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Interseismic Deformation in the Gulf of Aqaba from GPS Measurements

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Summary

Although the Dead Sea Transform fault system has been extensively studied in the past, little has been known about the present-day kinematics of its southernmost portion that is offshore in the Gulf of Aqaba. Here we present a new GPS velocity field based on three surveys conducted between 2015 and 2019 at 30 campaign sites, complemented by 11 permanent stations operating near the gulf coast. Interseismic models of strain accumulation indicate a slip rate of $4.9^{+0.9}_{-0.6}$ mm/yr and a locking depth of $6.8^{+3.5}_{-3.1}$ km in the gulf's northern region. Our results further indicate an apparent reduction of the locking depth from the inland portion of the Dead Sea Transform towards its southern junction with the Red Sea rift. Our modelling results reveal a small systematic left-lateral residual motion that we postulate is caused by, at least in part, late postseismic transient motion from the 1995 M_W 7.2 Nuweiba earthquake. Estimates of the moment accumulation rate on the main faults in the gulf, other than the one

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that ruptured in 1995, suggest that they might be near the end of their current interseismic period, implying elevated seismic hazard in the gulf area.

Keywords: Continental tectonics: strike-slip and transform, Space geodetic surveys, Neotectonics, Transient deformation, Plate motions

1 1. Introduction

2 The highest seismic hazard in Saudi Arabia and Egypt belongs to the areas bor-
3 dering the Gulf of Aqaba (Fig. 1). This hazard is due to earthquakes on the off-
4 shore extension of the Dead Sea Transform (DST), a major fault system that ac-
5 commodates left-lateral motion between the Arabian plate and the Sinai-Levant
6 subplate (Freund et al., 1970; McKenzie, 1972). Several major earthquakes are
7 known to have occurred along this southernmost portion of the DST during
8 the last millennium, with the 1995 M_W 7.2 Nuweiba earthquake in the Gulf of
9 Aqaba being the most recent one (Klinger et al., 1999; Hofstetter, 2003). The
10 decision by the Kingdom of Saudi Arabia to build a megacity in the NEOM area
11 near the gulf has put this region under the spotlight, requiring further research
12 efforts to improve the seismic hazard knowledge of the gulf area.

13 The DST formed as a result of the mid-Cenozoic breakup of the Arabo-
14 African plate and today it extends roughly 1000 km from the Red Sea rift to
15 the East Anatolian fault (Garfunkel, 1981). The total estimated left-lateral offset
16 of the DST south of the Dead Sea is about 105 km (e.g., Quennell 1958, 1959;
17 Freund et al. 1970; McKenzie et al. 1970; Garfunkel 1981; Joffe and Garfunkel
18 1987) and the onset of its left-lateral strike-slip motion postdates 19-22 Ma (e.g.,
19 Bar et al. 1974; Steinitz et al. 1978; Bartov et al. 1980; Eyal et al. 1981; Joffe and
20 Garfunkel 1987; Ben-Avraham et al. 2012; Nuriel et al. 2017). These estimates
21 translate to an average slip rate between 4.8 mm/yr and 5.5 mm/yr since the
22 initiation of fault motion along the southern part of the DST comprising the
23 Jordan Valley and Wadi Arabah faults.

24 The geomorphology along the strike of the DST is marked by a series of pro-
25 nounced depressions, 5-20 km wide, corresponding to pull-apart basins. These

26 morphological depressions form at jogs between discontinuous segments of the
27 DST and are usually delimited by both normal and strike-slip faults (e.g., [Gar-](#)
28 [funkel 1981](#); [Ben-Avraham and Zoback 1992](#); [Klinger et al. 1999](#); [Ben-Avraham](#)
29 [et al. 2012](#)). From north to south, these depressions comprise the Ghab and
30 Hula basins, the Sea of Galilee, the Dead Sea and the Gulf of Aqaba (see inset
31 in Fig. 1). These series of pull-apart basins formed by left-lateral shear along
32 left-stepping en échelon fault segments, which results in rhomb-shaped grabens
33 subjected to left-lateral shear and extension ([Quennell, 1959](#); [Garfunkel, 1981](#)).
34 Despite being primarily a transtensional fault system ([Garfunkel, 1981](#)), a ma-
35 jor ~ 200 km long restraining bend characterises the central part of the DST
36 in Lebanon and Syria (e.g., [Heimann and Ron 1987](#); [Gomez et al. 2007](#)). This
37 segment deviates $20^\circ - 30^\circ$ from the main trend of the DST and accounts for
38 ~ 1.6 mm/yr of convergent motion between the Sinai and Arabian plates at that
39 latitude (e.g., [Elias et al. 2007](#); [Gomez et al. 2007](#)).

40 Historical records, paleoseismic trenching and geomorphological studies in-
41 dicate that many destructive earthquakes have occurred along the DST over
42 the past two millennia (e.g., [Ambraseys 2009](#); [Agnon 2014](#); [Meghraoui 2015](#);
43 [Klinger et al. 2015](#); [Lefevre et al. 2018](#)). Two successive earthquakes occurred
44 in 363 AD, affecting more than 20 towns in Palestine and Syria ([Ambraseys,](#)
45 [2009](#)). According to [Thomas et al. \(2007\)](#), these earthquakes caused extensive
46 destruction at several localities in southern Wadi Arabah. This observation is in
47 agreement with [Klinger et al. \(2015\)](#), who reported an event horizon in a trench
48 located 30 km north of the city of Aqaba (Jordan) that could be related to these
49 earthquakes.

50 A major historical earthquake occurred in 1068 AD in the gulf area, de-
51 stroying the town of Eilat ([Ambraseys, 2009](#)). [Zilberman et al. \(2005\)](#) reported
52 that the earthquake was strong enough to rupture the surface along the Eilat
53 fault, affecting an irrigation system. Historical records document another violent
54 earthquake in 1212 AD. This earthquake was strongly felt in Eilat, and caused
55 damages to the monastery of St Catherine in the Sinai Peninsula ([Ambraseys,](#)
56 [2009](#)). This event was identified in a trench close to the village of Qatar in

57 southern Jordan (Klinger et al., 2015), and a brecciated layer associated with
58 this earthquake was also reported by Kagan et al. (2011) in the Dead Sea basin.
59 The last major historical earthquake reported in the gulf area occurred in 1588.
60 Despite its location and magnitude are a subject of debate, intense shaking
61 was documented as far as Cairo and Madinah, suggesting an epicentre in the
62 northern Red Sea region (Ambraseys, 2009). Finally, the largest instrumental
63 earthquake along the entire DST was the M_W 7.2 Nuweiba Earthquake, which
64 struck the Gulf of Aqaba area on November 22, 1995, causing significant dam-
65 age in nearby coastal communities in Egypt, Saudi Arabia, Jordan and Israel
66 (Klinger et al., 1999; Hofstetter, 2003).

67 The gulf itself is about 180 km long and up to 25 km wide (Ben-Avraham
68 et al., 1979; Garfunkel, 1981), and its structure is dominated by three left-
69 stepping en échelon faults bounding a series of pull-apart basins (Ben-Avraham
70 et al. 1979; Ben-Avraham 1985; Ribot et al. 2021, Fig. 1). The northernmost
71 fault segment within the gulf is the Eilat fault. This fault is approximately 59 km
72 long and constitutes the western boundary of the Eilat deep. The central fault
73 segment is the Aragonese fault, which extends approximately 53 km, bounding
74 the Eilat deep to the east and the Aragonese deep to the west. The Arnona fault
75 constitutes the southernmost segment, extending roughly 83 km and forming the
76 western boundary of the Dakar and Tiran deeps. Compelling evidence indicates
77 that the Aragonese fault segment ruptured during the Nuweiba earthquake,
78 while the other segments have not released significant seismic moment in the
79 past several centuries (Shamir et al., 2003; Hofstetter, 2003; Baer et al., 2008;
80 Ribot et al., 2021).

81 The seismic activity in the gulf is characterised by the occurrence of swarms,
82 which have been observed since the deployment of the first regional seismic
83 networks in the area (El-Isa et al. 1984; Alamri et al. 1991 as cited in Klinger
84 et al. 1999). Significant swarm activity was reported in 1983, 1990 and 1993,
85 reaching M_W 6.1 (El-Isa et al. 1984; Pinar and Türkelli 1997; Klinger et al.
86 1999, Fig. 2). Most earthquakes in the gulf exhibit predominantly left-lateral
87 motion. However, some of them show significant normal motion, as evidenced by

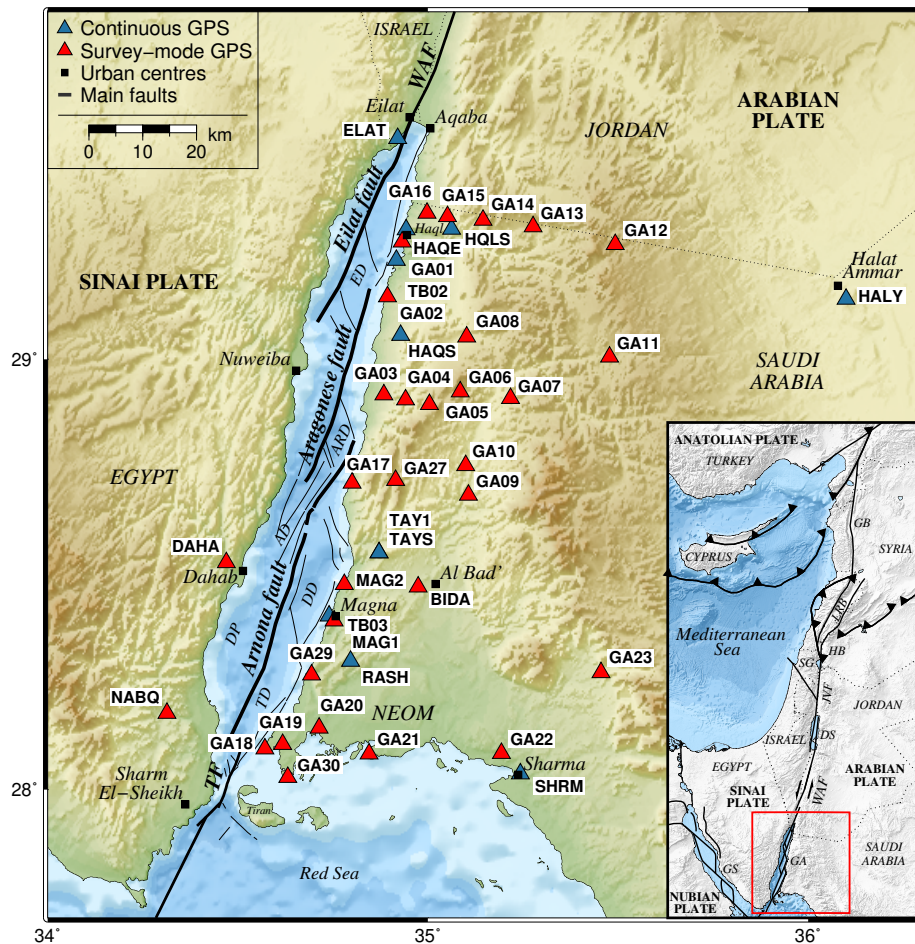


Figure 1: Mapped faults in the Gulf of Aqaba along with the GPS stations used in this study. Permanent and campaign GPS sites are shown as blue and red triangles, respectively. Fault traces and pull-apart basins are based on Ribot et al. (2021). The inset shows the area of interaction of the Arabian, Nubian, Sinai and Anatolian plates. Shaded topography is based on the SRTM30 PLUS digital elevation model (Becker et al., 2009) and regional-scale fault traces in the inset are compiled from Le Béon et al. (2008). TD, Tiran Deep; TF, Tiran Fault; DP, Dahab Plateau; DD, Dakar Deep; AD, Arnona Deep; ARD, Aragonese Deep; ED, Eilat Deep; GS, Gulf of Suez; GA, Gulf of Aqaba; WAF, Wadi Arabah Fault; DS, Dead Sea; JVF, Jordan Valley Fault; SG, Sea of Galilee; HB, Hula Basin; LRB, Lebanese Restraining Bend; GB, Ghab Basin.

88 the pure normal slip reported for the largest event during the 1993 earthquake
 89 swarm (Klinger et al., 1999; Hofstetter, 2003). More recently, in June 2015,
 90 a M_W 5.2 strike-slip earthquake struck the gulf area, causing strong shaking
 91 in nearby communities and an aftershock sequence that lasted for more than

92 two weeks (El-Aal and Badreldin, 2016). Recent seismic data collected by the
 93 Saudi Geological Survey (SGS) show that earthquakes concentrate along the
 94 gulf faults, (see as shown in Fig. 2).

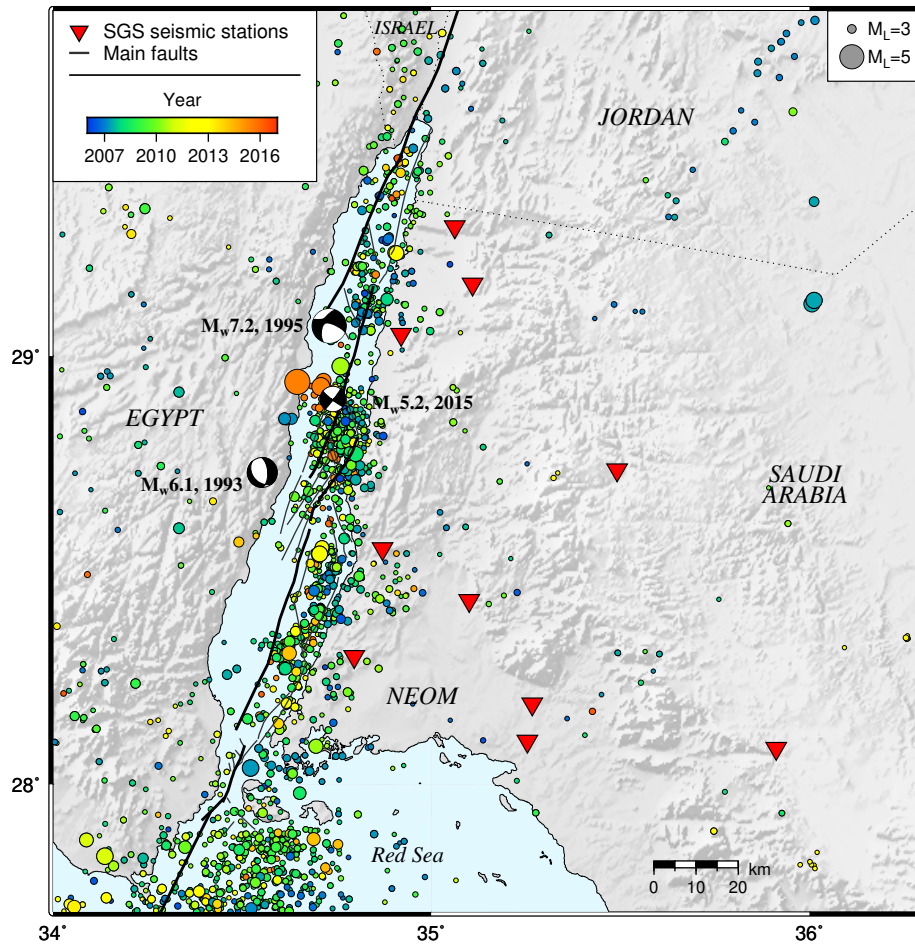


Figure 2: Earthquake locations ($M_L \geq 1.5$) in the Gulf of Aqaba region from January 2006 to March 2016 with circle sizes scaled according to earthquake magnitude and colours according to date. Focal mechanisms of the 1993 M_W 6.1, 1995 M_W 7.2 and 2015 M_W 5.2 earthquakes are from the National Earthquake Information Center (NEIC, <https://earthquake.usgs.gov>), the Global Centroid-Moment-Tensor (GCMT) catalog (Dziewonski et al., 1981; Ekström et al., 2012) and El-Aal and Badreldin (2016), respectively. Seismic stations operated by the Saudi Geological Survey (SGS) are shown as red inverted triangles. Note that the one-sided geometry of the SGS seismic network might bias epicentral locations.

95 The present-day crustal deformation along the DST north of the Gulf of
 96 Aqaba has been extensively analysed in terms of interseismic strain accumu-

107 lation (e.g., [McClusky et al. 2003](#); [Wdowinski et al. 2004](#); [Gomez et al. 2007](#);
108 [Le Béon et al. 2008, 2010, 2012](#); [Al Tarazi et al. 2011](#); [Sadeh et al. 2012](#); [Mah-](#)
109 [moud et al. 2013](#); [Masson et al. 2015](#); [Hamiel et al. 2016, 2018a](#)). However, only
110 four **GPS** studies have provided slip rate estimates across the fault system in the
111 gulf area. By implementing a back-slip model, [Mahmoud et al. \(2005\)](#) derived a
112 slip rate of 4.4 ± 0.3 mm/yr, assuming a fixed locking depth of 13 km. A similar
113 approach was followed by [Reilinger et al. \(2006\)](#), who reported a slip rate of 4.5
114 ± 0.3 mm/yr, holding the locking depth fixed to 12 km. [ArRajehi et al. \(2010\)](#)
115 estimated slip rates of 4.8 ± 0.2 mm/yr in the gulf, and, more recently, [Gomez](#)
116 [et al. \(2020\)](#) implemented a back-slip model and reported a slip rate of $4.9 \pm$
117 0.1 mm/yr. Unfortunately, these studies were unable to provide locking depth
118 estimates due to the limited geodetic coverage near the gulf.

119 In this study, we use previously unpublished geodetic data to derive an up-
120 dated crustal motion velocity field of the gulf. We further implement standard
121 models of interseismic deformation to investigate the fault kinematics along
122 the offshore segments of the DST fault system. Finally, we consider four pos-
123 sible model scenarios of time-dependent viscoelastic deformation produced by
124 the Nuweiba earthquake and discuss whether postseismic transient motions are
125 evidenced in the GPS data acquired near the gulf.

116 2. GPS Data and Analysis

117 A new survey-mode GPS network was installed in 2014 and surveyed three times
118 in February 2015, March 2017 and May 2019 (see Tab. S1 and Fig. S1). This
119 network consists of 27 GPS markers distributed over an area of 150 km by 70
120 km on the eastern gulf coast, extending from the Strait of Tiran in the south to
121 the Jordan border in the north (Fig. 1). During the surveys, we also occupied
122 three older markers installed by MIT near the municipalities of Al Bad' and
123 Magna (sites BIDA, MAG1 and MAG2). These stations were measured annually
124 between 2010 and 2014 ([Gomez et al., 2020](#)), and their data were reprocessed
125 in our study. We also analysed the GPS data collected between 2000 and 2005

126 at the campaign sites DAHA and NABQ, which are located near the gulf coast
127 on the Sinai Peninsula (McClusky et al., 2003; Mahmoud et al., 2005; Gomez
128 et al., 2020).

129 To better understand the relative motion between the Sinai-Levant subplate
130 and the Arabian plate, we processed data from 20 stations of the GIL network
131 of continuous GPS monitoring in Israel (Wdowinski et al., 2001), covering a
132 19-year analysis period from 2000 to 2019 (see Fig. S2). Considering the rapid
133 expansion of continuous GNSS GPS networks in Saudi Arabia, we incorporated
134 data from additional 11 permanent stations near the gulf (HAQS, HQLS, TAY1,
135 HAQE, SHRM, BEJD, TB02, TB03, HALY, RASH and TAYS). Our research
136 significantly improves the GPS station coverage near the Gulf of Aqaba com-
137 pared to previous studies, which relied on sparse GPS data in this area (e.g.,
138 McClusky et al. 2003; Mahmoud et al. 2005; Reilinger et al. 2006; ArRajehi
139 et al. 2010; Gomez et al. 2020). Nevertheless, near-fault GPS observations are
140 limited to distances greater than 8 km due to the presence of the gulf itself and
141 the lack of offshore geodetic measurements.

142 We analysed the GPS data using the GAMIT/GLOBK software package
143 version 10.70 (Herring et al., 2018) following a three-step approach described
144 in detail by Dong et al. (1998). Daily carrier phase data from campaign and
145 permanent stations were analysed separately to improve the processing efficiency
146 (e.g., Bock et al. 1997). Following this strategy, we obtained two sets of loosely
147 constrained solution vectors and covariance matrices. Both solutions share a set
148 of 16 reference sites with well-determined coordinates and velocities, which we
149 used to tie our final combined solution to the International Terrestrial Reference
150 Frame (ITRF) 2014 (Altamimi et al., 2017). Campaign solutions were then
151 merged with the continuous ones to get a final solution with ambiguities resolved
152 and loose constraints on estimated parameters.

153 Combined solutions were used to estimate position time series and site ve-
154 locities in GLOBK. The reference frame realisation was achieved by imposing
155 tight constraints to the coordinates and velocities of 16 sites that comprise the
156 reference network (ZIMM, GRAZ, ONSA, JOZE, WSRT, KIT3, IISC, POLV,

157 ARTU, POL2, MAS1, RABT, YEBE, DGAR, MBAR, NKLG), minimising in
158 an iterative scheme their position adjustments with respect to their ITRF2014
159 a priori coordinates by estimating a six-parameter Helmert transformation. Fi-
160 nally, we accounted for time-correlated errors in the position time series by
161 characterising their noise spectrum using the Hector software package version
162 1.7.2 (Bos et al., 2013). Since campaign stations do not have enough obser-
163 vations to estimate the time-correlated noise in their position time series, we
164 used the mean random walk values obtained for the permanent stations in the
165 study area ($4.5 \cdot 10^{-7} \text{ m}^2/\text{yr}$ and $1.5 \cdot 10^{-6} \text{ m}^2/\text{yr}$ for the horizontal and vertical
166 velocity components, respectively). Following this approach, the resulting hori-
167 zontal velocity uncertainties for stations within the new campaign GPS network
168 deployed near the gulf vary between 0.5 and 1.0 mm/yr (see Tab.S2).

169 3. GPS Velocity Field

170 Time series and site velocities realised in ITRF2014 were rotated into stable
171 Sinai reference frame. To perform this rotation, we estimated the Euler pole of
172 the Sinai subplate with respect to ITRF2014 using five reliable stations located
173 on the Sinai subplate at distances greater than 40 km from the DST (CSAR,
174 ALON, YRCM, BSHM and RAMO), as shown in Fig. S2. The SINAI-ITRF2014
175 Euler pole derived here is $54.7 \pm 0.7^\circ \text{N}$, $347.8 \pm 4.0^\circ \text{E}$, $\omega = 0.417 \pm 0.021^\circ \text{Ma}^{-1}$,
176 which is consistent with Euler vectors estimated in earlier studies (e.g., Wdowin-
177 ski et al. 2004; Le Béon et al. 2008; Sadeh et al. 2012; Hamiel et al. 2016, 2018b,c;
178 Hamiel and Piatibratova 2019). Observed GPS horizontal velocities for the sites
179 on the Arabian plate show NNE motion relative to Sinai (Fig. 3), as reported in
180 previous geodetic investigations (e.g., ArRajehi et al. 2010; Sadeh et al. 2012;
181 Gomez et al. 2020). The velocity field thus reflects the left-lateral motion of the
182 DST fault system. The increasing DST-parallel velocities towards the interior of
183 the Arabian plate in this Sinai-fixed reference frame (Figs. 3-5a), suggest that
184 the coastal areas bordering the gulf faults are currently undergoing interseismic
185 strain accumulation.

186 Fig. 4 shows profiles of fault-parallel and fault-normal components of GPS-
 187 derived velocities relative to the Sinai-Levant subplate. These projections were
 188 performed assuming a single vertical fault with a strike of N16°E extending from
 189 the Strait of Tiran in the south to the northernmost tip of the gulf (dashed yel-
 190 low line in Fig. 3). Increasing DST-parallel velocities towards the Arabian plate
 191 support active interseismic strain accumulation along the coastal areas adjacent
 192 to the gulf (Fig. 4a). We further note that the mean fault-normal velocity con-
 193 sidering all the stations on the Arabian side of the gulf is $\sim 0.8 \text{ mm/yr}$ (Fig. 4b).
 194 This observation is consistent with a partly divergent motion between the Sinai
 195 and Arabian plates at this latitude, as suggested in previous investigations (e.g.,
 196 Mahmoud et al. 2005; Reilinger et al. 2006; Gomez et al. 2020). However, the
 197 scatter observed in both velocity profiles suggests that the assumption of a sin-
 198 gle fault along the full extension of the gulf may be inadequate. This observation
 199 led us to consider different fault traces within the gulf based on the structural
 200 maps by Ben-Avraham (1985) and Ribot et al. (2021), see below.

201 4. Elastic Dislocation Model

202 We conducted an initial kinematic assessment of the GPS velocity field across
 203 the gulf by implementing a screw dislocation model (e.g., Weertman and Weert-
 204 man 1964; Savage and Burford 1973). Following this approach, theoretical fault-
 205 parallel velocities v can be expressed as follows:

$$v = a + \frac{b}{\pi} \tan^{-1} \left(\frac{x}{D_L} \right) \quad (1)$$

206 where a defines the vertical origin of the arctangent curve, and x is the per-
 207 pendicular distance measured from the fault trace. The fault is assumed to
 208 be locked during the interseismic period at depths extending from the surface
 209 down to a depth D_L (often referred to as locking depth), below which free slip
 210 occurs at a rate equivalent to the secular relative plate motion b . GPS measure-
 211 ments at distances within one-half the locking depth are required to accurately
 212 constrain fault locking depths, whereas far-fault observations are important to

213 constrain secular slip rates (Smith-Konter et al., 2011). This simple interseismic
 214 model has been extensively used to investigate kinematic parameters of many
 215 strike-slip and transform faults globally (e.g., Lisowski et al. 1991; Wright et al.
 216 2001; Reilinger et al. 2006; Sadeh et al. 2012).

217 Here we analyse two velocity profiles across the Eilat fault in the north and
 218 the Arnona and Tiran faults in the south, which include the largest number of

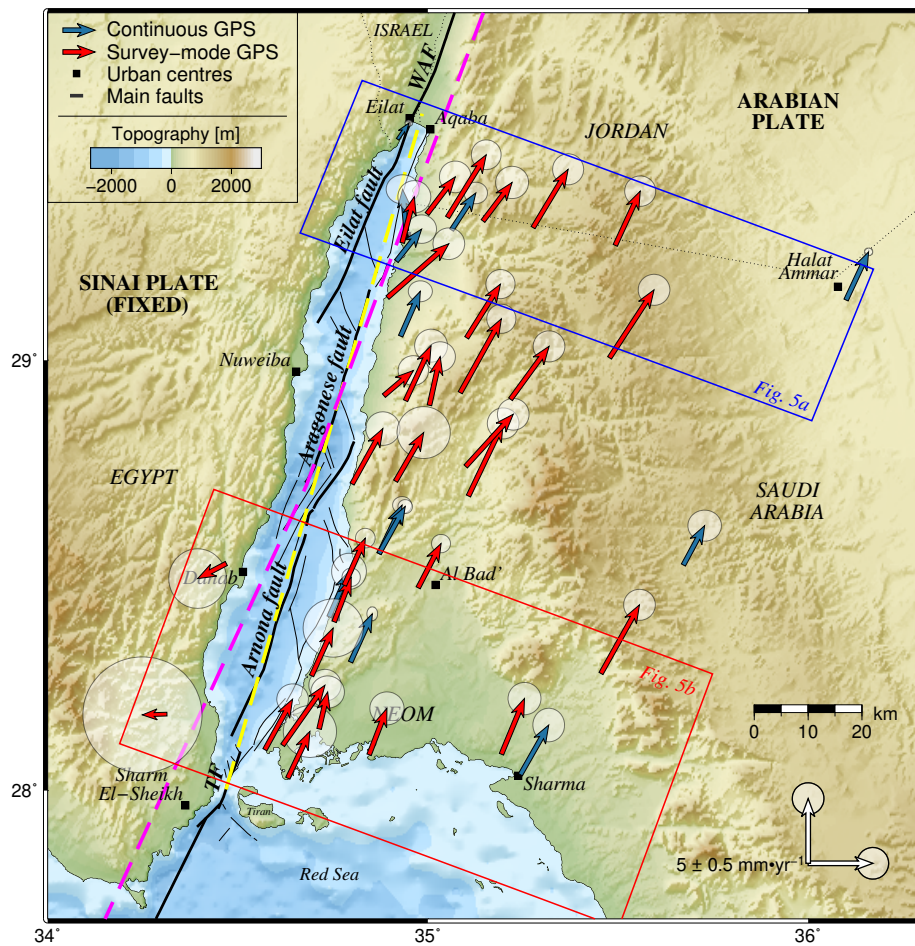


Figure 3: Horizontal GPS velocities relative to Sinai with 95% confidence ellipses. Permanent and campaign site velocity vectors are coloured in blue and red, respectively. The blue and red boxes indicate the location of the velocity profiles shown in Fig. 5a-b. The dashed yellow line corresponds to the assumed fault location used in Fig. 4a-b. The dashed magenta arc delineates the small circle around the Arabia-Sinai Euler pole derived in this study.

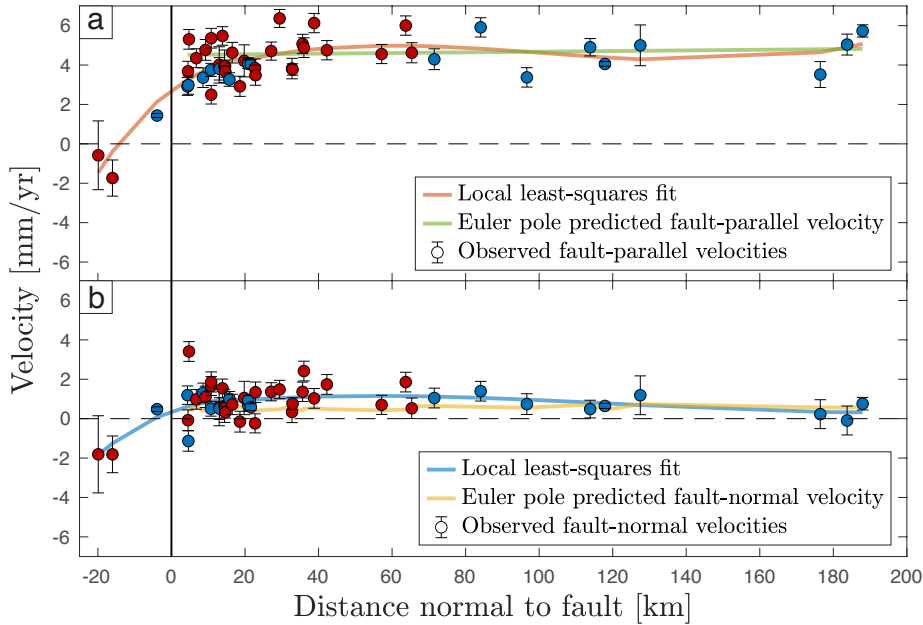


Figure 4: Profiles of a) fault-parallel and b) fault-normal velocities across Gulf of Aqaba assuming a single vertical fault striking N16°E along the gulf (dashed yellow line in Fig. 3). Site velocities are expressed in a Sinai-fixed reference frame, i.e., positive velocity values on the Arabian plate ($x > 0$) represent left-lateral motion in a) and opening in b) across the gulf. The solid orange and blue lines correspond to the best local least-squares fit to fault-parallel and fault-normal velocities, respectively. The least-squares fit was performed using a third-order Savitzky-Golay filter considering 39 data points per window. The pale green and yellow lines represent the fault-parallel and fault-normal velocities predicted by the Arabia-Sinai Euler Pole. Permanent and campaign site velocities are coloured in blue and red, respectively.

219 near-fault GPS stations. The northern profile extends from the city of Eilat to
 220 the northern tip of the gulf to the municipality of Halat Ammar located 120
 221 km inside the Arabian plate (Fig. 5a), assuming a fault strike of N22°E. The
 222 southern profile extends from Sharm El-Sheikh in the Sinai Peninsula to the city
 223 of Sharma in Saudi Arabia (Fig. 5b), and assumes a fault strike of N20°E. The
 224 sampling width of these profiles is based on the extension of the fault segments
 225 within the gulf, which overlap in certain areas forming pull-apart basins, as
 226 shown in Fig. 1. Following this criterion, the southern Eilat and the Aragonese
 227 pull-apart basins correspond to the limits of the northern and southern profiles,
 228 respectively. Unlike the northern and southern profiles, the central area across
 229 the Aragonese fault does not include enough near- and far-field GPS stations to

230 reliably constrain slip rates and locking depths. Moreover, there are no stations
231 constraining GPS rates on the Sinai Peninsula west of the Aragonese fault. For
232 these reasons, the central segment was not considered for slip-rate and locking
233 depth inversions.

234 We evaluated if observed fault-parallel velocities resemble the theoretical
235 pattern of strain accumulation given by Eq. 1, and excluded from the analy-
236 sis those sites showing anomalous rates with no clear tectonic explanation (grey
237 dots in Fig. 5a-b). We then constrained model parameters (i.e., a , b and D_L) and
238 their uncertainties using a grid-search algorithm coupled with a Monte Carlo
239 procedure that accounts for uncertainties in the fault position and stations'
240 velocities. The best fit corresponds to the set of model parameters that results
241 in the lowest root mean squared (RMS) error relative to the actual GPS obser-
242 vations (Fig. 5a-b). To estimate uncertainties in slip rates and locking depths,
243 we conducted 10^4 Monte Carlo simulations, perturbing the location of the fault
244 randomly within a range of ± 0.5 km and adding Gaussian noise to the GPS
245 velocities within their 1-sigma uncertainties. In each Monte Carlo simulation,
246 the grid-search explores the misfit (RMS) of the model for a range of values in
247 the parameter domain, and selects the combination of model parameters that
248 results in the lowest misfit. The strike (azimuth) of the fault segments was held
249 fixed during the slip rate and locking depth inversions. However, we explored the
250 effect of small perturbations (± 5 degrees) in fault strike on the final inversion
251 results. We found that small variations in fault strike do not have a statistically
252 significant impact on the slip rate and locking depth estimates, as shown in
253 Fig. S3.

254 After performing the Monte Carlo inversion, we derived a 2×2 covariance
255 matrix from the resulting distribution of slip rates and locking depths. From this
256 covariance matrix, we estimated the width (ω) and height (h) of the confidence
257 level ellipse, which are given by $\omega = 2 \times nstd \times \sqrt{\lambda_1}$ and $h = 2 \times nstd \times \sqrt{\lambda_2}$,
258 where $nstd$ represents the desired confidence level (in standard deviations) and
259 $\lambda_{1,2}$ are the eigenvalues of the covariance matrix. Finally We followed this ap-
260 proach to derive , we estimated the eigenvalues and eigenvectors for the 1-sigma

261 (68%) confidence level ellipses, as shown in Fig. 5c-d.

262 Our results yielded a slip rate of $4.9^{+0.9}_{-0.6}$ mm/yr and a locking depth of
 263 $6.8^{+3.5}_{-3.1}$ km for the northern profile across the Eilat fault, we derived (red curve
 264 Fig. 5a). For the southern profile Further south, across the Arnona and Tiran
 265 faults, we derived a slip rate of $5.5^{+1.3}_{-0.9}$ mm/yr and a locking depth of $0.8^{+3.4}_{-0.8}$ km
 266 (red curve in Fig. 5b). Fault parallel rates at Stations GA12, GA13 and GA15 in
 267 the northern profile and GA19, GA20 and GA23 in the southern profile exhibit
 268 anomalous rates relative to nearby stations with no obvious explanation. These
 269 outliers could be related to local effects that may include unmodelled tropo-
 270 spheric signals, local deformation and seasonal loading, which are not modelled
 271 for the campaign stations. The inclusion of sites with anomalous behaviour in
 272 the slip rate and locking depth inversion is undesirable because they adversely
 273 affect fit to the data (e.g., Sadeh et al. 2012). For this reason, these stations
 274 were dismissed from the analysis, yielding the results given by the red curves
 275 in Fig. 5a-b. disagreed with the expected 1D screw elastic dislocation model
 276 within their 1-sigma uncertainties and were dismissed from the analysis.

277 Fault parallel rates at stations GA12, GA13 and GA15 disagreed with the
 278 expected 1D screw dislocation model within their 1-sigma uncertainties and were
 279 dismissed from the analysis. We further implemented an alternative approach,
 280 solving for the best fitting model for a fixed slip rate of 4.5 mm/yr, which cor-
 281 responds to the fault parallel rate predicted by the Arabia-Sinai Euler pole at
 282 the fault trace (see Fig. 4a). Following this approach, for the northern profile,
 283 we obtained a best fitting locking depth of 6.2 km (yellow curve in Fig. 5a).

284 We also obtained a very shallow locking depth estimate (< 2 km) for a fixed
 285 slip rate of 4.5 mm/yr (yellow curve in Fig. 5b). To further analyse the effect
 286 of near-fault station rates on the final locking depth estimates, we considered
 287 an additional scenarios assuming a fixed slip rate of 4.5 mm/yr and omitting
 288 stations GA01 and GA16 in the northern profile and GA18 in the southern
 289 profile (blue curves in Fig. 5a-b), which moves at full plate rate despite be-
 290 ing located ~ 8 km from the fault trace. Our results show that the omission
 291 of sites GA01 and GA16 in the northern profile has a minor effect on the slip

292 rate estimate but results in a deeper locking depth of 10.5 km. Similarly, ex-
293 cluding of station GA18 in the southern profile results in a deeper locking depth
294 estimate of 3.2 km compared with the 0.8 km obtained when including GA18
295 in the inversion (green line in Fig. 5b). The 68% confidence level ellipses shown
296 in Fig. 5c-d are elongated, showing a clear trade-off between slip rates and fault
297 locking depths. As demonstrated in several studies, higher slip rates require
298 larger deeper locking depths (e.g., Wright et al. 2001; Meade and Hager 2005;
299 Smith-Konter et al. 2011). This correlation arises from the tight constraint on
300 the near-field velocity gradient imposed by the dislocation model (see Wright
301 et al. 2001 for details).

302 Finally, we compared our results with the slip rates and locking depths de-
303 rived by Li et al. (2021) based on the implementation of Sentinel 1 burst-over-
304 lap interferometry (BOI) in the lands bordering the gulf (light orange curves in
305 Fig. 5a-b). Our Taken together, both results indicate shallower locking depths
306 from north to south, towards the Red Sea rift. Our results further suggest that
307 a very shallow locking depth or creeping is required in the southern part of the
308 gulf to account for the low strain accumulation observed in this area. However,
309 the current near-fault GPS data are insufficient to unambiguously discriminate
310 between partial shallow locking and full creep scenarios along the southern fault
311 segments. These results motivated us to implement a more complex model of
312 strain accumulation, which not only considers fault-parallel motions but also
313 allows for tensile motions and rigid block rotations.

314 5. Back-Slip Model

315 Interseismic deformation studies conducted further north along the DST in Is-
316 rael, Jordan, Syria and Lebanon have shown that GPS-derived velocities can be
317 well explained by rigid block movement bounded by a narrow deformation zone
318 that accommodates the relative motion between them (e.g., Mahmoud et al.
319 2005; Reilinger et al. 2006; Gomez et al. 2007, 2020; Al Tarazi et al. 2011).
320 This view is also supported by the distribution of instrumental and historical

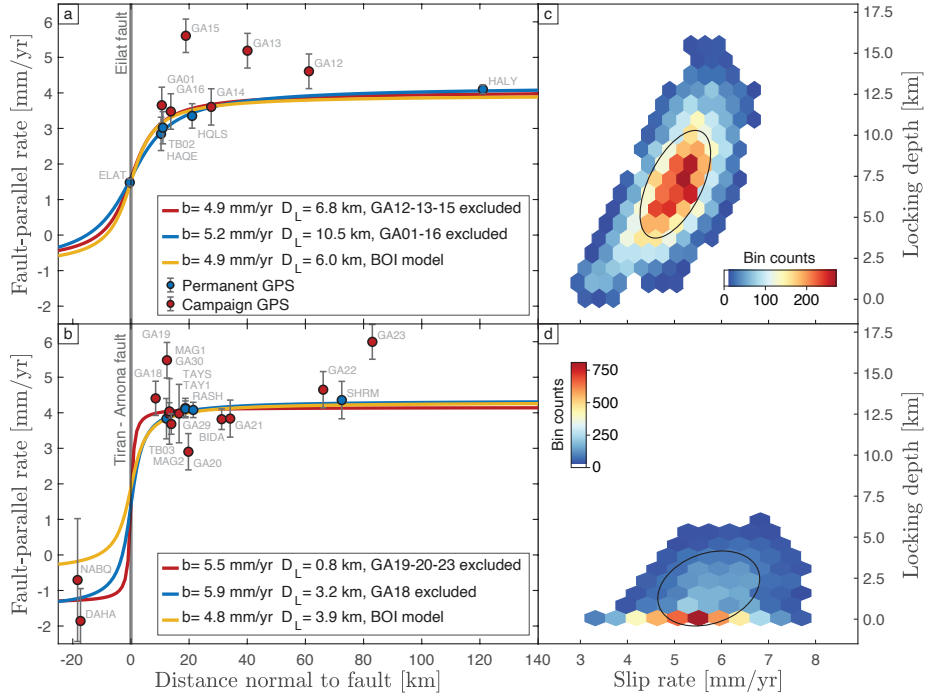


Figure 5: Observed fault-parallel velocities with 1-sigma uncertainties across the Eilat fault (panel a) and Tiran-Arnona faults (panel b) in comparison with predictions from elastic dislocation models in a Sinai-fixed reference frame. Panels c and d depict the results of the slip rate and locking depth inversions and 68% confidence level ellipses. The red curve in both profiles represents the global minimum considering a , b and D_L as free parameters. Results from Li et al. (2021), based on burst-overlap interferometric (BOI) observations, are shown in light orange. Outliers are shown as grey dots and were dismissed from analysis, as discussed in the text.

321 seismicity, which concentrates along the main fault strands of the DST fault
 322 system (Hofstetter et al., 2014). We thus next investigated whether the velocity
 323 field in the Gulf of Aqaba area can also be explained using a similar back-slip
 324 model.

325 We adopted the approach described by McCaffrey (2005, 2009) and defined
 326 two blocks representing the Sinai and Arabian plates. The boundary between
 327 them was modelled as a set of linear fault segments representing the main fault
 328 strands of the gulf mapped by Ben-Avraham (1985) and Ribot et al. (2021),
 329 with short relay faults connecting the segments (Fig. 7a). Model-predicted ve-
 330 locities result from superimposing linear rates due to block rotations and those

331 associated with the back-slip motion applied on block boundaries, which are
332 estimated following [Okada \(1985\)](#).

333 **The number of near-fault GPS observations is insufficient to reliably invert**
334 **for the locking fraction distribution along the fault segments within the gulf,**
335 **even for sparse discretisations of the fault planes. Hence, to further investigate**
336 **the fault kinematics in the gulf's southern region, we considered three forward**
337 **back-slip model scenarios defining the locking fraction of the fault segments in**
338 **this area a priori.** In the first scenario, we imposed full interseismic locking to
339 all the faults in the gulf (Fig. 6a). In the second scenario, we considered creep
340 on the Arnona and Tiran faults (Fig. 6b). Finally, in the third scenario, we
341 imposed creep on the Tiran fault **only** (Fig. 7). Locked faults were modelled as
342 near vertical extending down to 8.7 km , based on our best-fit estimate across the
343 Eilat fault. Creeping faults were modelled as free-slipping boundaries that do
344 not produce any elastic strain. **To evaluate the effect of implementing different**
345 **fault geometries in our back-slip models, we ran an additional scenario assuming**
346 **a dip angle of 65° for the Aragonese fault, as proposed by [Baer et al. \(2008\)](#). By**
347 **comparing the resulting back-slip model residuals with those obtained assuming**
348 **a vertical fault (Fig. 7), we show that the variance difference is not large enough**
349 **to be statistically significant, as depicted in Fig. S5.**

350 Considering that most stations in the gulf area have short time series (< 5
351 years), we did not optimise for rotation parameters during the back-slip mod-
352 elling. Instead, we defined the Arabia-Sinai Euler pole a priori using the Sinai-
353 ITRF2014 Euler vector derived in this study and the Arabia-ITRF2014 rotation
354 pole reported by [Altamimi et al. \(2017\)](#). In this way, the back-slip applied on
355 each locked fault in the gulf was determined by the Arabia-Sinai rotation pole
356 (35.00°N , 8.78°E , $\omega=0.105^\circ\text{Ma}^{-1}$), assuming that no aseismic slip occurs dur-
357 ing the interseismic period on those faults. Finally, we compared observed and
358 back-slip predicted velocities in the gulf area and analysed the residual rates.

359 Our modelling results indicate that the coastal areas bordering the Arnona
360 fault are undergoing strain accumulation. This is supported by the good fit to
361 the data **on the Arabian side of the gulf** obtained by imposing full interseismic

locking to the Arnona fault in the first model scenario (Fig. 6a) and the small but
coherent southward residuals obtained in the second scenario, where creeping
is assumed along the Arnona and Tiran faults (Fig. 6b). We analysed these
residuals quantitatively, showing that when the Arnona fault is assumed to be
fully locked, near-fault velocity residuals are close to zero (panels a and b in
Fig. S6). In contrast, full-creep on the Arnona fault leads to a systematic bias
of ~ 0.4 mm/yr and ~ 0.2 mm/yr in the North and East velocity components,
respectively (panels c and d in Fig. S6). We also noted that the fit to the data
across the Tiran fault degrades if this fault is assumed to be locked during the
interseismic period. This observation suggests that creeping might be required
to explain the low interseismic strain accumulation observed across this fault.
We found that the best fit to the data is obtained by implementing scenario
3, which considers full locking along all the fault segments except Tiran fault,
which is modelled as a creeping fault. Scenario 3 explains the interseismic strain
accumulation observed across the Arnona fault and the full-plate rate exhibited
by near-fault GPS stations across Tiran fault (see Fig. 7a).

Our results further suggest a fault-parallel slip-rate of ~ 4.5 mm/yr and a
tensile component of ~ 0.3 mm/yr along the main fault strands in the gulf, as
determined from the the relative plate motion (see Fig. 7a). The back-slip model
succeeds in fitting 13 out of 17 GPS rates roughly 80% (76%) of observed veloci-
ties within their 95% confidence ellipses in the southern region of the gulf (east of
the Arnona and Tiran fault traces). Nevertheless, our modelling results reveal an
apparent reduction in the quality of the fit for most sites located in northern Gulf
of Aqaba (east of the Aragonese fault trace), where less than 60% of residuals lie
within velocity uncertainties only 12 out of 23 observations (52%) of observed
velocities were fitted within their 95% confidence ellipses. Interestingly, residual
velocities are not randomly distributed but are instead systematically directed
towards the East-Northeast on the Arabian side of the gulf. This observation led
us to investigate in more detail the nature of these residuals and their spatial
distribution. For this purpose, we estimated descriptive statistics of the misfits,
discriminating between stations in the northern and southern regions of the gulf.

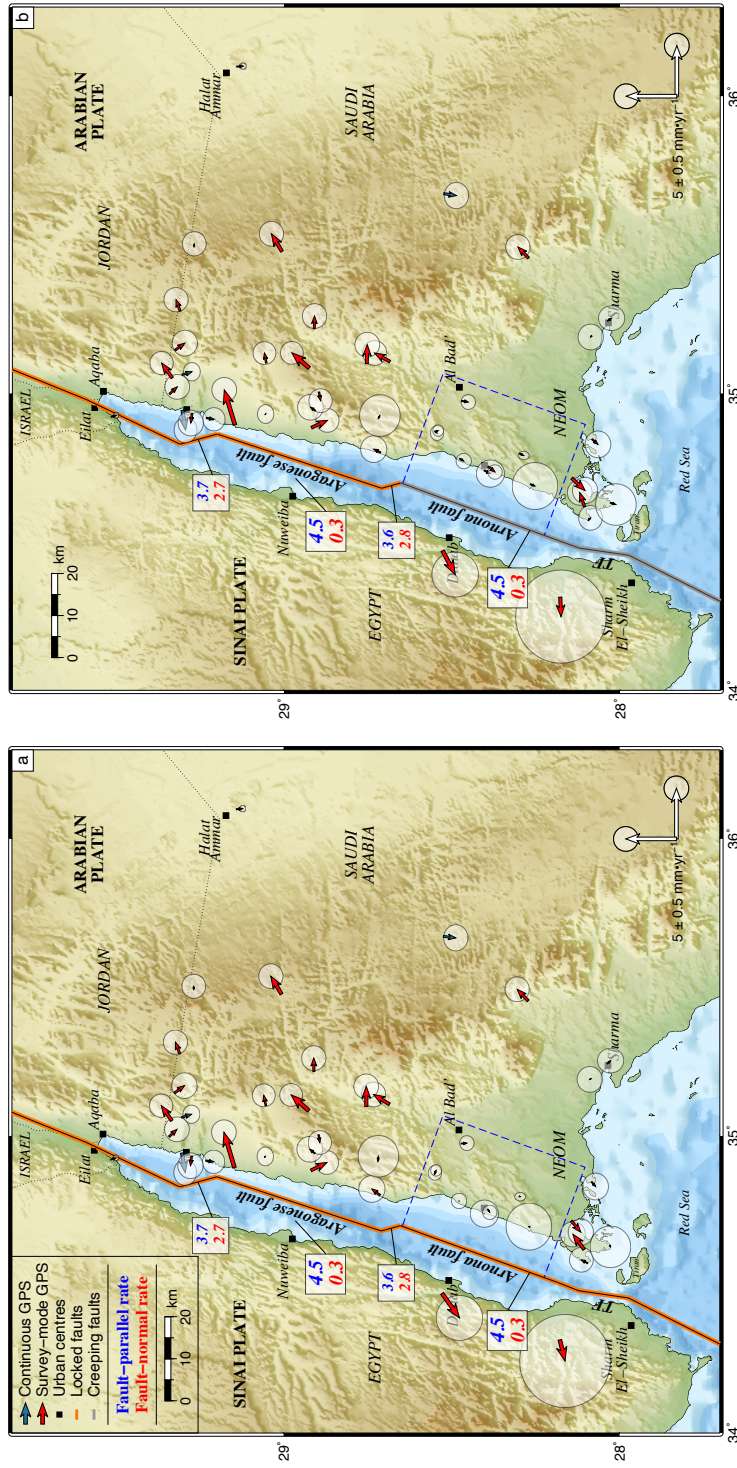


Figure 6: Panel a shows residual velocities after implementing the first back-slip model scenario assuming full interseismic locking on all the fault traces. Panel b shows the residuals obtained after implementing the second scenario assuming full creep along the Arnona fault and the Tiran fault (TF). Orange and grey lines represent locked and creeping fault traces, respectively. Residual rates of seven near-fault stations within the blue dashed boxes are analysed in Fig. S6.

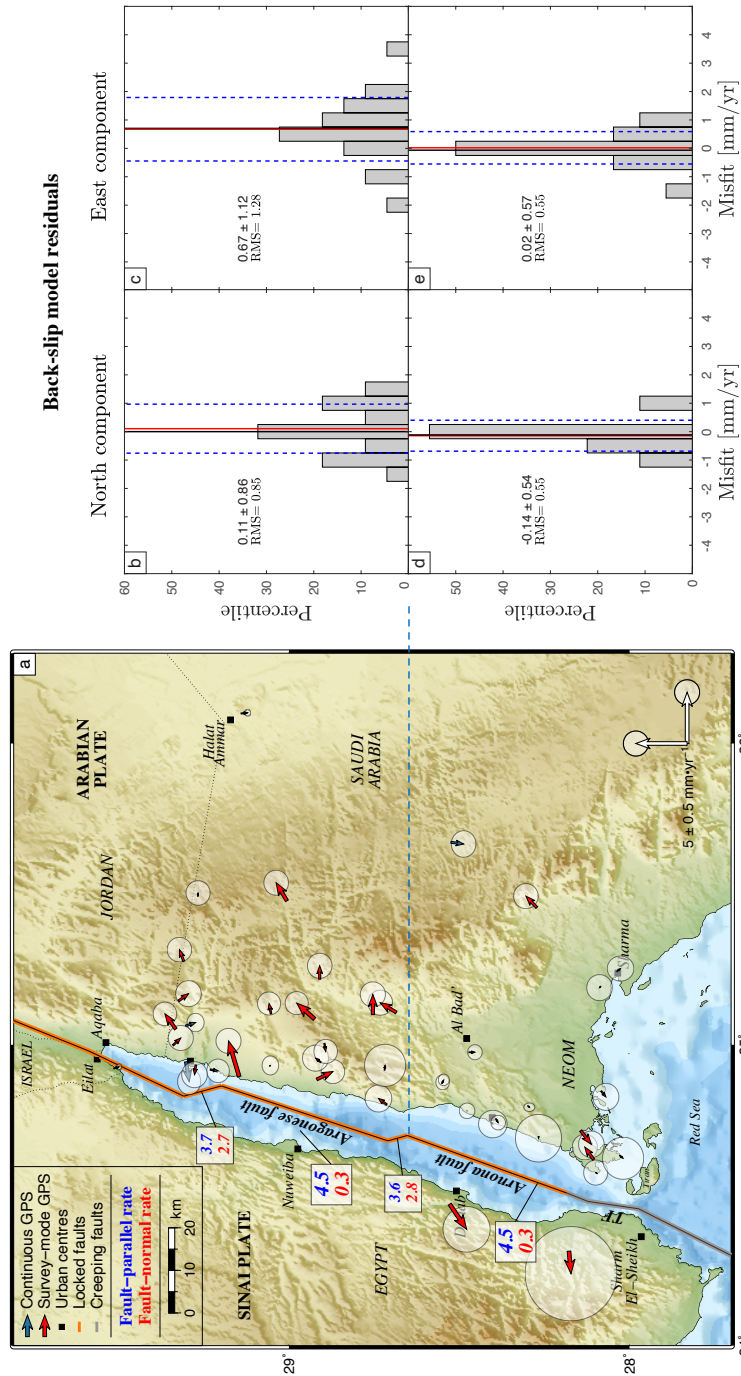


Figure 7: Residual velocity vectors after implementing the third back-slip model scenario (panel a), which assumes full creep on only the Tiran fault (TF). Histograms of fault parallel and perpendicular residuals for the northern part of the gulf (north of the blue dashed line) are shown in panels b-c and for the southern part in panels d-e, with both the bias and RMS values reported.

393 The misfit distributions depicted in Fig. 7b-e quantitatively demonstrate
394 that the magnitude of residual velocities differs between the northern and south-
395 ern regions on the Arabian side of the gulf. Notably, the average magnitude of the
396 east component of residual velocities in the northern area is about 0.7 mm/yr .
397 This value contrasts with that observed east of the Arnona fault in the southern
398 region of the gulf, where the east component of residual velocities is roughly zero.
399 Similarly, but to a lesser extent, we observe a small discrepancy in the north
400 component of residual velocities. The mean magnitude of the north component
401 of residual velocities east of the Aragonese fault is 0.15 mm/yr , while the
402 average value of the same velocity component observed east of the Arnona fault
403 in the southern region of the gulf is -0.11 mm/yr . Our results thus suggest
404 the existence of a left-lateral residual motion across the gulf that cannot be
405 explained by block rotations and interseismic strain accumulation alone. These
406 results drove us to study the potential influence of postseismic transient motions
407 caused by the 1995, $M_w 7.2$ Nuweiba earthquake in the lands bordering the gulf.

408 Fig. 7a also shows consistent misfits for stations DAHA and NABQ in the
409 Sinai Peninsula. Both stations show significant residuals directed towards the
410 west-southwest.

411 6. Postseismic Modelling

412 Transient surface deformation has been systematically observed following large
413 earthquakes spanning periods from months to decades (Scholz, 2002). Three
414 time-dependent mechanisms are most often tested to explain such deformation
415 transients: Afterslip on and/or around the ruptured fault plane (e.g., Savage
416 et al. 1994; Bürgmann et al. 1997), poroelastic rebound (e.g., Peltzer et al.
417 1996; Jónsson et al. 2003) and viscoelastic relaxation of the lower crust and
418 upper mantle (e.g., Pollitz 1997; Freed and Bürgmann 2004). The last one is the
419 only process capable of producing postseismic deformation spanning both large
420 spatial and long temporal scales (Vergnolle et al. 2003 and references therein).
421 Despite considerable efforts aimed at elucidating the underlying nature of the

422 mechanisms controlling postseismic deformation, discriminating among them
423 using geodetic measurements is challenging and is presently an active area of
424 research (e.g., Jónsson 2008; Liu et al. 2020).

425 Considering the limited GPS measurements in the years following the 1995
426 Nuweiba earthquake, the data are insufficient to invert for rheological param-
427 eters. Hence, we set up a series of forward models to analyse the postseismic
428 surface deformation resulting from different lithospheric structures for the Gulf
429 of Aqaba. For this purpose, we studied the viscoelastic relaxation following the
430 methodology proposed by Wang et al. (2006) for a layered viscoelastic Earth
431 model accounting for gravity effects. We used a simplified 1D velocity model of
432 the gulf area from Tang et al. (2016) to define a layered Earth structure consist-
433 ing of a Burgers viscoelastic half-space underlying a 30 km thick elastic layer,
434 representing the Earth’s upper mantle and crust, respectively (see Fig. 8b-d). We
435 also considered a three-layer Earth structure consisting of a 30 km elastic upper
436 crust underlain by a biviscous lower crust and upper mantle (see Fig. 8e). The
437 source of the $M_W 7.2$ Nuweiba earthquake was modelled using the distributed
438 slip model derived by Baer et al. (2008) from ERS-1 and ERS-2 Interferometric
439 Synthetic Aperture Radar (InSAR) data (see Fig. 9).

440 We explored four model scenarios by varying the transient and steady-state
441 viscosities of the Burgers solids. In the first three scenarios, the viscosity of
442 the underlying half-space was set based on the best-fit viscosity reported by
443 Piersanti et al. (2001) for the Sinai-Suez area ($\sim 10^{18} Pa \cdot s$). In the fourth
444 scenario, we considered a thin crème brûlée rheological structure consisting on
445 a 30 km elastic upper crust underlain by a biviscous lower crust and upper
446 mantle. Following the considerations above, in scenario 1, we assigned values of
447 $1 \cdot 10^{18}$ and $1 \cdot 10^{19} Pa \cdot s$ to the transient (η_K) and steady-state (η_M) viscosities,
448 respectively. In scenario 2, we fixed η_K to $2 \cdot 10^{18} Pa \cdot s$ and η_M to $2 \cdot 10^{19} Pa \cdot s$.
449 In scenario 3, we considered a higher transient viscosity of $1 \cdot 10^{19} Pa \cdot s$ and a
450 steady-state viscosity of $1 \cdot 10^{20} Pa \cdot s$. Finally, in scenario 4, we defined steady-
451 state viscosities of $2 \cdot 10^{18} Pa \cdot s$ and $4 \cdot 10^{19} Pa \cdot s$ for the lower crust and upper
452 mantle, respectively (transient/steady-state viscosity = 0.1). We held fixed the

453 ratio between effective and unrelaxed shear modulus (α) to a value of 0.5 in the
 454 first three scenarios, and 0.7 in the last model (see below).

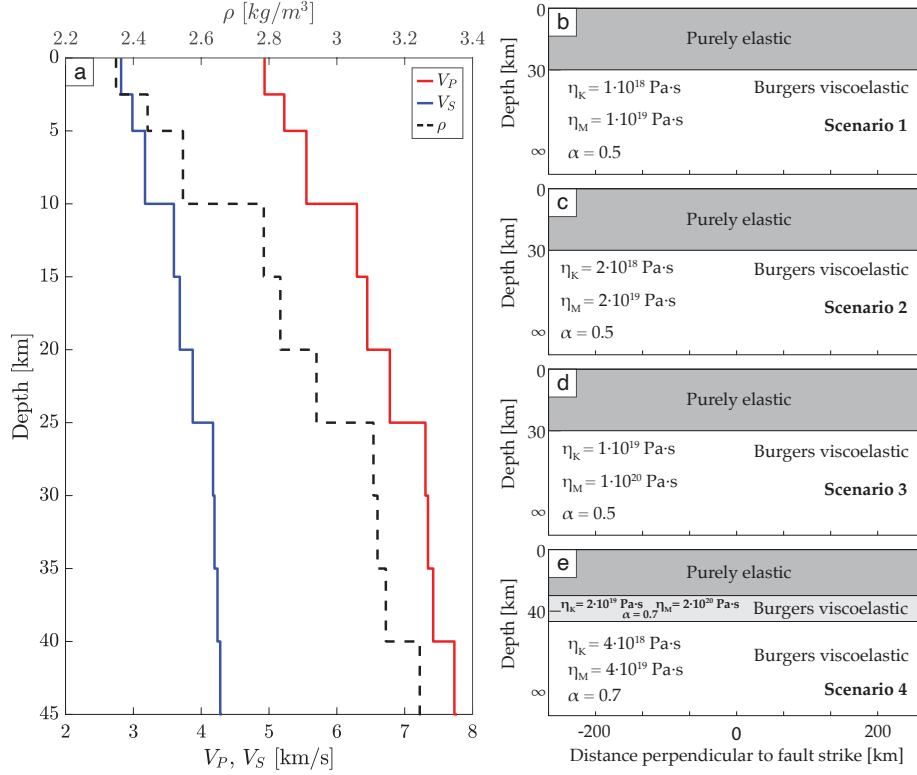


Figure 8: Panel a shows the simplified 1D layered Earth model of the Gulf of Aqaba (adapted from Tang et al. 2016). Panels b, c, d and e are schematic representation of the lithosphere for model scenarios 1, 2, 3 and 4, respectively. Viscoelastic parameters of the Kelvin and Maxwell elements are also given for each scenario: η_K is the transient viscosity (dashpot of the Kelvin solid), η_M represents the steady-state viscosity (dashpot of the Maxwell fluid). α corresponds to the ratio between effective and unrelaxed shear modulus $\frac{\mu_K}{\mu_K + \mu_M}$. The unrelaxed shear modulus (μ_M) can be derived from the values of S wave velocities (V_s) and densities (ρ) provided in panel a.

455 For each model scenario, we computed time series of postseismic displace-
 456 ments for the period spanning from the origin time of the 1995 Nuweiba earth-
 457 quake to the year 2020. Postseismic velocities were then derived from the average
 458 displacement rate over the observation period of each GPS station. As can be
 459 seen in Fig. 9, model scenarios implementing lower transient and steady-state
 460 viscosities result in slightly higher postseismic rates compared with those includ-
 461 ing higher viscosities. The pattern of modelled postseismic velocities in all four

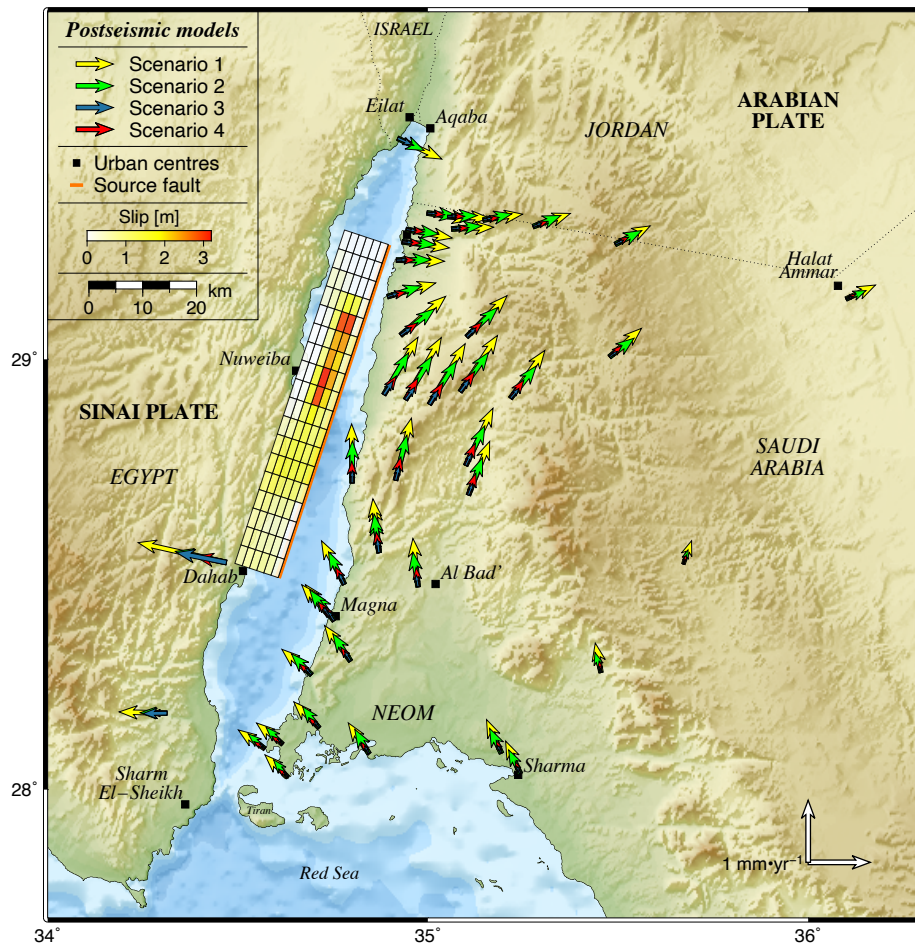


Figure 9: Modelled postseismic velocities computed at the location of GPS stations in the study area. Yellow, green, blue and red vectors represent modelled postseismic velocities for scenarios 1, 2, 3 and 4, respectively. **Postseismic rates were estimated for the observation period at each GPS station (see Fig. S1 for details).** The colour-coded grid represents the finite-source model of the Nuweiba earthquake derived by Baer et al. (2008). Note the change in velocity scale from previous figures.

462 cases is similar and shows left-lateral motion, which agrees with the coseismic
 463 slip prescribed in the finite-fault model. However, unlike the modelled interseismic
 464 mic velocities, the predicted postseismic deformation fields exhibit maximum
 465 fault-parallel velocities about 30 km away from the fault trace, with lower rates
 466 next to the fault and in the far-field.

467 To determine whether the back-slip model residuals shown in Fig. 7 can

468 be explained by postseismic transient motions from the Nuweiba earthquake,
469 we estimated the misfits resulting from subtracting modelled postseismic rates
470 from the block model residuals. Fig. 10a depicts the misfit vectors obtained
471 for the four postseismic model scenarios considered in our study. As done in
472 the back-slip model, we discriminated between GPS stations located in the
473 northern and southern regions adjacent to the gulf on the Arabian plate. The
474 misfit distributions and RMS values for each postseismic model are shown in
475 Fig. 10b-e.

476 Overall, the lowest RMS misfits east of the Aragonese fault were obtained by
477 implementing model scenario 2 (green vectors in Fig. 10a). Modelled postseismic
478 velocities in the north (east of Aragonese fault **trace**) lead to a reduction in
479 the RMS misfit by $\sim 11\%$ and $\sim 4\%$ of the east and north **rate velocity**
480 components, respectively. In contrast, the quality of the fit in the southern region
481 of the gulf (east of the Arnona and Tiran fault **traces**) degraded by 4% and 41%
482 in east and north velocity components, respectively. The reduction in the quality
483 of the fit in this area is due to northwestward modelled postseismic velocities,
484 which are not observed in the block-model residuals. As shown in Fig. 7, GPS-
485 derived velocities are well-fit by the back-slip model in the gulf's southern region
486 located east of the Arnona and Tiran faults. Hence, the consideration of any
487 additional postseismic signal in ~~that area~~ **the southern region of the gulf** results
488 in a degradation of the fit.

489 Our results suggest that modelled postseismic velocities induced by the 1995,
490 $M_W 7.2$ Nuweiba earthquake explain part of the left-lateral residual motion ob-
491 served in the lands bordering the gulf. However, the forward models fail to
492 satisfactorily reproduce the magnitudes of the residual velocity field by under-
493 estimating somewhat the back-slip model residuals in the northeast of the gulf
494 while overshooting observed residuals to the southeast.

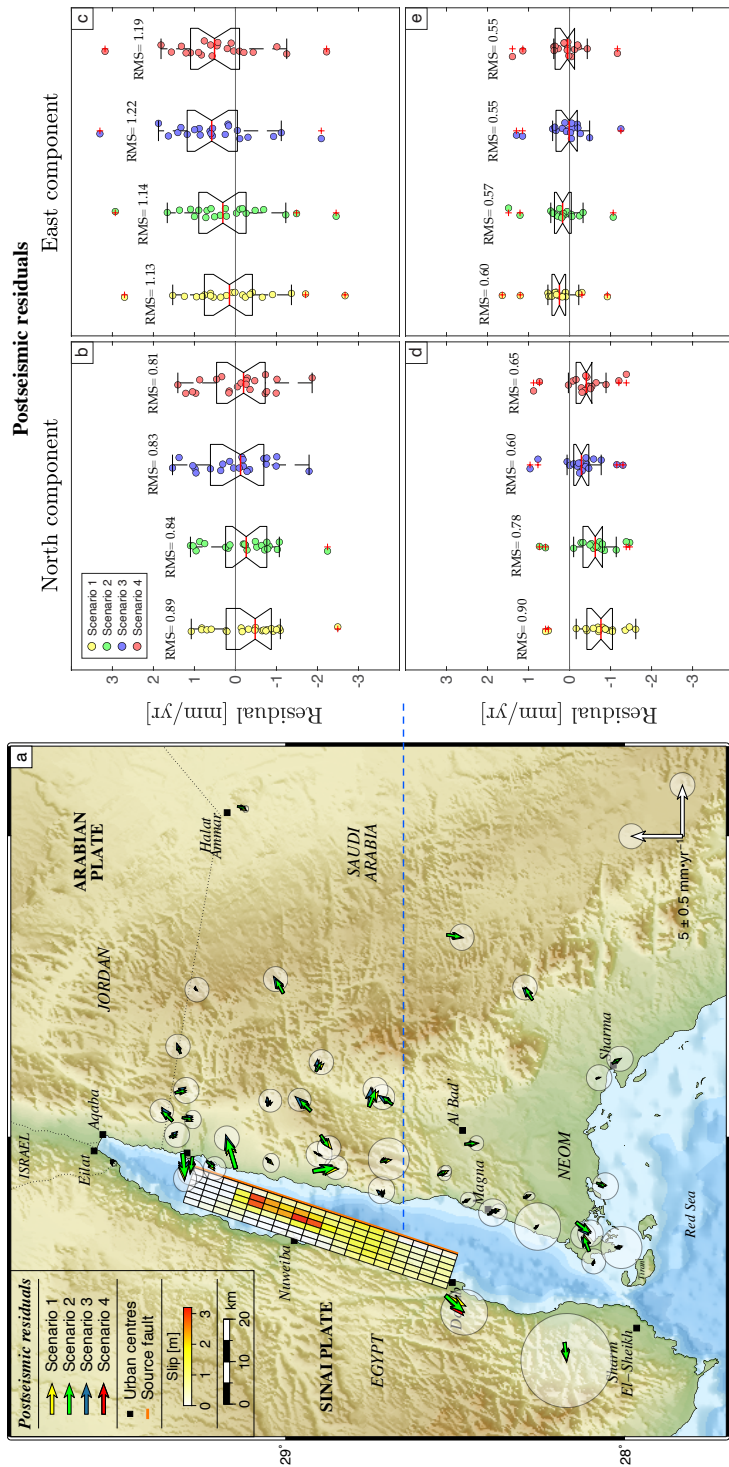


Figure 10: Residual velocities after implementing four postseismic model scenarios discussed in the text, with the orange line delineating the 1995 co-seismic fault rupture of the Aragonese fault (after Baer et al. 2008). For visualisation purpose, 95% confidence ellipses are shown only for the best fitting postseismic model (scenario 2 + 4). The contrast between the residuals of stations in the **North** and **South** (East of the Aragonese fault **trace**) is illustrated in panels b-e, with **notched** boxes enclosing data within 25th and 75th percentiles. For each scenario, the median of the misfit distribution is shown as a solid red line and whiskers extend to the most extreme data points not considered as outliers, which are represented by the red crosses.

495 **7. Discussion**

496 Previous estimates of the relative motion between the Sinai and Arabian plates
497 along the Gulf of Aqaba relied on sparse observations at a limited number of
498 GPS sites located on each side of the gulf (e.g., Mahmoud et al. 2005; Reilinger
499 et al. 2006; ArRajehi et al. 2010; Gomez et al. 2020). However, the establishment
500 of 27 survey-mode sites in 2014 and the rapid expansion of permanent geodetic
501 networks in Saudi Arabia during the last decade have allowed us to derive an
502 updated GPS velocity field of this region.

503 By implementing an elastic half-space model, we derived a slip rate of
504 $4.9_{-0.6}^{+0.9}$ mm/yr and a locking depth of $6.8_{-3.1}^{+3.5}$ km across the Eilat fault. We
505 also estimated a slip rate of $5.5_{-0.9}^{+1.3}$ mm/yr and a locking depth of $0.8_{-0.8}^{+3.4}$ km
506 further south across the Arnona and Tiran faults (see red curves in Fig. 5a-b).
507 While the slip rates derived here agree with previous studies, the locking depth
508 estimates are shallower than those reported along the Wadi Arabah fault
509 north of the gulf, which range between 12 and 20 km (e.g., Le Béon et al. 2008;
510 Sadeh et al. 2012; Masson et al. 2015; Hamiel et al. 2018b; Gomez et al. 2020).
511 Further south, stations' rates show little evidence of strain accumulation
512 despite their proximity to the Tiran fault. This pattern agrees with a very
513 shallow locking depth, which might indicate potential creep along this fault
514 segment (see Fig. 5b).

515 To evaluate the impact of outliers in the slip rate and locking depth inver-
516 sions, we ran additional tests considering all the stations in the profiles including
517 stations with anomalous rates. Inversions including all the GPS stations in the
518 northern profile yielded a slip rate of $6.7_{-0.7}^{+0.7}$ mm/yr and a locking depth of
519 $8.8_{-2.2}^{+2.2}$ km. Further south, across the Arnona and Tiran faults, this test re-
520 sulted in a slip rate of $5.9_{-0.7}^{+1.3}$ mm/yr and a locking depth of $2.5_{-2.2}^{+3.2}$ km. Our
521 results thus suggest that the slip rates seem to be more affected by the outlier
522 rates in the mid- and far-field, than the locking depths (see Fig. S4). Although
523 the slip rates and locking depths for the southern profile are not as well con-
524 strained as for the northern profile, our modelling results show that the buried

525 screw dislocation model provides consistent locking depth estimates even in the
526 presence of the outliers, indicating shallower locking depths in the south.

527 Previous geophysical studies conducted along the on-land portion of the DST
528 and its offshore extension in the Gulf of Aqaba indicate a systematic thinning of
529 the lithosphere towards the Red Sea (e.g., Ginzburg et al. 1979; Ben-Avraham
530 1985; Mohsen et al. 2006; El Khrepy et al. 2016a). Ben-Avraham (1985) con-
531 ducted gravity, magnetic and seismic surveys in the gulf, and proposed a thinner
532 crust in its southern region. This observation agrees with earlier investigations
533 indicating higher heat flow values and gradual crustal thinning towards the
534 south (e.g., Ginzburg et al. 1979; Ben-Avraham and Von Herzen 1987). Simi-
535 larly, tomographic inversions and S-receiver function studies revealed that the
536 lithospheric thickness decreases from 80 *km* north of the Dead Sea to 65 *km*
537 in the Gulf of Aqaba, indicating stronger asthenospheric upwelling in the south
538 (Mohsen et al., 2006; El Khrepy et al., 2016a). More recently, Li et al. (2021) im-
539 plemented along-track Sentinel-1 burst-overlap interferometric time-series anal-
540 ysis and reported decreasing locking depths from 15.9 ± 1.9 km of the Wadi
541 Arabah fault to only 3.9 ± 1.5 km in the southern part of the Gulf of Aqaba.
542 Hence, the locking depth reduction from north to south along the Wadi Arabah
543 and faults within the gulf might be explained by progressive crustal thinning
544 towards the Red Sea.

545 We estimated a slip rate of ~ 4.5 *mm/yr* along the main fault segments in
546 the gulf based on a forward back-slip model, which implemented the Arabia-
547 Sinai Euler pole derived in this study. This estimate is lower than the best-fit
548 inverted in our screw dislocation models, and could be attributed to potential
549 internal deformation or fragmentation of the Sinai plate as has been reported
550 in recent studies (e.g., Gomez et al. 2020). The present-day slip rate derived
551 from geodetic data in the gulf area agrees well with the long-term geological
552 estimates along the southern on-land portion of the DST, which converge to an
553 average value of 5 ± 1 *mm/yr* (e.g., Le Béon et al. 2010, 2012). We can thus
554 assume that the present-day slip rate has remained relatively constant over
555 the last millennium. Based on this assumption, the Eilat and Arnona faults

556 have accumulated significant moment since the last historical earthquakes that
557 struck the gulf area in 1212 AD and 1588 AD (e.g., [Ambraseys 2009](#); [Agnon](#)
558 [2014](#); [Klinger et al. 2015](#); [Lefevre et al. 2018](#)).

559 We estimated the slip deficit and moment accumulation rate on the major
560 strike-slip faults in the gulf, assuming that they are fully locked during the in-
561 terseismic period. We also considered an additional scenario assuming that the
562 Tiran fault in the southern region of the gulf creeps aseismically. Considering
563 that the 1212 AD and 1588 AD earthquakes cannot be unambiguously assigned
564 to specific fault traces within the gulf, we tested both as potential last earth-
565 quakes on the Eilat and Arnona faults. For this purpose, we evaluated lower
566 and upper bound estimates based on the fault kinematic parameters derived
567 in this research and previous investigations along the southern on-land portion
568 of the DST, corresponding to slip rates ranging between 4.4 and 5.2 mm/yr
569 (e.g., [Mahmoud et al. 2005](#); [Reilinger et al. 2006](#); [Le Béon et al. 2008](#); [ArRajehi](#)
570 [et al. 2010](#); [Al Tarazi et al. 2011](#); [Sadeh et al. 2012](#); [Masson et al. 2015](#); [Hamiel](#)
571 [et al. 2016, 2018a,b](#); [Gomez et al. 2020](#)). For the locking depths, we considered
572 the lower and upper bounds from the 1-sigma uncertainty derived in the gulf's
573 northern region, corresponding to locking depths ranging between 4 and 10 *km*
574 (see Fig. 5a).

575 Tab. 1 summarises the parameters considered in our calculations (fault length
576 L , locking depth L_D and slip rate \dot{u}) as well as estimates of the slip deficit u ,
577 geodetic moment M_0 , moment accumulation rate \dot{M}_0 and the associated mo-
578 ment magnitude M_W for each fault segment, which was computed based on
579 the scaling regression by [Hanks and Kanamori \(1979\)](#). The slip deficit and mo-
580 ment accumulation rates estimated for the Eilat and Arnona faults suggest that
581 impending earthquakes could reach M_W 6.7-7.3, assuming that the moment ac-
582 cumulated since the last known historical or instrumental earthquakes ("Last
583 EQ") would be released by an earthquake on each fault strand. Estimates of the
584 slip deficit and moment accumulated on the Aragonese fault are much smaller,
585 as expected, considering that this fault strand ruptured during the 1995 M_W 7.2
586 Nuweiba earthquake ([Klinger et al., 1999](#); [Shamir et al., 2003](#); [Hofstetter, 2003](#);

587 Baer et al., 2008; Ribot et al., 2021). Our results thus agree with previous stud-
588 ies suggesting that the Eilat and Arnona faults might be in a late stage of their
589 current interseismic period (e.g., Hamiel et al. 2018b; Kanari et al. 2020; Ribot
590 et al. 2021).

Table 1: Fault parameters and estimated earthquake potential for the main strike-slip fault segments in the Gulf of Aqaba.

Fault	L [km]	L_D [km]	\dot{u} [mm/yr]	Last EQ	u [m]	M_0 [Nm]	\dot{M}_0 [Nm/yr]	M_W
Eilat	59	4	4.4	1212	3.56	2.517×10^{19}	3.115×10^{16}	6.9
		10	5.2		4.20	7.437×10^{19}	9.204×10^{16}	7.2
	59	4	4.4	1588	1.90	1.346×10^{19}	3.115×10^{16}	6.7
10		5.2	2.25		3.976×10^{19}	9.204×10^{16}	7.0	
Aragonese	53	4	4.4	1995	0.11	6.996×10^{17}	2.798×10^{16}	5.9
		10	5.2		0.13	2.067×10^{18}	8.268×10^{16}	6.2
Arnona	83	4	4.4	1212	3.56	3.541×10^{19}	4.382×10^{16}	7.0
		10	5.2		4.20	1.046×10^{20}	1.295×10^{17}	7.3
	83	4	4.4	1588	1.90	1.893×10^{19}	4.382×10^{16}	6.8
10		5.2	2.25		5.593×10^{19}	1.295×10^{17}	7.1	
Tiran	50	Creep	4.4	-	-	-	-	-
			5.2					

591 Our modelling results reveal a residual left-lateral motion across the gulf
592 that cannot be resolved by standard models of interseismic strain accumulation
593 (Figs. 7 and 10). The nature of these velocity residuals is a subject of debate.
594 While some studies speculate that they could be caused by interseismic elastic
595 strain accumulation along the faults bounding the pull-apart basins in Gulf of
596 Aqaba (e.g., Pietrantonio et al. 2016), others point out that they could be due to
597 postseismic motions induced by the 1995 Nuweiba Earthquake (e.g., Piersanti
598 et al. 2001; Pe’eri et al. 2002; Riguzzi et al. 2006; Gomez et al. 2020).

599 We consider four possible explanations for the left-lateral residual motion
600 observed across the gulf. The first possibility is that anomalous misfits are due
601 to poor Euler vector parameters for the Sinai subplate relative to ITRF2014.
602 However, this explanation is not supported by the velocity field obtained for
603 stations of the GIL network in Israel, which exhibit residual velocities close to
604 zero, evidencing the robustness of the Euler vector parameters derived in our
605 study (see Tab. S2 and Fig. S2). A second possibility is that the resulting mis-
606 fits reflect the difference in temporal baselines between stations situated in the

607 northern and southern regions of the gulf as shown in Fig. S1. This hypothesis
608 might explain the small residuals obtained for stations with longer time series
609 located East of the Arnona fault (sites MAG1, MAG2 and BIDA, RASH, TAYS,
610 TAY1, TB03, SHRM). Nevertheless, it fails to explain the systematic nature of
611 residual motions.

612 A third hypothesis is that the relative motion between the Arabian and Sinai
613 plates along Gulf of Aqaba is accommodated by a wide deformation zone extend-
614 ing up to ~ 50 km across both plates. Several studies have identified two shear
615 belts composed of anastomosing faults on both sides of the gulf (e.g., Freund
616 et al. 1970; Bartov et al. 1980; Eyal et al. 1981; Garfunkel 1981; Lyberis 1988).
617 Earlier investigations indicate that these faults were active primarily during the
618 first slip stages along the DST and do not affect post-Miocene stratigraphic
619 units (e.g., Bartov et al. 1980; Eyal et al. 1981; Garfunkel 1981). A more recent
620 study reported small-scale normal faults affecting post-Miocene strata and Pleis-
621 tocene coral terraces indicating ENE-WSW extension (Bosworth et al., 2017).
622 However, the seismicity recorded in the gulf area shows that most earthquakes
623 concentrate along the gulf itself and the coastal plains (e.g., Ben-Avraham et al.
624 1979; Bartov et al. 1980, Fig. 2), suggesting that recent deformation is confined
625 to those areas.

626 Finally, the systematic left-lateral residual signal could reflect ongoing post-
627 seismic deformation caused by the Nuweiba Earthquake. A recent study by Liu
628 et al. (2020) found that the lower crust exhibits ductility at decadal time scales,
629 and thus postseismic transient motions may be larger and more enduring than
630 previously thought. As noted by Piersanti et al. (2001), the motion of stations
631 DAHA and NABQ on the Sinai side of the gulf is puzzling considering that
632 they exhibit motions that are opposite to the direction predicted by interseis-
633 mic models of strain accumulation (see Fig. 3). Similarly, the updated velocity
634 field presented in this study allowed us to quantitatively identify northeastward
635 residuals east of the Aragonese and Eilat faults, which agree with the overall
636 pattern of modelled postseismic rates.

637 We speculate that the misfit between the magnitudes of theoretical postseis-

mic motions and residuals from the back-slip model may arise from structural and rheological complexities in the gulf (e.g., Ben-Avraham 1985; El Khrepy et al. 2016a,b), which were not considered in our forward models. Therefore, postseismic motions and relaxation times could exhibit significant spatial variations depending on the lithospheric structure in different regions of the gulf and the slip distribution used to model the Nuweiba earthquake.

8. Conclusions

The establishment of 27 survey-mode GPS stations in 2014 and the analysis of up to 19 years of geodetic observations on the Arabian and Sinai-Levant plates allowed us to derive an updated crustal motion velocity field of the lands bordering the Gulf of Aqaba. By implementing an elastic interseismic model, we derived a slip rate of $4.9^{+0.9}_{-0.6} \text{ mm/yr}$ and a locking depth of $6.8^{+3.5}_{-3.1} \text{ km}$ across the Eilat fault in the gulf's northern region. Our modelling results suggest that either a shallow locking depth or creeping might be required to explain the low interseismic strain accumulation observed in the gulf's southern region across Tiran fault. We identified a systematic left-lateral residual motion across the gulf, characterised by northeastward residuals east of the Aragonese fault and southwestward residuals west of the Arnona fault. We postulate that postseismic transient motions caused by the 1995, $M_W 7.2$ Nuweiba earthquake could potentially explain these misfits. Estimates of the geodetic moment accumulated on the Eilat and Arnona faults since the last historical earthquakes that struck the gulf area in 1212 AD and 1588 AD indicate that impending earthquakes on these faults could potentially reach $M_W 6.7-7.3$, posing a significant hazard to the urban centres along the gulf coast. Future studies would benefit from incorporating additional GPS stations on the Sinai side of the gulf, refined slip models of the Nuweiba earthquake and more detailed information about historical earthquakes in the gulf region.

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676 **Data Availability**

677 The **GPS** data underlying this article will be made available on the UNAVCO
678 archive, at <https://www.unavco.org/data/dai/>.

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