

Strain budget of the Ecuador-Colombia subduction zone: a stochastic view

B. Gombert^{a,*}, Z. Duputel^a, R. Jolivet^b, M. Simons^c, J. Jiang^d, C. Liang^e, E. Fielding^e, L. Rivera^a

^a*Institut de Physique du Globe de Strasbourg, UMR7516, Université de Strasbourg, EOST/CNRS*

^b*Laboratoire de géologie, Département de Géosciences, École Normale Supérieure, PSL Research University, CNRS UMR 8538, Paris, France*

^c*Seismological Laboratory, Geological and Planetary Sciences, California Institute of Technology, Pasadena, California, USA*

^d*Institute of Geophysics and Planetary Physics, University of California, San Diego, La Jolla, California, USA*

^e*Jet Propulsion Laboratory, California Institute of Technology, Pasadena, California, USA*

Abstract

The 2016 Pedernales earthquake ($M_W=7.8$) ruptured a portion of the Colombia-Ecuador subduction interface where several large historical earthquakes have been documented since the great 1906 earthquake ($M=8.6$). Considering all significant ruptures that occurred in the region, it has been suggested that the cumulative moment generated co-seismically along this part of the subduction over the last century exceeds the moment deficit accumulated inter-seismically since 1906. Such an excess challenges simple models with earthquakes resetting the elastic strain accumulated inter-seismically in locked asperities. These inferences are however associated with large uncertainties that are generally unknown. The impact of spatial smoothing constraints on co-seismic and inter-seismic models also prevents any robust assessment of the strain budget. We propose a Bayesian kinematic slip model of the 2016 Pedernales earthquake using the most comprehensive dataset to date including InSAR and

*Corresponding author

Email address: gombert@unistra.fr (B. Gombert)

GPS offsets, tsunami waveforms, and kinematic records from high-rate GPS and strong-motions. In addition, we use inter-seismic geodetic velocities to produce a probabilistic inter-seismic coupling model of the subduction interface. Our stochastic co-seismic and inter-seismic solutions include the ensemble of all plausible models consistent with our prior information and that fit the observations within uncertainties. The analysis of these model ensembles indicates that an excess of co-seismic moment during the 1906 - 2016 period is likely in Central Ecuador only if we assume that 1942 and 2016 earthquakes are collocated. If this assumption is relaxed, we show that this conclusion no longer holds given uncertainties in co- and inter-seismic processes. The comparison of 1942 and 2016 teleseismic records reveals large uncertainties in the location of the 1942 event, hampering our ability to draw strong conclusions on the unbalanced moment budget in the region. Our results also show a heterogeneous coupling of the subduction interface that coincides with two slip asperities in our co-seismic model for the 2016 Pedernales earthquake and with the location of historical ruptures in 1958, 1979 and 1998. The spatial variability in coupling and complexity in earthquake history suggest strong heterogeneities in frictional properties of the subduction megathrust.

Keywords: Ecuador-Colombia subduction zone, Strain budget, Bayesian inversion, Kinematic source model, Geodetic coupling model

1. Introduction

A long standing question is the existence of persistent fault segments remaining locked in the inter-seismic period and failing suddenly during earthquakes while the surround-

4 ing interface creeps continuously. This conceptual model predicts so-called “characteristic”
5 earthquakes repeatedly rupturing the same locked fault segments with either periodic, time-
6 predictable or slip-predictable behaviours (Shimazaki and Nakata, 1980; Schwartz and Cop-
7 persmith, 1984). This paradigm is contradicted by an increasing number of observations
8 showing that the same fault area can break entirely in a single large earthquake ($M_W > 8.5$)
9 but also in a series of smaller ruptures. A remarkable example of such behaviour is the
10 Colombia-Ecuador subduction zone that experienced a complex sequence of earthquakes
11 since the beginning of the 20th century (see Figure 1). In 1906, the great $M_W = 8.6$ earth-
12 quake ruptured a ~ 500 -km-long segment of the subduction interface (Gutenberg and Richter,
13 1949; Ye et al., 2016). Several decades later, the same area was re-ruptured by a series of
14 smaller $M_W \leq 8.2$ events in 1942, 1958, 1979 and 1998 (Kanamori and McNally, 1982; Beck
15 and Ruff, 1984; Mendoza and Dewey, 1984; Chlieh et al., 2014). In April 2016, the region
16 in the vicinity of the 1942 Ecuador event was again ruptured by the $M_W = 7.8$ Pedernales
17 earthquake (Ye et al., 2016; He et al., 2017; Nocquet et al., 2017; Yi et al., 2018). Such vari-
18 ability among successive ruptures is also observed in other regions like Japan and Sumatra
19 where recent $M_W \sim 9$ megathrust earthquakes ruptured large fault segments that previously
20 experienced a serie of smaller events (Simons et al., 2011; Lay, 2015).

21 In addition to such variability among successive ruptures, major earthquakes in the
22 Colombia-Ecuador subduction zone seem to be clustered in time. Specifically, it has been
23 recently suggested that the seismic moment of the 1942, 1958 and 1979 earthquakes exceeds
24 the deficit accumulated since 1906 and that the 2016 Pedernales event may be associated with

25 more fault slip than the deficit accumulated since the 1942 earthquake (Nocquet et al., 2017).
26 Similar observations are reported in other regions, for example in 1797 and 1833 earthquakes
27 in Sumatra (Sieh et al., 2008), 1812 and 1857 earthquakes in California (Jacoby et al., 1988;
28 Heaton, 1990), and for the 2003 $M_W=7.6$ and 2013 $M_W=7.8$ Scotia sea earthquakes (Vallée
29 and Satriano, 2014). Such spatial and temporal clustering can be caused by spatial vari-
30 ations of fault coupling associated with heterogeneous frictional properties (Kaneko et al.,
31 2010). Moreover, there can be fluctuations in the patterns of inter-seismic fault coupling
32 before large earthquakes (Perfettini and Avouac, 2004; Mavrommatis et al., 2014; Yokota
33 and Koketsu, 2015) or during the post-seismic response of nearby large earthquakes (Heki
34 and Mitsui, 2013; Melnick et al., 2017).

35 Although the existence of an anomalously large co-seismic slip associated with a su-
36 percycle behaviour is plausible, other studies suggest that the seismic moment of the 2016
37 Pedernales earthquake is actually consistent with the strain accumulated in the region since
38 the 1942 and 1906 earthquakes (e.g., Ye et al., 2016; Yoshimoto et al., 2017; Yi et al., 2018).
39 These contrasting statements partly results from the ill-posed nature of inter- and co-seismic
40 slip inversions used to evaluate the strain budget along the megathrust. Such inferences
41 are affected by the lack of resolution near the trench during the inter-seismic period but
42 also by non-physics-based smoothing constraints used to regularize slip inversions. In ad-
43 dition, inter- and co-seismic estimates usually do not incorporate rigorous uncertainties (or
44 very often, no uncertainty at all), which complicates a quantitative assessment of the overall
45 strain budget. Strain budget analyses also suffer from the lack of information about past

46 earthquakes (Yi et al., 2018). Incorrect considerations on the size and position of historical
47 events can strongly affect the conclusion on the strain state of the plate boundary.

48 We propose a probabilistic exploration of the Colombia-Ecuador earthquake sequence,
49 fully accounting for uncertainties, including measurement errors, modeling errors, but also
50 uncertainties in the location or magnitude of past events. Using a Bayesian framework,
51 we explore both the inter-seismic geodetic coupling of the subduction interface and the co-
52 seismic slip distribution of the $M_W=7.8$ Pedernales earthquake. These estimates do not rely
53 on any spatial smoothing and provide full posterior probability distributions describing the
54 ensemble of plausible models that fit the observations and are consistent with simple prior
55 constraints (e.g., slip positivity in the direction of convergence).

56 **2. Geodetic coupling**

57 *2.1. Stochastic inter-seismic modeling*

58 We first compute a stochastic model of geodetic coupling along the Ecuadorian subduc-
59 tion interface. We use inter-seismic GPS velocities computed by Chlieh et al. (2014) from
60 29 stations installed in Ecuador and Colombia. The fault geometry is based on a 3D surface
61 following the Slab1.0 interface and discretized in triangles (c.f., Fig. S1 in the electronic
62 supplements). Using a back-slip approach (Savage, 1983), we invert for the inter-seismic
63 slip rate along the direction of convergence between Nazca and North Andean Sliver (NAS)
64 plates at each of the triangle knots assuming a barycentric interpolation scheme within the
65 triangles. This approach avoids unphysical slip discontinuities associated with traditional
66 parameterizations based on sub-faults with piecewise constant slip (Ortega Culaciati, 2013).

67 In our Bayesian inversion framework, the solution is the posterior ensemble of all plausible
68 inter-seismic slip rate models ($\mathbf{m}_{\mathcal{I}}$) that fit the GPS data ($\mathbf{d}_{\mathcal{I}}$) and that are consistent
69 with our prior hypotheses. This solution does not rely on any smoothing regularization
70 and is based on a simple uniform prior for the inter-seismic slip-rate that writes $p(\mathbf{m}_{\mathcal{I}}) =$
71 $\mathcal{U}(-0.05 \cdot V_p, 1.05 \cdot V_p)^M$ where V_p is the plate rate and M is the number of triangle knots
72 (260 knots). We thus restrict our posterior PDF to models in which slip on the fault aligns
73 with the direction of plate motion. Following Bayes' theorem, the posterior PDF is given by

$$p(\mathbf{m}_{\mathcal{I}}|\mathbf{d}_{\mathcal{I}}) \propto p(\mathbf{m}_{\mathcal{I}}) \exp \left[-\frac{1}{2}(\mathbf{d}_{\mathcal{I}} - \mathbf{G}_{\mathcal{I}}\mathbf{m}_{\mathcal{I}})^T \mathbf{C}_{\mathcal{I}}^{-1}(\mathbf{d}_{\mathcal{I}} - \mathbf{G}_{\mathcal{I}}\mathbf{m}_{\mathcal{I}}) \right] \quad (1)$$

74 where $\mathbf{G}_{\mathcal{I}}$ is the Green's function matrix and $\mathbf{C}_{\mathcal{I}}$ is the misfit covariance matrix combining
75 observational errors and prediction uncertainties. Green's functions are computed for a semi-
76 infinite stratified elastic medium derived from regional velocity models shown in Fig. S2
77 (Béthoux et al., 2011; Vallee et al., 2013; Nocquet et al., 2017). We account for prediction
78 uncertainties due to inaccuracies in this layered model using the approach of Duputel et al.
79 (2012, 2014). The uncertainty on the elastic structure, presented as grey histograms in
80 Fig S2, is estimated by comparing previously published models in the region.

81 We sample the posterior PDF $p(\mathbf{m}_{\mathcal{I}}|\mathbf{d}_{\mathcal{I}})$ using AlTar, a parallel Markov Chain Monte
82 Carlo (MCMC) algorithm following the CATMIP algorithm (Minson et al., 2013). More
83 details on the application of AlTar to investigate inter-seismic deformations can be found in
84 Jolivet et al. (2015b) and Klein et al. (2017). The resulting posterior ensemble of slip-rate
85 models in eq. (1) is then converted into stochastic coupling maps ($\mathbf{m}_{\mathcal{C}}$) using $\mathbf{m}_{\mathcal{C}} = 1 - \mathbf{m}_{\mathcal{I}}/V_p$.

86 *2.2. Geodetic coupling results*

87 Using our Bayesian framework, we generate 160 000 models corresponding to the posterior
88 information on geodetic coupling given measured inter-seismic velocities. We find that this
89 number is large enough to converge toward the posterior probability density. Representing
90 the ensemble of posterior models is challenging for multidimensional problems such as those
91 addressed in this study. To represent an ensemble solution, a common choice is to compute
92 the posterior mean (i.e., the average of all model samples). The posterior mean coupling
93 model is shown in Fig. 1 and Fig. 2a along with the associated $2\text{-}\sigma$ posterior uncertainties
94 in Fig. 2b. The posterior median model available in Fig. S3 is very similar to the posterior
95 mean, confirming that most marginal PDFs are nearly Gaussians. The variability of the
96 model population composing the solution is shown in supplementary movie M1.

97 Several features in our solution can be observed in previously published geodetic coupling
98 models (e.g., Nocquet et al., 2014; Chlieh et al., 2014). In the South, there is a very clear
99 high-coupling area offshore the Manta peninsula. This region has been previously associated
100 with transient slow-slip events (Vallee et al., 2013; Nocquet et al., 2014). As shown in Fig. 2a
101 and Fig. 2c, this area is associated with small model uncertainties probably because a GPS
102 station is located on La Plata Island, right above the coupled asperity. This coupled patch is
103 bounded to the north by a low-coupling corridor that might have acted as a creeping barrier
104 for the 1906, 1942, 1998 and 2016 earthquakes (cf., Fig. 1; Chlieh et al., 2014).

105 North of Bahía de Caráquez, we infer multiple patches of high geodetic coupling. Other
106 coupled patches can be identified offshore of Bahía de Caráquez, North and South of Ped-

107 ernales, and far offshore Esmeraldas. To first order, such heterogeneity is consistent with
 108 the "unsmoothed" solution of (Chlieh et al., 2014). This is unsurprising since our model-
 109 ing approach is not affected by any spatial smoothing. The high coupling asperity directly
 110 offshore of Bahía de Caráquez probably ruptured individually during the 1998 $M_W=7.2$
 111 earthquake while the coupled areas closer to Pedernales could have failed during the 1942
 112 and 2016 earthquake (Fig. 1). On the other hand, the large region of high coupling between
 113 Esmeraldas and Cap Manglares could be involved in the 1958 and 1979 ruptures (cf., Fig. 1).

114 However, we observe larger model uncertainties in this northern part due the lack of
 115 offshore measurements (Fig. 2b). This is quite clear in Fig. 2d showing that marginal PDFs
 116 close to the trench are nearly uniform. To quantify the robustness of our coupling map, we
 117 calculate the information gain from prior to posterior marginal PDFs using the Kullback-
 118 Leibler divergence, defined as:

$$D_{KL}^i = \int p(m_c^i | \mathbf{d}_c) \log_2 \frac{p(m_c^i)}{p(m_c^i | \mathbf{d}_c)} dm_c^i \quad (2)$$

119 where m_c^i is the coupling sampled in i -th knot of the triangular mesh. The resulting map
 120 shown in Fig. S4, indicates how much information is gained from the data in different regions
 121 of the model. It illustrates the difficulty to infer coupling properties close to the trench using
 122 land-based geodetic data. Still, the information gain remains significant within 30-40 km of
 123 the coast, and even sometimes almost up to the trench (e.g., offshore of the Manta peninsula
 124 and between Esmeraldas and Cap Manglares). This suggests that aforementioned asperities
 125 are reliable features of our solution.

126 **3. Rupture process of the 2016 Pedernales earthquake**

127 *3.1. Data overview*

128 We use several geodetic datasets covering both near-field and far-field static displace-
129 ments (cf., Fig. 3a). We gather GPS data from 12 campaign stations, 14 permanent stations
130 with daily solutions (CGPS), and 8 high-rate stations (HRGPS). Static offsets from cam-
131 paign and permanent stations are provided by Nocquet et al. (2017). We estimate our own
132 static displacements from HRGPS by measuring co-seismic offsets from the position before
133 and after the event. We use $1\text{-}\sigma$ errors provided by Nocquet et al. (2017) for the campaign
134 and CGPS and estimate uncertainties for HRGPS offsets from the standard deviation mea-
135 sured in 20 seconds pre- and post-event time windows. Vertical components of campaign
136 GPS are not used in the inversion as they show large uncertainties. In addition, we use three
137 interferograms derived from ALOS-2 wide-swath descending acquisitions, from ALOS-2 strip
138 map descending acquisitions and from Sentinel 1 descending acquisitions (cf., Fig. 4). Un-
139 wrapped interferograms are downsampled using a quad-tree algorithm (cf., Fig. S5; Lohman
140 and Simons, 2005). We estimate uncertainties related to atmospheric noise by estimating
141 empirical covariance functions for each interferogram (Jolivet et al., 2012, 2015a). Estimated
142 parameters are summarized in Table S1 and covariance functions are available in Fig S6.

143 Three nearby DART buoys (Deep-ocean Assessment and Reporting of Tsunamis) recorded
144 the tsunami generated by this event. Unfortunately, the waveform recorded by the closest
145 station (D32067) is unusable for modeling because of multiple data gaps and contamina-
146 tion by seismic waves. We use tsunami waveforms recorded at DART stations D32413 and

147 D32411 (cf., Fig. 3b), as they provide important constraints on the up-dip part of the
148 rupture. To remove tidal signals and reduce high-frequency noise, we band-pass filter the
149 waveforms between 8 min and 3 hours using a third order Butterworth filter. We derive
150 observational uncertainties from standard deviations computed in 140 and 100 min windows
151 before the first arrivals respectively for buoys D32413 and D32411.

152 We also include near-field seismic waveforms recorded by 10 strong-motion accelerometers
153 and 8 HRGPS stations (c.f. Figs. 5 and 6). We integrate the accelerometric data twice and
154 downsample them to 1 sps to match the HRGPS sampling rate. Waveforms are bandpass
155 filtered between 0.015 Hz and 0.08 Hz, except for a few noisy records for which we increased
156 the lower corner frequency to 0.037 Hz (Table S2). Waveforms are inverted in a 150 s-long
157 time window starting from the origin time of the mainshock (23:58:36 UTC).

158 *3.2. Stochastic co-seismic modeling*

159 Our kinematic modeling of the 2016 Pedernales earthquake is based on a non-planar fault
160 geometry in which the dip varies from 10° to 27° between 10 and 50 km depth, following the
161 bending of the Slab1.0 model (cf., Fig. S7; Hayes et al., 2012). The fault is discretized in
162 15×15 km patches in which we sample static (\mathbf{m}_S) and kinematic (\mathbf{m}_K) model parameters.
163 The static model vector \mathbf{m}_S includes two components of static slip in each patch (i.e., the
164 final integrated slip) and extra nuisance parameters to account for InSAR orbital errors (i.e.,
165 3 parameters per interferogram to model a linear function of range and azimuth). The two
166 components of static slip are U_{\parallel} , aligned with the direction of convergence between Nazca
167 and NAS plates, and U_{\perp} , which is perpendicular to U_{\parallel} . The vector of kinematic parameters

168 $\mathbf{m}_{\mathcal{K}}$ includes rupture velocity and rise time in each patch, along with hypocenter coordinates
 169 (i.e., the point of rupture initiation). Each point on the fault is only allowed to rupture once
 170 during the earthquake and we prescribe a triangular slip velocity function.

171 Following the approach of Minson et al. (2013), we first solve the final static slip distribu-
 172 tion (i.e., $\mathbf{m}_{\mathcal{S}}$) given available static observations ($\mathbf{d}_{\mathcal{S}}$), i.e., InSAR, GPS offsets and tsunami
 173 data. Using AlTar, we thus sample the posterior distribution:

$$\begin{aligned}
 p(\mathbf{m}_{\mathcal{S}}|\mathbf{d}_{\mathcal{S}}) &\propto p(\mathbf{m}_{\mathcal{S}}) p(\mathbf{d}_{\mathcal{S}}|\mathbf{m}_{\mathcal{S}}) \\
 &\propto p(\mathbf{m}_{\mathcal{S}}) \exp \left[-\frac{1}{2}(\mathbf{d}_{\mathcal{S}} - \mathbf{G}_{\mathcal{S}}\mathbf{m}_{\mathcal{S}})^T \mathbf{C}_{\mathcal{S}}^{-1}(\mathbf{d}_{\mathcal{S}} - \mathbf{G}_{\mathcal{S}}\mathbf{m}_{\mathcal{S}}) \right]
 \end{aligned}
 \tag{3}$$

174 where $\mathbf{G}_{\mathcal{S}}$ is the matrix including Green’s functions that are computed using the same
 175 layered elastic medium than the one used for the inter-seismic coupling model (cf., section 2).
 176 Tsunami waveforms are simulated using COMCOT (Liu et al., 1998) assuming a time step of
 177 1 sec and a 30-arc second GEBCO (General Bathymetric Chart of the Oceans) bathymetry.
 178 (Weatherall et al., 2015). As in eq. (1), the misfit covariance $\mathbf{C}_{\mathcal{S}}$ describes observational errors
 179 and prediction uncertainties due to inaccuracies of the assumed elastic structure Duputel
 180 et al. (2012, 2014). As we want to promote a dominant thrust motion while allowing local
 181 variations of the slip direction, the prior PDF $p(\mathbf{m}_{\mathcal{S}})$ includes uniform prior $\mathcal{U}(-1 \text{ m}, 15 \text{ m})$
 182 along the direction of convergence (U_{\parallel}) and Gaussian prior $\mathcal{N}(0, 0.5 \text{ m})$ in the perpendicular
 183 direction (U_{\perp}).

In a second step, we address the full joint inversion problem by incorporating kinematic
 observations $\mathbf{d}_{\mathcal{K}}$. HRGPS and strong motion data provide information on kinematic pa-
 rameters $\mathbf{m}_{\mathcal{K}}$ and bring additional constraints on $\mathbf{m}_{\mathcal{S}}$. The posterior PDF is then given by

(Minson et al., 2013):

$$\begin{aligned}
 p(\mathbf{m}_S, \mathbf{m}_K | \mathbf{d}_S, \mathbf{d}_K) &\propto p(\mathbf{m}_K) p(\mathbf{m}_S | \mathbf{d}_S) p(\mathbf{d}_K | \mathbf{m}_S, \mathbf{m}_K) \\
 &\propto p(\mathbf{m}_K) p(\mathbf{m}_S | \mathbf{d}_S) \exp \left[-\frac{1}{2} (\mathbf{d}_K - \mathbf{g}_K(\mathbf{m}_S, \mathbf{m}_K))^T \mathbf{C}_K^{-1} (\mathbf{d}_K - \mathbf{g}_K(\mathbf{m}_S, \mathbf{m}_K)) \right]
 \end{aligned}
 \tag{4}$$

184 where $\mathbf{g}_K(\mathbf{m}_S, \mathbf{m}_K)$ is the (non-linear) forward predictions for HRGPS and strong motion
 185 waveforms that are based on the Herrmann (2013) implementation of the discrete wave-
 186 number method (Bouchon and Aki, 1977). As in eq. (3), \mathbf{C}_K is the misfit covariance de-
 187 scribing measurement errors and predictions uncertainties due to Earth model inaccuracies.
 188 The prior $p(\mathbf{m}_K)$ is a combination of uniform priors $\mathcal{U}(1\text{ s}, 12\text{ s})$ and $\mathcal{U}(1\text{ km/s}, 4\text{ km/s})$ for
 189 rise-time and rupture velocity and a Gaussian PDF $\mathcal{N}(\mathbf{x}_h, \sigma = 5\text{ km})$ for the hypocenter
 190 coordinates (\mathbf{x}_h).

191 3.3. Co-seismic modeling results

192 The Pedernales rupture is mainly unidirectional with a significant southward directivity
 193 (see posterior mean model in Fig. 7, cumulative slip snapshots in Fig. 8, and supplementary
 194 movie M2; Ye et al., 2016; Nocquet et al., 2017; Yi et al., 2018). The inverted hypocenter
 195 (0.31° N , -80.15° W , depth=19.6 km; indicated by the red star in Fig. 7), is consistent with
 196 estimates from the Instituto Geofísico de la Escuela Politécnica Nacional (0.35° N , -80.16° W ,
 197 depth=17.0 km; <http://www.igepn.edu.ec>). Our solution depicts two large slip asperities
 198 separated by ~ 500 km that coincides roughly with two high-coupling zones north and south
 199 of the equator in Fig. 1. The first asperity is located close to the epicenter and fails within
 200 15 s after the origin time (Fig. 8). The second slip asperity ruptures about 10 s later and
 201 contributes to more than 60% of the total seismic moment. The rupture directivity and the

202 location of the southernmost asperity, with slip up to 8 m below the coastline, can probably
203 explain the large damages that have been reported south of the city of Pedernales (Nocquet
204 et al., 2017).

205 Posterior model uncertainties indicate that we have good constraints on slip amplitude
206 through the fault plane (Fig. 7b and supplementary movie M3). Moreover, stochastic rupture
207 fronts presented in Fig. 7a show that rupture initiation times are well resolved in large
208 slip areas. There is however a tradeoff between rupture initiation times and rise times as
209 illustrated in Fig. 7c-d. This is because our seismic observations are mostly sensitive on
210 subfault centroid times rather than on rupture times and rise times, resulting in a negative
211 correlation with a -1 slope between the two later parameters.

212 The southward directivity is clearly visible on HRGPS and strong motion data that show
213 large ground motion amplitudes south of the rupture. This is well captured by our stochastic
214 model predictions (Figs. 5 and 6). Some discrepancies are visible in the late arrivals, which
215 are probably due to unaccounted 3D heterogeneities. Geodetic measurements provide good
216 constraints on the static slip pattern, with large static displacements observed above the
217 large slip asperity in the south. Our solution is able to predict GPS measurements (Fig. 3a)
218 and InSAR data, with small residuals for Sentinel and ALOS-2 data (Fig. 4). We notice
219 larger misfits for the ALOS-2 descending track, probably due to atmospheric noise since
220 this interferogram is associated with significant spatially-correlated observational noise (cf.,
221 Fig. S6). Our solution also provides satisfactory fit to tsunami waveforms despite their
222 relatively small amplitude (<1 cm, Fig. 3b). These tsunami observations are important since

223 they clearly show the absence of slip in the shallow portion of the fault (shallow slip would
 224 produce large amplitude waves arriving too early at DART stations). This is also reported by
 225 Ye et al. (2016) that conducted trial and error teleseismic inversions, progressively removing
 226 shallow rows of patches to match the onset of tsunami signals.

227 **4. Strain budget along the Colombia-Ecuador subduction zone**

228 The Colombia-Ecuador subduction zone provides an outstanding opportunity to study
 229 the behaviour of a megathrust fault over multiple earthquake cycles. As mentioned above,
 230 before the 2016 Pedernales earthquake, the subduction interface experienced a sequence of
 231 megathrust ruptures that started with a large $M_W=8.6$ event in 1906 followed by a series
 232 of smaller earthquakes in 1942, 1958, 1979 and 1998. Because these events seem to cluster
 233 in time, it has been suggested that strain released by most recent earthquakes exceeds the
 234 deformation that accumulated inter-seismically since 1906 (Nocquet et al., 2017; Yi et al.,
 235 2018).

236 The strain budget along the megathrust can be investigated by comparing the co-seismic
 237 moment generated by earthquakes with the moment deficit accumulated during previous
 238 inter-seismic periods. We define the moment deficit accumulated over an inter-seismic time-
 239 span T over an area A of a fault as:

$$M_0^{deficit} = T V_p \iint_A \mu(\mathbf{x}) m_C(\mathbf{x}) d\mathbf{x} \quad (5)$$

240 where V_p is the long-term convergence rate, $\mu(\mathbf{x})$ is the shear modulus along the subduction
 241 interface and $m_{\mathcal{I}}(\mathbf{x})$ is the coupling model introduced in section 2. Using such approach,

242 Nocquet et al. (2017) propose that the co-seismic moment of the 1942 and 2016 earthquakes
243 are much larger than the deficit accumulated since the 1906 earthquake (by a factor of 3 to
244 5 times for the 1942 event and 1.3 to 1.6 times for 2016). This seems also true for northern
245 segments and in particular for the 1958 earthquake that has a seismic moment 1.5 to 1.8
246 times larger than the moment deficit estimated from the modeling of geodetic coupling. As
247 discussed by Nocquet et al. (2017) and Yi et al. (2018), these estimates remain questionable
248 given uncertainties on co-seismic slip and inter-seismic coupling.

249 Hereafter, we use our stochastic co-seismic and inter-seismic solutions to fully account
250 for posterior uncertainties and address the strain budget probabilistically. We assume a
251 magnitude of $M_W=7.8 \pm 0.2$ for the 1942 earthquake (Swenson and Beck, 1996; Ye et al.,
252 2016). We compare the probability distributions of seismic moment generated by the 1942
253 and 2016 earthquakes with the moment deficit accumulated since 1906 (Fig. 9). Assuming
254 the two events are co-located, maximum a posteriori models indicate that the seismic moment
255 for the 1942 and 2016 events are larger than the accumulated deficit by a factor of 2.0 and 1.2,
256 respectively. Taken together for the 1906-2016 period, the moment generated co-seismically
257 is 1.3 times larger than the moment deficit accumulated inter-seismically. Those estimates
258 are subject to considerable uncertainties reflected by the overlap between the PDFs (Fig. 9).
259 Although this overlap is not negligible, there is a relatively small probability of about 5% to
260 have a moment deficit larger or equal than the cumulative seismic moment of the 1942 and
261 2016 earthquakes. In this scenario, an excess of co-seismic moment since 1906 is likely given
262 available observations.

263 This conclusion only holds if the 2016 rupture largely overlaps with the 1942 earth-
264 quake, whose location is still debated. In particular, Yi et al. (2018) suggests that the
265 1942 earthquake occurred at shallower depth than the 2016 rupture from the comparison
266 of macroseismic maps of 1942 and 1958 events (Swenson and Beck, 1996). Therefore we
267 test the alternative hypothesis suggesting that the 1942 earthquake occurred between lat
268 0.5°S - 0.5°N at a depth shallower than 40 km (Nocquet et al., 2017). Fig. 10a,b shows that
269 the negative moment balance no longer holds. In this case, the probability of having a deficit
270 equivalent or larger than the co-seismic moment is larger than 70%. Fig. 10c,d shows that
271 this remains true if we further restrain the location of the 1942 event to be located updip of
272 the 2016 earthquake (as proposed by Yi et al., 2018).

273 We conducted a similar analysis for the 1958 northern Ecuador earthquake, assuming a
274 magnitude $M_W=7.6\pm 0.2$ (according to Ye et al., 2016). Maximum a posteriori models in
275 Fig. S9b show that the seismic moment generated by the 1958 earthquake is quite similar
276 to the accumulated deficit between 1906 and 1958. This contradicts with Nocquet et al.
277 (2017) that estimated that the 1958 earthquake had a seismic moment exceeding by 50% to
278 180% the moment accumulated inter-seismically. In our case, we clearly see that the PDF
279 of the moment deficit falls within uncertainties of the 1958 co-seismic moment. As shown
280 in Fig. S9, this still holds if we assume different location for the 1958 earthquake, which
281 discards the negative moment balance issue reported for 1942 and 2016 earthquakes.

282 5. Discussion and Conclusion

283 We develop stochastic models of the inter-seismic slip-rate along the Colombia-Ecuador
284 subduction and of the 2016 Pedernales earthquake, which provide new constraints on uncer-
285 tainties of inter- and co-seismic slip processes. Our results are to first order consistent with
286 some previously published models (e.g., Nocquet et al., 2017; Chlieh et al., 2014). In partic-
287 ular, our coupling model is similar to the "unsmoothed" model of Chlieh et al. (2014) since
288 it is not affected by smoothing regularization. Despite large uncertainties due to the lack of
289 geodetic observations far offshore, our solution clearly depicts a heterogeneous coupling of
290 the subduction interface (cf., Fig. 2).

291 Such heterogeneity correlates with the spatial complexity of the 2016 earthquake revealed
292 by our co-seismic solution. Our results indicate a unidirectional rupture towards the South
293 with two large slip zones that coincide with two high-coupling asperities in the inter-seismic
294 solution (cf., Fig. 1). We evaluate the possibility that the seismic moment generated by
295 the 1942 and 2016 earthquakes is larger than the moment deficit accumulated since the
296 great 1906 earthquake (as suggested by Nocquet et al., 2017). Our analyses show that this
297 conclusion only holds if we assume that there is a large overlap between the 1942 and 2016
298 ruptures. If this particular assumption is loosened, results indicate that such an unbalanced
299 moment budget is no longer required by observations. North of the Pedernales rupture, we
300 also show that the seismic moment of the 1958 earthquake is not necessarily larger than the
301 deficit accumulated since 1906 given uncertainties in co- and inter-seismic processes. The
302 question therefore entirely lies within the accuracy of the location and extent of historical

303 earthquakes.

304 One of the previously mentioned argument favouring an overlap between 1942 and 2016
305 earthquakes comes from the analysis of teleseismic waveforms recorded at a similar location
306 for both events (see details in supplementary text T1). Ye et al. (2016) showed that 1942 and
307 2016 waveforms at the station DBN (De Bilt, Netherlands) present significant dissimilarities.
308 Fig. S10 shows that such discrepancies can be explained by differences in the hypocenter
309 location with the same slip distribution for both events (as previously suggested by Nocquet
310 et al., 2017). However, the shape of the observed teleseismic P-wave is mostly controlled
311 by the relative location between the hypocenter and the main slip asperities (i.e., by the
312 corresponding apparent moment-rate function). In fact, Fig. S10c shows that the 1942
313 DBN waveform could be explained equally well if we assume that the 1942 rupture occurred
314 updip of the 2016 earthquake as suggested by focal depth and macro-iseismic maps of
315 the 1942 event (Yi et al., 2018). In this scenario, there is a probability of $\sim 70\%$ to have
316 a balanced moment budget since 1906 (i.e., a moment deficit that is larger or equal to
317 the seismic moment of 1942 and 2016 events). On the contrary, if there is a large overlap
318 between both earthquakes, our results show that there is a 95% probability that the moment
319 generated by 1942 and 2016 ruptures is larger than the moment deficit accumulated since
320 1906. In this case such an unbalanced moment budget can possibly be explained by temporal
321 variations in strain accumulation, which have been observed for example before and after the
322 2011 $M_W=9.0$ Tohoku earthquake (e.g., Mavrommatis et al., 2014; Heki and Mitsui, 2013)
323 and after the 2010 Maule earthquake (Melnick et al., 2017; Loveless, 2017). Alternatively,

324 Nocquet et al. (2017) propose a "supercycle" model where the apparent excess of co-seismic
325 moment results from the fact that the 1906 and 1942 earthquakes did not release all of the
326 accumulated strain along the megathrust. This is consistent with the modeling of historic
327 tsunami records suggesting that the 1906 earthquake mainly ruptured the shallow part of
328 the subduction without involving much slip close to the 2016 Pedernales event (Yoshimoto
329 et al., 2017). However, these estimates might be biased by the poor sensitivity of tsunami
330 data to deep slip, which can explain the relatively low magnitude of their resulting model
331 ($M_W=8.4$). The fact that the surface wave magnitude $M_s=8.6$ is otherwise consistent with
332 M_W also suggests that the 1906 earthquake is not a typical "tsunami" earthquake and is
333 therefore probably not associated with a predominantly shallow rupture (Kanamori, 1972).

334 The complex behavior of the Colombia-Ecuador subduction can be related to the large
335 heterogeneity of our coupling solution, which suggests significant spatial variability of fault
336 friction properties (Fig. 1). As shown for example by Kaneko et al. (2010), such heterogeneity
337 can produce earthquakes of different sizes re-rupturing the same fault region at short time
338 intervals. Complex earthquake sequences may also be promoted by partial stress drop of past
339 events that produces significant stress heterogeneity along the fault (Cochard and Madariaga,
340 1996). The fact that large earthquakes (like the 1906 event) are rapidly followed by sequences
341 of smaller ruptures (e.g., in 1942, 1958, 1979, 1998 and 2016) can then be understood if static
342 stress drop of the smaller events is small compared to the increase of dynamic stresses at
343 rupture fronts (Heaton, 1990; Melgar and Hayes, 2017).

344 As instrumental observations accumulate, there is a growing record of large earthquakes

345 that break portions of faults that experienced previously documented large ruptures. These
346 earthquakes continuously provide new observations suggesting complex earthquake sequences
347 with substantial spatial and temporal variability among successive ruptures of the same fault
348 system. As shown here, the study of long-term earthquake sequences and the associated
349 strain budget still relies on many assumptions and are affected by large uncertainties. To ad-
350 dress the seismogenic behaviour of active faults, we need to quantify how large observational
351 and modelling uncertainties are and how much information we have gained in comparison
352 to our preconceptions. Inaccuracies on historical earthquakes size and position can be sub-
353 stantial and also need to be properly considered. Such quantitative analysis is essential to
354 understand how strain accumulates inter-seismically and is released by earthquakes, thereby
355 improving seismic hazard assessment along subduction zones.

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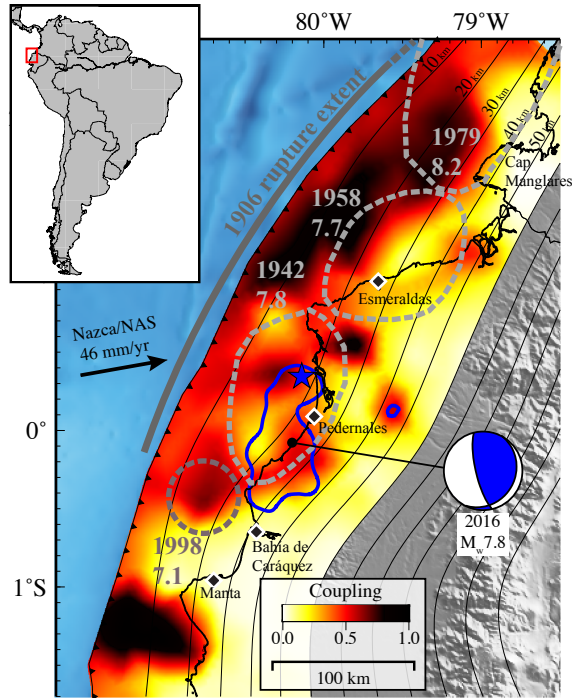


Figure 1: **Interseismic coupling and historical earthquakes.** The colour scale indicates the geodetic coupling of the subduction interface obtained from inter-seismic GPS velocities (cf. section 2). Blue line and blue star are respectively the 2 m isocontours of co-seismic slip and hypocenter obtained for the 2016 Pedernales earthquake (cf., section 3). Grey dashed lines show the approximate extent of the 1942, 1958, 1979, and 1998 events (Kanamori and McNally, 1982; Chlieh et al., 2014). The thick gray line shows the along-strike extent of the 1906 $M_W = 8.6$ earthquake. The focal mechanism of the 2016 Pedernales earthquake is presented in blue. Thin black lines are isocontours of the slab depth. The line with the adjacent black triangles shows the location of the trench. The black arrow illustrates the convergence direction of the Nazca plate toward the North Andean Silver plate (NAS, Chlieh et al., 2014).

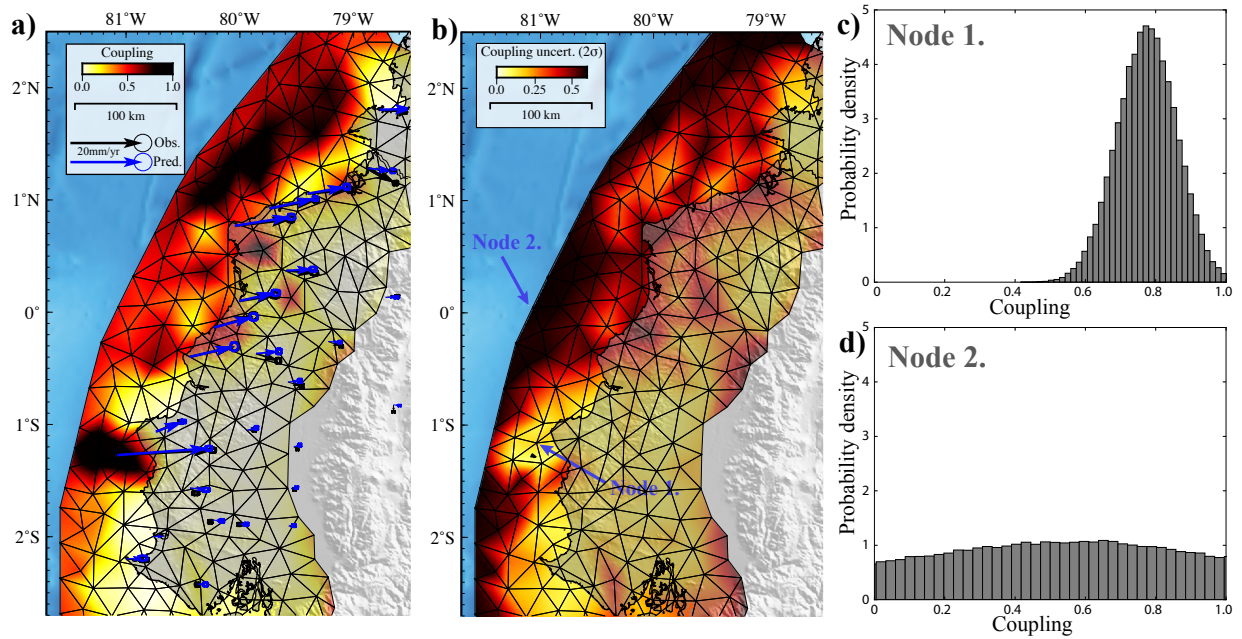


Figure 2: **Interseismic coupling of the Ecuadorian subduction margin.** a) Posterior mean coupling model. Thin black lines represent the fault parametrization. Coupling values are inverted at each triangle knot. Interseismic GPS displacement and model predictions are plotted as black and blue arrows, respectively. b) 2- σ uncertainties of the coupling model. c) et d) Marginal probability densities for the two nodes pointed out in b).

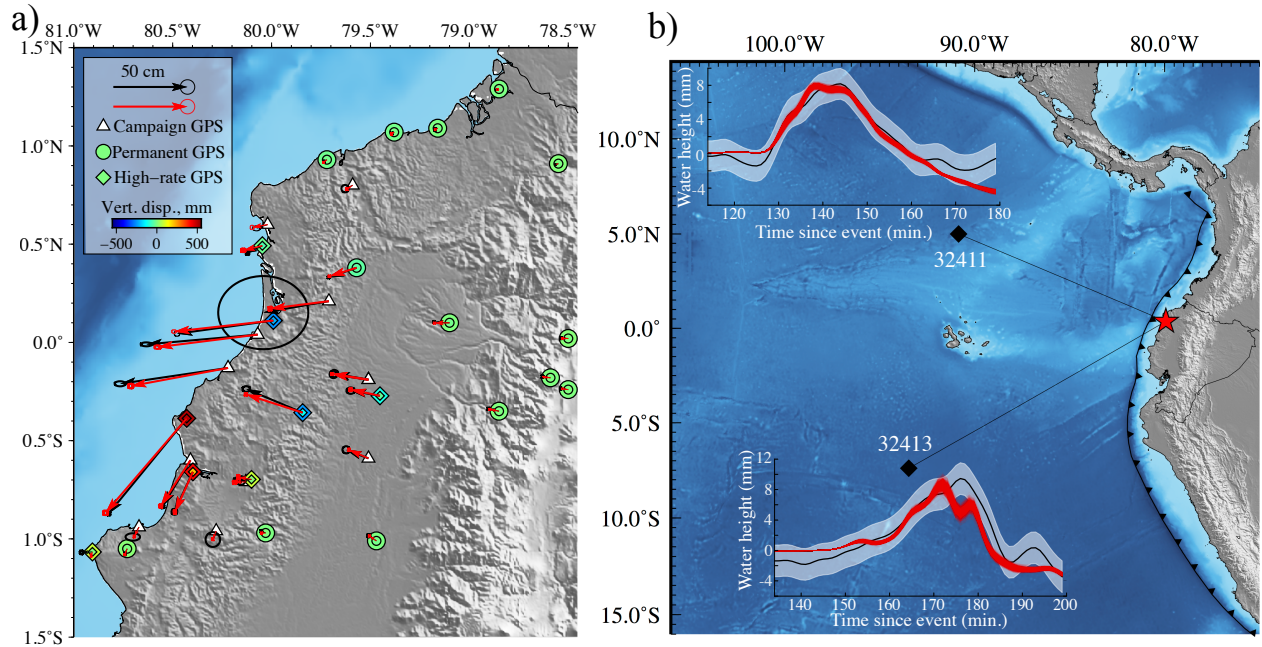


Figure 3: **GPS and tsunami observations used in this study.** **a)** GPS data and model predictions. Black and red arrows show observed and predicted GPS horizontal displacements along with their 95%-confidence ellipses (representing observational and prediction uncertainties, respectively). For the permanent and High-rate GPS, the symbol colour represents the vertical displacement. The outer symbol is the observation while the inner symbol is the mean model prediction. **b)** Observed and predicted tsunami waveforms. The red star defines the event epicenter while black diamonds are the locations of the two DART buoys that recorded the tsunami. For each of them, the amplitude of the first arrival is plotted as a thick black line. The surrounding shaded area marks the $2\text{-}\sigma$ confidence interval. Stochastic forward model predictions are plotted in red.

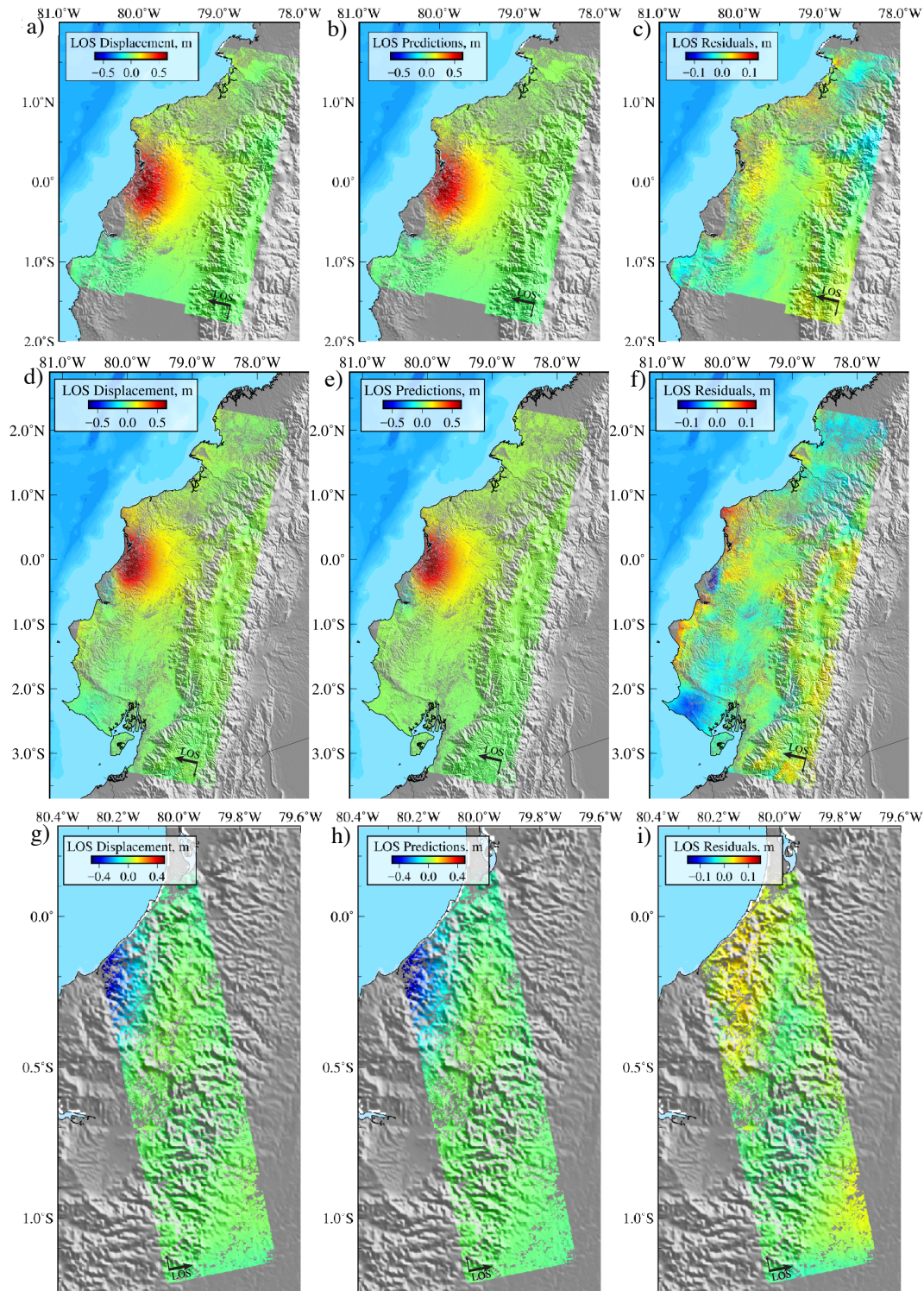


Figure 4: **Model performance for InSAR.** (a, d, g) InSAR observations. (b, e, h) Predictions for the posterior mean model. (c, f, i) Residuals of the Sentinel (top row), descending ALOS-2 (middle row), and ascending ALOS-2 (bottom row) interferograms. Decimated observations, predictions, and residuals are shown in Fig. S5

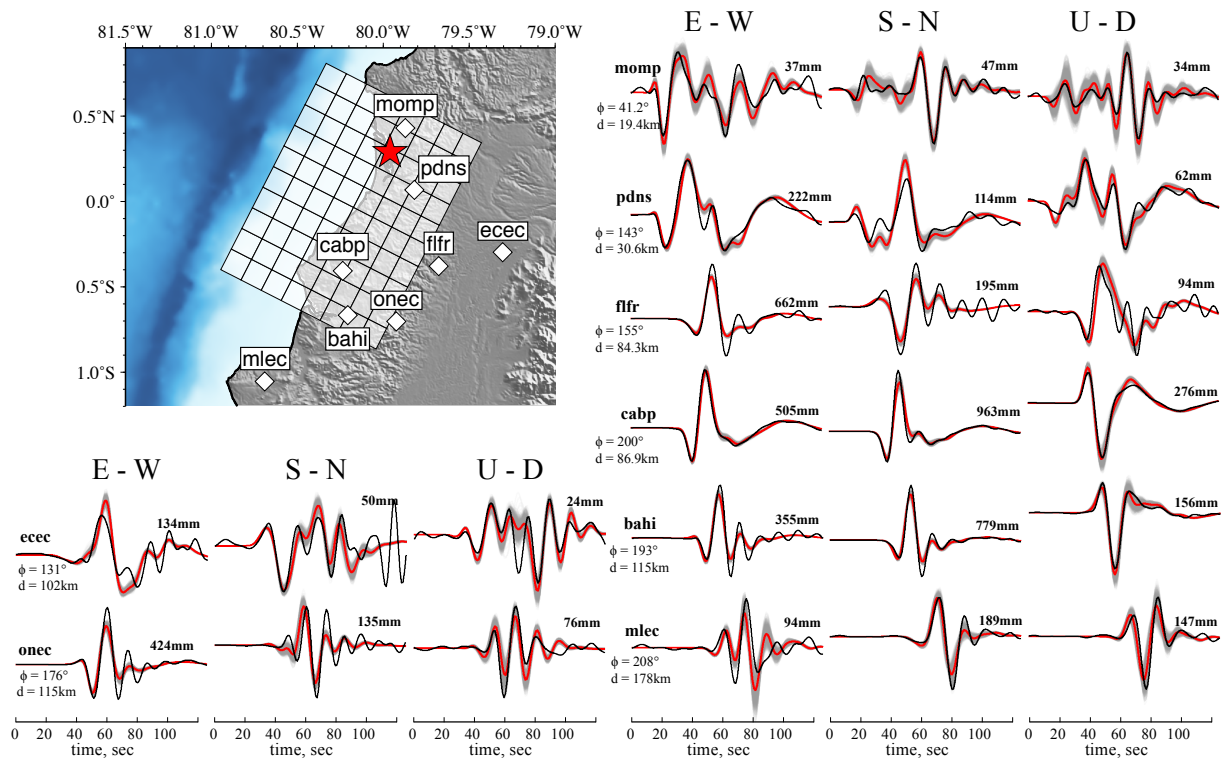


Figure 5: **High-rate GPS observations and model predictions.** The white diamonds on the top-left map indicate the position of the stations. The red star marks the inverted epicenter location. White rectangles are the fault parametrization. The East, North, and vertical components of each station are plotted around the map. For each waveform, the bold number indicates its maximum amplitude. The station azimuth Φ and distance d to the epicenter are also given. The black line is the recorded waveform. The gray lines are the stochastic predictions for our posterior model. The red line is the mean of the stochastic predictions.

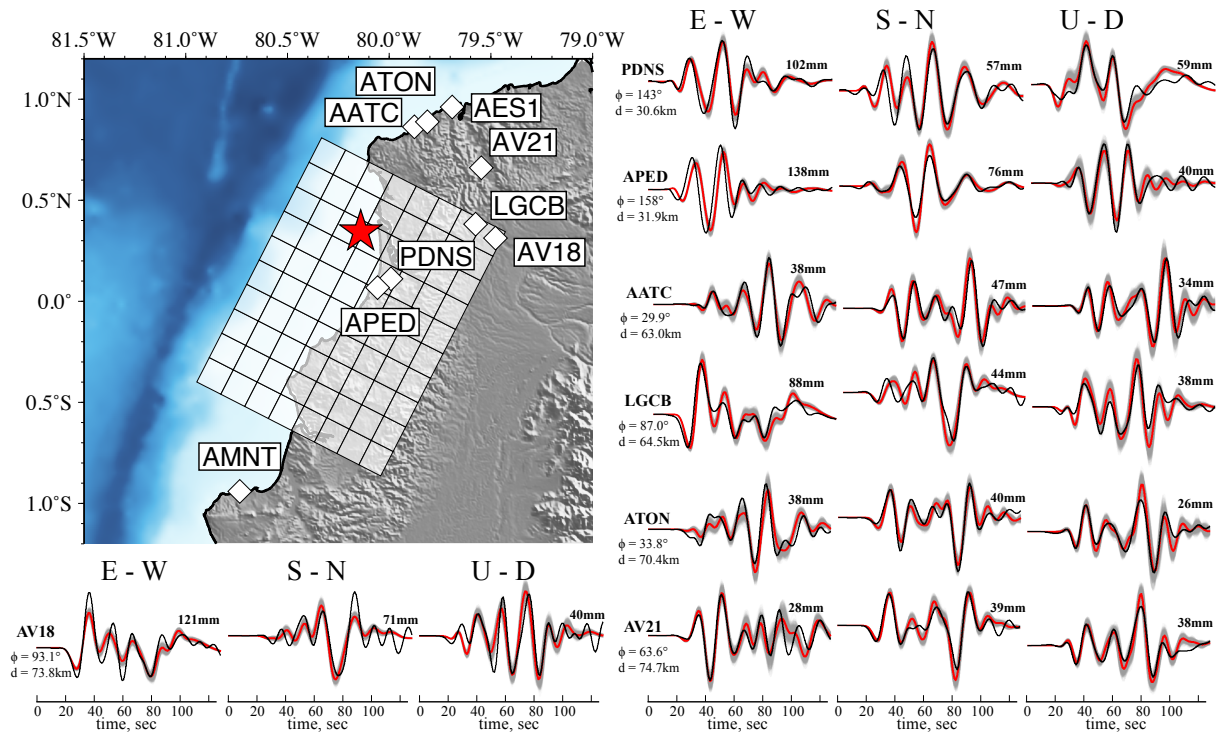


Figure 6: **Strong-motion observations and model predictions.** Same as Fig. 5. We show only the stations where three components are available. The remaining 5 waveforms from 3 additional stations and the associated model predictions are shown in Fig. S8

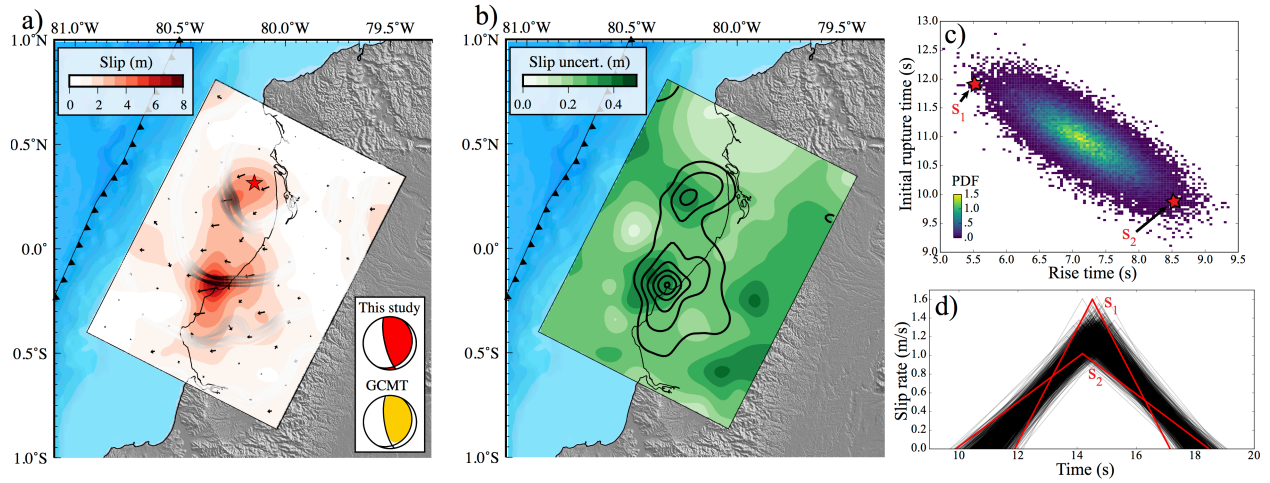


Figure 7: **Final co-seismic slip distribution.** **a)** The colour and arrows on the fault plane indicate the amplitude and direction of slip, respectively. Gray-scale lines are stochastic rupture fronts inferred from our model population plotted at 10s, 20s, and 30s. The darker the lines, the larger the slip at that location. The red star marks the hypocenter location. **b)** Slip uncertainty. The colour on the fault represents the absolute slip uncertainties. Black contour lines show the co-seismic slip every 1 m, starting from 2 m. **c)** Marginal probability distribution of rise time and initial rupture time in the first slip asperity (located close to the hypocenter). **d)** Posterior ensemble of source time functions at the same location of the fault. The source time functions labeled s_1 and s_2 in (d) correspond to rupture initiation times and rise times that are indicated with red stars in (c).

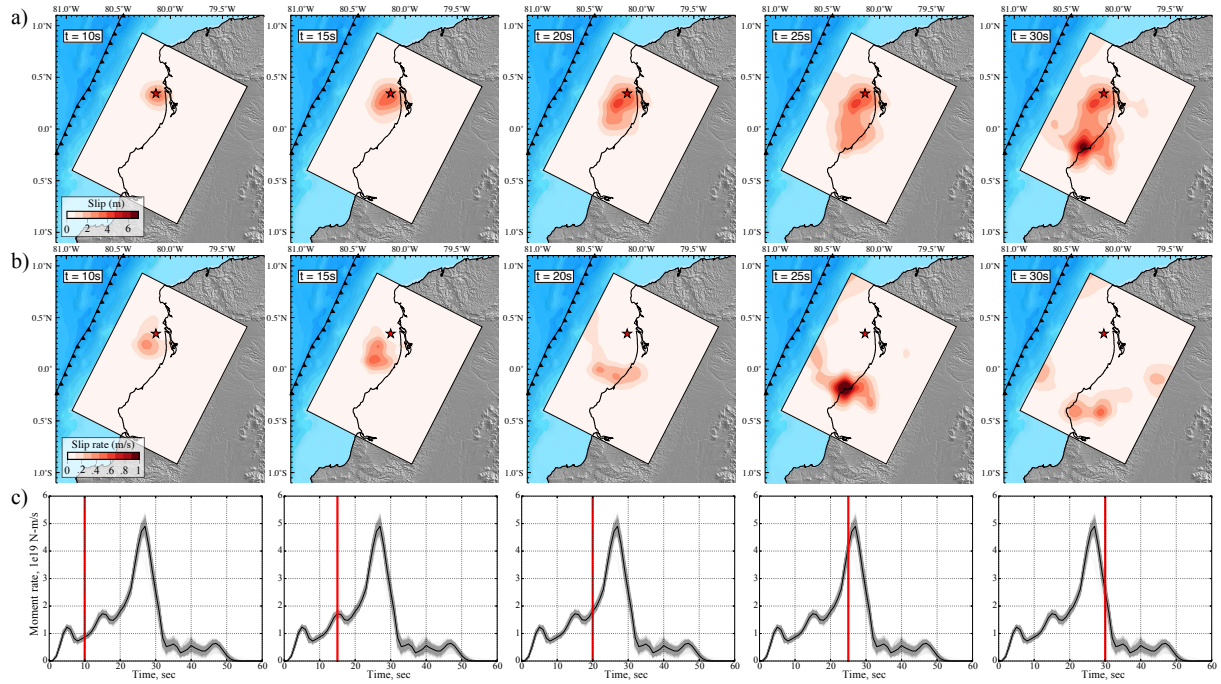


Figure 8: **Temporal evolution of co-seismic slip.** a) Cumulative slip on the fault 10 s, 15 s, 20 s, 25 s, and 30 s after the origin time. The red colour-scale indicates slip amplitude. The red star marks the epicenter location. b) Evolution of slip rate on the fault. c) Source time function (STF) of the event. Grey lines are stochastic STFs inferred from our model population while the black curve represents the posterior mean STF. Vertical red lines indicate the temporal position of each one of the snapshots

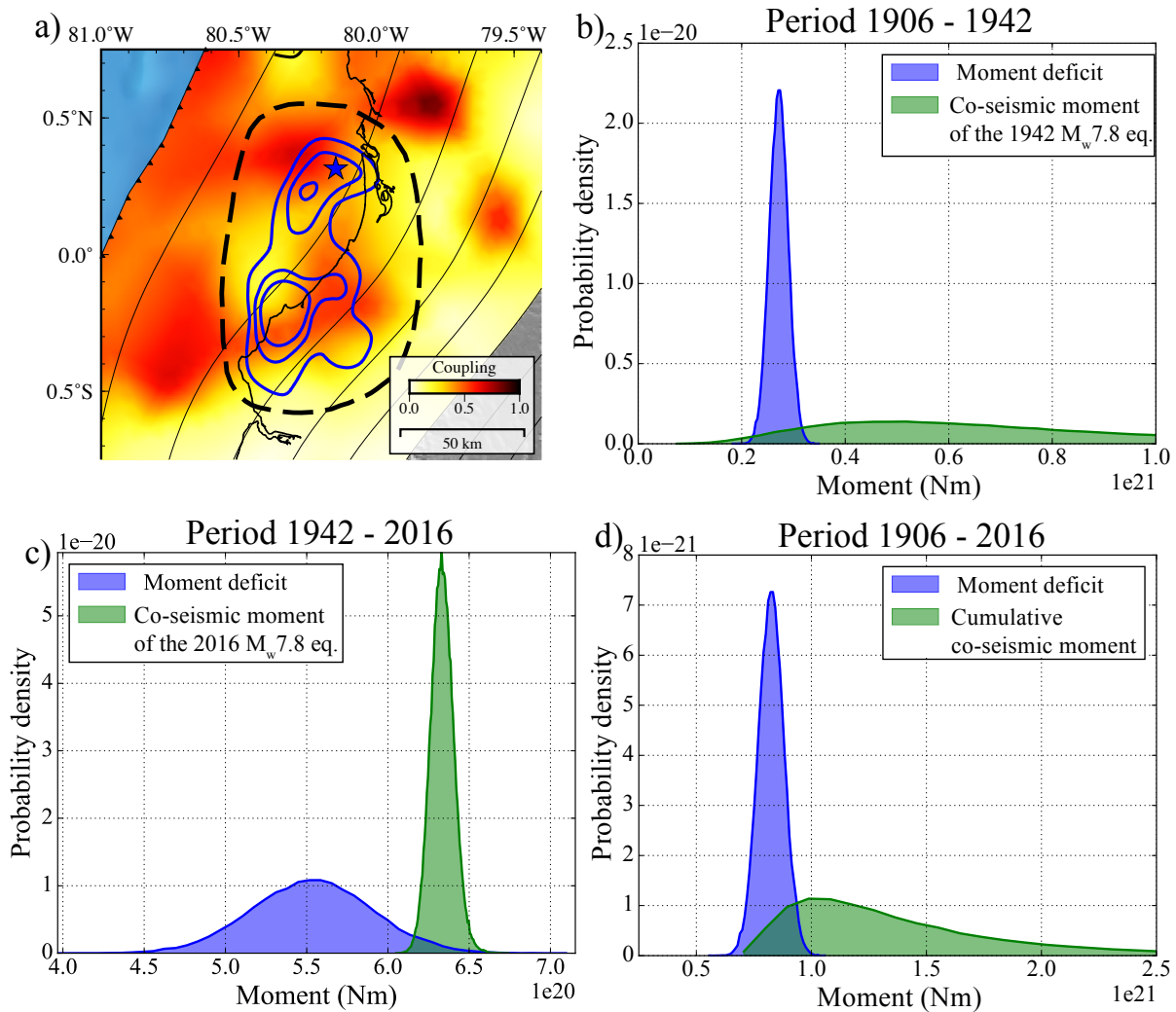


Figure 9: **Comparison of co-seismic moment and moment deficit.** a) The background colour represents the coupling posterior mean model. The blue stars shows the hypocentre location. Blue lines are the 2m, 3m, and 4m co-seismic slip isocontours. The black dashed line delimits the area where the co-seismic moment and moment deficit are computed. b) Probability densities of the co-seismic moment released by the 1942 earthquake and the moment deficit accumulated between 1906 and 1942 within the dashed ellipse shown in a). c) Probability densities of the co-seismic moment released by the 2016 earthquake and the moment deficit accumulated between 1942 and 2016. d) Probability densities of the co-seismic moment released by the sum of the 1942 and 2016 events, and of the moment deficit during the 1906 - 2016 period.

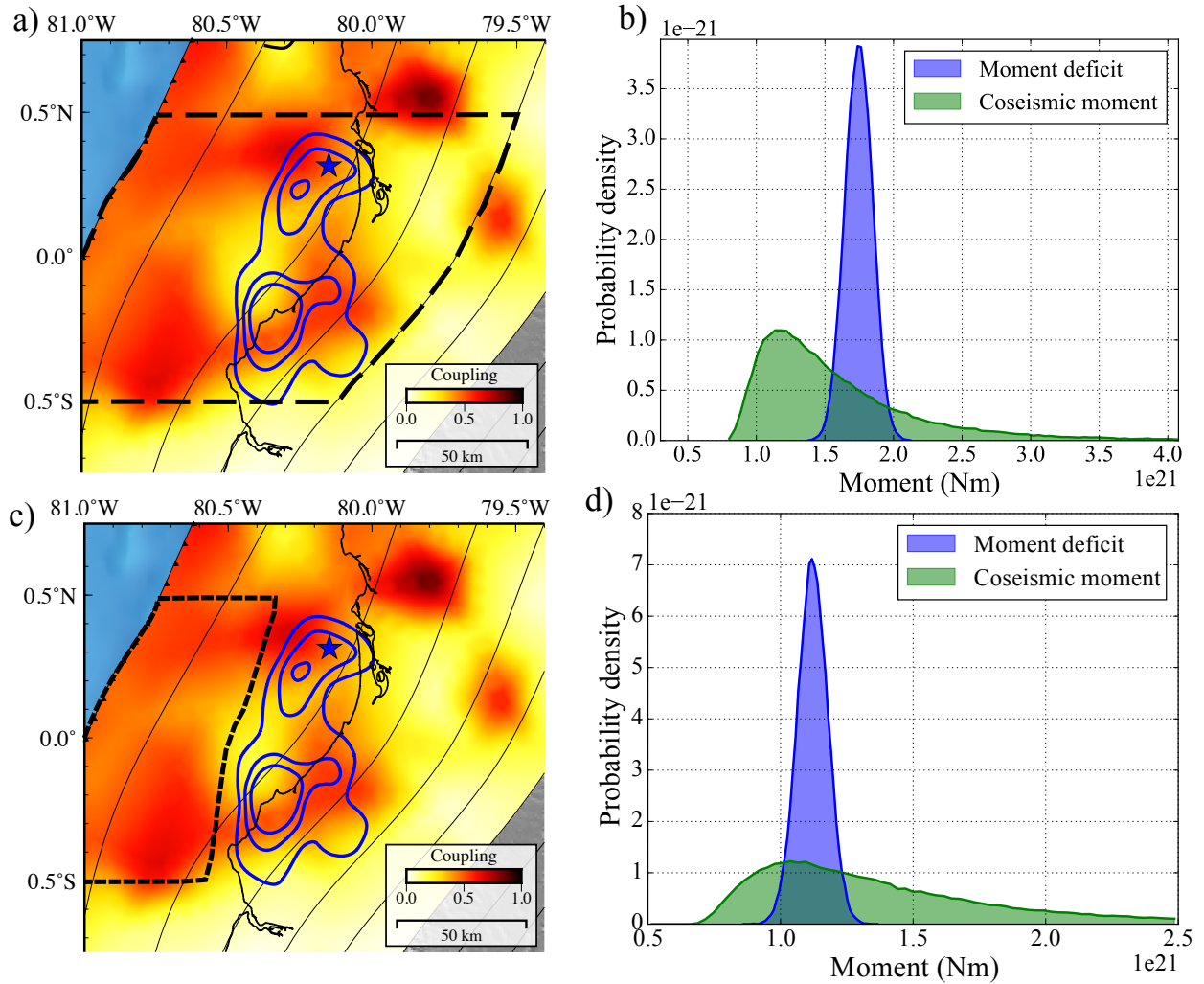


Figure 10: **Comparison of co-seismic moment and moment deficit considering the 1942 event happened at a different location.** **a)** Same as Figure 9a. The black dashed line delimits the area between 0.5°S and 0.5°N where the co-seismic moment and moment deficit are computed. **b)** Probability densities of the co-seismic moment and moment deficit in the 1906-2016 period. The co-seismic moment is the sum of the 1942 and 2016 events moment. **c)** Same as a), but the dashed black area shows where the 1942 earthquake could have been located. **d)** Probability densities of the co-seismic moment and moment deficit. The co-seismic moment is the sum of the 1942 and 2016 events moment. The moment deficit is the sum of the moment deficits computed in the updip section (shown in c)) for the 1942 - 2016 period and in the downdip section (ellipse in 9a) for the 1906 - 2016 period.

Supporting information for the main manuscript:

”Strain budget of the Ecuador-Colombia subduction zone: a stochastic view”

B. Gombert¹, Z. Duputel¹, R. Jolivet ², M. Simons³, J. Jiang⁴, C. Liang⁵, E. Fielding⁵, and L. Rivera¹

¹ Institut de Physique du Globe de Strasbourg, UMR7516, Université de Strasbourg, EOST/CNRS

² Laboratoire de géologie, Département de Géosciences, École Normale Supérieure, PSL Research University, CNRS UMR, Paris, France

³ Seismological Laboratory, Geological and Planetary Sciences, California Institute of Technology, Pasadena, California, USA

⁴ Institute of Geophysics and Planetary Physics, University of California, San Diego, La Jolla, California, USA

⁵ Jet Propulsion Laboratory, California Institute of Technology, Pasadena, California, USA

This electronic supplement is a collection of additional figures referenced in the main article. These figures were added to ensure the precision of the description of our method and results.

Supplementary text T1

Following Ye et al. (2014) and Nocquet et al. (2017), we compare waveforms of the 1942 earthquake recorded at the DBN station (De Bilt, Netherlands) with stochastic waveform predictions at the same station for the 2016 Pedernales slip distribution.

We compute displacement Green's functions for each subfault patch using the Kikuchi-Kanamori program (Kikuchi and Kanamori, 2003; Kikuchi, Masayuki and Kanamori, Hiroo, 1982). For comparison, we then convolve predicted stochastic waveforms with the instrumental response of the Galitzin seismometer that recorded the 1942 earthquake (pendulum and galvanometer periods $T_p=T_g=25$ s and gain factor $V_m=310$; Charlier and Van Gils, 1953).

In Fig. S10a, we first compare 1942 waveforms with predictions of the kinematic slip model (i.e., for the posterior distributions of slip, rise-times, rupture velocities) and hypocenter location obtained for the 2016 Pedernales earthquake. Model predictions show poor fit to the 1942 earthquake waveform. In Fig. S10b, we then compute predictions for the same kinematic slip distribution, but with a hypocenter location between the two slip asperities. With that hypocenter location, model predictions have a very good fit to the 1942 waveform. Finally, in Fig. S10c, we predict waveforms for a slip distribution on the megathrust interface, but updip of the actual 2016 rupture. Notice, that the dip is different due to the variation of the slab interface geometry with depth. We also correct the slip amplitude for the variation of shear modulus in our velocity model (cf., Fig. S2). Similarly to the previous case, the hypocenter is located between the two slip asperities. In this scenario, we are also able to explain the 1942 waveform. It illustrates that the teleseismic P-waveform is mostly sensitive to the relative location of the hypocenter and slip asperity rather than the absolute location of the earthquake.

Poles and zeros of the DBN Galitzin seismometer:

| | |
|-----------|--------------|
| CONSTANT | 2.399052e+02 |
| ZEROS | 3 |
| POLES | 4 |
| -0.251327 | 0.000000 |
| -0.251327 | 0.000000 |
| -0.251327 | 0.000000 |
| -0.251327 | 0.000000 |

Supplementary movie M1: Variability in the Ecuador-Colombia geodetic coupling solution The animation is made with 100 models randomly selected in the posterior population.

Supplementary movie M2: Temporal evolution of co-seismic slip of the 2016 Pedernales earthquake. **(left)** Posterior mean model of the cumulative slip. The bottom-right inset shows the stochastic source time function. **(right)** Incremental slip on the fault. The red star marks the inverted posterior mean hypocenter location.

Supplementary movie M3: Variability in the 2016 Pedernales earthquake co-seismic slip distribution solution The animation is made with 200 models randomly selected in the posterior population.

Table S1: InSAR observations used in this study.

| Satellite | Orbit | Acquisition dates | N° of data | Std. | Corr. length |
|------------------|--------------|--------------------------|-------------------|-------------|---------------------|
| ALOS-2 | ascending | 07/02/16 - 01/05/16 | 130 | 5.3 mm | 2.88 km |
| ALOS-2 | descending | 01/04/16 - 29/04/16 | 483 | 9.2 mm | 11.90 km |
| Sentinel-1A | descending | 12/04/16 - 24/04/16 | 380 | 5.0 mm | 15.0 km |

Table S2: Seismological data and filtering used in this study. We use a 4th order Butterworth bandpass filter.

| Station | Type | Filter corner frequencies | | |
|---------|---------------|---------------------------|------------------|------------------|
| | | East | North | Up |
| bahi | HRGPS | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz |
| cabp | HRPGS | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz |
| ecec | HRPGS | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz |
| fffr | HRPGS | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz |
| mlec | HRPGS | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz |
| momp | HRPGS | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz |
| onec | HRPGS | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz |
| pdns | HRPGS | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz |
| ISPT | Strong motion | N/A | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz |
| PDNS | Strong motion | 0.037Hz - 0.08Hz | 0.037Hz - 0.08Hz | 0.015Hz - 0.08Hz |
| LGCB | Strong motion | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz |
| AATC | Strong motion | 0.028Hz - 0.08Hz | 0.028Hz - 0.08Hz | 0.032Hz - 0.08Hz |
| AES1 | Strong motion | N/A | N/A | 0.015Hz - 0.08Hz |
| AMNT | Strong motion | N/A | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz |
| APED | Strong motion | 0.035Hz - 0.08Hz | 0.035Hz - 0.08Hz | 0.035Hz - 0.08Hz |
| ATON | Strong motion | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz |
| AV18 | Strong motion | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz |
| AV21 | Strong motion | 0.032Hz - 0.08Hz | 0.015Hz - 0.08Hz | 0.015Hz - 0.08Hz |

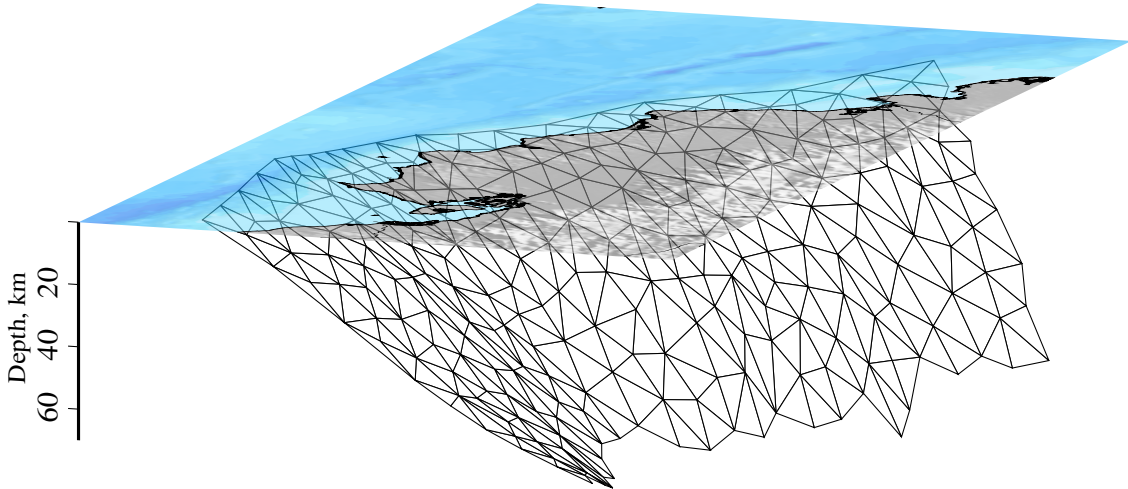


Figure S1: **Parametrization of the megathrust interface used for the coupling inversion.** Coupling value is inverted at each nodes

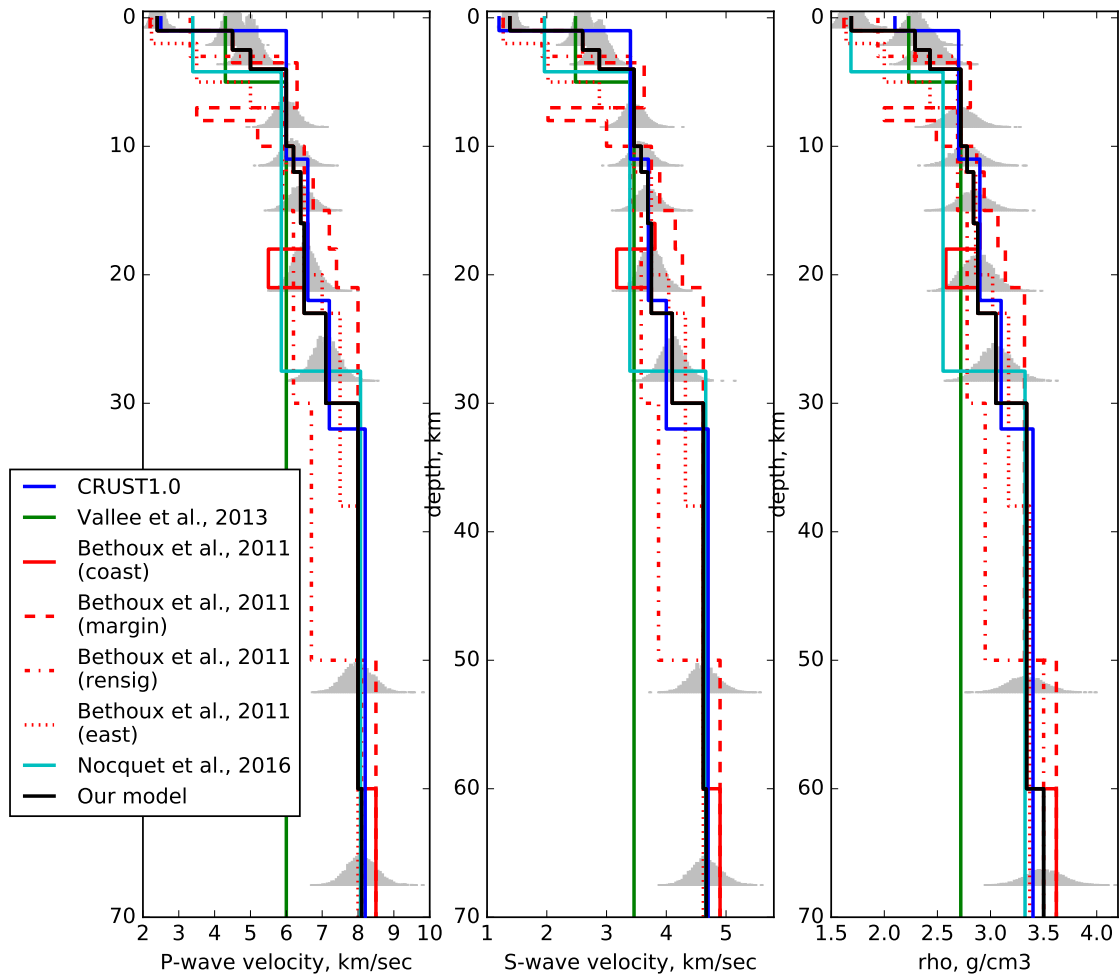


Figure S2: **Different models variability of the P-wave, S-wave, and density as a function of depth in central Ecuador.** A layered model used in this study for Green's function [GF] calculations is plotted as a solid black line. The blue line represents the CRUST2.0 model in the area (<http://igppweb.ucsd.edu/~gabi/rem.html>). The other models are from (Vallee et al., 2013; Bethoux et al., 2011; Nocquet et al., 2017). Grey histograms are the probability density function representing our confidence level on the elastic properties, as used to build the model prediction error.

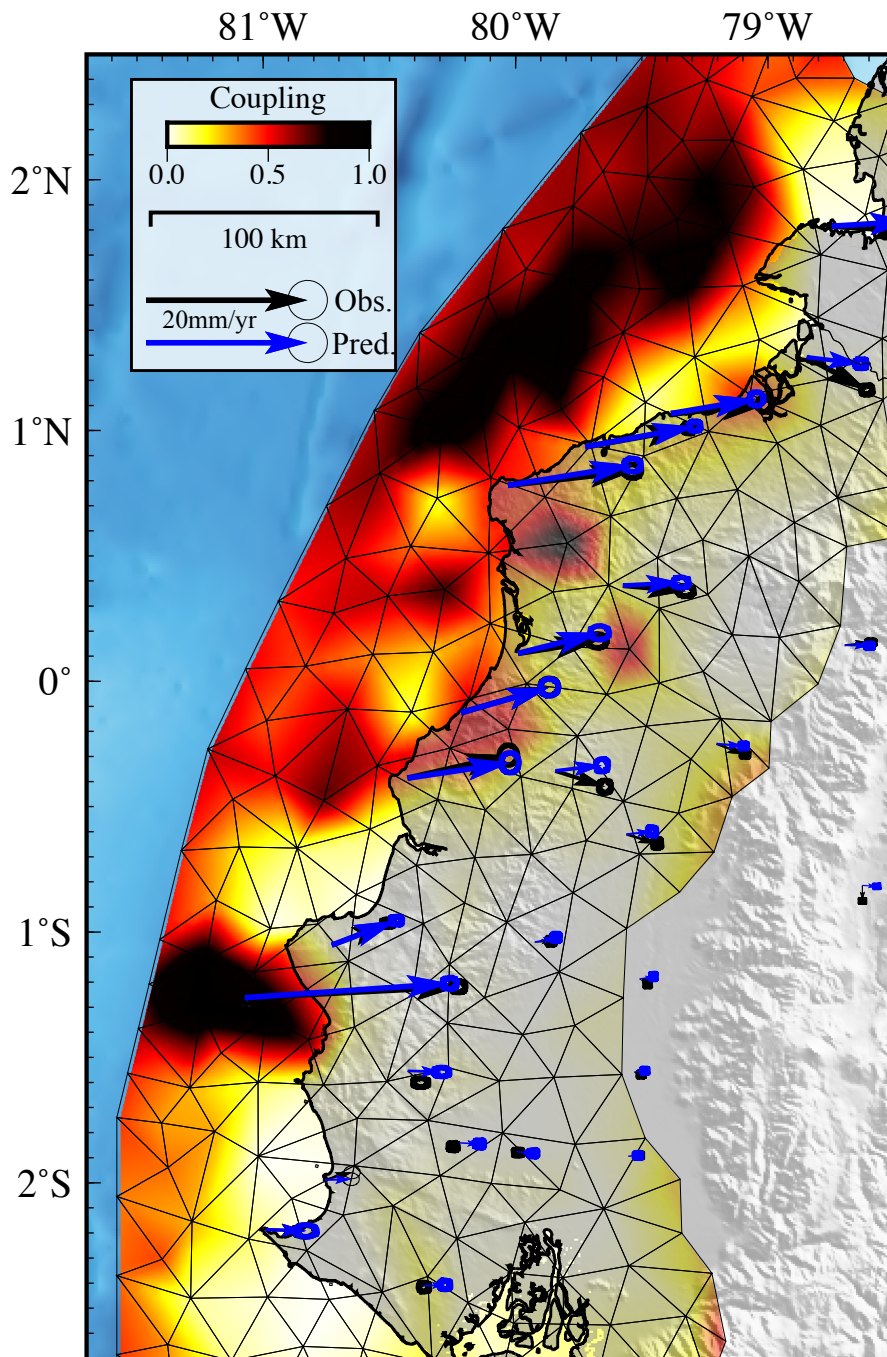


Figure S3: **Posterior Median coupling model.** Thin black lines represent the fault parametrization. Coupling values are inverted at each nodes. Interseismic GPS displacement and predictions for the median model are plotted as black and blue arrows, respectively.

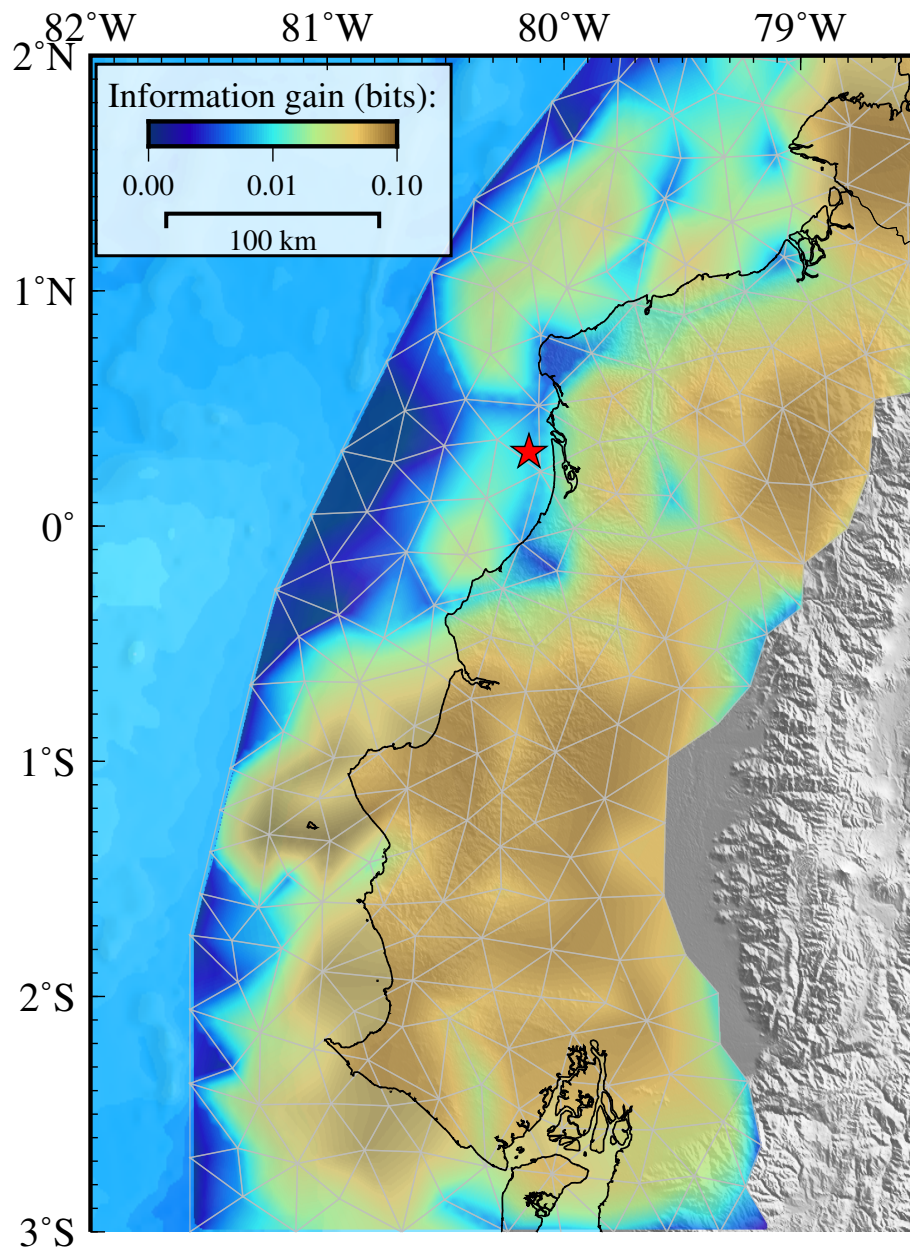


Figure S4: Kullback-Leibler divergence between the posterior and prior PDFs of coupling. Higher values indicate regions where the gain of information of the posterior PDF is significant relative to the prior distribution.

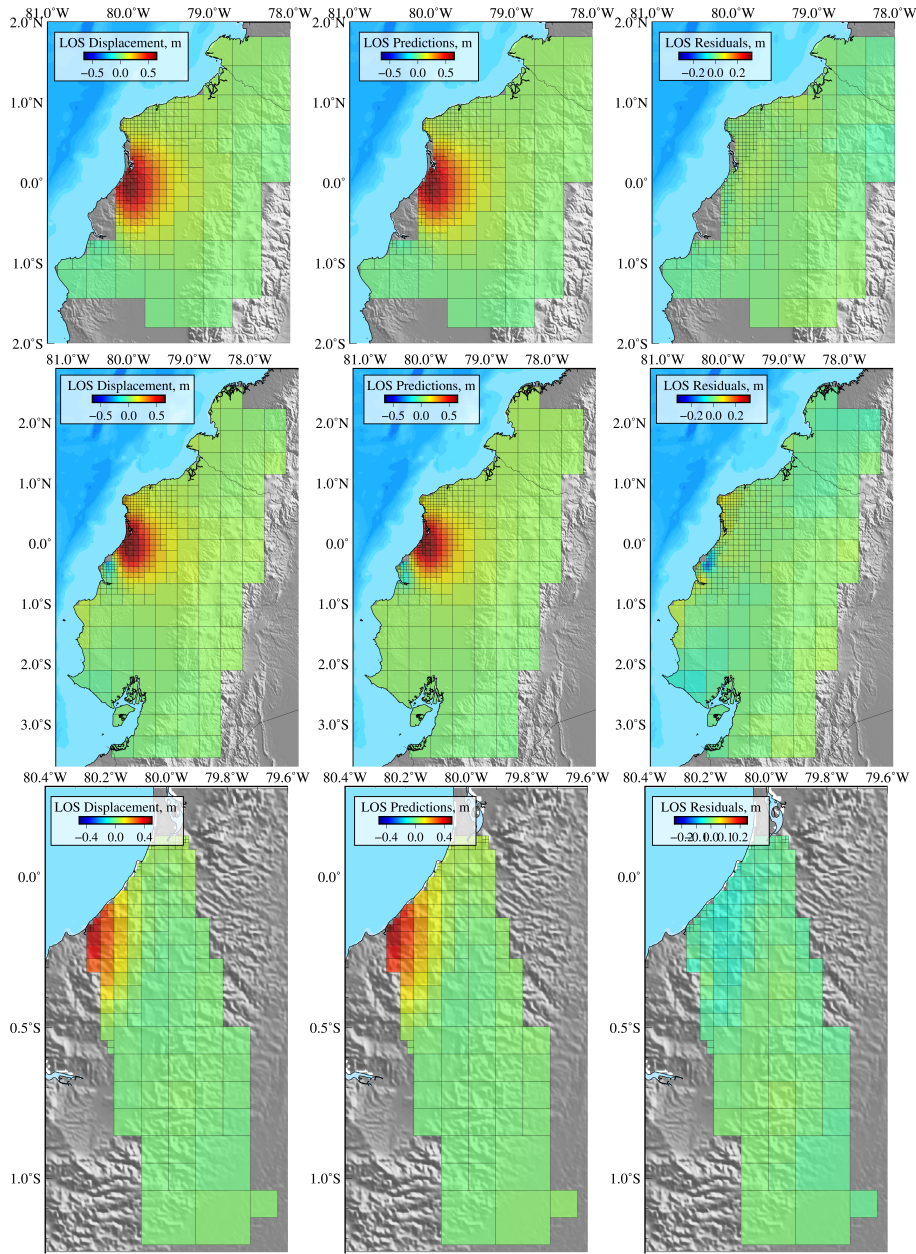


Figure S5: **Decimated InSAR observations, predictions, and residuals.** (a, d, g) Decimated InSAR observations inverted in this study. (b, e, h) Predictions for the posterior mean model. (c, f, i) Residuals of the Sentinel (top row), descending ALOS-2 (middle row), and ascending ALOS-2 (bottom row) interferograms.

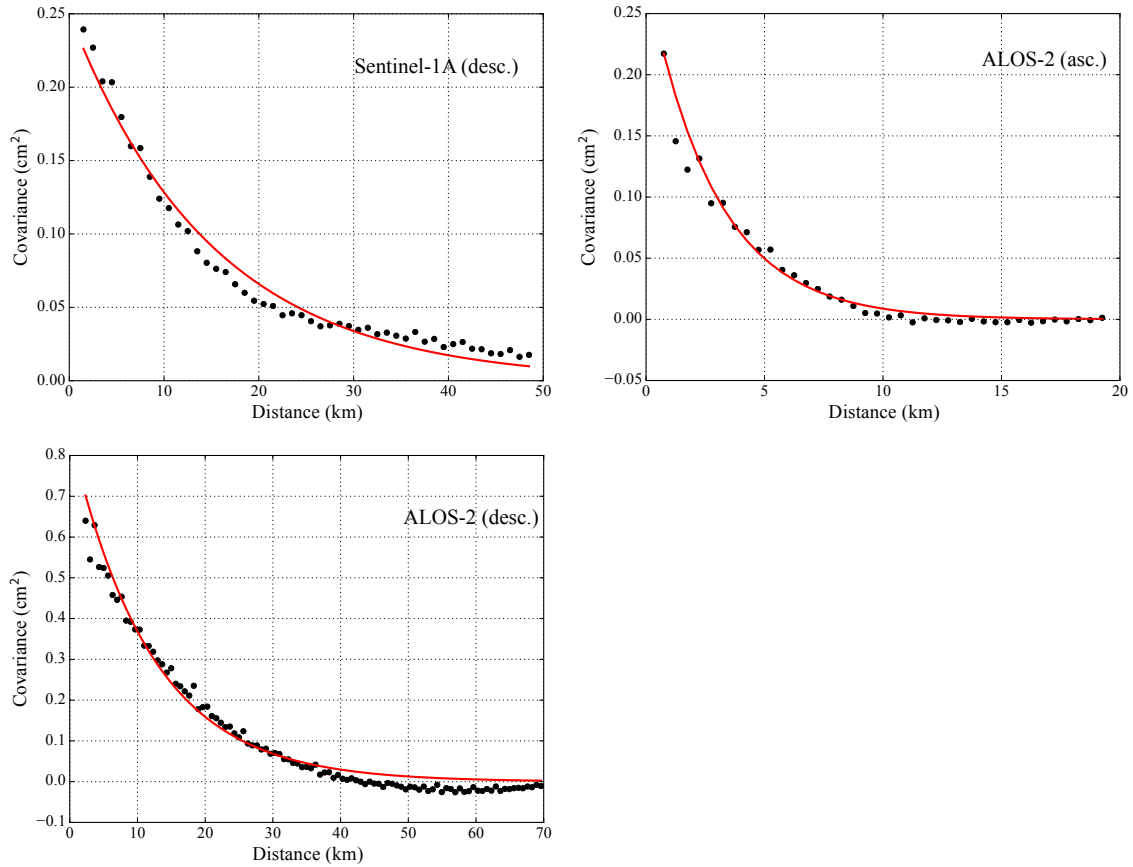


Figure S6: **Empirical covariance functions for the InSAR observations** 1D empirical covariance functions and the associated best-fit exponential function for each tracks. For each image, we compute the empirical covariance as a function of the distance between pixels and then fit an exponential function to these covariances (Jolivet et al., 2012). This exponential function is then used to build the data covariance matrix used in the inversion.

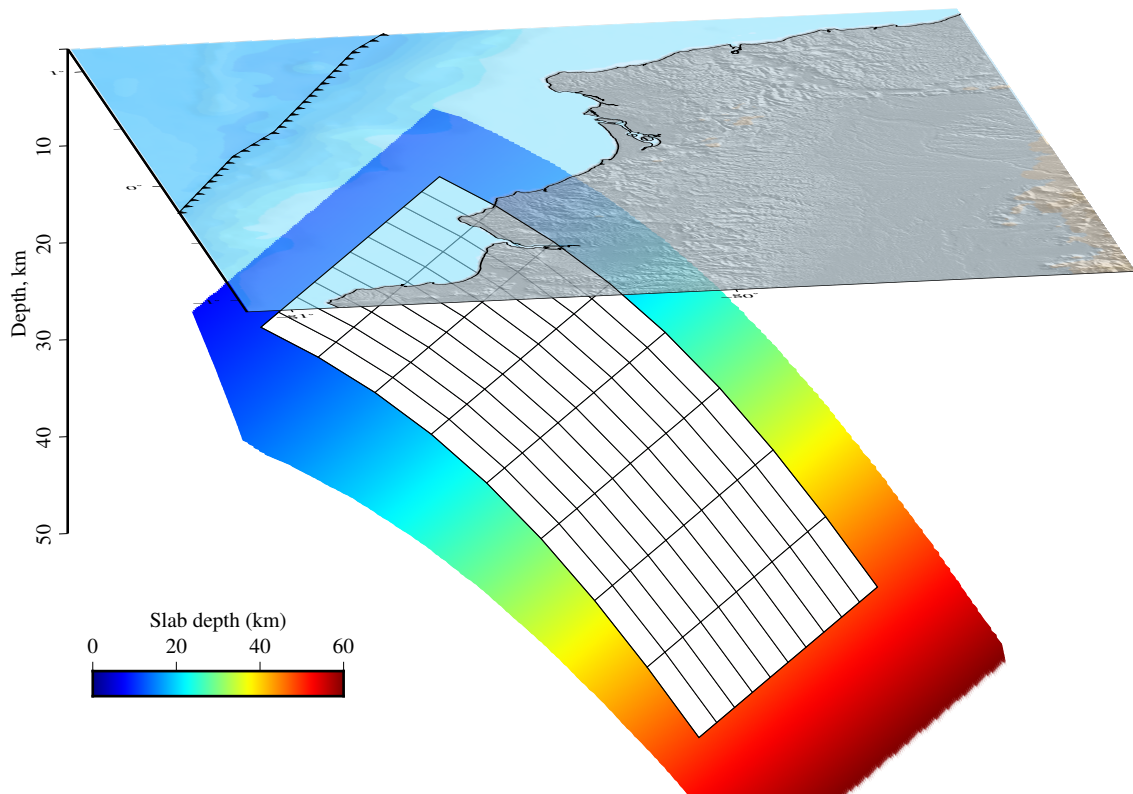


Figure S7: **Parametrization of the megathrust interface used for the co-seismic inversion**
 The coloured plane represent the slab1.0 model (Hayes et al., 2012). Each subfault patch is a 15 km x 15 km square

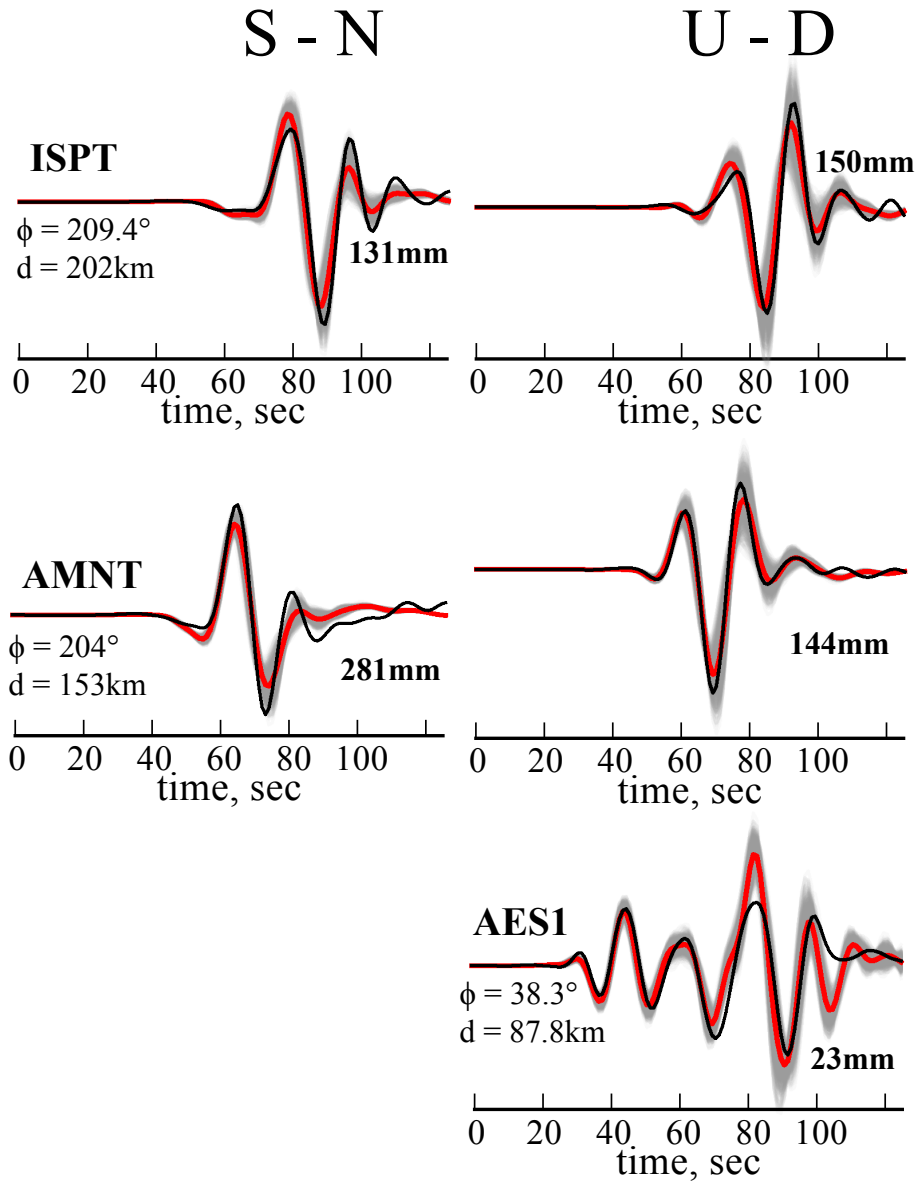


Figure S8: **Strong-motion observations and model predictions not presented in Figure 6 in the main text.** The North (left) and vertical (right) components of each station are plotted around the map. For each waveform, the bold number indicates it's maximum amplitude. The station azimuth Φ and distance d to the epicenter are also given. The black line is the recorded waveform. The gray lines are the stochastic predictions for our posterior model. The red line is the mean of the stochastic predictions.

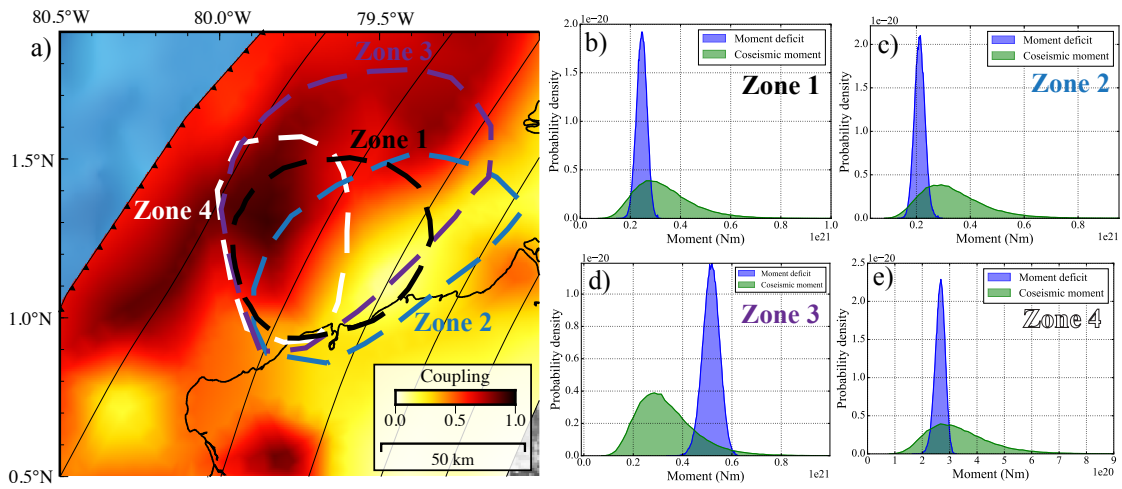


Figure S9: **Comparison of co-seismic moment and moment deficit in the 1958 earthquake region.** a) The background colour represents the coupling posterior mean model. The black dashed lines delimit four different areas where the co-seismic moment of the 1958 event and moment deficit for the 1906 - 1958 period are computed. b-e) Probability densities of the co-seismic moment released by the 1958 earthquake and the moment deficit accumulated between 1906 and 1958 in the different dashed area shown in a).

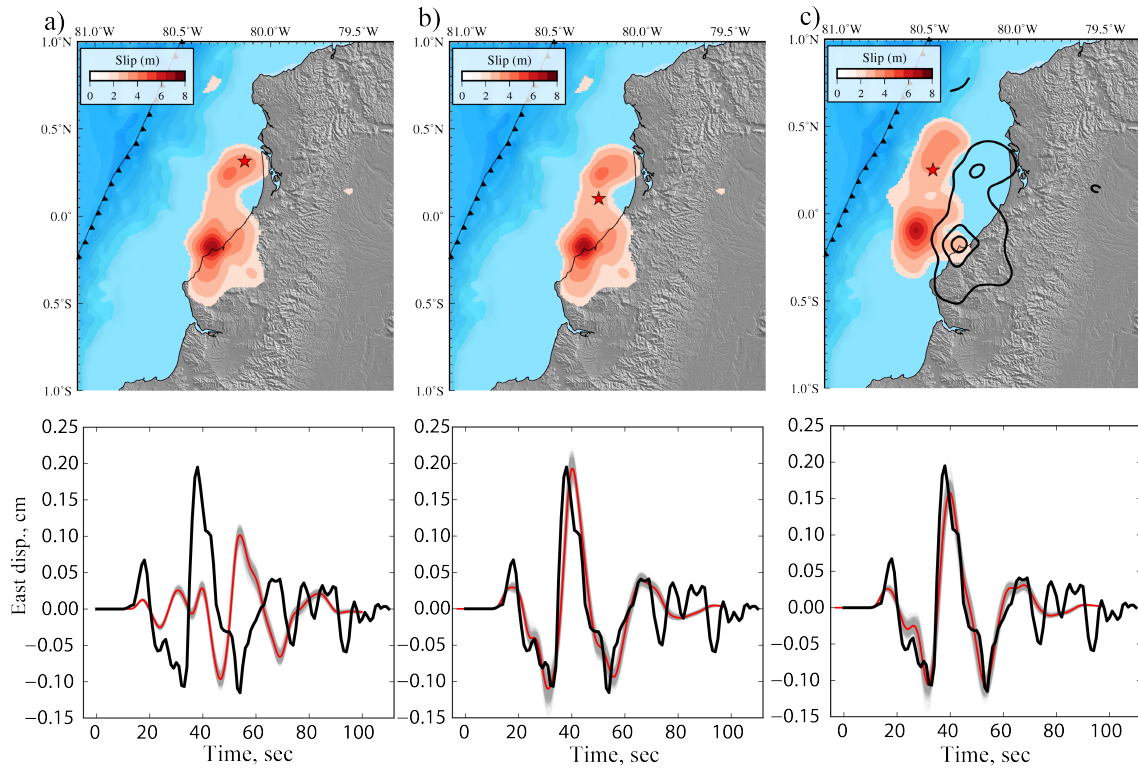


Figure S10: **Comparison of model predictions and 1942 earthquake waveform recorded in the DBN station, Netherlands.** (top) Slip model and hypocenter location (red star) used to compute the predictions shown in the bottom row. The model presented in a) results from the kinematic slip inversion of the 2016 earthquake. The models in b) and c) use a different hypocenter located between the two main slip asperities. The slip model in c) is the same as in a) and b), but located updip along the megathrust interface. Black lines in c) are slip contours of the original slip model. (bottom) East component waveform recorded at DBN for the 1942 earthquake (in black) and stochastic predictions (in grey) for the model shown on top. The red line is the posterior mean prediction. Predictions were convolved with the instrumental response of the Galitzin that recorded the event.

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