Dynamic stress estimation during the M5.7 remotely triggered earthquake caused by the 2016 M7.0 Kumamoto earthquake, Kyushu, Japan, using dense strong motion and high-rate GNSS data

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Declaration of Competing Interests

The authors acknowledge that there are no conflicts of interest recorded.

Abstract

The 2016 Kumamoto earthquake of M7.0 in Kyushu, Japan, was accompanied by an isolated remote earthquake of M5.7 located 80 km northeast of the mainshock. This event may be one of largest remotely triggered earthquake ever reported. Using a new analysis approach, dynamic Coulomb failure stress changes (ΔCFF) at the epicenter of the triggered earthquake were estimated from strong motion waveforms and high-rate Global Navigation Satellite System (GNSS) data, successfully revealing dynamic stress perturbation. The estimated dynamic maximum ΔCFF reached approximately +730 kPa, significantly exceeding the earthquake triggering threshold and aligning closely with result from a full waveform simulation by Miyazawa (2016). The triggered event occurred almost simultaneously with the passage of the maximum stress amplitude. The spatiotemporal distribution of dynamic ΔCFF, which was larger in the northeast and smaller in the southwest of the mainshock fault, reflected radiation and rupture directivity. These dynamic stress characteristics, along with the favorable geological conditions in the focal region, met the required criteria for triggering the event.

1. Introduction

The Mw 7.0 Kumamoto earthquake occurred at 16:25 (UTC) on April 15, 2016, in Kyushu, southwestern Japan (Fig. 1). This was a shallow crustal earthquake with a depth of 12km. The earthquake involved right-lateral strike faulting on the Hinagu and Futagawa active faults, with rupture propagating from southwest to northeast (Fig. 1, e.g., Uchide et al., 2016). Notably, the earthquake was accompanied by a remote $M_{JMA}5.7$ event in central Oita prefecture, approximately 80 km away from the Kumamoto earthquake epicenter (Fig. 1). This triggered earthquake, hereafter referred to as the “Oita earthquake”. The epicenter of this remote event was isolated from the aftershock region of the Kumamoto earthquake (e.g., Kobayashi, 2016). A clear seismic gap was observed between the aftershock area and triggered region (Fig. 1). This spatial feature suggested that it was not a typical aftershock of the M7.0 mainshock but rather a remotely triggered earthquake. The 5.7 magnitude was significantly larger than common triggering earthquakes (M<5) and was one of largest remotely triggered earthquake ever reported by instrumental observation (Hill and Prejean, 2015).

The onset time of the triggered earthquake was estimated to be 33 sec after the mainshock origin time (Nakamura and Aoi, 2017; Miyazawa, 2016). Seismic records near the epicenter of triggering event clearly indicated that the earthquake occurred during the passage of the S-wave train from mainshock (Nakamura and Aoi, 2017). This immediate response is notable because large remotely triggered earthquakes typically occur with a delay of several tens of minutes to hours (Hill et al., 1993; Parsons and Velasco, 2011; Yukutake et al., 2013).

Dynamic strain and stress characteristics are fundamental for understanding the mechanism of remote triggering. Estimation procedures for strain and stress using surface wave waveforms have been proposed and applied to several triggered events (Rubinstein et al. 2007, Miyazawa and Brodsky, 2008; Hill, 2012; Miyazawa, 2015, Enescu et al., 2016, Takeda et al., 2024, An et al., 2024). However, applying these methods to the Oita earthquake was challenging because the seismic wave train at the onset of this event might have consisted of complex body and surface waves. Miyazawa (2016) evaluated the dynamic Coulomb failure stress change (ΔCFF) at the triggered event's epicenter using full wavefield simulation, estimating the maximum ΔCFF during the S-wave train passage to be +780 kPa. This significant positive stress change strongly suggests that dynamic stress changes played a crucial role in the triggering process. It is important to confirm above simulated results with observed data.

Enescu et al. (2016) conducted a nationwide investigation across Japan into triggered earthquakes caused by the Kumamoto Earthquake, using high-sensitivity and broadband seismometer network to estimate stress changes. However, there was no mention of the largest triggered earthquake, the Oita Earthquake. There has been no previous instance of estimating the dynamic stress changes that caused the Oita Earthquake based on observational data. In Japan, dense nationwide strong motion seismograph and high-rate GNSS networks are in operation, fully recording seismic waves during earthquake triggering event. In this study, we evaluate dynamic stress change in focal regions based on observed seismic and geodetic data.

2. Data and analysis

2.1 **Strong motion seismograph data**

Seismic waves generated by the M7.0 Kumamoto mainshock were fully recorded by the KiK-net nationwide strong motion seismograph network, operated by the National Research Institute for Science and Disaster Resilience (NIED) (2019). This network is equipped with acceleration sensors featuring flat response characteristics below 30Hz, installed both at the surface and several hundred meters underground. For analysis, only waveform data from borehole sensors were used due to the stable integration of acceleration records (Hirai and Fukuwa, 2012). The acceleration seismogram data were converted into displacement waveforms by integrating the record twice, following the procedure proposed by Hirai and Fukuwa (2012). Initially, the mean DC offset before the arrival of the seismic wave train was subtracted from the entire waveform to correct for baseline offset. This seismic observation system has a flat response characteristic from 30 Hz to DC. In this study, we primarily discuss seismic waves with periods longer than 10 seconds; therefore, no instrument response correction was applied to the seismic observation system. Additionally, the sampling interval of the high-rate GNSS data is 1 second, meeting the conditions required for discussing seismic waves with periods longer than 10 seconds. Subsequently, the velocity record was calculated by integrating the acceleration seismograms. Artificial flexure signals in the velocity waveform were removed using quadratic function correction (Boore et al., 2002). Finally, displacement seismograms were derived by integrating the velocity records. These procedures were applied at the 54 KiK-net seismic stations in the Kyushu area, resulting in the acquisition of displacement seismic waveforms.

2.2 GNSS observation network data

Seismic waves from the Kumamoto earthquake were also captured by the GEONET nationwide GNSS network, operated by the Geographical Information Authority of Japan (GSI). The coordinate timeseries at each sampling epoch represent displacement waveform. The 1Hz high-sampling GNSS data collected between 15:00 and 18:00 (UTC) on April 15, 2016, were analyzed. The GNSS data was analyzed using the GIPSY/OASIS II software version 6.4 (e.g., Lichten and Border 1987), which processed the 1Hz GNSS raw phase and code data. The Kinematic Precise Point Positioning (kPPP) strategy was employed. Reference GPS satellite orbits and high-rate (5s) clock information were obtained from the Center for Orbit Determination in Europe (CODE) final products. The site coordinates were estimated every second, assuming a white noise stochastic model with a fixed process noise value of 10^{-2} km. Single receiver carrier phase ambiguities were not resolved. To remove repeated multipath noise from the 1Hz coordinate timeseries, pure sidereal filtering (236 s shift) was applied.

2.3 Waveform comparison between strong motion seismogram and GNSS coordinate timeseries

Displacement waveforms were independently derived from the KiK-net strong motion seismographs and GNSS receivers. Comparing waveform from these two different sensors validates the integration process and GNSS analysis. We selected pairs of strong motion and GNSS stations located within 1.5 km of each other for this comparison, resulting in five pairs meeting this criterion (Fig. 2a). All waveform pairs showed good agreement in both amplitude and phases (Fig. 2b). For instance, the KMMH02 KiK-net and the 0699 GNSS stations, the nearest pair to the mainshock epicenter, recorded synchronized waveforms with a maximum displacement amplitude of 40cm. Although the KGSH03 and 1090 GNSS stations, the furthest pair from the mainshock epicenter, showed the smaller displacement less than 5cm, their waveforms were still almost identical in phase and amplitude. These results confirmed that the displacement waveform generation processes for both the strong motion and GNSS data are valid and accurately represent ground displacement time series.

2.4 Dynamic stress change estimation

Dynamic stress change is a crucial factor in earthquake triggering (e.g., Hill, 2012). A method for estimating horizontal strain using the surface wave portion of seismic waves has been proposed (e.g., Gomberg and Agnew (2008), Miyazawa and Brodsky (2008)). However, the epicenter of the triggered earthquake in this case is approximately 80 km from the mainshock, which is only about 2.3 times the fault length of the mainshock (35 km), making it relatively close. Uchide (2016) mentioned that the low-frequency, large-amplitude seismic waves observed before and after the triggered earthquake are S-waves or Rayleigh waves, suggesting the difficulty of clear separation. Additionally, Miyazawa

(2016) noted that it is unclear whether the triggering of this earthquake was caused by body waves or surface waves. To address this, we applied a new strategy to estimate the dynamic stress change using displacement waveforms from three stations. Time series of stress is derived using the following procedure.

First, a triangle is formed from three adjacent stations equipped with the same sensor. Next, the horizontal strain tensor within the triangle is calculated using the two-component horizontal displacements from the three observation points. The horizontal stress tensors ($\sigma_{xx}, \sigma_{yy}, \sigma_{xy}$) are then obtained by multiplying the horizontal strain tensor (e_{xx}, e_{yy}, e_{xy}) by the elastic constant ($\lambda = \mu =$ 30GPa). The summation of shear and normal stress changes on the fault plane is a suitable indicator for shear faulting and is represented as the change in the Coulomb Failure Function (ΔCFF). Time series of stress tensor data is transformed into ΔCFF waveforms as follows:

$$\Delta\mathrm{CFF} = \Delta\tau + \mu'\Delta\sigma_n$$

Effective friction coefficient μ' is set to 0.4. The terms $\Delta\tau$ and $\Delta\sigma_n$ represent the shear stress change in the slip direction and the normal stress change on the fault plane (positive for extension), respectively. The ΔCFF is calculated using the given receiver fault parameters of strike, dip, and rake along with the horizontal stress tensor. The estimated ΔCFF values correspond to those at the ground surface, as the observation were made approximately at ground level.

3. **Results**

Fig. 3 shows the ΔCFF time series for triangles that include or are near the foci of triggered earthquake. Two triangles were evaluated for both the strong motion and GNSS stations. While receiver fault parameters are necessary for estimating ΔCFF, the focal mechanism of the triggered earthquake was not well determined due to the superimposed wave train from mainshock and triggered earthquake (Nakamura and Aoi, 2017). Miyazawa (2016) inferred the receiver fault parameters based on the focal mechanisms of the largest aftershock and its aftershock distribution. We used the parameters determined by Miyazawa (2016) (Table 1) to compare the simulation and observation results. The effects of uncertainly in the fault parameters were tested by varying the strike value from $\mathrm{N}220^{\circ}\mathrm{E}$ to $\mathrm{N}260^{\circ}\mathrm{E}$, dip from 60° to 80°, and rake from -160° to -120°.

The ΔCFF seismograms shown in Fig. 3 display a dominant wave pattern with one to one and a half cycle and a period of 10-15 s. The waveforms derived from strong motion data exhibit higher frequency content compared to those from GNSS data. This difference is likely due to the varying data sampling rate, with strong motion recorded at 100 Hz and GNSS at 1 Hz. However, this relatively short-period disturbance had a significantly smaller amplitude compared to the major longer-period wave and may be less effective in triggering earthquake.

The maximum ΔCFF in the western triangle, formed by the KiK-net strong motion sensors, reached approximately +530 kPa five seconds before the triggered earthquake occurred (Fig. 3b). At the origin time of the triggered earthquake, the ΔCFF value was around -500 kPa. The maximum positive stress value ranged from 80 kPa to 750 kPa, depending on uncertainty of the fault parameters (Fig. 3b). The highest value of 750 kPa was associated with a receiver fault characterized by a lower dip and higher rake, which aligns with the active fault geometry in focal region (Headquarters for Earthquake Research Promotion, 2020a). In the eastern triangle, the maximum value exceeded +200 kPa (Fig. 3c).

The ΔCFF waveforms from two adjacent triangles formed by the GEONET GNSS stations showed a high degree of coincidence in both amplitude and phase (Fig. 3e and f). The estimated maximum positive value exceeded +720 kPa and was synchronized with the onset time of triggered earthquake in both triangles. The effects of receiver fault uncertainly were also tested, with the maximum positive value ranging from approximately 350 kPa to 850 kPa. While the uncertainly in receiver fault parameters affected the waveform amplitude, it did not alter the phase characteristics in either the strong motion or GNSS data.

The dominant wave observed in this study is estimated to have a period of approximately 10–15 seconds, corresponding to a wavelength of about 35–53 km. To prevent the effects of spatial aliasing, it is required that the side length of the triangle used for stress estimation be less than half of the wavelength. The maximum side length of triangles formed by KiK-net observation points is 49 km, which exceeds half of the wavelength, indicating that stress estimation was affected by spatial aliasing.

The maximum side length of triangles formed by GNSS observation points is 28 km, which is approximately half of that formed by KiK-net observation points. However, it is slightly longer than half of the wavelength for a dominant wave with a 15-second period. For waves with shorter periods, the side length becomes less than half of their wavelengths. Therefore, the measurements from triangles formed by GNSS observation points are also considered to be affected by spatial aliasing. Spatial aliasing tends to result in an underestimation of the estimated stress values. Therefore, it is important to note that the actual stress values may be greater than those estimated in this study.

4. Discussion

The estimated maximum ΔCFF of over +700 kPa from GNSS data and possible over +200kPa from strong motion data in the focal region greatly exceeds the +30 kPa threshold for dynamic triggering proposed by Gomberg and Johnson (2005). van der Elst and Brodsky (2010) suggested lower triggering peak dynamic strain criterion of 0.1KPa for California and 30kPa strain for Japan. The dynamic stress changes were more than an order of magnitude larger than the static stress change caused by the Kumamoto mainshock (21 kPa) and ordinary tidal stress changes (10 kPa) (Miyazawa, 2016). These results indicate that the dynamic ΔCFF in the focal region far exceeded the necessary threshold for dynamic triggering. The maximum side length of the GNSS triangles used in this study was 28 km, which was longer than half of the observed dominant wavelength of 53 km, indicating that the measurements may be affected by spatial aliasing. Since spatial aliasing tends to cause underestimation of stress values, the result indicates that the estimated stress values significantly exceed the threshold remains valid.

The occurrence time of the triggered earthquake is indicated by a vertical line in Fig. 3. The waveforms following this point may contain the influence of seismic waves generated by the triggered earthquake. However, the maximum stress value appears either simultaneously with or before the occurrence of the triggered earthquake.

The Oita earthquake occurred in the Beppu-Haneyama active fault zone (Headquarters for Earthquake Research Promotion, 2020a). It is notable that only the Oita earthquake was selectively triggered among the many active faults in the Kyushu region. The dynamic ΔCFF at the Nishiyama and Hinagu active faults, with epicentral distances from the Kumamoto mainshock nearly the same as that of the Oita triggered event, was evaluated (Fig. 4). The receiver fault parameters were sourced from Headquarters for Earthquake Research Promotion (2020b) (Table 1). Fig. 4 compares the ΔCFF time series in three regions: the triggered Oita area and two active fault areas, using the strong motion data. The maximum amplitude in the Oita region, at 530 kPa, was more than twice that of the other two regions. These findings suggest a possible azimuthal dependence of dynamic stress changes. The rupture process analysis revealed that the Kumamoto earthquake mainshock propagated from southwest to northeast (e.g. Uchide et al., 2016). The focal region of the Oita triggered earthquake was located to the northeast, ahead of this propagation direction. These data indicate that the radiation characteristics, fault geometry, and rupture directivity selectively generated a large dynamic stress change at the Oita event hypocenter. Effective triggering due to source radiation and rupture directivity characteristics has been reported in other earthquakes, such as the 2011 Tohoku M9.0 earthquake (Miyazawa, 2011), the 1992 Landers M7.3 earthquake (Hill et al., 1993), and the 2004 Denali M7.9 earthquake (Eberhart-Phillips et al., 2003).

The focus of the triggered earthquake was located near active faults, active volcanoes and a high heat flow region (Headquarters for Earthquake Research Promotion, 2020a; Tanaka et al., 2004). These geologic and geothermal environments might provide favorable and sufficient conditions for triggering (e.g. Hill and Prejean, 2015). The onset of triggered earthquake occurred almost simultaneously with the passage of maximum amplitude (Fig. 3). However, delays in triggering are frequently observed. This immediate response might be related to the aforementioned geological circumstances.

Miyazawa (2016) estimated the ΔCFF using full wavefield simulation. It is notable that the maximum amplitude and full waveforms of the ΔCFF derived from the GNSS data, shown in Fig. 3 (e) and (f) were well consistent with the Miyazawa’ s (2016) simulation results. The ΔCFF estimated from this study corresponds to the ground surface, whereas Miyazawa (2016) presented the waveform and its amplitude at a hypocenter depth of 8.5km. These finding suggest that the analysis strategy proposed in this study is valid for stress evaluation at shallow depths.

Although the ΔCFF positive peak point from the GNSS data coincided with the onset of the triggered earthquake (Fig. 3e and f), the peak from strong motion sensors occurred approximately ~4 s earlier. The excellent agreement of displacement waveforms between the strong motion seismographs and GNSS receivers, as confirmed in Fig. 2, suggests that this difference is not due to sensor characteristics. However, the triangles formed by the stations differed in shape, size, side lengths and location between the strong motion and GNSS data due to station distribution. The GNSS triangles are smaller than those from the strong motion sensors, allowing for more stable stress estimation. These geometric parameters from triangulation may affect the phase characteristics of the strain and stress waveforms.

Based on the results in §2.3, it has been demonstrated that the displacement-converted seismic waveforms from strong motion seismographs align extremely well with the high-rate GNSS data, validating the integration method used for the strong motion waveforms in this study. The dynamic stress change estimation method employed in this study requires forming triangles using observation points. To avoid spatial aliasing, it is essential to create triangles that are as small and equilateral as possible. Generally, the density of seismic and GNSS observation points is limited. By combining strong motion and GNSS observation points, it is expected that smaller triangles can be formed, which would help mitigate spatial aliasing.

5. Conclusion

Dynamic stress change in the focal region of the M5.7 large remotely triggered earthquake, associated with the Kumamoto M7.0 earthquake, was evaluated using data from a dense strong motion seismographic network and a high-rate GNSS network. The estimated dynamic ΔCFF waveforms from independent sensors showed similarity and indicated a possible maximum dynamic stress change of +730 kPa at the focal region. This maximum stress greatly exceeded the known earthquake triggering threshold of +30 kPa. The triggered event occurred almost simultaneously with the passage of the maximum stress amplitude. The smaller ΔCFF amplitude in other regions with the same epicentral distance reflects the radiation and directivity effects from the mainshock rupture characteristics. This is consistent with the selective occurrence of the Oita triggered earthquake.

Data and Resources

Strong motion seismographic data used in this study were downloaded from the K-NET and KiK-net operated by the National Research Institute for Earth Science and Disaster Resilience, Japan (https://doi.org/10.17598/NIED.0004). The GNSS observation data observed by the Geospatial Information Authority of Japan were purchased through the Nippon GPS Data Service Corporation provided. Hypocenter data were from the Japan Meteorological Agency's published catalog (https://www.data.jma.go.jp/eqev/data/bulletin/index.html).

Acknowledgments

We express our gratitude to Dr. Kunikazu Yoshida, Dr. Michiko Shigefuji and Dr. Takahiro Shiina for their assistance with strong motion data analysis. This study was supported by the Ministry of Education, Culture, Sports, Science and Technology (MEXT) of Japan, under it's the First, Second and Third Earthquake and Volcano Hazards Observation and Research Program (Earthquake and Volcano Hazard Reduction Research), and JSPS KAKENHI(16H06298). We are grateful to anonymous reviewers for insightful comments and suggestions.

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Figure caption list

Figure 1. Aftershocks of the M7.0 Kumamoto earthquake sequence from 14 April to 31 May 2016 are shown from the JMA unified earthquake catalog with M>2.5. The open black star and open red star show the Kumamoto earthquake mainshock and the triggered Oita M5.7 earthquake, respectively. The first-motion focal mechanism for the mainshock and the triggered earthquake are from JMA and Miyazawa (2016), respectively. The black-outlined rectangles indicate the mainshock faults zones estimated from Uchide et al. (2016).

Figure 2. (a) Location map of observation sites used for waveform comparison between strong motion sensor and GNSS. Open red squares and open blue circles denote KiK-net strong motion seismograph site and GEONET GNSS sites, respectively, with site names. (b) The NS component of displacement time series for five pairs of KiK-net and GEONET site as shown in Fig. 2(a). Waveforms are arranged in order of increasing epicentral distance from top to bottom. Red and blue lines correspond to KiK-

net (100Hz) and GEONET (1Hz), respectively.

Figure 3. (a) Triangles used for the ΔCFF calculation with KiK-net strong motion data are shown. The black and red stars are epicenters of mainshock of the Kumamoto earthquake and the triggered earthquake, respectively. The focal mechanism for the triggered earthquake is taken from Miyazawa (2016). (b) Stress changes calculated in the region shown as (b) in Fig. 4(a). The black line represents changes in Coulomb failure stress (ΔCFF). The red and blue lines represent shear stress ($\Delta\tau$) and normal stress ($\Delta\sigma_n$), respectively. The shaded region indicates the range of possible ΔCFF values due to variation in fault parameters. The gray vertical bar marks the origin time of the triggered earthquake. (c) Stress changes calculated in the region shown as (c) in Fig. 4(a). (d) Triangles used for the ΔCFF calculation with GEONET GNSS data. (e) Stress changes calculated in the region shown as (e) in Fig. 4(d). (f) Stress changes calculated in the region shown as (f) in Fig. 4(d).

Figure 4. (a) Triangles used for ΔCFF calculation for three receiver faults using KiK-net strong motion data. (b) ΔCFF values calculated near the epicenter of the triggered earthquake and two other active faults. The black line represents ΔCFF for the fault plane of the triggered earthquake, as shown in Fig.3(b). The red and blue lines represent the ΔCFF for the Hinagu and the Nishiyama faults, respectively. The gray vertical bar indicates the origin time of the triggered earthquake.