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Time-dependent postseismic slip following the 1978 M_w 7.3 Tabas-e-Golshan, Iran earthquake revealed by over 20 years of ESA InSAR observations

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Abstract

We use over 20 years (1996–2017) of the European Space Agency’s (ESA) radar interferometry (InSAR) observations to investigate the postseismic deformation of the Tabas fold segment following the 1978 M_w 7.3 Tabas-e-Golshan earthquake in eastern Iran. We generated maps of satellite line-of-sight (LOS) velocity using two ERS descending tracks (1996-1999), one Envisat descending track (2003-2010), one Sentinel-1A descending track (2014-2017) and one Sentinel-1A ascending track (2014-2017). The LOS velocity shows afterslip continuing for at least 40 years after the earthquake. Elastic dislocation modelling based on the InSAR measurements reveals a decrease in postseismic velocities from $\sim 4.2\text{--}5.0$ mm/yr in 1996-1999 to 3.9 ± 0.6 mm/yr in 2003-2005, to 3.0 ± 0.4 mm/yr in 2006-2010, and to a present rate of 2.3-2.7 mm/yr in 2014-2017. The rates are consistent with our previous empirical model (afterslip rate $v = 88t^{-0.98}$ mm/yr), suggesting a $\sim 1/t$ decay in postseismic velocities. We then combine the InSAR rates and our previous estimates of the total earthquake slip derived from optical image matching and DEM differencing to explore the frictional behaviour of the Tabas fold. We obtained a rate-and-state parameter $a - b$ ranging from 0.063 to 0.0, indicating rate-strengthening frictional behaviour of the Tabas fault. We also inferred a minimum value of 4.7 m for coseismic slip, which fits the field measurements of the surface displacements (at least 1.5 m vertical motion by summing up the bedding plane slips). The results imply that coseismic slip and afterslip can occur at the same place. One possible mechanism to explain such phenomenon is that the frictional parameter $a - b$ is small enough to allow dynamic ruptures to propagate in rate-strengthening regions. The other possibility is that preseismic rate-weakening behaviour may transit to postseismic rate-strengthening due to coseismic shear heating; this mechanism questions the common assumption that rate-and-state frictional

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28 properties of faults do not change on the scale of available geodetic observations.

29 *Keywords:*

30 InSAR, Sentinel-1, Time-dependent postseismic slip, 1978 Tabas-e-Golshan earthquake,

31 Rate-and-state friction

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32 1. Introduction

33 Recent advances in seismic and geodetic techniques have enabled us to document the broad
34 variability in slip behaviours of active faults that have caused large earthquakes (e.g., *Avouac*,
35 2015, *Miyazaki et al.*, 2011, *Obara*, 2002, *Segall et al.*, 2000, *Thomas et al.*, 2014a). Within the
36 seismogenic depth, fault slip may be dominantly seismic in some places, i.e., displacement occurs
37 during earthquakes (slip rate of m/s), and dominantly aseismic in others as a result of steady
38 (mm/yr to cm/yr) or transient creep ($\mu\text{m/s}$ to mm/yr). In order to fully understand active
39 faulting, it is therefore of prime importance to measure deformation accurately throughout the
40 entire earthquake cycle.

41 Laboratory experiments suggest that, depending on the rock type, effective normal stress and
42 temperature, fault can exhibit either a rate-strengthening behaviour, producing stable sliding, or
43 a rate-weakening behaviour, leading to “stick-slip” motion, (e.g. *Dieterich*, 1979, *Marone*, 1998,
44 *Ruina*, 1983). In the framework of rate-and-state friction, earthquakes can only nucleate in rate-
45 weakening regions, whereas rate-strengthening regions are dominantly characterised by aseismic
46 creep. However, the complex behaviour of faults not only depends on the frictional properties of
47 the fault interface, but also arises from the interaction of dynamic stress transfer, levels of prestress,
48 and the evolution of the elastic properties of the surrounding medium by coseismic damage (e.g.,
49 *Hillers et al.*, 2006, *Kaneko et al.*, 2010, *Kato*, 2004, *Noda and Lapusta*, 2013, *Thomas et al.*,
50 2017a). For example, although laboratory experiments suggest a rate-strengthening frictional
51 behaviour at shallow depths (*Blanpied et al.*, 1995, *Kohlstedt et al.*, 1995), some active faults
52 seem to be locked near the surface and slip seismically. A striking case is the 2011 Tohoku-Oki
53 earthquake for which a large coseismic slip has been observed on the fault segments that were
54 thought to be rate-strengthening (*Noda and Lapusta*, 2013). Numerical models suggest that rate-
55 strengthening regions can be locked due to elastic interactions with the adjacent rate-weakening
56 regions, accumulating stresses during the interseismic period, which allows seismic rupture to
57 propagate through (*Kaneko et al.*, 2010, *Thomas et al.*, 2014b).

58 To better understand the frictional behaviour of faults in nature, it is necessary to combine
59 geodetic observations and physical models of fault slip. In this paper, we use radar interferometry

(InSAR) to study the shallow afterslip following the 1978 M_w 7.3 Tabas-e-Golshan earthquake in eastern Iran. Previous studies based on the European Space Agency’s (ESA) ERS and Envisat data (Copley, 2014), and optical imagery (Zhou *et al.*, 2016) have found long-lasting afterslip almost 40 years after the main shock. The rate is suggested to decay as $\sim 1/t$ (Zhou *et al.*, 2016). Here, we present deformation time series between 1996 and 2017 derived from ESA’s ERS (1996-1999), Envisat (2003-2010) and Sentinel-1A (2014-2017) data. The long time series allows us to explore fault frictional properties based on laboratory-derived rate-and-state friction laws. To conclude, we discuss the possible mechanisms for seismic/postseismic overlap as observed for the 1978 Tabas-e-Golshan earthquake.

2. Overview of the 1978 Tabas-e-Golshan earthquake

Iran is very seismically active due to the convergence between the Arabian and Eurasian plates (Figure 1a). The convergence rate is estimated to be ~ 25 mm/yr at longitude 60° (Vernant *et al.*, 2004). Six $M_w \geq 7.0$ earthquakes have been recorded in Iran since 1976 (from the GCMT catalogue) (Ekström *et al.*, 2012), five of which are in the east (Figure 1b). The 16 September, 1978 Tabas-e-Golshan earthquake (hereafter referred to as the 1978 Tabas earthquake) is the deadliest among them, killing $\sim 11,000$ people (Berberian, 1979). It occurred at the northern end of the Nayband fault (Figure 1b) where north-south right-lateral strike-slip motion turns into a zone of distributed thrusting that is accommodated on a series of folds within Neogene basin deposits (Figure 1c) (Zhou *et al.*, 2016).

To study the 1978 earthquake, Walker *et al.* (2003) analysed P and SH seismic waveforms, and found that the main shock occurred on a shallow (16°) dipping thrust with a centroid depth of ~ 9 km. Applying the fault scaling laws of Scholz (1982), Walker *et al.* (2003) calculated an average slip of ~ 3.3 m on the shallow dipping thrust. On the surface, the maximum coseismic displacement on the frontal thrust was found at the western margin of the Kurit fold segment and measured to be ~ 35 cm (vertical) (Berberian, 1979). Berberian (1979) also found extensive bedding plane slips along the 85 km-long system of folds (Figure 1c), with an overall vertical throw of ~ 1.5 m in a 5 km-wide zone. Using ERS and Envisat data, Copley (2014) found long-lasting postseismic slip on a steep ramp beneath the Tabas fold (hereafter referred to as the “Tabas ramp”) 30 years

88 after the main shock. The rate of afterslip was estimated to be ~ 5 mm/yr in the time interval of
89 1996–1999 from modelling ERS observations (*Copley, 2014*).

90 Recently, *Zhou et al. (2016)* used both historical and modern optical imagery to measure the
91 surface deformation associated with the 1978 earthquake. Image correlations and DEM differencing
92 revealed a total earthquake slip of ~ 7 m on the Tabas ramp between 1978 and 2013, 6.5 m of
93 which occurred between 1978 and 1991 (*Zhou et al., 2016*). Using *Zhou et al. (2016)*’s optical
94 measurements and *Copley (2014)*’s InSAR rate (5 mm/yr between 1996 and 1999), *Zhou et al.*
95 (2016) derived an empirical power law model, $s = 4.4t^{0.02}$, to relate the cumulative afterslip s
96 (metres) and time t (years after the main shock). This simple empirical model shows that the rate
97 of afterslip ($v = \frac{ds}{dt} = 88t^{-0.98}$ mm/yr) decays as $\sim 1/t$, which conforms to “Omori-like decay” of
98 postseismic velocities as discussed by *Ingleby and Wright (2017)*.

99 Models of fault slip based on both the optical image matching and InSAR results suggest that
100 the Tabas ramp does not reach the surface as a single main fault, but a discrete plane from a depth
101 of 0.4 ± 0.2 km (detailed in Section 4). This conforms to the geological observations of the width
102 of the deformation at the surface, i.e. ~ 220 m wide zone of bedding plane slip seen in the field
103 (*Berberian, 1979*) and optical imagery (*Zhou et al., 2016*). Bedding plane slips are complicated
104 because the motion at sub-surface is not planar, and may therefore be largely “invisible”, which
105 creates great challenges to field investigations of earthquake deformation. Nonetheless, they are
106 very common in blind-thrust fault systems. For example, *Philip and Meghraoui (1983)* found
107 a 200-300 m wide zone of bedding plane slip in the 1980 El-Asnam earthquake (similar to the
108 Tabas earthquake); *Berberian et al. (1992)* observed similar coseismic flexural-slip faulting on
109 bedding planes during the 1990 M_w 7.3 Rudbar-Tarom earthquake. Field-based studies could have
110 underestimated the coseismic slip of these earthquakes. In the following sections, we combine
111 geodetic observations and geophysical models, trying to provide a possible means of re-examining
112 the coseismic slip of some older earthquakes similar to the 1978 Tabas earthquake.

113 3. InSAR measurements of postseismic surface deformation

114 *Copley (2014)* stacked ERS (1996-1999) and Envisat (2003-2010) interferograms to measure
115 the postseismic deformation. In this paper, we reprocessed the ERS and Envisat data and used

the Small Baseline Subset (SBAS) method (*Berardino et al.*, 2002) to construct deformation time series. *Zhou et al.* (2016)’s empirical model of afterslip decay ($v = 88t^{-0.98}$ mm/yr) predicts the rate of afterslip between 2014 and 2017 to be ~ 2.6 mm/yr. In order to test and improve this empirical model, we also collected Sentinel-1A data to measure the actual rate during this time period. All the interferograms were made and unwrapped using the GAMMA software (*Werner et al.*, 2000). The ERS and Envisat interferograms were multi-looked with a factor of 5 in range and 25 in azimuth, and the Sentinel-1A interferograms with a factor of 20 in range and 4 in azimuth. We discarded interferograms with a perpendicular baseline of larger than 300 m (Figure 2). We then used the Generic InSAR Analysis Toolbox (GIAnt) (*Agram et al.*, 2013) to build the deformation time series.

Note that the deformation time series revealed afterslip on a series of folds that ruptured in the 1978 earthquake (not only the Tabas fold), but here we mainly focus on the Tabas ramp where we can combine *Zhou et al.* (2016)’s estimates of the total earthquake slip to investigate the frictional properties. The InSAR-derived postseismic deformation on other folds has been modelled and discussed in detail by *Copley* (2014). Note also that the measured surface deformation is very little sensitive to the motion on the shallow dipping thrust at depth (*Zhou et al.*, 2016), so the deformation time series essentially shows the motion on the ramp.

3.1. ERS: 1996-1999

The ERS data were collected from two descending tracks (435 and 206). Although the ERS acquisitions are very limited (Figure 2), fortunately all the interferograms (2 from ERS track 435 and 4 from 206) show a very distinctive tectonic signal across the fault. For ERS track 435, we simply stacked the two interferograms to get the rate map in Figure 3a. Swath profile A-A’ (2 km wide) shows that the average rate between 1996 and 1999 in the satellite line-of-sight (LOS) direction is ~ 2.7 mm/yr (Figure 3c). For the four interferograms from track 206, we applied empirical tropospheric corrections, removed orbital errors, and used the SBAS method to obtain the rate map as shown in Figure 3b. The average LOS rate is ~ 2.5 mm/yr, similar to that from track 435 and *Copley* (2014)’s estimate (2.4 mm/yr at roughly the same location).

143 3.2. *Envisat: 2003-2010*

144 We collected 22 acquisitions from Envisat descending track 435, and generated 462 interfero-
145 grams. For the time series analysis, we selected 43 interferograms (Figure 2) that show an evident
146 tectonic signal across the fault. After applying atmospheric and orbital corrections, and using the
147 SBAS method, we derived two rate maps (Figure 4). We separated the Envisat data (2003-2010)
148 into two periods (2003-2005 and 2006-2010) considering the facts that (1) we would need 2-3 years
149 of data for the small tectonic signal (on the order of 1-2 mm/yr) to exceed the non-tectonic noises,
150 and (2) there is a large acquisition gap in the data set between 2006 and 2009, making it difficult
151 to estimate the slip rate in 2007-2008. Swath profile A-A' in Figure 4 reveals an average LOS rate
152 of 2.2 mm/yr in 2003-2005, and 1.6 mm/yr in 2006-2010. The rates, though in the LOS direction,
153 show a decay with time.

154 3.3. *Sentinel-1A: 2014-2017*

155 Sentinel-1A was launched on the 3rd April, 2014; the first image over the Tabas area was
156 acquired in October 2014. We collected 38 acquisitions from descending track 166, and 32 from
157 ascending track 159. We formed interferograms by pairwise combination using the COMET InSAR
158 processing software LiCSAR (*Spaans et al.*, 2017), which is built around the GAMMA InSAR
159 software package. It is worth noting that although most Sentinel-1 interferograms have a spatial
160 baseline of less than 100 m, care needs to be taken when using the data because the wider coverage
161 of Sentinel-1 may also result in more “irregular”, short-wavelength atmospheric signals within
162 one scene. From our first inspection of the preliminary interferograms, we found that many of
163 the acquisitions were heavily contaminated by atmospheric delays (ionosphere in particular). The
164 errors are shown by the same fringe patterns in the interferograms with those acquisitions as either
165 the master or slave image. Since we are attempting to measure very small tectonic deformation,
166 we discarded all the contaminated acquisitions in order to minimise the influence of atmosphere.
167 We selected 20 (out of 38) acquisitions from descending track 166 and 21 (out of 32) from track
168 159 to form the networks in Figure 2 for time series analysis. Figure 5 shows the deformation time
169 series based on the Sentinel-1A data. Swath profiles in Figures 5c-d reveal 1.0 mm/yr (descending
170 track 166) and 1.7 mm/yr (ascending track 159) afterslip in the LOS direction between 2014 and

171 2017. The LOS rate derived from the descending track is smaller than the ascending track because
172 the projections of the horizontal and vertical motions onto the LOS direction of the descending
173 track are in the opposite sense and will cancel out, whereas for the ascending track, they add up.

174 4. Models of fault slip

175 InSAR measurements of surface deformation are in the satellite LOS direction (Figures 3-5), so
176 in order to compare all the rates of afterslip derived from different satellites and orbits, we need to
177 calculate the slip on the fault plane. To do this, we used an elastic dislocation approach (*Okada*,
178 1985) that assumes constant slip on a rectangular plane. We selected profile A-A' in the modelling
179 so as to avoid the strong subsidence signal in the north (see Figure 3), which is possibly caused by
180 groundwater pumping. The free parameters are: slip rate, the dip of the fault, the depth to the
181 top and bottom of the fault plane that has slipped. Because we do not have any reference point
182 in the LOS measurements, a linear trend (an offset of the data relative to zero and a gradient
183 along the profile) also needs to be calculated. The strike and rake were fixed to 2° and 90° (i.e.
184 pure thrust motion). We performed a grid search through the free parameters: varying the slip
185 rate from 0-10 mm/yr in steps of 0.1 mm/yr, the dip from $0-90^\circ$ in steps of 5° , the depth to the
186 top of the fault plane from 0-1.0 km in steps of 0.1 km, and the depth to the bottom from 4-10
187 km in steps of 1 km. In the calculations, we noticed a trade-off between the fault slip and dip;
188 this is because InSAR measures deformation in the LOS direction only, and cannot distinguish
189 between the horizontal and vertical components. Given that the fault dip was well constrained
190 by *Zhou et al.* (2016)'s optical image matching and DEM differencing results, we fixed the dip
191 of the Tabas ramp to 50° (also consistent with *Copley* (2014)) to obtain a more robust estimate
192 of the slip rate. We repeated the grid search and computed the residuals between each Okada
193 model and the measured profile. We fitted a linear trend to the residuals and removed it from
194 the measured profile, and then recalculated the residuals between the Okada models and the new
195 profile. The procedure was undertaken for all the profiles shown in Figures 3-5. The solutions with
196 the minimum misfit (the standard deviation of the residuals, RMSE) are shown in Figure 6 with
197 the parameters listed in Table 1.

198 All the best-fitting models in Table 1 gave a bottom depth of 5 ± 1 km, consistent with 4–5 km
 199 in *Copley* (2014) and 6 km in *Zhou et al.* (2016). The top depth, 0.4 ± 0.2 km, although slightly
 200 larger than 0.1 km in *Zhou et al.* (2016), confirms that the deformation is distributed near the
 201 surface (*Zhou et al.*, 2016). The fault slip rates are the most well-constrained parameters in the
 202 inversions. The ERS rates between 1996 and 1999 from different tracks, 5.0 ± 0.8 mm/yr (ERS
 203 D435) and 4.2 ± 0.5 mm/yr (ERS D206), agree well with *Copley* (2014)’s estimate (5 ± 1 mm/yr).
 204 The afterslip rate shows an evident decrease from ~ 4.5 mm/yr in 1996–1999, to a present rate of
 205 ~ 2.5 mm/yr.

206 Based on the empirical model of afterslip decay ($v = 88t^{-0.98}$ mm/yr) from *Zhou et al.* (2016),
 207 we can calculate the average afterslip rate in 2003–2005 (3.6 mm/yr), 2006–2010 (3.1 mm/yr), and
 208 2014–2017 (2.6 mm/yr) respectively. These estimates are solely based on the optical image match-
 209 ing results and *Copley* (2014)’s ERS measurements, and are therefore completely independent
 210 from the Envisat and Sentinel-1A measurements. Nonetheless, we found good agreement between
 211 them: 3.6 mm/yr (predicted) versus 3.9 mm/yr (measured) in 2003–2005, 3.1 mm/yr (predicted)
 212 versus 3.0 mm/yr (measured) in 2006–2010, and 2.6 mm/yr (predicted) versus 2.7 (and 2.3) mm/yr
 213 (measured) in 2014–2017. We attempt to use the new deformation time series to refine our previous
 214 empirical model.

215 5. Discussion

216 In this section, we combine our InSAR time series of afterslip and *Zhou et al.* (2016)’s image
 217 matching results of the earthquake slip to investigate the “rate-and-state” nature of fault friction.

218 5.1. Rate-and-state friction laws and the analytical solutions

219 Laboratory-derived rate-and-state friction laws (*Dieterich*, 1979, *Marone*, 1998, *Ruina*, 1983)
 220 have been extensively used to reproduce fault slip behaviour (e.g., *Avouac*, 2015, *Hillers et al.*,
 221 2006, *Kato*, 2004, *Noda and Lapusta*, 2013, *Perfettini and Avouac*, 2004, *Thomas et al.*, 2017b).
 222 Here we adopt the model with the aging law proposed by *Dieterich* (1979) and *Ruina* (1983):

$$223 \quad \tau_f = \bar{\sigma} \left[\mu_* + a \ln \left(\frac{V}{V_*} \right) + b \ln \left(\frac{\theta V_*}{D_c} \right) \right] \quad (1)$$

$$\frac{d\theta}{dt} = 1 - \frac{V\theta}{D_c} \quad (2)$$

where τ_f is the shear stress, $\bar{\sigma}$ is the effective normal stress, V is the slip rate, V_* is the reference slip rate, μ_* is the coefficient of friction for $V = V_*$, θ is the state variable, D_c is the critical slip for state variable evolution, and $a > 0$ and $b > 0$ are the constitutive parameters. At constant velocity, θ evolves to its steady state value L/V and the shear stress τ_f depends purely on the slip rate:

$$\tau_f = \bar{\sigma} \left[\mu_* + (a - b) \ln \left(\frac{V}{V_*} \right) \right] \quad (3)$$

Hence, the value of the parameter combination $a - b$ defines the fault behavior at steady state: $a - b > 0$ corresponds to velocity-strengthening friction properties, which lead to stable slip with the imposed loading rate, while $a - b < 0$ defines potentially seismic, rate-weakening fault behaviour.

Using a spring-slider model with a spring stiffness k , *Perfettini and Avouac* (2004) derived an analytical solution to estimate postseismic slip $U(t)$ using the rate-and-state framework:

$$U(t) = \alpha \ln \left[1 + \frac{\beta V_i}{V_*} \left(\exp \left(\frac{t}{t_r} \right) - 1 \right) \right] \quad (4)$$

where

$$\alpha = \frac{(a - b)\bar{\sigma}}{k} \quad (5)$$

$$t_r = \frac{\alpha}{V_*} \quad (6)$$

$$\beta = \exp \left(\frac{\Delta\tau}{(a - b)\bar{\sigma}} \right) \quad (7)$$

The parameter α corresponds to the length scale over which the elastic stress will reach the order of the frictional stress. k is the equivalent stiffness (*Marone et al.*, 1991, *Perfettini and Avouac*, 2004). t_r is the characteristic relaxation time. $\Delta\tau$ is the static Coulomb stress change induced by the coseismic event at time $t = 0$, $\Delta CFF = \Delta\tau - \mu_*\Delta\sigma$ (with the geomechanical convention that normal stress is positive in compression and friction coefficient μ approximated as μ_*). If positive, it will induces an increase in the sliding rate V_i by a factor β in a rate-strengthening area. Therefore βV_i is the initial rate at the beginning of the postseismic period.

Taking the derivative of $U(t)$ with respect to time t , we obtain the rate of afterslip:

$$V(t) = \beta V_i \frac{\exp\left(\frac{t}{t_r}\right)}{1 + \frac{\beta V_i}{V_*} \left(\exp\left(\frac{t}{t_r}\right) - 1\right)} \quad (8)$$

Assuming $\frac{\beta V_i t}{\alpha} \gg 1$ and $t < t_r$, and neglecting the higher-order terms, we can simplify Equations (4) and (8):

$$U(t) \approx \alpha \ln \left[1 + \frac{\beta V_i t}{\alpha} \right] \quad (9)$$

$$V(t) \approx \frac{\alpha}{t} \quad (10)$$

5.2. Afterslip rate and coseismic slip

As described above, the rate of afterslip should follow Equation (10). The measured rates are plotted in Figure 7. Using a weighted least-squares adjustment, we solved Equation (10) and obtained a best-fitting parameter of $\alpha = 90.6$ mm. The InSAR measurements show good agreement with the model $V(t) = \frac{90.6}{t}$ mm/yr. Since we also have an estimate of the slip that occurred in 1978-1991 (6.5 m in 13 years) and 1978-2013 (7.0 m in 35 years) from *Zhou et al.* (2016), we can use these values to calculate the other parameters in Equation (9). Note that the slip estimates are a combination of coseismic (U_{CS}) and postseismic ($U_{POST}(t)$) motion. For simplicity we also made the assumption that no significant preseismic slip has happened between the image acquisitions. Therefore:

$$U_{CS} + U_{POST}(1978 - 1991) = 6.5 \pm 0.5 \text{ m} \quad (11)$$

$$U_{CS} + U_{POST}(1978 - 2013) = 7.0 \pm 0.5 \text{ m} \quad (12)$$

Based on (9), (11) and (12), we can estimate parameters U_{CS} and βV_i . The best-fit solutions gave $U_{CS} = 5.8$ m and $\beta V_i = 3.5 \times 10^{-6}$ m/s. Because the optical image matching results contain relatively larger errors compared to InSAR, we also tried different combinations of U_{CS} and βV_i . Note that $\alpha = 90.6$ mm depends purely on the decay of the afterslip rate, i.e. the InSAR measurements, which we are confident of. To do so, we varied U_{CS} from 0-7 m and computed βV_i . Figure 8 shows that a series of U_{CS} and βV_i can fit the observations reasonably well, and βV_i

is inversely proportional to U_{CS} (Figure 8b). Because βV_i corresponds to the initial rate at the beginning of the postseismic period, it cannot be too large, i.e., it cannot exceed the coseismic slip rate, which is on the order of 1 m/s. Therefore, in order to get a physically meaningful value of βV_i , the coseismic slip has to be at least ~ 4.7 m. This modelled coseismic slip (~ 5 m) is consistent with the geological observations by *Berberian* (1979) (see Appendix A).

It is interesting to note how fast afterslip could occur (Figure 8). Using the lower bound of the coseismic slip (~ 4.7 m), rate-and-state friction laws predict ~ 1.1 m afterslip in 0.5 days immediately after the main shock, which is $\sim 60\%$ of the total afterslip (1.8 m). This raises some concerns about using InSAR and optical images to measure coseismic deformation. Since the first post-earthquake image is usually acquired a few days after an event, “coseismic displacements” derived from InSAR and optical images may contain a significant amount of afterslip in some cases.

To calculate the characteristic relaxation time t_r , we use a reference slip rate of $V_* = 1.0$ mm/yr. We do not have any direct measurements of V_* , and the value was estimated based on the assumption that the long-term slip rate agrees with the geological slip rate of ~ 1.5 mm/yr on the right-lateral strike-slip Nayband fault to its south (*Foroutan et al.*, 2014). The relaxation time $t_r = 91$ years, though much larger than the typical value of 7-11 years (*Perfettini and Avouac*, 2004), is expected given the surprisingly long timescale of the observed afterslip.

5.3. Frictional properties of the Tabas ramp

We used our InSAR measurements of afterslip on the Tabas ramp to estimate the frictional properties of the fault by applying two slightly different methods.

For the first method we made some assumption about the elastic properties of the medium. Based on Equation (5), we estimated the constitutive parameter $(a - b)\bar{\sigma} = 0.18$ MPa for an equivalent stiffness of $k = G/h$, where $G = 10$ GPa is the shear modulus of the medium near the surface and $h = 5$ km corresponds to the thickness of the rate-strengthening region (i.e. the bottom depth of the Tabas ramp). The friction parameter $(a - b)$ can then be computed if the effective normal stress is known. The distribution of effective normal stress $\bar{\sigma}$ with depth is usually calculated assuming hydrostatic pore pressure and for a rock density of 2700 kg/m³, typical of the

continental crust. This leads to $\bar{\sigma} \simeq 50$ MPa for a representative depth value of the ramp ($z = 3$ km) and consequently to $(a - b) = 0.0036$.

For the second method we had to make some inference about the static Coulomb stress change induced by the coseismic event. Using a simple fault geometry that follows the Tabas fault trace in the north, we estimate how much coseismic slip must have occurred on the detachment to get a $M_w 7.3$ earthquake if the ramp has slipped 4.7 m. For this basic model we assume that coseismic slip only happened in the northern segment of the fault, where most the deformation has been observed (Berberian, 1979, Zhou *et al.*, 2016), and that homogeneous slip occurred on the ramp and the detachment. This leads to a coseismic slip of 7.1 m on the detachment. We then computed the Coulomb stress change induced by the detachment on the ramp, which is ~ 3.5 MPa. Based on Equation (7), we estimated the constitutive parameter $(a - b)\bar{\sigma}$. To calculate $\beta = V_i/V_{pre}$, we used the geological slip rate of $V_{pre} \simeq 1.5$ mm/yr (Foroutan *et al.*, 2014), which gives $\beta = 2.1 \times 10^{10}$ if the coseismic slip on the ramp is 4.7 m (Figure 8a). As a consequence we obtain a value of 0.15 MPa for the constitutive parameter $(a - b)\bar{\sigma}$. Using the same estimate of effective normal stress as described above, we inferred that $(a - b) = 0.0029$ on the ramp, which leads to an equivalent stiffness of $k = 1.6 \times 10^6$ of the same order a magnitude than the value we inferred for the first method ($k = 2 \times 10^6$).

5.4. Possible mechanisms for seismic/postseismic slip overlap

The geodetic studies of the 1978 Tabas-e-Golshan earthquake and the postseismic slip that followed, combined with the analysis of the frictional properties using the rate-and-state friction framework, suggest that both coseismic slip and afterslip occurred at the same location. Although this phenomenon appears controversial, it has been inferred for other earthquakes, such as the 2003 Chengkung (Thomas *et al.*, 2017b) and 2010 Maule earthquakes (Bedford *et al.*, 2013). Here, we discuss the possible mechanisms that can explain such behaviour.

The first scenario is that the frictional properties of the Tabas ramp remain rate-strengthening through the entire seismic cycle, and the dynamic rupture during the 1978 earthquake propagates inside a rate-strengthening region. The first hypothesis is to assume that the frictional properties we infer for the Tabas ramp are stable through the entire seismic cycle and therefore, that the dy-

332 namic rupture during the 1978 Tabas-e-Golshan earthquake propagates inside a rate-strengthening
 333 region. Numerical models of earthquakes sequences based on rate-and-state friction laws have sug-
 334 gested that, for high pre-level of stress and/or for small positive values of the frictional parameter
 335 $a - b$, dynamic rupture may propagate inside a rate-strengthening region (*Kaneko et al.*, 2010,
 336 *Thomas et al.*, 2014b). Our inferred values of $(a - b) = 0.0036$ or $(a - b) = 0.0029$ are consis-
 337 tent with these models. Indeed, based on the study of *Thomas et al.* (2014b), for a 5 km large
 338 segment with $(a - b) = 0.003$ there is a $\sim 80\%$ chance that an earthquake propagates through
 339 the rake-strengthening region. Moreover, at seismic slip rate (~ 1 m/s), laboratory experiments
 340 have shown that the dynamic coefficient of friction may drop drastically (e.g., *Di Toro et al.*, 2011,
 341 *Wibberley et al.*, 2008), which favours seismic propagation in rate-strengthening areas. Additional
 342 friction weakening mechanisms advocated to explain such behaviour are related to shear heating
 343 that unavoidably occurs during fast sliding that accumulates significant slip. The first mechanism,
 344 described as “flash heating”, has both theoretical and experimental support (e.g., *Goldsby and*
 345 *Tullis*, 2011, *Rice*, 2006, *Tsutsumi and Shimamoto*, 1997). During seismic faulting, fault gouge
 346 grains heat up at asperity contacts and substantially weaken. Pore fluid pressurisation is another
 347 shear-heating-related weakening mechanism that might take place during seismic slip (e.g., *Rice*,
 348 2006, *Sibson*, 1973). In that case, pore fluid expands faster in the shearing layer than the sur-
 349 rounding porous space, which increases the pore fluid pressure and hence decreases the effective
 350 normal stress. This mechanism has been advocated for the 2011 Tohoku-Oki earthquake to ex-
 351 plain the dynamic propagation of the seismic rupture up to the fault trench (*Noda and Lapusta*,
 352 2013). Other weakening processes, including frictional melting (e.g., *Hirose and Shimamoto*, 2005,
 353 *Tsutsumi and Shimamoto*, 1997) and gel formation (e.g., *Goldsby and Tullis*, 2002) have also been
 354 suggested.

355 Another hypothesis is to assume that the frictional properties of the fault interface vary with
 356 time. This idea has been explored theoretically by *Thomas et al.* (2017b), who proposed that
 357 the coseismic/postseismic slip overlap may occur by shear-heating-induced transient change of
 358 rate-and-state properties. The dependence of quasi-static (slow-rate) friction properties on bulk
 359 temperature has been well-established (e.g., *Blanpied et al.*, 1995, *den Hartog et al.*, 2012, *Verberne*

360 *et al.*, 2014). Typically, at high enough temperatures, the rate-and-state properties can transit from
 361 rate-weakening to rate-strengthening. This feature is often encapsulated in the depth dependence
 362 of friction properties in the numerical models of fault slip. *Thomas et al.* (2017b) hypothesise
 363 that such transition from rate-weakening to rate-strengthening can also transiently occur within a
 364 rate-weakening region due to coseismic shear heating. For this particular scenario, it would mean
 365 that, before the 1978 Tabas-e-Golshan earthquake, the properties of the ramp were rate-weakening,
 366 which explains the large coseismic slip inferred. The (rapid) coseismic shear heating would have
 367 then triggered a change in the properties of the fault interface leading to a rate-strengthening
 368 behaviour. If the coulomb stress changes due to the coseismic rupture are positive, this would
 369 trigger afterslip on the ramp. Over time, temperature evolves through off-fault diffusion of heat
 370 and the rate-and-state parameter $a - b$ slowly changes accordingly towards its preseismic rate-
 371 weakening value. This scenario is consistent with the small $a - b$ values we infer from InSAR
 372 measurements, more than 19 years after the Tabas-e-Golshan earthquake.

373 Both scenarios are plausible, but since we are missing preseismic measurements and data right
 374 after the earthquake, it is difficult to favour one model over the other. If the second hypothesis
 375 is valid, it questions the common assumption that rate-and-state frictional properties of faults
 376 do not change on the scale of available geodetic observations. This emphasises the importance of
 377 measuring deformation accurately, with a closed time-stepping throughout the entire seismic cycle,
 378 to better understand fault behaviour.

379 6. Conclusions

380 We have investigated the postseismic deformation along the Tabas fold segment after the 1978
 381 M_w 7.3 Tabas earthquake using over 20 years of ESA's InSAR data. The deformation time series
 382 shows a $\sim 1/t$ decay in postseismic velocities from $\sim 4.2\text{--}5.0$ mm/yr in 1996-1999 to 3.9 ± 0.6
 383 mm/yr in 2003-2005, to 3.0 ± 0.4 mm/yr in 2006-2010, and to a present rate of $2.3\text{--}2.7$ mm/yr in
 384 2014-2017. Combining the InSAR rates and *Zhou et al.* (2016)'s estimates of the total earthquake
 385 slip, we analysed the frictional properties of the Tabas fault using the rate-and-state friction
 386 framework. The rate-and-state parameter $a - b = 0.005\text{--}0.01$ indicates rate-strengthening frictional
 387 behaviour. Meanwhile, we inferred ≥ 4.7 m coseismic slip (rate-weakening) at the same location.

388 This phenomenon, where coseismic slip overlaps with postseismic slip, appears controversial, but
 389 it has been observed in other places, e.g. the 2003 Chengkung (*Thomas et al.*, 2017b) and 2010
 390 Maule earthquakes (*Bedford et al.*, 2013). One possible mechanism is that the frictional parameter
 391 $a - b$ is small enough (close to velocity neutral) to allow dynamic ruptures to propagate in rate-
 392 strengthening regions (*Kaneko et al.*, 2010, *Thomas et al.*, 2014b). The other possibility is that
 393 preseismic rate-weakening behaviour may transit to postseismic rate-strengthening due to coseismic
 394 shear heating (*Thomas et al.*, 2017b), which questions the common assumption that rate-and-state
 395 frictional properties of faults do not change on the scale of available geodetic observations.

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Figure 1: Tectonic background of the 1978 Tabas-e-Golshan earthquake. (a) Iran is a seismically active country due to the convergence between the Arabian and Eurasian plates. (b) Active faults (black lines) in Iran, from *Walker et al.* (2003). Black dots are $M_w \geq 7.0$ earthquakes recorded in eastern Iran since 1976, from the GCMT catalogue (*Ekström et al.*, 2012). (c) Topography of the area inside the box on (b). Surface ruptures (red lines) observed in the field by *Berberian* (1979). Yellow star indicates the epicentre of the 1978 main shock. Red filled triangles denote thrust faulting and open triangles are high-angle reverse faulting. Black box marks the Tabas fold (Figures 3–5).

Figure 2: Baseline plots of selected small baseline interferograms from two ERS tracks (ERS 435 and 206), one Envisat track (Envisat 435) and two Sentinel-1 tracks (S1 D166 and S1 A159). Circles are selected SAR acquisitions: 3 from ERS track 435, 4 from ERS track 206, 22 from Envisat track 435, 20 from Sentinel-1A track 166, and 21 from Sentinel-1A track 159. Lines are the interferograms used in our time-series analysis.

Figure 3: Average LOS velocity between 1996 and 1999 from stacks of ERS interferograms. The satellite LOS direction is shown by the black arrow marked “LOS”. See Figure 2c for location. (a) Rate map derived from ERS track 435. (b) Rate map derived from ERS track 206. (c) and (d) Swath profiles (black box, 2 km wide) of motions (blue points) and topography (black lines) along A-A'. The average LOS rates are estimated to be 2.7 mm/yr and 2.5 mm/yr for tracks 435 and 206 respectively. Positive rates indicate motion towards the satellite.

Figure 4: Average LOS velocity between 2003 and 2010 from stacks of Envisat interferograms. The satellite LOS direction is shown by the black arrow marked “LOS”. See Figure 2c for location. (a) Average LOS velocity between 2003 and 2005. (b) Average LOS velocity between 2006 and 2010. (c) and (d) Swath profiles (black box, 2 km wide) of motions (blue points) and topography (black lines) along A-A'. The average LOS rates are estimated to be 2.2 mm/yr and 1.6 mm/yr for 2003–2005 and 2006–2010. Positive rates indicate motion towards the satellite.

Figure 5: Average LOS velocity between 2014 and 2017 from stacks of Sentinel-1A interferograms. The satellite LOS direction is shown by the black arrow marked “LOS”. See Figure 2c for location. (a) Rate map derived from Sentinel-1A track 166. (b) Rate map derived from Sentinel-1A track 159. (c) and (d) Swath profiles (black box, 2 km wide) of motions (blue points) and topography (black lines) along A-A'. The average LOS rates are estimated to be 1.0 mm/yr and 1.7 mm/yr for descending track 166 and ascending track 159 respectively. Positive rates indicate motion towards the satellite.

Figure 6: Models of fault slip that best fit the ground motion (red) measured using SAR data (blue points) from (a) ERS track 435 (1996–1999), (b) ERS track 206 (1996–1999), (c) Envisat track 435 (2003–2005), (d) Envisat track 435 (2006–2010), (e) Sentinel-1A track 166 (2014–2017), and (f) Sentinel-1A track 159 (2014–2017).

Figure 7: Rates of afterslip decay as $\sim 1/t$. Coloured dots are InSAR-derived rates, with short lines indicating the uncertainties in both rate and time. Black line is the best-fitting model to the measurements. The gradient, i.e. α , is 90.6 mm. See Section 5 for details.

Figure 8: (a) Estimates of afterslip based on different combinations of assumed coseismic slip and βV_i . Coloured dots are the optical measurements, with coloured lines showing the best-fitting models. (b) The best-fitting parameter βV_i is inversely proportional to the assumed coseismic slip. See Section 5.3 for details.

Table 1: Details of the grid search parameters that best fit the InSAR data. See text for descriptions.

Parameters	ERS D435 (1996-1999)	ERS D206 (1996-1999)	Envisat D435 (2003-2005) (2006-2010)		S1 D166 (2014-2017)	S1 A159 (2014-2017)	Optical Matching ⁵ (1974-1991) (1974-2013)		DEM ⁶ (1956-2013)
incidence (°)	28.7	21.0	25.7		39.0	36.0	/	/	/
heading (°)	-166	-166	-168		-167	-13	/	/	/
slip rate (mm/yr)	5.0	4.2	3.9	3.0	2.7	2.3	6.5 m	7.0 m	7.0 m
RMSE (mm/yr)	0.8	0.5	0.6	0.4	0.5	0.6	0.5 m	0.5 m	0.5 m
depth to top (km)	0.4	0.4	0.2	0.2	0.6	0.4	0.1	0.1	0.1
depth to bottom (km)	5	6	4	4	4	5	6	6	6
location (km) ¹	3.3	3.5	3.4	3.4	3.6	3.5	/	/	/
dip (°) ²	50	50	50	50	50	50	50	50	50
gradient (mm/yr/km) ³	0.193	0.126	-0.0258	0.0548	-0.0489	0.0928	/	/	/
offset (mm/yr) ⁴	-2.88	-1.57	-2.87	-0.705	-0.923	-0.56	/	/	/

¹ The along-profile position of the projection of the fault plane onto the surface.

² The dip of the fault was fixed to 50° (Copley, 2014, Zhou et al., 2016). See Section 4 for details.

³ Velocity gradient along the profile.

⁴ Offset of the measurements relative to zero.

⁵ Earthquake slip derived from optical image matching (Zhou et al., 2016).

⁶ Earthquake slip derived from DEM differencing (Zhou et al., 2016).

524 Appendix A. Reconciling modelled coseismic slip and geological observations

525 We use a simple model to explain how the modelled coseismic slip fits the geological observations
526 by *Berberian* (1979). As shown in Figure A.9, we consider the near-surface structure of the Tabas
527 ramp as a “domino” shear zone. The depth and width of the shear zone are $D = 400$ m and
528 $W = \frac{D}{\tan 50^\circ} = 335$ m. For simplicity, the rotating blocks within the shear zone are assumed to
529 be perpendicular to the fault plane, and regularly spaced with an average spacing of 20 m (based
530 on our visual analysis of the bedding planes seen in the SPOT-6 imagery). The length of the
531 blocks $L = W \sin 50^\circ = 257$ m. The rotation of each block driven by a coseismic slip of 5.8 m
532 on the fault at depth $\theta = \frac{5.8}{L} = 1.3^\circ$, and the resulting “displacement” (i.e. bedding plane slip
533 on the surface) $d = 5.8 \sin(\theta) = 0.13$ m (roughly the vertical displacement). The total vertical
534 “displacement” within the shear zone is $\frac{335}{20} \times d = 2.2$ m. As some bedding planes would be buried,
535 field measurements of the earthquake motion is only part of the total displacement. We simply
536 use the ratio of the width of the bedding planes seen on the surface (220 m) to the width of
537 the shear zone (W) to calculate the vertical throw that can be potentially measured in the field:
538 $v = \frac{220}{335} \times 2.2 \text{ m} = 1.44$ m. The prediction agrees well with the geological observations (at least 1.5
539 m), suggesting that the modelled coseismic slip on the Tabas ramp (5.8 m) is reasonable.

Figure A.9: Cartoon showing the sub-surface structure of the Tabas ramp. The tip of the discrete fault plane acts as a shear zone near the surface (at a depth of $0-0.4 \pm 0.2$ km). Bedding plane slips measured in the field are the total deformation within the shear zone (short green lines, d). θ is the rotation due to shearing driven by coseismic slip on the fault at depth.