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1 **Deep postseismic viscoelastic relaxation excited by an intraslab normal fault**
2 **earthquake in the Chile subduction zone**

3

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9

10 **Abstract**

11

12 The 2005 M_w 7.8 Tarapaca earthquake was the result of normal faulting on a west-
13 dipping plane at a depth of ~ 90 km within the subducting slab down-dip of the North
14 Chilean gap that partially ruptured in the 2014 M 8.2 Iquique earthquake. We use
15 Envisat observations of nearly four years of postseismic deformation following the
16 earthquake, together with some survey GPS measurements, to investigate the
17 viscoelastic relaxation response of the surrounding upper mantle to the coseismic
18 stress. We constrain the rheological structure by testing various 3D models, taking
19 into account the vertical and lateral heterogeneities in viscosity that one would expect
20 in a subduction zone environment. A viscosity of $4 - 8 \times 10^{18}$ Pa s for the continental
21 mantle asthenosphere fits both InSAR line-of-sight (LOS) and GPS horizontal
22 displacements reasonably well. In order to test whether the Tarapaca earthquake and
23 associated postseismic relaxation could have triggered the 2014 Iquique sequence, we

24 computed the Coulomb stress change induced by the co- and postseismic deformation
25 following the Tarapaca earthquake on the megathrust interface and nodal planes of its
26 M 6.7 foreshock. These static stress calculations show that the Tarapaca earthquake
27 may have an indirect influence on the Iquique earthquake, via loading of the M 6.7
28 preshock positively. We demonstrate the feasibility of using deep intraslab
29 earthquakes to constrain subduction zone rheology. Continuing geodetic observation
30 following the 2014 Iquique earthquake may further validate the rheological
31 parameters obtained here.

32

33 **Keywords:** Tarapaca, subduction zone rheology, viscoelastic relaxation, normal
34 faulting earthquake, geodetic observation

35

36 **1. Introduction**

37 One of the key factors that limit our understanding of the physics governing
38 megathrust earthquake cycles is a lack of knowledge of subduction zone rheology.
39 Theoretically, stresses induced by megathrust earthquakes will be relaxed in
40 thermally weakened layers. Postseismic deformation produced by this viscoelastic
41 relaxation process (VER) may be modeled to constrain the rheology. Recent
42 advances in the spatial and temporal coverage of geodetic measurements have
43 allowed for transient deformation following several megathrust earthquakes being
44 explicitly investigated to infer rheological properties at various subduction zones (e.g.

45 Pollitz et al., 2008; Hu & Wang 2012; Sun et al., 2014; Trubienko et al., 2014; Klein
46 et al., 2016).

47

48 Intermediate depth earthquakes represent another type of event in subduction zones
49 that ruptures within the subducting slab at depths of 70 to 300 km. Large
50 intermediate-depth earthquakes also induce stresses, which will gradually be relaxed
51 by VER in the adjoining mantle, and produce transient deformation that could be
52 indicative of subduction zone rheology. To test the feasibility of using intermediate
53 depth earthquakes to constrain subduction zone rheology, we investigate InSAR and
54 GPS observations following the June 13, 2005 magnitude 7.8 Tarapaca earthquake,
55 which occurred ~ 100 km inland from the coast in northern Chile (Fig. 1).

56 Seismological and geodetic studies identified this event as a slab-pull normal faulting
57 earthquake on a shallowly west-dipping intraslab fault at ~ 90 km depth (Peyrat et al.,
58 2006; Delouis & Legrand, 2007).

59

60 In this region, the Nazca plate subducts beneath the South American plate at a rate of
61 ~ 7 cm/yr (Argus et al., 2011). Studies of interseismic deformation show that this
62 segment of the plate boundary is overall highly locked, with a local decrease of
63 coupling in front of Iquique (e.g. Chlieh et al., 2011; Béjar-Pizarro et al., 2013;
64 Métois et al., 2013). In April 2014, the magnitude 8.2 Iquique earthquake occurred ~
65 100 km offshore and partially released the strain accumulated on the shallow
66 interface since the last big earthquake in 1877 (e.g. Ruiz et al., 2014; Schurr et al.,
67 2014; Bürgmann 2014). Two weeks before the mainshock, a M6.7 foreshock

68 ruptured at shallow depth of ~ 12 km (Fuenzalida et al., 2014). The spatial and
69 temporal closeness of the 2014 Iquique earthquake and the 2005 Tarapaca earthquake
70 raised the question about whether there is any link between the intraslab earthquakes
71 and megathrust earthquakes. We assess the static stress change brought about by the
72 2005 Tarapaca earthquake and subsequent VER, and explore whether the 2014
73 Iquique earthquake, or the energetic foreshock sequence in the preceding months,
74 may have been triggered by the 2005 Tarapaca earthquake.

75

76 **2. Surface deformation data**

77 **2.1 InSAR data**

78 We collect all postseismic SAR images from the C-band Envisat satellite spanning \sim
79 4 yrs from July 2005 to August 2009. In total, 29 SAR acquisitions from track 96 are
80 processed, each with a swath nearly 500 km long, large enough to give a good
81 coverage of the potential VER-induced long-wavelength ground displacement. Given
82 the relatively arid regional environment and sparse vegetation, the coherence is
83 generally good, leading to a total of 135 interferograms produced using the
84 JPL/Caltech software ROI_PAC (Rosen et al., 2004). Topographic phase is removed
85 using the 3-arc-second DEM from the Shuttle Radar Topography Mission (Farr et al.,
86 2007). The interferograms are unwrapped using a branch-cut method (Goldstein &
87 Werner, 1998).

88

89 The detection of low-amplitude, long-wavelength postseismic signals has in general
90 been limited by atmospheric delays and imprecise orbits (Jolivet et al., 2014). To
91 mitigate the atmospheric delay effect, we adopt the method described by Walters et
92 al. (2013) and Jolivet et al. (2014). We estimate the phase delay caused by water
93 vapor difference using data from the Medium-Resolution Imaging Spectrometer
94 (MERIS) instrument aboard the Envisat satellite, and that caused by spatial variation
95 of atmospheric pressure using the ERA-Interim global atmospheric model provided
96 by the European Centre for Medium-Range Weather Forecasts (ECMWF). The
97 effective use of MERIS data requires largely cloud-free ($< 25\%$ cloud) weather
98 conditions, which constrains the number of usable interferograms to 45 (Fig. S1 and
99 Table S2). After correction for atmospheric noise, uncertainties associated with the
100 satellite orbits are removed assuming a linear phase ramp across the interferogram.
101 Removal of a linear ramp should not compromise postseismic viscoelastic signal,
102 given the linear assumption along the whole track of data, although the effect of plate
103 interface creep may be reduced.

104

105 Given the depth of the earthquake (~ 90 km), any postseismic signal is expected to be
106 small (i.e. few millimetres). Considering the low signal-to-noise ratio, we chose to
107 build a deformation rate map, rather than a time-series. The rate map is produced by
108 averaging the added phase of 45 noise-corrected interferograms over their total time
109 span. Out of the 45 interferograms, pixels coherent in at least 40 interferograms are
110 stacked to produce a rate map.

111

112 The rate map shows two notable features (Fig. 2). One is the circular negative line-of-
113 sight change associated with activity at Sillajhuay volcano, with a peak rate of ~ 1.4
114 cm/yr. The other feature is a broad-scale negative millimeter-level line-of-sight
115 change over the area of the coseismic rupture. The postseismic range decrease is in
116 contrast to the circular zone of subsidence that occurred coseismically (Peyrat et al.,
117 2006). This contrast of coseismic and postseismic displacement is a first hint that the
118 postseismic relaxation mechanism is VER. In Section 3 below, we run models to
119 explore whether VER is a viable mechanism.

120

121 **2.2 GPS data**

122 In this study, we take advantage of the long-standing campaign measurements of the
123 GPS network installed by Chilean and French teams in North Chile starting in the
124 1990's (Ruegg et al., 1996). One profile composed of nine benchmarks traverses the
125 epicentral area of the Tarapaca earthquake (Ruegg et al., 1996; Chlieh et al., 2004;
126 Métois et al., 2013) from the coast to the vicinity of Sillajhuay volcano in a northeast
127 direction (Fig. 1). Measurements have been conducted on this network before the
128 Tarapaca earthquake, in 1996 and 2000, with an additional measurement in 2002 for
129 two stations (Table S3). The next survey was carried out in the month following the
130 Tarapaca earthquake, and subsequent measurements were conducted in 2010 and
131 2012 (Table S4).

132

133 We process all these data following the method described in Métois et al. (2013)
134 using the GAMIT-GLOBK software (King & Bock 2002), and obtain horizontal
135 velocities before and after the 2005 Tarapaca earthquake (Fig. 3a, and Table S3 &
136 S4). Velocities calculated first in the ITRF 2008 (Altamimi et al. 2011) are then
137 rotated into a fixed South America reference frame as defined by the pole from the
138 NNR Nuvel-1A model (DeMets Gordon 1994).

139

140 The large uncertainties associated with the pre-Tarapaca velocities (Table S3) are
141 mostly due to the bad quality of orbits and reference-frame stations prior to 1997, to
142 poor-quality antenna calibration models, and to the fact that the observation sessions
143 were often shorter than an entire day (in particular for the first 1996 campaign).
144 However, the large time span covered by the campaigns gives us confidence in these
145 velocities. We observe a significant difference between the pre- and post-Tarapaca
146 velocities (Fig. 3b and Table S5). The residual velocities are systematic and produce
147 a divergent pattern away from the epicentral area (Fig. 3b). We interpret this change
148 in velocity before and after the Tarapaca earthquake as mostly due to the postseismic
149 VER following the mainshock. However, part of this residual motion could be due to
150 afterslip following the mainshock, or due to changes in the degree of interseismic
151 coupling on the plate interface. The divergent pattern centred on the Tarapaca
152 epicentre speaks in favor of a signal dominated by postseismic VER, and we will test
153 this hypothesis in the following modelling section.

154

155 **3. Modelling**

156 **3.1 VER modelling**

157 Seismic tomography studies in this region reveal two important features of the
158 subduction zone rheology. The first is a uniform low P-wave attenuation for the
159 forearc, extending eastward to longitude -69° . In accordance with low surface heat
160 flow values, the low attenuation indicates a cold and stagnant forearc. The second is a
161 ~ 45 km-thick layer beneath the magmatic arc with high P-wave attenuation,
162 interpreted as being composed of partial melts ascending from mantle asthenosphere
163 (Schurr et al., 2003). The 2005 Tarapaca earthquake occurred within the subducted
164 rigid oceanic lithosphere. Given its proximity to the weak layer, there is a good
165 chance that any coseismic stress perturbation excited a VER process. To model the
166 VER-induced surface displacement, we used the software RELAX, a 3-D semi-
167 analytic package which can incorporate lateral rheological heterogeneity (Barbot &
168 Fialko 2010).

169
170 Our model configuration is shown on a cross section perpendicular to the trench in
171 Fig. 4. The elastic layers of the continental and subducted oceanic lithosphere are
172 both set to have a thickness of 40 km. A Poisson's ratio of 0.25 is assumed, and a
173 uniform shear modulus of 63.4 GPa is set for the entire model (Hetland & Zhang
174 2014). As an input, the coseismic slip model we adopted is from Peyrat et al. (2006),
175 as summarized in Table S1. Assuming a Maxwell viscoelastic rheology, we fixed the
176 oceanic mantle viscosity at 1×10^{20} Pa s, which is similar to the global mantle
177 average (Moucha et al., 2007), as used by Wang et al. (2012). Forward modelling

178 tests show that decreasing the viscosity of the oceanic mantle increases the RMS
179 misfit (Fig. S2). The rheological parameters we aimed to constrain include the size of
180 the strong forearc H_A (zone A), the thickness H_B and viscosity η_B of the weak layer
181 below the magmatic arc (zone B), and the viscosity of the area η_C (zone C) beneath
182 zone B (Fig. 4).

183
184 In order to obtain a model that fits the first-order spatial pattern as revealed by InSAR
185 observations, we progressively test various 3D rheological models in four steps (Fig.
186 5). For each step, the root-mean-square (RMS) misfit is plotted against the parameter
187 we are trying to constrain. Here, the area of Sillajhuay volcano is masked out before
188 calculating RMS, to avoid the potential influence of volcanic deformation on the
189 RMS computation. In the first step, we infer the best-fit size of zone A, treated as
190 elastic, while keeping the viscosities η_B of zone B and η_C of zone C as 4×10^{18} Pa s.
191 Then, in the second step, we vary the thickness H_B of zone B and find that a value
192 between 40 km and 50 km fits the observations well. In this step, zone B is taken to
193 be an elastic layer. We note that, the best fit thickness found in the second step is in
194 its lower limit, given the elastic assumption of zone B. Subsequently, we test
195 different viscosities η_C for zone C, and a viscosity of 8×10^{18} Pa s gives a minimum
196 RMS (Fig. 5c). Finally, we iterated the modelling with various viscosities for zone B,
197 and found a lower bound viscosity η_B of 5×10^{19} Pa s, since lower values rapidly
198 increase the RMS misfit (Fig. 5d). After finding the optimal values for each
199 parameter, we fixed the η_B and η_C as 5×10^{19} Pa s and 8×10^{18} Pa s, respectively, and
200 repeated the first step, to identify if the thickness for zone A stays at the optimal

201 value. This test (Fig. S3) shows a minimum RMS misfit at 45 km, suggesting
202 consistency with our optimal model. We also test for a viscoelastic zone A, by
203 decreasing its viscosity to the same value (4×10^{18} Pa s) as zones B and C. In this
204 case, the predicted displacement pattern differs markedly from the observations,
205 showing an opposite sign of range change (Fig. S4).

206
207 To validate our model, we compare the forward-modelled horizontal displacements
208 with the GPS data, which is the difference between the interseismic velocity before
209 and postseismic velocity after the Tarapaca earthquake. As shown in Fig. 6(a), the
210 InSAR-derived viscosity η_C of 8×10^{18} Pa s for zone C gives a very good azimuthal
211 fit to the GPS displacements, but the amplitudes are under-predicted. The azimuthal
212 alignment demonstrates that VER contributed at least part of the GPS-recorded
213 postseismic deformation. A model conducted with a lower viscosity η_C of 4×10^{18} Pa
214 s better fits the amplitude of the GPS data, but produces a systematic anti-clockwise
215 bias in azimuth for sites located northeast of the fault (Fig. 6b). We note that the
216 RMS calculated for the InSAR data differ less than 0.025 cm/yr for models with
217 viscosities η_C of 4×10^{18} and 8×10^{18} Pa s. With a lower viscosity in zone C, the
218 predicted InSAR displacement is slightly larger in amplitude in comparison to the
219 observations (Fig. S5). We further discuss this relative inconsistency between InSAR
220 and GPS data in Section 4.1.

221

222 **3.2 Static stress change modelling**

223 Static stress transfer is one of the mechanisms that can explain the occurrence of
224 earthquakes following other events. In addition to coseismic stress changes,
225 postseismic processes such as viscoelastic relaxation also modify the stress loading
226 on surrounding faults (Steacy et al., 2005). The 2014 Iquique earthquake and its main
227 M 6.7 preshock occurred about 9 years after the 2005 Tarapaca earthquake. The
228 proximity of these events both in space (Fig. 1) and time offers a good opportunity to
229 evaluate the possibility of static stress triggering.

230

231 To do so, we calculate the Coulomb stress change caused by the Tarapaca earthquake
232 and its postseismic VER process on the subduction shallow interface and preshock
233 nodal planes, based on the equation:

234

$$235 \Delta\sigma_f = \Delta\tau + \mu' \Delta\sigma_n$$

236

237 where $\Delta\tau$ and $\Delta\sigma_n$ are shear and normal stress changes, and μ' is the apparent friction.

238 We assume a value for μ' of 0.4, and note that changing this value does not affect
239 qualitatively the stress loading pattern. Details of coseismic rupture and viscoelastic
240 relaxation induced stress change are discussed in Section 4.3.

241

242 We choose the 2014 Iquique rupture plane from the coseismic model proposed by
243 Hayes et al. (2014). As there is currently no consensus on the rupture plane of the M
244 6.7 preshock, we projected stress changes onto both nodal planes derived from
245 moment tensor inversion by Fuenzalida et al. (2014). Fig. 7(a) shows the Coulomb

246 stress change superimposed on the slip models for the Iquique earthquake and its
247 largest aftershock, and the aftershock distribution from Schurr et al. (2014). It is
248 apparent that most aftershocks and the majority of slip locate in a negative stress
249 zone, implying a lack of direct triggering effect for the mainshock. Stresses resolved
250 onto the two pre-shock nodal planes show a positive loading on the shallow NNE-
251 dipping one (Fig. 7c). Hayes et al. (2014) propose that the M 6.7 preshock imparted a
252 positive stress on the plate interface where the M 8.2 Iquique earthquake occurred.
253 We further propose here that the 2005 Tarapaca earthquake may have acted as an
254 indirect trigger of the 2014 Iquique earthquake, via loading on the preshock.

255

256 **4. Discussion**

257 **4.1 Likely sources of uncertainties**

258 In this study, we have used an intermediate-depth intraslab normal faulting
259 earthquake to constrain the northern Chile subduction zone rheological structure. Our
260 models successfully retrieve the first-order spatial pattern of the geodetically-
261 recorded surface deformation. While a model with a viscosity of 8×10^{18} Pa s for
262 zone C fits well the InSAR displacement and GPS horizontal direction, it
263 underpredicts the GPS amplitude. Decreasing the viscosity for zone C to 4×10^{18} Pa s
264 increases the RMS misfit between InSAR model and data by ~ 0.025 cm/yr, but the
265 fit in GPS amplitude is improved. This inconsistency may be accounted for by
266 several sources of uncertainty.

267

268 The first source of uncertainty comes from the modelling assumption. We notice that,
269 the difference between observed and modelled GPS interseismic rate shows an
270 increasing trend in amplitude towards the coastal stations (Fig. 6b), indicating a
271 contribution from shallower process than deep VER. Given our model is constrained
272 only by InSAR data, of which the observation of postseismic deformation starts more
273 than one months after the earthquake, the early afterslip following the Tarapaca
274 earthquake may have much less effect on the InSAR postseismic observations. GPS
275 campaign observations start one week following the earthquake, which have greater
276 chance in mapping early afterslip on the normal fault. In addition to afterslip, the
277 assumption that the GPS-recorded difference of interseismic velocity is merely
278 caused by the post-Tarapaca VER effect excludes a possible change in subduction
279 zone coupling. Temporal variation in interseismic coupling has been seen in other
280 subduction zones, for example, on the Nicoya subduction interface (Feng et al., 2012)
281 and at northeastern Japan (e.g. Mavrommatis et al., 2014; Nishimura et al., 2004,
282 Loveless & Meade 2016), and could occur also in North Chile as suggested by Ruiz
283 et al. (2014). In our case, the degree of coupling may change as a result of stress
284 perturbation due to the Tarapaca earthquake and its postseismic VER process, and/or
285 the physical properties of the plate interface itself. Based on the above discussion, our
286 modelling strategy in this study using only InSAR postseismic deformation to
287 constrain the rheology brings down the effect from other postseismic processes
288 and/or temporal variations in coupling on the plate interface.

289

290 The second source of uncertainty is related to the regional complexity in rheology. In
291 our model, we did not consider possible viscosity heterogeneity in the along-strike
292 direction. As shown in seismic tomography studies (see fig. 5 in Schurr et al., 2003),
293 seismic attenuation varies at different latitudes, indicating heterogeneous rheology.
294 The heterogeneity can also be inferred from profiles drawn parallel to the satellite
295 flight direction across the deformed area. As shown in Fig. 8, decreasing the viscosity
296 for zone C produces a better fit to the data on the left hand-side (corresponding to the
297 southern part), while on the right hand side (northern side), the model deviates from
298 the data. It suggests a likely increase of viscosity from south to north. Our models
299 with uniform asthenospheric viscosity 8×10^{18} Pa s and 4×10^{18} Pa s could
300 potentially act as two end-member situations, demonstrating that a slight change in
301 viscosity can alter the observable deformation.

302

303 Third, given the low signal-to-noise ratio, any correction for noise in the InSAR data
304 could potentially introduce extra errors by overestimating/underestimating the
305 atmospheric contribution and phase ramp. As suggested by Bennartz & Fischer
306 (2001), the theoretical accuracy of MERIS water vapor retrieval is 0.16 cm, which
307 equals ~ 1.1 cm of uncertainty in the Envisat look direction. Li et al. (2006) also find
308 that the standard deviation of the difference between MERIS water vapor retrieval
309 and GPS-measured zenith delay is 0.11 cm, corresponding to ~ 0.74 cm of
310 atmospheric delay. Stacking of interferograms further reduces the atmospheric noise,
311 and thus improves the theoretical accuracy of the rate map by $N^{-0.5}$, where N denotes

312 the number of interferograms being stacked (Zebker et al., 1997). Assuming a similar
313 error for each interferogram used in this study, the theoretical error of our rate map is
314 ~ 0.17 cm/yr or ~ 0.12 cm/yr, based on the MERIS accuracy reported by Bennartz &
315 Fischer (2001) and Li et al. (2006), respectively (Fig. 8). In comparison to the
316 reported uncertainties of MERIS water vapor retrieval, a RMS misfit of ~ 0.15 cm/yr
317 between our model using 4×10^{18} Pa s for zone C and the rate map is not significant.
318 Taking the above factors into consideration, in addition to the large uncertainties
319 associated with the pre-Tarapaca GPS velocities, we suggest a likely range of
320 viscosity of $4 - 8 \times 10^{18}$ Pa s rather than stating a definite value.

321

322 **4.2 Subduction zone rheology**

323 Despite the uncertainties discussed above, our modelling constrains the local
324 rheological structure in several aspects. The forearc (zone A), bordered to the east by
325 a relatively weak lower crust and mantle lithosphere, must have high strength. Our
326 model also requires that the lower crust and mantle lithosphere are at least an order of
327 magnitude stronger than the asthenosphere beneath. Synthetic tests with low viscosity
328 for zones A and B produce a different sense of motion at the surface, incompatible
329 with the observations (Fig. S4).

330

331 Our inference of an asthenospheric viscosity of $4 - 8 \times 10^{18}$ Pa s is consistent with
332 estimates from recent postseismic studies of megathrust earthquakes. Klein et al.
333 (2016) investigated the GPS-recorded postseismic deformation following the 2010
334 Maule earthquake in Chile, and obtained a Maxwell viscosity of 3×10^{18} Pa s for the

335 asthenospheric mantle. In the study of postseismic deformation following the 1960 *M*
336 9.5 Valdivia, Chile earthquake, Ding & Lin (2014) obtained four increasing
337 asthenosphere viscosities corresponding to four different observation periods after the
338 earthquake. Their minimum value of 2×10^{18} Pa s, derived from the first four years
339 of observation right after the earthquake, agrees well with the viscosity we inferred
340 from the same period of observation after the Tarapaca earthquake. This consistency
341 validates our estimation of the short-term viscosity by fitting the averaged early
342 postseismic deformation over a short time scale (~ 4 yrs in both cases), although we
343 agree with the likely non-linear stress-dependency of viscosity in the long term
344 (Bürgmann & Dresen, 2008). An effective way to further test and modify current
345 models of rheological structure would be to investigate postseismic deformation
346 following the 2014 Iquique earthquake.

347

348 The regional rheological model inferred from geodetic observation shows
349 consistency with independent constraints from seismic attenuation studies, especially
350 in the contrast between the viscosity of zone A and B. Schurr & Rietbrock (2004)
351 proposed that a strong forearc nose (zone A) acts as a barrier that obstructs the
352 trenchward flow of hot asthenospheric mantle, and thus restricts the volcanic front to
353 the east. This is consistent with the numerical thermal modelling by Wada & Wang
354 (2009), which highlights the role of decoupling between slab and mantle wedge at
355 depth of ~ 80 km in the formation of stagnant cold fore arc. According to the linearity
356 of the volcanic front along the slab depth contour of ~ 90 km in northern Chile, we

357 suggest that, in this part of the subduction zone, a strong forearc nose may be a
358 common structure that sits on top of the subducted slab to a depth of ~ 90 km.

359

360 **4.3 Interplay of different earthquake types in a subduction zone environment**

361 Stress transfer has been successful in explaining the occurrence of megathrust
362 earthquakes in subduction zone earthquake cycles (e.g. Ding & Lin 2014). When
363 considering stress transfer between different phases of each cycle, it is also necessary
364 to consider the roles played by other types of earthquakes that occur within the
365 subduction zone. Large intraslab normal fault earthquakes are expected at the
366 downdip edge of the coupled subduction segment (Astiz et al., 1988; Lay et al.,
367 1989). Previous studies focused mainly on how the normal fault events themselves
368 interact with the megathrust earthquake cycle (e.g. Kausel et al., 1992; Gardi et al.,
369 2006). Here, we also investigated the stress loading due to VER processes following
370 large intraslab events.

371

372 In the case of the Tarapaca earthquake, the coseismic-only stress loading is
373 heterogeneous across the area of the coseismic rupture in the Iquique earthquake
374 (Fig. S6a). Postseismic VER exerts a positive stress change that overlaps with the
375 Iquique rupture, but is mostly downdip of it (Fig. S7a). For the plate interface down
376 to a depth of 40 km, stress loading due to postseismic VER is negative. Taken
377 together, the combined stress loading is mostly negative on the Iquique mainshock
378 rupture, showing that the postseismic Tarapaca VER process dominates the Coulomb
379 stress change across the Iquique rupture on the plate interface. On the nodal planes of

380 the M 6.7 preshock, the coseismic stress loading (Fig. S6b and c) from the Tarapaca
381 earthquake is ~ 8 times larger than that from the postseismic VER process (Fig. S7b
382 and c). The combined stress change projected on the shallow-dipping nodal plane of
383 the preshock is positive and reaches ~ 5 kPa. As mentioned earlier, the M 6.7
384 preshock positively loaded the 2014 Iquique rupture plane (Hayes et al., 2014).
385 Together, these lines of enquiry imply that, via loading of the M 6.7 preshock fault,
386 the static stress change from the Tarapaca earthquake may have acted as an indirect
387 trigger for the Iquique earthquake. At a late stage of the earthquake cycle when
388 interseismic stress accumulation is high, even a small stress perturbation may initiate
389 the subsequent failure of a long-coupled segment.

390

391 **5. Conclusions**

392 We take advantage of InSAR and GPS measurements covering the 2005 intraslab
393 Tarapaca earthquake epicentral area in North Chile, before and after the main shock,
394 to investigate the related postseismic relaxation effect. Our study demonstrates that
395 such a deep normal faulting intraslab earthquake generates a measurable deformation
396 at the surface that helps constrain the subduction zone rheology. We show that a
397 continental asthenosphere with viscosity of $4 - 8 \times 10^{18}$ Pa s underlying the lower
398 crust, together with a mantle lithosphere of viscosity $> 5 \times 10^{19}$ Pa s and a strong
399 forearc zone for the continental part, retrieve well the observed deformation pattern
400 in North Chile. Calculation of Coulomb stress change on the 2014 Iquique rupture
401 and its preshock nodal planes indicates that the static stress change from the Tarapaca

402 earthquake and its postseismic VER may have acted as an indirect trigger for the
403 Iquique earthquake, shedding new light on the overall sequence of seismic activity.

404

405

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411 who subsequently developed the network. We thank Roland Burgmann for valuable
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413

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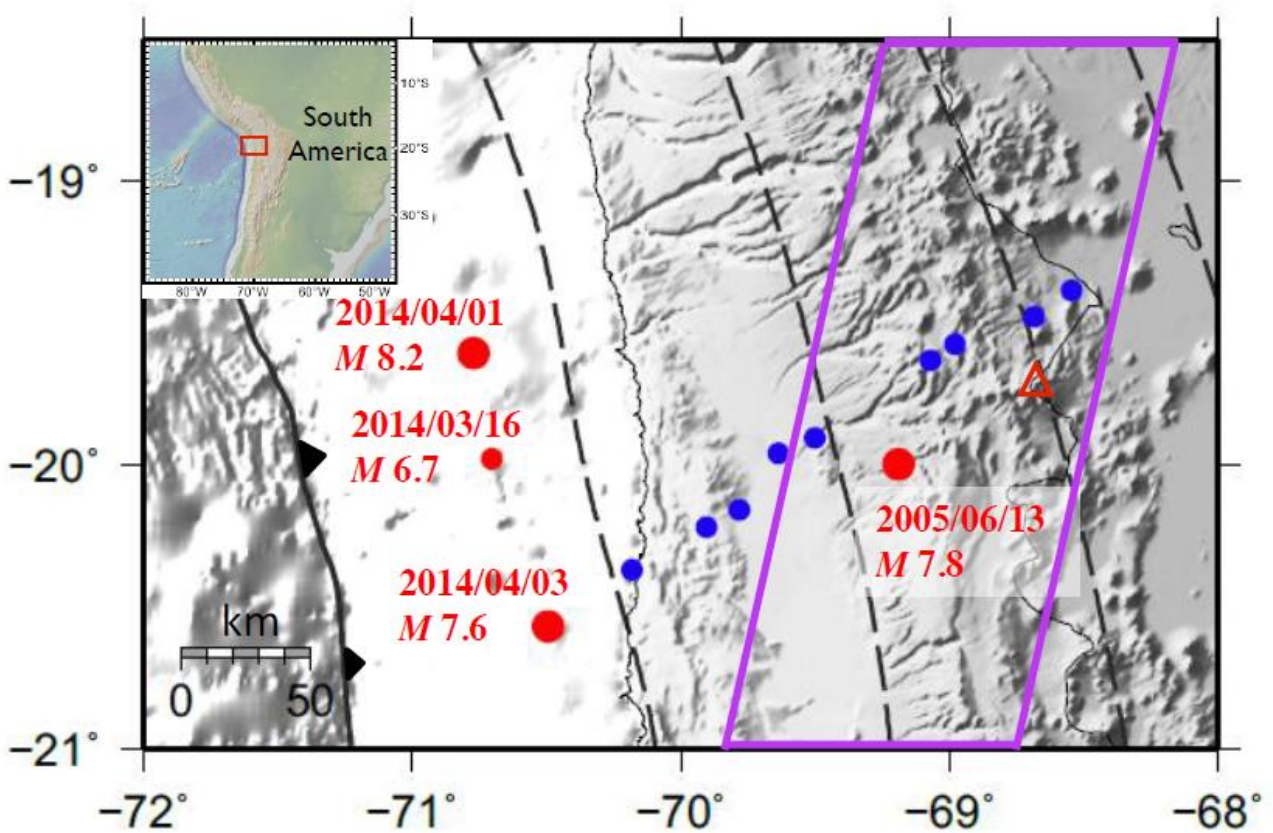
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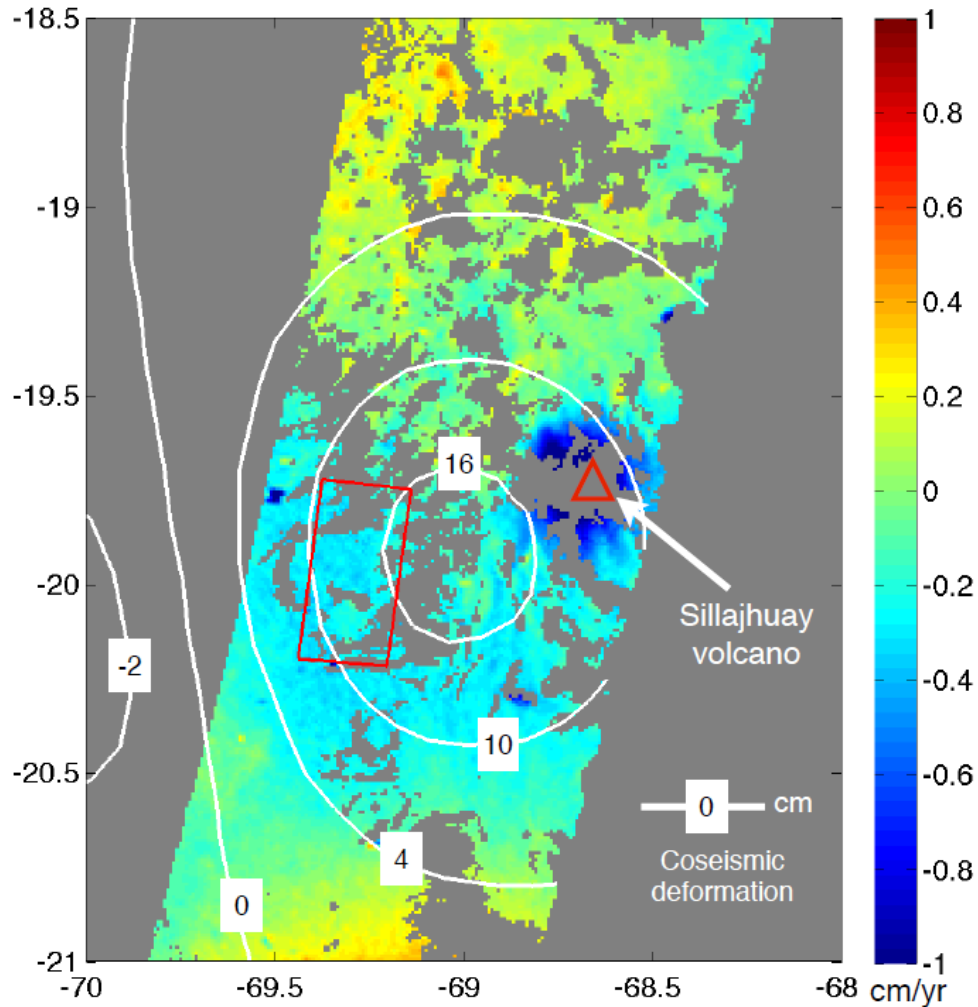
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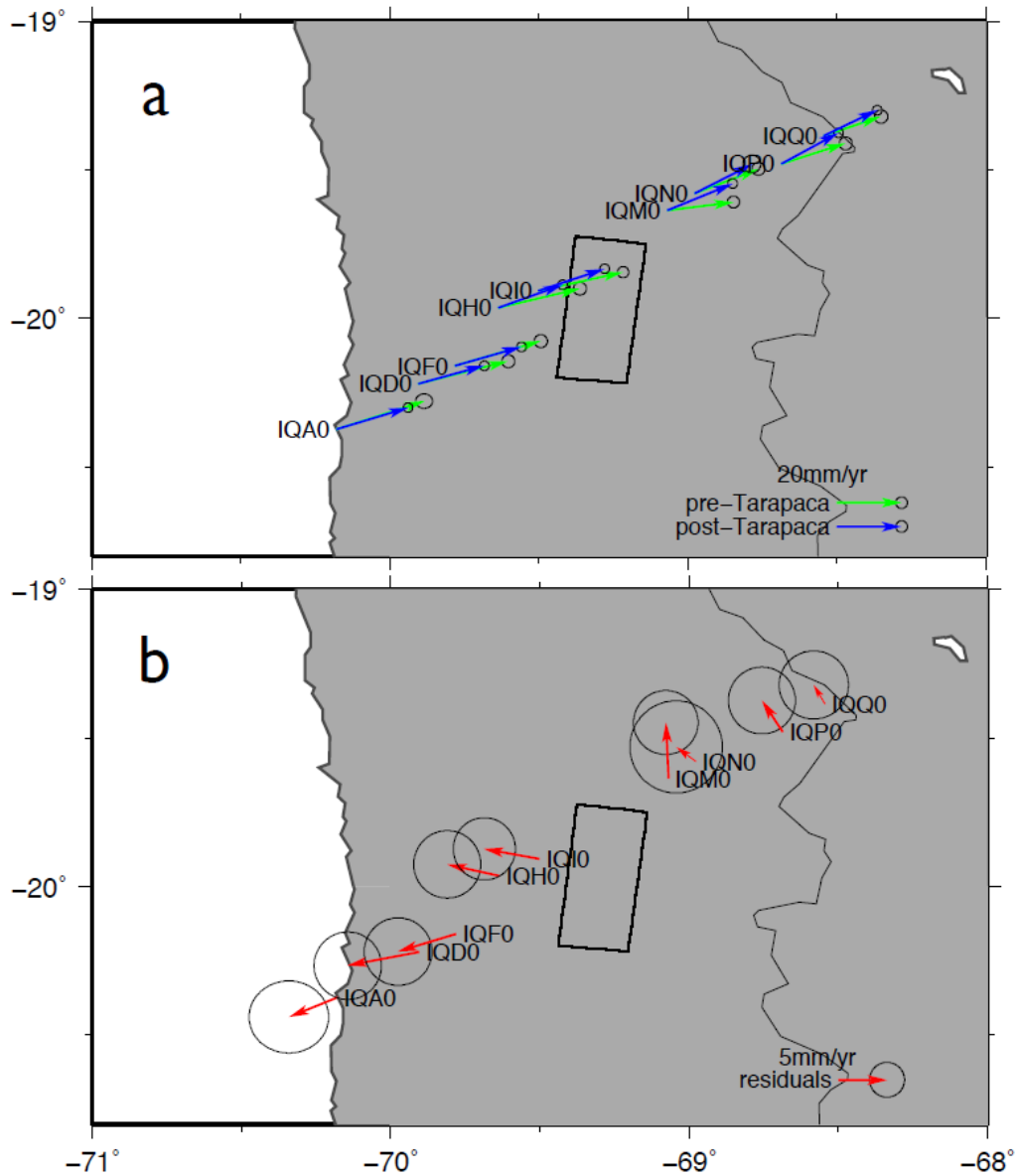
597 Fig. 1. Topographic map of the northern Chile subduction zone. Grey dashed lines
598 delineate the slab contours at 40 km depth intervals (Hayes et al., 2012). Red dots
599 mark the locations of earthquakes in this study. Red triangle marks the location of the
600 Sillajhuay volcano. Blue dots are survey-mode GPS locations (Métois et al., 2013).
601 Purple lines delineate the area covered by InSAR data shown in Fig. 2. Inset figure
602 shows the study region relative to South America.



603

604

605 Fig. 2. Rate map constructed from postseismic Envisat interferograms from June
 606 2005 to March 2009, as listed in Table S2. Warm colours indicate LOS motion away
 607 from the satellite. Coseismic deformation of the Tarapaca earthquake is shown by
 608 white contours, with positive values representing motion away from the satellite. Red
 609 rectangle is the surface projection of the Tarapaca rupture from Peyrat et al. (2006).
 610 Red triangle marks the location of the Sillajhuay volcano.



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612

613 Fig. 3. (a) Horizontal velocities relative to stable South America from survey mode

614 GPS measurements before (1996, 2000) and after (2005, 2010, 2012) the Tarapaca

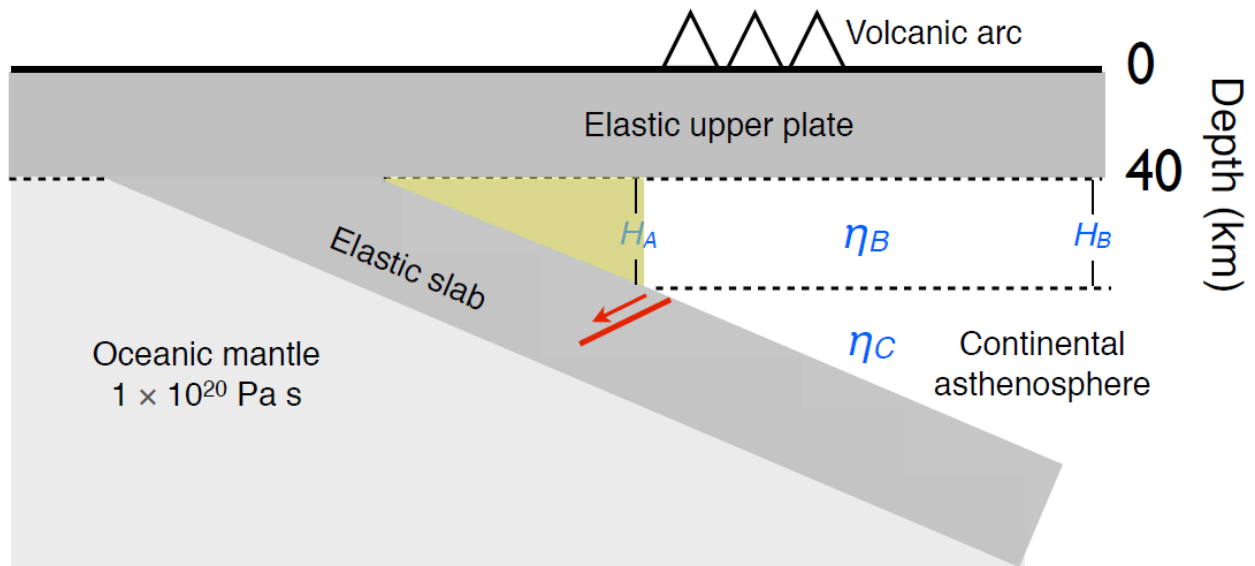
615 earthquake. Velocity vectors are tipped by the 80% confidence ellipses. (b)

616 Difference between the pre- and post-mainshock displacements shown in (a) (see

617 Table S3 - S5). Black rectangle is the surface projection of the Tarapaca rupture from

618 Peyrat et al. (2006)

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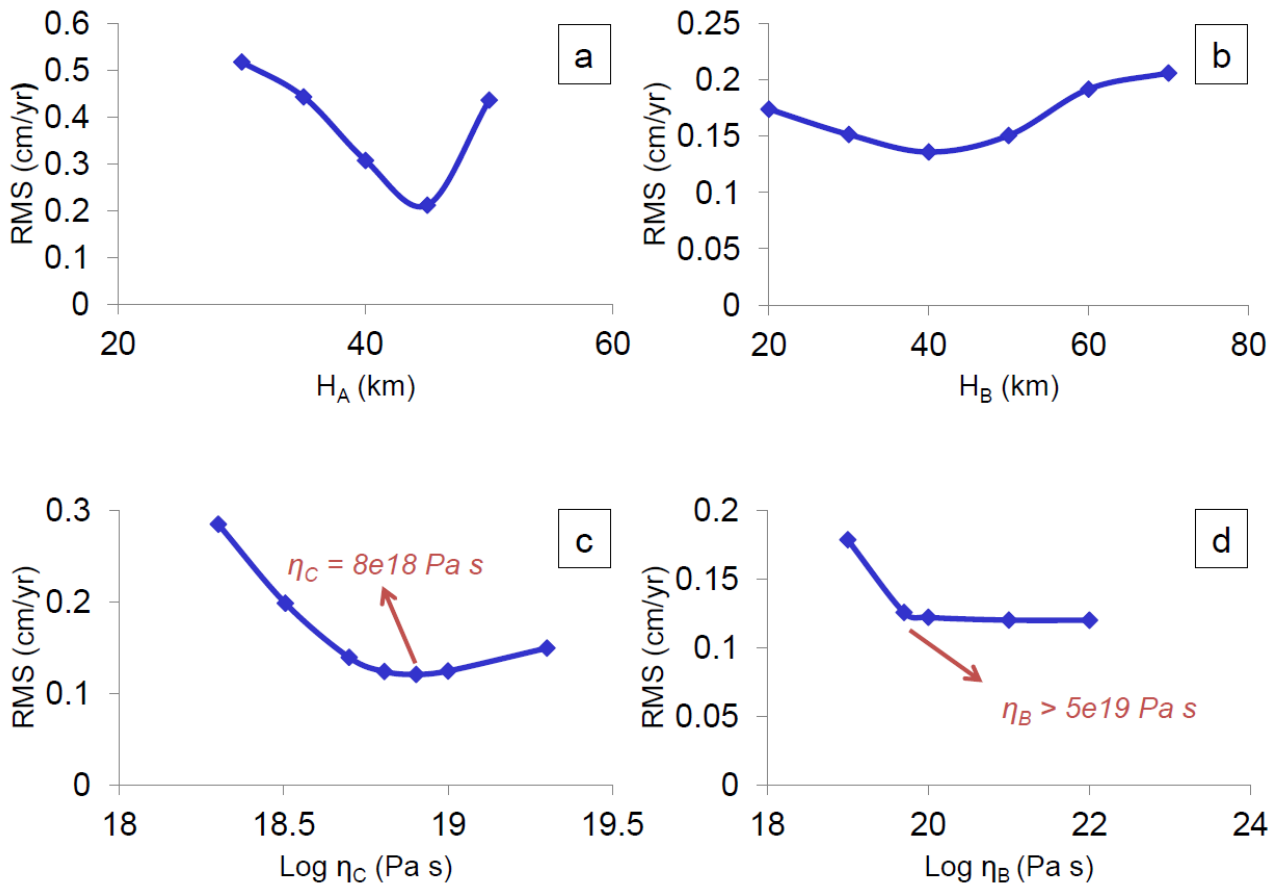
621

622 Fig. 4. Schematic cross-section perpendicular to the trench, showing the rheological
 623 structure constrained in this study. Yellow triangle represents the cold and stagnant
 624 part of the forearc (zone A). Parameters (H_A , H_B , η_B and η_C) in blue are variables to
 625 infer. Oceanic mantle viscosity is fixed, but its effect on surface deformation is tested
 626 and shown in Fig. S2.

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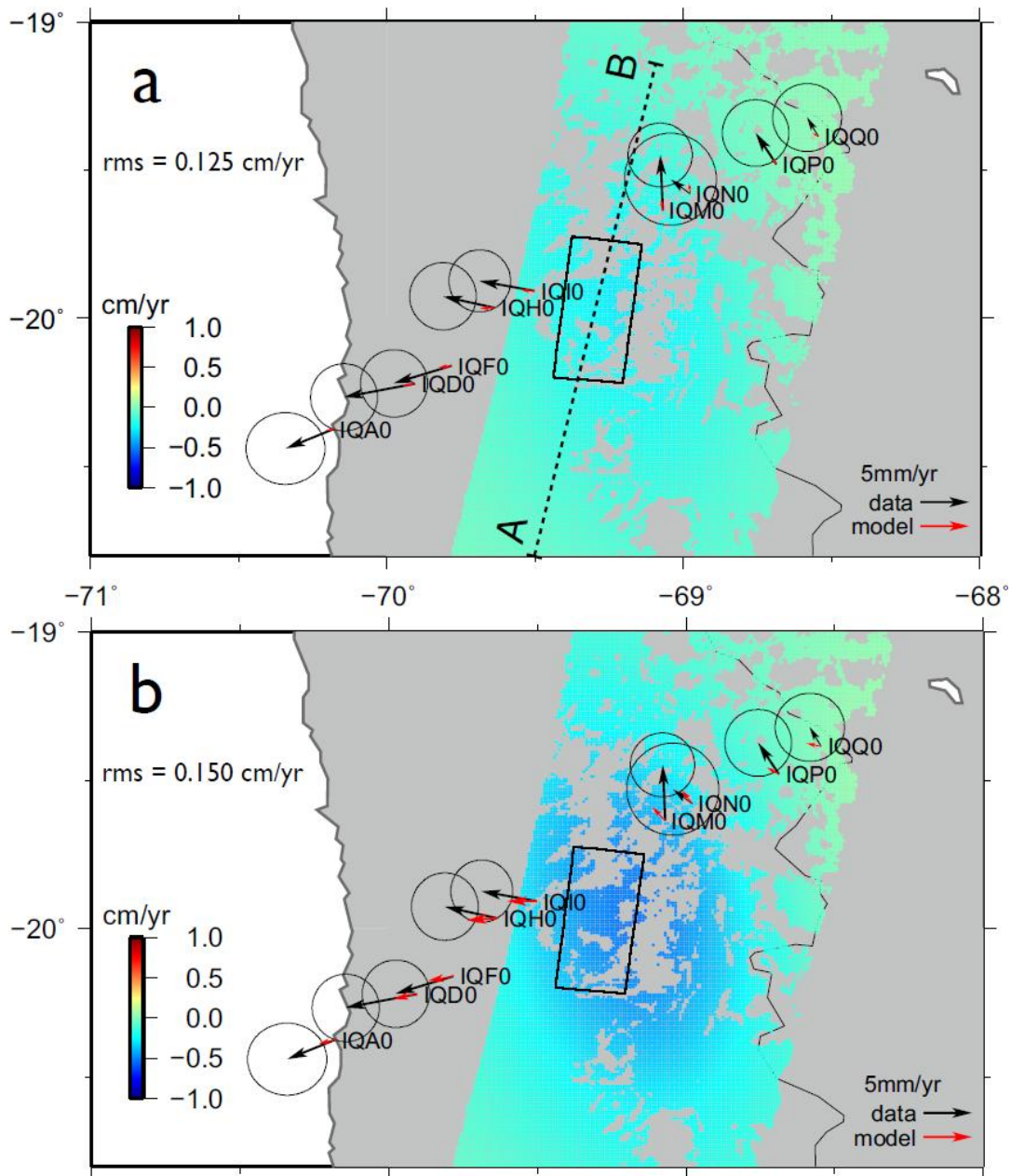


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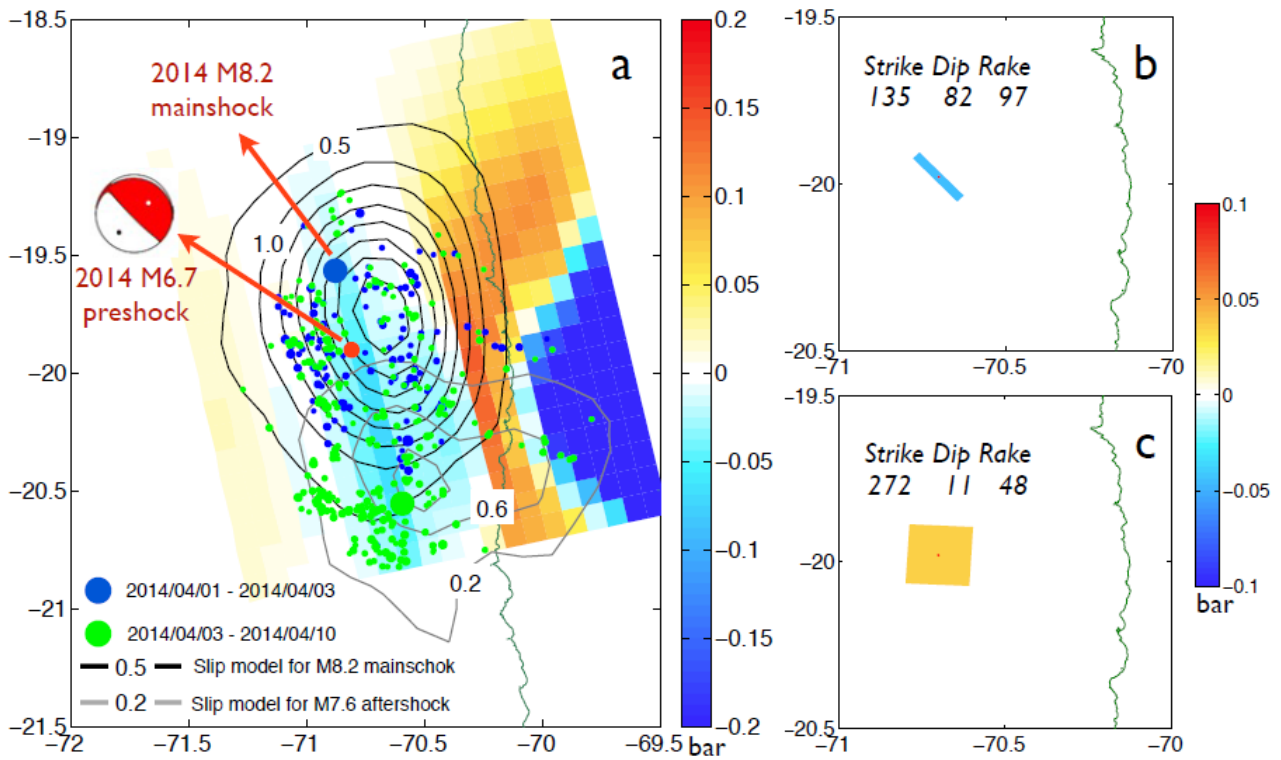
632 Fig. 5. RMS misfit curves derived during modelling. (a) RMS misfit versus thickness
 633 of zone A. Zone B and C are viscoelastic layers with fixed viscosity of 4×10^{18} Pa s.
 634 (b) RMS misfit versus thickness of zone B. Zone B is fixed as being elastic, while η_C
 635 is 4×10^{18} Pa s. (c) RMS misfit versus viscosity of zone C, holding A and B fixed at
 636 thicknesses and viscosities preferred in (a) and (b). (d) RMS misfit versus viscosity of
 637 zone B. Viscosity of zone C is fixed at 8×10^{18} Pa s, as derived in Fig 5(c).

638



639

640 Fig. 6. Modelled rate map of InSAR LOS displacement and GPS horizontal
 641 displacements (red arrows) with zone C viscosity η_C of: (a) 8×10^{18} Pa s; (b) 4×10^{18}
 642 Pa s. Other parameters are: $H_A = H_B = 45$ km, $\eta_B = 5 \times 10^{19}$ Pa s. Black arrows are the
 643 same as shown in Fig. 3(b). Black dashed line marked as A-B in (a) denotes the
 644 profile in Fig. 8. RMS misfit between model and InSAR data is given in units of
 645 cm/yr. Fig. 6(a) is reproduced in Fig. S8, but with the modelled GPS displacement
 646 scaled by 4 times, in order to clearly show the agreement in azimuth.

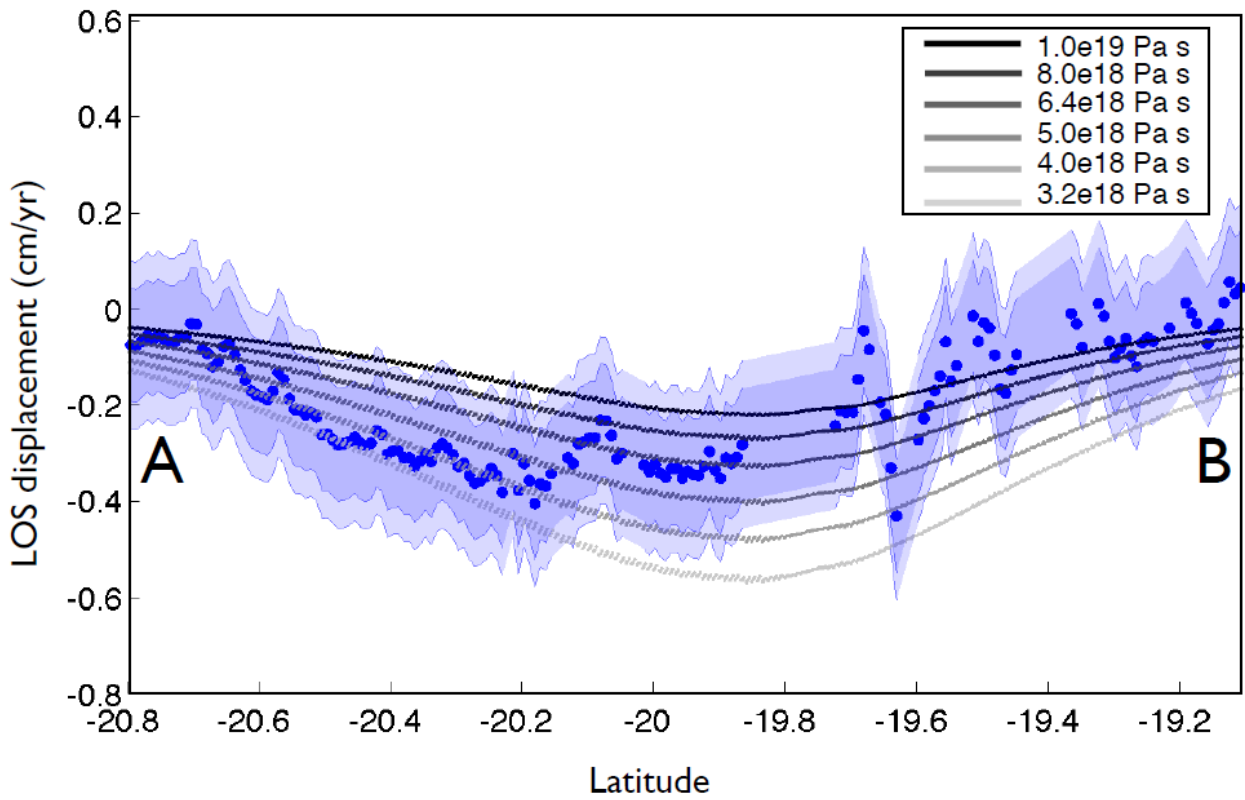


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649

650 Fig. 7. Coulomb stress changes induced by the co- and postseismic deformation
 651 associated with the Tarapaca earthquake. (a) Stress change on the subduction
 652 interface. Aftershock distribution and slip models for the M 8.2 Iquique earthquake
 653 and its M 7.6 aftershock, are from Schurr et al. (2014). (b) and (c) Stress change on
 654 the nodal planes of the M 6.7 foreshock.

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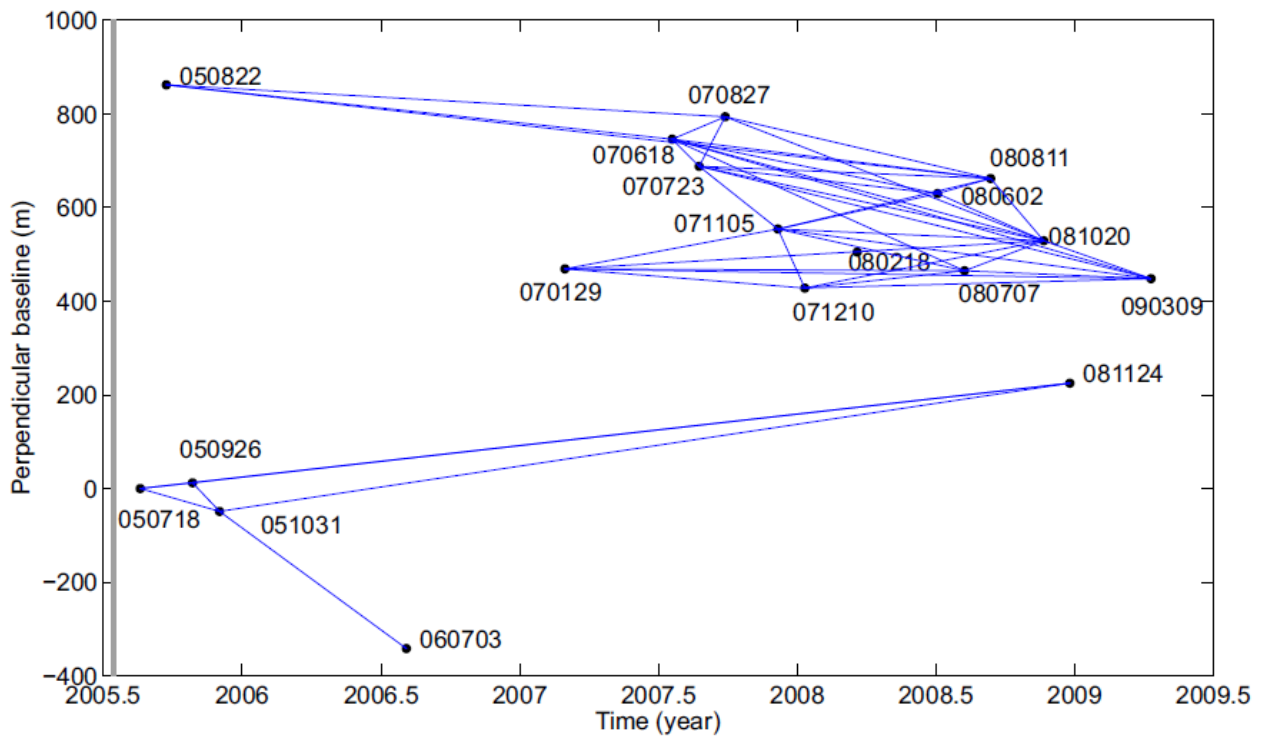


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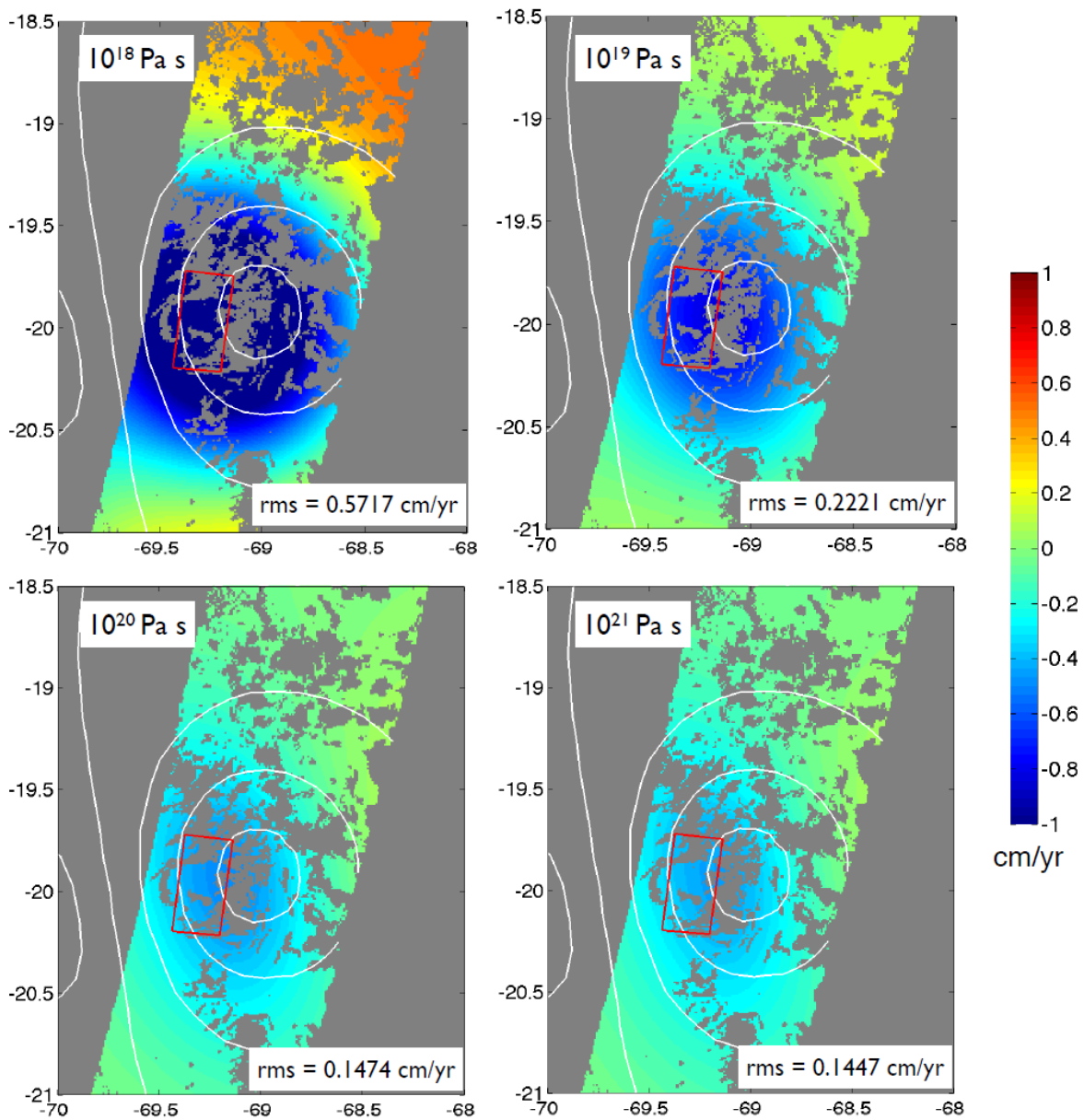
658 Fig. 8. LOS displacement changes with different viscosity for the continental
 659 asthenosphere along the A-B profile shown in Fig. 6(a). Blue dots show the LOS
 660 displacement rate along the profile. Higher viscosities provide a better a better fit to
 661 the northern part of the zone. The inner error bound (deep blue area) is a theoretical
 662 standard deviation derived from estimates by Li et al. (2006) of the MERIS data
 663 accuracy, and the outer bound (light blue area) is from Bennartz & Fischer (2001).

664



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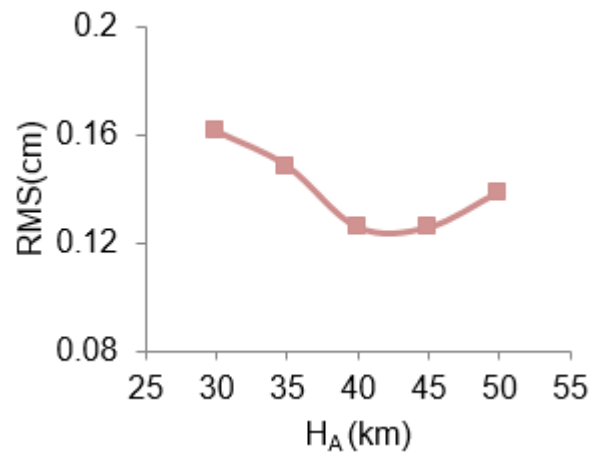
667 Fig. S1. Perpendicular baseline-time plots for the 45 Envisat interferograms (Table
 668 S2) used for rate map construction. Grey line marks the Tarapaca earthquake on 13
 669 June 2005.



670

671

672 Fig. S2. Modelled LOS displacements with various viscosity inputs for the oceanic
 673 mantle. Oceanic mantle viscosity is shown in top-left corner of each subfigure. Other
 674 parameters are same as used for Fig. 6(b). Warm colours indicate LOS motion away
 675 from the satellite. Coseismic deformation of the Tarapaca earthquake is shown by the
 676 same white contours as in Fig. 2. Red rectangle is the surface projection of the
 677 Tarapaca rupture from Peyrat et al. (2006). RMS misfit between InSAR model and
 678 data is given for each case in the lower right corner.



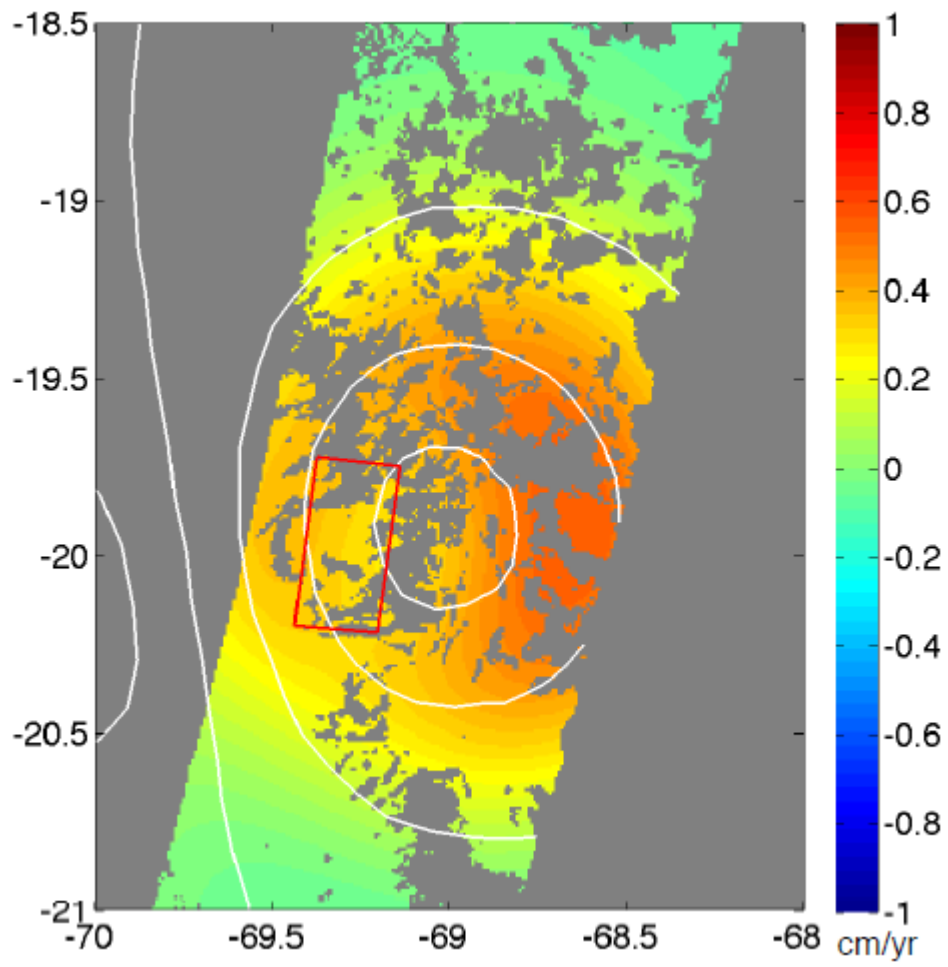
679

680 Fig. S3. RMS misfit versus thickness of Zone A. We assign 5×10^{19} Pa s and 8×10^{18}
681 Pa s for η_B and η_C respectively, as derived from progressive tests shown in Fig. 5.

682 This plot shows convergence with result in the first step, where the optimal thickness
683 for zone A is 45 km.

684

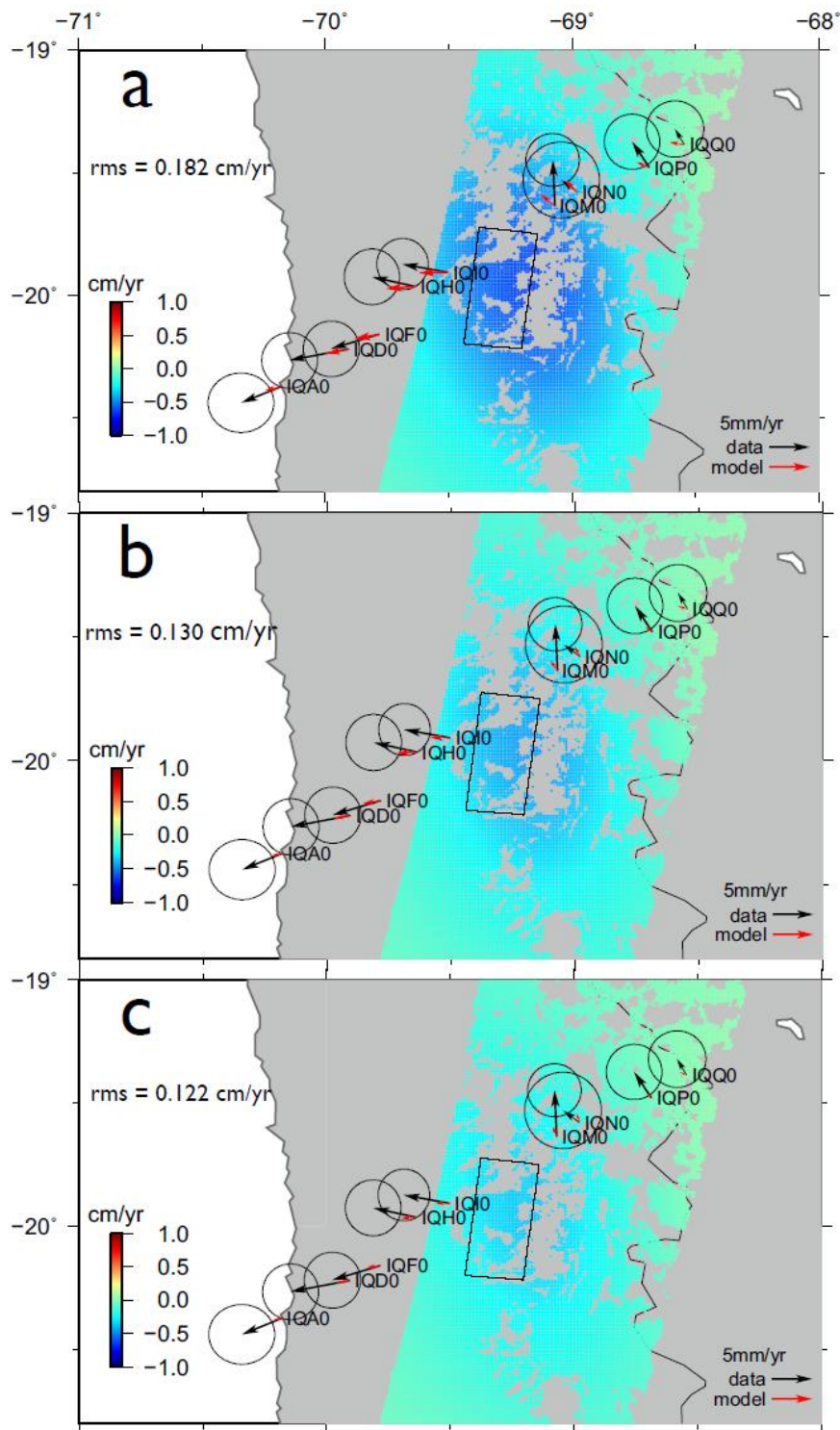
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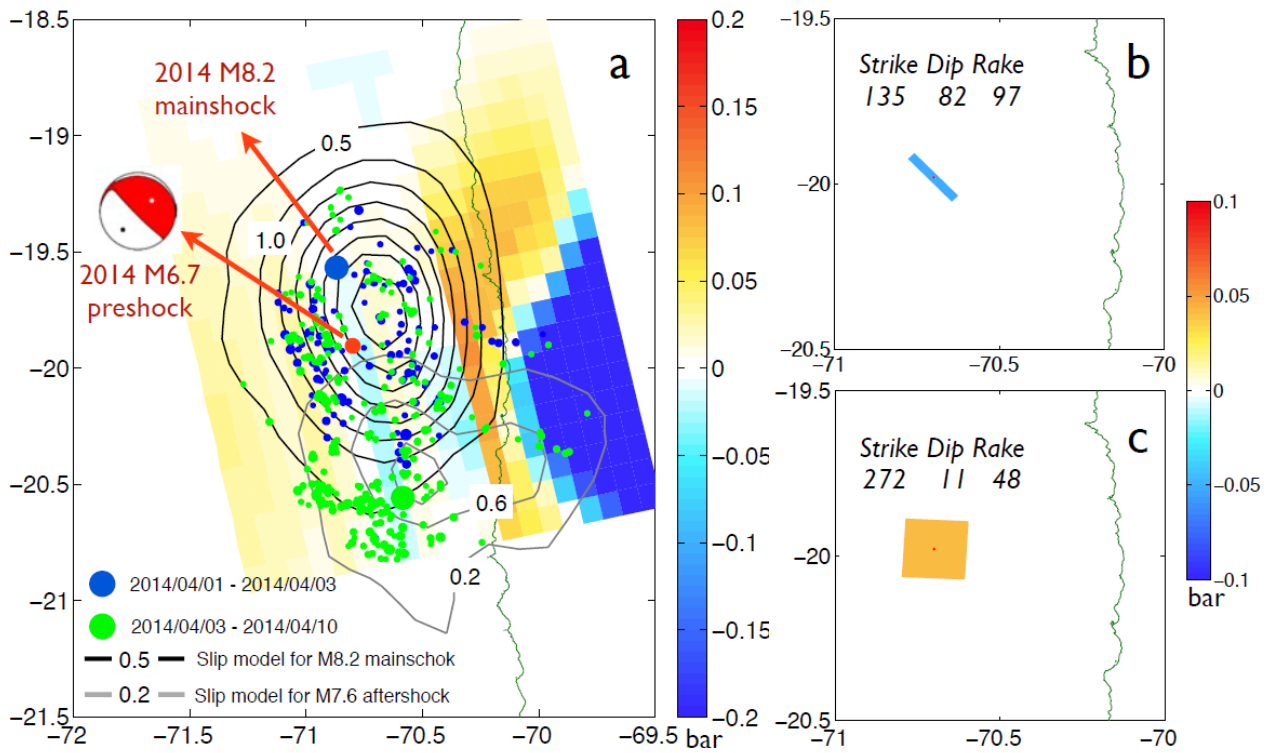
687

688 Fig. S4. Forward modeled LOS displacement with a viscoelastic zone A, sharing
689 same viscosity as zone B and C (4×10^{18} Pa s).



690

691 Fig. S5. Modelled rate map of InSAR LOS displacement and GPS horizontal
 692 displacements (red arrows) with zone C viscosity of: (a) 3.2×10^{18} Pa s; (b) 5×10^{18}
 693 Pa s; (c) 6.4×10^{18} Pa s. Other parameters are: $H_A = H_B = 45$ km, $\eta_B = 5 \times 10^{19}$ Pa s.
 694 Black arrows are the same as shown in Fig. 3(b). RMS misfit between model and
 695 data is given with unit cm/yr.

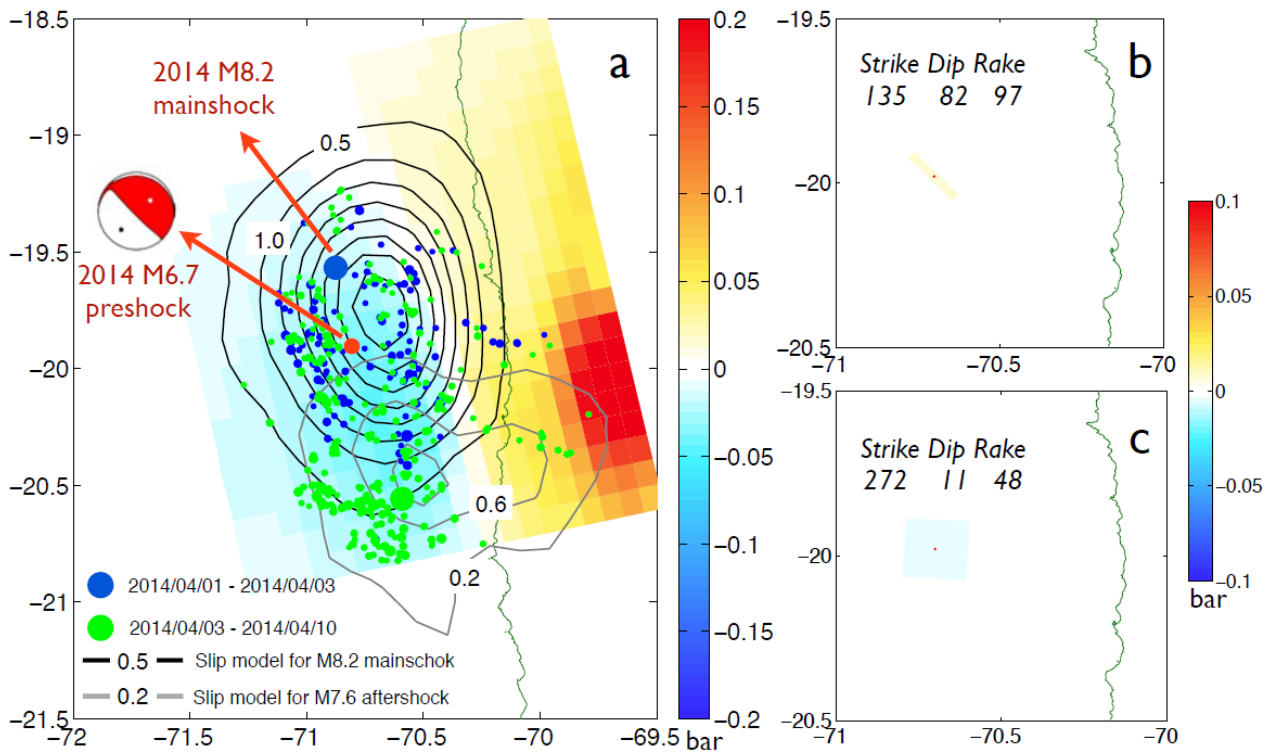


696

697

698 Fig. S6. Coulomb stress changes induced by the coseismic rupture of the Tarapaca
 699 earthquake only. (a) Stress change on the subduction interface. Aftershock
 700 distribution, and slip models for M 8.2 Iquique earthquake and its M 7.6 aftershock,
 701 are from Schurr et al. (2014). (b) and (c) Stress change resolved on the nodal planes
 702 of the M 6.7 preshock.

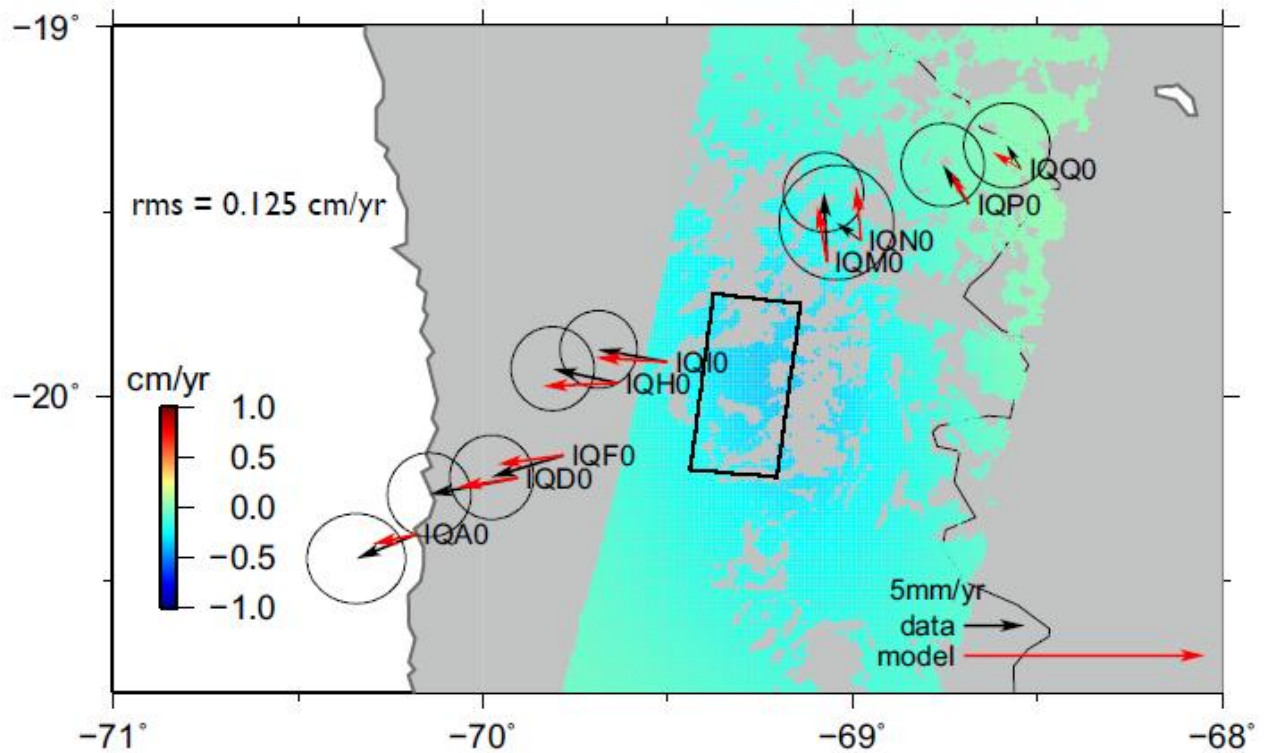
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704

705 Fig. S7. Coulomb stress changes induced by the postseismic VER after the Tarapaca
 706 earthquake only. (a) Stress change on the subduction interface. Aftershock
 707 distribution, and slip models for M 8.2 Iquique earthquake and its M 7.6 aftershock,
 708 are from Schurr et al. (2014). (b) and (c) Stress change resolved on the nodal planes
 709 of the M 6.7 preshock.

710



711

712 Fig. S8. A reproduction of Fig. 6. Modelled rate map of InSAR LOS displacement
 713 and GPS horizontal displacements (red arrows) with zone C viscosity of 8×10^{18} Pa s.
 714 In comparison to Fig. 6(a), modelled GPS horizontal displacements in Fig. S8 are
 715 scaled up by 4 times, in order to clearly show the agreement with data in azimuth.

716

717

718 **Tables**

719

720 Table S1. Source parameters of the 2005 Mw 7.8 Tarapaca earthquake (Peyrat et al.,
721 2006), used as input for VER modelling.

722

723

Length (km)	Width (km)	Top depth (km)	Strike	Dip	Rake	Slip (m)
54	24	90	189	24	-74	6.5

724

725

726

727

728

729 Table S2. List of postseismic interferograms stacked for rate map.

730

Postseismic interferogram pairs (yymmdd - yymmdd)
050718 - 050926
050718 - 051031
050718 - 081124
050822 - 070618
050822 - 070827
050822 - 080811
050926 - 051031
050926 - 081124
051031 - 060703
051031 - 081124
070129 - 071105
070129 - 071210
070129 - 080707
070129 - 081020
070129 - 090309
070618 - 070723
070618 - 070827

070618 - 080602
070618 - 080707
070618 - 080811
070618 - 081020
070618 - 090309
070723 - 070827
070723 - 071105
070723 - 080602
070723 - 080811
070723 - 081020
070723 - 090309
070827 - 080811
070827 - 081020
071105 - 071210
071105 - 080602
071105 - 080707
071105 - 080811
071105 - 081020
071105 - 090309
071210 - 080707
071210 - 081020
071210 - 090309

080602 - 080811
080602 - 081020
080707 - 081020
080707 - 090309
080811 - 081020
081020 - 090309

731

732

733 Table S3. Pre-Tarapaca GPS velocities in mm/yr relative to stable South America as
 734 defined by NNR-Nuvel1A model. The original velocities were calculated in
 735 ITRF2008.

736

STATION	LON	LAT	Ve	Vn	σ_e	σ_n	Measurements	Time-span
IQA0	-70.18	-20.373	27.19	8.72	1.46	1.25	2	1996.915- 2000.8
IQD0	-69.904	-20.221	27.78	7.05	1.11	1.12	2	1996.915-2000.8
IQF0	-69.781	-20.160	26.51	7.70	1.12	1.12	2	1996.915-2000.8
IQH0	-69.636	-19.964	25.23	5.93	1.11	1.12	2	1996.915-2000.8
IQI0	-69.501	-19.907	26.13	5.75	0.96	0.97	3	1996.915-2002.556
IQM0	-69.07	-19.636	20.54	2.55	1.03	1.02	3	1996.915-2002.556
IQN0	-68.978	-19.579	19.78	7.56	1.12	1.12	2	1996.915-2000.8
IQP0	-68.686	-19.481	19.87	6.39	1.11	1.11	2	1996.915-2000.8
IQQ0	-68.544	-19.387	17.71	6.03	1.17	1.14	2	1996.915-2000.8

737

738

739 Table S4. Post Tarapaca GPS velocities in mm/yr relative to stable South America as
740 defined by NNR-Nuvel1A model. The original velocities were calculated in
741 ITRF2008.

742

STATION	LON	LAT	Ve	Vn	σ_e	σ_n	Measurements	Time-span
IQA0	-70.18	-20.373	22.22	6.70	0.82	0.82	3	2005.501-2012.288
IQD0	-69.904	-20.221	20.40	5.66	0.82	0.82	2	2005.501-2012.288
IQF0	-69.781	-20.160	20.54	5.89	0.82	0.82	3	2005.501-2012.288
IQH0	-69.636	-19.964	19.86	7.11	0.81	0.82	3	2005.501-2012.288
IQI0	-69.501	-19.907	20.43	6.78	0.82	0.82	3	2005.501-2012.288
IQM0	-69.07	-19.636	20.28	8.30	0.82	0.82	3	2005.501-2012.288
IQN0	-68.978	-19.579	17.76	9.05	1.53	1.53	2	2010.458-2012.288
IQP0	-68.686	-19.481	17.71	9.66	0.81	0.81	3	2005.501-2012.288
IQQ0	-68.544	-19.387	16.53	8.02	0.82	0.82	3	2005.501-2012.288

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745 Table S5. Residual velocities (post-pre Tarapaca earthquake) in mm/yr interpreted as
 746 postseismic deformation in this study. Velocities are relative to stable South America
 747 as defined by NNR-Nuvel1A model.

STATION	LON	LAT	Re	Rn	σ_{re}	σ_{rn}
IQA0	-70.18	-20.373	-4.97	-2.02	2.28	2.07
IQD0	-69.904	-20.221	-7.38	-1.39	1.93	1.94
IQF0	-69.781	-20.160	-5.97	-1.81	1.94	1.94
IQH0	-69.636	-19.964	-5.37	1.18	1.92	1.94
IQI0	-69.501	-19.907	-5.7	1.03	1.78	1.79
IQM0	-69.07	-19.636	-0.26	5.75	1.85	1.84
IQN0	-68.978	-19.579	-2.02	1.49	2.65	2.65
IQP0	-68.686	-19.481	-2.16	3.27	1.92	1.92
IQQ0	-68.544	-19.387	-1.18	1.99	1.99	1.96

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