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# Microseismic Constraints on the Mechanical State of the North Anatolian Fault Zone 13 Years After the 1999 M7.4 Izmit Earthquake

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1           **Microseismic Constraints on the Mechanical State of the**  
2           **North Anatolian Fault Zone Thirteen Years after the 1999**  
3           **M7.4 Izmit Earthquake**

4                           **Eric Beauce<sup>1,3</sup>, Robert D. van der Hilst<sup>1</sup>, Michel Campillo<sup>2,1</sup>**

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8           **Key Points:**

- 9           • We build an extensive earthquake catalog using our new fully automated method  
10           along the western North Anatolian Fault Zone (NAFZ).  
11           • We observe variable statistical properties (b-value, and temporal clustering) along  
12           the NAFZ.  
13           • High b-value and strong temporal clustering may suggest the occurrence of slow  
14           slip in the Lake Sapanca step-over.

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## Abstract

The 17 August 1999  $M_w$  7.4 Izmit earthquake ruptured the western section of the North Anatolian Fault Zone (NAFZ) and strongly altered the fault zone properties and stress field. Consequences of the co- and post-seismic stress changes were seen in the spatio-temporal evolution of the seismicity and in the surface slip rates. Thirteen years after the Izmit earthquake, in 2012, the seismic network Dense Array for North Anatolia (DANA) was deployed for 1.5 years. We built a new catalog of microseismicity ( $M < 2$ ) by applying our automated detection and location method to the DANA data set. Our method combines a systematic backprojection of the seismic wavefield and template matching. We analyzed the statistical properties of the catalog by computing the Gutenberg-Richter b-value and by quantifying the amount of temporal clustering in groups of nearby earthquakes. We found that the microseismicity mainly occurs off the main fault and that the most active regions are the Lake Sapanca step-over and near the Akyazi fault. Based on previous studies, we interpreted the b-values and temporal clustering *i*) as indicating that the Akyazi seismicity is occurring in high background stresses and is driven by the Izmit earthquake residual stresses, and *ii*) as suggesting evidence that an intricate combination of seismic and aseismic slip was taking place on heterogeneous faults at the eastern Lake Sapanca, near the brittle-ductile transition. Combined with geodetic evidence for enhanced north-south extension around Lake Sapanca following the Izmit earthquake, the seismicity supports the possibility of slow slip at depth in the step-over.

## Plain Language Summary

On 17 August 1999, a large M7.4 earthquake struck near the city of Izmit, in western Turkey, and caused important human and material losses. The earthquake resulted from the large and sudden displacement of crustal blocks along the North Anatolian Fault Zone (NAFZ). Transient changes in the crustal and fault properties are commonly observed following such large events. In this study, we analyze the statistical properties of microearthquakes, that is, of small earthquakes ( $M < 2$ ) typically too small to affect the surrounding population, to gain knowledge about the state of the NAFZ more than a decade after the Izmit earthquake. First, we address the challenge of locating microearthquakes, in space and time, by applying our automatic earthquake detection and location algorithm. Then, we use a statistical analysis to characterize physical properties of the NAFZ and, thus, to highlight the peculiar properties of faults near Lake Sapanca. We interpret that these faults are heterogeneous and slip both seismically and aseismically. Our study calls for taking a closer look at the extension across Lake Sapanca with different, complementary geophysical methods.

## 1 Introduction

The North Anatolian Fault Zone (NAFZ) is a 1,500 km long strike-slip fault that bounds the Anatolian plate to the south and the Eurasian plate to the north (Figure 1A). The fault slips, overall, in a right-lateral manner to accommodate the westward motion of Anatolia with respect to Eurasia due to the combination of the subduction along the Hellenic trench and the Cyprus trench in the southwest and the south and the collision with Arabia in the southeast (Le Pichon & Angelier, 1979; McClusky et al., 2000; Reilinger et al., 2006). Near the Gulf of Izmit, in western Turkey, the NAFZ splits into a northern strand and a southern strand. These two strands bound the Almacik mountains in the east and the Armutlu block in the west, and separate the Istanbul Zone in the north from the Sakarya Terrane in the south, which are the remains of the passive margin of the Intra-Pontide Ocean (see Figure 1B, e.g. Akbayram et al., 2013). Most of the deformation is accommodated on the northern strand (e.g. Meade et al., 2002; Reilinger et al., 2006).

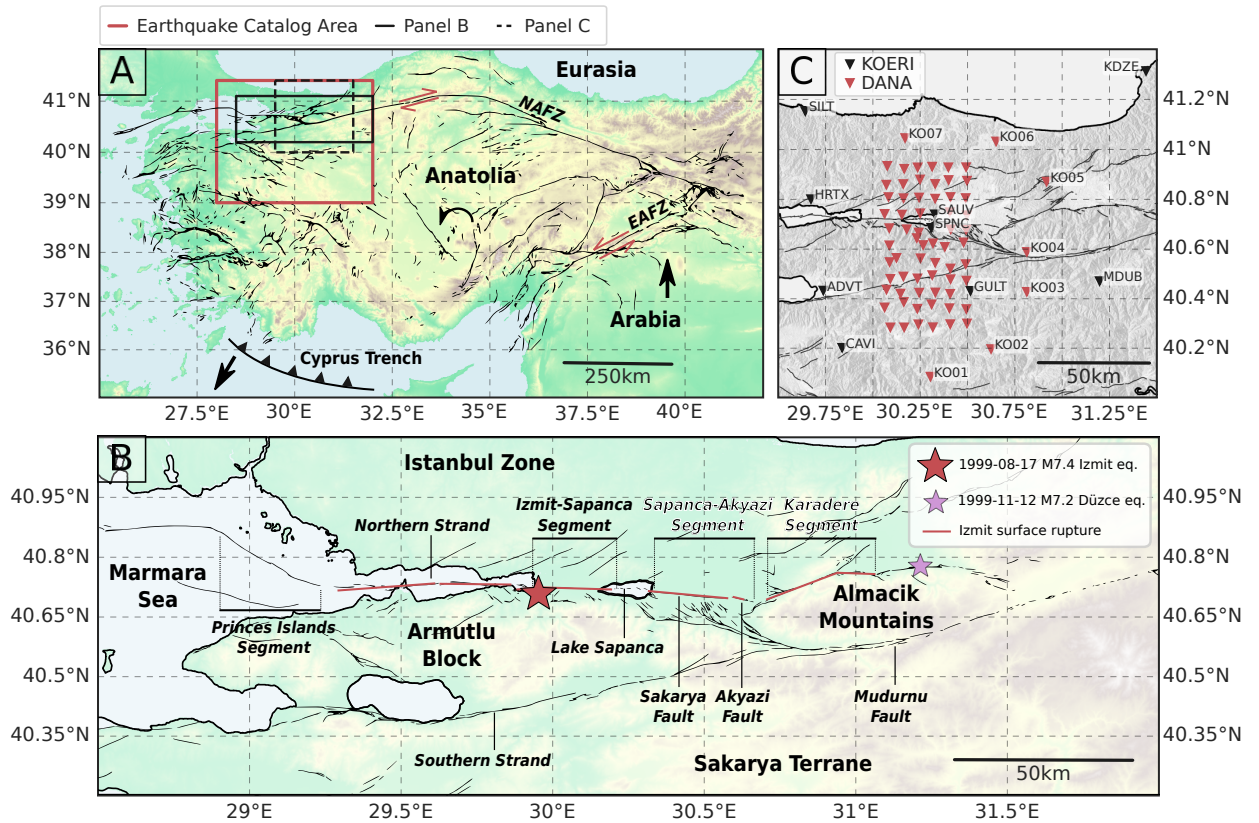
The 17 August 1999 M7.4 Izmit earthquake and the 12 November 1999 Düzce M7.2 earthquake are the most recent (as of the time of writing) events of a series of westward

65 migrating  $M > 7$  earthquakes that ruptured almost entirely the North Anatolian Fault Zone  
 66 (e.g. Toksöz et al., 1979; Stein et al., 1997). The Izmit earthquake nucleated near the Izmit  
 67 Bay, propagated bilaterally and broke a 150 km-long, almost vertical section of the fault  
 68 made of four, or five, segments along the northern strand (Toksoz et al., 1999; Barka et  
 69 al., 2002). To the east, the rupture propagated at super-shear speeds (Bouchon et al., 2001,  
 70 2011) and broke the Izmit-Sapanca, the Sapanca-Akyazi and the Karadere segments (cf.  
 71 names on Figure 1B). To the west, the rupture propagated along the Gölcük segment and  
 72 stopped on the Yalova segment (Langridge et al., 2002), increasing the probability of ma-  
 73 jor failure further west beneath the Marmara Sea (Parsons et al., 2000). The Düzce earth-  
 74 quake nucleated near the eastern termination of the Izmit earthquake, likely due to in-  
 75 creased Coulomb stress (Parsons et al., 2000; Utkucu et al., 2003). The co- and post-seismic  
 76 stress changes and the transient changes of the fault's mechanical properties caused by  
 77 the Izmit earthquake affected the local seismicity patterns and the focal mechanisms of  
 78 microearthquakes (e.g. Bohnhoff et al., 2006; Pınar et al., 2010; Ickrath et al., 2015). GPS  
 79 and interferometric synthetic aperture radar (InSAR) observations suggest that fast and  
 80 rapidly decaying afterslip occurred in the middle-to-lower crust in the months following  
 81 the Izmit-Düzce earthquake sequence (e.g. Reilinger et al., 2000; Bürgmann et al., 2002),  
 82 then relayed by slower post-seismic slip at depth (Ergintav et al., 2009; Hearn et al., 2009).  
 83 Patterns of surface displacement also suggest the existence of shallow creep along the  
 84 Izmit-Sapanca and the Sapanca-Akyazi segments (e.g. Çakir et al., 2012; Hussain et al.,  
 85 2016). Transient creep episodes have been identified more than a decade after the Izmit  
 86 earthquake (Aslan et al., 2019).

87 Despite the overall good understanding of the east-west motion along the west-  
 88 ern NAFZ, smaller scale, north-south extension at some locations remains enigmatic. Co-  
 89 and post-seismic slip on vertical fault segments seems unable to reproduce the patterns  
 90 of north-south extension observed in geodetic data (e.g. Ergintav et al., 2009; Hearn et  
 91 al., 2009). Even though refining the geometry of the main fault segments of the NAFZ  
 92 helps explain the observations (e.g. slightly north dipping faults, Cakir et al., 2003), mod-  
 93 els of the post- and inter-seismic deformation along the NAFZ would benefit from tak-  
 94 ing into account secondary structures, such as the faults in step-overs. Microseismicity  
 95 ( $M < 2$ ) provides information at small length scales at seismogenic depths and thus is com-  
 96 plimentary to geodetic data in building a better understanding of slip along the NAFZ (aseis-  
 97 mic vs seismic, distributed vs localized), that is, of its mechanical state.

98 The abundance of microearthquakes makes them well-suited for statistical analy-  
 99 ses. Of interest here are the  $b$ -value of the Gutenberg-Richter law (Gutenberg & Richter,  
 100 1941) that describes the frequency-magnitude distribution of a population of earthquakes,  
 101 and the fractal dimension  $D$  of the earthquake occurrence time series (Smalley Jr et al.,  
 102 1987; Beaucé et al., 2019) that quantifies the strength of temporal clustering. The  $b$ -value  
 103 acts as a stressmeter (Amelung & King, 1997; C. H. Scholz, 2015), and the fractal dimen-  
 104 sion  $D$  is related to the density of fractures and seismic-aseismic slip partitioning (C. Scholz,  
 105 1968; Dublanchet et al., 2013).

106 The dense seismic array DANA (Dense Array for North Anatolia DANA, 2012, see  
 107 Figure 1C) was deployed around the rupture trace of the 1999-08-17  $M 7.4$  Izmit earth-  
 108 quake, it operated from early May 2012 to late September 2013. These data enabled mul-  
 109 tiple studies that improved our understanding of the complex structures and seismicity  
 110 patterns in the region (e.g. Poyraz et al., 2015; Kahraman et al., 2015; Papaleo et al., 2018;  
 111 Taylor et al., 2019). Here, we study microearthquakes in order to improve our understand-  
 112 ing of the mechanical state of the North Anatolian Fault Zone more than a decade after  
 113 the Izmit earthquake. First, we briefly describe our automated earthquake detection and  
 114 location method (Section 3), and then present the earthquake catalog (Section 4.1) and  
 115 a statistical analysis of collective properties of earthquakes ( $b$ -value, Section 4.2, and tem-  
 116 poral clustering, Section 4.3). These observations allow a characterization of the phys-  
 117 ical environment in which seismicity takes place. We interpret and discuss our results to  
 118 question the role of secondary structures in the dynamics of NAFZ (Section 5).



**Figure 1.** **A:** Large scale view of the North Anatolian Fault Zone. Abbreviations: NAFZ - North Anatolian Fault Zone, EAFZ - East Anatolian Fault Zone. The red arrows indicate the direction of coseismic motion. Our study region is located at the western end of the NAFZ. **B:** Magnified view of the fault zone in our study region. Larger font names are the main geologic units: Istanbul Zone, Armutlu Block, Almacik Mountains and Sakarya Terrane. The smaller font, italic names are segments and faults of the NAFZ: the Izmit-Sapanca segment, the Sapanca lake step-over, the Sapanca-Akyazi segment (which together constitute the northern strand), the Karadere segment and the southern strand (names following Barka et al., 2002). The Sapanca-Akyazi segment is made of the Sakarya fault and the Akyazi fault. The flat area around the Akyazi fault is referred to as the Akyazi plain. Both Lake Sapanca and the Akyazi plain are pull-apart basins. The large red star indicates the epicenter of the  $M_w$  7.4 Izmit earthquake, and the small purple star indicates the epicenter of the  $M_w$  7.2 Düzce earthquake. **C:** The seismic stations used in this study are from the temporary experiment DANA (70 stations, red triangles; DANA, 2012) and the permanent network KOERI (9 stations, black triangles; Kandilli Observatory And Earthquake Research Institute, Boğaziçi University, 1971). Each column of the DANA array is indexed by a letter and each row is indexed by a number (DA01, DA02, ..., DB01, ...).

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## 2 Data

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The continuous seismic data were recorded by broadband stations from the temporary array DANA (70 stations) and the permanent network KOERI (9 stations, see the locations in Figure 1, and the Data and Resources section). The time period covered by this study is set by the duration of the DANA experiment: 2012-05-04 to 2013-09-20. Sampling rates are 50 Hz for all stations but SAUV, which samples at 100 Hz. We band-pass filtered the data between 2 Hz and 12 Hz to eliminate low frequency noise and to

126 allow us to downsample the time series to 25 Hz in order to make the computation less  
127 intensive.

### 128 3 Methodology

#### 129 3.1 Earthquake Detection, Location, and Magnitude Estimation

130 We analyzed the data for the entire time period with a fully automatized earthquake  
131 detection and location method. The core of the workflow, summarized in Figure 2, con-  
132 sists of three stages:

- 133 1. Backprojection (Section 3.1.1): The energy of the seismic wavefield is continuously  
134 backprojected onto a 3D grid of potential sources to detect coherent (earthquake)  
135 sources.
- 136 2. Relocation (Section 3.1.2): The P- and S-wave first arrivals of the previously de-  
137 tected events are identified with the automatic phase picker PhaseNet (Zhu & Beroza,  
138 2019), and the picks are used in the NonLinLoc earthquake location software (Lomax  
139 et al., 2000, 2009).
- 140 3. Template matching (Section 3.1.3): The successfully relocated earthquakes are used  
141 as template earthquakes in a matched-filter search to detect other, smaller earth-  
142 quakes in the same region using the Fast Matched Filter software (Beaucé et al.,  
143 2018).

144 The detection method is discussed in detail in Beaucé et al. (2019), but the relocation is  
145 now fully automated and includes PhaseNet and NonLinLoc. In an extra step, we further  
146 characterized the detected earthquakes by relocating them with the double-difference  
147 method (Section 3.1.4) and estimating their magnitude (Section 3.1.5).

##### 148 3.1.1 Backprojection and Location

149 We backprojected the energy of the seismic wavefield recorded at the array of seis-  
150 mic stations onto a 3D grid of potential sources beneath the study region, searching for  
151 the space-time locations of coherent sources. We computed the composite network re-  
152 sponse (CNR, Frank & Shapiro, 2014):

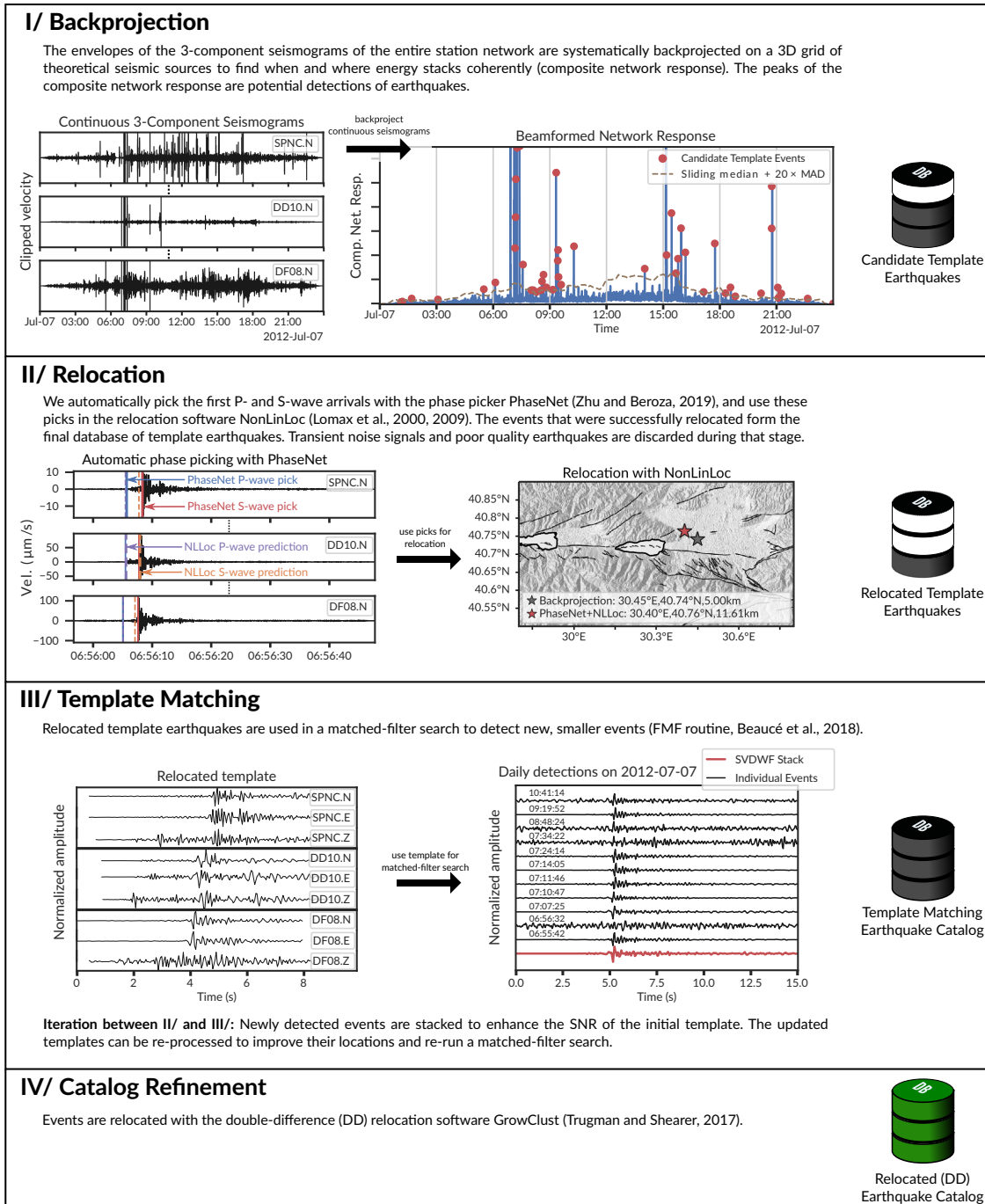
$$153 \text{CNR}(t) = \max_k \{ \text{NR}_k(t) \}; \quad \text{NR}_k(t) = \sum_{s,c} \text{env} \left( u_{s,c}(t + \tau_{s,c}^{(k)}) \right). \quad (1)$$

154 In this equation,  $t$  is the detection time and  $\text{NR}_k(t)$  is the network response for source  
155 location indexed by  $k$  at time  $t$ .  $\text{NR}_k(t)$  is the sum of the envelopes (the modulus of the  
156 analytical signal) of the seismograms  $u_{s,c}$  shifted in time by the moveout  $\tau_{s,c}^{(k)}$  on station  
157  $s$  and component  $c$ . The moveouts were computed using the ray-tracing software Pykonal  
158 (White et al., 2020) in the 1D velocity model from Karabulut et al. (2011) (see Table S1).  
159 We note that the use of a 1D velocity model in this region can introduce significant er-  
160 rors in the earthquake locations because of the strong lateral velocity variations, in par-  
161 ticular across the two strands of the NAFZ (e.g. Karahan et al., 2001; Kahraman et al., 2015;  
162 Papaleo et al., 2018). This velocity model produced a visually satisfying agreement be-  
163 tween earthquake epicenters and fault surface traces, and allowed consistency with a pre-  
164 vious study on the same data set (Poyraz et al., 2015). The backprojection method nat-  
165 urally provides an estimate of the location of each detected events. However, the net-  
166 work response finds the times that aligned the envelope maxima rather than the P- and  
167 S-wave arrivals, which results in approximate locations.

##### 168 3.1.2 Relocation

169 All the events detected through the CNR were processed with the deep neural net-  
170 work PhaseNet (Zhu & Beroza, 2019) to automatically pick the P- and S-wave first arrivals.

## Earthquake Detection and Location Workflow



**Figure 2.** Summary flowchart of the earthquake detection and location method. For clarity, only a subset of stations are shown in the above panels, but all the analysis is carried on the 79 stations together. Template matching is performed on the 10 stations closest to the source and the detection threshold is set to  $8 \times \text{RMS}$  of the correlation coefficients in a 30-minute sliding window. See Data and Resources for code availability.

171 These picks were then used by the location software NonLinLoc (Lomax et al., 2000, 2009)  
 172 to get the earthquake locations and their uncertainties given as  $1-\sigma$  intervals. We required

173 at least four P- and S-wave picks and a total minimum of 15 picks to relocate an event.  
 174 Requiring both P- and S-wave picks helps constrain the earthquake depth, and impos-  
 175 ing at least 15 picks efficiently reduced the number of solutions with very large uncer-  
 176 tainties. Events that could not be successfully relocated with NonLinLoc (e.g. noisy picks,  
 177 multiple sources recorded at the same time) were discarded. More information about the  
 178 input parameters used by PhaseNet and NonLinLoc can be found in Text S1 of Support-  
 179 ing Information S1.

### 180 **3.1.3 Template Matching**

181 Successfully relocated events were kept as templates and used in a matched-filter  
 182 search to detect new, smaller magnitude earthquakes. Template matching is a powerful  
 183 method for detecting low signal-to-noise ratio (SNR) events given prior knowledge of the  
 184 target seismicity (e.g. Gibbons & Ringdal, 2006; Shelly et al., 2007; Ross et al., 2019). It  
 185 consists of searching for all earthquakes with similar waveforms and moveouts to a known  
 186 earthquake, that is, earthquakes sharing a similar location and focal mechanism. The sim-  
 187 ilarity is measured by the network-averaged correlation coefficient (CC) between the tem-  
 188 plate waveforms  $T_{s,c}$  and the seismograms  $u_{s,c}$  shifted by the template moveout  $\tau_{s,c}$ :

$$189 \quad \text{CC}(t) = \sum_{s,c} w_{s,c} \sum_{n=1}^N \frac{T_{s,c}(t_n) u_{s,c}(t + t_n + \tau_{s,c})}{\sqrt{\sum_{n=1}^N T_{s,c}^2(t_n) \sum_{n=1}^N u_{s,c}^2(t + t_n + \tau_{s,c})}}, \quad (2)$$

190 where  $w_{s,c}$  is the weight attributed to station  $s$ , component  $c$ , and  $N$  is the length of the  
 191 template waveforms. We ran the matched-filter search on multiple nodes of a super-computer  
 192 equipped with Graphic Processing Units (GPUs) using the template matching software  
 193 Fast Matched Filter (Beaucé et al., 2018). We used a template window of 8 seconds start-  
 194 ing 4 seconds before the S wave on the horizontal components and 1 second before the  
 195 P wave on the vertical components. We used a detection threshold of 8 times the root  
 196 mean square (RMS) of the CC time series in a 30-minute sliding window ( $8 \times \text{RMS}\{\text{CC}(t)\}$ ).  
 197 The 8 s template duration is adequate given the signal duration of small magnitude earth-  
 198 quakes at  $\sim 10$ -50 km source-receiver distances. The  $8 \times \text{RMS}$  threshold is in the con-  
 199 servative range of commonly used threshold in template matching studies (e.g. Shelly et  
 200 al., 2007; Ross et al., 2019). Note that  $8 \times \text{RMS}$  is about 12 times the median absolute  
 201 deviation (MAD) for a gaussian distribution.

202 After a matched-filter search over the whole study period, each template earthquake  
 203 has detected potentially many new similar earthquakes. We leveraged the similarity of  
 204 the detected events to form higher SNR waveforms with the Singular Value Decompo-  
 205 sition and Wiener Filtering method (Moreau et al., 2017), for the efficient extraction of  
 206 coherent signal and waveform denoising. The new, higher SNR template earthquakes were  
 207 in turn used to refine the locations and run another iteration of the matched-filter search.  
 208 We iterated the detection-stacking-relocation workflow only once to trade-off SNR im-  
 209 provement and loss of high frequency information due to stacking.

210 Neighboring templates often detect the same events, therefore we kept a single event  
 211 out of all detections occurring within three seconds of each other, from templates whose  
 212 uncertainty ellipsoids were separated by less than 5 km, and with average waveform sim-  
 213 ilarity greater than 0.33. These thresholds were chosen based on physical considerations  
 214 (the time threshold 3 sec assumes location errors of up to 10-15 km, the space thresh-  
 215 old 5 km accounts for coherency of waves at 2 Hz, etc) and empirically by inspecting the  
 216 output catalog for duplicated events.

### 217 **3.1.4 Double-Difference Relative Relocation**

218 We refined the earthquake locations in the region of interest, near the NAFZ and  
 219 beneath the stations, with the double-difference relative relocation method (e.g. Poupinet  
 220 et al., 1984; Waldhauser & Ellsworth, 2000). P- and S-wave differential arrival times were  
 221 computed by finding the lag times that maximize the inter-event correlation coefficients



and summing them to the travel time differences. The differential times were then processed by the relocation software GrowClust (Trugman & Shearer, 2017, additional information on parameters are given in supplementary information, Text S1.4). GrowClust estimates location uncertainties with the non-parametric bootstrap resampling method (Efron & Tibshirani, 1986).

### 3.1.5 Magnitude Estimation

Local magnitudes were computed from the amplitude ratios of peak velocities. This required estimating the magnitude of at least one event per template to calibrate our local magnitude scale. Therefore, we computed the moment magnitude  $M_w$  by fitting the Brune model (Equation (3), Brune, 1970) to the multi-station average displacement spectra that satisfied an SNR criterion (see details in supplementary Text S1.6 and Figure S1).

$$|u_{\text{Brune}}(f)| = \frac{\Omega_0}{\left(1 + \frac{f}{f_c}\right)^2}. \quad (3)$$

In Equation (3),  $\Omega_0$  is the low-frequency plateau, which is proportional to the seismic moment  $M_0$ , and  $f_c$  is the corner frequency. Additional information on how we corrected the spectra for geometrical spreading and attenuation to compute  $M_0$  from  $\Omega_0$  is given in Text S1.6. The moment magnitude  $M_w$  is:

$$M_w = \frac{2}{3} (\log M_0 - 9.1). \quad (4)$$

Once moment magnitude estimates were available for at least one event in a template family, we estimated a local magnitude  $M_L$  for all other events based on log amplitude ratios (details in Text S1.6). Finally, we measured the scaling between  $M_w$  and  $M_L$  and built the calibration first-order relationship  $M_w = A + BM_L$  (see Figure S1B).

### 3.1.6 Identifying Mining-Related Seismicity

Template matching lends itself particularly well to identifying sources of mining-related earthquakes. We identified these by analyzing the distribution of detection times within the day. Templates that detected more than 80% of events between 6am and 6pm were categorized as mining-related templates (see Figure S3), since we do not expect natural seismicity to occur within preferred times.

## 3.2 Gutenberg-Richter b-value

The frequency-magnitude distribution of earthquakes typically follows the Gutenberg-Richter law (Gutenberg & Richter, 1941):

$$\log N(M) = a - bM. \quad (5)$$

In Equation (5),  $N(M)$  is the number of earthquakes exceeding magnitude  $M$ , the a-value depends on the total number of observed events, and the b-value controls how frequent larger earthquakes are (typically  $b \approx 1$ ). We estimated the b-value with the maximum likelihood technique (Aki, 1965):

$$b = \frac{1}{\ln(10) (\bar{M} - M_c)}. \quad (6)$$

Equation (6) is derived for continuous magnitudes  $M$  (no bias from binned magnitudes).  $M_c$  is the magnitude of completeness, *i.e.* the magnitude above which all events are detected. We computed  $M_c$  with the maximum curvature technique (*e.g.* Wiemer & Katsumata, 1999, , see Figure S2). Additional information on the computation of  $M_c$  and  $b$  is given in the supporting information (see Text S1.7 and Figure S2).

At each template location, we selected all the templates within a 5 km-radius and used the events they detected to compute  $b$  and  $M_c$ . Following Tormann et al. (2013), we imposed a minimum of 50 events to compute the  $b$ -value and, in addition, requested a minimum of 30 events above the magnitude of completeness. As these numbers are still low, we carefully estimated the uncertainties to assess the statistical significance of  $b$ -value differences between different groups following Utsu (1966). The statistical test is described in detail in Text S1.7 and illustrated in Figure S2.

### 3.3 Temporal Clustering

We refer to temporal clustering as the tendency of earthquakes to influence the timings (advance or delay) of future earthquakes, that is, the non-randomness of earthquake sequences (e.g. Gardner & Knopoff, 1974; Marsan & Lengline, 2008). We quantified the strength of temporal clustering in earthquake sequences by analyzing the statistical properties of the number of earthquakes per unit time, which we refer to as the earthquake occurrence time series:

$$e(t) = \text{Number of events} \in [t; t + \Delta t], \quad (7)$$

where  $\Delta t$  is a user-defined time bin duration, and  $t$  is the calendar time. An example is given in Figure 3A. Burst-like sequences covering wide intervals of recurrence times are not random (see Figure 3B,C) but clustered in time. Time clustered seismicity exhibits time scale invariant characteristics. The spectrum of the earthquake occurrence  $e(t)$  follows a power law of frequency ( $\propto f^{-\beta}$ , see Figure 3D), and the time series  $e(t)$  shows a fractal statistics (Figure 3E). We measured the fractal dimension of  $e(t)$  by subsequently dividing the time axis into smaller and smaller time bins (varying size  $\tau$ ), and counting the fraction of bins  $x$  that were occupied by at least one earthquake (Smalley Jr et al., 1987; Lowen & Teich, 2005). For a certain range of time bin sizes  $\tau$ , we observe:

$$x \propto \tau^{1-D}. \quad (8)$$

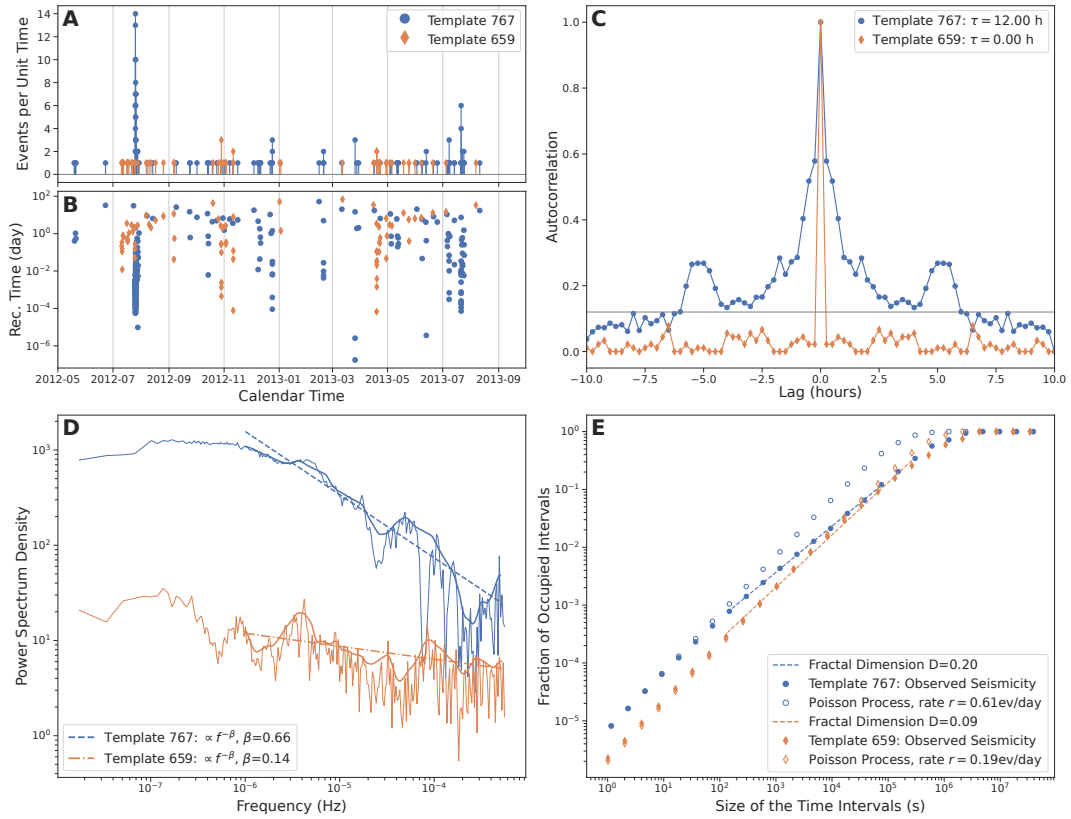
In Equation (8),  $D$  is the fractal dimension of the time series. The fractal dimension varies between the two end-members  $D = 0$  for a point process (e.g. Poisson point process for the background seismicity), and  $D = 1$  for a line (uninterrupted seismicity). A large fractal dimension ( $D > 0.2$ ) characterizes cascade-like activity where past events strongly influence the timings of future events. Fractal analysis has been used in multiple studies to characterize earthquake clustering (Smalley Jr et al., 1987; Lee & Schwarcz, 1995; Beaucé et al., 2019). Note that periodic seismicity does not follow a fractal behavior and cannot be characterized by this method. Building the  $x(\tau)$  curve (Equation (8), Figure 3E) is computationally more simple than estimating the spectrum (Figure 3D). Likewise, it is simpler to fit  $x(\tau)$ . Therefore, we chose to compute the fractal dimension  $D$  to characterize temporal clustering in the rest of this study.

The method described in Section 3.3 does not explicitly deal with space. However, we applied this analysis to subsets of the earthquake catalog containing neighboring earthquakes (as described for the  $b$ -value, see Section 3.2), and thus obtained a fractal dimension for each template.

## 4 Results

### 4.1 The Earthquake Catalog

Our earthquake catalog and detection and location codes are available online (see Data and Resources, and see the supplementary information for details on the structure of the catalog file).



**Figure 3.** Quantifying the strength of temporal clustering in a strongly clustered sequence (Template 767, blue dots) and a weakly clustered sequence (Template 659, orange diamonds). **A:** Number of earthquakes per unit time (referred to as earthquake occurrence, see Equation (7)). **B:** Recurrence times vs. origin times. **C:** Autocorrelation of the earthquake occurrence time series. The horizontal black line is the arbitrary threshold used to define the correlation time  $\tau$ . **D:** Power spectral density of the earthquake occurrence. The linear trend, in the log-log space, is the exponent of the power-law that indicates a scale invariant process. **E:** Fractal analysis of the earthquake occurrence (see text and Equation (8)). We measure the slope between  $dt_{\min}=100$  s and  $dt_{\max} = 1/r$ , where  $r = N/T$  is the average seismic rate (number of events  $N$  divided by time span  $T$ ). For reference, for each template we simulate the seismicity from a Poisson point process with average rate  $r$ . The slope of the Poisson point process gives a fractal dimension  $D = 0$  (i.e. dimension of a point).

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### 4.1.1 Regional Seismicity

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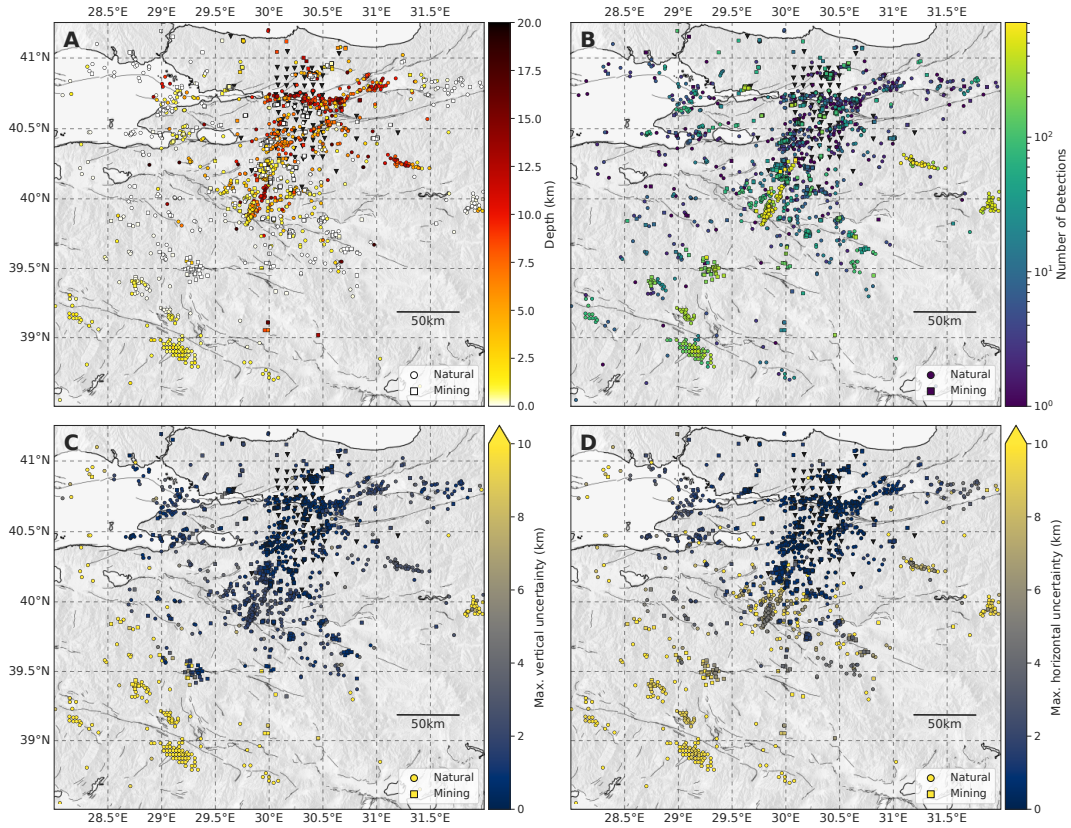
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Following the method described in Section 3.1, we built a database of 3,546 templates and with them detected 35,437 events, including both natural and anthropogenic seismicity. We applied our analysis between 38.50°N-41.50°N and 28.00°E-32.00°E (see Figure 1A). Figure 4 shows the locations of the 3,320 template earthquakes that are shallower than 20 km and have horizontal uncertainties less than 15 km, as well as the cumulative detection count per template over the whole study period. We purposely present an earthquake catalog for this region that extends far beyond the NAFZ to provide a comprehensive description of the earthquake signals found in the data set. We found that most of the 1,972 events detected with templates located deeper than 20 km originated far outside the study region, in particular in the Hellenic and Cyprus subduction zones in the southwest and south of the study region, respectively. Therefore, we discarded these deeper templates for any further analysis. Furthermore, we found that about half



**Figure 4.** Map view of the locations of the template earthquakes detected and used in this study. Only templates with maximum horizontal uncertainty less than 15 km and depth less than 20 km are shown (total of 3,320 templates). Filled dots are for natural earthquakes (1,471 templates), and squares are for mining-related events (1,849 templates; see text for details about identifying templates as mining templates). **A:** Event depths. **B:** Cumulative number of event detections per template. Most of the detected earthquakes actually originate from outside the North Anatolian Fault Zone. **C:** Maximum vertical uncertainty  $v_{\max}$ , i.e. depth range spanned by the projection of the uncertainty ellipse onto a vertical plane. **D:** Maximum horizontal uncertainty  $h_{\max}$ , i.e. length of the major semi-axis of the projection of the uncertainty ellipse onto the horizontal plane.

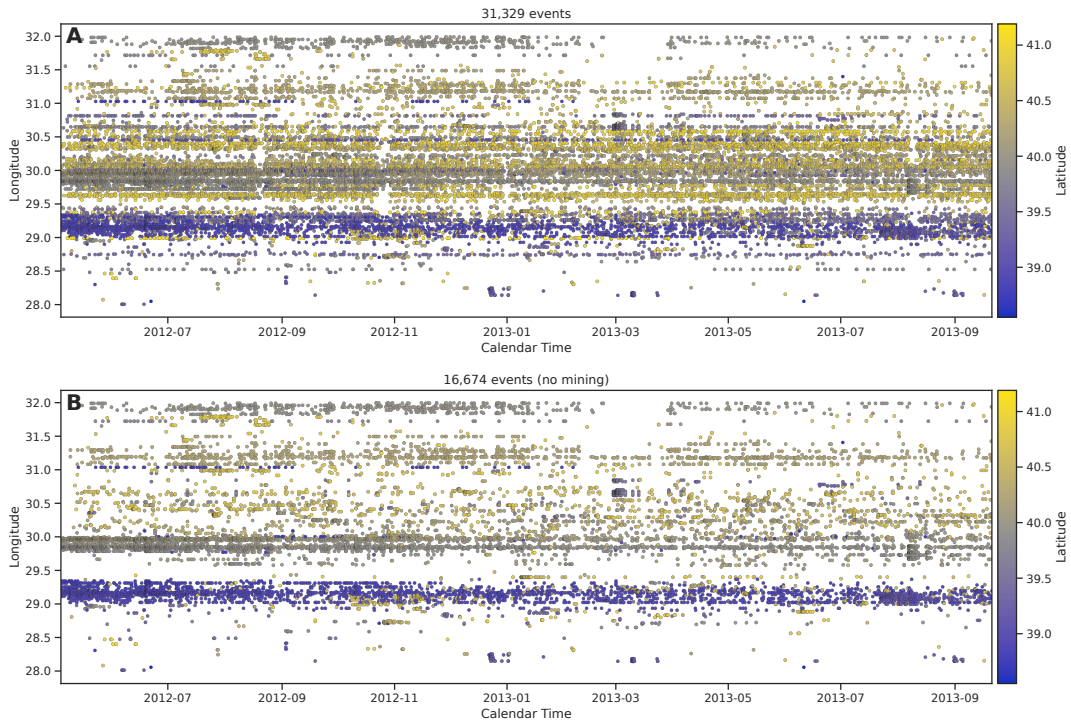
321 of the detected seismicity was due to mining activity (see Section 3.1.6): among the 31,329  
 322 earthquakes detected with the 3,320 templates, we identified 16,674 natural earthquakes  
 323 and 14,655 mining-related earthquakes. The locations of mining activity that we iden-  
 324 tified (see Figure 4) agree well with the analysis of Poyraz et al. (2015) (their Figure 3) whereas  
 325 the Kandilli catalog (see Data and Resources) tends to report less explosions, in partic-  
 326 ular beneath the DANA array (see Figure S4).

327 The majority of earthquakes occurred outside the station array and not in the NAFZ  
 328 itself, that is, north of 40.80°N or south of 40.30°N (see Figure 4B). Location uncer-  
 329 tainties increase with increasing distance from the DANA array: inside 40.30°N-41.00°N and  
 330 30.00°E-30.50°E, the average horizontal uncertainty is  $\bar{h}_{\max} = 0.97$  km and the average  
 331 vertical uncertainty is  $\bar{v}_{\max} = 0.74$  km, whereas these uncertainties increase to  $\bar{h}_{\max} =$   
 332  $8.53$  km and  $\bar{v}_{\max} = 4.57$  km outside this box (see Figure 4C,D).

333 Since accurate moment magnitude estimation rely on correct source-receiver dis-  
 334 tances (see Section 3.1.5), we only computed moment magnitudes for events with  $h_{\max} <$   
 335  $5$  km (see Figure 4D). After the SNR criterion, we could estimate moment magnitudes within

336 168 template families of natural seismicity, from which we computed 1,929 local mag-  
 337 nitudes. These magnitudes range from -1 to 4, and we obtained  $b = 0.85$  and  $M_c = 1.18$   
 338 (see Section 3.2 and Figure S5). We computed a  $M_w - M_L$  calibration close to identity,  $M_w =$   
 339  $0.15 + 0.93M_L$  (see Section 3.1.5 and Figure S1B). The magnitude of completeness of  
 340 our catalog indicates that we were not able to estimate magnitudes below  $M \approx 1$  but  
 341 still detected them: only 12% of the detected seismicity has a magnitude estimate. There-  
 342 fore,  $M_c = 1.18$  is only an upper bound to the magnitude of completeness of the whole  
 343 catalog. For reference, we estimated a b-value and magnitude of completeness of  $b =$   
 344  $0.91$  and  $M_c = 1.05$  with the catalog published in Poyraz et al. (2015), with magnitudes  
 345 ranging from 0 to 4 (see Figure S5). Our magnitudes seemed to be systematically larger  
 346 than theirs for smaller events, with an average difference of 0.5 unit over all compared  
 347 events (see Figure S5C). Detailed b-values and magnitudes of completeness are presented  
 348 in Section 4.2.

349 We present the spatio-temporal distribution of the seismicity in Figure 5. An over-  
 350 all decaying activity of natural earthquakes is superimposed to a uniform mining-related  
 351 activity (compare Figure 5A vs. B). We observe two sequences of slowly decaying activ-  
 352 ity below  $39^\circ\text{N}$  and around  $40^\circ\text{N}$ . The southernmost earthquake sequence ( $39^\circ\text{N}$ ) is part  
 353 of the aftershock activity of the M5.1 2012-05-03  $39.18^\circ\text{N}/29.10^\circ\text{E}/5.4$  km earthquake  
 354 (just before the deployment of DANA). The  $40^\circ\text{N}$  sequence is not featured in the Kandilli  
 355 nor in the United States Geological Survey catalog.



**Figure 5.** Spatio-temporal distribution of the earthquake activity in the study region. The longitude of each event is shown against its origin time, and the color codes the latitude. **A:** We detected 31,329 events with the 3,320 template earthquakes presented in Figure 4 from 2012-05-04 to 2013-09-20. **B:** The templates due to natural seismicity detected 16,674 earthquakes. The seismic activity taking place on the NAFZ (latitudes  $40.35^\circ\text{N}$ - $40.80^\circ\text{N}$ ) represents a small amount of the total seismicity ( $\sim 2,000$  events).

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#### 4.1.2 Seismicity of the North Anatolian Fault Zone

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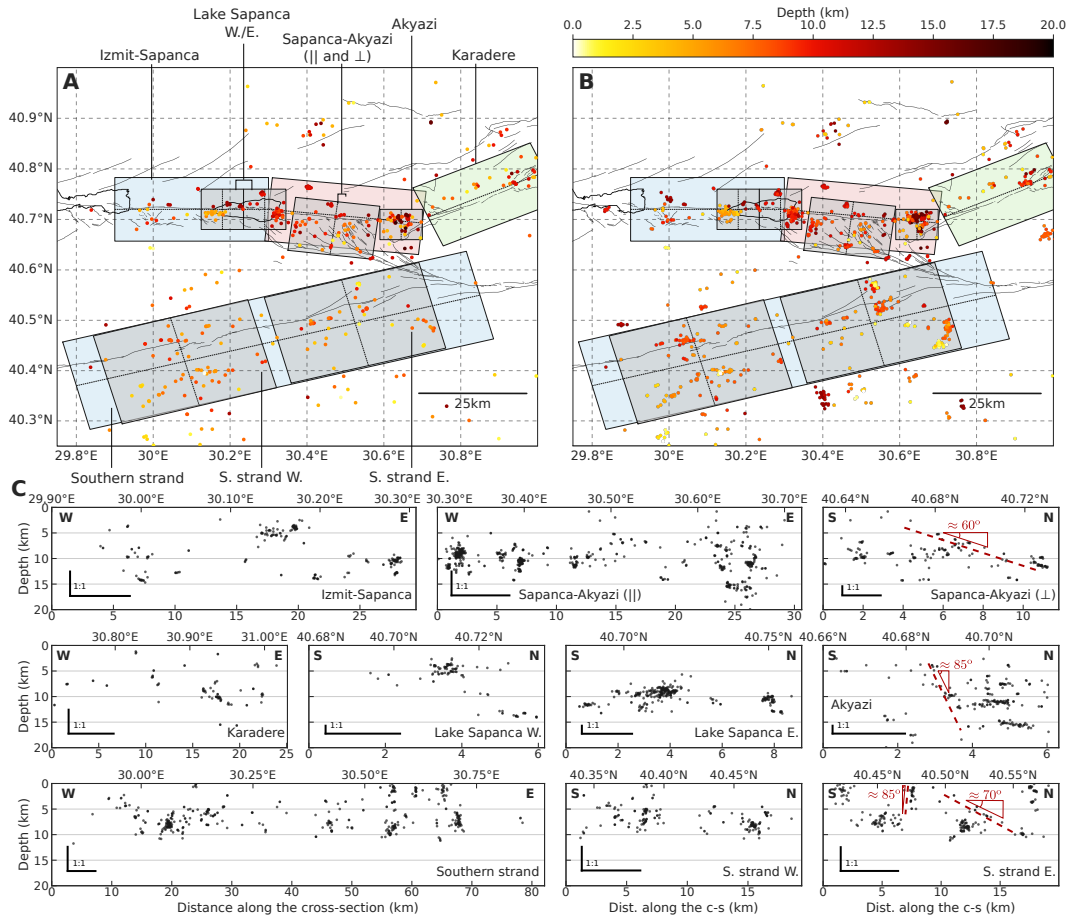
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Here, we focus on the template earthquakes located in the vicinity of the NAFZ and near the station array (40.25°N-41.00°, 29.80°E-31.00°E). Figure 6 shows the locations of these template earthquakes, as well as the 2,141 earthquakes relocated with the double-difference method (see Section 3.1). The median horizontal and vertical errors on relative locations are 73 m and 91 m, respectively, meaning that they can reliably be interpreted in terms of active structures. Earthquake hypocenters reveal a complex network of faults, with much of the seismicity occurring on secondary faults rather than on the NAFZ itself. We divided the fault zone into nine subregions (*cf.* Figure 6A) whose names we will keep referring to in this manuscript. These are organized into four along-strike sections: Izmit-Sapanca, fault-parallel Sapanca-Akyazi, Karadere, and the entire southern strand, and six fault-perpendicular sections: Lake Sapanca west and east, fault perpendicular Sapanca-Akyazi, Akyazi, and the southern strand west and east. The northern strand is overall more active than the southern strand, and the Sapanca-Akyazi segment hosts the densest activity. In particular, both terminations of the segment, the eastern side of Lake Sapanca and the area around the Akyazi fault, host strong seismicity. The Akyazi region features the deepest seismicity in the vicinity of the NAFZ (down to 20 km). The group of earthquakes located at the northernmost of the Sapanca-Akyazi region (Figure 6A-B) are part of the 2012-07-07  $M_L$ 4.1 Serdivan earthquake sequence. Most of the seismicity along the southern strand occurs in areas where surface fault traces indicate more structural complexity. Note that the relocated seismicity tends to be distributed in patches, which is partly due to the detection method. Indeed, template matching tends to detect groups of collocated earthquakes, whereas small events located in between template earthquakes may remain undetected.

The fault parallel and fault perpendicular cross-sections in Figure 6C show the events' depth distribution. The seismicity is enhanced in the lower half of the seismogenic zone: 7-15 km along the northern strand, and even deeper than 15 km around the Akyazi fault, and 5-10 km depth along the southern strand. The main exception to that depth distribution are the earthquakes at the western side of Lake Sapanca, with hypocenters clustered around 5 km depth. The Lake Sapanca W. cross-section (see Figure 6C) shows that this shallow seismicity seems restricted to the southern side of the fault, namely the Armutlu Bloc.

The map views and cross-sections in Figure 6 suggest a narrower deformation zone in the north where seismicity is mostly distributed within 5-10 km of the main fault trace, whereas we observe a wider deformation zone along the southern strand with seismicity distributed within 15-20 km of the fault trace. We emphasize that the detected microseismicity illuminates the deformation zone associated with the NAFZ rather than the main fault itself. The Sapanca-Akyazi and Akyazi fault perpendicular cross-sections could indicate a north dipping deformation zone with a 60° dip angle in the middle of the Sapanca-Akyazi segment, and 85° near the Akyazi fault. However, these trends are weak and hypocenters mostly show horizontal alignments. Identifying a global dip direction of the deformation zone along the southern strand is equally ambiguous. In the east, one could either identify slightly south dipping structures ( $\sim 85^\circ$ ) or more strongly north dipping structures ( $\sim 70^\circ$ ).

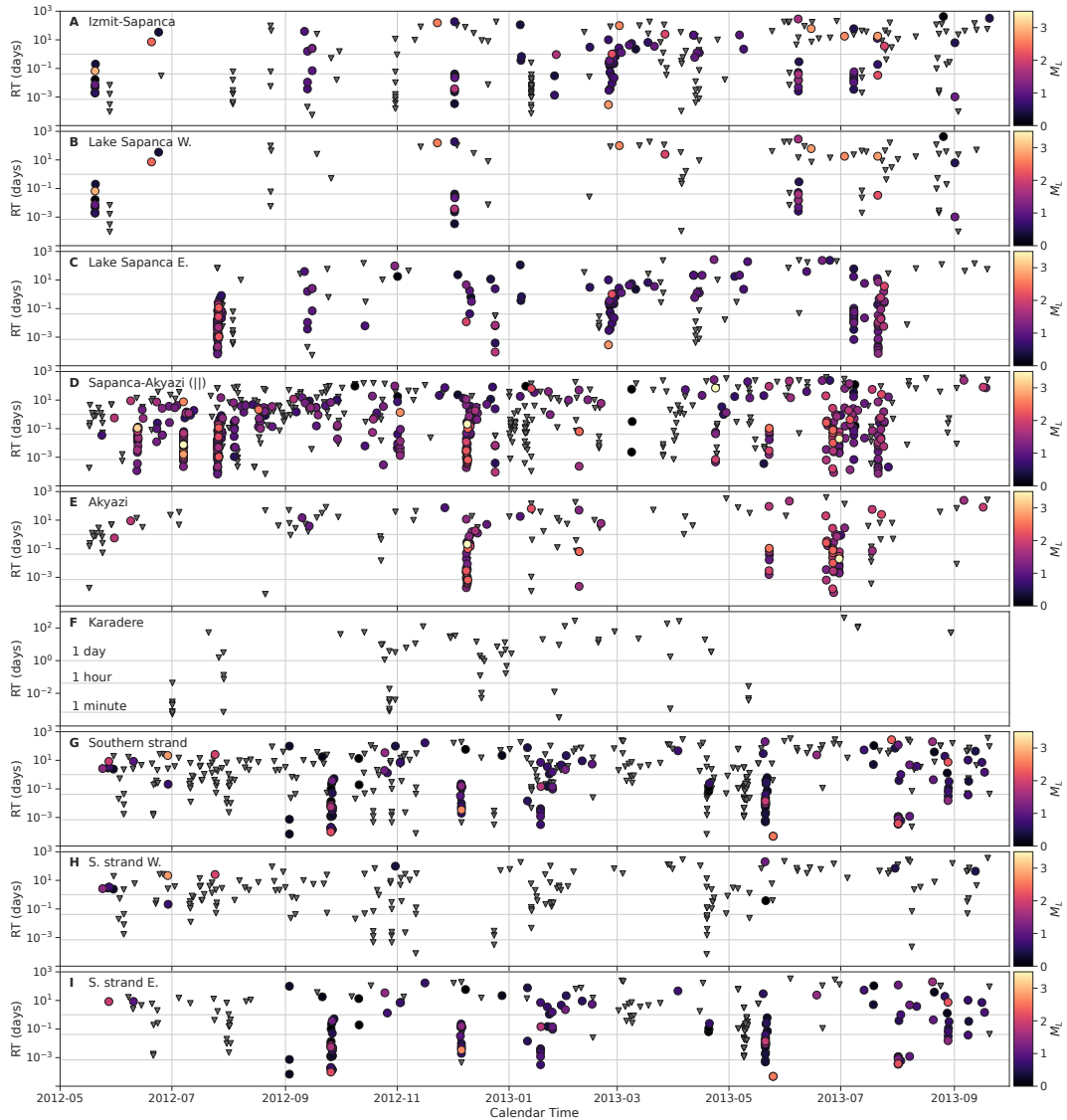
We present the temporal distribution of the seismicity in Figure 7. The recurrence times are given against their detection times for each of the nine cross-sections introduced above. The recurrence time is the time interval between two consecutive co-located earthquakes. In practice, recurrence times are computed as the time intervals between consecutive events detected by a same template. The most striking feature of Figure 7 is the organization of some earthquake sequences into bursts of seismicity with recurrence times spanning many orders of magnitude. These sequences are time clustered (see Section 3.3). These bursts are usually associated with sequences of foreshocks-mainshock-aftershocks, although in general earthquake sequences can have no clear mainshock (that is, an event of magnitude larger than all other events of the sequence) and still exhibit a strong burst-like behavior. The seismicity at the eastern end of Lake Sapanca and near



**Figure 6.** Earthquakes in the North Anatolian Fault Zone. **A:** Locations of the template earthquakes with color coded depths. We define nine subregions along the different segments of the fault. Only in this figure the Sapanca-Akyazi region is subdivided into a fault parallel and a fault perpendicular sections. The thin black dotted lines inside each colored box define either fault parallel or fault perpendicular cross-sections (see bottom panels, C). The color shading of each box is only to help distinguish between them. **B:** Earthquake hypocenters successfully relocated with the double-difference method and color coded by depth. Events for which relocation was not successful were attributed the template location. **C:** Depth cross-sections of the different areas introduced above. The earthquake locations contained in the boxes are projected onto the boxes' central axis. The bottom x-axes are distances along the cross-section axes in kilometers, and the top x-axes are the geographic coordinates relevant to each cross-section (either longitude or latitude). Note that the aspect ratio across cross-sections varies. The 1:1 aspect ratio is drawn in the lower left corner of each cross-section. The dashed red lines and angles are given for reference but are not our take-away message.

411 Akyazi is almost exclusively organized into such sequences of burst-like seismicity, whereas  
 412 the southern strand hosts much less of these burst-like episodes. Figure 7 also reports  
 413 the local magnitudes (see Section 3.1.5). The Sapanca-Akyazi segment and its vicinity is  
 414 the most active region with the largest magnitude events observed during the study per-  
 415 iod. Among the nine  $M_L \gtrsim 3$  natural earthquakes we detected, three occurred near  
 416 each other, close to the city of Serdivan, including the largest event of the study: the 2012-  
 417 07-07  $M_L 4.1$  Serdivan earthquake ( $30.404^\circ\text{E}/40.763^\circ\text{N}/11.3$  km). The area around the Akyazi

418 fault also produced four  $M_L > 3$  earthquakes, whereas earthquakes near Lake Sapanca  
 419 did not exceed  $M_L = 3$ .



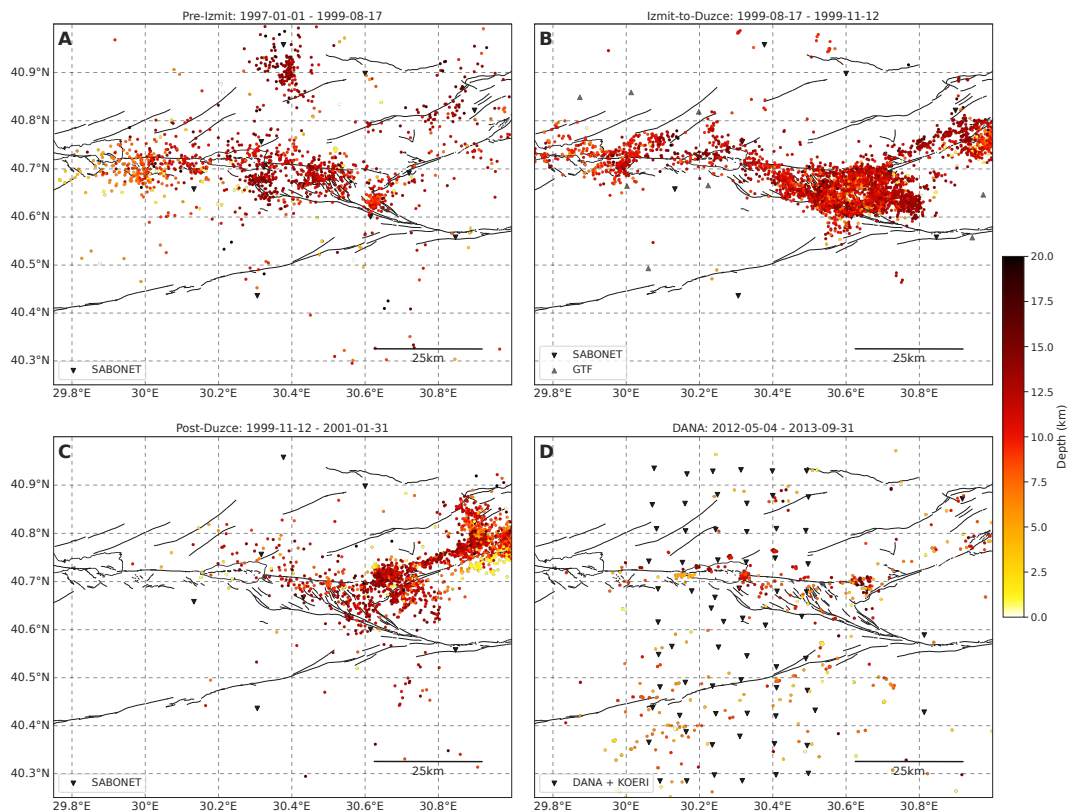
**Figure 7.** Time evolution of the earthquake recurrence times for different subsets of the earthquake catalog (refer to Figure 6 for the name of the areas). The recurrence time is the time between two consecutive events detected by a same template. Note that the y-axis is in log scale and that some seismic episodes span many orders of magnitude of recurrence time. These episodes are characteristic of burst-like, or cascade activity (see text). The color scale indicates the local magnitude, and inverted grey triangles are events for which no reliable estimates were obtained.

#### 4.1.3 Comparison with Past Seismicity

420 We combined different earthquake catalogs to compare the 2012-2013 detected  
 421 seismicity with the pre-Izmit, Izmit-Düzce, and early post-Düzce seismicity (Bulut et al.,  
 422 2007; Ickrath et al., 2015; Bohnhoff et al., 2016, and see Figure 8). We note that the Izmit-Düzce  
 423 earthquake catalog is more complete in the west (around the Izmit-Sapanca seg-  
 424 ment) than the pre-Izmit and early post-Düzce catalogs due to the higher number of sta-  
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426 tions used in this time period (see, e.g. Ickrath et al., 2015, and Figure 8B). It is also worth  
 427 mentioning that these three catalogs show both natural and mining-related seismicity whereas  
 428 we have discarded the man-made seismicity to the best of our ability (see Section 3.1.6).  
 429 During these three time periods, the (moment) magnitudes of completeness of these cat-  
 430 alogs are  $M_c = 1.56$ , 1.69, and 1.44, respectively, and few  $M_w < 1$  earthquakes are re-  
 431 ported (see Figure S6). Using our  $M_w$ - $M_L$  calibration to convert our local magnitudes to  
 432 effective moment magnitudes, we obtained  $M_c = 1.18$  and 27% of the earthquakes con-  
 433 tributing to the frequency-magnitude distribution have  $M_w < 1$  (see Figure S6D). There-  
 434 fore, also recalling that most of the detected events are too small to be characterized by  
 435 their magnitude, our catalog reports smaller events than the catalogs we compare it with.



**Figure 8.** Comparison of the **A:** pre-Izmit, **B:** Izmit-Düzce, the SABONET stations were comple-  
 mented by 21 temporary stations from the German Task Force (GTF, grey triangles), **C:** early post-  
 Düzce, and **D:** late post-Düzce seismicity. The inverted black triangle are the seismic stations and the  
 colored dots are the earthquake locations.

436 The middle sections of the Izmit-Sapanca and Sapanca-Akyazi segments were particu-  
 437 larly active seismically before the Izmit earthquake, and some clusters of earthquakes  
 438 were observed beneath Lake Sapanca (Figure 8A). The Izmit earthquake is known to have  
 439 nucleated near a swarm of seismicity that was active before the M7.4 event (Crampin et  
 440 al., 1985; Lovell et al., 1987; Ito et al., 2002). In the three months between the Izmit earth-  
 441 quake and the Düzce event, the seismic activity was strongest in the area around the triple  
 442 junction between the Sapanca-Akyazi segment, the Karadere segment, and the Mudurnu  
 443 fault (Figure 8B). The Izmit hypocentral region remained active and, comparatively, lit-  
 444 tle activity was detected near Lake Sapanca. After the Düzce earthquake, most activity  
 445 along the Izmit-Sapanca and Sapanca-Akyazi terminated, and seismicity concentrated along  
 446 the Karadere segment (Figure 8C). The Akyazi region, where little coseismic slip was ob-  
 447 served (Ozalaybey et al., 2002; Bohnhoff et al., 2006, 2008), hosted a cluster of strong

448 activity, possibly driven by the Izmit residual stresses. Note that no seismicity was de-  
 449 tected near Lake Sapanca. About 13 years after the Izmit and Düzce earthquakes, we de-  
 450 tected the strongest activity at the eastern side of Lake Sapanca, and near the Akyazi fault  
 451 (Figure 8D). If not due to the absence of  $M < 1$  earthquakes in these catalogs, the lack  
 452 of intense seismicity near Lake Sapanca in the early post-Düzce period suggests that faults  
 453 near Lake Sapanca did not slip during the afterslip-driven aftershock sequence with Omori-  
 454 like decaying seismicity (Perfettini & Avouac, 2004). Moreover, the Omori law predicts  
 455 a seismicity rate about four orders of magnitude lower 13 years after the mainshock (using  
 456 Omori law parameters from Bayrak & Öztürk, 2004), therefore the seismic activity near  
 457 Lake Sapanca should have been high after the Izmit earthquake if the 2012-2013 seis-  
 458 micity were to be remnants of aftershocks. The 2012-2013 Lake Sapanca seismicity also  
 459 appears much stronger than the pre-Izmit seismicity (Figure 8A).

## 460 4.2 Observed b-values

461 Computed b-values and magnitudes of completeness (see Section 3.2) are presented  
 462 in Figure 9. Of most interest to this study, we see that earthquakes at the eastern side  
 463 of Lake Sapanca exhibit higher b-values ( $b \approx 1.1$ ) than earthquakes near the Akyazi fault  
 464 ( $b \approx 0.8$ , see Figure 9D). The magnitude of completeness varies from  $M_c \approx 1.3$  near  
 465 Akyazi to  $M_c \approx 1.0$  near Lake Sapanca. We note that visual checking of the frequency-  
 466 magnitude distributions showed that, in general, they follow the Gutenberg-Richter law  
 467 well, except for the Serdivan earthquakes where a peak around  $M_L \approx 2.5$  can be ob-  
 468 served. The significance of the b-value difference between the eastern Lake Sapanca and  
 469 Akyazi was assessed by applying the statistical test described in (Utsu, 1966) and we found  
 470 that the difference was significant at the 96% confidence level (see Text S1.7).

## 471 4.3 Observed Temporal Clustering

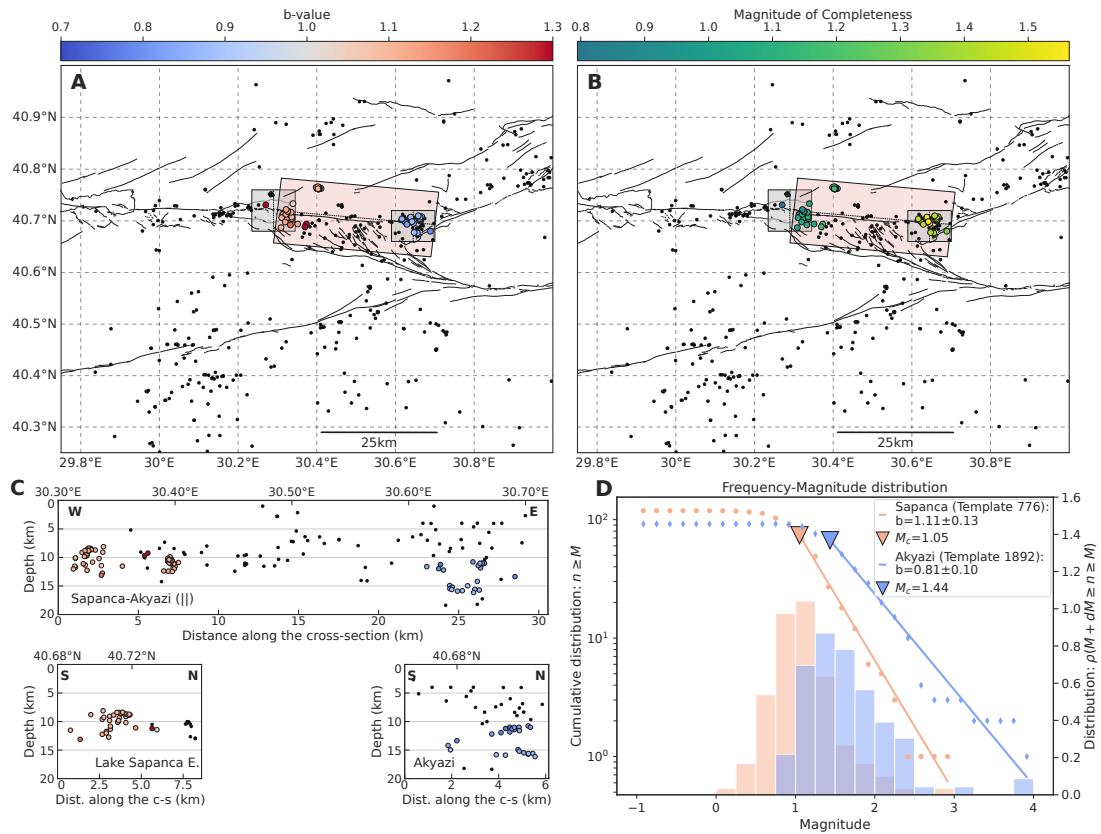
472 We characterized temporal clustering as a function of space (see Figure 10) follow-  
 473 ing the method described in Section 3.3. The strongest temporal clustering (fractal di-  
 474 mension  $D > 0.20$ ) is observed on the eastern side of Lake Sapanca, beneath the so-  
 475 called Rangefront trace. Other areas of strong activity, like the Serdivan earthquakes (around  
 476  $30.404^\circ\text{E}/40.763^\circ\text{N}$ ) and the Akyazi area, only show small-to-moderate temporal cluster-  
 477 ing ( $D < 0.14$ ), thus confirming the outstanding character of the eastern Lake Sapanca.  
 478 We note that while the temporal organization of recurrence times shown in Figure 7 in-  
 479 dicated burst-like seismicity in all of the above mentioned areas, this quantitative anal-  
 480 ysis was necessary to distinguish between strongly and moderately time clustered se-  
 481 quences. A few other isolated locations exhibit strong temporal clustering, and seem to  
 482 be systematically occurring near the bottom of the seismogenic zone (cf. Figure 10C). Com-  
 483 paring the cumulative number of detections per template and their fractal dimension shows  
 484 that there is no trivial correlation between the two (see Figure 10A vs. B). We note that  
 485 we did the same fractal analysis on all templates of the study region and found another  
 486 region of strong temporal clustering on the NAFZ, in the eastern Marmara Sea, where  
 487 the 1999 Izmit earthquake arrested (see Figure S7).

## 488 5 Interpretation and Discussion

489 In the present section, we discuss implications of the observed spatial earthquake  
 490 distribution for the mechanical state of the NAFZ (Section 5.1), we interpret the b-values  
 491 in terms of low and high stresses (Section 5.2), and we explain how temporal clustering  
 492 can be related to fault rheology (Section 5.3).

### 493 5.1 Spatial Distribution of Seismicity

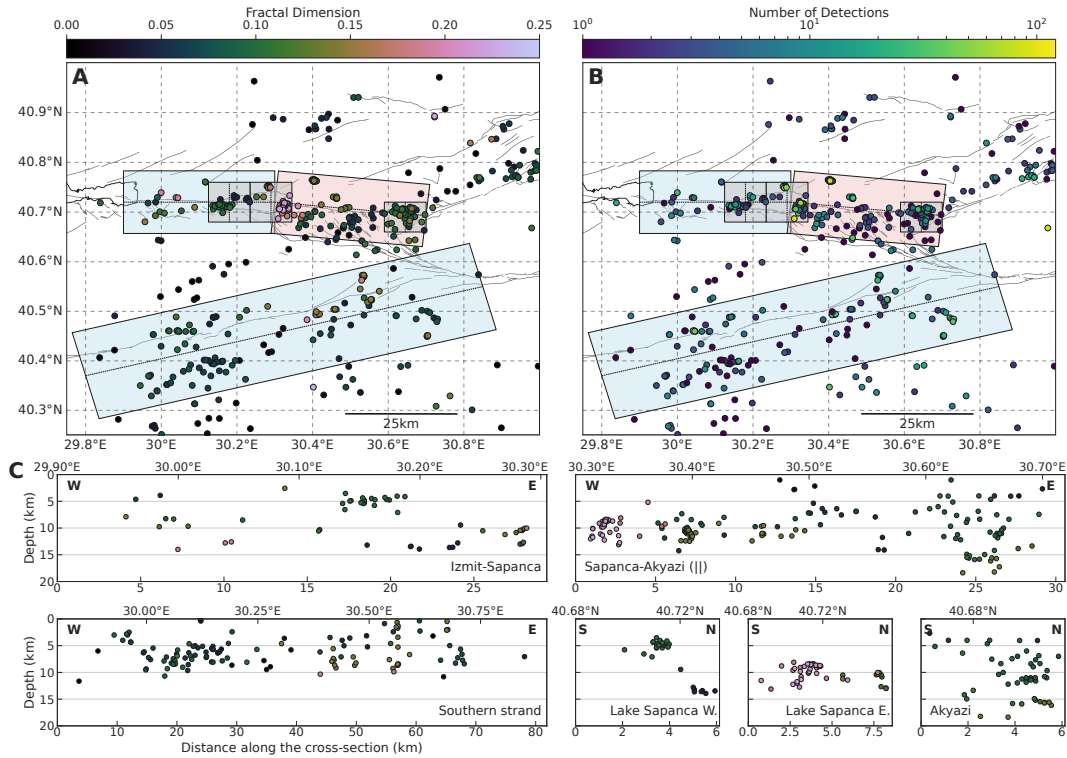
494 Apart from the Karadere segment, the seismicity is taking place off the main fault  
 495 on a complex network of secondary faults, similarly to the Izmit-Düzce aftershocks (e.g.



**Figure 9.** **A:** Map view of template earthquakes with color coded Gutenberg-Richter  $b$ -value. Smaller black dots are event families for which we could not compute moment magnitudes (see text). **B:** Map view of template earthquakes with color coded magnitude of completeness. In both top panels, the shaded areas refer to the regions introduced in Figure 6. **C:** Template earthquakes with color coded  $b$ -value on fault parallel and fault perpendicular cross-sections. Hypocenters are projected along the dotted axes shown on the map view. **D:** Estimation of the  $b$ -value and its uncertainties for two earthquake populations near Lake Sapanca (red,  $b \approx 1.1$ ) and near Akyazi (blue,  $b \approx 0.8$ ). Cumulative (scatter plot) and non-cumulative (histogram) frequency-magnitude distributions. The dashed curves are the kernel density estimate of the non-cumulative probability density functions (pdf). The mode of the pdf is used as the magnitude of completeness (maximum curvature method). The  $b$ -value is computed with the maximum likelihood estimate (MLE, Equation (6)).

496 Ozalaybey et al., 2002; Bulut et al., 2007, and see Figure 6). This feature is in stark contrast  
 497 with the simplicity of the Izmit and Düzce earthquakes, which occurred on simple  
 498 fault segments (Barka et al., 2002; Langridge et al., 2002). Off-fault seismicity has also  
 499 been observed to be a characteristic of fault zones early in their seismic cycle (Ben-Zion  
 500 & Zaliapin, 2020) and might be due to off-fault, distributed deformation contributing to  
 501 accommodate slip deficits resulting from heterogeneous slip along the fault (Dolan & Har-  
 502 avitch, 2014).

503 Shallow creep has been observed along the Izmit-Sapanca and the Sapanca-Akyazi  
 504 segments (e.g. Çakir et al., 2012; Hussain et al., 2016; Aslan et al., 2019). The creep rates  
 505 and creep along the Izmit rupture have evolved with time (e.g. Bürgmann et al.,  
 506 2002). Aslan et al. (2019) use 2011-2017 InSAR data and 2014-2016 GPS data and, thus,  
 507 is the closest study to ours in time. The authors' model shows shallow creep down to 5 km  
 508 along the Izmit-Sapanca segment and down to 2 km at the western end of the Sapanca-



**Figure 10.** **A:** Map view of template earthquakes with color coded fractal dimension (*cf.* Equation (8)) showing the strength of temporal clustering. **B:** Map view of template earthquakes with color coded cumulative number of detections. In both top panels, the shaded areas refer to the regions introduced in Figure 6. **C:** Template earthquakes with color coded fractal dimension on fault parallel and fault perpendicular cross-sections. Hypocenters are projected along the dotted axes shown on the map view. High fractal dimensions mean strongly time clustered activity (*i.e.* past events strongly influence the timings of future events).

509 Akyazi segment. Such shallow creep should drive microseismicity in the vicinity of the  
 510 creeping fault sections (*e.g.* Lohman & McGuire, 2007). The depth cross-sections (Fig-  
 511 ure 6C) only show shallow seismicity at the western Lake Sapanca ( $\approx 5$  km depth). Although  
 512 these depths are consistent with the creep depth given in Aslan et al. (2019), a direct causal-  
 513 ity link to the shallow creep is not straightforward because it is taking place off-fault (Fig-  
 514 ure 6B). Seismicity along the Sapanca-Akyazi segment do not support creep-driven activ-  
 515 ity at shallow depths ( $\approx 2$  km). However, hypocenters suggest that, at the time of the  
 516 study, the base of the seismogenic zone is around 10-15 km, which is in good agreement  
 517 with Aslan et al. (2019).

518 Comparing our catalog with the seismicity in the past (see Section 4.1.3) showed  
 519 that the eastern Lake Sapanca did not appear to be a particularly active area, either be-  
 520 fore or right after the Izmit earthquake. However, this comparison relies on catalogs made  
 521 with different methods and station coverage, and, consequently, different magnitudes  
 522 of completeness. The seismic activity at the eastern Lake Sapanca may be a permanent  
 523 feature of the step-over that can only be observed with low magnitude of completeness  
 524 catalogs ( $M_c \lesssim 1.0$ ). We further discuss in Section 5.4 whether the seismicity at the east-  
 525 ern Lake Sapanca is a new feature of the fault zone caused by the post-Izmit deforma-  
 526 tion or is a constant phenomenon that could not be observed in such detail in the past.  
 527 Variations in seismicity along the southern strand are harder to interpret because the lack

528 of earthquakes in previous catalogs (see Figure 8) is partly due to the absence of stations  
 529 in the past.

## 530 5.2 Gutenberg-Richter $b$ -value

531 Laboratory experiments have shown that the  $b$ -value seems to be controlled by the  
 532 state of stress, specifically that  $b$  decreases with increasing differential stress (e.g. C. H. Scholz,  
 533 1968; Amitrano, 2003). Decreasing  $b$ -value with depth (Mori & Abercrombie, 1997; Wiemer  
 534 & Wyss, 1997) and high  $b$ -value along creeping sections (e.g. Amelung & King, 1997; Wiemer  
 535 & Wyss, 1997) also support the negative correlation of  $b$  with stress. Thus, the  $b$ -value  
 536 can be used as a stressmeter.

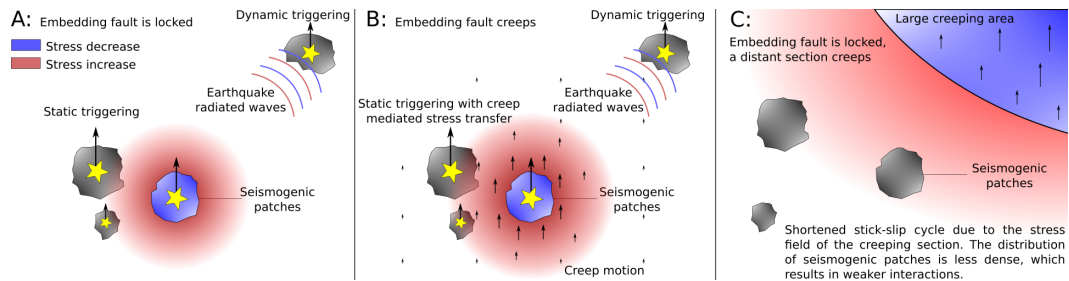
537 Our results (Figure 9) show a clear difference in  $b$ -values between the eastern Lake  
 538 Sapanca ( $b \approx 1.1$ ) and the Akyazi ( $b \approx 0.8$ ) seismicity. We recall that this difference is  
 539 significant at the 96% confidence level (see Section 4.2). We interpret the higher  $b$ -values  
 540 at Lake Sapanca as an indication of low background stresses, while we interpret the lower  
 541  $b$ -values at Akyazi as indicating high background stresses. Low stress at the eastern Lake  
 542 Sapanca suggests that aseismic slip might play a role in driving the seismicity, implying  
 543 that, there, faults have weak sections. High stress near Akyazi can be understood as re-  
 544 sulting from the stress concentration that occurred during the Izmit earthquake, when  
 545 little co-seismic slip occurred along the Akyazi fault and the Akyazi gap.

## 546 5.3 Temporal Clustering, Earthquake Interactions, and Fault Mechanical Proper- 547 ties

548 Strongly time clustered seismicity with a wide range of recurrence times, as presented  
 549 in Section 4.3, cannot be explained only by fluctuations of the background seismicity rate,  
 550 for example due to the injection of fluids at depth. Indeed, a Poisson point process with  
 551 a transient increased rate only shifts the distribution of recurrence times towards shorter  
 552 times but does not widen the distribution and does not have a large fractal dimension (see  
 553 Figure S8). Temporal clustering, that is, cascading of events, emerges when different faults  
 554 or sections of a fault interact (e.g. Burridge & Knopoff, 1967; Marsan & Lengline, 2008;  
 555 Fischer & Hainzl, 2021). Earthquakes can trigger each other due to the static stress changes  
 556 induced by the co- and postseismic displacements (e.g. King & Cocco, 2001), but also due  
 557 to the dynamic stress changes induced by the elastic waves radiated by the rapid coseis-  
 558 mic motions (e.g. Fan & Shearer, 2016). Furthermore, because of the stress redistribu-  
 559 tion following any slip motion (not necessarily at seismic speeds), interaction can occur  
 560 between a seismogenic asperity and its creeping surroundings: accelerated creep (e.g. af-  
 561 terslip) increases the stressing rate on the asperity (e.g. Cattania, 2019; Cattania & Segall,  
 562 2021). In realistic, complex conditions where seismic and aseismic slip co-occurs on short  
 563 length scales (e.g. Collettini et al., 2011), numerical models show that both co-seismic and  
 564 creep mediated stress changes are important factors controlling the clustering of earth-  
 565 quakes (Dublanche et al., 2013; Cattania & Segall, 2021). The contribution of creep me-  
 566 diated stress transfers to temporal clustering might even be more important than static  
 567 stress changes due to the breaking of asperities (Dublanche, 2019). In fact, this means  
 568 that both seismic and aseismic events can cluster in time, but that earthquake catalogs  
 569 only capture the seismic signature of temporal clustering. Effectively, these interacting  
 570 stress fields result in a clock advance or delay in the cycle of the earthquake sources (e.g.  
 571 Harris et al., 1995; Gomberg et al., 1998) and thus in non-random earthquake sequences.

572 Figure 11 sketches different earthquake interaction scenarios explaining tempo-  
 573 ral clustering: in a locked fault Figure 11A, and with creep mediated stress transfers Fig-  
 574 ure 11B. Note that remote creep acting on a sparse asperity population (Figure 11C) would  
 575 produce Poissonian seismicity (e.g. Lohman & McGuire, 2007). Thus, areas of strong tem-  
 576 poral clustering (see Figure 10) indicate faults with intrinsic properties: heterogeneous  
 577 rheology resulting in juxtaposed seismic and aseismic slip, and rough or densely fractured  
 578 fault zone providing many seismogenic asperities. These properties enhance interaction-

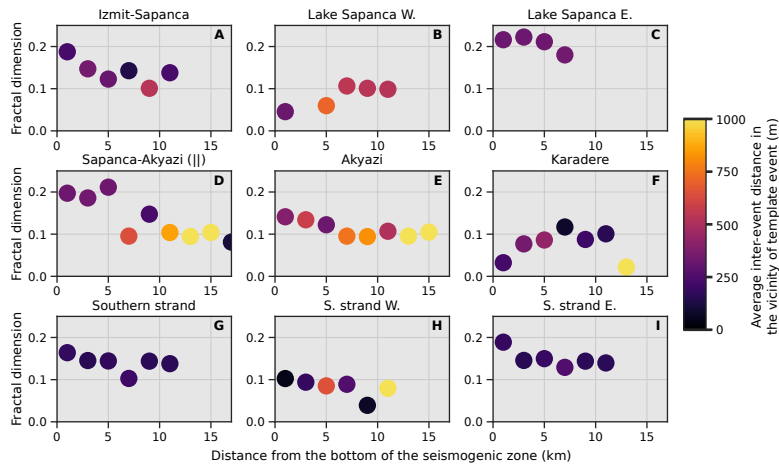
579 driven seismicity, that is, driven by the redistributed stresses of past events. However,  
 580 the long time-scale behavior of clustered seismicity may be modulated by time-dependent  
 581 remote forcing.



**Figure 11.** Sketch of different earthquake interaction scenarios. **A:** Seismogenic asperities embedded in a locked fault. **B:** Seismogenic asperities embedded in a creeping fault. In **A** and **B**, the color shows the stress change due to rupture of the seismogenic patch. The triggered ruptures occur with some delay. **C:** Seismogenic asperities embedded in a locked fault, but stressed by a remote creeping section of the fault. The asperities are not close enough to the creeping patch to strongly interact via static stress changes. The spatial configuration of asperities does not promote strong interactions.

582 Where rheology transitions from brittle to ductile, for example at the base of the  
 583 seismogenic zone, faults are likely to host both unstable, seismic slip and stable, aseis-  
 584 mic slip (C. H. Scholz, 1998; Skarbek et al., 2012). Therefore, seismicity near the bottom  
 585 of the seismogenic zone would be expected to display temporal clustering because, there,  
 586 interacting asperities are likely to be embedded in a creeping fault (cf. Figure 11B, Dublanchet  
 587 et al., 2013). We investigated the relationship between temporal clustering and the prox-  
 588 imity to the bottom of the seismogenic zone to elucidate the role of fault stability in our  
 589 observations (i.e. scenario Figure 11A vs. 11B). The results, in Figure 12, indicate that,  
 590 as expected, seismicity tends to get more time clustered as it gets closer to the brittle-  
 591 ductile transition and that strong clustering almost always happens at the bottom of the  
 592 seismogenic zone. Exceptions are at the western side of Lake Sapanca (Figure 12B) where  
 593 results might be biased due to the absence of significant seismicity at depth, and along  
 594 the Karadere segment (Figure 12F) where large source-receiver distances yield poor hypocen-  
 595 tral depth resolution and thus low confidence results.

596 We also investigated a possible correlation between the proximity to the brittle-ductile  
 597 transition and the density of seismic sources, which could as well explain the increase in  
 598 temporal clustering. We took the average inter-event distance within neighboring earth-  
 599 quake subcatalogs as a proxy for asperity density. We note that this measure of asper-  
 600 ity density is imperfect because a single asperity can break repeatedly. The smaller num-  
 601 ber of detected earthquakes along the southern strand might also be insufficient to com-  
 602 pute a meaningful average inter-event distance. We do not observe a clear systematic  
 603 increase in asperity density with decreasing distance from the bottom of the seismogenic  
 604 zone, but the observational limits mentioned above prevent us from drawing definite con-  
 605 clusions. Figure 12 rather shows that both the proximity to the brittle-ductile transition  
 606 and a large event density favor temporal clustering. Our observations therefore support  
 607 that dense asperity populations along with creep mediated stress transfers do promote  
 608 strong temporal clustering (cf. Figure 11B, Dublanchet et al., 2013). Thus, this study sug-  
 609 gests that faults at the eastern side of Lake Sapanca are in heterogeneous stability regimes  
 610 allowing unstable (seismic) and stable (aseismic) slip.



**Figure 12.** Clustering vs. depth vs. event density. Inside each region, templates are binned per distance from the bottom of the seismogenic zone and the fractal dimension is averaged among the 10% largest values, resulting in a "soft" maximum of each bin. The location of the bottom of the seismogenic zone is approximated by the depth of the locally deepest template. Dots are colored according to the average inter-event distance within the neighboring earthquake subcatalogs; this is a proxy for asperity density. Darker colors mean higher density. Strongest clustering tends to occur at the bottom of the seismogenic zone, *i.e.* at the transition zone between unstable (brittle) and stable (ductile) sliding.

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#### 5.4 Implications for the Lake Sapanca Step-Over

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In summary, the Gutenberg-Richter  $b$ -values (see Section 5.2) and temporal clustering (see Section 5.3) point to the role of different rheological properties in producing earthquakes between the two sides of Lake Sapanca. At the western side, the shallow active sections seem incapable of producing strongly time clustered seismicity. At the eastern side, the depth distribution, the strong temporal clustering (Figure 10C), and the relatively high  $b$ -values (Figure 9) suggest that thin along-dip fault sections slip in a mixed seismic and aseismic mode.

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Heterogeneous faults near the brittle-ductile transition have stable and unstable sections (e.g. Collettini et al., 2011). Weakly unstable sections may produce transient episodes of slow slip (e.g. Bürgmann, 2018). The temporal distribution of earthquakes at the eastern Lake Sapanca (see Figure 7C) suggests that faults are slipping during intermittent episodes of deformation. Thus, the weakest sections of the faults at the eastern Lake Sapanca might be intermittently driven to slowly slip and, in turn, activate the seismogenic asperities (Skarbek et al., 2012; Cattania & Segall, 2021). The lack of seismicity along the up-dip sections suggests they are either fully locked or fully creeping, but there is no evidence for such a large creeping section in geodetic data (e.g. Aslan et al., 2019). Geologic data suggest that the so-called Sapanca Complex, constituted of weak serpentinites and strong metabasites (Akbayram et al., 2013, and references therein), might reach the southeastern side of Lake Sapanca at depth where we observe the highly clustered seismicity. Such lithology is consistent with the scenario of strong asperities embedded in a weak, stable fault. How much seismic moment is released through (partial) aseismic slip during these episodes of strong microseismicity remains an open question.

Whether these intermittent episodes of deformation are a permanent feature of Lake Sapanca or result from the mechanical changes that faults underwent because of co- and post-seismic stress changes is hard to elucidate entirely since our comparison with the past seismicity relies on unequal catalogs (see Section 4.1.3). However, we know from

geodetic data that north-south extension around the Lake Sapanca step-over accelerated considerably following the Izmit earthquake (Ergintav et al., 2009; Hearn et al., 2009). Stress analyses have also shown that the NAFZ weakened after the Izmit-Düzce earthquake sequence (e.g. Pinar et al., 2010; Ickrath et al., 2015). Given that deformation in the step-over is faster than before the Izmit earthquake, (micro)seismicity should also be stronger. Early post Izmit-Düzce seismicity (before early 2001) is either lacking from the catalogs due to insufficient detection capability, or the increase of seismicity occurred later due to postseismic relaxation processes such as enhanced slip rates below seismogenic depths.

The postseismic response of at least two releasing step-overs of the NAFZ, Lake Sapanca and another one in the eastern Marmara Sea, has been shown to produce substantial north-south extension following the Izmit earthquake (Ergintav et al., 2009; Hearn et al., 2009). Ergintav et al. (2009) have shown that models of postseismic slip on the main fault do not account well for the north-south extension in these two step-overs, in particular after the first three years. We further compared these step-overs by extending our temporal clustering analysis further along the NAFZ and found that the eastern Marmara Sea was also hosting clustered seismicity at the eastern termination of the Princes Islands segment (cf. Figure S7). Large earthquake location uncertainties prevented us from carrying the same detailed study but this section has been identified as an area of high  $b$ -value (Raub et al., 2017). We can hypothesize that Lake Sapanca and the eastern Marmara Sea behave similarly. In both cases, fault heterogeneities, and perhaps their stress history, could explain an hybrid seismic and aseismic slip regime (Collettini et al., 2011). As to how much slip is accommodated seismically vs. aseismically and whether the aseismic part is related to the deformation missing from the current models has to be addressed by the means of geodesy.

## 6 Concluding Remarks

We found that the patterns of seismicity have changed after the Izmit-Düzce earthquake sequence (see Section 4.1.3). We observed two regions of high seismicity, the Akyazi fault and the Lake Sapanca step-over (Section 4.1), with different statistical properties (see Sections 4.2 and 4.3). Our interpretation is that the Akyazi seismicity was driven by the high residual stresses left by the absence of co-seismic during the 1999 Izmit earthquake whereas the Lake Sapanca seismicity was caused by mixed seismic and aseismic slip on heterogeneous faults at the brittle-ductile transition (Section 5).

Our study emphasizes the important role of secondary structures in the late post-seismic stage of the NAFZ, and possibly through the interseismic phase. The structural complexity of these structures appears in stark contrast to the relatively simple co-seismic dynamics of the Izmit earthquake (rupture on almost straight and vertical fault segments). Given that the north-south extension across the Lake Sapanca step-over accelerated following the Izmit earthquake, we question whether the proposed seismic-aseismic heterogeneous slip regime could be related to this deformation (see Section 5.4). The seismicity supports, but not prove, the possibility of slow slip in the step-over. We suggested that the releasing step-over in the Marmara Sea, with similar temporal clustering and accelerated extension following the Izmit earthquake, could behave analogously to the Lake Sapanca step-over. The present study does not provide the means to relate the observed surface deformation to slip on specific faults, but it does encourage the search for slow slip on normal faults in these step-overs. Finally, our study emphasizes that slip may not always happen in well separated seismic and aseismic sections but, instead, may happen over complex, intricate unstable and stable domains.

## 7 Data and Resources

The earthquake catalog is available at the Zenodo data set repository (DOI: 10.5281/zenodo.6362973). We used the version 1.0.1 of our BPFM Python package for earthquake



689 detection and location, which is stored at <https://doi.org/10.5281/zenodo.6780316>  
690 (last accessed December 2021). The last version is maintained on Github at [https://github](https://github.com/ebeauce/Seismic_BPMF)  
691 [.com/ebeauce/Seismic\\_BPMF](https://github.com/ebeauce/Seismic_BPMF).

692 The topographic data used for the maps were taken from the Shuttle Radar Topo-  
693 graphic Mission (SRTM) 90-m database ([https://cgiarcsi.community/data/srtm-90m-](https://cgiarcsi.community/data/srtm-90m-digital-elevation-database-v4-1/)  
694 [digital-elevation-database-v4-1/](https://cgiarcsi.community/data/srtm-90m-digital-elevation-database-v4-1/), last accessed December 2021). The maps were  
695 made with the Cartopy Python library (version 0.18.0, last accessed December 2021, Met  
696 Office, 2010 - 2015). The seismic data were recorded by the temporary array DANA (DANA  
697 , 2012, DOI: [https://doi.org/10.7914/SN/YH\\_2012](https://doi.org/10.7914/SN/YH_2012)) and by the permanent KOERI sta-  
698 tions (Kandilli Observatory And Earthquake Research Institute, Boğaziçi University, 1971,  
699 DOI: <https://doi.org/10.7914/SN/KO>).

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