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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 ~ 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, 1st, 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 25th of
101 August and the 25th 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 25th and
104 the 31st 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 km² 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° ; 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° 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 ± 0.07 , 6.29 ± 0.04 , 6.42 ± 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 1st). 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°S and 28.5°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°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.m²). 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_μ), 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 2nd and the 24th 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°S), and the Chañaral segment (North
 398 of 27°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 $\sim 100 \times 100$ km², 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
 477 closely in the next future.

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496 [com/JMNocquet/pyacs36](http://www.geologie.ens.fr/~joliviet/csi), CSI (<http://www.geologie.ens.fr/~joliviet/csi>) and
497 Pyrocko (<https://pyrocko.org/>)

498 **Data availability:** Seismic data collected are available through the Incorporated Re-
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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
504 upon request.

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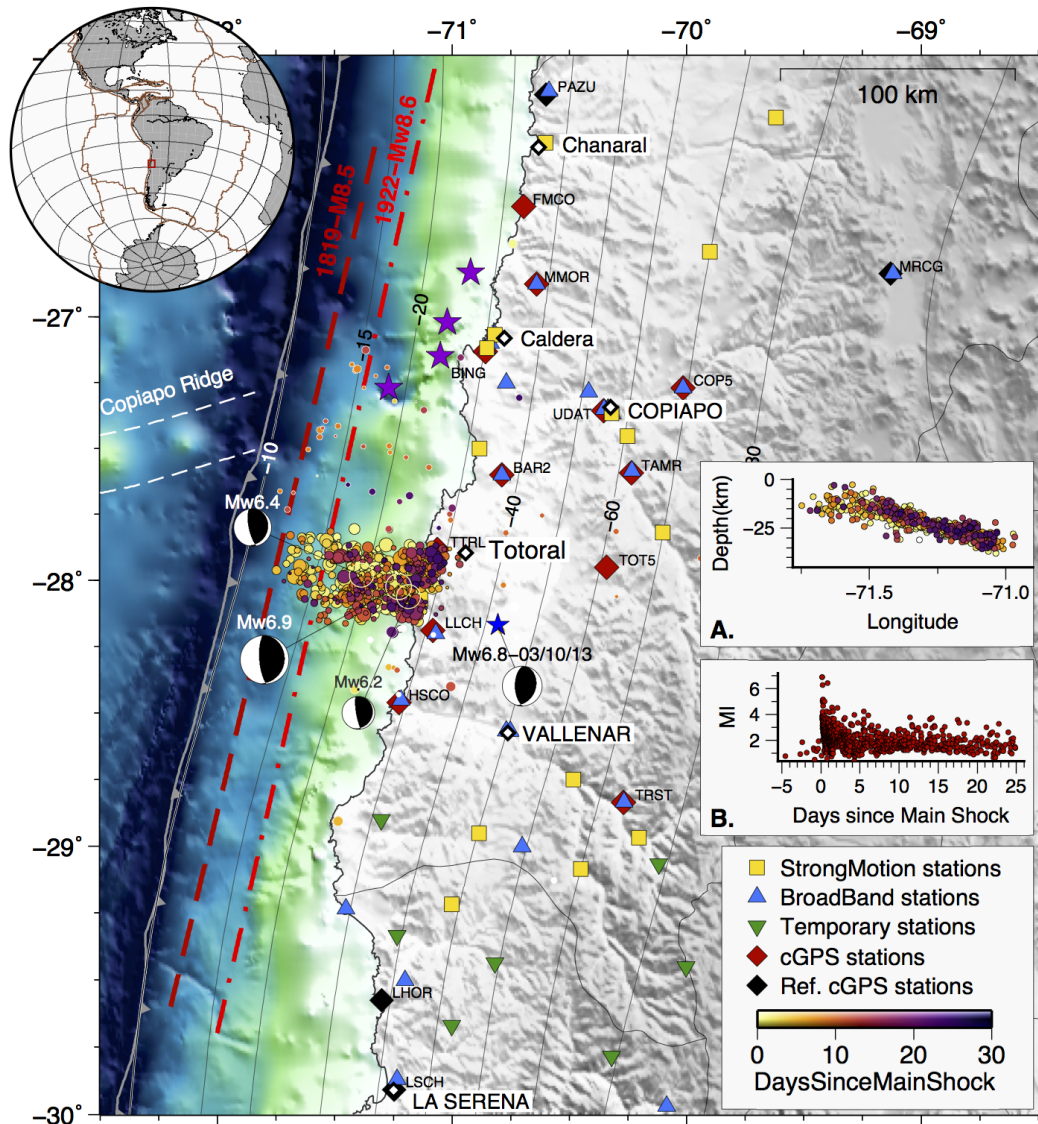
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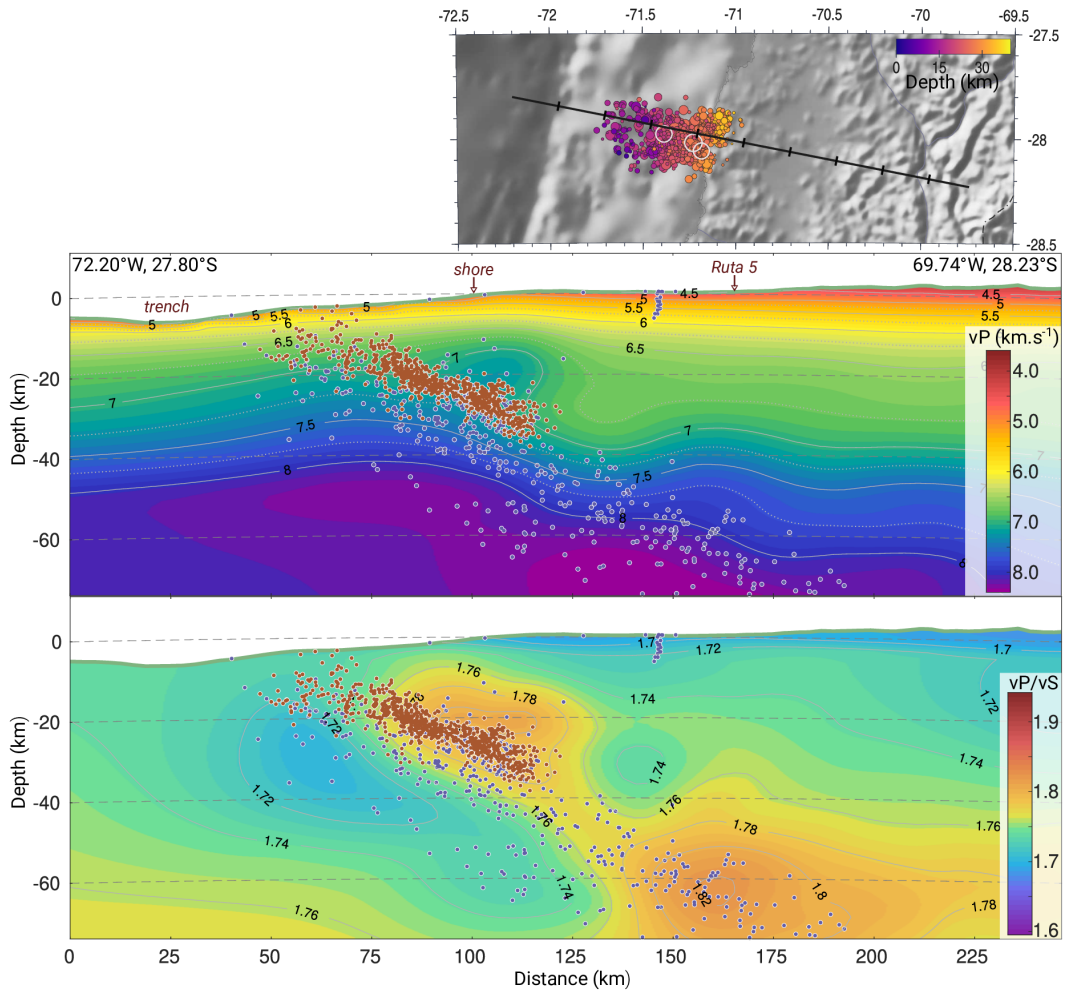
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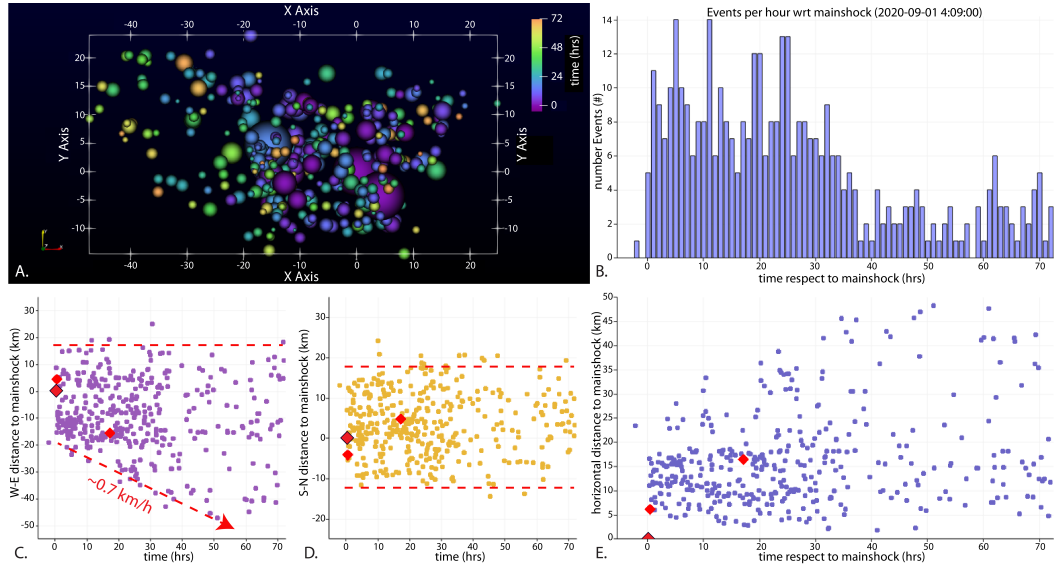
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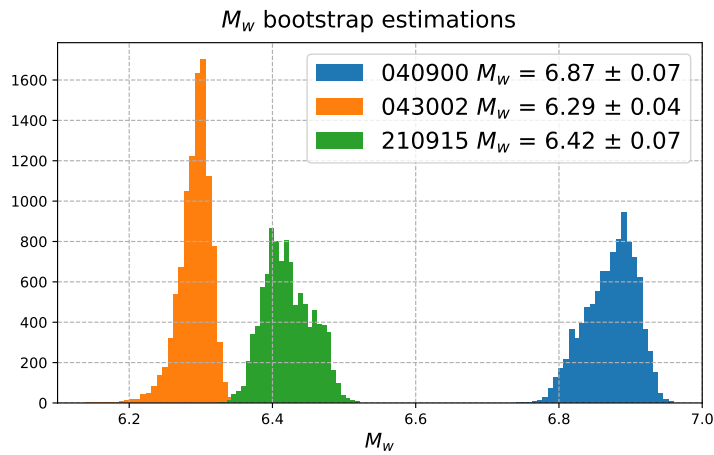
679 **Figure 1.** Overall context of the sequence of September 2020 in the Atacama region of Chile. The re-
 680 located earthquakes catalog is plotted as a function of time since the mainshock (in days since the mainshock).
 681 Events represented with white contours were relocated outside of the core 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.



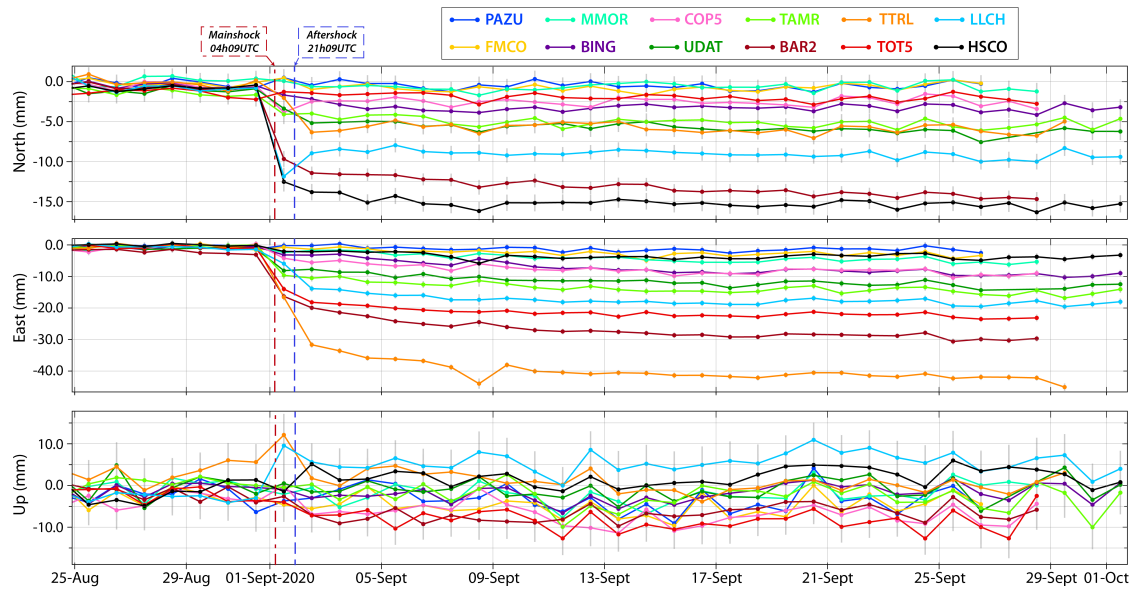
687 **Figure 2.** Vertical cross-section of tomography model for both P-wave velocity and V_P/V_S velocity ratio.
 688 Cross-section oriented perpendicular to the trench, across the sequence. Blue dots represent background
 689 seismicity in a 50 km range from the cross-section (CSN catalog, 2013 – 07/2020). Red dots represent the
 690 seismic sequence between August 25th and September 25th 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.



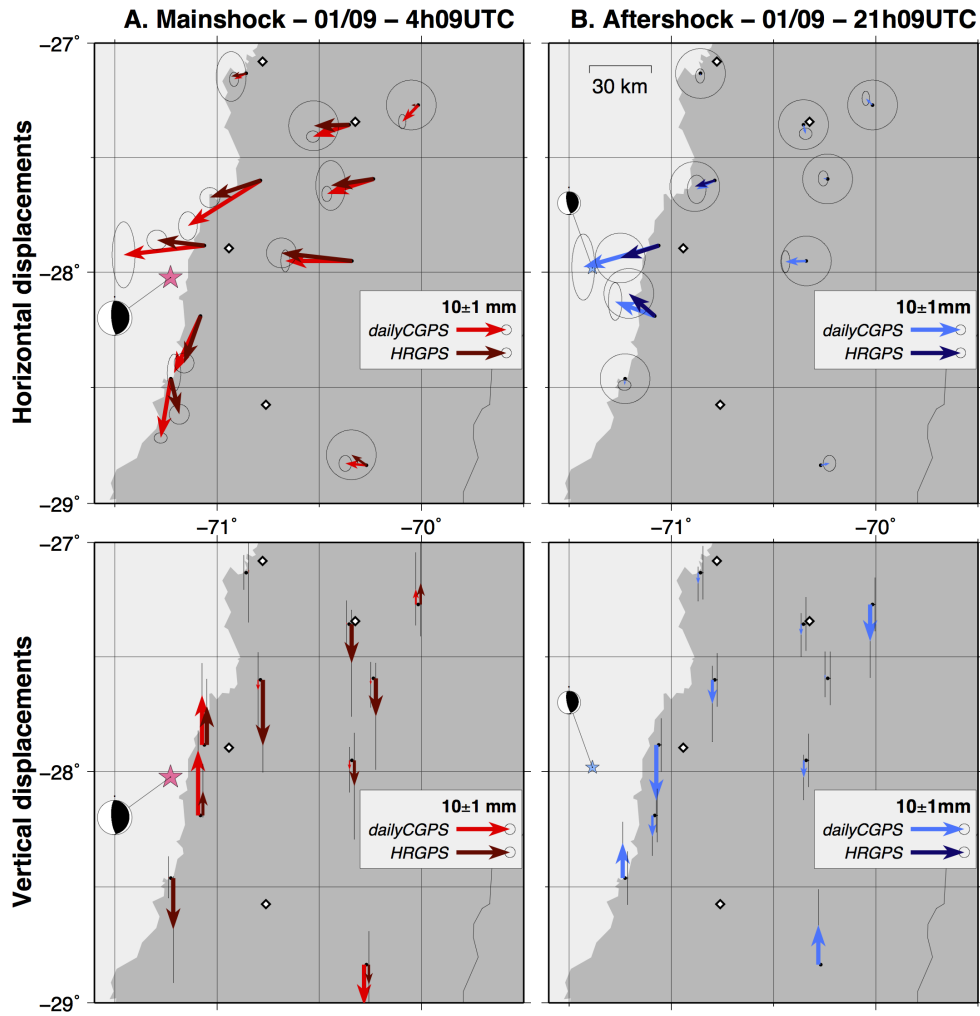
693 **Figure 3.** Analysis of the relocated catalog over the first 72 h after the mainshock; A. map view, distances
 694 are in km, coordinate (0,0) correspond to the mainshock and colours represent time; B. number of events/h,
 695 time is relative to the mainshock origin time. Bins are centered on 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.



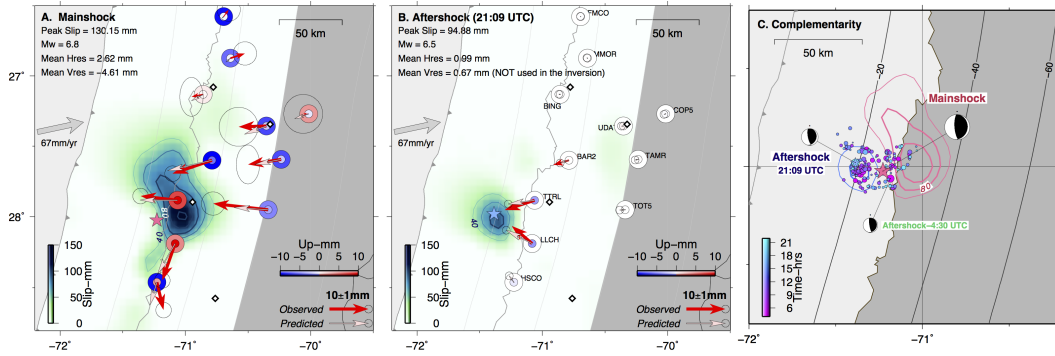
700 **Figure 4.** Histograms of the bootstrap analysis for the moment magnitude of the three largest events with
 701 10^5 inversions. Average M_w and $\pm 2\sigma$ uncertainties are given in the legend.



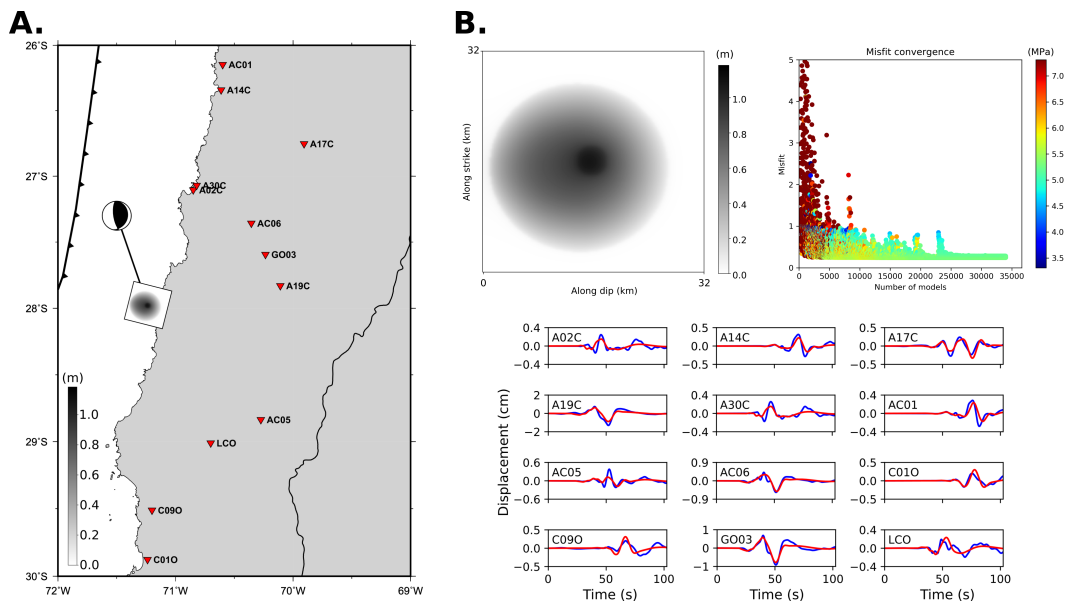
702 **Figure 5.** Time series of GPS daily positions from stations in the region of the sequence on the 3 compo-
 703 nents. The vertical black lines flag the exact time of the 2 events of September 1st, at the beginning of the
 704 sequence.



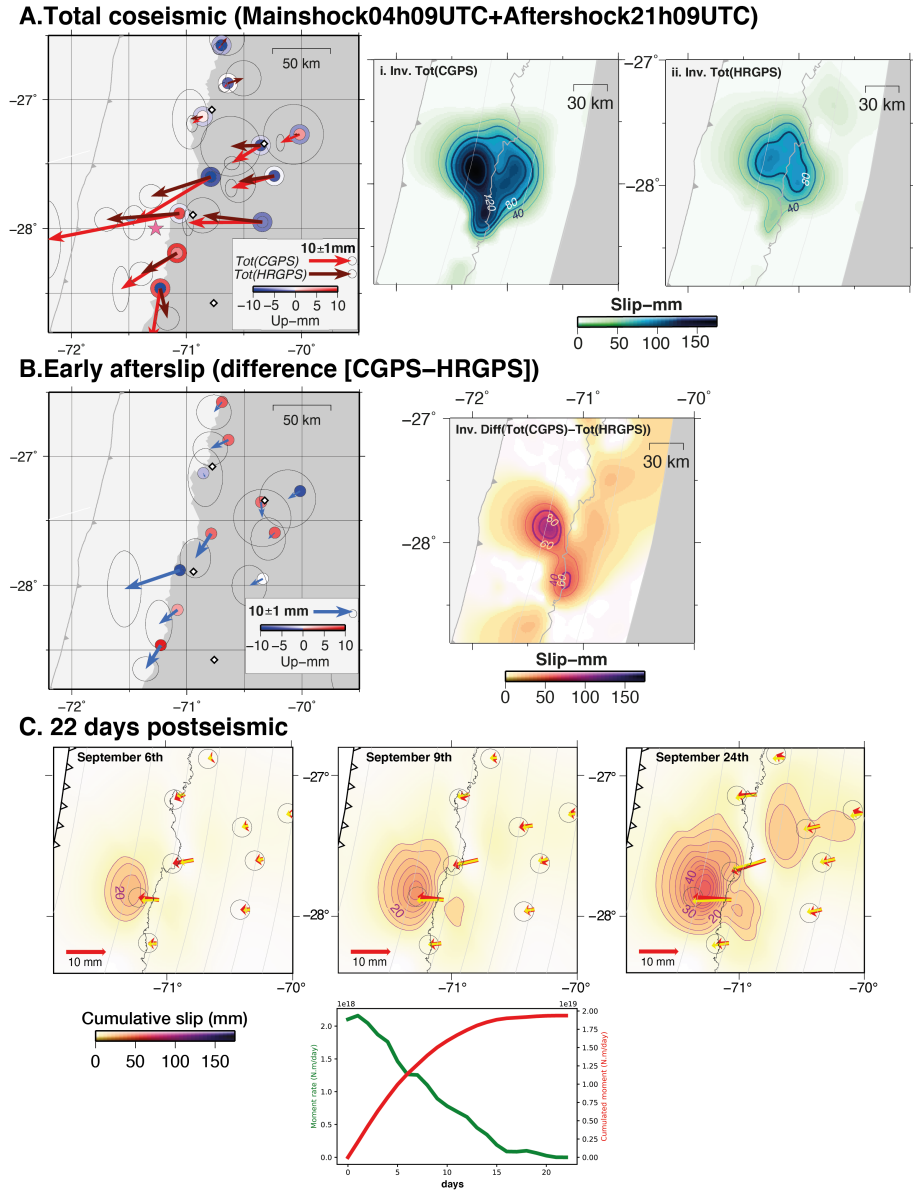
705 **Figure 6.** Comparison of 24 hours CGPS and HRGPS static co-seismic offset estimations; A. mainshock of
 706 04:09 (reddish vectors) and B. aftershock of 21:09 (blueish vectors). Horizontal top row, vertical bottom row.
 707 Earthquakes' locations from the relocated catalog and mechanisms from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



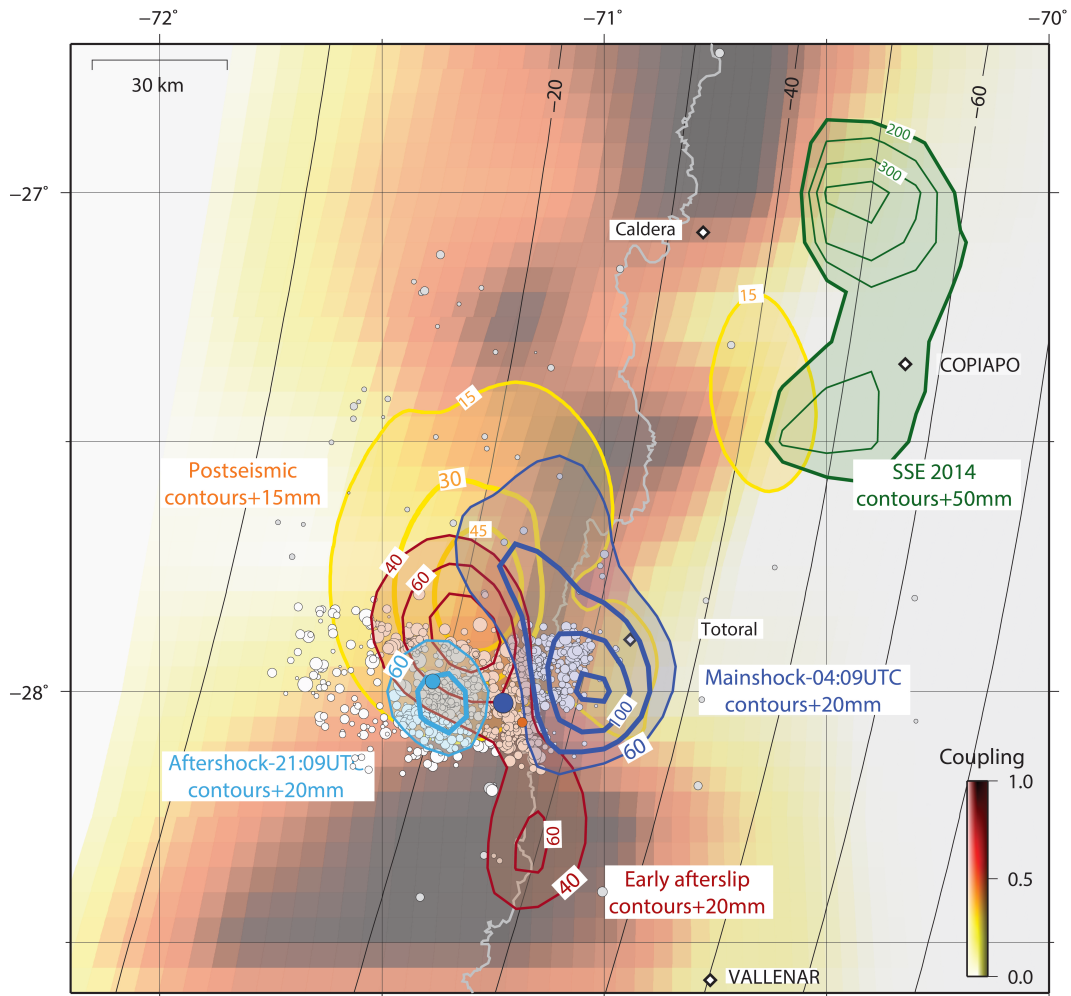
708 **Figure 7.** Slip distributions of A. the mainshock at 4:09 UTC inverted from the HRGPS (Fig. 6-A); B. the
 709 aftershock at 21:09 UTC inverted from the HRGPS (Fig. 6-B); Distributions are represented as the blue color
 710 scale (in mm), blue isolines are represented every 20 mm; Horizontal coseismic 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, 1st (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.