

**Dynamic Triggering of Earthquakes in Northeast Japan Before and After the
2011 M9.0 Tohoku-oki Earthquake**

Yuki Takeda* (1), Bogdan Enescu** (1), Masatoshi Miyazawa (2) and Like An (1)

(1) Department of Geophysics, Graduate School of Science, Kyoto University, Kyoto, Japan

(2) Disaster Prevention Research Institute, Kyoto University, Kyoto, Japan

*Now at the: Japan Weather Association, Tokyo, Japan

**Corresponding author:

Bogdan Enescu

Department of Geophysics, Graduate School of Science, Kyoto University

Kitashirakawa, Oiwake-cho, Sakyo-ku, Kyoto 606-8502, Japan

benescu@kugi.kyoto-u.ac.jp

Bogdan Enescu is also affiliated with the: National Institute for Earth Physics (NIEP), Magurele,

Romania

20 *Key Points:*

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22 We analyze earthquake triggering-ability in five swarm areas of NE Japan by estimating transient
23 stresses.

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25 Our results suggest that triggering thresholds decreased by a few kPa after the 2011 Tohoku-oki
26 earthquake.

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28 The findings may imply increased triggering-ability in volcanic regions after large megathrust
29 earthquakes.

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31 *Abstract.* Previous studies documented a relative scarcity of remote dynamic triggering of
32 earthquakes in Japan and suggested that this could be related to Japan's predominantly compressive
33 tectonic regime and/or the more frequent occurrence of large earthquakes in Japan. For example,
34 remote triggering in California, characterized by extensional tectonics, occurs at levels of stress change
35 as small as 0.1 kPa, while in Japan transient stresses of 30 kPa or larger are required. However, the
36 dynamic triggering threshold in Japan, following the 2016 Mw7.0 Kumamoto earthquake, has been
37 found to be of just a few kPa, significantly smaller than reported previously. It was proposed that a
38 decrease in the triggering threshold may have taken place in Japan, in particular at volcanic/geothermal
39 areas, after the 2011 Mw9.0 Tohoku-oki earthquake. In this study we analyze the possible change in
40 dynamic triggering conditions in five areas in northeast Japan, where swarm earthquakes have
41 occurred immediately after the Tohoku-oki earthquake. The triggering thresholds in these five areas
42 have been estimated based on the analysis of waveform recordings of 49 teleseismic earthquakes that
43 occurred between 2004 and 2020. A decrease of the triggering threshold (or triggering-ability) is
44 apparent in all but one region. However, a statistical significance K-S test could not reject, at a 5%
45 level, the null hypothesis stating that "the distribution of dynamic stress changes for triggering
46 earthquakes that occurred before and after the Tohoku-oki event is the same". We interpret the changes
47 in the triggering threshold to be related to the pore pressure increase (thus fault strength decrease) in

48 the crust, following the Tohoku-oki earthquake. Our results also indicate that dynamic triggering in
49 Japan is more common than reported previously.

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53 **Introduction**

54 An earthquake (mainshock) changes the static and/or dynamic stress field in the surrounding areas,
55 thus being able to trigger other earthquakes. Dynamically triggered earthquakes are induced by the
56 transient stress changes caused by the passage of seismic waves from the mainshock, while static
57 triggered earthquakes are the result of permanent stress changes caused by the slip on the mainshock
58 fault plane. Static stress changes may play the main role in the triggering of earthquakes in the near-
59 field (i.e., in the vicinity of the mainshock, up to distances of a few fault-lengths away from the
60 mainshock) (e.g., King et al., 1994), while in the far-field (i.e., at teleseismic distances of more than a
61 few fault-lengths from the mainshock), the dynamic stress changes induced by the mainshocks' surface
62 waves are considered to be mainly responsible for triggering other earthquakes, since static stress
63 changes attenuate significantly faster with distance (e.g., Freed, 2005).

64 The occurrence of dynamically triggered earthquakes is primarily observed during the passage of
65 the surface waves (Rayleigh or Love waves) from the mainshock, but may sometimes be significantly
66 delayed (i.e., minutes to days after the passage of mainshock surface waves, e.g., Prejean and Hill,
67 2018). For example, the earthquakes triggered dynamically by the 2016 Mw7.0 Kumamoto earthquake
68 in Japan occurred mainly when the Rayleigh or Love wave amplitude was relatively large (Enescu et
69 al., 2016). It is documented that dynamic triggering of earthquakes is more likely to occur in volcanic
70 and geothermal areas (e.g., Prejean and Hill, 2018), as the surface waves may cause changes in the

pore fluid pressure and stimulate fluid flow. Moreover, the remote triggering of earthquakes has been also observed to correlate with the tectonic regime, with more abundant triggering documented in extensional and trans-tensional tectonic regions (e.g., Prejean and Hill, 2018).

Previous studies (e.g., Harrington and Brodsky, 2006) reported that dynamic triggering of earthquakes might be relatively scarce in Japan compared to other seismic regions worldwide, like California and Greece, and suggested that this is related with Japan's predominantly compressive tectonic regime – in contrast to the extensional regime of California and Greece - and/or the more frequent occurrence of large earthquakes in Japan. Van der Elst and Brodsky (2010) and Miyazawa et al. (2021) reported that in California dynamic triggering occurs at dynamic stress change (σ_d) levels of about 0.1 kPa and larger. However, in Japan, stress changes of more than ~30 kPa are required to trigger earthquakes (Van der Elst and Brodsky, 2010). Widespread dynamic activation of seismicity in Japan following the 2011 Tohoku-oki earthquake (Miyazawa, 2011) might have been the result of large dynamic stresses (of 500 kPa and above) following the megathrust event.

Enescu et al. (2016), who investigated the dynamic triggering in Japan following the 2016 Mw7.0 Kumamoto earthquake, reported that dynamic triggering occurred at levels of a few kPa and above in regions located in central Japan, Tohoku region and Hokkaido. The clearest cases of triggering were reported in volcanic and geothermal regions, as well as along some active faults. They hypothesized that the widespread earthquake triggering observed after the Kumamoto earthquake might be related,

89 at least in part, to a decrease of dynamic stress triggering levels (or “triggering-ability” levels) in
90 northern Japan, after the 2011 Mw9.0 Tohoku-oki earthquake.

91 Here we analyze the occurrence of dynamically triggered earthquakes in five inland areas of the
92 Tohoku and Hida regions, northeast Japan, after large teleseismic events that occurred from 2004 to
93 2020, and investigate whether the dynamic triggering threshold (or the “earthquake triggering-ability”
94 level) changed after the 2011 Mw9.0 Tohoku-oki earthquake. We chose these areas because swarm
95 earthquakes have been reported to occur there after the 2011 Tohoku-oki earthquake (Okada et al.,
96 2011; Kato et al., 2013; Okada et al., 2015; Yoshida et al., 2017), which may imply that the triggering
97 environment has changed. Note that according to Okada et al. (2015) there were no notable seismic
98 swarms before the large megathrust earthquake in these investigated areas. Understanding the changes
99 of dynamic triggering levels may lead to better constraints of the crustal stress/strain state of a region
100 (Enescu et al., 2016; Prejean and Hill, 2018). Moreover, clarifying the conditions under which dynamic
101 triggering occurs is important from an earthquake hazard point of view, as it has been reported that in
102 some cases relatively large earthquakes may be dynamically triggered (Pollitz et al., 2012, Tamura and
103 Miyazawa, 2017). Note that although dynamic triggering by body-waves has been reported in the
104 literature (e.g., Miyazawa, 2012), such observations are relatively rare and more difficult to confirm.
105 Therefore, we mainly discuss in this paper the triggering of earthquakes by surface-waves, which is
106 most commonly observed.

107

108 **Data**

109 We use Hi-net, V-net, and F-net seismic networks data, provided by the National Research
110 Institute for Earth Science and Disaster Resilience (NIED) in six observation areas in northeast Japan
111 (Figure 1): a) at the border of Yamagata and Fukushima prefectures, b) southwest of Akita-
112 Komagatake volcano, c) Hida area (at the border between Nagano and Gifu prefectures), d) Iwaki
113 area (in Fukushima prefecture), e) central Yamagata area (in Yamagata prefecture) and f) northeastern
114 Miyagi prefecture. The first five regions were chosen for their post-Tohoku-oki swarm activity, as
115 explained in the **Introduction**. The sixth region is an area where no swarm earthquakes have occurred
116 and will be presented, for comparison, in the **Discussion** section. Figure 2 shows the seismic stations
117 used in each area.

118 The High-Sensitivity Seismograph Network (Hi-net, NIED) consists of borehole-installed three-
119 component high-frequency velocity seismographs, installed with a spacing of about 20 km all over
120 Japan (e.g., Aoi et al., 2020). To complement the Hi-net stations recordings, we have also used for
121 our observations of local events the V-net (NIED) high-frequency seismograph sensors that were
122 installed in relatively shallow boreholes at some volcanoes located inland Japan. Finally, the Full
123 Range Seismograph Network (F-net) is the broad-band seismic network of Japan, with well isolated

instruments installed in underground tunnels. The F-net broad-band recordings were used for computing the dynamic stress changes.

The teleseismic earthquakes investigated for potential triggering in northeast Japan occurred from 2004 (year from when the continuous Hi-net waveforms can be downloaded from the Hi-net website) to 2020 and met the following three conditions: 1) the earthquakes' depth was 100 km or less (since deeper earthquakes have less pronounced surface waves) and the earthquakes' magnitude M_w was 6.0 or larger, 2) the earthquakes occurred at distances of four fault-lengths or more from the observation areas; this threshold distance is selected because, at distances larger than a few fault lengths from the mainshock, dynamic triggering is more likely than the static one, which becomes negligible due to high-attenuation with distance - see, for example, Fan and Shearer (2016), for a discussion, and 3) the maximum dynamic stress change (σ_d) calculated from the surface waves amplitudes is 1.0 kPa or more because at smaller levels of dynamic stress the triggering is unlikely or much more difficult to observe and quantify. Because we are mostly interested in triggering due to relatively low dynamic stress changes and the remote earthquakes that produced large stress changes were characterized by much noisier local seismograms, we have excluded from our analysis any earthquakes that produced changes of σ_d larger than 50 kPa. There were 49 earthquakes that satisfied the depth condition (condition 1). Information concerning these earthquakes is shown in Table 1, while Figure 3 shows their epicentral location. Among these earthquakes, we investigated for

potential triggering the mainshocks that satisfy the distance and maximum dynamic stress change (σ_d) conditions (conditions 2 and 3) for each observation area. The number of teleseismic events analysed for possible triggering in each of the six studied regions is 44 in Yamagata-Fukushima, 47 in Hida, 48 in Iwaki, 45 in Yamagata and 42 in Miyagi. While Table 1 presents the focal mechanism of the investigated teleseismic events, the correlation between focal mechanism and local triggering is not analysed in this study.

Methods

In this section, we explain the performed analyses, which essentially consist in identifying potentially triggered earthquakes (steps 1 and 2 below), calculate the β -value to confirm statistically whether the identified earthquakes were caused by dynamic stress changes after the analyzed mainshocks (step 3), calculate dynamic stress changes (step 4) and, finally, analyse statistically the change in triggering-ability due to the 2011 Tohoku-oki earthquake (step 5).

The analysis steps are as follows:

- 1) Hi-net and V-net network waveforms were filtered at 10 to 30 Hz to detect local earthquakes, and F-net broadband waveforms were filtered at 0.01 to 0.2 Hz to observe surface waves from teleseismic events; this frequency range is confirmed from the examination of the dispersive surface waves. The F-net sensors also have a flat instrument response in this range of frequencies.

The F-net broadband waveforms were checked for saturation (clipping) before selection and, afterwards, instrument correction was applied. The N-S and E-W F-net waveforms were rotated to obtain the radial and transverse components, in the case of each teleseism. This is because Rayleigh waves are predominant on the radial and vertical components, while Love waves are predominant on the transverse component. After a careful visual inspection of waveforms, we eliminated a few stations from further analysis for which the level of background seismicity or consistent local noise was considered too high, potentially masking the triggered earthquakes. Available nearby stations were used instead of the noisier ones.

Because the hypocenter of a triggered earthquake is in general not co-located to the Hi-net (or V-net) recording station or the nearby F-net station (used for dynamic stress estimations), time-shifts (e.g., Aiken and Peng, 2014; Enescu et al., 2016) were applied to the waveforms. Such time-shifts are especially important when discussing, for example, whether the triggered earthquakes occurred during the Rayleigh wave or Love wave trains. As our focus in this study is different, the potential time bias of a few seconds that could be introduced by not shifting the waveforms does not effect the findings of this study, as we further explain in the **Results** section.

- 2) We detect local earthquakes for which the P-wave and S-wave can be clearly confirmed (i.e., their amplitudes are above a certain noise level (see below)) and their “S–P time” is up to 5 seconds (or roughly they occur within 40 km from the recording Hi-net or V-net station, using the rule of

thumb, distance (km) = $t_s - t_p$ (s) x 8 (km/s)). We do not evaluate the magnitude of the detected earthquakes because in many cases the potentially triggered events can be clearly observed at only one station, which is insufficient for a reliable magnitude determination; further bias (i.e., overestimation) of the magnitude estimation is also associated with the overlap of seismic waves of teleseismic and triggered earthquakes (Miyazawa, 2011).

We searched for local earthquakes in two time-windows - the ‘before’ and ‘after’ windows around the time of a triggering teleseismic event – to assess whether or not triggering occurred. The “before” time window spans 5 hours prior to the earliest arrival of the P-wave from a triggering teleseismic event (Aiken and Peng, 2014) for a certain observation area. In the case of the 2004 Mw7.4 Southeast off the Mie earthquake (Table 1), an Mw7.2 earthquake occurred nearby, about 5 hours before the Mie earthquake; we therefore use only the data in a time period of 2 hours before the P-wave arrival to account for the possible triggering effect of the Mw7.2 earthquake. The “after” time window is defined (i) from the arrival of a 5km/s surface wave until the arrival of a 2km/s surface wave (Aiken and Peng, 2014), or (ii) 1000 s from the arrival of the 5km/s wave when the length of the window calculated at (i) is less than 1000 s. The reason why we use window-lengths of 1000 s or larger is that for far-away, large teleseismic events the amplitude of surface waves can be relatively large for long periods. In addition, after the arrival of seismic waves from the mainshock the noise level may become higher, which may influence

visual identification of earthquakes. Therefore, we visually defined a noise window within the “after” window, which includes the largest observed noise amplitudes. We then used the largest amplitude in the noise window to define the noise level and count only those signals (P-wave and S-wave arrival of local earthquakes) with an amplitude of 1.5 times higher than the noise level (Enescu et al., 2016). In a few cases, for which high noise levels were observed before the teleseismic mainshock, the “before” window was used to define the noise level.

- 3) In the third step, we calculated the β -value statistics at the stations where earthquakes were detected during the previous step. We used the same method of Aiken and Peng (2014), which defines:

$$\beta = \frac{Na - N\left(\frac{Ta}{T}\right)}{\sqrt{N\left(\frac{Ta}{T}\right)\left(1 - \frac{Ta}{T}\right)}} \quad (1)$$

where Na is the number of earthquakes that occurred in the “after” time window, N is the total number of earthquakes that occurred in the time windows “before” and “after”, and Ta is the length of the “after” time window. T represents the total length of the time windows “before” and “after”. When $\beta \geq 1.96$, it is considered that the increase in seismic activity in the “after” time window compared to the “before” time window is statistically significant at a confidence level of 95% of the standard normal distribution (Prejean and Hill, 2018).

4) We calculated the maximum dynamic stress change (σ_d) at F-net stations nearby Hi-net (or V-net) stations where dynamic triggering of local earthquakes was investigated. From Jaeger and Cook (1979), σ_d is approximately represented by:

$$\sigma_d = \frac{Gu'}{V_{ph}} \quad (2)$$

Here, G is the shear modulus, u' is the observed peak particle velocity and V_{ph} is the phase velocity. The calculations were made assuming that G is 30 GPa, the phase velocity of the Love-wave is 4.1 km/s, and the phase velocity of the Rayleigh-wave is 3.5 km/s, which are the same as those used by Aiken and Peng (2014) and Enescu et al. (2016). The maximum dynamic stress change was estimated using the filtered F-net waveforms (as explained in Step 1); in the case of the triggered or potentially triggered earthquakes, the amplitudes in a time window surrounding the events are considered because we do not perform any time-shift, as explained previously.

5) Finally, we investigated the change in σ_d threshold before and after the 2011 Tohoku-oki earthquake for each observation area. More precisely, we are interested in observing whether the triggering due to teleseismic earthquakes takes place at smaller dynamic stress change levels following the large megathrust event than before it or, in other words, if there is a change in the triggering-ability. The Kolmogorov-Smirnov (K-S) two-sample test was also performed to investigate whether there was a statistically significant change of the triggering environment in the six regions after the 2011 Tohoku-oki earthquake, compared to the years before the megathrust

event. The K-S two sample test is one of the non-parametric tests that determine whether two samples are different by comparing the empirical Cumulative Distribution Functions (CDFs) of the two samples. Using the values of σ_d that were confirmed (by the β -value test, at Step 3) of causing dynamically triggered earthquakes after large, remote earthquakes that occurred before and after the Tohoku-oki earthquake, we made the null hypothesis that “the distribution of σ_d for triggering earthquakes that occurred before and after the Tohoku-oki earthquake is the same”. The K-S test can show whether this hypothesis can be rejected statistically or not. Note that failing to reject the null hypothesis does not imply that the null hypothesis is true, but it simply means that there is not enough evidence to conclude that the alternative hypothesis is true.

Results

Before quantifying the earthquake triggering in each of the selected areas, we will show earthquake triggering examples (Figure 4), as well as a visual definition of the time windows explained in the **Methods** section (Figure 5).

We show in Figure 4a an example of a waveform observed in the Yamagata-Fukushima border area, at the NAZH station; a triggered earthquake was detected, following the occurrence of the 2017 Mw 7.7 Commander Islands (Russia) earthquake. Figure 4b shows the arrival of seismic waves of the 2013 Mw7.7 Pakistan earthquake, recorded at the YNZH station, in the same area. Because a time-

248 shift should usually be applied to the waveform data (see **Methods** section), we estimated that this
249 time-shift ranges from about 1 s to 13 s. As the length of the time windows used for the analysis is
250 much longer than this time-shift, the bias introduced by not shifting the waveforms is not significant.
251 In Figure 4a, the local earthquake occurred when the surface wave amplitude (especially Love waves),
252 from the teleseismic event, was large. In Figure 4b as well, the earthquake occurred when the surface
253 wave amplitude (especially Rayleigh waves) was large. Note that since the “ t_s - t_p ” time of both local
254 earthquakes is 5 seconds or less, it is considered that the local earthquakes occurred near the two
255 stations, respectively. The periods of the teleseismic waves responsible for the triggering in Figure 4
256 range from around 10 to 25 s (similar to those in Enescu et al., 2016).

257 Figure 5 shows an example of defining the “before” and “after” time windows (Figure 5a), visually
258 defining the noise-window within the “after” window (Figure 5b), reading the largest noise amplitude
259 within the noise-window (Figure 5c) and counting events in the “before” and “after” windows (Figures
260 5d and e) for the case of the 2016 Mw7.0 Kumamoto earthquake, at the Hi-net station ASNH, located
261 in Hida area.

262 In the next sub-sections, we quantify the earthquake triggering in each of the investigated areas.
263 We note here that most of our detected earthquakes are not included in the standard earthquake catalogs
264 since they are very small and observed at a limited number of stations that are not sufficient for a
265 reliable location.

266

267 *The Yamagata-Fukushima border area*

268 Figure 6a shows a graph summarizing the triggering results for this area, more precisely values of
269 maximum dynamic stress change, calculated using equation (2), associated with the teleseismic,
270 potentially triggering events. The minimum value of σ_d when triggering occurred before the 2011
271 Tohoku-oki earthquake is 8.7 kPa, and the minimum value of σ_d when triggering occurred after the
272 Tohoku-oki earthquake is 1.1 kPa, indicating the triggering threshold decreased following the
273 megathrust event. We note that triggered earthquakes do not always occur if the value of σ_d exceeded
274 this minimum value; however, in this study, we call this minimum-value “threshold”. In addition, when
275 the K-S two sample test was performed using the distribution of σ_d of triggered earthquakes (i.e., red
276 triangles), the p -value is 0.085, which implies that the null hypothesis (i.e., “the distribution of σ_d for
277 triggering earthquakes before and after the Tohoku-oki earthquake is the same”) cannot be rejected at
278 the significance level of 5%.

279

280 *The southwest part of the Akita-Komagatake area*

281 Figure 6b shows the dynamic stress changes for this area. The triggering threshold, as expressed
282 by σ_d , before the Tohoku-oki earthquake is 8.7 kPa, and the threshold of σ_d after the Tohoku-oki

283 earthquake decreases to 1.5 kPa. When the K-S two sample test is performed for this area, the p -value
284 is 0.33, so the null hypothesis cannot be rejected.

285

286 *The Hida area*

287 Figure 6c shows dynamic stress changes for this area. Comparing the triggering thresholds of σ_d
288 before and after the Tohoku-oki earthquake, the thresholds are 3.8 kPa and 1.3 kPa, respectively, i.e.
289 the threshold has decreased. Note that after the 2011 Tohoku-oki earthquake three teleseismic events
290 with a value of σ_d smaller than the triggering values obtained before the megathrust event were found
291 to have dynamically triggered earthquakes (Figure 6c). However, when the K-S two sample test was
292 performed for this area, the p -value is 0.63, and the null hypothesis cannot be rejected.

293

294 *The Iwaki area*

295 The graph in Figure 6d shows the dynamic stress changes for this area. Before the Tohoku-oki
296 earthquake, there were no cases of detected earthquakes. However, after the Tohoku-oki earthquake,
297 potentially triggered earthquakes were detected for all the teleseismic mainshocks. Before the Tohoku-
298 oki earthquake, the threshold of σ_d could not be measured because there were no triggered
299 earthquakes, i.e. the maximum value of σ_d is 31.6 kPa, and therefore, the threshold should be larger
300 than 31.6 kPa. On the other hand, the threshold after the Tohoku-oki earthquake is 1.6 kPa. So one can

infer that there was a change in the triggering environment. The K-S two sample test could not be performed because earthquakes were not detected before the Tohoku-oki earthquake.

303

The Central Yamagata area

Figure 6e shows the dynamic stress changes in this area. There is only one case of triggering detected before the 2011 Tohoku-oki earthquake. Note that the triggering dynamic stress in this case was calculated to be 1.6 kPa and no other locally triggered earthquakes at larger transient stresses were detected, which may suggest that the only observed triggered earthquake may have occurred by chance. Because we cannot evaluate the triggering environment based on such limited data, we conclude that the change in the triggering environment in this area is unknown.

311

Discussion

Dynamic stress changes, σ_d , in the studied areas before and after the 2011 Tohoku-oki earthquake

Table 2 summarizes the changes in the threshold σ_d for the regions we have investigated in northeast Japan and the p -values calculated by the K-S two-sample test. The hyphens “—” represent places where σ_d values or p -values could not be measured because no local earthquake triggering occurred.

The triggering threshold before the 2011 Tohoku-oki earthquake at the Yamagata-Fukushima border, Hida area, and southwest of Akita-Komagatake area are smaller than 30 kPa, an approximate threshold for Japan, as reported by van der Elst and Brodsky (2010). The reasons for the smaller thresholds found in this study might be related to the procedure followed in this research, which focused on detecting triggered earthquakes for small dynamic stress changes, and/or the inclusion in this study of triggered earthquakes that were slightly delayed (up to about one hour after the arrival of surface waves from the teleseismic events). In addition, in four observation areas (all except Central Yamagata), the dynamic stress triggering threshold after the Tohoku-oki earthquake is around 1.5 kPa, which is smaller than the triggering cases before the Tohoku-oki earthquake.

When the K-S test was applied as described in the **Results** section, we could not confirm the change of the triggering environment at a significance level of 5%. The smallest p -value of 0.085 is found for the case of the Yamagata-Fukushima border area. Note that such a value would lead to the rejection of the null hypothesis if the significance level is relaxed (to 10%, for example). Nevertheless, the number of samples used for the p -value computation are very small: (4 , 8), (3, 6) and (7, 10) for the Yamagata-Fukushima, Akita-Komatagatake and Hida regions, respectively, where the first and second number in the parantheses indicate the number of triggering teleseismic events in each region, before and after the 2011 Tohoku-oki earthquake, respectively. Since the reliability of a statistical test depends on the number of samples available, the K-S results for such small samples should be

regarded with caution (Hermann et al., 2022). Some studies recommend using more than 50 samples for obtaining robust statistical results (e.g., Mishra et al., 2019). Note that for the Iwaki area, although the K-S test could not be applied (as explained in the Results section), the change in the triggering environment is supported by the results. Due to the relatively small number of samples, applying the K-S test to subsets of data would be affected even more by this issue.

The “after” time window used in this study might be too short to capture the delayed earthquake triggering, since dynamic triggering may occur minutes to days after the mainshock (e.g., Prejean and Hill, 2018). Aiken and Peng (2014) calculated the β -value for immediate triggering using an “after” window that spans from the arrival of the 5 km/s wave to the arrival of the 2km/s wave (similar with this study), while the β -value for delayed triggering was estimated for an “after” time window that spans from the arrival of the 5 km/s wave until five hours later. In this research, we limited our focus to the immediate triggering largely because the delayed triggering is more subtle and therefore more difficult to detect due to the mixture of triggered and background seismicity. Tests that we have made using longer “after” windows confirmed this difficulty.

We have also employed a Z-value statistical test (e.g., Pankov and Kilb, 2020) to quantify triggering, instead of the β -value statistic. The Z-value test is more conservative, resulting in fewer confirmed triggered events. However, the main findings reported in this study do not change based on this analysis.

Evaluating the cause of dynamic stress threshold changes for the occurrence of triggered earthquakes

The changes in the dynamic stress triggering thresholds might be related to the existence of pressurized crustal fluids in the investigated areas. Figure 7, modified from Brenguier et al. (2014), shows the seismic velocity susceptibility calculated from the coseismic seismic velocity changes in the crust, induced by the 2011 Tohoku-oki earthquake, with superposed continuous-line rectangles indicating the five areas investigated in this study. A large change in velocity susceptibility (red areas in Figure 7) implies the effective pressure has decreased significantly (Brenguier et al., 2014) following the 2011 Tohoku-oki earthquake. Indeed, NE Japan has experienced extensional forcing, as a result of the occurrence of the 2011 megathrust earthquake (Ozawa and Fujita, 2013), which in turn may lead to the opening of cracks, in a vertical direction, and promoting the rise of fluids from the deeper parts of the crust or from the upper mantle (Prejean and Hill, 2018). As a result, the pore fluid pressure may have increased and the fault strength decreased, which in turn may have encouraged the occurrence of earthquakes, according to the Coulomb failure criterion. Terakawa et al. (2013) documented the effect of pore fluid pressure changes on the seismic activation following the 2011 Tohoku-oki earthquake.

Note that according to our results, there was no occurrence of triggered earthquakes after the 2011 Tohoku-oki earthquake in central Yamagata (box e in Figure 7). One possibility is that despite

pore fluid pressure increase, the pressurized fluid was redistributed, and therefore, it did not lead to local earthquake triggering. Note that the data after the Tohoku-oki earthquake used to produce the results shown in Figure 7 (Brenguier et al., 2014) spans for 5 days and the earthquake swarm activity identified by Okada et al. (2015) has subsided after about 360 days (1 year) from the Tohoku-oki earthquake. These intervals are relatively short compared to our study period (2004 – 2020). Therefore, we cannot exclude that changes in the triggering environment in central Yamagata may have occurred after the 2011 Tohoku-oki earthquake, but those changes may have only lasted for a relatively short time.

In addition, although the Iwaki area is not in a velocity susceptible region (box d in Figure 7), our results suggest that the triggering condition has changed. According to Imanishi et al. (2012), there is a "branching fault", from the subduction zone, near Iwaki, to the Fukushima-Ibaraki border. After the April 11, 2011 Mw6.6 Iwaki earthquake, the outflow of hot springs from depth in this area were stimulated (Sato et al., 2020), and the branching fault might have experienced fluid movement from the subduction area (Togo et al., 2014). Therefore, we may infer that the pore fluid pressure in this area increased and triggered earthquakes were more likely to occur. Note that unlike the other four regions, the Iwaki area is not a volcanic region (see also Figure 2). This might be a factor to explain why the Iwaki area does not correspond to a red region in Figure 7; we speculate that the approach of Brenguier et al. (2014), while suitable to map pore pressure changes in volcanic areas (at

approximate depths between 0 – 10 km), may have been less sensitive to detect similar anomalies in the Iwaki area due to different depth-range characteristics of pressurized fluids, crack orientations, or a combination of these.

Investigation of an area without earthquake swarms

We have investigated so far the possible change of the triggering environment in areas where swarm earthquakes have occurred in northeast Japan, which are predominantly located within regions of increase velocity susceptibility (red regions in Figure 7). In this section, we investigate the possibility of triggered seismicity in northeastern Miyagi Prefecture (dashed box f in Figure 7). The area is seismically active and relatively close to the megathrust fault. Because this area is in a purple region in Figure 7, there is almost no change in seismic velocity susceptibility. The stations we used and the dynamic stress changes for this area are shown in Figures 2f and 6f, respectively. One locally triggered earthquake was observed following a teleseismic event that occurred before the Tohoku-oki earthquake and another locally triggered earthquake followed a teleseismic event after the 2011 megathrust earthquake. The σ_d values were 3.3 kPa and 6.4 kPa, respectively. Triggered earthquakes were not observed at dynamic stress changes, σ_d , larger than the above values, so it cannot be excluded that the observed triggering might be, in fact, a coincidence. In any case, the σ_d threshold does not decrease following the 2011 megathrust event. Since the triggered earthquakes occurred

once before and once after the Tohoku-oki earthquake, the K-S two sample test could not be performed.

Limitations of current research and analysis to be considered in future studies

From all the observations, we can infer that the change in the triggering thresholds found in this study might be related to the increase in pore fluid pressure, although the K-S statistical test could not confirm a change in the triggering environment, at a 5% significance level (or statistical testing could not be applied due to insufficient data). It is desirable to increase the number of samples in future studies, for example by increasing the number of areas where the change in triggering-ability is investigated. Newer data, after the year 2020, should also be investigated. In addition, machine learning techniques can be applied for automatic event detection; this could increase the number of detected locally triggered earthquakes and may permit to improve their characterization, including magnitude determination and precise location. Last but not least, new approaches for the quantification of earthquake remote dynamic triggering (e.g., van der Elst and Brodsky, 2010; Brodsky and van der Elst, 2014; Wang et al., 2018, Miyazawa et al., 2021) and estimation of dynamic stress changes (e.g., Gonzalez-Huizar and Velasco, 2011; Yoshida et al., 2020) should be considered.

Conclusions

We have investigated the occurrence of dynamically triggered earthquakes in several areas in northeast Japan, following 49 remote mainshocks that occurred from 2004 to 2020.

The investigation of triggering in five areas (or six if we include the northeastern Miyagi prefecture area) in northeast Japan shows that the threshold of σ_d is on the order of one digit, which is smaller than the threshold of 30 kPa for Japan but still larger than the threshold of 0.1 kPa for California reported previously (van der Elst and Brodsky, 2010; Miyazawa et al., 2021).

Before and after the Tohoku-oki earthquake, the dynamic stress change threshold for seismicity activation decreased in the areas of Yamagata-Fukushima border, southwest of Akita-Komagatake volcano, Hida, and Iwaki. The change in the triggering environment, however, cannot be confirmed at a 5% significance level, using the K-S test in the first 3 regions, while in the Iwaki area the K-S two sample test could not be performed because there were no triggered earthquakes found before the 2011 M9.0 Tohoku-oki earthquake. However, the clear triggering of this region by teleseismic events observed after the 2011 event indicates that the triggering environment has changed. Finally, in the Central Yamagata region, the data was insufficient for any inference on the change of the triggering environment.

Because the dynamic stress change threshold decreased mainly in volcanic areas, we conclude that the increase in pore fluid pressure (likely a consequence of the extensional deformation in northeast Japan caused by the 2011 Tohoku-oki earthquake) might have caused a change in the

earthquake triggering conditions. These findings may contribute to a better understanding of the complex underlying physical processes of inland seismicity activation following major megathrust earthquakes.

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Data and Resources

We use the NIED's Hi-net, F-net, and V-net networks data available at <https://doi.org/10.17598/nied.0003>, <https://doi.org/10.17598/nied.0005>, and <https://doi.org/10.17598/nied.0006>, respectively, and the JMA seismic network data, as well as the earthquake catalog data of JMA (https://www.data.jma.go.jp/eqev/data/bulletin/index_e.html), NCEDC (<https://www.ncedc.org/ncedc/catalog-search.html>) and the ANSS Comprehensive Earthquake Catalog (<https://earthquake.usgs.gov/earthquakes/search/>).

Declaration of Competing Interests

The authors declare no competing interests.

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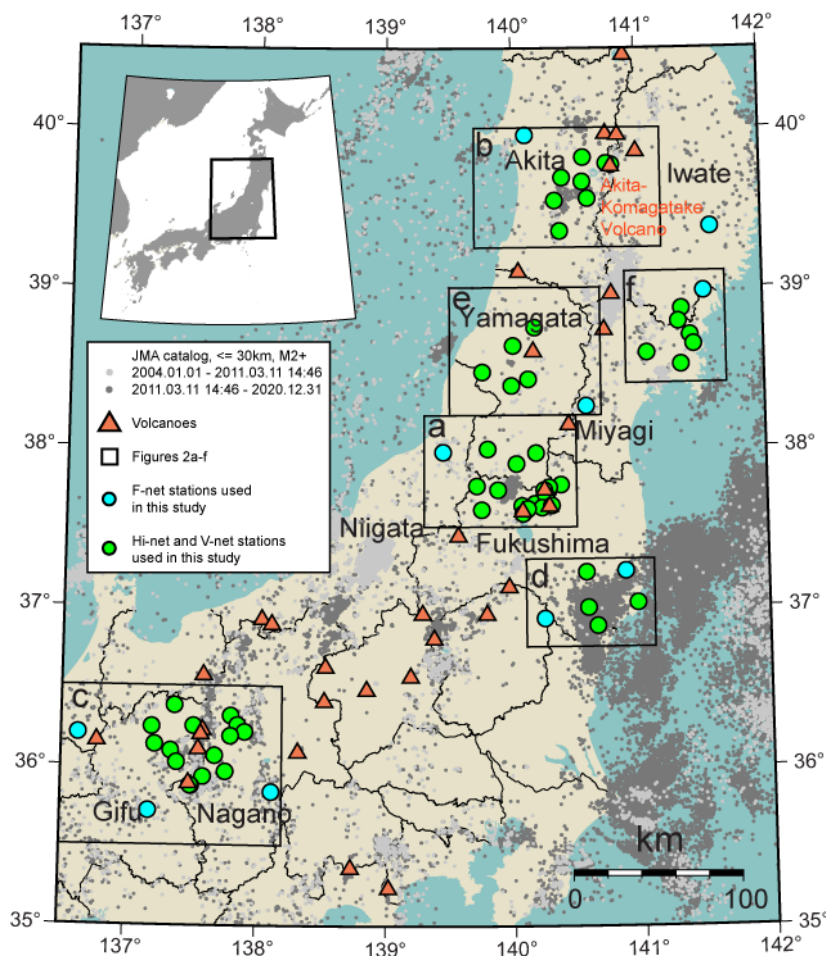
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567 **Figure 1. Study areas in NE Japan to detect possible changes in the dynamic stress triggering**
568 **levels.**

569 The black boxes indicate the study areas, and the red triangles are active volcanoes. Light grey dots
570 indicate the shallow (hypocentral depths of 30 km or less), $M \geq 2.0$ earthquakes that occurred from
571 January 1, 2004 to 2011 Tohoku-oki earthquake and the dark grey dots indicate the earthquakes
572 occurred from the 2011 Tohoku-oki earthquake to December 31, 2020. Black lines mark the prefecture
573 boundaries and the mentioned prefecture names are given in black. The green circles are Hi-net/V-net

574 stations, the light blue circles are F-net broad-band stations (a) Yamagata-Fukushima border area, (b)
575 Southwest of Akita-Komagatake area, (c) Hida area, (d) Iwaki area, (e) Central Yamagata area. (f)
576 northeastern Miyagi area. The names of seismic stations in each area are shown in Figure 2.

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587 boxes in Figure 2a. The light and dark gray dots have the same explanation as in Figure 1.

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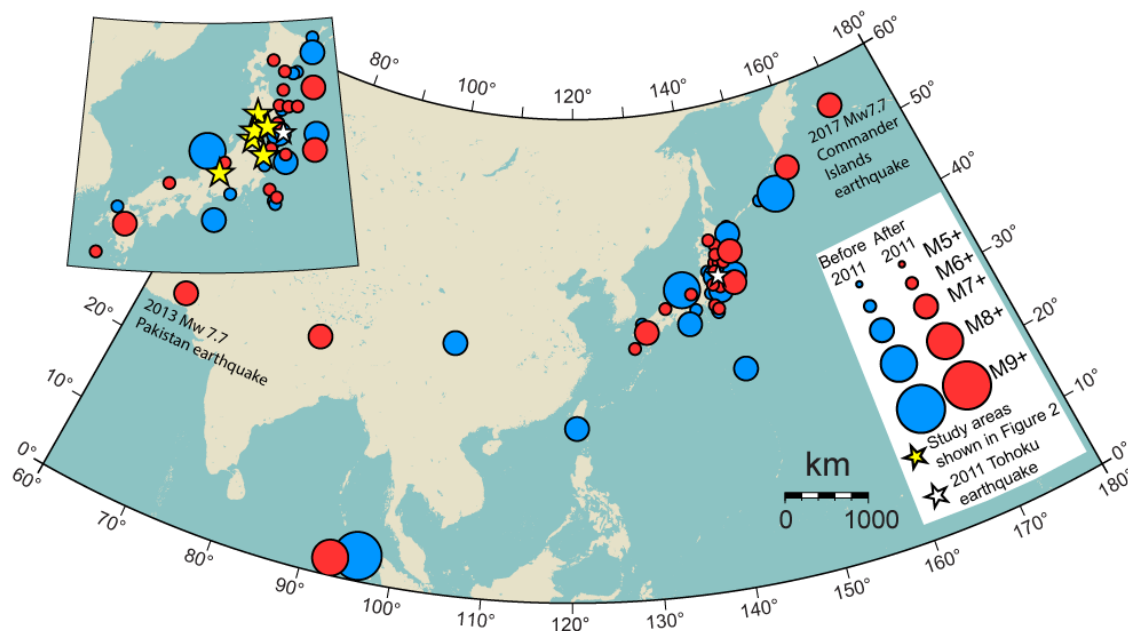


Figure 3. Map showing the potentially triggering teleseismic earthquakes.

The blue circles are the earthquakes that occurred before the 2011 Tohoku-oki earthquake (25 events), the red circles are the earthquakes that occurred after the Tohoku-oki earthquake (24 events), and the yellow stars indicate the study areas in NE Japan (shown in Figure 1) while the white star indicates the 2011 Tohoku-oki earthquake. The moment tensor solution provided by the Northern California Earthquake Data Center (NCEDC) and the Advanced National Seismic System (ANSS) for each event that occurred before and after the 2011 Tohoku-oki earthquake are listed in Table 1. Two events that were used in Figure 4 were provided with names, 2013 Mw 7.7 Pakistan earthquake and 2017 Mw 7.7 Commander Islands earthquake.

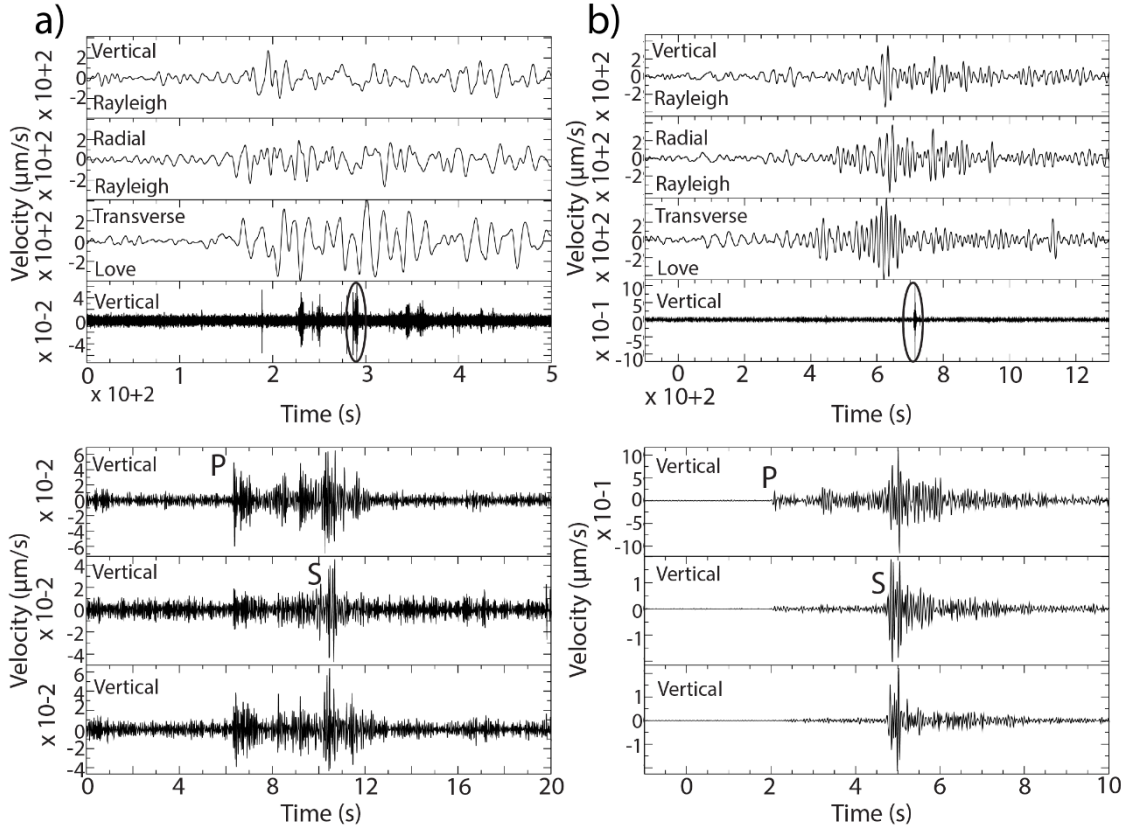
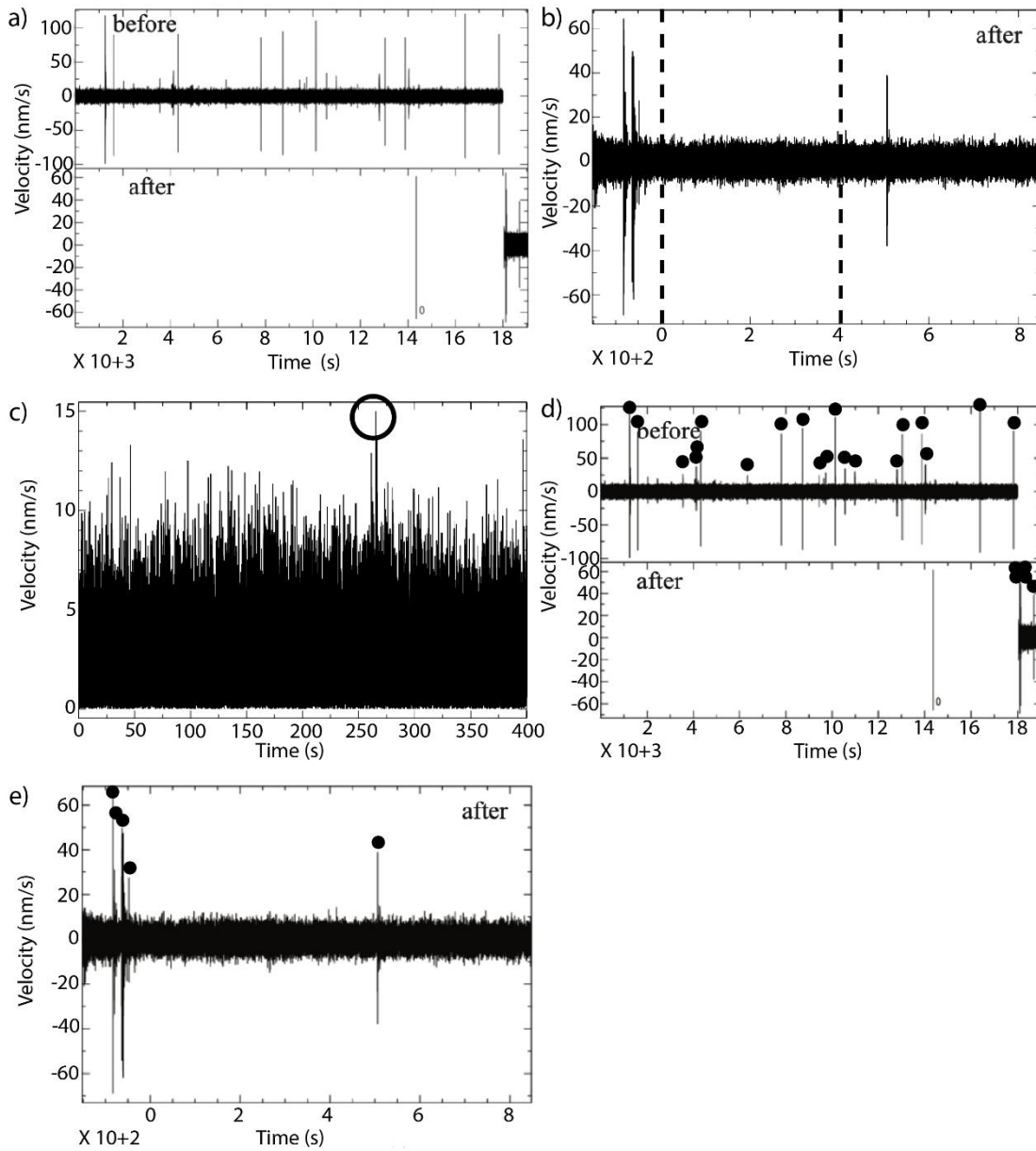


Figure 4. An example of earthquake triggering in the Yamagata-Fukushima border area.

Seismograms of the (a) 2017 Mw7.7 Commander Islands (Russia) earthquake observed at Hi-net NAZH and F-net SBTf stations and (b) 2013 Mw7.7 Pakistan earthquake observed at Hi-net YNZH and F-net SBTf stations. SBTf is closest F-net station to NAZH and YNZH. Waveforms in (a) and (b) are vertical, radial, and transverse F-net station components and vertical Hi-net station component (order from top). Insets at bottom of (a) and (b) are zoomed vertical, N-S and E-W Hi-net waveform components (order from top), showing triggered events (black circles in (a) and (b)). Horizontal axis origin time is: (a) Top: 08:42:20, July 17, 2017, (a) bottom: 08:47:00 July 17, 2017 and (b) top: 21:00:00, September 24, 2017, (b) bottom: 21:11:50, September 24, 2017. Vertical axis is velocity ($\mu\text{m/s}$). P and S letters indicate P- and S-wave onsets of the triggered events, respectively.



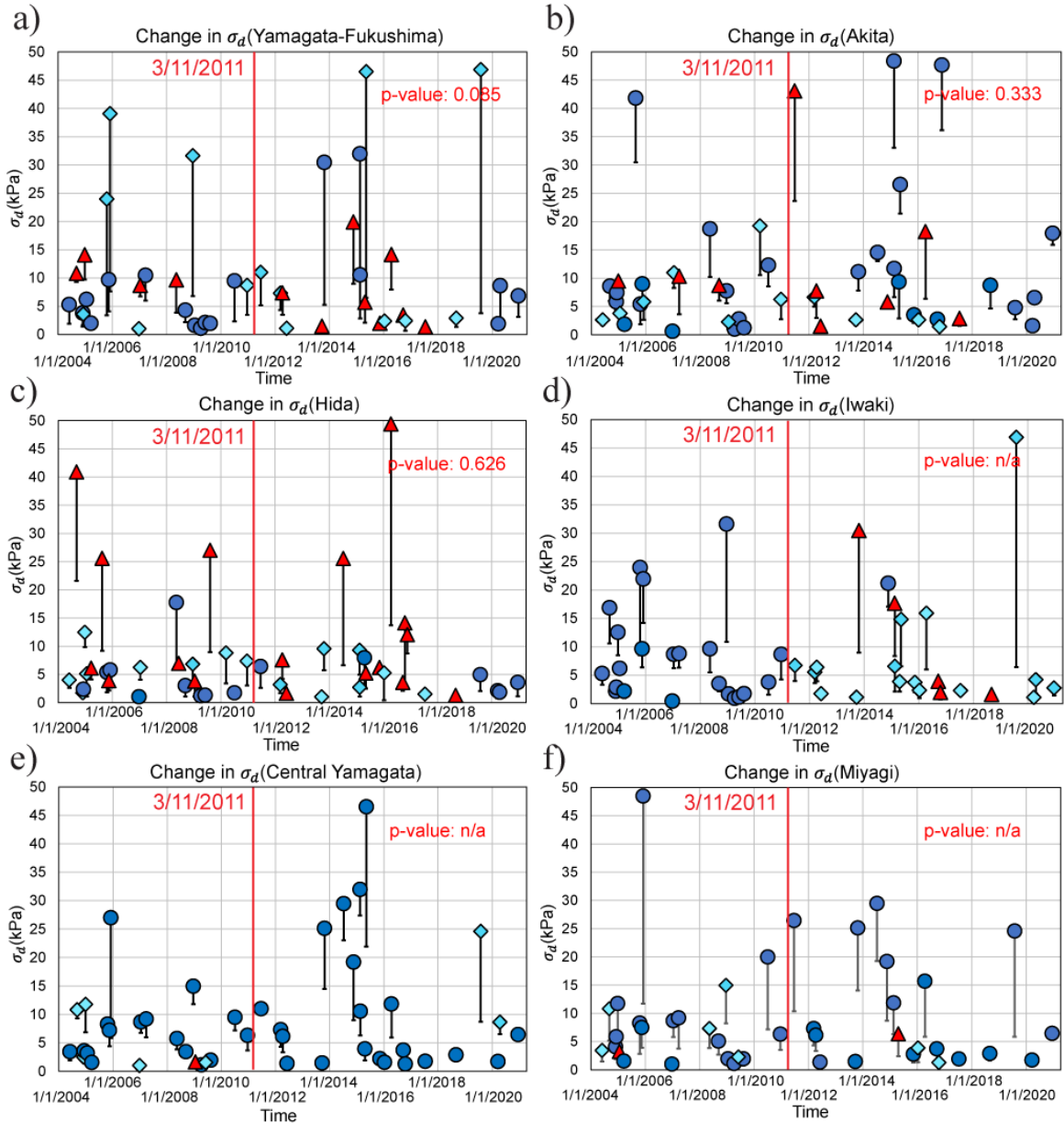
609 **Figure 5. Counting earthquakes for the β -value estimation.**

610 (a) Waveform cut-off, showing the time window “before” and “after” for the 2016 Mw7.0 Kumamoto
611 earthquake, at the Hi-net ASNH station located in Hida area. The seismogram is band-pass filtered
612 from 10 to 30 Hz; the upper side of the plot represents the “before” window and the lower side
613 represents the “after” window. The vertical axis represents velocity (nm/s) and the horizontal axis

614 represents time (s), from 20:26:39, on April 15, 2016 (JST). (b) Waveform cut-off of the “after”
615 window. The horizontal axis represents time (s), from 01:29:59, on April 16, 2016. The noise window
616 is shown between the black dotted lines. (c) The absolute values of the waveform between the black
617 dotted lines in (b). The black circle indicates the largest amplitude in the window, which is also
618 considered as the noise level for this study case (noise = 15 nm/s). (d) The count of the number of
619 earthquakes (signals above the threshold level defined as $\text{signal} \geq \text{noise} \times 1.5$). (e) Enlarged part of the
620 “after” time window in (d). At this station, there are 20 earthquakes detected in the “before” window
621 and five in the “after” window, and the β -value is 3.3, which means that dynamically triggered
622 earthquakes have occurred near this station.

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626 **Figure 6. Change of σ_d in (a) Yamagata-Fukushima border area, b) southwest of Akita-**

627 **Komagatake area, c) Hida area), d) Iwaki area, e) central Yamagata area, f) northeastern**

628 **Miyagi), before and after the 2011 Tohoku-oki earthquake.**

629 Red triangles indicate that triggered earthquakes occurred, light blue rhombs indicate that potentially
630 triggered earthquakes were detected, but β -value < 1.96 , and blue circles indicate that no earthquakes
631 were detected in the “after” window. The horizontal axis is time, the vertical axis is σ_d , and the red
632 vertical line indicates the March 11, 2011 Tohoku-oki earthquake. The bars represent the span of σ_d
633 (values calculated from the maximum amplitudes of the vertical, radial, and transverse components
634 recorded at F-net stations), with the circles representing the maximum value and the bottom of the
635 vertical line the minimum value from all those calculated. Since is reasonable to consider that an
636 earthquake was triggered locally by the maximum σ_d value, colored circle indicates the maximum
637 value. The p -values obtained using the K-S test (see text) are also indicated in the corresponding panels.
638 For the p -value calculation we used only confirmed triggered earthquakes (red triangles in the figure).
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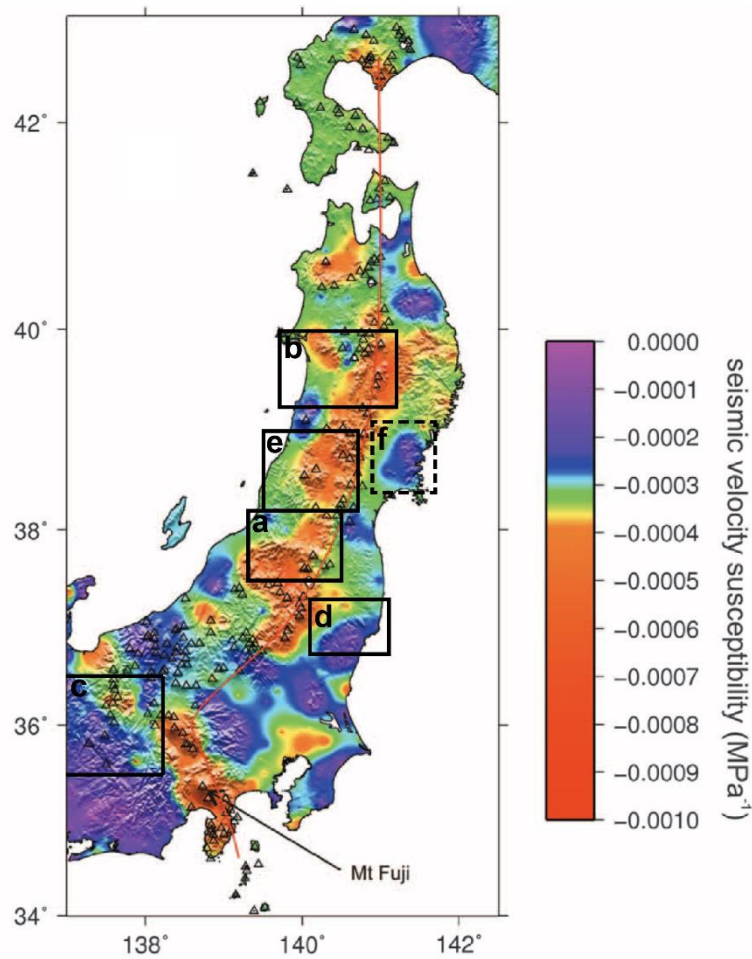


Figure 7. Seismic velocity susceptibility in NE Japan (modified from Brenguier et al., 2014; permission obtained from the authors and the journal Science).

Seismic velocity susceptibility estimated from the decrease of seismic velocity after the 2011 Tohoku-oki earthquake. Black contour rectangles indicate areas discussed in the text: (a) Yamagata-Fukushima border area, (b) Southwest of Akita-Komagatake area, (c) Hida area, (d) Iwaki area, (e) Central Yamagata area, and (f) Northeastern Miyagi area (dotted-line contour rectangle).

650 **Table 1. Teleseismic earthquakes used for investigating triggering in NE Japan.**

651 We use information from the NCEDC, from 2004 to 2012, and the ANSS from 2013. For each moment
 652 tensor solution, strike, dip and rake are provided for both nodal planes.

Date	North latitude, East longitude	Location	Mw	Moment tensor solution
2004/5/30	(34.251, 141.406)	eastern off the Kanto	6.5	 P1 183 16 86 P2 8 74 91
2004/9/5	(33.184, 137.071)	southeast off the Mie	7.4	 P1 285 56 110 P2 71 39 63
2004/11/29	(43.006, 145.119)	off the Kushiro	7.0	 P1 240 26 123 P2 25 68 75
2004/12/6	(42.9,145.228)	off the Kushiro	6.8	 P1 241 27 119 P2 29 67 76
2004/12/26	(3.295, 95.982)	off the Sumatra	9.0	 P1 336 7 114 P2 132 84 87
2005/1/19	(34.064, 141.491)	eastern off the Kanto	6.6	 P1 6 71 99 P2 160 21 66
2005/3/20	(33.807, 130.131)	western off the Fukuoka	6.6	 P1 28 76 152 P2 125 63 16
2005/8/16	(38.276,142.039)	off the Miyagi	7.2	 P1 198 21 86 P2 22 69 92
2005/10/19	(36.405,140.839)	off the ibaraki	6.3	 P1 207 22 97 P2 19 68 87

2005/11/15	(38.107, 144.896)	off the Sanriku	7.0	 P1 189 57 -91 P2 11 33 -89
2005/12/2	(38.089,142.122)	off the Miyagi	6.5	 P1 197 20 84 P2 23 70 92
2006/12/26	(21.799, 120.547)	near Taiwan	7.1	 P1 171 24 -61 P2 319 69 -102
2007/1/13	(46.243, 154.524)	eastern of the Chishima Islands	8.1	 P1 222 52 -199 P2 84 46 -58
2007/3/25	(37.336, 136.588)	off the Noto Peninsula	6.7	 P1 196 67 83 P2 32 24 105
2008/5/12	(31.002, 103.322)	Wenchuan	7.9	 P1 222 29 152 P2 338 77 64
2008/6/14	(39.030,140.881)	Iwate-Miyagi Nairiku	6.9	 P1 195 51 81 P2 29 40 101
2008/9/11	(41.892, 143.754)	off the Tokachi	6.8	 P1 235 15 116 P2 28 76 83
2008/12/20	(36.541,142.425)	eastern off the Kanto	6.3	 P1 341 61 70 P2 199 35 122
2009/1/16	(46.857,155.154)	eastern of the Chishima Islands	7.4	 P1 204 53 100 P2 7 38 77
2009/4/7	(46.049,151.548)	eastern of the Chishima Islands	6.9	 P1 230 26 105 P2 34 65 83
2009/6/5	(41.824, 143.445)	off the Tokachi	6.4	 P1 221 16 99 P2 32 74 87

2009/8/11	(34.743, 138.264)	Suruga Bay	6.2	 P1 307 47 199 P2 88 51 63
2010/3/14	(37.745,141.59)	off the Fukushima	6.5	 P1 205 19 93 P2 22 71 89
2010/7/5	(39.697,142.369)	off the Iwate	6.3	 P1 189 13 90 P2 10 77 90
2010/12/22	(26.901, 143.698)	near the Chichizima	7.4	 P1 345 57 -56 P2 113 46 -131
2011/6/23	(39.955,142.205)	off the Iwate	6.7	 P1 189 20 91 P2 8 70 90
2012/3/14	(40.887, 144.944)	off the Sanriku	7.0	 P1 262 49 -99 P2 96 42 -80
2012/4/11	(2.327, 93.063)	western off the northern Sumatra	8.6	 P1 359 72 -27 P2 98 65 -160
2012/6/6	(34.943, 141.132)	eastern off the Tiba	6.1	 P1 315 51 -177 P2 223 88 -39
2013/9/24	(26.951, 65.501)	Pakistan	7.7	 P1 228 46 8 P2 134 86 136
2013/10/26	(37.156,144.661)	off the Fukushima	7.1	 P1 180 46 -100 P2 15 45 -79
2014/7/12	(37.005,142.453)	off the Fukushima	6.5	 P1 244 24 -80 P2 20 67 -95
2014/11/22	(36.641,137.888)	northern Nagano	6.2	 P1 236 51 110 P2 26 43 67

2015/2/17	(39.8558,142.8808)	off the Sanriku	6.7	 P1 192 16 87 P2 15 74 91
2015/2/20	(39.8244,143.5871)	off the Sanriku	6.2	 P1 186 12 80 P2 16 78 92
2015/4/25	(28.231, 84.731)	Nepal	7.8	 P1 290 7 101 P2 99 83 89
2015/5/13	(38.906,142.032)	off the Miyagi	6.8	 P1 190 20 79 P2 23 71 94
2015/11/14	(31.001, 128.873)	western off the Satsuma Peninsula	6.7	 P1 190 85 -170 P2 99 80 -5
2016/1/14	(41.972, 142.781)	off the Urakawa	6.7	 P1 216 20 107 P2 18 71 84
2016/4/16	(32.791, 130.754)	Kumamoto	7.0	 P1 224 66 -152 P2 122 64 -27
2016/9/23	(34.462, 141.637)	eastern off the Kanto	6.2	 P1 5 70 100 P2 158 23 64
2016/10/21	(35.374, 133.809)	Central Tottori	6.2	 P1 338 82 -19 P2 71 71 -171
2016/11/22	(37.393,141.387)	off the Fukushima	6.9	 P1 238 42 -78 P2 42 49 -101
2017/7/18	(54.443, 168.857)	Russia Commander Islands	7.7	 P1 305 86 163 P2 37 73 4
2018/9/6	(42.686, 141.929)	middle Eastern Iburi	6.6	 P1 333 61 83 P2 167 30 102

2019/8/4	(37.759, 141.603)	off the Fukushima	6.3	 P1 179 25 66 P2 26 67 101
2020/3/25	(48.964, 157.696)	northwest Pacific	7.5	 P1 201 47 88 P2 23 43 92
2020/4/20	(38.8953,142.0049)	off the Miyagi	6.3	 P1 186 30 77 P2 20 61 97
2020/12/21	(40.8667,142.5814)	eastern off the Aomori	6.3	 P1 14 74 99 P2 162 18 59

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655 **Table 2. The change of σ_d threshold and the p-value for each of the studied regions (including**
656 **the Miyagi region).**

657 A summary of the σ_d thresholds before and after the Tohoku-oki earthquake and the p-value obtained
658 from the K-S two sample test. The hyphens “—” represent areas where the corresponding values could
659 not be computed, as explained in the text.

Location	Dynamic stress threshold before the Tohoku-Oki (kPa)	Dynamic stress threshold after the Tohoku-Oki (kPa)	p-value
Yamagata-Fukushima border	8.7kPa	1.1kPa	0.085
southwest of Akita- Komagatake	8.7kPa	1.5kPa	0.333
Hida	3.8kPa	1.3kPa	0.626
Iwaki	—kPa	1.6kPa	—
Central Yamagata	1.6kPa	—kPa	—
Miyagi	3.3kPa	6.4kPa	—

660

661

662