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Interseismic loading of subduction megathrust drives long term uplift in northern Chile

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Key Points:

- We propose an approach to quantify the ratio between elastic and inelastic deformation in subduction forearcs
- Two distinct correlations can be observed between interseismic and quaternary uplift rates
- We propose that 4 to 8% of interseismic uplift rates translates into persistent deformation in Northern Chile

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18 **Abstract**

19 Large earthquakes are the product of elastic stress that has accumulated over decades to
20 centuries along segments of active faults. Assuming an elastic crust, one can roughly esti-
21 mate the location and rate of accumulation of elastic stress. However, this general frame-
22 work does not account for inelastic, irrecoverable deformation, which results in large scale
23 topography. We do not know over which part of the earthquake cycle such deformation
24 occurs. Using InSAR and GNSS measurements, we report on a potential correlation be-
25 tween long-term, inelastic and short-term, interseismic vertical rates in northern Chile.
26 Approximately 4 to 8% of the geodetically-derived interseismic vertical rates translates
27 into permanent deformation, suggesting topography of the forearc builds up during the
28 interseismic period. This observation provides a quantitative basis for an improved under-
29 standing of the interplay between short-term and long-term dynamics along convergent
30 plate boundaries.

1 Introduction

Along a subduction interface, the quasi-continuous motion of converging plates leads to the build up of elastic stress where the interface is locked [Savage, 1983]. Large earthquakes, occurring as slip on the megathrust interface, redistribute elastic energy accumulated within the surrounding medium. In addition to these sudden changes in stress, parts of the subduction interface may slip aseismically during the interseismic period, either episodically during slow slip events [Hirose *et al.*, 1999; Dragert *et al.*, 2001] or over decades to centuries [Mazzotti *et al.*, 2000; Metois *et al.*, 2016]. Averaged over the interseismic period, a coupling coefficient, γ , is usually used to describe the inferred local slip deficit normalized by the long term convergence rate along the subduction megathrust (Fig. 1). $\gamma = 1$ corresponds to a fully locked interface, with a potential for slip during future earthquakes. $\gamma = 0$ indicates an inferred slip rate equal to the local plate convergence rate [Savage, 1983]. Subduction interfaces are paved with locked sections (i.e. portions of the interface with $\gamma \simeq 1$) separated by creeping sections [e.g. Konca *et al.*, 2008; Nocquet *et al.*, 2014] (i.e. $\gamma \ll 1$).

Large megathrust earthquakes uplift the surface located above a patch of slip and induce subsidence around it [Savage, 1983; Vigny *et al.*, 2011; Simons *et al.*, 2011]. To the extent that the coupling coefficient, γ , is stable through time, segments that ruptured during a large event reload during the interseismic period causing surface deformation of opposite sign to that occurring during the earthquake. In the most simplistic model, the amount of slip during earthquakes should be balanced by the slip deficit that builds up in the interim period, at least once integrated over many earthquake cycles. However, along several subduction zones, the coastal domain, a region defined to extend from the coast to the continental shelf (Fig. 1), experiences long-lived vertical displacement, either subsidence or uplift, over geological times (i.e. 10^5 - 10^6 years), for example as inferred from the geometry of raised terraces in Chile [Saillard *et al.*, 2009; Regard *et al.*, 2010], Japan [Matsu'ura, 2015] or Greece [Mouslopoulou *et al.*, 2016] or the erosion patterns along rivers in Mexico [Ramírez-Herrera *et al.*, 2018]. Such long-term, irrecoverable deformation can be modeled with a purely plastic rheology, for instance using critical taper theory [Davis *et al.*, 1983]. Such models, however, cannot address the question of when inelastic deformation cumulates in the forearc with respect to the different phases of the earthquake cycle.

63 For comparison, typical vertical displacement rates over geological times for the
64 coastal domain are generally one order of magnitude lower than typical interseismic uplift
65 or subsidence rates [a fraction of millimeter per year vs several millimeters per year, e.g.
66 *Béjar-Pizarro et al.*, 2013; *Jolivet and Simons*, 2018; *Regard et al.*, 2010; *Melnick*, 2016;
67 *Hashima and Sato*, 2017, Fig. 1]. Comparatively, large megathrust earthquakes will gen-
68 erate meter-scale vertical displacements every time they occur [e.g. *Vigny et al.*, 2011; *Si-*
69 *mons et al.*, 2011]. We cannot yet assess how much permanent uplift or subsidence occurs
70 during earthquakes [*Baker et al.*, 2013; *Melnick*, 2016] or during the interseismic period
71 [*Béjar-Pizarro et al.*, 2013; *Saillard et al.*, 2017], as our understanding of the earthquake
72 cycle is limited to simplified elastic and visco-elastic models which do not allow for per-
73 manent deformation to build up within the upper crust.

74 Focusing on the interseismic period, correlations between distributions of coupling
75 coefficients along megathrusts and morphotectonic features [*Victor et al.*, 2011; *Saillard*
76 *et al.*, 2017] or large amplitude gravity anomalies [*Song and Simons*, 2003; *Wells et al.*,
77 2003] suggest that interseismic fault locking is to first order a long-term feature, stable
78 over many seismic cycles. Interseismic fault locking is thought to imprint its signature
79 in the morphology of the subduction forearc. For instance, a qualitative relationship has
80 been described between the maximum depth of high coupling coefficient along subduc-
81 tion megathrust and the position of the coastal domain [*Béjar-Pizarro et al.*, 2013; *Rousset*
82 *et al.*, 2015; *Mouslopoulou et al.*, 2016; *Saillard et al.*, 2017] (Fig. 1). If part of the uplift
83 measured during the interseismic period is not recovered elastically during earthquakes oc-
84 ccurring along the megathrust, then anelastic deformation accumulates and the coastal do-
85 main will uplift over geological times [*Song and Simons*, 2003; *Mouslopoulou et al.*, 2016;
86 *Saillard et al.*, 2017]. Because most of the slip is often located offshore, earthquakes along
87 subduction zones tend to lower the coastal domain [*Simons et al.*, 2011; *Duputel et al.*,
88 2015]. However, if coseismic slip reaches depths below the coastline, the coastal domain
89 will uplift coseismically [*Grandin et al.*, 2016; *Vigny et al.*, 2011].

90 The residual permanent vertical displacement will therefore depend on the relative
91 balance between coseismic slip and interseismic slip deficit and on the position of the
92 hinge line with respect to the coastal domain [*Saillard et al.*, 2017]. Contradictory conclu-
93 sions have been reached considering the contribution of earthquakes and interseismic load-
94 ing to the permanent deformation of the forearc. For instance, *Melnick* [2016] proposes
95 that intermediate-depth earthquakes are responsible for the uplift of the northern Chilean

96 coast while he discards the influence of interseismic fault locking. Other studies point to
97 spatial relationships between patterns of coupling coefficients and long term features of
98 the forearc in support of the influence of interseismic fault locking on topography building
99 [*Béjar-Pizarro et al.*, 2013; *Rousset et al.*, 2015; *Mouslopoulou et al.*, 2016; *Saillard et al.*,
100 2017].

101 **2 Data and Methods: The case of northern Chile**

102 Here, we use direct estimates of interseismic displacement rates and compare these
103 with long term uplift rates in subduction regions. Geodetic measurements of surface dis-
104 placement rates, including GNSS (historically referred to as GPS) and InSAR (Synthetic
105 Aperture Radar Interferometry) measurements, are available over many subduction fore-
106 arcs. InSAR provides measurements of surface deformation along the direction of the
107 Line-Of-Sight (LOS) of a satellite. For SAR acquisitions from the Envisat satellite, for
108 instance, the LOS incidence angle is approximately 23° resulting in a high sensitivity to
109 vertical displacements. However, InSAR is also sensitive to horizontal displacements con-
110 founding attempts to isolate a purely vertical component of displacement. GNSS provides
111 time series of three-component displacements at sparsely distributed locations. However,
112 uncertainties on GNSS-derived vertical displacement rates typically reach several millime-
113 ters per year, making it a challenge to measure the sub-millimetric displacement rates ex-
114 pected from permanent deformation [*Melnick*, 2016].

115 One solution is to estimate a model of the distribution of coupling coefficients along
116 the megathrust that captures all measurements of vertical, horizontal and LOS displace-
117 ments along with their respective uncertainties in order to produce an a posteriori estimate
118 of vertical interseismic motion, and compare those predictions with long-term displace-
119 ment rates. Along subduction zones, where locking of the megathrust is mostly offshore,
120 it is possible to infer locked asperities with geodetic measurements made onshore [*Béjar-*
121 *Pizarro et al.*, 2013; *Lin et al.*, 2015]. However, because we are generally limited to ob-
122 servations made at the surface and onshore, inferred distributions of coupling coefficient
123 often differ significantly due to different modeling strategies and different regularizations
124 of the inverse problem. In the present study, we do not particularly care about this non-
125 uniqueness since we only use these models to provide the best possible estimate of verti-
126 cal displacement rates over the interseismic period consistent with the available geodetic
127 data. Effectively, models are used to interpolate the data.

128 Considering the duration of the modern instrumental period with respect to both
129 the characteristic time of the earthquake cycle and the longer time required to accumulate
130 significant anelastic deformation, it is crucial to focus on a region where we will be able
131 to measure surface displacements that are unambiguously related to interseismic loading
132 of the megathrust. In particular, the relaxation period following a large earthquake might
133 extend over decades and could strongly influence what is considered as interseismic de-
134 formation [Trubienko *et al.*, 2013; Hashima and Sato, 2017]. Finally, this region must be
135 densely sampled by GNSS measurements and its geographical characteristics must be opti-
136 mal for InSAR.

137 Northern Chile has been hit by several earthquakes, including the large, M_w 7.7, To-
138 copilla earthquake [Bejar-Pizarro *et al.*, 2010; Schurr *et al.*, 2012] in 2007 and the great,
139 M_w 8.1, Iquique earthquake [Duputel *et al.*, 2015] in 2014. There, inferences of along-
140 strike variations of coupling coefficients are broadly consistent between models [Metois
141 *et al.*, 2016; Schurr *et al.*, 2014]. The last great megathrust event prior to the 2014 Iquique
142 earthquake occurred in 1877 with an estimated magnitude of 8.6 [Comte and Pardo, 1991;
143 Metois *et al.*, 2013]. Therefore, we assume that surface displacement rates in this region
144 prior to the 2014 Iquique earthquake are not significantly changing due to postseismic vis-
145 cous relaxation and truly reflect the effect of locking along the megathrust.

146 The relative aridity of northern Chile makes it an ideal target for InSAR measure-
147 ments [e.g. Pritchard and Simons, 2002; Béjar-Pizarro *et al.*, 2013]. We use SAR data ac-
148 quired by the Envisat satellite over the 2003-2010 period to derive a map of LOS ground
149 velocity (Fig. 2). We use a multi-pixel method to infer displacement time series, displace-
150 ment rates averaged over the observation period and earthquake-related offsets accounting
151 for spatial covariances as well as other nuisance parameters [Jolivet and Simons, 2018].
152 Our approach resolves long wavelength signals in SAR acquisitions, hence does not re-
153 quire input from GNSS data. The resulting map of LOS displacement rates highlights a
154 variable uplift rate along the coastal domain, with LOS rates spanning a 0-4 mm/yr range.
155 Measurements of horizontal interseismic displacement rates are available thanks to the
156 dense continuous GNSS measurements that have been deployed since 2000 [Simons *et al.*,
157 2010] and are complementary to our InSAR displacement rate maps [Metois *et al.*, 2016].

158 We use a backslip approach to estimate the coupling coefficient from geodetic dis-
159 placement rates [Savage, 1983] and apply a Bayesian formalism to explore the range of

160 possible models given the GNSS and InSAR surface rates [Jolivet *et al.*, 2015]. The mean
 161 interseismic coupling model is broadly consistent with the most recent published studies
 162 [Li *et al.*, 2015; Metois *et al.*, 2016]. We infer a highly coupled segment extending north-
 163 wards from, at least, Antofagasta in the south up to 20.5°S and a relatively less coupled
 164 segment offshore Iquique (Fig. 3). The mean model has lower coupling coefficients at the
 165 trench but this feature is not resolved given that we are limited to observations made on-
 166 shore (see supplementary materials). There is a clear and robust separation between the
 167 coupled segments, with a narrow barrier with a coupling coefficient of 0 (maximum *a pos-*
 168 *teriori*, mean equals 0.1 ± 0.1). Contrary to recently published models [Li *et al.*, 2015;
 169 Metois *et al.*, 2016], we do not infer fault locking underneath the south American conti-
 170 nent. This inference of no kinematic locking extending below the coastline is driven by
 171 the InSAR observations indicating only onshore uplift [Béjar-Pizarro *et al.*, 2013].

172 Our goal is to compare the short-term vertical displacements predicted by our cou-
 173 pling models to long-term estimates of coastal uplift. Figure 4 shows the vertical displace-
 174 ment rates over the forearc predicted by our mean coupling model. To first order, topog-
 175 raphy shows striking similarities with this pattern of vertical rates, as maximum uplift is
 176 located in the coastal cordillera. However, topography is a signal integrated over millions
 177 of years, amalgamating the effects of tectonics, mantle dynamics and erosion. Therefore,
 178 topography alone cannot be simply used to study the influence of short term megathrust
 179 dynamics on the building of topography.

180 The presence of marine terraces all along the Chilean coast documents the long term
 181 uplift of the coastal domain over the quaternary [Regard *et al.*, 2010; Saillard *et al.*, 2017].
 182 Using a landscape evolution model that accounts for variations in sea level to reproduce
 183 the topography of the coastal domain and, in particular, the shape of the so-called coastal
 184 rasa, based on the competition of wave erosion and uniform uplift, Melnick [2016] de-
 185 termined uplift rates over the quaternary with dense spatial sampling. These rates quan-
 186 tify the uplift of the Chilean coast at an average rate of 0.13 ± 0.04 mm/yr since, at least,
 187 1 Myr.

188 Lateral variations within these quaternary uplift rates are visible along the coast.
 189 Melnick [2016] finds no obvious correlation between the long-term rates and interseismic
 190 rates. However, the interseismic model used to predict interseismic rates assumes fully
 191 homogeneous kinematic locking of the megathrust down to 35 km-depth without along-

192 strike variations, which is inconsistent with our observations and model of coupling co-
193 efficient (Fig. 3) as well as with some previously published models [e.g. *Li et al.*, 2015;
194 *Metois et al.*, 2016].

195 **3 Results: comparing short and long term uplift rates**

196 In contrast to previous analysis, we observe a slight correlation (Pearson product-
197 moment correlation coefficient of 0.5) between the distribution of coupling coefficients
198 offshore (i.e. 50 km offshore in the direction of convergence) and quaternary uplift of the
199 coastal domain (Fig. 4). Quaternary uplift appears faster where coupling coefficient is high
200 along the megathrust. However, because of the inherent limitations in inferred models of
201 coupling (i.e. non-uniqueness), we rather rely on the relationship between predicted inter-
202 seismic and long-term uplift. Effectively, as alluded to earlier, we consider our coupling
203 model as no more than a physics-based interpolation, including uncertainties, of interseis-
204 mic surface displacements along the coast. Although we do not include vertical GNSS-
205 derived velocities in our inversion, a first order comparison of the predicted interseismic
206 uplift at available GNSS stations suggests a relatively correct agreement with our predic-
207 tions [see supp. mat. and *Blewitt et al.*, 2016]. Such validation against an independent
208 data set should be taken with caution as formal uncertainties in GNSS-derived velocities
209 might be underestimated, especially for vertical rates of motion. In addition, we need to
210 consider that numerous active faults have been recognized in the forearc, in particular over
211 the regions of Tocopilla, Salar Grande and North of Pisagua [*Allmendinger and González*,
212 2010]. All types of faulting, reverse, normal and strike slip, active over the neogene and
213 quaternary have been identified in specific regions along the coastline, which could indi-
214 cate along strike variations in the long term behavior of the forearc.

215 Plotting the interseismic, predicted, uplift rates with respect to the modeled quater-
216 nary uplift rates, it appears that points can be separated in two groups (please refer to sup-
217 plementary materials for a plot without our interpretation). Within each group, short and
218 long term rates correlate. After separating these points in two distinct groups, we find that
219 there is a spatial coherence within these groups. In a first group (colored dots on figure
220 4), we consider points from regions that have experienced extensive neogene and quater-
221 nary active faulting according to *Allmendinger and González* [2010], while in the second
222 group, we consider points from regions devoid of active faulting (black dots in Fig. 4).
223 Points are not randomly taken in a group or the other and rather cluster spatially along the

224 coastline. Taken independently, the correlation between short-term and quaternary uplift
 225 shows correlation coefficients of 0.74 and 0.76 for the areas with and without recent ac-
 226 tive faulting, respectively. While the separation might sound ad hoc at first, we believe the
 227 spatial coherence of these two groups of points is the signature of an along-strike variable
 228 behavior of the forearc subjected to interseismic loading.

229 For a given interseismic uplift rate, estimates of long term coastal uplift are consis-
 230 tently slower in areas that have experienced significant quaternary faulting. We fit both
 231 data sets separately using a Bayesian regression technique (see supp. mat.) and infer the
 232 distribution of plausible linear relationships between quaternary and short-term uplift. In
 233 both cases, we observe that inferred quaternary uplift rates correspond to 4 to 8% of our
 234 estimated interseismic rates in northern Chile. In addition, it appears that, while a zero-
 235 valued interseismic uplift rate corresponds to a zero-valued quaternary uplift rate for the
 236 regions affected by recent active faulting, this is not the case for regions without identi-
 237 fied active faulting. In the later case, a zero-valued interseismic uplift rate corresponds to
 238 a 0.1 ± 0.05 mm/yr quaternary uplift rate.

239 **4 Discussion: Internal anelastic deformation and the role of faulting**

240 We interpret the observed correlation(s) between million-year time scale uplift and
 241 present-day interseismic uplift as the long-term signature of interseismic deformation.
 242 We explain the distribution of short vs. long term uplift rates as the combination of three
 243 overlapping mechanisms. First, a broad regional scale uplift of 0.1 ± 0.05 mm/yr affects
 244 the whole coastline in northern Chile. Second, 4 to 8% of the interseismic uplift is not
 245 balanced by co- and post-seismic slip related to large megathrust earthquakes. Third, a
 246 systematic offset of about 0.1 mm/yr is observed between regions that have and have not
 247 experienced active faulting over the quaternary. Our interpretation therefore suggests that
 248 interseismic and coseismic deformations do not cancel each other over many cycles and
 249 that approximately 4 to 8% of the vertical interseismic uplift is permanent. While the im-
 250 print of fault locking on the forearc morphology has been previously hypothesized based
 251 on the trench parallel gravity anomaly [e.g. *Song and Simons*, 2003; *Wells et al.*, 2003], we
 252 provide here a quantitative estimate of the amount of permanent deformation for northern
 253 Chile.

254 Then, two hypotheses could explain the systematic difference of uplift rates in re-
255 gions with and without recent (neogene to quaternary) active faulting. First, slip along
256 shallow structures within the forearc may have dissipated part of the energy that should
257 have transferred into gravitational potential energy (i.e. topography), hence a lesser effi-
258 ciency at building topography. Second, the presence of such pervasive faulting could im-
259 ply variations in forearc material properties along the coast. These hypotheses now need
260 to be tested against in-depth modeling.

261 Essential to our interpretation is that forearc materials must deform beyond their
262 elastic limit in between large megathrust earthquakes at characteristic interseismic strain
263 rates of 10^{-17} to 10^{-15} s⁻¹ (i.e. 0.1 to 10 nanostrain per year), and upper crustal temper-
264 atures and pressures below $\sim 400^\circ\text{C}$ and ~ 200 MPa, respectively. Among possible candi-
265 date mechanisms, pressure solution processes enable viscous-like creep under low de-
266 viatoric stress and temperature, particularly in shallow porous rock units (e.g., sedimen-
267 tary layers) [Niemeijer *et al.*, 2002]. Under greater deviatoric stresses, low-porosity rocks
268 are more likely to deform inelastically through brittle creep – the macroscopic manifesta-
269 tion of grain-scale cracks that nucleate from pre-existing defects and grow sub-critically
270 through stress- and temperature-activated chemical processes [Paterson and Wong, 2005;
271 Brantut *et al.*, 2013].

272 We illustrate the possible behavior of the forearc through the example of quartz, a
273 material typical of a continental crust, which can deform visco-plastically through dislo-
274 cation glide at low temperature [Bhat *et al.*, 2011]. Under a compressive load, such mate-
275 rial behaves elastically at low strain regime and starts to deviate from elasticity at higher
276 strain and the style of deformation (i.e. elastic or inelastic) depends on the ambient strain
277 rate (Fig. 5). For strain rates on the order of those expected for the interseismic period
278 within the forearc (i.e. 10^{-17} to 10^{-15} s⁻¹), quartz behaves elastically for stresses lower
279 than about 100 MPa, with inelastic strain increasing at higher stresses. Such a visco-plastic
280 material would allow for interseismic loading up to absolute stresses on the order of those
281 of a typical stress drop (i.e. less than 100MPa), while accumulating permanent strain. In
282 addition, because quartz tends to be elastic at high strain rates, we can assume the forearc
283 would behave elastically during the coseismic (and maybe immediate post-seismic) period.

284 Our interpretation focuses only on the behavior of the upper crust of the overrid-
285 ing plate, and thus does not account for the viscous rheology of the upper mantle over

286 the earthquake cycle [Wang *et al.*, 2012], which may affect surface strain rates and es-
287 timates of coupling along the megathrust [Li *et al.*, 2015]. Also, we do not account for
288 other potential transient deformation events. Post-seismic stress relaxation following large
289 earthquakes, for instance, occurs through aseismic slip [e.g. Hsu *et al.*, 2006] and viscous
290 flow within the upper-mantle [e.g. Trubienko *et al.*, 2013]. The influence of mantle relax-
291 ation considering a crust that can deform inelastically has not been extensively modeled
292 yet. That said, we believe that accounting for viscous processes in the mantle and afterslip
293 would not modify our conclusions based purely on the comparison of surface deformation.
294 First, given the geological time scales required to build topography, we can consider large
295 earthquakes and their corresponding afterslip as single slip events. Second, post-seismic
296 viscous flow within the upper-mantle following a large earthquake puts the forearc in com-
297 pression for a period of time that depends on the characteristic time involved [Sun *et al.*,
298 2014]. If that characteristic time is small with respect to large megathrust cycles, it should
299 be negligible.

300 Other examples worldwide do not necessarily compare simply with the behavior we
301 observe in northern Chile. For instance, in Japan, the coastline of Tohoku subsides both
302 during large earthquakes [e.g. Simons *et al.*, 2011] and during the interseismic period [e.g.
303 Hu *et al.*, 2016]. Only during the post-seismic period, a selection of GNSS-derived ground
304 velocities indicates uplift of the forearc. Although all phases of the earthquake cycle have
305 been recorded in this region, the summation of observed coseismic, modeled postseismic
306 and observed interseismic motion cannot explain the long-term uplift observed in the area.
307 Visco-elastic models of interseismic motion allow for a potential reversal of rates of ver-
308 tical motion during the interseismic period [Trubienko *et al.*, 2013; Hashima and Sato,
309 2017], which could reconcile long and short term observations in this part of Japan. In
310 northern Chile, the only available surface deformation measurements are during the inter-
311 seismic period and for moderate sized earthquakes and their respective post-seismic se-
312 quences. Considering the example of Tohoku, surface displacements related to the largest
313 possible event (and its postseismic displacements) along the megathrust completely over-
314 ride signals from other smaller magnitude events. Therefore, a complete budget of vertical
315 displacements is impossible here and we can only hypothesize based on the observed cor-
316 relation. Systematic mapping of short- and long-term vertical motion of subduction fore-
317 arcs is therefore required to determine globally how much of plate convergence actually
318 ends up in permanent deformation of subduction forearcs worldwide in order to constrain

319 future geodynamical models attempting at bridging time scales, from seconds to millions
320 of years.

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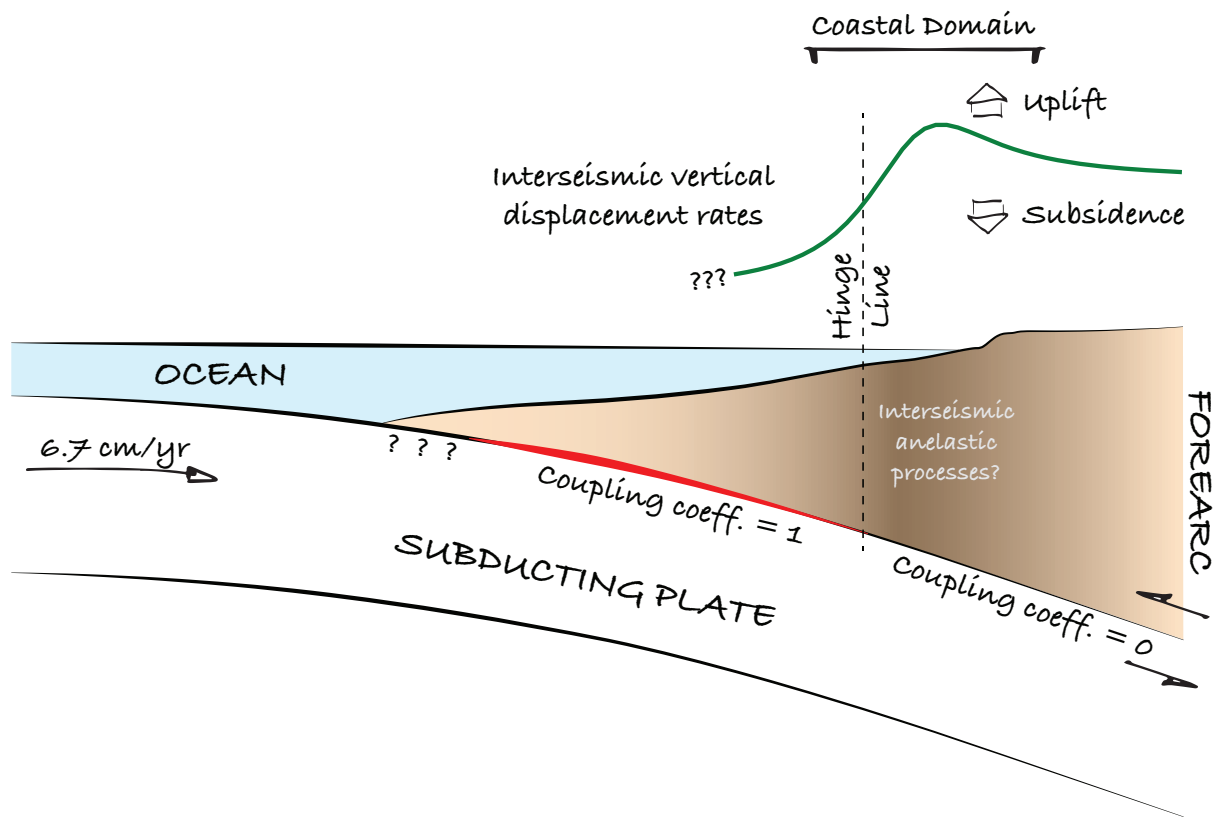
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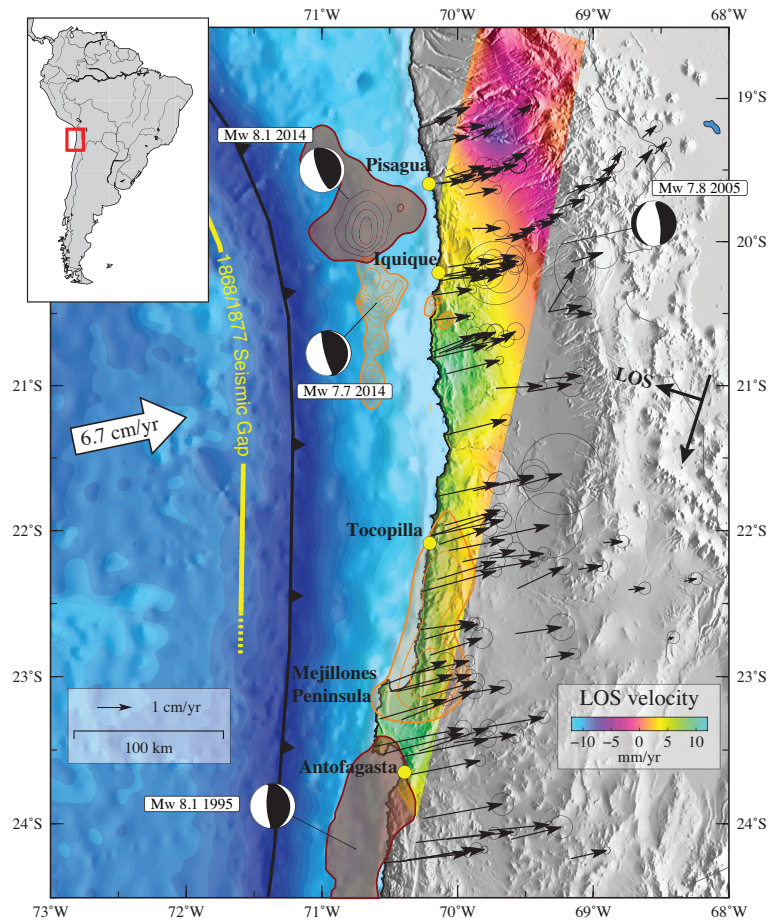
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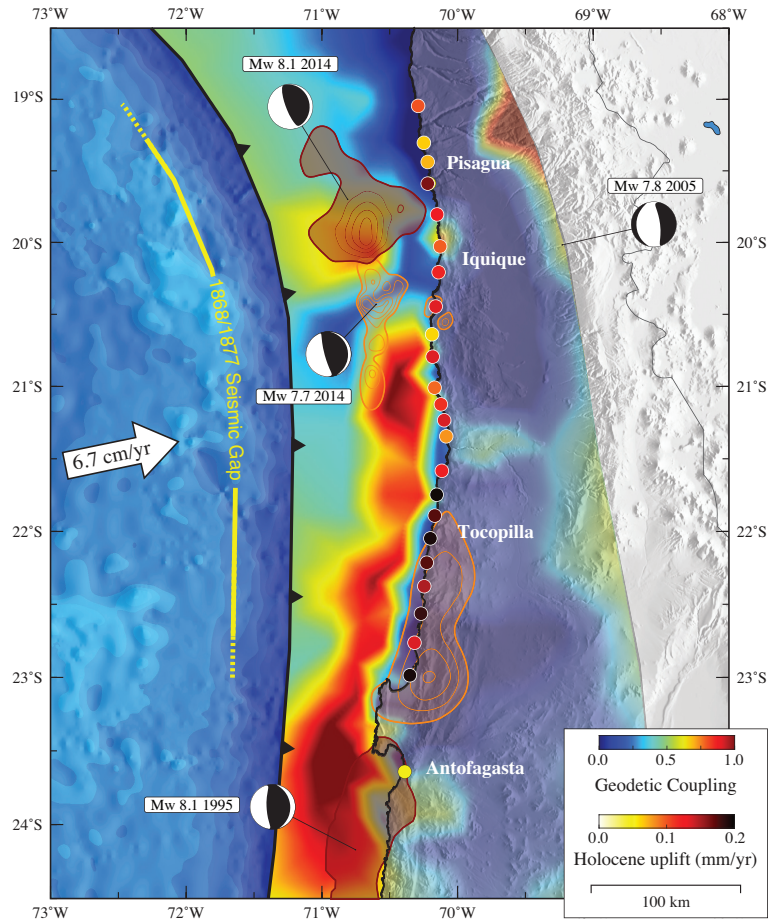
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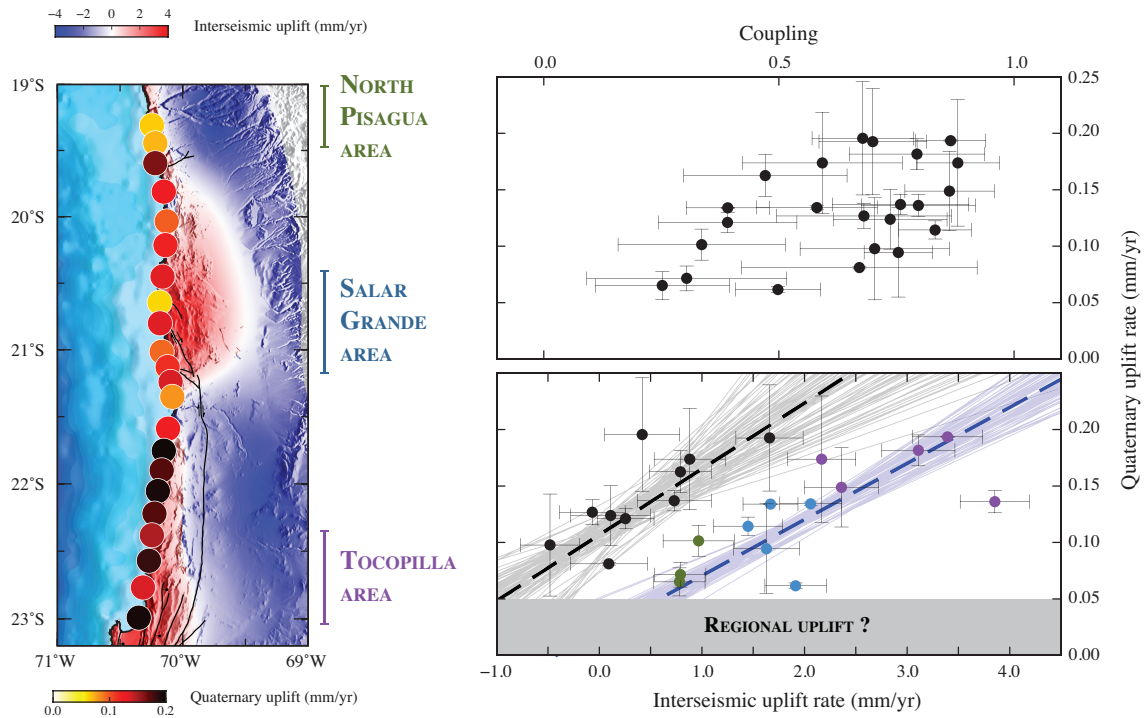
501 **Figure 1. Simplified seismotectonic setting of the subduction forearc** – An oceanic plate slides under-
 502 neath the subduction forearc. The interface is divided between a locked section and a creeping, unlocked
 503 section. The coupling coefficient is the parameter describing the apparent slip deficit normalized by the long-
 504 term plate convergence rate. A coupling coefficient of 0 indicates slip at plate rate along the interface while
 505 a coupling coefficient of 1 indicates full kinematic locking. The coastal domain is the region proximal to the
 506 coastline that is submitted to changes in sea level. During the interseismic period, uplift is maximum at the
 507 hinge line, which corresponds to the surface projection of the transition between the locked and unlocked
 508 sections of the interface at depth.



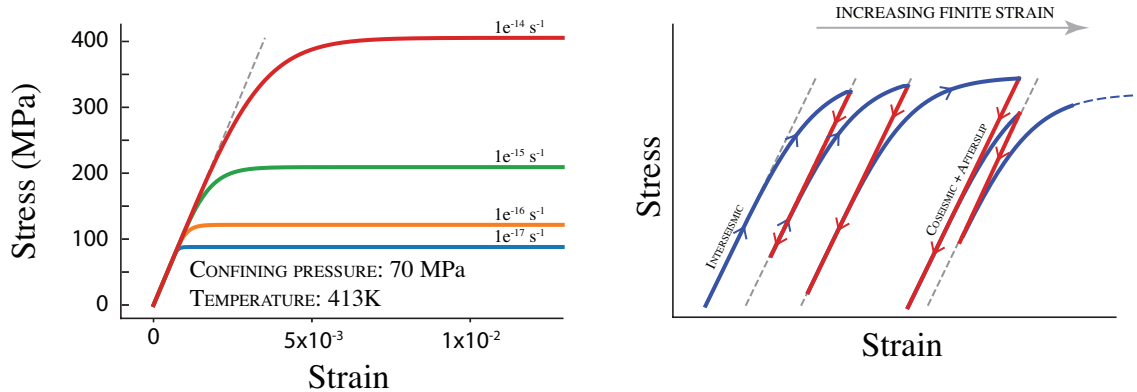
509 **Figure 2.** Tectonic setting and data set – Maps of the northern Chilean subduction zone. Arrows on the
 510 top map are GNSS-derived velocities from *Metois et al.* [2016] for the interseismic period preceding the
 511 2014 earthquake. Color indicates ground velocity in the direction of the satellite line-of-sight for Envisat
 512 acquisitions along track 96 [*Jolivet and Simons, 2018*]. Dark red contour lines are 2 m slip contours for earth-
 513 quakes of magnitude larger than 8, including the Antofagasta, Mw 8.1, 1995 event [*Pritchard et al., 2002*]
 514 and the Iquique, Mw 8.1, 2014 event [*Duputel et al., 2015*]. Orange contour lines are 0.5 m slip contours
 515 for earthquakes of magnitude between 7 and 8, including the 2007, Mw 7.7, Tocopilla event [*Béjar-Pizarro*
 516 *et al., 2013*] and the 2014, Mw 7.7, Iquique’s biggest aftershock [*Duputel et al., 2015*]. Focal mechanisms for
 517 earthquakes larger than magnitude 7.5 since 1995 are from the Global CMT project [*Ekström et al., 2012*].
 518 Topography is from SRTM [*Farr and Kobrick, 2000*].



519 **Figure 3.** Megathrust coupling and slip distributions – Mean of the posterior PDF of the coupling coeffi-
 520 cient along the megathrust. Color indicates whether the fault is kinematically coupled (red) or creeping at the
 521 plate rate (blue). Colored circles are the inferred quaternary uplift rates from *Melnick* [2016]. Earthquake slip
 522 distributions and moment tensors are similar as in Figure 2



523 **Figure 4.** Comparison between short term interseismic fault behavior and quaternary coastal uplift – Map
 524 shows quaternary coastal uplift rate [Melnick, 2016] (colored circles) and interseismic uplift rates predicted
 525 from our model (background color). (Top right) Quaternary coastal uplift as a function of coupling coefficient
 526 offshore, 50 km from the coast in a direction parallel to relative plate motion. (Bottom right) Quaternary
 527 coastal uplift as a function of interseismic uplift with our interpretation of the relationship between interseis-
 528 mic and quaternary uplift. Colored dots indicate locations where quaternary active faulting has been identified
 529 [please refer to the map for the location of these colored points; Allmendinger and González, 2010]. Thick
 530 blue and black lines are the linear trends predicted from the Bayesian regression for the region where quater-
 531 nary active faulting has been identified and not identified, respectively. Thin lines show a hundred realizations
 532 of the posterior PDF derived by the Bayesian regression, reflecting the range of possible trends for each group
 533 of points. See supplementary materials for a similar plot without interpretation, for a figure highlighting the
 534 position of the various points on a map and for a description of the regression applied.



535 **Figure 5.** Schematic model for the accumulation of permanent strain throughout the earthquake cycle –
 536 (Left) Stress-strain relationship for the visco-plastic low temperature deformation of quartz at 70 MPa and
 537 413 K, conditions corresponding to a depth of about 7 km. As the visco-plastic response of quartz is strain
 538 rate dependent, we show the corresponding behavior for 4 strain rates corresponding to a range of interseismic
 539 strain rates. (Right) Schematic behavior of a visco-plastic material to the cycles of loading and unloading
 540 imposed by the succession of interseismic and coseismic phases on the megathrust. During the interseismic
 541 phase, stress increases in the forearc, at a strain rate imposed by the slip rate on the megathrust, hence the
 542 variable irrecoverable strain that builds up laterally. During the coseismic phase, stress drops to a lower level,
 543 leaving persistent strain within the forearc.