.8$  and Omori-Utsu parame-  
 162 ters mentioned above. Results shown in Fig. S1 indicate that there is only a probability of  
 163 0.3% of having at least two aftershocks of magnitude  $M_w \geq 6.3$  shortly after the mainshock.  
 164 However, this estimate depends on the assumed b-value. If we consider  $b = 0.7$  as for the  
 165 Atacama sequence, the aforementioned probability increases to about 4%.

### 166 **3 GPS data analysis**

167 Early 2019, in order to densify the CSN network [*Báez et al.*, 2018], we installed 5  
 168 continuous GPS (cGPS) stations in the Atacama region. Three of them were collocated with  
 169 broad-band seismometers (see section 2.1). Overall, we benefit from 12 cGPS stations lo-

170 cated in the area of the sequence complemented by 5-6 stations further away for the refer-  
 171 ence (Fig. 1). In this study, we use both the stations positions obtained from 24-hours daily  
 172 solutions throughout the whole duration of the sequence and the high rate (1 Hz, hereafter  
 173 HRGPS) data that allow to decipher the successive displacements during the first day.

### 174 3.1 24-hours daily solutions

175 In addition to the data from the national Chilean network [Báez *et al.*, 2018] and from  
 176 the 5 additional stations, we include data of the Argentinian RAMSAC network [Piñón *et al.*,  
 177 2018] and of the Brazilian RBMC network. We also include all the IGS stations available on  
 178 the South American continent. This dataset is processed using the GAMIT/GLOBK software  
 179 following the classical MIT methodology [Herring *et al.*, 2010a,b]. In a second step, we pro-  
 180 duce daily time series by constraining continental stations to their well-known coordinates in  
 181 the ITRF2014 [Altamimi *et al.*, 2017] with the PYACS toolbox.

182 A specific difficulty needs to be addressed when several large earthquakes occur during  
 183 the same day. If a single coordinate is calculated for the entire day, it will end up being any-  
 184 where between the pre- and post-earthquakes coordinate, depending on different parameters:  
 185 when exactly the earthquakes occur during the day, which data segment (before, between and  
 186 after the earthquakes) is the longest, and how the filter will handle data that does not fit the  
 187 obtained average position of the day. In order to eliminate the pre-seismic observation (be-  
 188 fore 4:30 UTC) and to separate the two events in the data (see Fig. S2), we consider at which  
 189 time the two main events occurred ( $M_w = 6.9$  at 4:09 UTC and  $M_w = 6.4$  at 21:09 UTC)  
 190 and the day of the earthquake was processed using only the observations acquired between  
 191 4:30 UTC and 21:00 UTC. Therefore, this day's position corresponds to an averaged position  
 192 of the station after the first event ( $M_w = 6.9$ ) and before the second event ( $M_w = 6.4$ ). Note  
 193 that the selected time window also allows us to exclude the  $M_w = 6.3$  aftershock. Because  
 194 only 25 min separates this aftershock from the mainshock, the potential deformation gener-  
 195 ated by this aftershock is most likely impossible to differentiate from the mainshock, using  
 196 daily GPS solutions.

197 Time series reveal significant displacements on at least 7 stations (Fig. 5). Steps be-  
 198 tween days 244 and 245 (resp. 245 and 246) correspond to the coseismic displacements  
 199 generated by the first (resp. the second) event, both occurring during day 245 (September  
 200 1<sup>st</sup>). The typical curvature of the time series of the stations nearest the events during the



201 days (possibly weeks) following the mainshock also reveals postseismic deformation. This  
 202 deformation seems unusually large ( $\sim 100\%$  in only a couple of days) at the nearest station  
 203 (TTRL). The estimation of the coseismic displacement of the first event of  $M_w = 6.9$  at 04:09  
 204 (the mainshock) is obtained by differentiating between the position at midday 245 (between  
 205 4:30 and 21:00 UTC) and the position of the day before (244) (Fig. 6-A, vectors in light red).  
 206 It includes part, but not all, of the post-seismic deformation occurring during the 15 hours  
 207 time span between the mainshock and the large aftershock at 21:09, which is potentially a  
 208 combination of rapid after-slip and a-seismic deformation, but also potential deformation  
 209 due to the  $M_w = 6.3$  aftershock of 04:30. The estimation of the co-seismic displacement of  
 210 the second event ( $M_w = 6.4$ ) is obtained by difference between the position of the following  
 211 day (246 - 02/09/2020) and the position of the day of the 2 earthquakes previously described  
 212 (midday 245, between 4h30 and 21:00 UTC). In a similar way, it also includes a combination  
 213 of rapid after-slip and potential a-seismic deformation that might have occurred after both  
 214 events (Fig. 6-B, vectors in light blue).

### 215 3.2 High rate GPS observations

216 High rate data are processed with Track software from MIT (T. Herring) which is a  
 217 double-difference software, meaning that we compute the motion of a "rover" station relative  
 218 to a "fixed" station. In this processing we use 5 "fixed" stations surrounding the area of inter-  
 219 est (represented by black diamonds on Fig. 1): LSCH (La Serena) and LHOR (LosHornos)  
 220 to the South; PAZU (Pan de Azucar) to the North; ALUM in Argentina and MRCG (Mar-  
 221 icunga) to the East and North-East. We use the tropospheric zenith delays (ZTD) gener-  
 222 ated by the 24 h static solution (one delay estimated every 2 hours at every site) to constrain  
 223 the tropospheric delay in the kinematic processing to the static value. For the three largest  
 224 events, we generate motograms (high rate evolution of position with time, from the latin  
 225 word "moto" for motion) of one hour spanning the events (see Fig. S5 for the mainshock at  
 226 4:09 UTC, Fig. S6 for the largest aftershock at 21h09 UTC and Fig. S7 for the smaller after-  
 227 shock at 4:30 UTC). For all motograms, we built a sidereal filter by simply stacking the 1-  
 228 hour data segments, of 3 (or 6) days before the earthquake with a 4 m 7 s time delay everyday  
 229 following *Choi et al.* [2004]. We then filter the co-seismic motogram, by simply subtract-  
 230 ing this common mode to the original data. Then, the co-seismic jump is simply estimated  
 231 as the offset between the 3-minutes data segment before and after the time of the earthquake  
 232 (Fig. 6). Uncertainties are estimated visually from the motograms and range between 1 and

233 5 mm for the horizontal components and 5 and 10 mm for the vertical component. We are  
 234 able to identify clear co-seismic jumps at most stations for the mainshock, small but discern-  
 235 able jumps at several stations for the largest aftershock, but nothing for the smaller aftershock  
 236 of 4:30 UTC. This is an indication of the threshold detection of our current cGPS network :  
 237 between magnitude 6.3 and 6.4.

238 Comparing coseismic offsets extracted from both the daily solution and from the HRGPS  
 239 solution offers some confidence. Although the HRGPS is associated with larger uncertainties  
 240 (5 mm) than daily solutions (1-2 mm), both solutions appear very consistent and show very  
 241 similar offset. Specifically, stations located more than 50 km from the epicenter compare  
 242 very well (BING, MMOR, UDAT, TAMR, TOT5, TRST). However, for both events, near-  
 243 field stations (TTRL, BAR2, and LLCH) exhibit a smaller HRGPS coseismic offset (smaller  
 244 by 50%) than the daily solution one. This is very significant and indicates additional defor-  
 245 mation is present immediately after the earthquake occurrence.

## 246 **4 Analysis of major earthquakes**

### 247 **4.1 Coseismic slip static inversions**

248 We built a fault geometry with triangular patches based on Slab2.0 [*Hayes et al.*, 2018]  
 249 between 26.5°S and 29°S and down to 60 km depth. We evaluated the slip distributions gen-  
 250 erated by the two largest earthquakes by inverting the coseismic displacements estimated  
 251 from the HRGPS. We compute constrained least squares inversions using the *CSI* toolbox  
 252 [*Gombert et al.*, 2019]. For both models, we apply as little smoothing as possible and we  
 253 forbid back slip in the thrust direction. We assume only one slip component which direc-  
 254 tion is fixed parallel to the plate convergence [convergence vector from *Klein et al.*, 2018a].  
 255 Green functions are calculated at each node of the fault plane, assuming a homogeneous elas-  
 256 tic half-space [*Meade*, 2007].

257 10 to 14 stations spanning the area were used in the inversion (Fig. 6). Resolution tests  
 258 are fully described in the Supplemental material. They show that (1) a good recovery for  
 259 ~40 x40 km patterns is found between 15-55 km depth even with conservative noise budget  
 260 for co-seismic offsets; (2) a very good (1-2 km) ability to locate the area of maximum slip;  
 261 (3) peak-slip amounts are recovered within 10-30% and magnitude by 0.1; (4) extent of slip  
 262 might be smeared by a few km.

263 For the mainshock, we find a slip distribution spreading over a rather large surface of  
 264  $80 \times 40 \text{ km}^2$ , between  $27.5^\circ\text{S}$  and  $28.5^\circ\text{S}$ . This surface seems too large for a  $M_w$  6.9 earth-  
 265 quake, but the bulk of the slip is concentrated in a much smaller area of only about  $25 \times 20 \text{ km}^2$   
 266 (Fig. 7-A). There is a trade-off between the quantity of slip and the size of the rupture zone.  
 267 We test several models in which we concentrate larger slip amount in a narrower zone (for  
 268 ex. within the region currently yielding more than 60 mm, or more than 80 mm of slip, see  
 269 Fig. S8). Southward offsets can be reproduced by a larger amount of slip in the north (see  
 270 Fig.S8-B). But reducing the rupture zone to  $\sim 30 \times 30 \text{ km}^2$  leads to significantly larger resid-  
 271 uals at closest stations (BAR2 and TTRL, Fig. S8-C). Therefore, the extension of the rupture  
 272 zone to the north is required by the observations at more than 50 km, yielding significant  
 273 westward coseismic offsets which are not converging toward a pin point. The best fit model  
 274 includes a narrow strip of slip, elongated below the coastline south of the high slip area. This  
 275 feature depicts only several cm of slip and is requested only by millimetric variations at a  
 276 few stations. It may be beyond the resolution of our data and modelling. The deep extension  
 277 of slip, reaching 40 km down, observed at  $28^\circ\text{S}$  seems required both by the large coseismic  
 278 displacement measured at station TOT5 located some 75 km away from the epicenter, and  
 279 by the coseismic uplift measured at BAR2 and LLCH. Although vertical data do not appear  
 280 essential since an inversion considering only the horizontal coseismic displacements pro-  
 281 duces similar slip pattern. We tested models with pure dip slip direction perpendicular to the  
 282 trench, and models with two slip directions, but neither provides satisfying results (see sup-  
 283 porting information for more details). We estimate a seismic potency of  $4.14 \cdot 10^8 \text{ m.m}^2$ ,  
 284 which corresponds to a moment of  $2.01 \cdot 10^{19} \text{ N.m}$  ( $M_w = 6.8$ ) using a shear modulus of  
 285  $4.9 \cdot 10^{10} \text{ Pa}$  (which is the value used for the W-phase). The geodetic moment appears slightly  
 286 smaller than the seismic moment re-estimated at long-period using the W-phase but still lies  
 287 within the error bar. Considering the size of this event, we made the approximation of a ho-  
 288 mogeneous half-space for all our inversions, which could account for part of the difference.

289 For the largest aftershock, because it generates smaller displacements than the main-  
 290 shock at many stations, we dispose of less well determined co-seismic vectors. In particular  
 291 we do not use any vertical displacement in the inversion of the aftershock slip distribution.  
 292 Also, considering that we have very few observations, we decreased the uncertainties of non-  
 293 zero vectors to 1mm, in order to strongly encourage the model to fit these. We find a circu-  
 294 lar slip distribution, significantly smaller with about  $30 \times 30 \text{ km}$  overall (only  $10 \times 10 \text{ km}$   
 295 for the bulk of the slip), with a peak slip at 95 mm (Fig. 7-B). For this event as well, the

geodetic moment also appears slightly larger than the seismologic one, with  $6.04 \cdot 10^{18}$  N.m ( $M_w = 6.5$ , corresponding to a seismic potency of  $1.24 \cdot 10^8$  m.m<sup>2</sup>). Finally, the epicenter of the mainshock is located on the updip-western edge of the rupture zone, suggesting a downdip-bilateral propagation. The aftershock slip distribution is located updip the mainshock rupture zone and shows a striking complementary (Fig. 7-C). The aftershock lies in the hole left by the bean-shaped mainshock. Together they homogenise the slip over a larger and rounder area.

## 4.2 Dynamic inversions

We used seismic waveforms from strong motion stations deployed in the area (Fig. 1) to estimate the dynamic properties of the coseismic rupture. The low-frequency source properties (e.g., average slip and stress drop) of the largest event were estimated using an elliptical patch approach (e.g., [Ruiz and Madariaga, 2011; Herrera et al., 2017]). In this model, the rupture nucleates within a circular area and then propagates through a larger elliptical area. This rupture process is controlled by the friction law proposed by *Ida* [1972]. Hence, in addition to the geometric parameters defining the circle and the ellipse, this dynamic model also includes the stress drop ( $T_e$ ), the yield stress ( $T_\mu$ ), and the slip-weakening distance ( $D_c$ ). We used strong-motion records integrated to displacement and band-pass filtered in low frequency (0.02-0.2 Hz for the mainshock). The AXITRA code [Bouchon, 1981; Coutant, 1989] was used to simulate the source-to-receiver wave propagation via an appropriate 1-D velocity model for the area, which was extracted from [Potin et al., 2019]. The inversion of the dynamic model was performed using the Neighborhood Algorithm [Sambridge, 1999], which finds the model that best fits the observed waveform data. The misfit between observed and modeled waveforms was calculated using an L2 norm.

The best solution for the mainshock converged toward an elliptical rupture of 24.4 km by 26 km (Fig. 8), with a minimum misfit of 0.24 (Figs. 8 and S11), a maximum slip of 1.1 m and a  $M_w = 6.7$ , which is similar to the solution obtained from HRGPS (Fig. 7). Also, the associated dynamic parameters are  $T_e = 5.3$  MPa,  $T_\mu = 5.59$  MPa, and  $D_c = 0.72$  m. These dynamic parameters are similar to those obtained for inter plate events along Chilean subduction [Ruiz et al., 2017; Otarola et al., 2021] and the stress drop parameters are in the average of thrust earthquakes occurring on a subduction interface [Kanamori and Anderson, 1975].

## 5 Interplay between seismic and aseismic slip

### 5.1 Static inversion of afterslip on the day of the mainshock

On one hand, we compute the total co-seismic motion due to both events, both quantified by daily cGPS between the 30/08/2020 and the 02/09/2020 (shown in light red on Fig. 9-A). This calculation includes both events and the total amount of aseismic slip that occurred over the two days. On the other hand, we compute the total displacements measured by HRGPS (shown by dark red arrows on Fig. 9-A). Considering that the HRGPS allows to extract the pure co-seismic motion over a couple of minutes around the earthquakes, the difference between the total daily cGPS co-seismic estimates and the total HRGPS estimates (Fig. 9-B) should highlight the amount of early afterslip during the day of the earthquakes. Indeed, this difference shows a significant westward motion at TTRL and BAR2, similar to the post-seismic motion observed over the following days (Fig. 9-C).

Using the same methodology and parameters as previously (section 4.1), we compute static inversions of the different displacements fields. Unsurprisingly, the slip distribution inferred from the total daily cGPS displacements (Fig. 9-Ai, noted in following Ai) shows significantly more slip than the slip distribution inferred from the total HRGPS displacements (Fig. 9-Aii, noted in following Aii). In particular the peak slip of (Ai) reaches 17 cm compared to only 10 cm for (Aii). But both distributions show very similar patterns over a somehow circular area extending from 27.4°S to 28.5°S. The distribution of early postseismic shows slip occurring on a significantly smaller, narrow peanut-shape area elongated along a roughly NS direction (Fig. 9-B). Part of this slip could be coseismic slip due the  $M_w = 6.3$  aftershock which occurred at 4:30 UTC and that we were not able to extract from HRGPS. Small amount of slip observed at greater depth is most likely unresolved.

### 5.2 Time-dependent inversion of the postseismic deformation

Significant displacements are observed on the cGPS time series during a 22-days period, between the 2<sup>nd</sup> and the 24<sup>th</sup> of September. In order to quantify the slip evolution after the second large aftershock, we perform a kinematic inversion of the cGPS times series. We invert for slip on the subduction interface (following *Rolandone et al. [2018]* and *Bletery and Nocquet [2020]*). We find that the best fit to the time series is obtained with a smoothing parameter  $\sigma = 20 \text{ mm} \cdot \sqrt{\text{day}}$  and a correlation distance between subfaults of  $D_{corr} = 35 \text{ km}$ . The total slip after 22 days is equivalent to  $M_w = 6.8$ . Overall, it spreads over roughly the

358 same area as the area ruptured by the mainshock and its largest aftershock, between 27.5°S  
 359 and 28°S (Fig. 9-C). A static inversion of the cumulative postseismic displacement (follow-  
 360 ing the same methodology as the coseismic static inversions) over the same period yields  
 361 a very similar same pattern (Fig. S12). Regarding its spatio-temporal evolution, the post-  
 362 seismic slip begins offshore and starts developing onshore and deeper after 6 days (Fig. 9-C1  
 363 & 9-C2). At a later stage, on the ninth day, a dissociated smaller patch begins more to the  
 364 North, between 27°S and 27.5°S (Fig. 9-C3). It is deeper - at a depth of approximately 35 to  
 365 55 km - and localised in the updip vicinity of the 2014 slow slip event [*Klein et al.*, 2018b].  
 366 The northward migration of post-seismic slip is associated with a northward rotation of post-  
 367 seismic vectors wrt. co-seismic vectors at several stations near the epicenter area (BAR2,  
 368 LLCH, TTRL) and the development of Westward vectors North of the epicenter area (BING,  
 369 MMOR, UDAT). The source time function associated to this inversion shows a quasi-steady  
 370 decrease in the slip's intensity. Then, negligible slip is found to occur after approximately 18  
 371 days. A movie of the postseismic slip evolution is provided in the supporting information.

## 372 **6 Discussion and Conclusions**

### 373 **6.1 General agreement and small discrepancies**

374 Concerning the mainshock, the different types of modelling presented here are in good  
 375 agreement, with some discrepancies regarding the magnitudes, the size of slip distributions  
 376 and the peak slip. The dynamic model yields a smaller magnitude ( $M_w = 6.7$ ) than the one  
 377 inferred from the W-phase ( $M_w = 6.9$ ). This is common and due to the simple elliptical ge-  
 378 ometry used for the dynamic inversion, which can therefore not fully capture the correct slip  
 379 distribution and concentrate the solution. GPS constrained slip models yield a magnitude of  
 380 6.8, slightly smaller than the W-phase magnitude, but the difference is within the error bar  
 381 (cf section 2.2 and table S1, same observation for the difference in magnitude of the largest  
 382 aftershock between the static inversion and the W-phase analysis). Slip models inferred from  
 383 GPS show a quite larger rupture zone, which could have several origins. First, it could be an  
 384 artefact imposed by wrongly detected small displacements at stations located farther away  
 385 from the epicenter, although this should mostly be taken into account by the uncertainties.  
 386 Second, the geodetic models might be contaminated by inaccuracies in the Earth model as  
 387 we assume an homogeneous half-space and neglect topography [e.g., *Duputel et al.*, 2014;  
 388 *Langer et al.*, 2020]. Finally, the model resolution is limited by the number of observations,  
 389 resulting in a trade-off between the amount of slip and the size of the rupture. Eventually,

390 from both analyzes, we are confident that the greatest slip is well concentrated in an area of  
 391  $30 \times 30 \text{ km}^2$ , associated with the rupture of a single asperity. The HRGPS inversion shows a  
 392 more extended rupture area, the lesser slip regions probably being at the resolution limit of  
 393 our data.

## 394 **6.2 Relation with Coupling on the interface**

395 We compare the two slip distributions with the coupling distribution proposed in the  
 396 region by *Klein et al.* [2018a] (Fig. 10. The whole September sequence takes place in be-  
 397 tween the highly coupled Atacama segment (South of  $28^\circ\text{S}$ ), and the Chañaral segment (North  
 398 of  $27^\circ\text{S}$ ). There, in the so-called Baranquilla inter-segment, we observe a narrow strip highly  
 399 coupled connecting the 2 segments with significantly lower coupling on both the shallower  
 400 and deeper part of the interface. We find that most of the slip due to the 01/09/2020 main-  
 401 shock (dark blue contour) occurred downdip of its epicenter (dark blue dot), mostly overlap-  
 402 ping the narrow strip of higher coupling. The largest aftershock at 21:09 UTC (light blue  
 403 contour) shows a striking complementarity with the mainshock, occurring updip and ex-  
 404 tending in the low coupled region (Fig. 10). Early afterslip that occurs during the 17 hours  
 405 between the mainshock and the largest aftershock (Fig. 9-B), is located mostly between the  
 406 rupture zones of the two earthquakes, in a peanut-like shape (Fig. 10). Part of the obtained  
 407 slip could be coseismic due to the 4:30 UTC  $M_w = 6.3$  aftershock, and part indeed due to  
 408 aseismic slip.

## 409 **6.3 Interplay of seismic and aseismic slip in an area of heterogeneous coupling**

410 We showed that the probability of having at least two aftershocks of magnitude  $M_w >$   
 411 6 within 24 hours is quite low. This leads us to question whether it is a simple mainshock-  
 412 aftershock sequence or a seismic swarm, which is commonly defined as an increase of seis-  
 413 micity rate without a clear mainshock earthquake [*Holtkamp et al.*, 2011]. It could also mean  
 414 that there is room for other processes that could have triggered these earthquakes so shortly  
 415 after the mainshock.

416 During the first 17 hours, seismicity spread updip the mainshock epicenter and out-  
 417 side its rupture zone, into what later became the largest aftershock rupture zone, (Fig. 7-C).  
 418 The asymmetry observed between the updip and downdip propagations of seismicity over  
 419 the first 72 h (Fig. 3-C) is most likely driven by a specific source. Incidentally, the quantity

420 of aseismic slip occurring directly after the mainshock, and over the following 20 days, is  
 421 abnormally high. There is some overlap between the afterslip distribution, and the the co-  
 422 seismic slip distributions of the mainshock and the largest aftershock (Fig. 10). But the bulk  
 423 of the distributions are disconnected and the overlap lies within the regions of lesser slip.  
 424 We suggest that slow slip could be responsible for increasing shear stress at the front of the  
 425 slip zone, propagating updip at approximately 0.7 km/hour ([which is within the range of  
 426 slow slip propagation speeds observed elsewhere, *Gao et al.*, 2012] - Fig. 3-C), until sur-  
 427 rounding a locked asperity which eventually triggered the  $M_w = 6.4$  aftershock, 17 hours af-  
 428 ter the mainshock. Such a relation between seismicity at the front of the slip has been pro-  
 429 posed in various context, during the interseismic phase but also during SSE, associated or  
 430 not with non-volcanic tremors [*Bartlow et al.*, 2011, 2014; *Vaca et al.*, 2018; *Bletery and*  
 431 *Nocquet*, 2020], and is consistent with numerical models of seismicity driven by slow slip  
 432 [e.g. *Ariyoshi et al.*, 2012; *Yingdi and Ampuero*, 2017; *Wynants-Morel et al.*, 2020]

433 The equivalent moment released over a period of 22 days, following the mainshock,  
 434 reaches more than 80% of the coseismic moment, spreading in a much broader region than  
 435 the coseismic rupture zone where the coupling is lower, as well as in a broader region than  
 436 the aftershocks area. Usually, postseismic deformation reaches around 25% of the co-seismic  
 437 deformation after a month. However, several cases have been documented where moder-  
 438 ate size earthquakes are followed by abnormally large afterslip in Japan [*Yagi et al.*, 2001;  
 439 *Suito et al.*, 2011] and northern Peru where moment released through aseismic slip during  
 440 a sequence was several time larger (3 to 14) than the moment released through earthquakes  
 441 [*Villegas Lanza et al.*, 2015]. For the latter, it has been suggested that additional processes  
 442 - i.e. not only an earthquake but also, for example, one or several slow slip events - were  
 443 involved to explain such a large amount of afterslip. A similar hypothesis was proposed to  
 444 explain the abnormally rapid and large early afterslip following the 2016  $M_w$  7.8 Ecuador  
 445 earthquake [*Rolandone et al.*, 2018]. Complex sequence with large afterslip occurring very  
 446 close a the recurrent SSE patch was also observed in Mexico [*Radiguet et al.*, 2016]. The At-  
 447 acama region seems propitious to slow slip events, while such an event was observed in the  
 448 region in 2014 [*Klein et al.*, 2018b]. Here, overall, we estimated from the geodetic models  
 449 that the sequence released a total moment of  $4.94 \cdot 10^{19}$  N.m ( $M_w = 7.1$ ), with close to 60%  
 450 through earthquakes and 40% through aseismic slip. Slip occurred spread over an area of  
 451  $\sim 100 \times 100$  km<sup>2</sup>, much larger than expected for  $M < 7$  earthquakes, also highlighting the role  
 452 of aseismic slip during the sequence. Postseismic slip migrates to greater depth 6 days after



453 the mainshock, reaching eventually the 2014 slow slip area. Therefore, the Baranquilla LCZ  
454 seems prone for aseismic processes, potentially recurrent at depth as observed in the past,  
455 and favors large postseismic slip.

#### 456 **6.4 Considerations on seismic hazard in the area**

457 Considering the historical seismicity in the region, *i.e.* the  $M_w \geq 8.5$  mega-earthquakes  
458 of 1819 and 1922, and the high coupling imaged in the Atacama and Chañaral segments,  
459 we previously suggested that a joint rupture of these two segments was highly plausible in  
460 the future [Klein *et al.*, 2018b]. Both segments have indeed accumulated enough deforma-  
461 tion since 1922 to generate a  $M_w \geq 8$  earthquake [Klein *et al.*, 2018a]. What is the impact  
462 of this sequence regarding scenarios for future megathrust ruptures in the region ? Differ-  
463 ent scenarios seem plausible. On one hand, the whole September 2020 sequence is likely to  
464 have increased the stress at the edges of the highly locked Atacama and Chañaral segments,  
465 promoting future rupture(s) there. In particular the whole sequence occurred very near the  
466 northern edge of the Atacama segment. Could this initiate the destabilization of this highly  
467 locked patch and trigger a rupture of this segment already ? And would a rupture of the Ata-  
468 cama segment trigger in turn the rupture of the Chanaral segment, initiating a 1819 or 1922  
469 like megathrust earthquake ? On the other hand, this same sequence may have released a sig-  
470 nificant amount of stress in the Baranquilla LCZ, which could in turn decrease the potential  
471 for a joint rupture of the Atacama and Chañaral segments by reinforcing its ability to act as  
472 a barrier for megathrust rupture propagation. In this scenario, Atacama and Chañaral seg-  
473 ments could rupture independently, at different times and with smaller earthquakes than in  
474 1819 and 1922. It is difficult to decipher between these scenarios, but the occurrence of a  
475 seismic sequence between two highly locked patches identified to be responsible for devas-  
476 tating earthquakes 100 and 200 years ago is a clear sign that this region should be monitored  
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502 The coseismic offset tables extracted from daily and HRGPS presented in the study are  
503 in the supporting information. Position time series of the sequence can be made available  
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## 505 **References**

- 506 Aki, K. (1965), Maximum likelihood estimate of  $b$  in the formula  $\log N = a - bM$  and its confi-  
507 dence limits, *Bull. Earthq. Res. Inst., Tokyo Univ.*, 43, 237–239.
- 508 Altamimi, Z., L. Métivier, P. Rebischung, H. Rouby, and X. Collilieux (2017), Itrf2014 plate  
509 motion model, *Geophysical Journal International*, 209(3), 1906–1912.
- 510 Ariyoshi, K., T. Matsuzawa, J.-P. Ampuero, R. Nakata, T. Hori, Y. Kaneda, R. Hino, and  
511 A. Hasegawa (2012), Migration process of very low-frequency events based on a chain-  
512 reaction model and its application to the detection of preseismic slip for megathrust earth-  
513 quakes, *Earth, planets and space*, 64(8), 693–702.

- 514 Báez, J., F. Leyton, C. Troncoso, F. del Campo, M. Bevis, C. Vigny, M. Moreno, M. Si-  
515 mons, E. Kendrick, H. Parra, et al. (2018), The Chilean GNSS network: Current status  
516 and progress toward early warning applications, *Seismological Research Letters*.
- 517 Barrientos, S. (2018), The seismic network of Chile, *Seismological Research Letters*, 89(2A),  
518 467–474.
- 519 Bartlow, N. M., S. Miyazaki, A. M. Bradley, and P. Segall (2011), Space-time correlation of  
520 slip and tremor during the 2009 Cascadia slow slip event, *Geophysical Research Letters*,  
521 38(18).
- 522 Bartlow, N. M., L. M. Wallace, R. J. Beavan, S. Bannister, and P. Segall (2014), Time-  
523 dependent modeling of slow slip events and associated seismicity and tremor at the Hiku-  
524 rangi subduction zone, New Zealand, *Journal of Geophysical Research: Solid Earth*,  
525 119(1), 734–753.
- 526 Beck, S., S. Barrientos, E. Kausel, and M. Reyes (1998), Source characteristics of historic  
527 earthquakes along the central Chile subduction zone, *Journal of South American Earth  
528 Sciences*, 11, 115–129.
- 529 Blertey, Q., and J.-M. Nocquet (2020), Slip bursts during coalescence of slow slip events in  
530 Cascadia, *Nature communications*, 11(1), 1–6.
- 531 Bloch, W., J. Kummerow, P. Salazar, P. Wigger, and S. A. Shapiro (2014), High-resolution  
532 image of the North Chilean subduction zone: seismicity, reflectivity and fluids, *Geophysi-  
533 cal Journal International*, 197(3), 1744–1749.
- 534 Bloch, W., T. John, J. Kummerow, P. Salazar, O. S. Krüger, and S. A. Shapiro (2018), Watch-  
535 ing dehydration: Seismic indication for transient fluid pathways in the oceanic mantle of  
536 the subducting Nazca slab, *Geochemistry, Geophysics, Geosystems*, 19(9), 3189–3207,  
537 doi:10.1029/2018GC007703.
- 538 Bouchon, M. (1981), A simple method to calculate Green's functions for elastic layered me-  
539 dia, *Bulletin of the Seismological Society of America*, 71(4), 959–971.
- 540 Choi, K., A. Bilich, K. M. Larson, and P. Axelrad (2004), Modified sidereal filtering: Impli-  
541 cations for high-rate GPS positioning, *Geophysical Research Letters*, 31(22).
- 542 Comte, D., and G. Suarez (1994), An inverted double seismic zone in Chile: Evidence of  
543 phase transformation in the subducted slab, *Science*, 263(5144), 212–215.
- 544 Comte, D., M. Pardo, L. Dorbath, C. Dorbath, H. Haessler, L. Rivera, A. Cisternas, and  
545 L. Ponce (1992), Crustal seismicity and subduction morphology around Antofagasta, Chile:  
546 preliminary results from a microearthquake survey, *Tectonophysics*, 205(1-3), 13–22.

- 547 Comte, D., H. Haessler, L. Dorbath, M. Pardo, T. Monfret, A. Lavenu, B. Pontoise, and  
548 Y. Hello (2002), Seismicity and stress distribution in the Copiapo, northern Chile subduc-  
549 tion zone using combined on- and off-shore seismic observations, *Physics of the Earth and*  
550 *Planetary Interiors*, 132(1–3), 197 – 217, doi:http://dx.doi.org/10.1016/S0031-9201(02)  
551 00052-3, subduction Zone Structure and Megathrust Earthquakes.
- 552 Coutant, O. (1989), Programme de simulation numerique AXITRA, *Rapport LGIT*.
- 553 Duputel, Z., L. Rivera, H. Kanamori, and G. Hayes (2012), W phase source inversion for  
554 moderate to large earthquakes (1990–2010), *Geophysical Journal International*, 189(2),  
555 1125–1147, doi:10.1111/j.1365-246X.2012.05419.x.
- 556 Duputel, Z., P. S. Agram, M. Simons, S. E. Minson, and J. L. Beck (2014), Accounting for  
557 prediction uncertainty when inferring subsurface fault slip, *Geophys. J. Int.*, 197(1), 464–  
558 482.
- 559 Efron, B., and R. J. Tibshirani (1993), *An introduction to the Bootstrap*, Chapman &  
560 HALL/CRC.
- 561 Gao, H., D. A. Schmidt, and R. J. Weldon (2012), Scaling relationships of source parameters  
562 for slow slip events, *Bulletin of the Seismological Society of America*, 102(1), 352–360.
- 563 Gombert, B., Z. Duputel, E. Shabani, L. Rivera, R. Jolivet, and J. Hollingsworth (2019), Im-  
564 pulsive source of the 2017 mw= 7.3 ezgeleh, iran, earthquake, *Geophysical research let-  
565 ters*, 46(10), 5207–5216.
- 566 Hayes, G. P., G. L. Moore, D. E. Portner, M. Hearne, H. Flamme, M. Furtney, and G. M.  
567 Smoczyk (2018), Slab2, a comprehensive subduction zone geometry model, *Science*,  
568 362(6410), 58–61.
- 569 Herrera, C., S. Ruiz, R. Madariaga, and P. Poli (2017), Dynamic inversion of the 2015 Jujuy  
570 earthquake and similarity with other intraslab events, *Geophysical Journal International*,  
571 209(2), 866–875.
- 572 Herring, T., R. King, and S. C. McClusky (2010a), *GAMIT : GPS Analysis at MIT, release*  
573 *10.4*.
- 574 Herring, T., R. King, and S. C. McClusky (2010b), *GLOBK : Global Kalman filter VLBI and*  
575 *GPS analysis program release 10.4*.
- 576 Holtkamp, S. G., M. E. Pritchard, and R. B. Lohman (2011), Earthquake swarms in South  
577 America, *Geophysical Journal International*, 187(1), 128–146, doi:10.1111/j.1365-246X.  
578 2011.05137.x.

- 579 Ida, Y. (1972), Cohesive force across the tip of a longitudinal-shear crack and Griffith's spe-  
580 cific surface energy, *Journal of Geophysical Research*, 77(20), 3796–3805.
- 581 Kanamori, H., and D. L. Anderson (1975), Theoretical basis of some empirical relations in  
582 seismology, *Bulletin of the seismological society of America*, 65(5), 1073–1095.
- 583 Kanamori, H., L. Rivera, L. Ye, T. Lay, S. Murotani, and K. Tsumura (2019), New con-  
584 straints on the 1922 atacama, chile, earthquake from historical seismograms, *Geophysical*  
585 *Journal International*, 219(1), 645–661.
- 586 Klein, E., M. Métois, and V. C. D. A. Meneses, G. (2018a), Bridging the gap between North  
587 and Central Chile : insight from new GPS data on coupling complexities and the Andean  
588 sliver motion, *Geophysical Journal International*, 213(3), 1924 – 1933.
- 589 Klein, E., Z. Duputel, D. Zigone, C. Vigny, J.-P. Boy, C. Doubre, and G. Meneses (2018b),  
590 Deep transient slow slip detected by survey gps in the region of atacama, chile, *Geophys-*  
591 *ical research letters*, 45(22), 12–263.
- 592 Langer, L., T. Ragon, A. Sladen, and J. Tromp (2020), Impact of topography on earthquake  
593 static slip estimates, *Tectonophysics*, 791, 228,566.
- 594 Leyton, F., A. Leopold, G. Hurtado, C. Pastén, S. Ruiz, G. Montalva, and E. Saez (2018),  
595 Geophysical characterization of the chilean seismological stations: First results, *Seismo-*  
596 *logical Research Letters*, 89(2A), 519–525.
- 597 Lomnitz, C. (2004), Major earthquakes of Chile: A historical survey, 1535-1960, *Seismolog-*  
598 *ical Research Letters*, 75, 368–378, doi:10.1785/gssrl.75.3.368.
- 599 Meade, B. J. (2007), Algorithms for the calculation of exact displacements, strains, and  
600 stresses for triangular dislocation elements in a uniform elastic half space, *Computers &*  
601 *Geosciences*, 33(8), 1064–1075.
- 602 Métois, M., A. Socquet, C. Vigny, D. Carrizo, S. Peyrat, A. Delorme, E. Maureira, M.-C.  
603 Valderas-Bermejo, and I. Ortega (2013), Revisiting the north Chile seismic gap segmenta-  
604 tion using GPS-derived interseismic coupling, *Geophysical Journal International*, 194(3),  
605 1283–1294, doi:10.1093/gji/ggt183.
- 606 Métois, M., C. Vigny, and A. Socquet (2016), Interseismic coupling, megathrust earthquakes  
607 and seismic swarms along the Chilean subduction zone (38–18° S), *Pure and Applied*  
608 *Geophysics*, 173(5), 1431–1449.
- 609 Omori, F. (1894), On the Aftershocks of Earthquakes, *Journal of the College of Science, Im-*  
610 *perial University of Tokyo*, 7, 111–200.

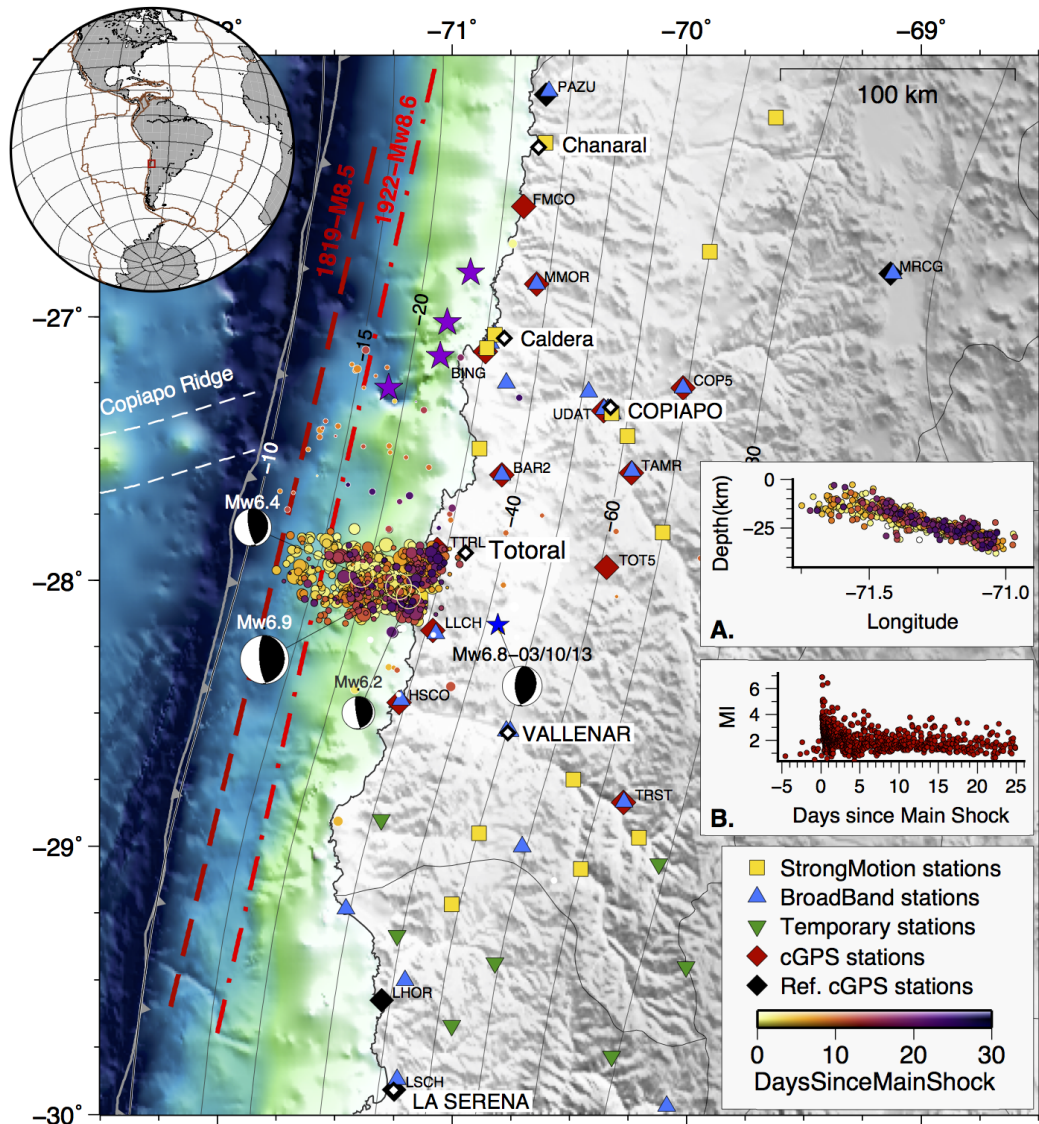
- 611 Otarola, C., S. Ruiz, C. Herrera, R. Madariaga, and C. Siegel (2021), Dynamic rupture of  
612 subduction earthquakes located near the trench, *Earth and Planetary Science Letters*, 562,  
613 116,842.
- 614 Pacheco, J. F., and L. R. Sykes (1992), Seismic moment catalog of large shallow earth-  
615 quakes, 1900 to 1989, *Bulletin of the Seismological Society of America*, 82(3), 1306–1349.
- 616 Pastén-Araya, F., P. Salazar, S. Ruiz, E. Rivera, B. Potin, A. Maksymowicz, E. Torres, J. Vil-  
617 larroel, E. Cruz, J. Valenzuela, D. Jaldín, G. González, W. Bloch, P. Wigger, and S. A.  
618 Shapiro (2018), Fluids Along the Plate Interface Influencing the Frictional Regime of the  
619 Chilean Subduction Zone, Northern Chile, *Geophysical Research Letters*, 45(19), 10,378–  
620 10,388, doi:10.1029/2018GL079283.
- 621 Pastén-Araya, F., B. Potin, S. Ruiz, L. Zerbst, F. Aden-Antoniów, K. Azúa, E. Rivera, A. Ri-  
622 etbrock, P. Salazar, and A. Fuenzalida (2021), Seismicity in the upper plate of the North-  
623 ern Chilean offshore forearc: Evidence of splay fault South of the Mejillones peninsula.,  
624 *Tectonophysics*, accepted.
- 625 Piñón, D. A., D. D. Gómez, R. Smalley Jr, S. R. Cimbaro, E. A. Lauría, and M. G. Bevis  
626 (2018), The history, state, and future of the Argentine continuous satellite monitoring net-  
627 work and its contributions to geodesy in Latin America, *Seismological Research Letters*,  
628 89(2A), 475–482.
- 629 Potin, B., S. Barrientos, B. Valette, and S. Ruiz (2019), Tomography of Chile, in *8th Interna-  
630 tional Symposium on Andean Geodynamics (ISAG)*.
- 631 Radiguet, M., H. Perfettini, N. Cotte, A. Gualandi, B. Valette, V. Kostoglodov, T. Lhomme,  
632 A. Walpersdorf, E. C. Cano, and M. Campillo (2016), Triggering of the 2014 Mw 7.3 Pa-  
633 panao earthquake by a slow slip event in Guerrero, Mexico, *Nature Geoscience*, 9(11),  
634 829.
- 635 Reasenberg, P. A., and L. M. Jones (1989), Earthquake hazard after a mainshock in califor-  
636 nia, *Science*, 243(4895), 1173–1176.
- 637 Rolandone, F., J.-M. Nocquet, P. A. Mothes, P. Jarrin, M. Vallée, N. Cubas, S. Hernandez,  
638 M. Plain, S. Vaca, and Y. Font (2018), Areas prone to slow slip events impede earthquake  
639 rupture propagation and promote afterslip, *Science Advances*, 4(1), eaao6596.
- 640 Ruiz, S., and R. Madariaga (2011), Determination of the friction law parameters of the mw  
641 6.7 michilla earthquake in northern chile by dynamic inversion, *Geophysical Research  
642 Letters*, 38(9).

- 643 Ruiz, S., and R. Madariaga (2018), Historical and recent large megathrust earthquakes in  
644 Chile, *Tectonophysics*, pp. –, doi:<https://doi.org/10.1016/j.tecto.2018.01.015>.
- 645 Ruiz, S., F. Aden-Antoniow, J. Baez, C. Otarola, B. Potin, F. Campo, P. Poli, C. Flores,  
646 C. Satriano, F. Leyton, et al. (2017), Nucleation phase and dynamic inversion of the Mw  
647 6.9 Valparaíso 2017 earthquake in Central Chile, *Geophysical Research Letters*, *44*(20).
- 648 Sambridge, M. (1999), Geophysical inversion with a neighbourhood algorithm—I. Searching  
649 a parameter space, *Geophysical journal international*, *138*(2), 479–494.
- 650 Sippl, C., B. Shurr, G. Asch, and J. Kummerow (2018), Seismicity structure of the northern  
651 Chile forearc from >100,000 double-difference relocated hypocenters, *Journal of Geo-*  
652 *physical Research: Solid Earth*, *123*(5), 4063–4087.
- 653 Soloviev, S., and C. Go (1976), Catalog of tsunamis on the western coast of the Pacific  
654 ocean, nauka, urss (in Russian).
- 655 Suito, H., T. Nishimura, M. Tobita, T. Imakiire, and S. Ozawa (2011), Interplate fault slip  
656 along the Japan trench before the occurrence of the 2011 off the Pacific coast of Tohoku  
657 earthquake as inferred from GPS data, *Earth, planets and space*, *63*(7), 19.
- 658 Utsu, T. (1957), Magnitude of Earthquakes and Occurrence of their Aftershocks, *Zisin (J.*  
659 *Seismol. Soc. Jap.)*, *10*, 35–45, doi:10.4294/jpe1952.43.1.
- 660 Vaca, S., M. Vallée, J.-M. Nocquet, J. Battaglia, and M. Régnier (2018), Recurrent slow slip  
661 events as a barrier to the northward rupture propagation of the 2016 Pedernales earthquake  
662 (central Ecuador), *Tectonophysics*, *724*, 80–92.
- 663 Villegas Lanza, J., J. Nocquet, F. Rolandone, M. Vallee, H. Tavera, F. Bondoux, T. Tran,  
664 X. Martin, and M. Chlieh (2015), A slow slip and seismic swarm sequence in a weakly  
665 coupled subduction zone in northern Peru, *AGUFM, 2015*, S31A–2735.
- 666 Wessel, P., W. H. F. Smith, R. Scharroo, J. Luis, and F. Wobbe (2013), Generic Mapping  
667 Tools: Improved version released, *Eos, Transactions American Geophysical Union*,  
668 *94*(45), 409–410, doi:10.1002/2013EO450001.
- 669 Willis, B. (1929), *Studies in Comparative Seismology: Earthquake Conditions in Chile*, 382,  
670 Carnegie Institution of Washington.
- 671 Wynants-Morel, N., F. Cappa, L. De Barros, and J.-P. Ampuero (2020), Stress perturbation  
672 from aseismic slip drives the seismic front during fluid injection in a permeable fault,  
673 *Journal of Geophysical Research: Solid Earth*, *125*(7), e2019JB019179.
- 674 Yagi, Y., M. Kikuchi, and T. Sagiya (2001), Co-seismic slip, post-seismic slip, and after-  
675 shocks associated with two large earthquakes in 1996 in Hyuga-Nada, Japan, *Earth, planets*

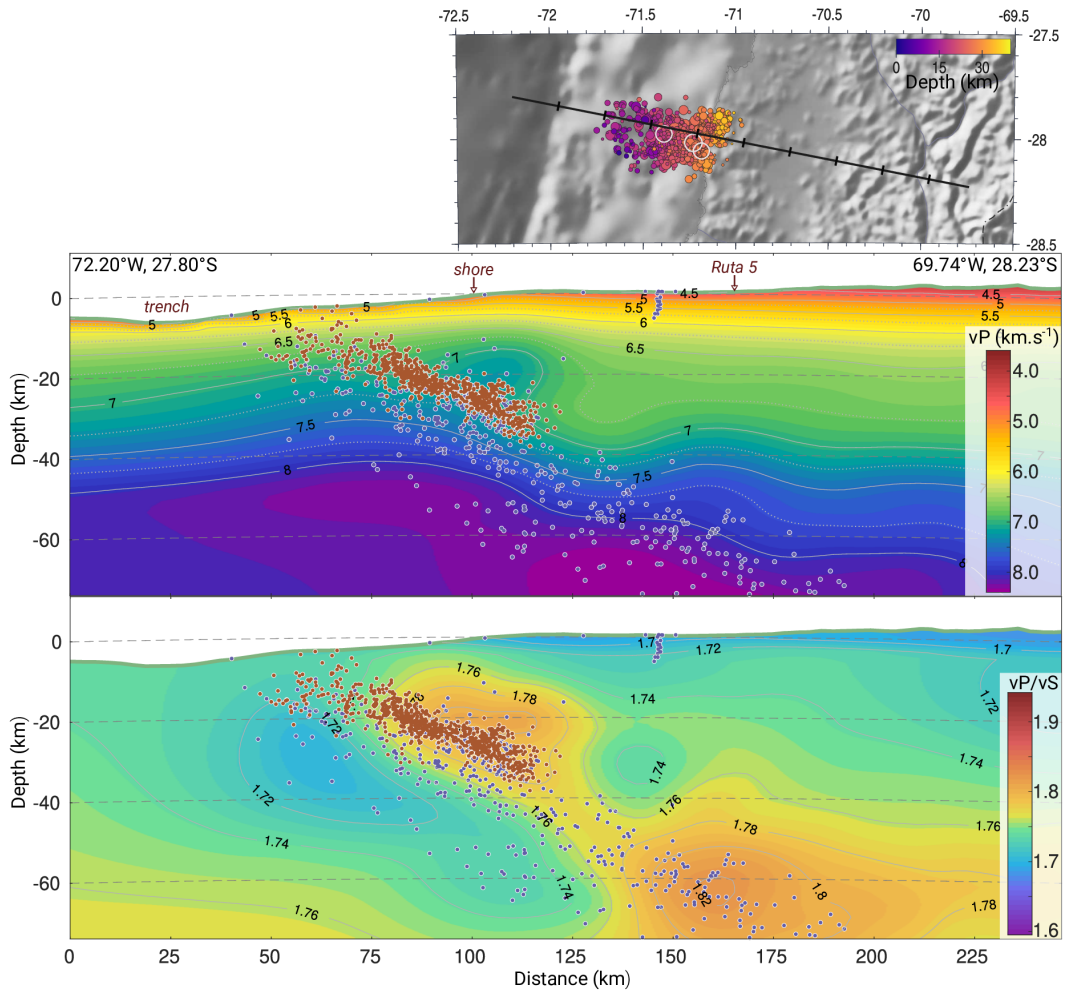
676 *and space*, 53(8), 793–803.

677 Yingdi, L., and J.-P. Ampuero (2017), Preprint: Tremor migration patterns and the collective  
678 behavior of deep asperities mediated by creep.

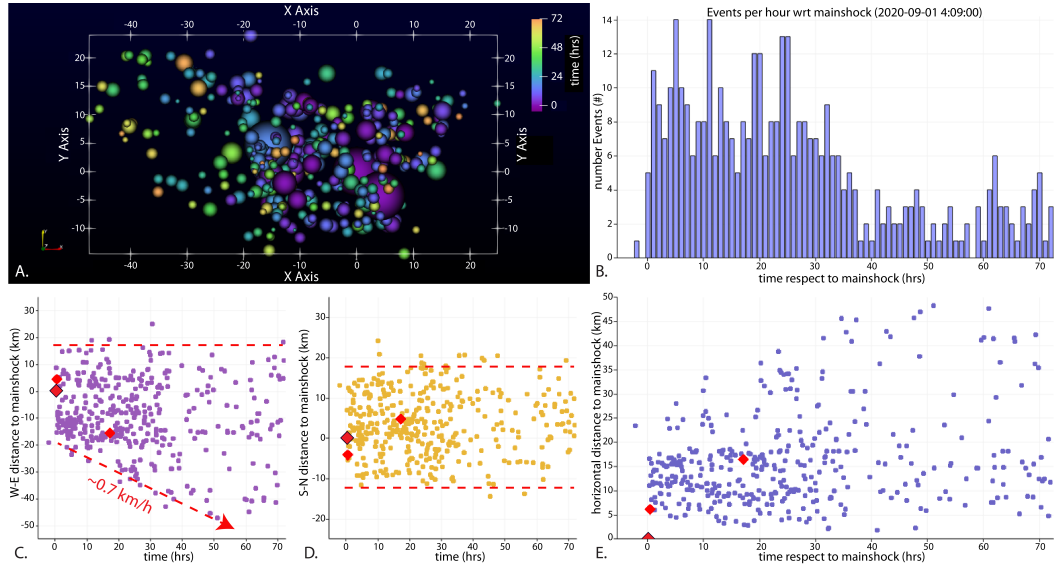




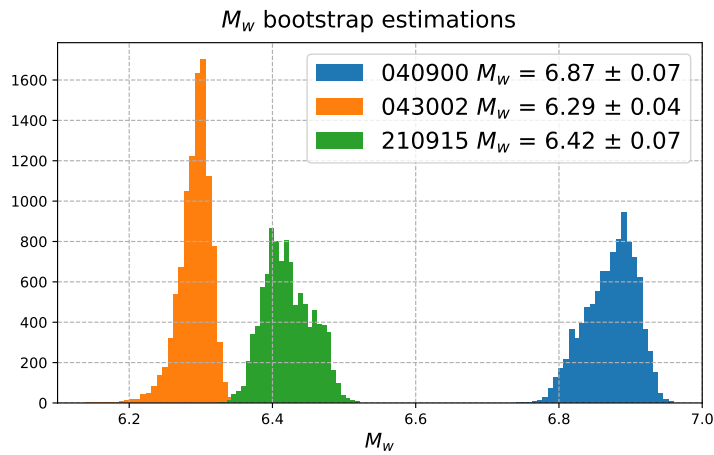
679 **Figure 1.** Overall context of the sequence of September 2020 in the Atacama region of Chile. The re-  
 680 located earthquakes catalog is plotted as a function of time since the mainshock (in days since the mainshock).  
 681 Events represented with white contours were relocated outside of the core sequence. Mechanisms and  $M_w$   
 682 of the 3 largest events are the re-estimated one. The different observation networks used in this study are re-  
 683 presented. A. Cross section of the relocated catalog of the core sequence, as function of depth, with the same  
 684 color scale function of time. B. Local magnitudes  $M_l$  of the relocated catalog of the core sequence as function  
 685 of time. Violet stars show swarms locations [Holtkamp *et al.*, 2011]. Slab isodepth from Hayes *et al.* [2018].  
 686 The dashed red lines illustrates the approximate length 1819 and 1922 earthquakes rupture zones.



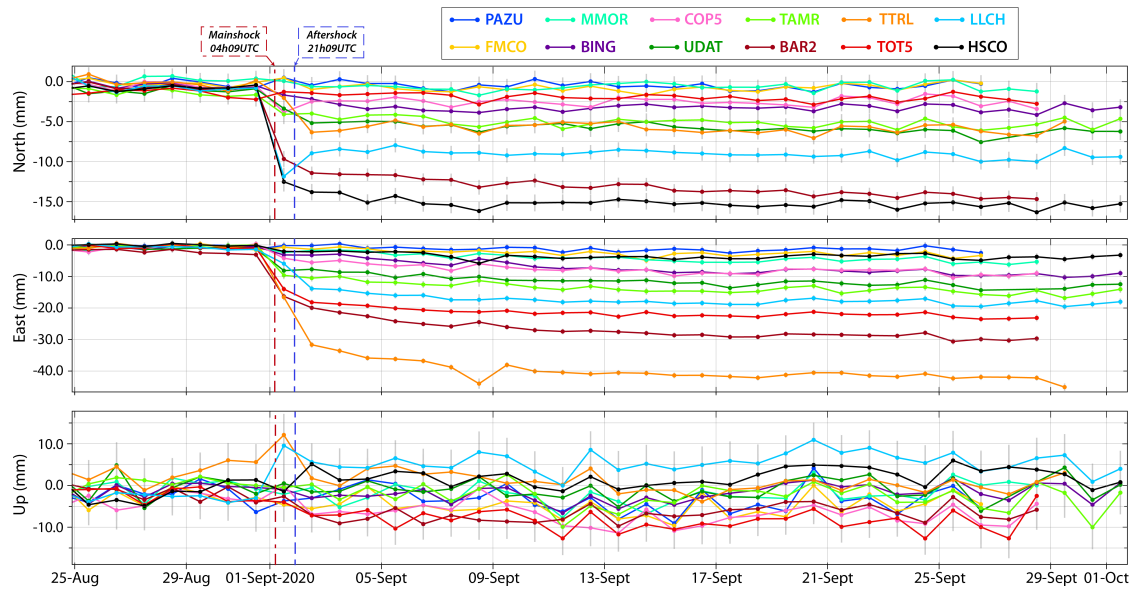
687 **Figure 2.** Vertical cross-section of tomography model for both P-wave velocity and  $V_P/V_S$  velocity ratio.  
 688 Cross-section oriented perpendicular to the trench, across the sequence. Blue dots represent background  
 689 seismicity in a 50 km range from the cross-section (CSN catalog, 2013 – 07/2020). Red dots represent the  
 690 seismic sequence between August 25<sup>th</sup> and September 25<sup>th</sup> included. All earthquakes were relocated in the  
 691 local 3D tomography model presented. Seismicity of the sequence spreads along subduction contact between  
 692 the trench and about 40 km at depth.



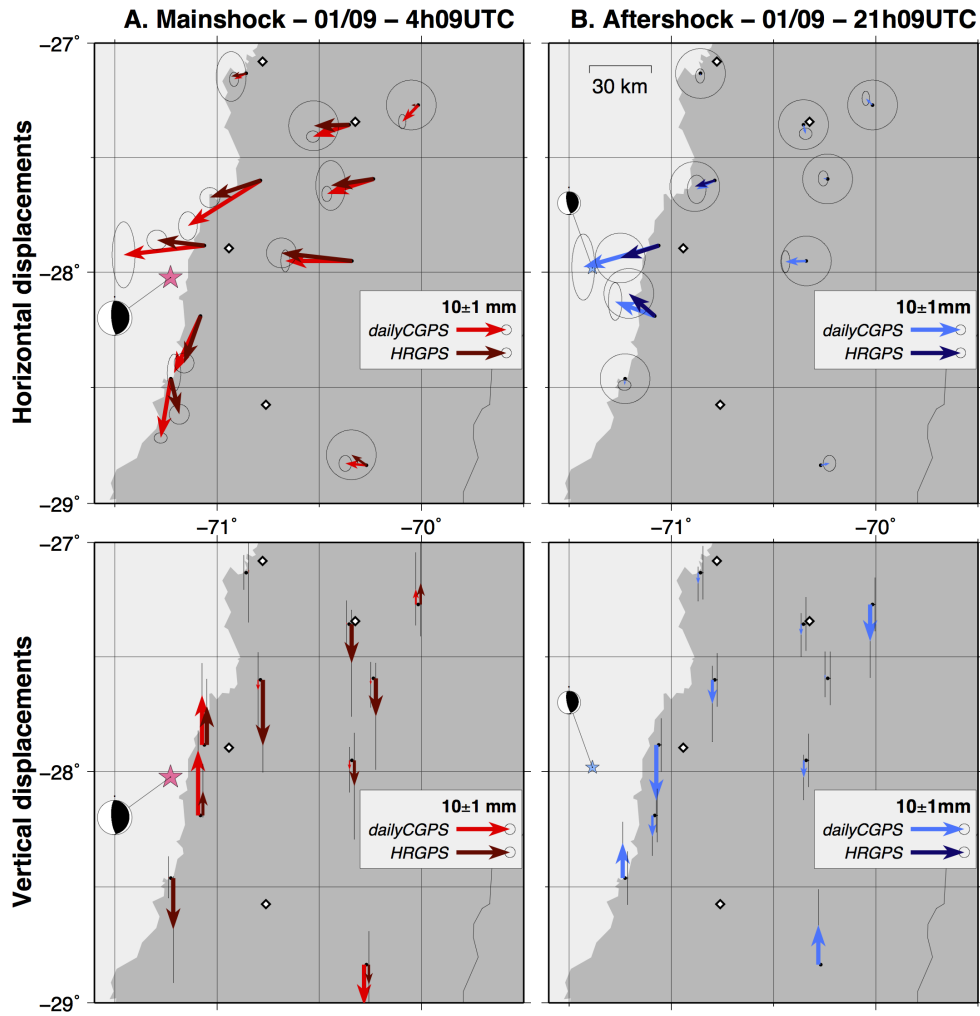
693 **Figure 3.** Analysis of the relocated catalog over the first 72 h after the mainshock; A. map view, distances  
 694 are in km, coordinate (0,0) correspond to the mainshock and colours represent time; B. number of events/h,  
 695 time is relative to the mainshock origin time. Bins are centered on the hour; C. distance to mainshock in the  
 696 West-East direction vs. time, in km; D. distance to mainshock in the South-North direction vs. time, in km; E.  
 697 horizontal distance to main shock in km. The 3 main events are highlighted by the red diamonds on subplots  
 698 C, D E, the mainshock is black contoured. The red dashed lines on subplots C and D depict the seismicity  
 699 boundaries.



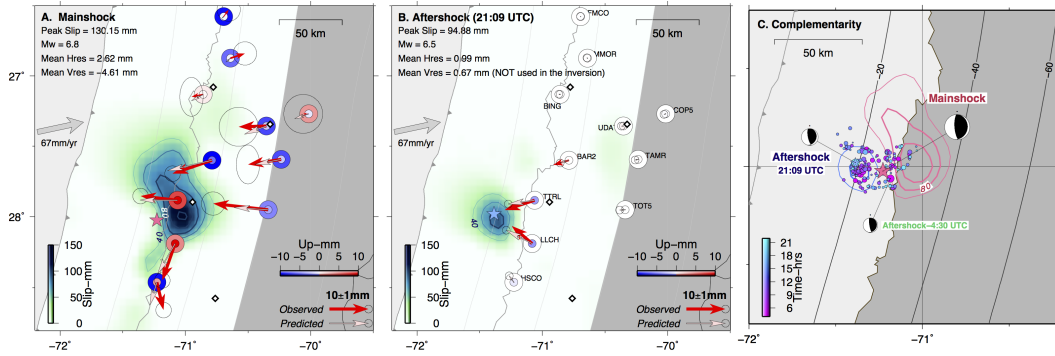
700 **Figure 4.** Histograms of the bootstrap analysis for the moment magnitude of the three largest events with  
 701  $10^5$  inversions. Average  $M_w$  and  $\pm 2\sigma$  uncertainties are given in the legend.



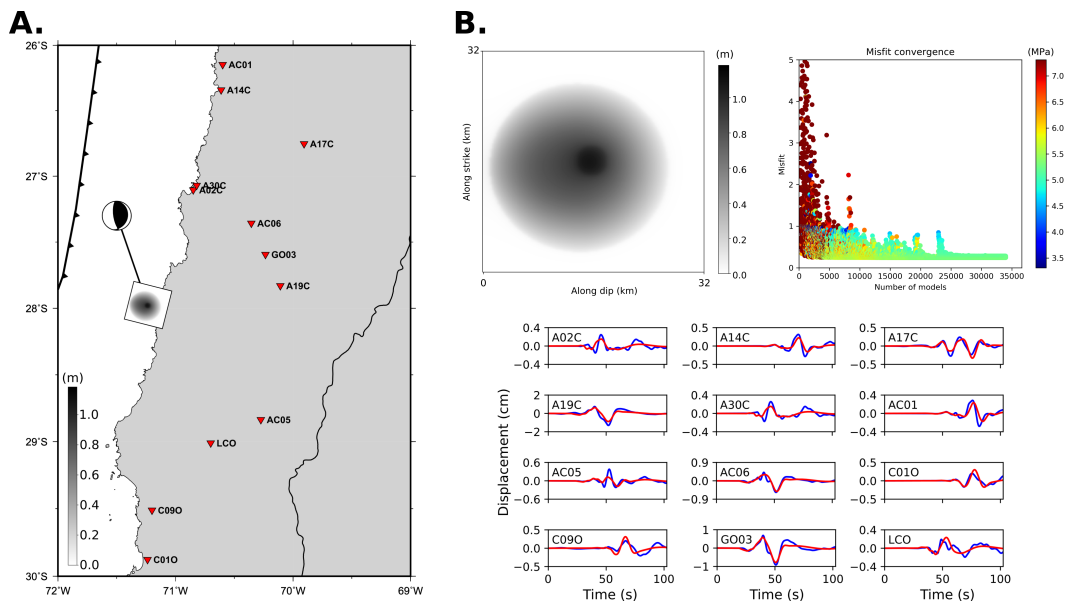
702 **Figure 5.** Time series of GPS daily positions from stations in the region of the sequence on the 3 compo-  
 703 nents. The vertical black lines flag the exact time of the 2 events of September 1<sup>st</sup>, at the beginning of the  
 704 sequence.



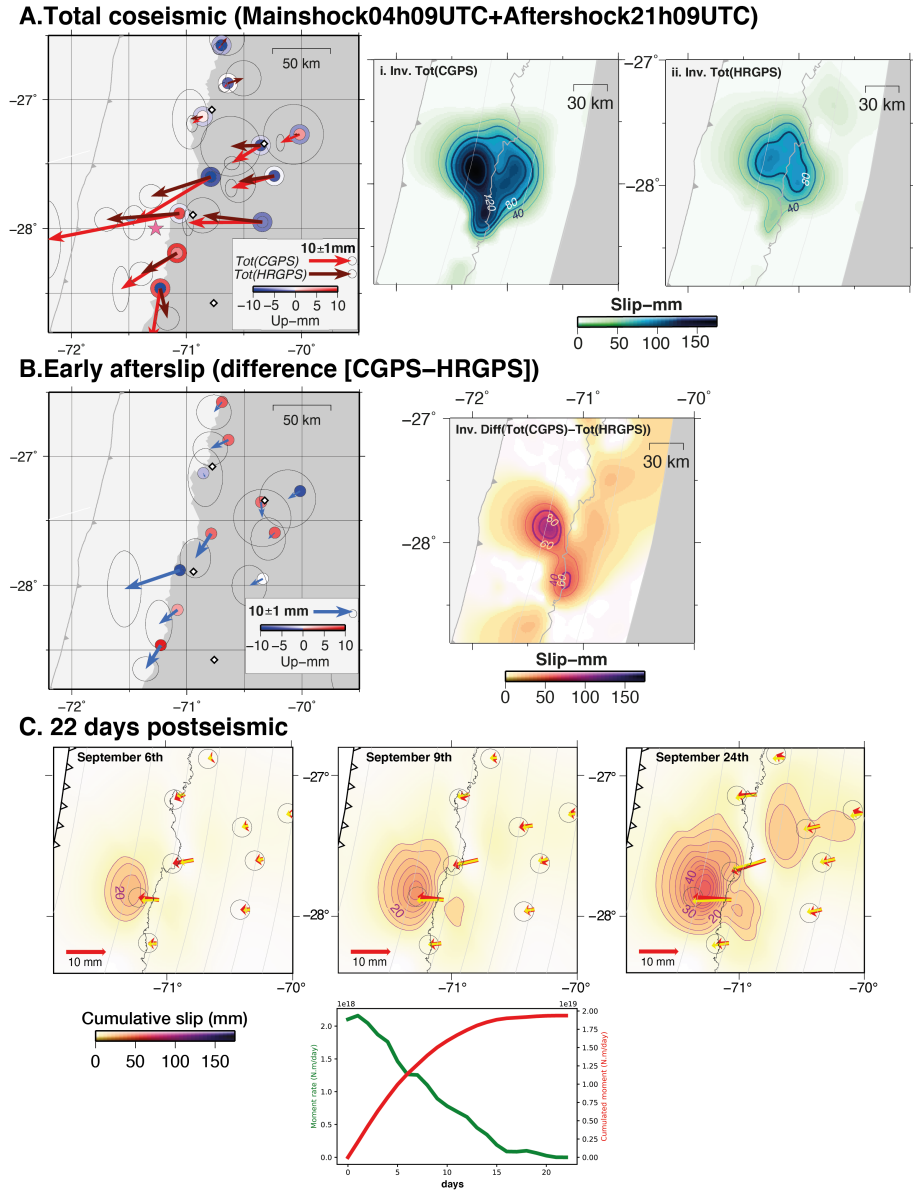
705 **Figure 6.** Comparison of 24 hours CGPS and HRGPS static co-seismic offset estimations; A. mainshock of  
 706 04:09 (reddish vectors) and B. aftershock of 21:09 (blueish vectors). Horizontal top row, vertical bottom row.  
 707 Earthquakes' locations from the relocated catalog and mechanisms from the W-phase analysis.



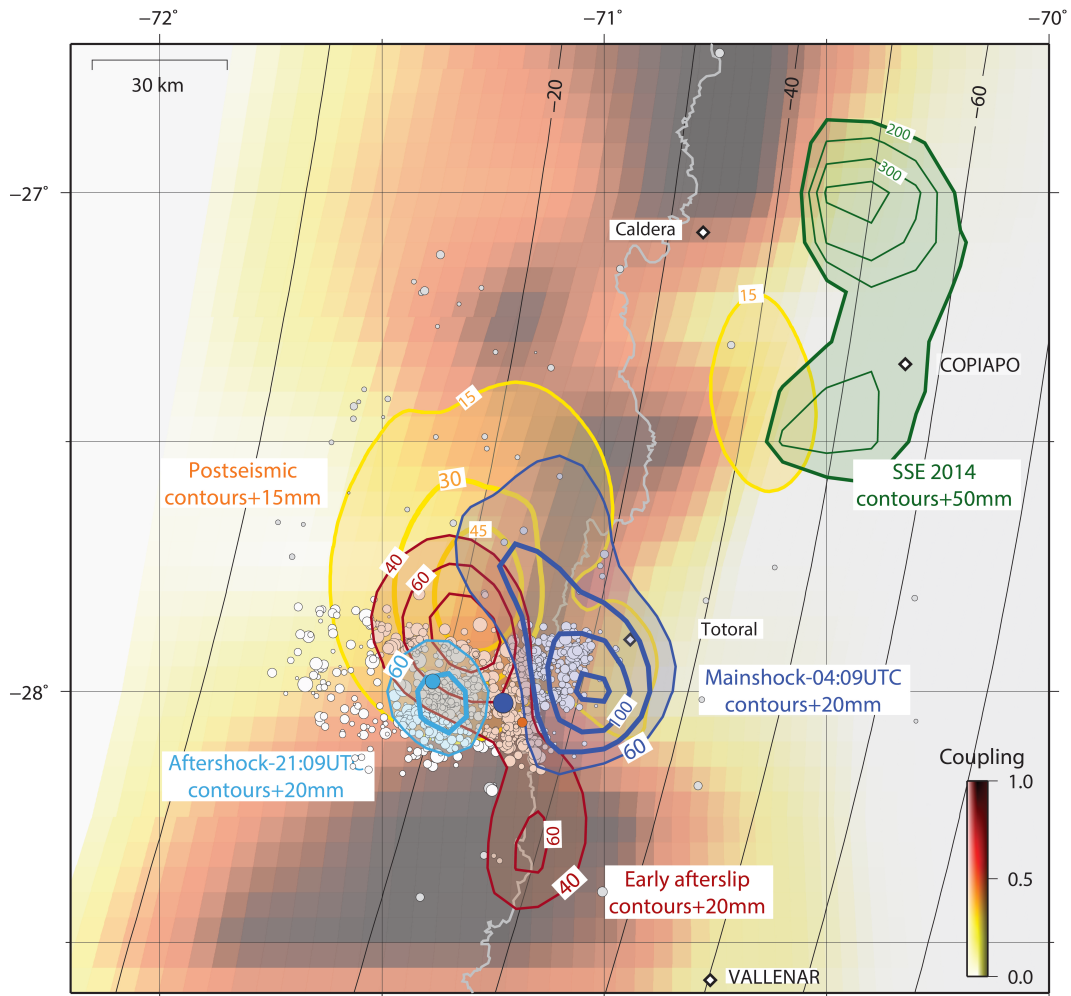
708 **Figure 7.** Slip distributions of A. the mainshock at 4:09 UTC inverted from the HRGPS (Fig. 6-A); B. the  
 709 aftershock at 21:09 UTC inverted from the HRGPS (Fig. 6-B); Distributions are represented as the blue color  
 710 scale (in mm), blue isolines are represented every 20 mm; Horizontal coseismic displacements are depicted  
 711 by arrows: Observations (red) vs predictions (pink); Vertical coseismic displacements are depicted by col-  
 712 ored dots : Observations (big circles) vs predictions (small circles) with amplitude represented with the polar  
 713 color scale; C. Zoom in to compare both slip distributions and the relocated catalog of aftershocks occurring  
 714 between the 2 events represented with the color scale. Isodepth from Slab2.0 [Hayes *et al.*, 2018]



715 **Figure 8.** Coseismic model of the mainshock obtained from the dynamic inversion. A. Geographic context  
 716 of the mainshock rupture and stations used for modeling. The moment tensor was obtained from GCMT. B.  
 717 Dynamic slip model on the fault plane and waveform misfit convergence colored with the stress drop. The  
 718 bottom plot shows the E-W observed (blue) and modeled (red) waveforms of the best dynamic model.



719 **Figure 9.** Slip history over the sequence: A. Total coseismic: vectors show the total coseismic displace-  
 720 ment on September, 1<sup>st</sup> (including both events and the aseismic slip that occurred during that period) mea-  
 721 sured by CGPS (light red) and the corresponding Slip distribution (i), compared with the total coseismic  
 722 displacement due to the 2 events measured by HRGPS (dark red) and the corresponding slip distribution (ii);  
 723 B. Early afterslip estimated from the difference between CGPS and HRGPS estimates and the corresponding  
 724 slip distribution; C. Slip-time dependent inversion of the postseismic deformation 22 days with 3 snapshots  
 725 of the cumulative slip distribution. Yellow and red arrows are respectively model-predicted and observed  
 726 displacements for CGPS sites recorded since the mainshock. Postseismic slip contours are every 10 mm. Gray  
 727 lines are Slab2.0 isodepth from *Hayes et al.* [2018].



728 **Figure 10.** Slip distributions of the  $M_w = 6.9$  mainshock (01/09/2020-4:09 UTC, dark blue contours every  
 729 20 mm starting at 60mm), the  $M_w = 6.4$  aftershock (01/09/2020-21:09 UTC, light blue contours every 20 mm  
 730 starting at 60 mm); the rapid afterslip between the 2 events (red contours every 20 mm starting at 40 mm), and  
 731 1 month of postseismic slip (yellow contours every 15 mm starting at 15 mm). The epicenter of the  $M_w = 6.2$   
 732 aftershock (4:30 UTC) is depicted by the orange dot. Comparison with the coupling distribution in the re-  
 733 gion [Klein *et al.*, 2018a] and the 2014 SSE distribution [Klein *et al.*, 2018b, ], represented by the dark green  
 734 contours every 50 mm starting at 200 mm]). Background seismicity from relocated catalog depicted by white  
 735 dots. Slab2.0 isodepth from Hayes *et al.* [2018] every 10 km.