Non-triggering and then triggering of a repeating aftershock sequence in the Dead Sea by the 2023 Kahramanmaraş earthquake pair: Implications for the physics of remote delayed aftershocks

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June 23, 2023

Abstract

Most aftershocks occur in areas experiencing large co-seismic stress changes, yet some occur long after the mainshock in remote lightly-stressed regions. The triggering mechanism of these remote delayed aftershocks is not well understood. Here, we study aftershocks occurring in the Dead Sea (DS) area following the 2023 Mw7.8 and Mw7.6 Kahramanmaraş earthquakes. Most aftershocks cluster along previously quiescent structures off the main the DS fault strand. Visual inspection disclosed three aftershocks instantaneously triggered by the Mw7.6 in the Northern DS basin, and match-filtering revealed a delayed aftershock. Waveform similarity and temporal clustering suggest the northern DS aftershocks re-rupture a stick-slip patch loaded by surrounding creep. Velocity-gradient seismograms show the Mw7.6 exerted larger transient stresses than the Mw7.8, which may explain triggering by the Mw7.6, but not by the Mw7.8. This account of instantaneously-triggered repeaters underscores the role of interactions between aseismic and seismic slip in remote triggering.

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² aftershock sequence in the Dead Sea by the 2023

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1. Abstract

Most aftershocks occur in areas experiencing large co-seismic stress changes, yet some occur 5 long after the mainshock in remote lightly-stressed regions. The triggering mechanism of these 6 remote delayed aftershocks is not well understood. Here, we study aftershocks occurring in the 7 Dead Sea (DS) area following the 2023 $M_w7.8$ and $M_w7.6$ Kahramanmaraş earthquakes. Most aftershocks cluster along previously quiescent structures off- the main the DS fault strand. Visual 9 inspection disclosed three aftershocks instantaneously triggered by the $M_w 7.6$ in the northern 10 DS basin, and match-filtering revealed a delayed aftershock. Waveform similarity and temporal 11 clustering suggest the northern DS aftershocks re-rupture a stick-slip patch loaded by surrounding 12 creep. Velocity-gradient seismograms show the $M_w7.6$ exerted larger transient stresses than the 13 $M_w7.8$, which may explain triggering by the Mw7.6, but not by the Mw7.8. This account of 14

¹⁵ instantaneously-triggered repeaters underscores the role of interactions between aseismic and
 ¹⁶ seismic slip in remote triggering.

2. Introduction

Aftershocks play an important role in the relaxation and redistribution of stress perturbations 17 caused by major earthquakes [Ben-Zion et al., 2003; Ziv and Schmittbuhl, 2003]. While the vast 18 majority of them occur long after the passage of the seismic waves, a tiny fraction may occur 19 during the surface-wave train [Hill et al., 1993; Brodsky et al., 2000]. From a physical standpoint, 20 it is useful to categorize aftershocks according to their time of occurrence with respect to the 21 passage of the seismic waves excited by the mainshock source. Here, aftershocks that occurred 22 during the wave-induced ground shaking are referred to as instantaneous aftershocks, and those 23 that occurred afterward are referred to as delayed aftershocks. It is well established that the 24 majority of the delayed aftershocks occur in places where the static (i.e., permanent) stress 25 changed in a manner that encourages fault slip [King et al., 1994]. Yet, some of them occur 26 in remote sites, where the stress changes caused by the mainshock are tiny [Gomberg et al., 27 2001; Kilb et al., 2000]. The physical mechanism that gives rise to remote delayed aftershocks is 28 not well understood. Furthermore, not all delayed aftershocks are caused directly by the stress 29 change imparted by the mainshock. Instead, a notable fraction of them is induced by multiple 30 stress transfers from early aftershocks to the site of subsequent ones [Helmstetter et al., 2003; 31 Marsan, 2005; Ziv, 2006a, b], or by remotely triggered fault creep [Inbal et al., 2017]. Aftershocks 32 of aftershocks and triggered creep may occur in areas where the stress perturbation (positive or 33 negative) induced by the mainshock is small, and may thus explain the occurrence of some 34 delayed aftershocks in remote sites [Ziv, 2006a, b; Inbal et al., 2017]. 35

The 2023 $M_w7.8$ and $M_w7.6$ Kahramanmaraş earthquake pair from Eastern Turkey, caused 36 a dramatic earthquake rate increase over vast areas (Figure 1a). This activity was very well 37 recorded by strong-motion sensors located along the northern Dead Sea (DS) basin, which are 38 part of the dense Israel Seismic Network (ISN) for earthquake early warning - the TRUAA 39 network (Kurzon et al., 2020). In this study we examine a sequence of four repeating aftershocks 40 in the northern DS basin, three of which are instantaneous aftershocks in the sense that they 41 occurred during the passage of the surface waves of the $M_w7.6$ earthquake (Figure 1b). The 42 importance of this finding is twofold. First, the triggering of aftershocks on a portion of the 43 Dead Sea Transform (DST) that has not ruptured in recent history, suggests that rupture of that 44 segment may have been promoted by the Kahramanmaras earthquakes. Second, the observations 45 shed new light on the triggering mechanism of remote aftershocks. 46

3. The remotely triggered Dead Sea aftershock sequence

3.1. Remote delayed aftershocks throughout Palestine and Israel

The DST, which accommodates 4-5 mm/yr left-lateral motion between the Sinai sub-plate and 47 the Arabian plate (Figure 1a), is the most active fault affecting the study area. Yet, the most 48 intense earthquake rate change due to the 2023 Kahramanmaraş earthquakes occurred off the 49 main DST strand. For example, just south of the Palestinian city of Nablus, the earthquake 50 rate has surged by more than two orders of magnitude (Figure 1a). Up to the time interval 51 reported here, fault slip-rates in that region were reported to be below 0.5 mm/year, and thus 52 these faults were considered to be seismically inactive [Sharon et al., 2020]. Another area of 53 notable aftershock activity is located in the Mediterranean Sea, offshore Haifa (Figure 1a). This 54 activity is likely associated with the seaward extension of the extensional Carmel Fault, or with 55 the southern extension of the Mount Lebanon Thrust. Two other notable but spatially-limited 56



Figure 1: Study area location maps and earthquake count. (a) A map of aftershock rate change, calculated for a sliding window of 0.25 by 0.25 degrees, using: $(N_{as}/T_{as})/(N_{bg}/T_{bg})$, where "as" and "bg" stand for aftershock and background intervals, respectively, N signifies earthquake count, T signifies the interval, and T_{as} and T_{bg} are equal to 32 days and 10 years, respectively. The dashed square indicates the location of the map shown in panel b, dashed line signifies political borders, and solid line marks the location of the DST. DST, Dead Sea Transform, DS, Dead Sea; MS, Mediterranean Sea; GE, Gulf of Eilat; N, Nablus; and H, Haifa. (b) A map showing the epicenters of the 4 repeating aftershocks (circles), and the TRUAA's strong-motion stations in that area (triangles). (c) Study area earthquakes count as a function of time with respect to the time of the 2023 $M_w 7.8$ earthquake. The circles mark the three local aftershocks with M>3.3, and the arrow marks a point of notable aftershock rate increase.

rate increases are resolved east of the southern DS basin and northeast of the Gulf of Eilat. In contrast, the northern DS basin, which is a major pull-apart basin located along the DST, has experienced only a modest earthquake rate increase. In light of these observations, we infer that the pre-stress levels on the structures on either side of the main DST strand are presently higher than on the DST system itself. The implications of this result for earthquake risk assessment, and the development of earthquake mitigation strategies are far-reaching and warrant further consideration.

A plot of cumulative earthquake count as a function of time discloses complex temporal cluster-64 ing, with large variations in the rate of earthquake occurrence (Figure 1c). Some of the jumps in 65 the earthquake rate are attributable to local M > 3.3 aftershocks that act locally as mainshocks 66 of small aftershock sequences (marked by empty circles on Figure 1c), whereas others, but most 67 notably the one marked by an arrow, can neither be associated with local mainshocks, nor with 68 regional ones. That notable increase is mostly attributable to a seismic swarm located to the 69 south of Nablus. Given the timing relative to the Kahramanmaras earthquakes, the pervasive 70 lack of instantaneous triggering, and the lack of notable local mainshocks, it is likely that addi-71 tional mechanisms triggered the seismicity showing up as unexplainable steps in the cumulative 72 earthquake count versus time diagram. One possibility is the Kahramanmaras earthquake pair 73 triggered aseismic fault slip, which may be detectable geodetically. 74

3.2. Discerning instantaneous aftershocks

In search of local earthquakes within the surface and coda waves of the Kahramanmaraş mainshocks, we scanned all the ISN recorded strong-motion and broad-band seismograms of the $M_w7.8$ and $M_w7.6$ earthquakes. By band-pass filtering the seismograms between 8 and 16 Hz, we identified 3 microearthquakes that occurred during passage of surface waves excited by the $M_w7.6$ (Figure 2a). These are best observed in the strong-motion seismograms recorded by the MSLM and DSI stations, and to a lesser extent also by TMR, OVNT, EFSH, and KEGD. The P- and S-phase arrivals of the earliest aftershock to MSLM are well resolved (Figure 2b, 2c, and S2), and constrain its hypocentral distance to about 16 kilometers. Low-pass filtered ground velocity (integrated accelerations) time-series at station MSLM are shown in Figure 2a. Inspection of these, as well as other, low-pass filtered accelerograms indicate that the strongest ground motion in the DS area is measured on the transverse component, which is associated with Love wave propagation. A detailed analysis of the surface-wave induced strain in the northern DS area is presented in Section 6.1.

3.3. Supplementing the Dead Sea aftershock catalog using Match Filter Analysis

To identify multiplets of event #1 (registered or not in the GSI catalog), we implemented 88 the template-based Match Filter Analysis (MFA; *Gibbons and Ringdal* [2006]). We extracted 89 a three-component template waveform, consisting of a portion of the P-wave train excited by 90 event #1. The template duration is 1.8 s, starting from 0.3 s prior to the P-wave arrival at 91 the four nearest stations (in Figure 1b), and filtered between 8 to 16 Hz. The MFA detection 92 process operates as follows: First, we correlate the template waveform across a 2.13 year-long 93 time-series (segmented into days) starting from January 1, 2021, and ending on February 21, 94 2023, with 90% overlap between adjacent temporal windows. Next, we correlate the correlations 95 of the vertical channels carrying the template across stations, thus deriving the inter-station time 96 offsets corresponding to the template. Finally, we shift the template cross-time three-component 97 correlations according to the cross-station offsets, and stack the cross-correlations. These were 98 scanned for multiplets. For a positive hit, we require the value of this function to exceed its 99 1-hour median by 9 times the Median Absolute Deviation of that hour. We turn each successful 100 hit into a template, and repeat the same procedure as for event #1. 101

¹⁰² Analysis of the 9 hour interval between the $M_w7.8$ and $M_w7.6$ mainshocks did not yield any ¹⁰³ positive detections, and the analysis of the 17 days following the $M_w7.6$ earthquake yielded 3 hits.



Figure 2: Ground acceleration and velocity recorded at stations MSLM and DSI showing the surface wave arrivals, the instantaneously triggered earthquake and the first two aftershocks. V, T, and R are for the vertical, transverse, and radial component of motion, respectively. The numbers refer to the events listed in Table 1. (a) Band- and low-pass filtered seismograms from MSLM. Top trace is band-pass filtered between 8 to 16 Hz, and the three bottom traces are low-pass filtered above 0.1 Hz. Time is referenced to the $M_w7.6$ origin time. Arrows show arrivals due to events #1 to #3. (b,c) Vertical accelerograms containing the P-wave train of the 4 earthquakes of the northern DS basin repeating aftershock. The traces are filtered between 8 to 16 Hz and aligned on the P-wave arrival. The vertical bars are for scale. Event #1 was instantaneously triggered during the passage of surface waves excited by the February 6, 2023 $M_w7.6$ Kahramanmaraş earthquake. Left panel: Station MSLM. Right panel: Station DSI.

We did not detect similar events occurring in the two years before the M_w 7.8. The newly detected 104 earthquake source parameters are listed in Table 1 (Supplementary Materials), and their P-wave 105 seismograms are shown in Figure 2. They exhibit great similarity to the instantaneously triggered 106 earthquake template (event #1), especially at station MSLM, located a few km from event #1107 epicenter (Figure 2). Between February 6 to February 17, 2023, the ISN located 6 earthquakes 108 occurring within 15 km of event #1, one of them passed our detection criteria (event #4, see 109 Table 1). Based on waveform similarity, we suggest that the 4 aftershocks re-ruptured an isolated 110 fault patch. The MFA further indicates that this patch was previously inactive for at least two 111 years. 112

4. Master event location

The four hypocenters were located in two steps. In the first, we implemented a grid-search 113 approach, and used the 1-D velocity model of *Gitterman et al.* [2003]. In the second step, we used 114 the master event location approach to get relative locations [*Fitch*, 1975]. We chose the latest 115 earthquake (event #4) as a master event, as its initial location was most tightly constrained. 116 High waveform similarity allowed us to improved the P-phase picking of the 3 slave events by 117 cross-correlation master event P-wave arrivals. Furthermore, we fixed the depths of the slave 118 events, to that of the master. The result of the master-slave event locations (Figure 1b) indicates 119 the four ruptures overlap, supporting the view of repeating aftershocks. 120

5. Temporal variations

5.1. Recurrence times

The spatial clustering of the four earthquakes, and their strong waveform similarity are consistent with the possibility that this series resulted from repeated slip events on an isolated

stick-slip patch that is embedded within an otherwise aseismic fault [Vidale et al., 1994; Nadeau 123 et al., 1994; Marone et al., 1995]. The effect of perturbing the stress of such a configuration is 124 to increase the creep rate on the aseismic segment. If the slip on that segment is resisted by a 125 logarithmic rate- strengthening friction, the rate of creep is expected to decay as 1/t following 126 a stress step, with t being the elapsed time since the stress perturbation [Schaff et al., 1998; 127 Chen and Lapusta, 2009. Consequently, the stressing rate on the stick-slip patch varies with the 128 reciprocal of time, and the following relation between the recurrence time, τ , and t is predicted 129 [Schaff et al., 1998]:130

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 $\frac{1}{\tau_i} = \frac{a}{t_i},\tag{1}$

where a is an empirically determined fitting coefficient, and i is the earthquake index. The right-132 hand side of (1) is similar to that of the well-known Omori law [Omori, 1894], which provides 133 good fit to the vast majority of aftershock sequences observed following major earthquakes. In 134 contrast to the Omori law, this function is equated here to the reciprocal of the recurrence time, 135 which may be thought of as the instantaneous rate of earthquake occurrence. If instead, the creep 136 is resisted by viscous stresses (with friction being linearly proportional to slip rate), the rate of 137 creep is expected to decay exponentially [Segall, 2010], and so is the instantaneous earthquake 138 rate: 139

$$\frac{1}{\tau_i} = a \exp(-bt),\tag{2}$$

with *b* being another fitting coefficient. Relations (1) and (2), hereafter refer to as the Omori and Exponential models, respectively, provide means for corroborating the isolated stick-slip patch model, and discriminating between the two friction models. The reciprocal of the repeating quakes recurrence time as a function of time since the first aftershock is shown in Figures 3a on a log-log space. Because the number of data is small, it is not evident which relation best fits the recurrence intervals. To quantify the degree to which the temporal evolution of this earthquake
 cluster is more consistent with relation (1) than (2), an Omori-versus-Exponential Fitting Index
 (FI) is introduced:

$$FI = \frac{L_{\rm Exp}^2 - L_{\rm Omori}^2}{L_{\rm Exp}^2 + L_{\rm Omori}^2},\tag{3}$$

where L^2_{Omori} and L^2_{Exp} are the norm-2 misfits to equations 1 and 2, respectively. This function 150 may range between -1 and +1, with FI=-1 if the $1/\tau$ time-series is perfectly modeled by the 151 exponential relation in (2), and FI=1 if it is perfectly modeled by the Omori function in (1). 152 The fitting index of the time-series in question is equal to 0.8, thus strongly suggesting that the 153 data are better fit with (1) than with (2). Yet, given the small dataset, it is important to assess 154 the extent to which this result is statistically meaningful. In the supplementary materials we 155 present the result of a Monte Carlo testing, showing that the possibility of FI=0.8 emerging by 156 chance may be rejected at greater than 96% confidence level. 157

5.2. Relative seismic moments

¹⁵⁸ We assume the earthquake spectra follow the well-known ω^2 model [*Aki*, 1967; *Brune*, 1970], ¹⁵⁹ and recover the source properties directly from the observed spectral amplitudes. The low-¹⁶⁰ frequency displacement spectral plateau Ω_0 and the seismic moment are related as [*Aki and* ¹⁶¹ *Richards*, 2002]:

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$$M_0 = \Omega_0 RC,\tag{4}$$

where R is the hypocentral distance and collects site and path parameters and the radiation pattern. Note that the seismic moment estimates are insensitive to the decay rate at frequencies well-above the source' characteristic frequency. The low-frequency spectral plateau, Ω_0 , may be computed in the frequency or time domain. Here, Ω_0 is estimated using the simple time-domain



Figure 3: Summary of temporal analysis. (a) Log-log diagram of the reciprocal of recurrence time as a function of time since the first aftershock. (b) Relative moment as a function of recurrence time for events 2-4, calculated separately for accelerograms recorded at stations MSLM (gray circles) and DSI (black squares). Red symbols correspond to the N + 1 predictions.

 $_{167}$ expression of *Lior and Ziv* [2020]:

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$$\Omega_0 = 2\sqrt{T} D_{rms}^{1.5} / V_{rms}^{0.5},\tag{5}$$

where T is the data interval, and D_{rms} and V_{rms} are the displacement and velocity root-meansquares (rms), respectively. The above expression is valid in the far-field, and does not require prior knowledge of the source corner frequency and attenuation effects. For these reasons, use of this time-domain expression is most advantageous for small earthquakes and/or data of limited frequency band-width [*Lior and Ziv*, 2020]. Because repeating earthquakes recorded at a given station share the same site and path parameters, the seismic moment of earthquake j relative to that of earthquake k, calculated using seismograms recorded by station i may expressed as:

$$\frac{M_0^{ij}}{M_0^{ik}} = \frac{(D_{rms}^{1.5}/V_{rms}^{0.5})_{ij}}{(D_{rms}^{1.5}/V_{rms}^{0.5})_{ik}}.$$
(6)

We calculated relative seismic moments from doubly-integrated accelerations at MSLM and 177 DSI using 4 seconds-long intervals starting at the S-phase onset. Because events 1 to 3 coincide 178 with the passage of the surface waves of the M_w 7.6 Kahramanmaraş earthquake, they may only be 179 studied in a limited frequency band. Therefore, a 6-18 Hz bandpass filter was applied following 180 each integration. The relative seismic moments exhibit a 20% steady increase of the relative 181 moment per 10-fold increase time elapsed since event 4. (Figure 3b). This trend is consistent 182 with experiments showing that friction increases with the logarithm of time since the previous 183 slip episode [Dieterich, 1972; Scholz et al., 1972; Scholz and Engelder, 1976]. 184

6. Discussion

The DS repeating aftershock sequence reported in this study differs from previously reported ones in two main respects. First, while previous repeaters were located in the near-field of the mainshock the DS sequence is clearly located at its far-field. Second, the triggering mainshock occurred just 9 hours after a larger mainshock that did not trigger instantaneous aftershocks in the study area or delayed aftershocks near the DS sequence. These differences create an opportunity for revisiting previous paradigms, and introducing new ones.

6.1. Triggering mechanism for delayed remote aftershocks

It is difficult to explain the triggering of remote delayed aftershocks by means of stress transfer because of the small permanent stress changes. In contrast, explaining the occurrence of instantaneous aftershocks is more straightforward because their magnitude seems sufficient for triggering of critically-stressed faults [*Gomberg and Johnson*, 2005]. Thus, if instantaneous triggering was a common phenomena, one could have imagined a stress transfer process, whereby delayed aftershocks are triggered by the instantaneous aftershocks that act locally as small main¹⁹⁷ shocks [Helmstetter et al., 2003; Ziv, 2006a; Heimisson, 2019]. Because, however, our search for
¹⁹⁸ instantaneous triggering yielded only 3 co-located aftershocks within the northern DS basin, this
¹⁹⁹ triggering mechanism cannot explain the widespread delayed aftershocks reported in this study
²⁰⁰ (Figure 1a). Here we suggest that the observations presented in this study shed new light on the
²⁰¹ triggering mechanism of remote aftershocks.

It is a widely held notion that the re-rupturing of a given fault patch, with recurrence intervals 202 that decay as the reciprocal of elapsed time since the first aftershock, requires strong interaction 203 with a velocity-strengthening creeping segment [Schaff et al., 1998]. As this sequence occurred 204 in the far-field of its trigger, and in an area that did not experience any earthquake rate increase 205 prior to its onset, we can further infer that remote delayed aftershocks result from the stress 206 change induced by local creep. That creep triggered by seismic waves excited by the mainshock. 207 Once in motion, they stress their surroundings, and trigger instantaneous or delayed aftershocks. 208 Indeed, the sequence documented in this study consists of four aftershocks, of which the first 209 three are instantaneous, and the fourth is delayed. The interaction between aseismic and seismic 210 slip is rather complex. The stick-slip patch is being stressed by the surrounding creep during 211 inter-event intervals, and is stressing it with each slip episode. Thus, deeper understanding of 212 this process may require forward modeling. 213

6.2. Implications of the non-triggering by the $M_w 7.8$ earthquake

That the repeating aftershock sequence was triggered instantaneously by a $M_w7.6$ earthquake just 9 hours after an equally distant $M_w7.8$ earthquake raises the question as to what may be learned from the non-triggering of that sequence by the largest of the two. The non-triggering by the $M_w7.8$ and the triggering by the $M_w7.6$ earthquakes just 9 hours later may be understood if the transient stress change induced by the latter was notably larger than that induced by the



Figure 4: Ground velocity and its spatial gradient as a function of time. \dot{u}_t denotes the transverse component of velocity. $\dot{u}_{t,r}$, $\dot{u}_{r,r}$, and $\dot{u}_{v,r}$ are for gradients of the transverse, radial, and vertical velocity components computed along the radial direction, respectively. Number in percentage denotes the difference between the maximum $M_w7.6$ and $M_w7.8$ gradient maxima relative to the maximum amplitude of the $M_w7.8$ for each component (i.e. positive is for increase in the $M_w7.6$ maxima gradient amplitude). a. Station DSI low-pass filtered above 1 Hz. b-d. The velocity gradient computed using differences between DSI and MSLM (see Figure 1 for station locations). Seismograms were band-pass filtered between 32 to 16 s before differentiation. Units are in micro-strain/s.

former. Indeed the Peak-Ground-Velocity (PGV) induced by the $M_w7.6$ in the DS area is by about 10% larger than that induced by the $M_w7.8$. Equating the peak strain to the product between the PGV and the slowness [e.g *Benioff*, 1935; *Mikumo and Aki*, 1964] yields an equally larger strain - and therefore also stress change - for the $M_w7.6$ than for the $M_w7.8$. It is difficult to attribute the notably different triggering effect of the two earthquakes to a 10% difference in the stress perturbations at the same site.

Gomberg and Aqnew [1996] reported large discrepancies between strain calculation using the 225 above approach and direct strain-meter measurements. Thus, to address the question as to 226 what may be learned from the non-triggering by the M_w 7.8, it is instructive to obtain a more 227 direct strain measurement. Since strain-meters are not available in the study area, we implement 228 an alternative approach that takes advantage of the network's alignment being sub-parallel to 229 the radial direction pointing from the northern DS basin to the $M_w7.6$ source. By taking the 230 difference between the same-component long-period (>10 s) velocity time-series measurements 231 at two nearly radially-aligned stations, and dividing them by the inter-station distance projected 232 onto the radial direction, we obtained timelines of the following velocity gradients: $\dot{u}_{r,r}$, $\dot{u}_{t,r}$, and 233 $\dot{u}_{v,r}$, where t, r, and v represent the transverse, radial, and vertical components, respectively. 234 In doing so, we require that the inter-station distances be less than about a quarter of the 235 seismic wave-length (after filtering). The main advantage of this approach with respect to the 236 one described above is that it does not rest on the assumption that the slowness between each 237 station pair equals the slowness measured across the network. Under the assumption that the 238 waves can be treated as perfectly planar, the strain rate tensor components can be approximated 239 from the measured gradients [Gomberg and Agnew, 1996]. In this case $\dot{\varepsilon}_{rr} \approx \dot{u}_{r,r}$, $\dot{\varepsilon}_{tr} \approx 1/2\dot{u}_{t,r}$, 240 and $\dot{\varepsilon}_{vr} \approx 1/2\dot{u}_{v,r}$, where $\dot{\varepsilon}_{rr}$ and $\dot{\varepsilon}_{tr}$ are the horizontal dilatational and shear strains, respectively, 241 and $\dot{\varepsilon}_{vr}$ is the shear strain acting on vertical planes parallel to radial direction. 242

The results of our new approach for deformation rate analysis are presented in Figure 4 for the two nearest stations to the DS aftershock sequence - MSLM and DSI. We find that the differences between $M_w7.8$ and the $M_w7.6$ are more notable in the deformation rate domain than they are in the raw velocity time-series. Comparing the $M_w7.8$ and the $M_w7.6$ deformation rate timeseries, we find that the most notable positive gradient increases are observed on the transverse

and radial components, whereas the peak M_w 7.6 vertical component gradient is reduced relative 248 to the M_w 7.8. The peak dynamic long-period (>10 s) transverse velocity gradient due to the 249 $M_w 7.6$ is about 30% larger than the $M_w 7.8$ peak gradient, however individual phases that differ 250 by up to a factor of two appear after differentiation (Figure 4b). We examined the deformation 251 rate gradients using all neighboring station pairs along the western shore of the DS lake (Figure 252 1a). This indicated that M_w 7.6 transient shear strain is increased by 30% to 40% relative to the 253 $M_w 7.8$ at all the northern DS basin station pairs, that dilatational strain peak for MSLM-DSI 254 increased by 30% but was reduced at the adjacent station pairs, and that the $M_w7.6$ transient 255 shear stress peaks halfway between MSLM and DSI, in agreement with the repeater location. 256

Two source-related effects may combine to produce 10% stronger ground shaking and 30% 257 to 40% strain rate for the M_w 7.6 than for the M_w 7.8. The first factor, affecting mainly the 258 long-period strain rates, is the azimuthal dependence of the Love wave radiation pattern. The 259 M_w 7.6 fault strikes approximately east-west, and is thus well oriented with respect to the Love-260 wave radiation towards the DS, while the $M_w 7.8$ strikes approximately north-north-east, thus 261 producing minimum Love-wave radiation in that direction. This eastward strike rotation is 262 sufficient in order to offset the level of seismic energy radiated from the $M_w7.6$ source to an 263 amplitude comparable to the one radiated by the $M_w 7.8$ source. A second factor, affecting 264 mostly the short-period ground shaking between the two mainshocks is their static stress drops. 265 The $M_w 7.8$ ruptured a relatively large fault area of about 1400 km² that slipped on average by 266 about 5 m, whereas the M_w 7.6 ruptured about half the M_w 7.8 fault area, and its average slip 267 is about 2.5 m [Barbot et al., 2023]. Since the stress drop is proportional to the ratio between 268 the average slip and the fault length, those differences decreased the $M_w 7.8$ stress drop by a 269 factor of about 3 relative to the M_w 7.6 stress drop. We suggest that the larger stress drop and 270

7. Conclusive remarks

The 2023 Kahramanmaraş earthquake pair triggered widespread aftershock activity throughout Palestine and Israel. Surprisingly, the strongest earthquake rate increase is observed on either side of the DST and in the Mediterranean Sea offshore Haifa, suggesting that secondary structures in the study area are presently more prestressed than the DST main strand. Earthquake count as a function of time reveals complex evolution, with activity bursts that are attributable to $M_w > 3$, which act as local mainshocks. Other seismicity rate jumps that cannot be explained by local triggering, may be indicative of aseismic deformations.

Visual inspection of high-pass filtered accelerograms of the M_w 7.8 and M_w 7.6 Kahramanmaraş earthquakes discloses three instantaneously triggered aftershocks of the latter in the northern DS basin. A delayed aftershock is revealed via MFA. Based on strong waveform similarity and co-location, with location uncertainties that are smaller than the estimated rupture dimensions, it is inferred that the four aftershocks correspond to a sequence of repeaters. Temporal analyses are presented, which indicate that the logarithm of repeat time and the seismic moment increase linearly with the logarithm of elapsed time since the first aftershock.

This first account of an instantaneously triggered repeating aftershock sequence at the farfield of its trigger underscores the role of the interaction between aseismic and seismic slip in remote aftershock triggering [*Inbal et al.*, 2017]. A comparison between peak ground shaking and spatial derivative of velocity seismograms of the two Kahramanmaraş earthquakes suggests larger transient stress change in the DS during the $M_w7.6$ than during the $M_w7.8$. This may explain why the DS aftershocks were triggered by the $M_w7.6$, but not the $M_w7.8$, despite the latter preceding the former by just a few hours.

Finally, given a sequence of $N \geq 3$ after shocks, with a τ versus t relation obeying Equation 294 1, the timing of the N + 1 aftershock may be predicted using: $t_{i+1} = t_i \frac{a}{a-1}$. Thus, should the 295 current $1/\tau$ trend continue, the next aftershock is expected to occur about 278 days after the 296 mainshock (using a=1.04 and $t_4=10.72$ days, see red open circle in Figure 3a). Furthermore, 297 from the relative moment versus time plot in Figure 3b, the size of that earthquake is expected 298 to be 20% larger than that of the largest aftershock. These predictions may require some revision 299 in case that the local stress field would be altered by some external source, such as a local or 300 regional earthquake, or a slow slip episode on a nearby segment. 301

302 Acknowledgments.

Open Research. Data from the IS network are available at https://doi.org/10.5281/zenodo.7980346. The seismic catalog is available at https://seis.gsi.gov.il/fdsnws/event/.

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Non-triggering and then triggering of a repeating

² aftershock sequence in the Dead Sea by the 2023

³ Kahramanmaraş earthquake pair: Implications for

the physics of remote delayed aftershocks

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1. Abstract

Most aftershocks occur in areas experiencing large co-seismic stress changes, yet some occur 5 long after the mainshock in remote lightly-stressed regions. The triggering mechanism of these 6 remote delayed aftershocks is not well understood. Here, we study aftershocks occurring in the 7 Dead Sea (DS) area following the 2023 $M_w7.8$ and $M_w7.6$ Kahramanmaraş earthquakes. Most aftershocks cluster along previously quiescent structures off- the main the DS fault strand. Visual 9 inspection disclosed three aftershocks instantaneously triggered by the $M_w 7.6$ in the northern 10 DS basin, and match-filtering revealed a delayed aftershock. Waveform similarity and temporal 11 clustering suggest the northern DS aftershocks re-rupture a stick-slip patch loaded by surrounding 12 creep. Velocity-gradient seismograms show the $M_w7.6$ exerted larger transient stresses than the 13 $M_w7.8$, which may explain triggering by the Mw7.6, but not by the Mw7.8. This account of 14

¹⁵ instantaneously-triggered repeaters underscores the role of interactions between aseismic and
 ¹⁶ seismic slip in remote triggering.

2. Introduction

Aftershocks play an important role in the relaxation and redistribution of stress perturbations 17 caused by major earthquakes [Ben-Zion et al., 2003; Ziv and Schmittbuhl, 2003]. While the vast 18 majority of them occur long after the passage of the seismic waves, a tiny fraction may occur 19 during the surface-wave train [Hill et al., 1993; Brodsky et al., 2000]. From a physical standpoint, 20 it is useful to categorize aftershocks according to their time of occurrence with respect to the 21 passage of the seismic waves excited by the mainshock source. Here, aftershocks that occurred 22 during the wave-induced ground shaking are referred to as instantaneous aftershocks, and those 23 that occurred afterward are referred to as delayed aftershocks. It is well established that the 24 majority of the delayed aftershocks occur in places where the static (i.e., permanent) stress 25 changed in a manner that encourages fault slip [King et al., 1994]. Yet, some of them occur 26 in remote sites, where the stress changes caused by the mainshock are tiny [Gomberg et al., 27 2001; Kilb et al., 2000]. The physical mechanism that gives rise to remote delayed aftershocks is 28 not well understood. Furthermore, not all delayed aftershocks are caused directly by the stress 29 change imparted by the mainshock. Instead, a notable fraction of them is induced by multiple 30 stress transfers from early aftershocks to the site of subsequent ones [Helmstetter et al., 2003; 31 Marsan, 2005; Ziv, 2006a, b], or by remotely triggered fault creep [Inbal et al., 2017]. Aftershocks 32 of aftershocks and triggered creep may occur in areas where the stress perturbation (positive or 33 negative) induced by the mainshock is small, and may thus explain the occurrence of some 34 delayed aftershocks in remote sites [Ziv, 2006a, b; Inbal et al., 2017]. 35

The 2023 $M_w7.8$ and $M_w7.6$ Kahramanmaraş earthquake pair from Eastern Turkey, caused 36 a dramatic earthquake rate increase over vast areas (Figure 1a). This activity was very well 37 recorded by strong-motion sensors located along the northern Dead Sea (DS) basin, which are 38 part of the dense Israel Seismic Network (ISN) for earthquake early warning - the TRUAA 39 network (Kurzon et al., 2020). In this study we examine a sequence of four repeating aftershocks 40 in the northern DS basin, three of which are instantaneous aftershocks in the sense that they 41 occurred during the passage of the surface waves of the $M_w7.6$ earthquake (Figure 1b). The 42 importance of this finding is twofold. First, the triggering of aftershocks on a portion of the 43 Dead Sea Transform (DST) that has not ruptured in recent history, suggests that rupture of that 44 segment may have been promoted by the Kahramanmaras earthquakes. Second, the observations 45 shed new light on the triggering mechanism of remote aftershocks. 46

3. The remotely triggered Dead Sea aftershock sequence

3.1. Remote delayed aftershocks throughout Palestine and Israel

The DST, which accommodates 4-5 mm/yr left-lateral motion between the Sinai sub-plate and 47 the Arabian plate (Figure 1a), is the most active fault affecting the study area. Yet, the most 48 intense earthquake rate change due to the 2023 Kahramanmaraş earthquakes occurred off the 49 main DST strand. For example, just south of the Palestinian city of Nablus, the earthquake 50 rate has surged by more than two orders of magnitude (Figure 1a). Up to the time interval 51 reported here, fault slip-rates in that region were reported to be below 0.5 mm/year, and thus 52 these faults were considered to be seismically inactive [Sharon et al., 2020]. Another area of 53 notable aftershock activity is located in the Mediterranean Sea, offshore Haifa (Figure 1a). This 54 activity is likely associated with the seaward extension of the extensional Carmel Fault, or with 55 the southern extension of the Mount Lebanon Thrust. Two other notable but spatially-limited 56



Figure 1: Study area location maps and earthquake count. (a) A map of aftershock rate change, calculated for a sliding window of 0.25 by 0.25 degrees, using: $(N_{as}/T_{as})/(N_{bg}/T_{bg})$, where "as" and "bg" stand for aftershock and background intervals, respectively, N signifies earthquake count, T signifies the interval, and T_{as} and T_{bg} are equal to 32 days and 10 years, respectively. The dashed square indicates the location of the map shown in panel b, dashed line signifies political borders, and solid line marks the location of the DST. DST, Dead Sea Transform, DS, Dead Sea; MS, Mediterranean Sea; GE, Gulf of Eilat; N, Nablus; and H, Haifa. (b) A map showing the epicenters of the 4 repeating aftershocks (circles), and the TRUAA's strong-motion stations in that area (triangles). (c) Study area earthquakes count as a function of time with respect to the time of the 2023 $M_w 7.8$ earthquake. The circles mark the three local aftershocks with M>3.3, and the arrow marks a point of notable aftershock rate increase.

rate increases are resolved east of the southern DS basin and northeast of the Gulf of Eilat. In contrast, the northern DS basin, which is a major pull-apart basin located along the DST, has experienced only a modest earthquake rate increase. In light of these observations, we infer that the pre-stress levels on the structures on either side of the main DST strand are presently higher than on the DST system itself. The implications of this result for earthquake risk assessment, and the development of earthquake mitigation strategies are far-reaching and warrant further consideration.

A plot of cumulative earthquake count as a function of time discloses complex temporal cluster-64 ing, with large variations in the rate of earthquake occurrence (Figure 1c). Some of the jumps in 65 the earthquake rate are attributable to local M > 3.3 aftershocks that act locally as mainshocks 66 of small aftershock sequences (marked by empty circles on Figure 1c), whereas others, but most 67 notably the one marked by an arrow, can neither be associated with local mainshocks, nor with 68 regional ones. That notable increase is mostly attributable to a seismic swarm located to the 69 south of Nablus. Given the timing relative to the Kahramanmaras earthquakes, the pervasive 70 lack of instantaneous triggering, and the lack of notable local mainshocks, it is likely that addi-71 tional mechanisms triggered the seismicity showing up as unexplainable steps in the cumulative 72 earthquake count versus time diagram. One possibility is the Kahramanmaras earthquake pair 73 triggered aseismic fault slip, which may be detectable geodetically. 74

3.2. Discerning instantaneous aftershocks

In search of local earthquakes within the surface and coda waves of the Kahramanmaraş mainshocks, we scanned all the ISN recorded strong-motion and broad-band seismograms of the $M_w7.8$ and $M_w7.6$ earthquakes. By band-pass filtering the seismograms between 8 and 16 Hz, we identified 3 microearthquakes that occurred during passage of surface waves excited by the $M_w7.6$ (Figure 2a). These are best observed in the strong-motion seismograms recorded by the MSLM and DSI stations, and to a lesser extent also by TMR, OVNT, EFSH, and KEGD. The P- and S-phase arrivals of the earliest aftershock to MSLM are well resolved (Figure 2b, 2c, and S2), and constrain its hypocentral distance to about 16 kilometers. Low-pass filtered ground velocity (integrated accelerations) time-series at station MSLM are shown in Figure 2a. Inspection of these, as well as other, low-pass filtered accelerograms indicate that the strongest ground motion in the DS area is measured on the transverse component, which is associated with Love wave propagation. A detailed analysis of the surface-wave induced strain in the northern DS area is presented in Section 6.1.

3.3. Supplementing the Dead Sea aftershock catalog using Match Filter Analysis

To identify multiplets of event #1 (registered or not in the GSI catalog), we implemented 88 the template-based Match Filter Analysis (MFA; *Gibbons and Ringdal* [2006]). We extracted 89 a three-component template waveform, consisting of a portion of the P-wave train excited by 90 event #1. The template duration is 1.8 s, starting from 0.3 s prior to the P-wave arrival at 91 the four nearest stations (in Figure 1b), and filtered between 8 to 16 Hz. The MFA detection 92 process operates as follows: First, we correlate the template waveform across a 2.13 year-long 93 time-series (segmented into days) starting from January 1, 2021, and ending on February 21, 94 2023, with 90% overlap between adjacent temporal windows. Next, we correlate the correlations 95 of the vertical channels carrying the template across stations, thus deriving the inter-station time 96 offsets corresponding to the template. Finally, we shift the template cross-time three-component 97 correlations according to the cross-station offsets, and stack the cross-correlations. These were 98 scanned for multiplets. For a positive hit, we require the value of this function to exceed its 99 1-hour median by 9 times the Median Absolute Deviation of that hour. We turn each successful 100 hit into a template, and repeat the same procedure as for event #1. 101

¹⁰² Analysis of the 9 hour interval between the $M_w7.8$ and $M_w7.6$ mainshocks did not yield any ¹⁰³ positive detections, and the analysis of the 17 days following the $M_w7.6$ earthquake yielded 3 hits.



Figure 2: Ground acceleration and velocity recorded at stations MSLM and DSI showing the surface wave arrivals, the instantaneously triggered earthquake and the first two aftershocks. V, T, and R are for the vertical, transverse, and radial component of motion, respectively. The numbers refer to the events listed in Table 1. (a) Band- and low-pass filtered seismograms from MSLM. Top trace is band-pass filtered between 8 to 16 Hz, and the three bottom traces are low-pass filtered above 0.1 Hz. Time is referenced to the $M_w7.6$ origin time. Arrows show arrivals due to events #1 to #3. (b,c) Vertical accelerograms containing the P-wave train of the 4 earthquakes of the northern DS basin repeating aftershock. The traces are filtered between 8 to 16 Hz and aligned on the P-wave arrival. The vertical bars are for scale. Event #1 was instantaneously triggered during the passage of surface waves excited by the February 6, 2023 $M_w7.6$ Kahramanmaraş earthquake. Left panel: Station MSLM. Right panel: Station DSI.

We did not detect similar events occurring in the two years before the M_w 7.8. The newly detected 104 earthquake source parameters are listed in Table 1 (Supplementary Materials), and their P-wave 105 seismograms are shown in Figure 2. They exhibit great similarity to the instantaneously triggered 106 earthquake template (event #1), especially at station MSLM, located a few km from event #1107 epicenter (Figure 2). Between February 6 to February 17, 2023, the ISN located 6 earthquakes 108 occurring within 15 km of event #1, one of them passed our detection criteria (event #4, see 109 Table 1). Based on waveform similarity, we suggest that the 4 aftershocks re-ruptured an isolated 110 fault patch. The MFA further indicates that this patch was previously inactive for at least two 111 years. 112

4. Master event location

The four hypocenters were located in two steps. In the first, we implemented a grid-search 113 approach, and used the 1-D velocity model of *Gitterman et al.* [2003]. In the second step, we used 114 the master event location approach to get relative locations [*Fitch*, 1975]. We chose the latest 115 earthquake (event #4) as a master event, as its initial location was most tightly constrained. 116 High waveform similarity allowed us to improved the P-phase picking of the 3 slave events by 117 cross-correlation master event P-wave arrivals. Furthermore, we fixed the depths of the slave 118 events, to that of the master. The result of the master-slave event locations (Figure 1b) indicates 119 the four ruptures overlap, supporting the view of repeating aftershocks. 120

5. Temporal variations

5.1. Recurrence times

The spatial clustering of the four earthquakes, and their strong waveform similarity are consistent with the possibility that this series resulted from repeated slip events on an isolated

stick-slip patch that is embedded within an otherwise aseismic fault [Vidale et al., 1994; Nadeau 123 et al., 1994; Marone et al., 1995]. The effect of perturbing the stress of such a configuration is 124 to increase the creep rate on the aseismic segment. If the slip on that segment is resisted by a 125 logarithmic rate- strengthening friction, the rate of creep is expected to decay as 1/t following 126 a stress step, with t being the elapsed time since the stress perturbation [Schaff et al., 1998; 127 Chen and Lapusta, 2009. Consequently, the stressing rate on the stick-slip patch varies with the 128 reciprocal of time, and the following relation between the recurrence time, τ , and t is predicted 129 [Schaff et al., 1998]:130

131

 $\frac{1}{\tau_i} = \frac{a}{t_i},\tag{1}$

where a is an empirically determined fitting coefficient, and i is the earthquake index. The right-132 hand side of (1) is similar to that of the well-known Omori law [Omori, 1894], which provides 133 good fit to the vast majority of aftershock sequences observed following major earthquakes. In 134 contrast to the Omori law, this function is equated here to the reciprocal of the recurrence time, 135 which may be thought of as the instantaneous rate of earthquake occurrence. If instead, the creep 136 is resisted by viscous stresses (with friction being linearly proportional to slip rate), the rate of 137 creep is expected to decay exponentially [Segall, 2010], and so is the instantaneous earthquake 138 rate: 139

$$\frac{1}{\tau_i} = a \exp(-bt),\tag{2}$$

with *b* being another fitting coefficient. Relations (1) and (2), hereafter refer to as the Omori and Exponential models, respectively, provide means for corroborating the isolated stick-slip patch model, and discriminating between the two friction models. The reciprocal of the repeating quakes recurrence time as a function of time since the first aftershock is shown in Figures 3a on a log-log space. Because the number of data is small, it is not evident which relation best fits the recurrence intervals. To quantify the degree to which the temporal evolution of this earthquake
 cluster is more consistent with relation (1) than (2), an Omori-versus-Exponential Fitting Index
 (FI) is introduced:

$$FI = \frac{L_{\rm Exp}^2 - L_{\rm Omori}^2}{L_{\rm Exp}^2 + L_{\rm Omori}^2},\tag{3}$$

where L^2_{Omori} and L^2_{Exp} are the norm-2 misfits to equations 1 and 2, respectively. This function 150 may range between -1 and +1, with FI=-1 if the $1/\tau$ time-series is perfectly modeled by the 151 exponential relation in (2), and FI=1 if it is perfectly modeled by the Omori function in (1). 152 The fitting index of the time-series in question is equal to 0.8, thus strongly suggesting that the 153 data are better fit with (1) than with (2). Yet, given the small dataset, it is important to assess 154 the extent to which this result is statistically meaningful. In the supplementary materials we 155 present the result of a Monte Carlo testing, showing that the possibility of FI=0.8 emerging by 156 chance may be rejected at greater than 96% confidence level. 157

5.2. Relative seismic moments

¹⁵⁸ We assume the earthquake spectra follow the well-known ω^2 model [*Aki*, 1967; *Brune*, 1970], ¹⁵⁹ and recover the source properties directly from the observed spectral amplitudes. The low-¹⁶⁰ frequency displacement spectral plateau Ω_0 and the seismic moment are related as [*Aki and* ¹⁶¹ *Richards*, 2002]:

162

$$M_0 = \Omega_0 RC,\tag{4}$$

where R is the hypocentral distance and collects site and path parameters and the radiation pattern. Note that the seismic moment estimates are insensitive to the decay rate at frequencies well-above the source' characteristic frequency. The low-frequency spectral plateau, Ω_0 , may be computed in the frequency or time domain. Here, Ω_0 is estimated using the simple time-domain



Figure 3: Summary of temporal analysis. (a) Log-log diagram of the reciprocal of recurrence time as a function of time since the first aftershock. (b) Relative moment as a function of recurrence time for events 2-4, calculated separately for accelerograms recorded at stations MSLM (gray circles) and DSI (black squares). Red symbols correspond to the N + 1 predictions.

 $_{167}$ expression of *Lior and Ziv* [2020]:

168

$$\Omega_0 = 2\sqrt{T} D_{rms}^{1.5} / V_{rms}^{0.5},\tag{5}$$

where T is the data interval, and D_{rms} and V_{rms} are the displacement and velocity root-meansquares (rms), respectively. The above expression is valid in the far-field, and does not require prior knowledge of the source corner frequency and attenuation effects. For these reasons, use of this time-domain expression is most advantageous for small earthquakes and/or data of limited frequency band-width [*Lior and Ziv*, 2020]. Because repeating earthquakes recorded at a given station share the same site and path parameters, the seismic moment of earthquake j relative to that of earthquake k, calculated using seismograms recorded by station i may expressed as:

$$\frac{M_0^{ij}}{M_0^{ik}} = \frac{(D_{rms}^{1.5}/V_{rms}^{0.5})_{ij}}{(D_{rms}^{1.5}/V_{rms}^{0.5})_{ik}}.$$
(6)

We calculated relative seismic moments from doubly-integrated accelerations at MSLM and 177 DSI using 4 seconds-long intervals starting at the S-phase onset. Because events 1 to 3 coincide 178 with the passage of the surface waves of the M_w 7.6 Kahramanmaraş earthquake, they may only be 179 studied in a limited frequency band. Therefore, a 6-18 Hz bandpass filter was applied following 180 each integration. The relative seismic moments exhibit a 20% steady increase of the relative 181 moment per 10-fold increase time elapsed since event 4. (Figure 3b). This trend is consistent 182 with experiments showing that friction increases with the logarithm of time since the previous 183 slip episode [Dieterich, 1972; Scholz et al., 1972; Scholz and Engelder, 1976]. 184

6. Discussion

The DS repeating aftershock sequence reported in this study differs from previously reported ones in two main respects. First, while previous repeaters were located in the near-field of the mainshock the DS sequence is clearly located at its far-field. Second, the triggering mainshock occurred just 9 hours after a larger mainshock that did not trigger instantaneous aftershocks in the study area or delayed aftershocks near the DS sequence. These differences create an opportunity for revisiting previous paradigms, and introducing new ones.

6.1. Triggering mechanism for delayed remote aftershocks

It is difficult to explain the triggering of remote delayed aftershocks by means of stress transfer because of the small permanent stress changes. In contrast, explaining the occurrence of instantaneous aftershocks is more straightforward because their magnitude seems sufficient for triggering of critically-stressed faults [*Gomberg and Johnson*, 2005]. Thus, if instantaneous triggering was a common phenomena, one could have imagined a stress transfer process, whereby delayed aftershocks are triggered by the instantaneous aftershocks that act locally as small main¹⁹⁷ shocks [Helmstetter et al., 2003; Ziv, 2006a; Heimisson, 2019]. Because, however, our search for
¹⁹⁸ instantaneous triggering yielded only 3 co-located aftershocks within the northern DS basin, this
¹⁹⁹ triggering mechanism cannot explain the widespread delayed aftershocks reported in this study
²⁰⁰ (Figure 1a). Here we suggest that the observations presented in this study shed new light on the
²⁰¹ triggering mechanism of remote aftershocks.

It is a widely held notion that the re-rupturing of a given fault patch, with recurrence intervals 202 that decay as the reciprocal of elapsed time since the first aftershock, requires strong interaction 203 with a velocity-strengthening creeping segment [Schaff et al., 1998]. As this sequence occurred 204 in the far-field of its trigger, and in an area that did not experience any earthquake rate increase 205 prior to its onset, we can further infer that remote delayed aftershocks result from the stress 206 change induced by local creep. That creep triggered by seismic waves excited by the mainshock. 207 Once in motion, they stress their surroundings, and trigger instantaneous or delayed aftershocks. 208 Indeed, the sequence documented in this study consists of four aftershocks, of which the first 209 three are instantaneous, and the fourth is delayed. The interaction between aseismic and seismic 210 slip is rather complex. The stick-slip patch is being stressed by the surrounding creep during 211 inter-event intervals, and is stressing it with each slip episode. Thus, deeper understanding of 212 this process may require forward modeling. 213

6.2. Implications of the non-triggering by the $M_w 7.8$ earthquake

That the repeating aftershock sequence was triggered instantaneously by a $M_w7.6$ earthquake just 9 hours after an equally distant $M_w7.8$ earthquake raises the question as to what may be learned from the non-triggering of that sequence by the largest of the two. The non-triggering by the $M_w7.8$ and the triggering by the $M_w7.6$ earthquakes just 9 hours later may be understood if the transient stress change induced by the latter was notably larger than that induced by the



Figure 4: Ground velocity and its spatial gradient as a function of time. \dot{u}_t denotes the transverse component of velocity. $\dot{u}_{t,r}$, $\dot{u}_{r,r}$, and $\dot{u}_{v,r}$ are for gradients of the transverse, radial, and vertical velocity components computed along the radial direction, respectively. Number in percentage denotes the difference between the maximum $M_w7.6$ and $M_w7.8$ gradient maxima relative to the maximum amplitude of the $M_w7.8$ for each component (i.e. positive is for increase in the $M_w7.6$ maxima gradient amplitude). a. Station DSI low-pass filtered above 1 Hz. b-d. The velocity gradient computed using differences between DSI and MSLM (see Figure 1 for station locations). Seismograms were band-pass filtered between 32 to 16 s before differentiation. Units are in micro-strain/s.

former. Indeed the Peak-Ground-Velocity (PGV) induced by the $M_w7.6$ in the DS area is by about 10% larger than that induced by the $M_w7.8$. Equating the peak strain to the product between the PGV and the slowness [e.g *Benioff*, 1935; *Mikumo and Aki*, 1964] yields an equally larger strain - and therefore also stress change - for the $M_w7.6$ than for the $M_w7.8$. It is difficult to attribute the notably different triggering effect of the two earthquakes to a 10% difference in the stress perturbations at the same site.

Gomberg and Aqnew [1996] reported large discrepancies between strain calculation using the 225 above approach and direct strain-meter measurements. Thus, to address the question as to 226 what may be learned from the non-triggering by the M_w 7.8, it is instructive to obtain a more 227 direct strain measurement. Since strain-meters are not available in the study area, we implement 228 an alternative approach that takes advantage of the network's alignment being sub-parallel to 229 the radial direction pointing from the northern DS basin to the M_w 7.6 source. By taking the 230 difference between the same-component long-period (>10 s) velocity time-series measurements 231 at two nearly radially-aligned stations, and dividing them by the inter-station distance projected 232 onto the radial direction, we obtained timelines of the following velocity gradients: $\dot{u}_{r,r}$, $\dot{u}_{t,r}$, and 233 $\dot{u}_{v,r}$, where t, r, and v represent the transverse, radial, and vertical components, respectively. 234 In doing so, we require that the inter-station distances be less than about a quarter of the 235 seismic wave-length (after filtering). The main advantage of this approach with respect to the 236 one described above is that it does not rest on the assumption that the slowness between each 237 station pair equals the slowness measured across the network. Under the assumption that the 238 waves can be treated as perfectly planar, the strain rate tensor components can be approximated 239 from the measured gradients [Gomberg and Agnew, 1996]. In this case $\dot{\varepsilon}_{rr} \approx \dot{u}_{r,r}$, $\dot{\varepsilon}_{tr} \approx 1/2\dot{u}_{t,r}$, 240 and $\dot{\varepsilon}_{vr} \approx 1/2\dot{u}_{v,r}$, where $\dot{\varepsilon}_{rr}$ and $\dot{\varepsilon}_{tr}$ are the horizontal dilatational and shear strains, respectively, 241 and $\dot{\varepsilon}_{vr}$ is the shear strain acting on vertical planes parallel to radial direction. 242

The results of our new approach for deformation rate analysis are presented in Figure 4 for the two nearest stations to the DS aftershock sequence - MSLM and DSI. We find that the differences between $M_w7.8$ and the $M_w7.6$ are more notable in the deformation rate domain than they are in the raw velocity time-series. Comparing the $M_w7.8$ and the $M_w7.6$ deformation rate timeseries, we find that the most notable positive gradient increases are observed on the transverse

and radial components, whereas the peak M_w 7.6 vertical component gradient is reduced relative 248 to the M_w 7.8. The peak dynamic long-period (>10 s) transverse velocity gradient due to the 249 $M_w 7.6$ is about 30% larger than the $M_w 7.8$ peak gradient, however individual phases that differ 250 by up to a factor of two appear after differentiation (Figure 4b). We examined the deformation 251 rate gradients using all neighboring station pairs along the western shore of the DS lake (Figure 252 1a). This indicated that M_w 7.6 transient shear strain is increased by 30% to 40% relative to the 253 $M_w 7.8$ at all the northern DS basin station pairs, that dilatational strain peak for MSLM-DSI 254 increased by 30% but was reduced at the adjacent station pairs, and that the $M_w7.6$ transient 255 shear stress peaks halfway between MSLM and DSI, in agreement with the repeater location. 256

Two source-related effects may combine to produce 10% stronger ground shaking and 30% 257 to 40% strain rate for the M_w 7.6 than for the M_w 7.8. The first factor, affecting mainly the 258 long-period strain rates, is the azimuthal dependence of the Love wave radiation pattern. The 259 M_w 7.6 fault strikes approximately east-west, and is thus well oriented with respect to the Love-260 wave radiation towards the DS, while the $M_w 7.8$ strikes approximately north-north-east, thus 261 producing minimum Love-wave radiation in that direction. This eastward strike rotation is 262 sufficient in order to offset the level of seismic energy radiated from the $M_w7.6$ source to an 263 amplitude comparable to the one radiated by the $M_w 7.8$ source. A second factor, affecting 264 mostly the short-period ground shaking between the two mainshocks is their static stress drops. 265 The $M_w 7.8$ ruptured a relatively large fault area of about 1400 km² that slipped on average by 266 about 5 m, whereas the M_w 7.6 ruptured about half the M_w 7.8 fault area, and its average slip 267 is about 2.5 m [Barbot et al., 2023]. Since the stress drop is proportional to the ratio between 268 the average slip and the fault length, those differences decreased the $M_w 7.8$ stress drop by a 269 factor of about 3 relative to the M_w 7.6 stress drop. We suggest that the larger stress drop and 270

7. Conclusive remarks

The 2023 Kahramanmaraş earthquake pair triggered widespread aftershock activity throughout Palestine and Israel. Surprisingly, the strongest earthquake rate increase is observed on either side of the DST and in the Mediterranean Sea offshore Haifa, suggesting that secondary structures in the study area are presently more prestressed than the DST main strand. Earthquake count as a function of time reveals complex evolution, with activity bursts that are attributable to $M_w > 3$, which act as local mainshocks. Other seismicity rate jumps that cannot be explained by local triggering, may be indicative of aseismic deformations.

Visual inspection of high-pass filtered accelerograms of the M_w 7.8 and M_w 7.6 Kahramanmaraş earthquakes discloses three instantaneously triggered aftershocks of the latter in the northern DS basin. A delayed aftershock is revealed via MFA. Based on strong waveform similarity and co-location, with location uncertainties that are smaller than the estimated rupture dimensions, it is inferred that the four aftershocks correspond to a sequence of repeaters. Temporal analyses are presented, which indicate that the logarithm of repeat time and the seismic moment increase linearly with the logarithm of elapsed time since the first aftershock.

This first account of an instantaneously triggered repeating aftershock sequence at the farfield of its trigger underscores the role of the interaction between aseismic and seismic slip in remote aftershock triggering [*Inbal et al.*, 2017]. A comparison between peak ground shaking and spatial derivative of velocity seismograms of the two Kahramanmaraş earthquakes suggests larger transient stress change in the DS during the $M_w7.6$ than during the $M_w7.8$. This may explain why the DS aftershocks were triggered by the $M_w7.6$, but not the $M_w7.8$, despite the latter preceding the former by just a few hours.

Finally, given a sequence of $N \geq 3$ after shocks, with a τ versus t relation obeying Equation 294 1, the timing of the N + 1 aftershock may be predicted using: $t_{i+1} = t_i \frac{a}{a-1}$. Thus, should the 295 current $1/\tau$ trend continue, the next aftershock is expected to occur about 278 days after the 296 mainshock (using a=1.04 and $t_4=10.72$ days, see red open circle in Figure 3a). Furthermore, 297 from the relative moment versus time plot in Figure 3b, the size of that earthquake is expected 298 to be 20% larger than that of the largest aftershock. These predictions may require some revision 299 in case that the local stress field would be altered by some external source, such as a local or 300 regional earthquake, or a slow slip episode on a nearby segment. 301

302 Acknowledgments.

Open Research. Data from the IS network are available at https://doi.org/10.5281/zenodo.7980346. The seismic catalog is available at https://seis.gsi.gov.il/fdsnws/event/.

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