

Distribution of interseismic coupling along the North and East Anatolian Faults inferred from InSAR and GPS data

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Key Points:

- Distribution of interseismic coupling on the North and East Anatolian Faults
- Quantification of uncertainties provided by a Bayesian inversion framework
- The 2020 M_w 6.8 Elazığ earthquake released 221.5 years (± 26) of accumulated moment

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Abstract

The North Anatolian Fault (NAF) is one of the most hazardous fault in the world. After decades of low seismicity, the M_w 6.8 Elazığ earthquake (January 24, 2020) has recently reminded us that the East Anatolian Fault (EAF) is also capable of producing large earthquakes. To better estimate the seismic hazard associated with these two faults, we jointly invert Interferometric Synthetic Aperture Radar (InSAR) and GPS data to image the spatial distribution of interseismic coupling along the eastern part of both the North and East Anatolian Faults. We perform the inversion in a Bayesian framework, enabling to estimate uncertainties on both long-term relative plate motion and coupling. We find that coupling is high and deep (0-20 km) on the NAF and heterogeneous and superficial (0-5 km) on the EAF. Our model predicts that the Elazığ earthquake released between 200 and 250 years of accumulated moment, suggesting a bi-centennial recurrence time.

Plain Language Summary

Earthquakes are thought to occur on coupled fault portions, which are “locked” during the time separating two earthquakes while tectonic plates are steadily moving. The spatial distribution of coupling has been imaged along numerous large faults in the world, but despite its considerable associated seismic hazard, not on the North Anatolian Fault (NAF). The recent M_w 6.8 Elazığ earthquake (January 24, 2020) has reminded us that the East Anatolian Fault (EAF) is also capable of producing large earthquakes. To better assess the seismic hazard associated with both the NAF and the EAF, we image the distribution of interseismic coupling along these faults. We find that the NAF is homogeneously and strongly coupled. On the opposite, coupling is shallow and heterogeneous along the EAF. The initiation of the Elazığ earthquake coincides with a strongly locked but narrow (5 x 14 km) and superficial patch. The rest of the rupture extends over moderately coupled fault portions. Several fault segments along the EAF present similar coupling distributions, suggesting that, provided enough time, they could host earthquakes of similar magnitude. We estimate that it took between 200 and 250 years to accumulate the moment released by the Elazığ event.

1 Introduction

Earthquakes are thought to rupture fault portions which have previously accumulated a deficit of interseismic slip over tens to thousands of years (e.g., Avouac, 2015). Quantifying the spatial distribution of interseismic coupling – i.e. the percentage of slip deficit with respect to the long-term drift of tectonic plates – along large faults is therefore crucial to anticipate earthquakes and better assess seismic hazard (e.g., Kaneko et al., 2010). The emergence of space geodetic techniques has enabled inferring interseismic coupling along a number of large faults during long quiescent periods of time separating one large earthquake to the next (e.g., Bürgmann et al., 2005; Moreno et al., 2010; Loveless & Meade, 2011; Protti et al., 2014; Jolivet et al., 2015; Metois et al., 2016; Nocquet et al., 2017). Though interseismic coupling models have been proposed to estimate the locking depth of the North and East Anatolian Faults (e.g., Tatar et al., 2012; Mahmoud et al., 2013; Cavalié & Jónsson, 2014; Aktug et al., 2016), none have quantified the lateral variations of coupling along these faults, which has prevented from studying the relationship between coupling and large earthquakes. The density of InSAR observations (Cavalié & Jónsson, 2014) combined with sparser GPS measurements allows to infer these lateral variations of coupling on the eastern part of the NAF-EAF system (Fig. 1).

The eastern part of the North Anatolian Fault (NAF) is known to produce large earthquakes (e.g., Ambraseys, 1971, 1989; Barka, 1996) and thought to be coupled from 0 to 15 km depth (Reilinger et al., 2006; Cavalié & Jónsson, 2014). On the other hand,

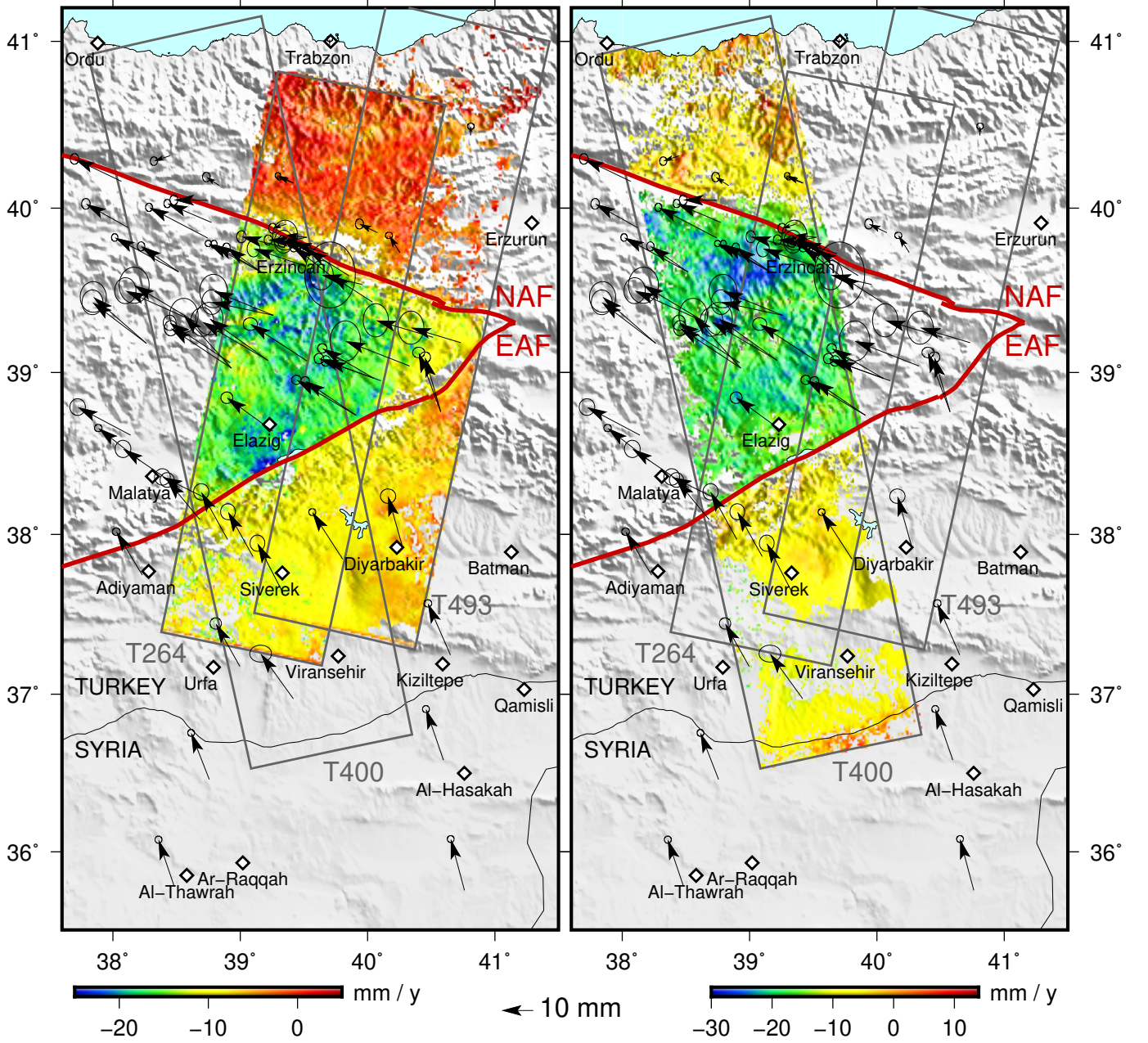


Figure 1. The NAF-EAF system (red lines) and available observations of surface deformation. Color maps show InSAR horizontal velocities (in a Eurasia-fixed reference frame) in the satellite line of sight (LOS) direction, $\sim 103^\circ$ N for descending tracks T264 and T493 (left), $\sim 77^\circ$ N for ascending track T400 (right) (Cavalié & Jónsson, 2014). Black arrows show GPS measurements and their 95% ellipses of uncertainty (Reilinger et al., 2006; Ozener et al., 2010; Tatar et al., 2012). White diamonds indicate large cities.

simple back slip models showed that the East Anatolian Fault (EAF) is weakly coupled and only in the first kilometers of the upper crust, from 0 to 5 km (Cavalié & Jónsson, 2014). This observation was in good agreement with the low seismicity rate recorded during the twentieth century (Burton et al., 1984; Jackson & McKenzie, 1988). For those reasons, the January 24 2020 M_w 6.8 Elazığ earthquake came as a surprise, on a segment where the last earthquake of comparable magnitude occurred in 1905 (Nalbant et al., 2002). To understand this unexpected event, and more generally the seismicity in the region, we infer here the spatial distribution of interseismic coupling along the eastern part of the NAF-EAF system using InSAR (Cavalié & Jónsson, 2014) and GPS measurements (Reilinger et al., 2006; Ozener et al., 2010; Tatar et al., 2012) of the long term surface deformation (Fig. 1).

Space geodesy provides measurements of the surface deformation which are informative of the fault behavior at depth. Extracting the information contained in geodetic observations of the Earth surface deformation – such as InSAR and GPS – to infer the interseismic coupling along faults requires solving an inverse problem which solution is non-unique (Tarantola & Valette, 1982; Nocquet, 2018). Most inversion techniques deal with this non-uniqueness by finding the solution that best fits the observations, in a least square sense. Because the best-fitting solution is usually very heterogeneous, a roughness penalty is usually applied so that typical published coupling (or slip) models are the smoothest best-fitting solutions among an infinity of possible models. We adopt here a Bayesian sampling approach, which does not invert for a specific “ambiguously-defined best solution” but for the entire solution space, sampled with respect to the likelihood of each model. This approach originally developed to invert for co-seismic slip models (Minson et al., 2013) enables to reliably estimate uncertainties on coupling distributions (Jolivet et al., 2015).

2 Bayesian inversion of rotation poles and interseismic coupling along two faults from InSAR and GPS data

We invert for the interseismic coupling along the North-East Anatolian fault system using InSAR and GPS measurements of the eastern Anatolia surface deformation (Fig. 1). Our InSAR dataset is composed of two descending and one ascending tracks all crossing both the North and East Anatolian faults near their junction in eastern Turkey (Cavalié & Jónsson, 2014). Our GPS dataset is composed of the horizontal components of 72 GPS stations located in the area (Reilinger et al., 2006; Ozener et al., 2010; Tatar et al., 2012).

We use a Bayesian sampling approach to estimate the probability density function (pdf) $p(\mathbf{m}|\mathbf{d})$ of a large number of likely models \mathbf{m} given our data \mathbf{d} . This pdf may be evaluated based on the ability of a model \mathbf{m} to predict the data \mathbf{d} (Minson et al., 2013)

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

where \mathbf{G} is the matrix of the Green’s functions and \mathbf{C}_χ is the misfit covariance.

Vector \mathbf{d} is composed of 144 GPS measurements (72×2 components) and a subset of InSAR pixels on the 3 tracks down-sampled using the Quadtree algorithm (Jónsson et al., 2002).

Because the inferred distribution of coupling is highly sensitive to the (usually) pre-determined tectonic block motion, especially in a case involving 3 plates, we do not impose pre-calculated plate rotations but invert for them simultaneously with the interseismic coupling. To do so, we express the rotation vector \mathbf{w}^p in Cartesian geocentric

coordinates

$$\mathbf{w}^p = \Omega^p \begin{pmatrix} \cos \phi^p \cos \lambda^p \\ \cos \phi^p \sin \lambda^p \\ \sin \phi^p \end{pmatrix} \quad (3)$$

where λ^p and ϕ^p are the longitude and latitude of the Euler pole of a plate p and Ω^p is its angular velocity (Bowring, 1985). We then discretize the eastern part of the North and East Anatolian faults into 110 subfaults of depth-dependent sizes (Table S1, S2) and invert for the model vector

$$\mathbf{m} = \begin{pmatrix} \mathbf{w}^1 \\ \mathbf{w}^2 \\ \mathbf{S} \end{pmatrix} \quad (3)$$

where ¹ stands for Anatolia with respect to Eurasia, ² for Arabia with respect to Eurasia, and \mathbf{S} is the back-slip on each subfault. Accordingly, we build \mathbf{G} so that

$$\mathbf{G} = (\mathbf{A} \quad -\mathbf{G}_S) \quad (4)$$

where \mathbf{A} is the linearized rotation matrix in Cartesian geocentric coordinates (see Appendix A) and \mathbf{G}_S is the classical matrix of the Green's functions computed using the analytical solution of a shear finite fault embedded in an elastic half space (Mansinha & Smylie, 1971; Okada, 1985).

\mathbf{C}_χ is the covariance matrix, which translates data and epistemic uncertainties into uncertainties on the inverted model \mathbf{m} (Duputel et al., 2014; Bletery et al., 2016; Ragon et al., 2018; Ragon, Sladen, & Simons, 2019; Ragon, Sladen, Bletery, et al., 2019). Here, we only account for data uncertainties. For GPS records, we fill \mathbf{C}_χ with the (squared) standard deviations and covariances between the east and north components of a given station provided in the GPS solutions. For InSAR pixels, we first remove the tectonic signal from the unsampled interferograms using a preliminary model and calculate the covariance across the pixels of the residual interferograms as a function of their distances. We fit an exponential function to the obtained cloud of points and express the covariance $C_{i,j}$ between 2 pixels as a function of their distance $D_{i,j}$

$$C_{i,j} = a^2 \exp\left(\frac{-D_{i,j}}{b}\right) \quad (5)$$

by applying a regression to the parameters a and b independently on the 3 tracks (Jolivet et al., 2015). We then use equation 5 to evaluate the covariance on the sub-sampled interferograms.

$p(\mathbf{m})$ is the pdf describing the prior information assumed on the different model parameters. We choose the less informative distributions for back-slip parameters \mathbf{S} , i.e. uniform distributions between 0 and the long term interplate velocities: 19.5 mm/y for the North Anatolian and 13 mm/y for the East Anatolian fault (Cavalié & Jónsson, 2014). For the rotation vectors \mathbf{w}^1 and \mathbf{w}^2 , we draw 100,000 sets of parameters ($\lambda^1, \phi^1, \Omega^1, \lambda^2, \phi^2, \Omega^2$) from log-normal distributions defined by means and standard deviations taken from previously published solutions (Le Pichon & Kreemer, 2010, summarized in Table 1). For each drawn set of parameters, we calculate the corresponding \mathbf{w}^1 and \mathbf{w}^2 . We obtain Gaussian-like distributions for each component of \mathbf{w}^1 (Fig. S1) and \mathbf{w}^2 (Fig. S2) and use them as prior pdfs.

3 Distribution of interseismic coupling in eastern Anatolia

We obtain a posterior marginal pdf for every inverted parameter in \mathbf{m} . We convert the inverted pdfs on the rotation vectors ($\mathbf{w}^1, \mathbf{w}^2$) (Fig. S3) into pdfs on the Euler pole coordinates and angular velocities (Fig. S4). The means and 2- σ standard deviations of the inverted pdfs are summarized in Table 1. They are close to the previously published values we used as a prior (Le Pichon & Kreemer, 2010) but not equal. The

Table 1. *A priori* (Le Pichon & Kreemer, 2010) and *a posteriori* Euler pole coordinates and angular velocities with respect to Eurasia. *A posteriori* parameters are the mean and 2- σ standard deviation (95% confidence) of the posterior pdfs (Fig. S4).

	Plate	Longitude ($^{\circ}$ E)	Latitude ($^{\circ}$ N)	Angular velocity ($^{\circ}$ /My)
<i>A priori</i>	Anatolia	31.96 ± 0.10	32.02 ± 0.10	1.307 ± 0.083
	Arabia	15.21 ± 0.10	28.31 ± 0.10	0.396 ± 0.010
<i>A posteriori</i>	Anatolia	34.22 ± 0.35	30.96 ± 0.60	1.087 ± 0.078
	Arabia	16.13 ± 0.52	27.08 ± 0.37	0.386 ± 0.008

likely explanation for this small discrepancy is that the plates are not strictly rigid and thus the rotations we invert from data concentrated near the faults are slightly different from those obtained for the rotations averaged over the entire plates. Our goal here is to infer the coupling distribution, and for that aim a refined estimate of the rotation parameters close to the fault is preferable to a plate-average solution, but one should be careful in using values in Table 1 for other purposes.

For each posterior Euler pole, we calculate the rotation predicted at the center of each patch and project the obtained vector on the strike direction to obtain posterior pdfs of the long-term rate along the faults (Fig. S5). These pdfs are consistent with a steady long term slip rate of ~ 20 mm / year along the entire studied segment of the NAF. On the EAF, they predict a rate of ~ 15 mm / year on the eastern end decreasing westward down to ~ 10 mm / year on the western end of the studied segment (Figs. 2, S5, Tables S1, S2).

We divide the back-slip obtained on each patch for each sampled model by the long-term fault rate predicted by each model on each subfault to obtain the posterior marginal pdfs on the coupling coefficients (Fig. S6, S7). We show these pdfs in the form of their means (Fig. 2) and standard deviations (Fig. 3). Although restrictive, this representation gives an approximate view of the coupling spatial distribution and its associated uncertainties. Uncertainty is high on the extreme west and – to a lesser extent – the extreme east parts of the fault system which are located outside of the InSAR tracks (Fig. 1). The standard deviation on most parts of the faults is $< 20\%$, much lower on many subfaults (Fig. 3).

We calculate the GPS and InSAR measurements predicted for every posterior sampled model. We plot the predicted GPS means (red arrows) and 2- σ standard deviations (red ellipses) on Fig. S8. For InSAR, we plot the mean predicted line of sight (LOS) displacements (Fig. S10-S12) and standard deviations (Fig. S13). The range of likely models that we found (Fig. S6-S7) is in very good agreement with both GPS and InSAR data. Residuals are small and show no coherent pattern (Fig. S9, S13). We find that every posterior sampled model predict very similar GPS and InSAR displacements; red ellipses are hardly visible on Fig. S8 and the standard deviations of the predicted InSAR LOS displacements are very small (Fig. S13). This highlights the limited resolution on the coupling model: if different models predict the same observations, discriminating between them is difficult.

We show focal mechanisms of $M > 4.8$ earthquakes in the studied area from the Global Centroid Moment Tensor (GCMT) catalog (Dziewonski et al., 1981; Ekström et al., 2012) for events posterior to 1976 and from a compilation of historical earthquakes (Tan et al., 2008) for earlier events (1938 – 1976) (Fig. 2). Focal mechanisms are represented at the location of their surface projections (i.e. at depth = 0). Colors indicate the dates of the events. The largest earthquake in the studied area is the 1939 M_S 8.0

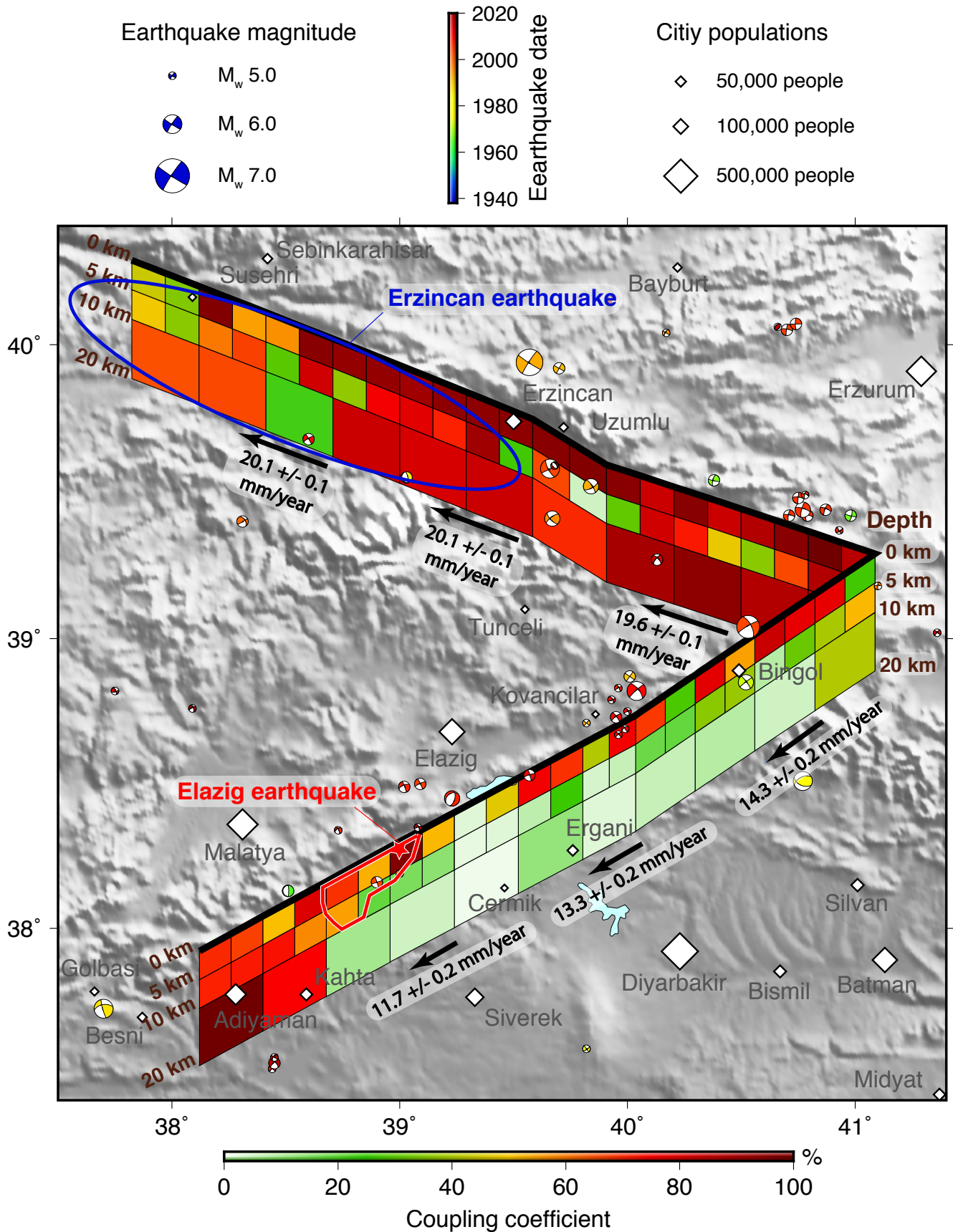


Figure 2. Interseismic coupling distribution inverted from InSAR and GPS data (mean of posterior pdfs in Fig. S6-S7). Black thick arrows indicate the long-term slip rate at depth derived from the inversion (mean and standard deviation of posterior pdfs in Fig S5). Focal mechanisms show $M > 4.8$ earthquakes (colors indicate event dates). Contours delineate the approximate rupture extent of the 1939 M_S 8.0 Erzincan earthquake and of the 2020 M_w 6.8 Elazığ earthquake (USGS finite fault solution). Red star indicates the epicenter of the Elazığ earthquake.

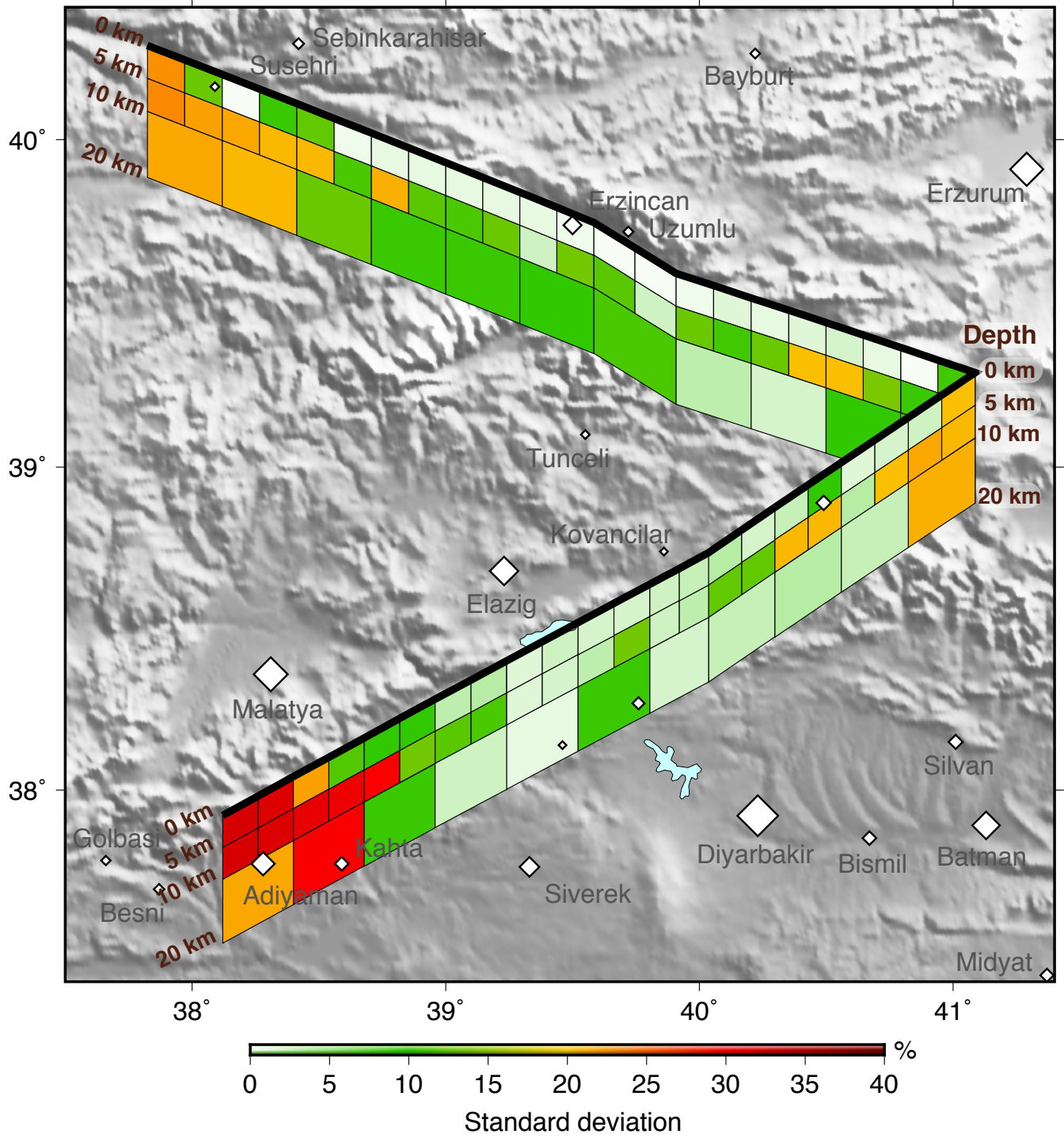


Figure 3. Standard deviation of the coupling posterior pdfs $p(S|d)$. The extreme west and – to a lesser extent – the extreme east parts of the fault system, presenting high standard deviations, are located outside of the InSAR tracks.

Erzincan earthquake which initiated near Erzincan and extended over the entire NAF segment west of Erzincan represented in Fig. 2 (Barka, 1996; Stein et al., 1997). We find that this entire section is strongly coupled, such as the rest of the studied NAF segment east of Erzincan. This easternmost segment of the NAF presents a moderate seismicity compared to the rest of the NAF. Our interseismic slip distribution suggests that it is as prone to generate large earthquakes as the rest of the NAF and as the Erzincan rupture segment in particular.

We find that locking on the EAF is much shallower with high coupling values limited to the first 5 km, consistently with previous studies (Cavalié & Jónsson, 2014). High coupling found at depth on the westernmost part of the fault has large uncertainty (Fig. 3). Furthermore, we find that coupling also varies within the shallowest portion of the fault, alternating strongly coupled segments with seemingly creeping sections (Fig. 2). We note that the few earthquakes recorded on the EAF coincides with relatively high coupling. Before the recent Elazığ earthquake, the two largest events occurred near the localities of Bingöl (M_w 6.3, 2003) and Kovancilar (M_w 6.1, 2010). The second one was followed by numerous aftershocks with magnitudes up to 5.6. All of these earthquakes occurred on $> 65\%$ coupled fault portions while fault segments with coupling $< 50\%$ do not appear to have hosted $M > 4.8$ earthquakes.

According to the USGS finite-fault model (USGS, 2020), the Elazığ earthquake initiated between Elazığ and Malatya (red star in Fig. 2) and propagated unilaterally westward (red contour in Fig. 2). The early part coincides with a strongly-locked (coupling coefficient: 96%) but narrow (13.7×5 km) patch. The rupture seems then to have propagated throughout moderately coupled (coupling coefficient: 50-80%) fault segments. Although the USGS model is preliminary, its contours correlate fairly well with the coupling distribution, suggesting that the rupture stopped when reaching $< 20\%$ coupled fault portions (Fig. 2).

The last $M > 6.6$ earthquake in the approximate region dates back to 1905 ($M_S = 6.7$) (Nalbant et al., 2002). This event was located west of the recent Elazığ earthquake (38.6° E, 38.1° N) (Nalbant et al., 2002) but, given location uncertainties, may have ruptured the same fault portion. We calculate, for each sampled coupling model, the accumulated moment inside the rupture contour of the Elazığ earthquake since 1905. To simplify the problem, we assume that the earthquake ruptured the entire surface of the 4 main subfaults inside the rupture contour and not more, i.e. the 3 shallowest subfaults plus the westernmost intermediate-depth one (Fig. 3). We obtain a pdf of the seismic moment accumulated since 1905 (Fig. 4.a). The pdf mean is 7.3×10^{18} N.m, its standard deviation 0.8×10^{18} N.m. According to the USGS solution, the seismic moment released during the Elazığ earthquake is 13.87×10^{18} N.m $\gg 7.3 \pm 0.8 \times 10^{18}$ N.m. This seems to indicate that the recent Elazığ earthquake did not rupture the same fault portion than the 1905 earthquake. We further calculate the pdf of the time necessary to accumulate the seismic moment which was released during the 2020 Elazığ earthquake (Fig. 4.b). The mean and standard deviation of the obtained pdf give a recurrence time for an Elazığ-type earthquake of 221.5 ± 26 years.

4 Conclusion

We inverted InSAR and GPS observations to image the interseismic coupling along the North and East Anatolian faults in eastern Turkey. We adopted a Bayesian sampling approach in order to estimate posterior uncertainties on the coupling distributions and on the long term fault rate. We did not impose a pre-calculated plate motion but inverted for the rotation of both the Anatolian and Arabian plates with respect to Eurasia, ensuring that the inferred coupling distribution is not biased in a systematic way by an inaccurate plate motion model. We found that the North Anatolian fault is homogeneously strongly coupled from 0 to 20 km depth while the East Anatolian fault is weakly cou-

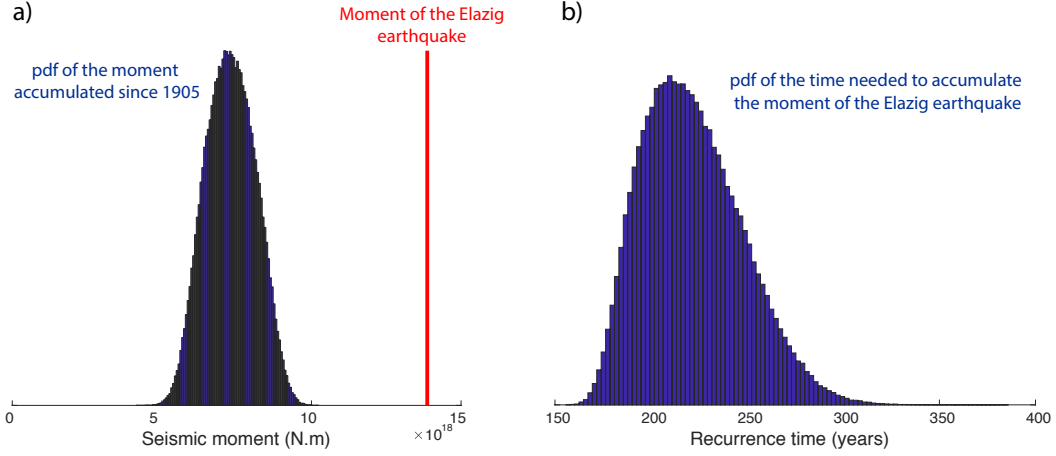


Figure 4. a) Pdf of the accumulated seismic moment on the 4 patches inside the Elazığ rupture since 1905. The red vertical line indicates the seismic moment of the Elazığ earthquake according to the USGS solution (13.87×10^{18} N.m). b) Pdf of the time necessary to accumulate the seismic moment which was released during the Elazığ earthquake.

pled for the most part with high coupling values limited to the shallowest part of the fault (0 to 5 km). Furthermore, we find that coupling is heterogeneous within this shallow portion, alternating seemingly creeping sections with strongly locked patches. Comparison between our interseismic coupling distribution and the preliminary finite-fault model of the USGS for the 2020 M_w 6.8 Elazığ earthquake reveals that this event likely initiated on one of this strongly locked (coupling coefficient: 96%) fault patch and then propagated into moderately coupled fault segments (coupling coefficient: 50-80%). Overall, we estimate that the Elazığ earthquake released $221.5 (\pm 26)$ years of accumulated moment, suggesting a recurrence time comprised between 200 and 250 years.

Appendix A Rotation matrix \mathbf{A}

We build the rotation matrix \mathbf{A} so that the motion due to the rotation of both the Anatolian and Arabian plates with respect to Eurasia is equal to $\mathbf{A} \cdot \mathbf{W}$, where

$$\mathbf{W} = \begin{pmatrix} \mathbf{w}^1 \\ \mathbf{w}^2 \end{pmatrix} \quad (\text{A1})$$

Sorting all data points located on the Eurasian plate at the beginning of \mathbf{d} , all data points located on the Anatolian plate in the middle and all data points located on the Arabian plate at the end, i.e. writing \mathbf{d} as

$$\mathbf{d} = \begin{pmatrix} \mathbf{d}_0 \\ \mathbf{d}_1 \\ \mathbf{d}_2 \end{pmatrix} \quad (\text{A2})$$

with \mathbf{d}_0 , \mathbf{d}_1 , \mathbf{d}_2 data points located on the Eurasian, Anatolian and Arabian plates respectively, we can write \mathbf{A} as a block matrix

$$\mathbf{A} = \begin{pmatrix} \mathbf{0} & \mathbf{0} \\ \mathbf{A}' & \mathbf{0} \\ \mathbf{0} & \mathbf{A}' \end{pmatrix} \quad (\text{A3})$$

so that $\mathbf{A} \cdot \mathbf{W}$ equals $\mathbf{0}$ for data points in Eurasia, $\mathbf{A}' \cdot \mathbf{w}^1$ in Anatolia and $\mathbf{A}' \cdot \mathbf{w}^2$ in Arabia. \mathbf{A}' is a transfer matrix relating the rotation vector in Cartesian geocentric co-

ordinates \mathbf{W} to the rotation block motion at each data point. It can be expressed at the location of an InSAR pixel or GPS station of longitude λ and latitude ϕ as

$$\mathbf{A}'_{\lambda,\phi} = \begin{pmatrix} -\sin \lambda & \cos \lambda & 0 \\ -\sin \phi \cos \lambda & -\sin \phi \sin \lambda & \cos \phi \\ \cos \phi \cos \lambda & \cos \phi \sin \lambda & \sin \phi \end{pmatrix} \cdot \begin{pmatrix} 0 & z & -y \\ -z & 0 & x \\ y & -x & 0 \end{pmatrix} \quad (\text{A4})$$

where

$$\begin{pmatrix} x \\ y \\ z \end{pmatrix} = R_e(1 - \epsilon \sin^2 \phi)^{-1/2} \begin{pmatrix} \cos \phi \cos \lambda \\ \cos \phi \sin \lambda \\ (1 - \epsilon) \sin \phi \end{pmatrix} \quad (\text{A5})$$

with $R_e = 6378.137$ km the Earth equatorial radius and $\epsilon = 0.00669438003$ the Earth eccentricity (Bowring, 1985).

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