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1 **Impulsive source of the 2017, $M_W=7.3$, Ezgeleh, Iran,**
2 **earthquake**

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12 **Key Points:**

- 13 • The Ezgeleh earthquake ruptured a flat thrust fault in the Zagros fold and thrust
14 belt
- 15 • Kinematic slip modelling reveals a highly impulsive source with southward direc-
16 tivity, possibly causing the large damage in the area
- 17 • The direction of co-seismic slip suggests a strain partitioning between thrust and
18 unmapped strike-slip faults

**This is the accepted version of the manuscript.
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Abstract

On November 12th 2017, a $M_W=7.3$ earthquake struck near the Iranian town of Ezgeleh, at the Iran-Iraq border. This event was located within the Zagros fold and thrust belt which delimits the continental collision between the Arabian and Eurasian Plates. Despite a high seismic risk, the seismogenic behaviour of the complex network of active faults is not well documented in this area due to the long recurrence interval of large earthquakes. In this study, we jointly invert InSAR and near-field strong-motions to infer a kinematic slip model of the rupture. The incorporation of these near-field observations enables a fine resolution of the kinematic rupture process. It reveals an impulsive seismic source with a strong southward rupture directivity, consistent with significant damage south of the epicenter. We also show that the slip direction does not match plate convergence, implying that some of the accumulated strain must be partitioned onto other faults.

Plain Language Summary

Iran is a very seismically active region. However, the 2017 Ezgeleh earthquake of magnitude 7.3 occurred in a region where large earthquakes have not been documented for several centuries. Our knowledge of fault locations, geometries, and seismic behaviours is therefore limited in this region. We use near-field seismological and satellite geodetic data to retrieve the spatial and temporal distribution of slip occurring on the fault during the Ezgeleh earthquake. We show that the high slip rate and Southward directivity of the rupture may have worsen damage South of the epicentre. We also observe that tectonic motion is partitioned between different type of faults. Although the Ezgeleh earthquake did release a significant part of that strain, other seismogenic faults in the region could represent an important hazard for nearby population.

1 Introduction

On November 12th, 2017, the Iranian province of Kermanshah and the Iraqi Kurdistan were shaken by a severe $M_W=7.3$ earthquake located south of the border. It caused the death of ~ 630 people and considerable damage, in particular in the Iranian city of Sarpol-e Zahab (c.f. Figure 1). The earthquake triggered numerous landslides and rock falls, including a massive 4x1 km landslide in Kermanshah (Miyajima et al., 2018).

49 The hypocenter is located within the Zagros Mountains near the Iranian town of
50 Ezgeleh, a tectonically active region that accommodates crustal shortening (e.g., Berberian & King, 1981) resulting from the collision between the Arabian Plate and the Eurasian
51 Plate. About a third to a half of current convergence is accommodated within the Za-
52 gros belt (Vernant et al., 2004). The belt hosts many moderate earthquakes ($M=5-6$)
53 with depths ranging from 4 km to 20 km, although this range is debated (e.g., Niazi et
54 al., 1978; Nissen, Tatar, Jackson, & Allen, 2011; Talebian & Jackson, 2004). Our knowl-
55 edge of the regional seismo-tectonics is further complicated by the very rare occurrence
56 of co-seismic surface rupture (Talebian & Jackson, 2004; Walker et al., 2005).
57

58 The Ezgeleh earthquake occurred at the transition between the Lorestan Arc in the
59 south-east and the Kirkuk Embayment in the north-west (c.f. Figure 1). The area is cov-
60 ered by a 8-13 km thick sedimentary cover heavily folded into numerous anticlines (e.g.,
61 Alavi, 2007; Falcon, 1969). Sediments are crossed by many thrust faults that flatten within
62 the basement (Sadeghi & Yassaghi, 2016; Tavani et al., 2018). As expected from the lack
63 of surface ruptures and fault scarps, most of these faults are blind, hence the difficulty
64 to infer their geometry. In this region, plate convergence is roughly north-south (c.f., Fig-
65 ure 1) with a rate between 19 mm/yr (Kreemer, Blewitt, & Klein, 2014) and 24 mm/yr
66 (DeMets, Gordon, & Argus, 2010). Slip is partitioned between thrust faults at the front
67 of the belt, such as the Mountain Front Fault, the High Zagros Fault and the Zagros Fore-
68 deep Fault, and the Main Recent Fault, a right-lateral strike-slip fault located at the back
69 of the belt (c.f., Figure 1; Berberian, 1995). This part of the Zagros belt hosts moder-
70 ate seismicity, but the last significant earthquakes ($5.9 \lesssim M \lesssim 6.4$) to strike the area hap-
71 pened in 958 and 1150 (Ambraseys & Melville, 2005). Therefore, our understanding of
72 the regional seismo-tectonic setting is obscured by the undersampled seismic cycle and
73 the absence of ground geodesy. The 2017 Ezgeleh earthquake highlighted the seismic haz-
74 ard in this portion of the Zagros belt. Its analysis hence provides a unique opportunity
75 to enrich our understanding of the region and the associated seismic hazard. In addi-
76 tion, the availability of near-field strong-motion records offers the possibility to closely
77 study the propagation of the rupture on the fault.

78 In this study, we propose a stochastic analysis of the 2017 earthquake source pro-
79 cess. We use a Bayesian framework to infer a population of co-seismic slip models that
80 fit available observations. While currently available studies were either limited to the static
81 final distribution of slip on the fault (Barnhart, Brengman, Li, & Peterson, 2018; He, Wen,

82 Xu, & Chen, 2018; Vajedian et al., 2018; Wanpeng et al., 2018; Yang et al., 2018) or used
83 far-field teleseismic data (Chen et al., 2018; Nissen et al., 2019), we jointly invert InSAR
84 and near-field strong-motion data which provide a better resolution (Anderson, 2003)
85 to propose a kinematic description of the earthquake source. We use a layered velocity
86 model that is routinely used to locate earthquakes by the IRSC (Iranian Seismological
87 Center), which ensures modelling is performed to the best of our knowledge (Supplemen-
88 tary Table T1).

89 **2 Inversion of co-seismic slip**

90 **2.1 Observations**

91 Due to the remote location of the event, the only available geodetic data come from
92 interferometric Synthetic Aperture Radar (InSAR). We use three SAR interferograms
93 computed from acquisition by the Sentinel-1 satellite, along two ascending and one de-
94 scending tracks (Figures 2a and S1-2). We use the ISCE software with precise orbits and
95 SRTM DEM to compute the co-seismic interferograms (Rosen, Gurrola, Sacco, & Ze-
96 bker, 2012). The coherence of the radar phase is excellent, likely due to the arid condi-
97 tions of this region. Acquisition dates are available in Table T2. We measure up to 80 cm
98 of ground displacement toward the satellite in the ascending tracks, suggesting uplift and/or
99 displacement toward the south-west. The number of data points in the unwrapped in-
100 terferograms is reduced using a recursive quad-tree algorithm (cf., Figure S1; Lohman
101 & Simons, 2005). We estimate uncertainties due to tropospheric perturbations in the phase
102 by estimating empirical covariance functions for each interferograms (Jolivet et al., 2014).
103 Estimated covariance parameters are summarized in Table T2.

104 We include near-field seismic waveforms recorded by 10 strong-motion accelerom-
105 eters from the Iran Strong Motion Network (ISMN) to constrain the temporal evolution
106 of slip during the earthquake rupture. Although located only on one side of the rupture,
107 all stations are within 102 km of the epicentre (c.f. Figure 2b). Details on strong mo-
108 tion data processing are given in Supplementary Text T1 (Ide, 2007; Lee & Lahr, 1972).
109 The east component of the two stations located south of the rupture (SPZ and GRS)
110 was not used due to the poor quality of the record. We integrate accelerometric data to
111 recover ground velocity, downsampled to 1 sps. Waveforms are bandpass filtered between

112 7 s and 50 s using a 4th order Butterworth band-pass filter, then windowed around the
 113 first arrivals.

114 **2.2 Estimation of the fault plane**

115 The two nodal planes of the global CMT mechanism (Ekström, Dziewonski, Mater-
 116 novskaya, & Nettles, 2005) are either a shallow north-east dipping plane (351° strike and
 117 11° dip) or a nearly vertical plane (121° strike and 83° dip). We conduct a grid-search
 118 on fault geometry parameters for each nodal plane. The goal is to discriminate between
 119 the two planes and to find the optimal fault geometry to limit forward modelling errors.

120 We grid-search the fault location and its strike and dip angles by inverting the In-
 121 SAR displacement to find the geometry that better explains the observations. For each
 122 tested geometry, slip is inverted on 96 subfault patches using a simple least-square tech-
 123 nique. More details on the method are given in Supplementary text T2 (Tarantola, 2005).
 124 We find that even the best sub-vertical plane has a RMS six times larger than the shallow-
 125 dipping plane (c.f. Figures S4 and S5). Although the sub-vertical plane is compatible
 126 with a back-thrust fault that may exist in the region (Tavani et al., 2018) or with the
 127 reactivation of steep normal faults (Jackson, 1980), the shallow dipping plane is in bet-
 128 ter agreement with receiver functions analysis (Paul, Hatzfeld, Kaviani, Tatar, & Péquegnat,
 129 2010) and the tectonic setting (e.g. Berberian, 1995; Vergés et al., 2011). Our optimal
 130 plane (351° strike, 14° dip, 13 km depth) agrees well with other studies using a similar grid-
 131 search approach (Barnhart et al., 2018; Wanpeng et al., 2018). In the following, we will
 132 consider that the Ezgeleh earthquake occurred on our optimum shallow dipping plane.

133 **2.3 Co-seismic slip modelling**

134 We use fault parameters inferred in section 2.2 to construct a planar fault and di-
 135 vide it in 96 subfault patches, each with a dimension of 7×7 km². Patch size was deter-
 136 mined through trial and error to limit correlation between slip on neighbouring parts of
 137 the fault. Source model parameters include total final slip, rupture velocity, and rise time
 138 for each patch along with hypocenter location. We define \mathbf{m}_S the vector including the
 139 two components of static slip (i.e. final integrated slip), and \mathbf{m}_K the vector of kinematic
 140 parameters describing the temporal evolution of slip.

141 We solve the problem in a Bayesian framework using AlTar, a Markov Chain Monte
 142 Carlo algorithm based on the algorithm described by Minson, Simons, and Beck (2013).
 143 It samples the full posterior probability distribution of the models that fit observations
 144 and are consistent with prior information. The strength of our solution is that it does
 145 not rely on any spatial smoothing and provides accurate estimates of the posterior slip
 146 uncertainty. We sample the posterior probability density $p(\mathbf{m}_S, \mathbf{m}_K | \mathbf{d}_S, \mathbf{d}_K)$ given by

$$147 \quad p(\mathbf{m}_S, \mathbf{m}_K | \mathbf{d}_S, \mathbf{d}_K) \propto p(\mathbf{m}_K) p(\mathbf{m}_S) p(\mathbf{d}_S | \mathbf{m}_S) p(\mathbf{d}_K | \mathbf{m}_S, \mathbf{m}_K) \quad (1)$$

148 where \mathbf{d}_S and \mathbf{d}_K are the InSAR and strong-motion observations, respectively. The prior
 149 PDFs $p(\mathbf{m}_S)$ and $p(\mathbf{m}_K)$ are mostly uniform distributions designed to prevent some model
 150 features such as back-slip. They are described in details in Table T3. For further details
 151 on the method, the reader can refer to Supplementary text T3, Duputel, Agram, Simons,
 152 Minson, and Beck (2014); Herrmann (2013); Minson et al. (2013) and Gombert et al. (2018).

153 **3 Results**

154 In the first seconds following the hypocentral time, slip propagates in every direc-
 155 tion around the hypocentre (c.f. Figure 3 and supplementary movie M1). Approximately
 156 5 seconds after origin, the rupture almost only propagates toward the south. The largest
 157 slip rate occurs roughly after 6 seconds, 20 km south of the epicentre. We observe a strong
 158 directivity toward the south, consistent with a shorter, higher amplitude signal at sta-
 159 tions SPZ and GRS compared to stations located in the north (c.f., Figures 2 and S3).
 160 In addition, we infer a large slip rate on the fault. As shown in Figures 4d-e and S6, slip
 161 rate increases up to more than 3 m/s where the slip is maximum. The slip rate functions
 162 of two fault patches presented here show the fast increase in slip rate associated with a
 163 short rise time ≤ 5 s, defining a sharp slip pulse (Heaton, 1990). Although larger than
 164 the values usually reported in kinematic slip models (usually ranging from 0.1 m/s to
 165 1 m/s), our slip rate estimates for this event are compatible with well documented earth-
 166 quakes (e.g., Cirella, Piatanesi, Tinti, Chini, & Cocco, 2012; Minson et al., 2014). The
 167 fast slip rate of the fault is reflected in the moment rate function (MRF) shown in Fig-
 168 ure 3c. To obtain the MRF, we first calculate the scalar moment function, $M_0(t)$, by sum-
 169 ming the moment tensor function of each subfault and using definition of the scalar mo-
 170 ment from Dahlen and Tromp (1998). The MRF is then obtained using the time-derivative
 171 of $M_0(t)$. As shown in Figure 3c, 90% of the moment was released within the first 14 sec-

172 onds of the rupture, depicting an overall impulsive earthquake. The mean rupture speed
173 along-strike is 3.0 ± 0.25 km/s (Figure S7), which is $\sim 0.9V_s$ at that depth.

174 The posterior mean model of the final cumulative slip is shown in Figure 4a. At
175 first order, this solution is in agreement with previously published static models (Barn-
176 hart et al., 2018; Wanpeng et al., 2018). We infer a ~ 50 km long and ~ 30 km wide rup-
177 ture, with a peak slip of $5.5 \text{ m} \pm 0.5 \text{ m}$. One difference arises as previous models proposed
178 that two distinct asperities ruptured during the earthquake. Our posterior mean model
179 does not show a clearly distinct rupture area in the north, closer to the hypocenter. How-
180 ever, roughly 20% of the models in our solution present such a feature (see Supplemen-
181 tary Movie M2). This indicates that it is in the realm of possibilities but available ob-
182 servations cannot entirely resolve it. The slip direction is constant along most of the fault,
183 with a $131.5^\circ \pm 0.8^\circ$ rake corresponding to a motion toward the south-west. The inferred
184 focal mechanism is therefore consistent with long-period moment tensor inversions.

185 Our Bayesian framework allows us to directly infer the posterior uncertainties as-
186 sociated with the model parameters. Slip uncertainties are represented in Figure 4a by
187 the 95% confidence ellipses. In addition, posterior marginal distributions after the static
188 and kinematic inversions of the along-rake slip of two fault patches are shown in 4b-c.
189 Unsurprisingly, the inclusion of kinematic observations reduces the posterior uncertain-
190 ties of those parameters. On the highest slipping patch for instance, the $1-\sigma$ posterior
191 uncertainty decreases from 0.82 m to 0.52 m. Over the fault, we observe a rather low pos-
192 terior uncertainty at shallow and intermediate depths, where slip is located. At depths
193 larger than 15 km, uncertainties become more significant. However, the inspection of each
194 model composing the solution reveals a good consistency in the slip distribution, with
195 nonetheless a larger variability in the northern part of the rupture (c.f., supplementary
196 movie M2). This is confirmed by our analysis of the model ensemble revealing that data
197 provides more information where the fault experienced large slip (cf., Supplementary text
198 T4 and Figure S8).

199 As shown in Figures S1, S2 and S9, model predictions fit Sentinel-1A observations
200 very well. Residuals are small over the three tracks, and they are consistent with the am-
201 plitude of 5-7 days of post-seismic signal (~ 10 cm; Barnhart et al., 2018). Stochastic model
202 predictions of the strong-motion data are shown in Figures 2 and S3. Overall, our so-
203 lution can explain the observations with a good accuracy. Posterior model predictions

204 of stations KAT, SNI and MHD suffer from larger uncertainties, likely explained by their
205 greater distance from the hypocenter.

206 **4 Discussion**

207 As suggested by previous studies (Barnhart et al., 2018; He et al., 2018; Wanpeng
208 et al., 2018), the Ezgeleh earthquake likely occurred on the Mountain Flexure Fault (some-
209 times referred as Main Front Fault, noted MFF in Figure 1). Along the major part of
210 the Zagros belt, the MFF follows a NW-SE axis with a $\sim 120^\circ$ azimuth and is aligned with
211 many topographic features (visible on the DEM in Figure 4). However, the strike of the
212 fault differs by about 50° with the topography orientation at the location of the earth-
213 quake. This discrepancy is explained by a major bend in the MFF at this location as
214 it transitions between the Lurestan Arc (LA) in the south and the Kirkuk Embayment
215 (KE) in the north (e.g., Koshnaw et al., 2017; Vergés et al., 2011). Interestingly, the fault
216 bend between the LA and KE corresponds to the northern bound of the rupture (Fig-
217 ure 3). This geometry change possibly stopped the rupture propagation, as suggested
218 by numerical models (Aochi, Fukuyama, & Matsu'ura, 2000). The rupture may also have
219 been halted by the 8 km to 10 km thick sediment cover, whose depth roughly corresponds
220 to the updip limit of slip. Although poorly constrained, these boundaries could nonethe-
221 less help to better assess the probable size for future large events in the region, a valu-
222 able element in seismic hazard assessment (e.g., Hetényi et al., 2016).

223 These sediments are heavily folded in the forearc basin and host many large an-
224 tyclines (e.g., Casciello et al., 2009; Kent, 2010). These folds are evidence for thin-skin
225 shortening occurring within the belt (Koshnaw et al., 2017). However, the slip of the 2017
226 earthquake occurred at larger depth, between 10 km and 15 km. This deeper co-seismic
227 deformation suggests that thick-skin shortening is also happening in this part of the Za-
228 gros range (Nissen et al., 2011; Vergés et al., 2011). The slip direction of the Ezgeleh earth-
229 quake on the MFF is nearly perpendicular to the alignment of the topographic features
230 mentioned above (cf., Figure 4a), creating a maximum 65 cm of uplift and 33 cm of sub-
231 sidence across the belt (c.f., Figure S10). Despite the relatively large depth of the Ezgeleh
232 earthquake, such co-seismic deformation may thus contribute to the growth of the Za-
233 gros topography. Afterslip might also contribute although it seems to occur on a shal-
234 low dipping decollement at the front of the mountain range (Barnhart et al., 2018).

235 An interesting feature of the Ezgeleh earthquake is the discrepancy between the
236 co-seismic slip direction and the current plate motion. Both the GSRM v2.1 model (Kreemer
237 et al., 2014) and the MORVEL model (DeMets et al., 2010) predict a nearly N-S plate
238 convergence (see Figure 1) while the overall co-seismic slip vector is oriented on a S 30° W
239 axis (see Figure 4).

240 This difference suggests that strain partitioning is occurring in this part of the Za-
241 gros belt, with a partial decoupling between the thrust and right-lateral strike-slip mo-
242 tion (McCaffrey, 1992; Platt, 1993). Strain partitioning in the Lurestan Arc and the Kirkuk
243 Embayment has been proposed before based on the analysis of regional focal mechanisms
244 (Talebian & Jackson, 2004). The Main Recent Fault (MRF; see Figure 1) is a major NW-
245 SE, 800 km long right-lateral strike-slip fault which accommodates some of the strain
246 (Tchalenko & Braud, 1974). It hosted several large earthquakes and has a ~50 km hor-
247 izontal offset (Talebian & Jackson, 2002). However, other structures may be accommo-
248 dating the strike-slip component of the convergence. Between July and November 2018,
249 three significant aftershocks with respective magnitudes of $M_W=5.8$, $M_W=6.0$, and $M_W=6.2$
250 occurred south of the mainshock epicenter (c.f. Figure 1b). These events present a right-
251 lateral strike-slip focal mechanism, but are located more than 100 km west of the MRF.
252 They could have ruptured the Khanaqin fault, a N-S strike-slip structure marking the
253 boundary between the Lurestan Arc and the Kirkuk Embayment (e.g., Berberian, 1995;
254 Blanc, Allen, Inger, & Hassani, 2003; Hessami, Koyi, & Talbot, 2001). However, there
255 is very limited evidence that the Khanaqin fault is actually a strike-slip fault. As a mat-
256 ter of fact, a recent study by Tavani et al. (2018) using reconstruction of seismic profiles
257 proposed that the Khanaqin fault is a back-thrust structure accommodating the SW-
258 NE motion. Therefore, undetected strike-slip faults may be accommodating some of the
259 strike-slip deformation closer to the forearc than the MRF. Throughout the Zagros, the
260 MRF is known to accommodate a major component of the northward motion via strain
261 partitioning (Talebian & Jackson, 2002). However, not all the regional shear is accom-
262 modated solely on the MRF, especially along its northwest limit, where strike-slip fault-
263 ing becomes increasingly distributed (Copley & Jackson, 2006). The additional contri-
264 bution from other less-known active structures in this region therefore poses an impor-
265 tant uncharacterized seismic risk for local populations.

266 While the rupture directivity imaged here is visible in published slip models (Chen
267 et al., 2018; Nissen et al., 2019), the sharpness of the slip pulse in these studies is blurred

268 by their use of far-field data and smoothing constraints. The good spatial and tempo-
269 ral resolution of our kinematic slip model enabled by the use of near-field observations
270 reveals interesting features. Figures 3 and S6 show that the rupture starts as a growing
271 crack that rapidly transition into a pulse with a rise time of about 4 sec. This crack-pulse
272 transition occurs within the first four seconds and less than 7 km from the hypocenter,
273 therefore away from the rupture boundaries. Day (1982) proposed that if the slip rate
274 is controlled by the fault geometry the rise time should be $T_R = 0.5 \times W/V_R$, with W
275 the fault width and V_R the rupture velocity. However, we infer T_R values systematically
276 higher than inverted rise times (Figure S11). This pulse-like behaviour is therefore un-
277 likely to result from healing phases emanating from the along-dip finiteness of the fault.
278 A rapid crack-pulse transition is in agreement with early observations by Heaton (1990)
279 and later studies (e.g., Beroza & Mikumo, 1996; Meier, Heaton, & Clinton, 2016; Wang,
280 Day, & Denolle, 2019). Such a pulse may result from a number of mechanisms such as
281 frictional self-healing, fault strength or stress heterogeneities, bimaterial effects and wave
282 reflections within low-velocity fault zones (e.g., Andrews & Ben-Zion, 1997; Huang & Am-
283 puero, 2011; Perrin, Rice, & Zheng, 1995). After this early transition from a growing crack,
284 the rupture continues its journey along-strike as a decaying pulse toward the north, and
285 a strong growing pulse toward the south.

286 This strong southward propagating pulse seems to have a significant impact in the
287 distribution of damage and landslides triggered by the earthquake. The Ezgeleh earth-
288 quake induced extensive destructions of dwellings in Iraqi Kurdistan, but mostly in the
289 Iranian province of Kermanshah. Figure 1b) shows the intensity of damage created by
290 the mainshock. It is obtained from field observations conducted by the International In-
291 stitute of Earthquake Engineering and Seismology of Iran (IIEES). Reported macroseis-
292 mic intensities are also shown in Figure S12. Damage intensity roughly follows the sur-
293 face projection of the slip distribution, but larger damage was reported in the south. In
294 addition to building damage, many rockfalls and landslides occurred south of the rup-
295 ture and up to 125 km from the centroid, including a large 4 km long and 1 km wide land-
296 slide (Miyajima et al., 2018). These observations may however be biased by the difficulty
297 to report such phenomena on the Iraqi side of the border. Many different factors can also
298 largely influence the effects of an earthquake, like soil nature or mountain slopes. In ad-
299 dition to rupture directivity, studies have suggested that the strong impulsiveness of the
300 source can intensify low-frequency ground shaking, particularly damaging to buildings

(Melgar & Hayes, 2017; Somerville & Graves, 2003). The large slip-rate and short rise-time of the southward propagating pulse may therefore have exacerbated the damage observed south-west of the Ezgeleh earthquake.

5 Conclusion

The 2017 Ezgeleh earthquake breaks a long hiatus on strong events affecting the Zagros thrust and fold belt in the Kermanshah province. The joint inversion of InSAR and near-field strong-motion observations reveals a predominantly thrust motion on a near-horizontal blind crustal fault. We also infer a highly impulsive source propagating toward the south. These kinematic properties may have played a role in the numerous slope instabilities and in the important damage that affected Iranian cities.

Furthermore, the misalignment between the plate convergence and the slip direction provide additional hints for a strain partitioning in this part of the Zagros belt between thrust motion on flat crustal faults and right-lateral strike-slip. As suggested by late aftershocks, unmapped dextral faults could be accommodating part of that shear strain, and therefore represent an important seismic risk for nearby populations.

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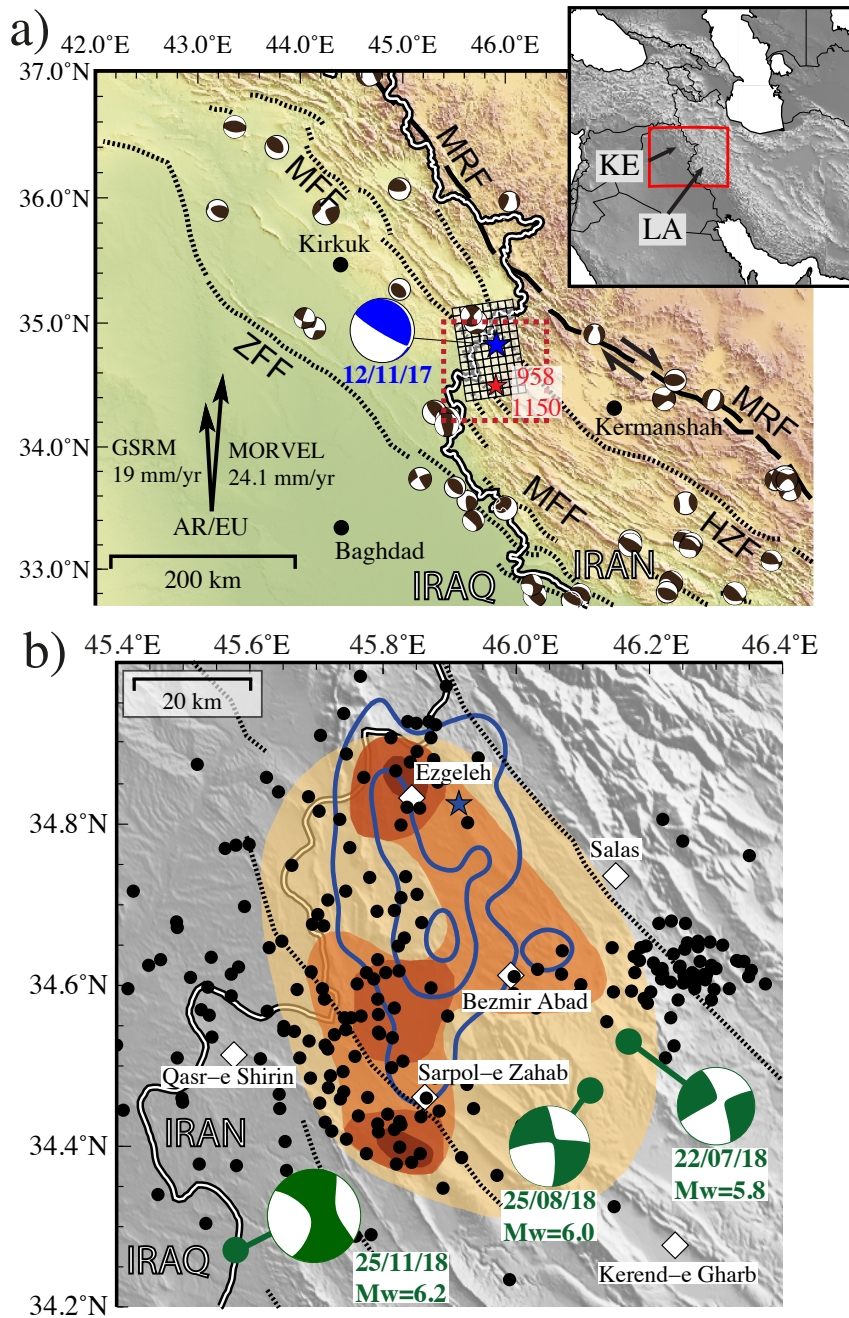
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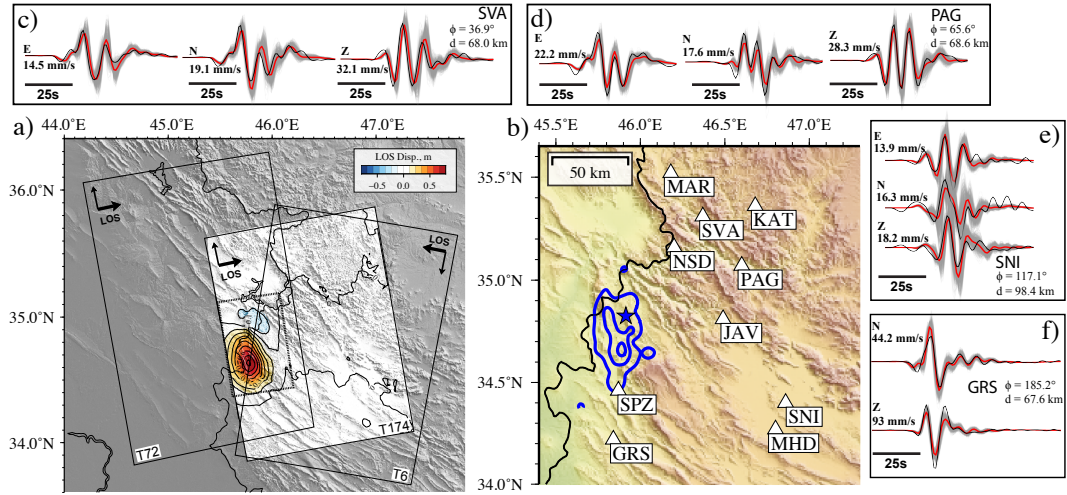
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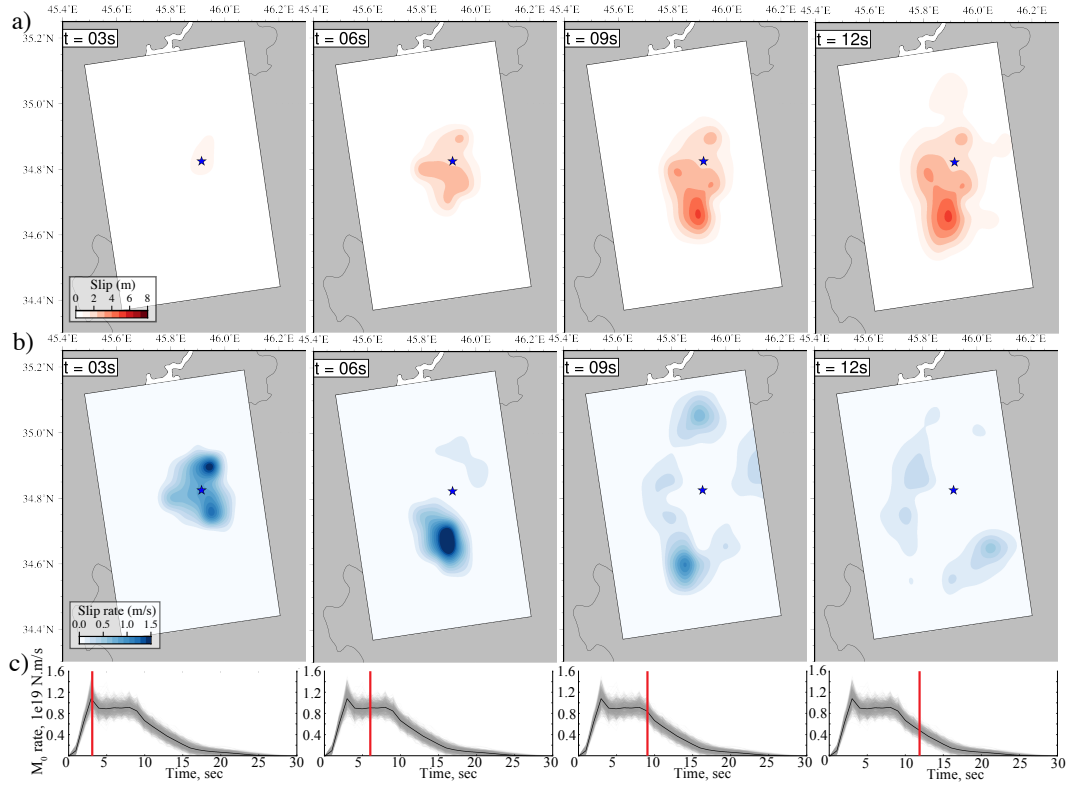
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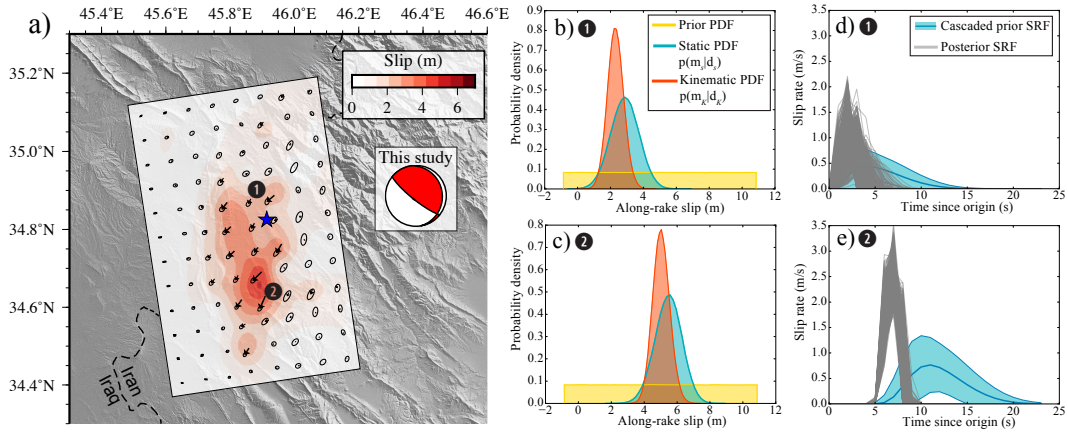
529 **Figure 1. Regional seismotectonic context and damage associated with the 2017 Ezgeleh earth-**
 530 **quake. a)** Blue star marks the epicentre location, and the squares represent the fault parametrisation. Blue
 531 beachball is the moment tensor estimated in this study and brown moment tensors are regional seismicity from
 532 the Global CMT catalogue (Ekström et al., 2012). Red star shows the approximate location of two historical
 533 earthquakes. Dashed black line is the Main Recent Fault (MRF) and dotted lines are supposed location of
 534 regional blind faults (MFF: Mountain Flexure Fault; HZF: High Zagros Fault; ZFF: Zagros Foredeep Fault;
 535 Berberian, 1995). Arrows indicate the convergence of the Arabian plate (AR) with respect to stable Eurasia
 536 (EU) from the GSRM v2.1 (Kreemer et al., 2014) and MORVEL (DeMets et al., 2010) models, computed with
 537 the UNAVCO Plate Motion Calculator. LA: Lorestan Arc. KE: Kirkuk Embayment. Red dashed rectangle
 538 indicates position of b). **b)** Black dots are aftershocks located by the International Institute of Earthquake
 539 Engineering and Seismology of Iran (IIEES). Focal mechanisms from the Global CMT catalogue of three large
 540 aftershocks are shown in green. Brown colours indicate the level of damage based on a compilation of destruc-
 541 tion rate and landslide activity interpolated from field surveys conducted by the Geological Survey of Iran (GSI,
 542 2017). The darker the colour, the more intense the damage. Blue lines are the 1.5 m co-seismic slip contour.



543 **Figure 2. Observations used in the inversion.** a) Unwrapped Sentinel-1A interfer-
 544 ograms showing surface displacement in LOS direction (Track 174). The footprint of one
 545 additional ascending and descending tracks are also shown. Data, predictions and model per-
 546 formance of the 3 interferograms are available in Figures S1-2. b) Location of strong-motion
 547 records (white triangles). c-f) Waveforms of four selected station around the epicenter. For each
 548 waveform, the bold number indicates its maximum amplitude. Φ and d are station azimuth
 549 and distance to epicentre, respectively. The black line is the recorded waveform, grey lines are
 550 stochastic predictions for our posterior model, and the red line is the mean of stochastic predic-
 551 tions. Remaining waveforms are shown in Figure S3.



552 **Figure 3. Temporal evolution of co-seismic slip.** a) Cumulative slip on the fault 3 s, 6 s,
 553 9 s, and 12 s after the origin time. The red colour-scale indicates slip amplitude. b) Evolution of
 554 slip rate on the fault. c) Source time function (STF) of the event. Grey lines are stochastic STFs
 555 inferred from our model population while the black curve represents the posterior mean STF.
 556 Vertical red lines indicate the time of each snapshot.



557 **Figure 4. Final co-seismic slip distribution** a) Colour and arrows on the fault plane
 558 indicate amplitude and direction of slip, respectively. Ellipses represent the 95% posterior un-
 559 certainty. Results presented in subfigures b-e) are obtained for patches labelled 1 and 2. The
 560 background topography comes from the Shuttle Radar Topography Mission (SRTM; Farr et
 561 al., 2007). b-c) Prior, posterior static PDF, and posterior kinematic PDF of along-rake slip in
 562 patches 1 and 2. d-e) Slip rate evolution in patches 1 and 2. Blue line is the mean prior Slip
 563 Rate Function (SRF) used in the sampling, surrounded by 1- σ uncertainties. Posterior SRFs in
 564 grey are from 1000 thousands models randomly selected from our solution.

Figure 1.

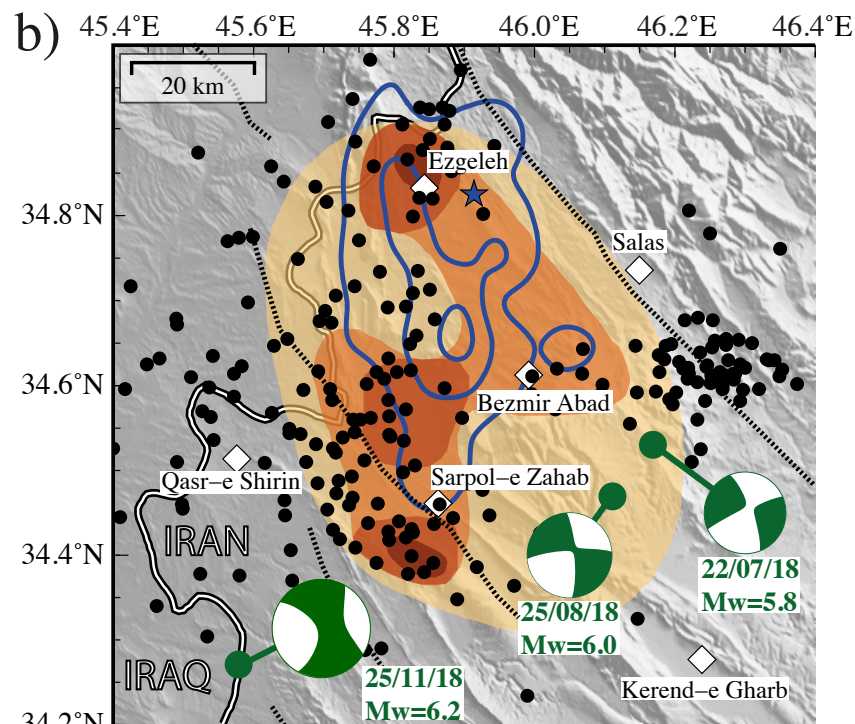
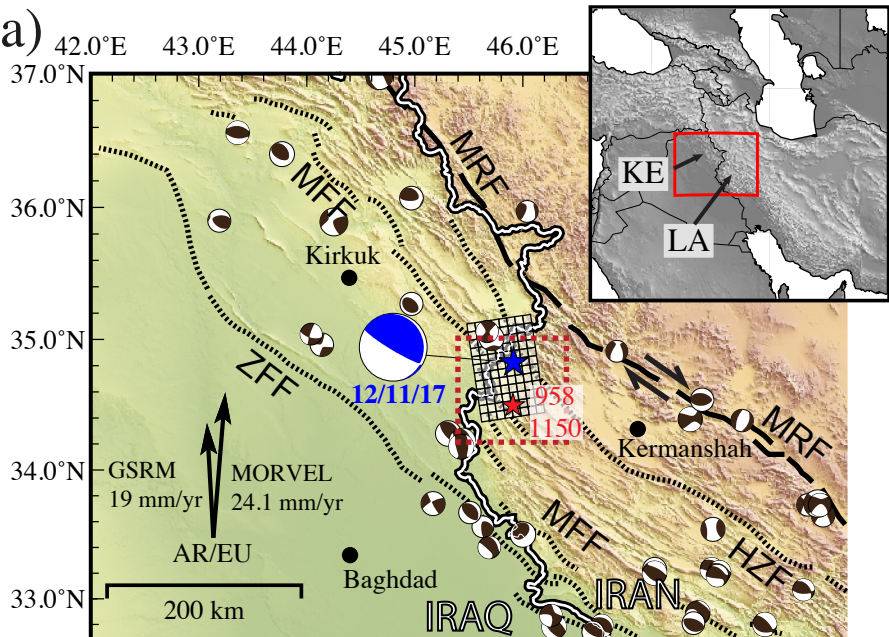


Figure 2.

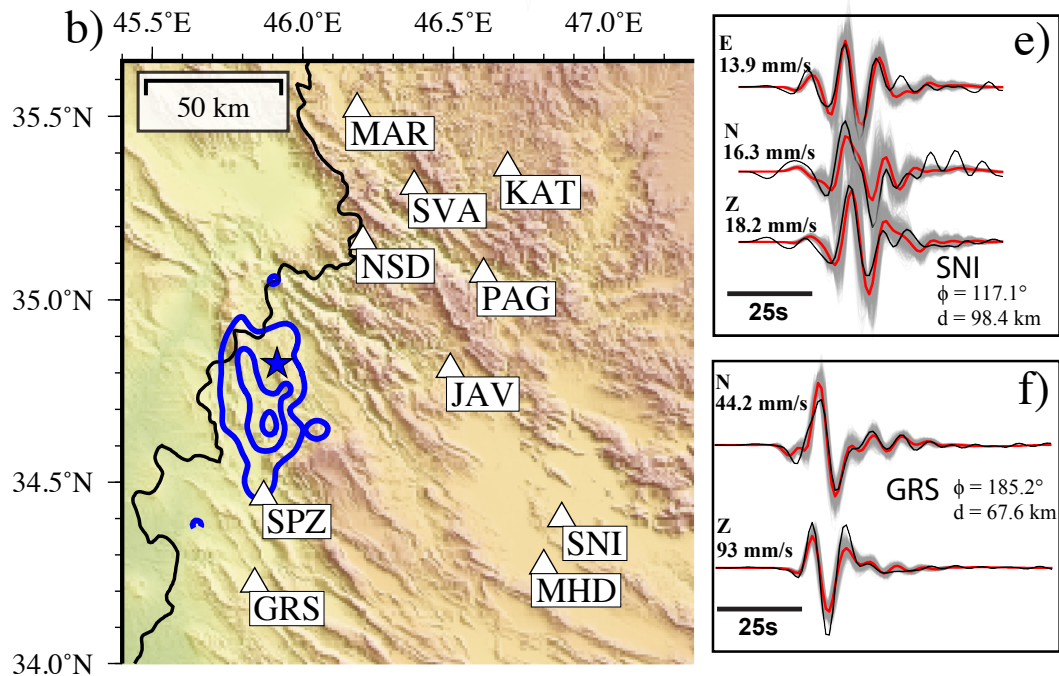
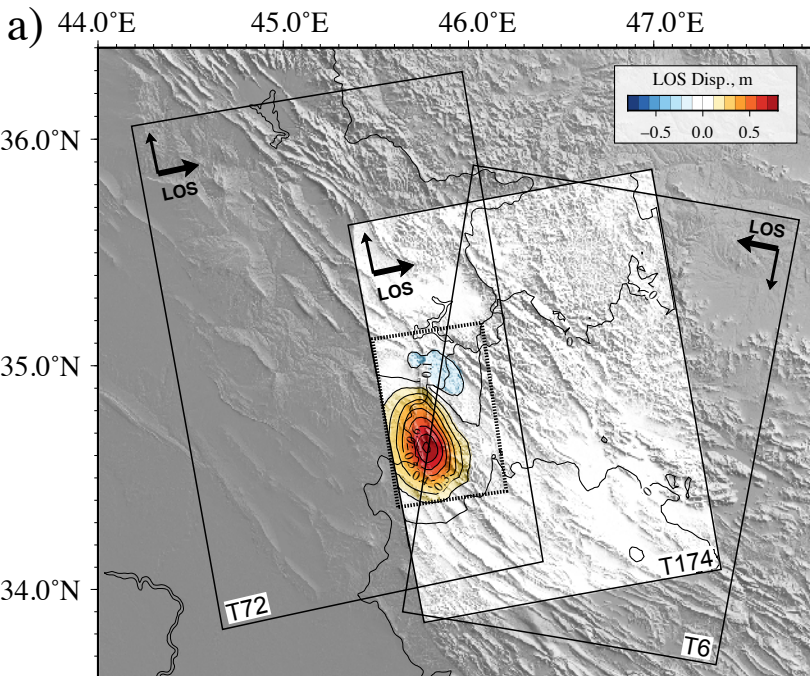
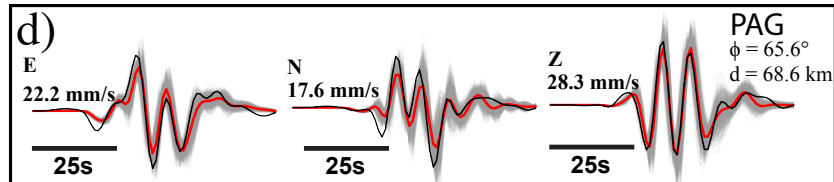
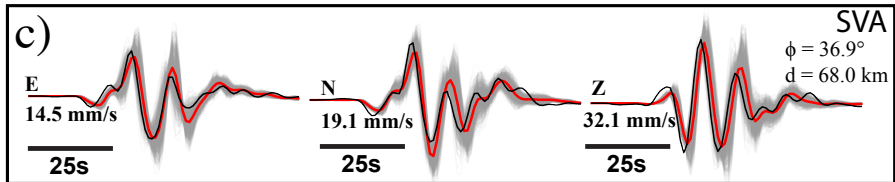


Figure 3.

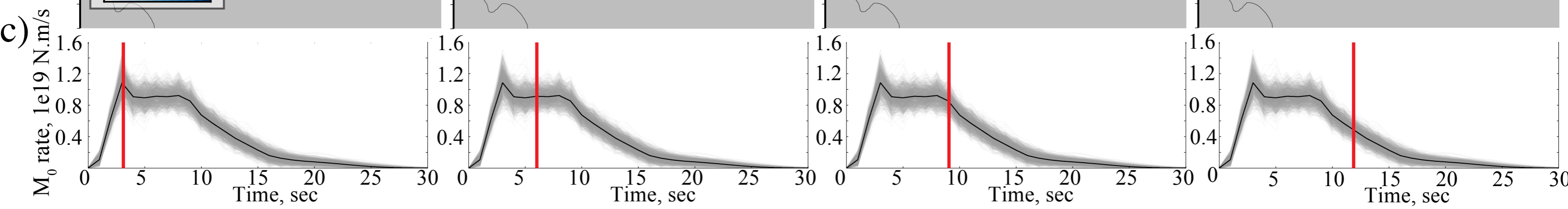
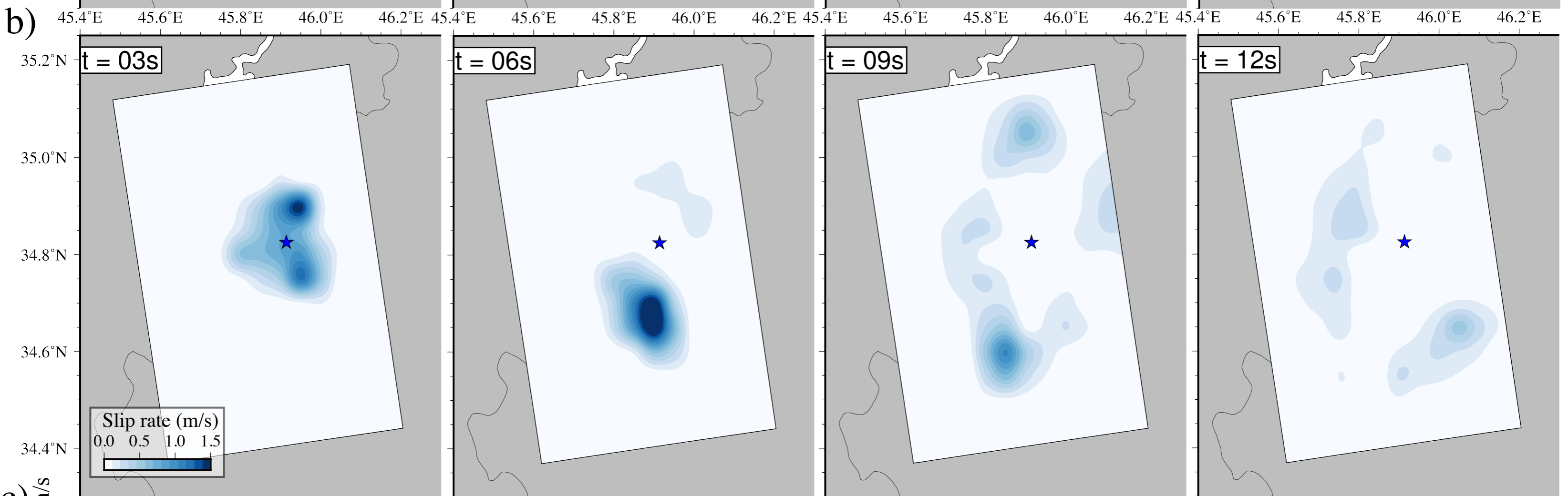
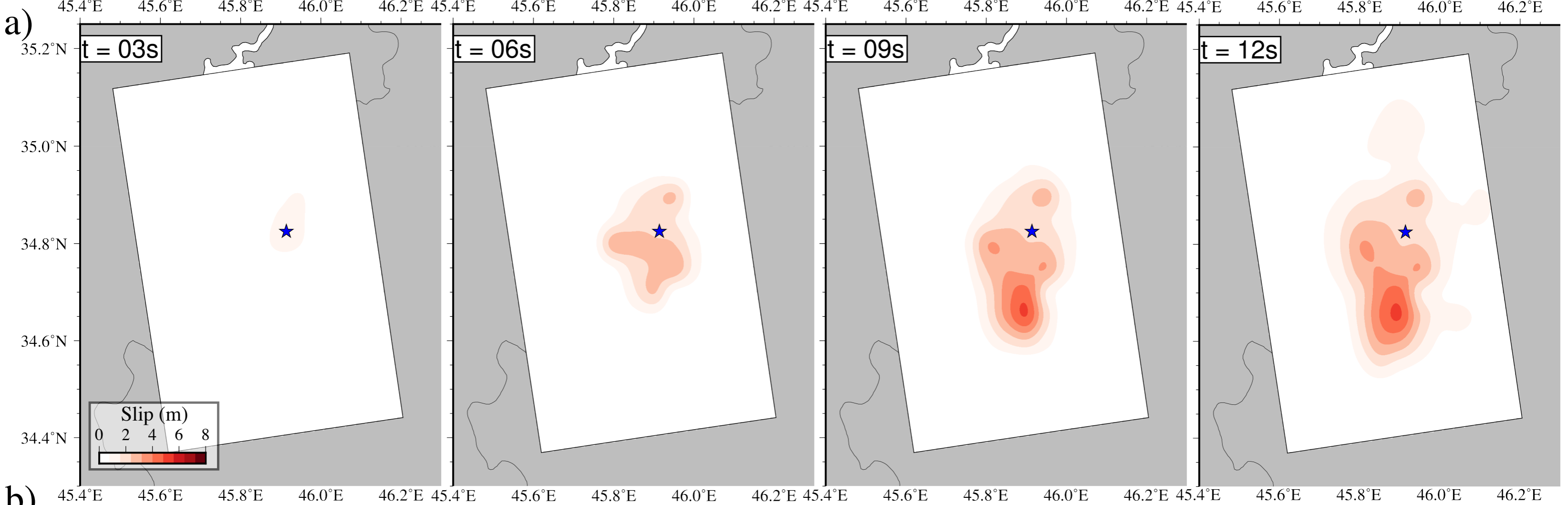


Figure 4.

