Stress state inferred from $b$-value and focal mechanism distributions in the aftershock area of the 2005 West Off Fukuoka Prefecture earthquake

Keita Chiba

Author#1: Keita Chiba, Institute of Seismology and Volcanology, Faculty of Science, Kyushu University, 744 Motooka, Nishi-Ku, Fukuoka, 819-0395 Japan, kchiba@sevo.kyushu-u.ac.jp, Tel: +81 92-802-4346,
ORCID: https://orcid.org/0000-0002-2841-3371

*Corresponding author: Keita Chiba
Abstract—The spatiotemporal stress states in the aftershock region of the 2005 west off Fukuoka prefecture earthquake are examined via an analysis of the $b$-values and focal mechanism solutions. The aftershocks are aligned roughly NW–SE, with the southeastern part of the aftershock region believed to correspond to Kego Fault, which extends beneath the Fukuoka metropolitan area. This study reveals depth-dependent $b$-values in the focal region, where the $b$-values ($b = 0.7$–$1.4$) are generally higher above the mainshock depth (9.5 km) and lower ($b = 0.5$–$1.0$) at greater depths. The shallower region possesses a significant temporal increase in $b$-values, whereas a lateral $b$-value heterogeneity is observed in the deeper region. The $b$-values ($b \sim 1.0$) near the mainshock are relatively high, whereas the northwestern and southeastern edges of the deep region have lower $b$-values ($b = 0.5$–$0.7$). On the other hand, many of the focal mechanisms for the $M \geq 3.5$ events are located in the low $b$-value area of the deep region. The stress-tensor inversion results reveal a change in stress state from strike-slip to strike-slip/normal faulting. These findings imply that the stress state remains high and/or slightly decreased in the northwestern and southeastern parts of the deep region. These results and the findings of previous research on this earthquake sequence suggest that the likelihood of future large earthquakes along the southeastern part of the aftershock region should be considered relatively high.

Keywords: $b$-value, Stress state, 2005 West Off Fukuoka Prefecture Earthquake, Focal mechanism, Stress tensor inversion

1. Introduction

The 2005 West Off Fukuoka Prefecture Earthquake ($M_\text{J} 7.0$) occurred in Genkai-nada, Fukuoka Prefecture, northern Kyushu, Japan, on 20 March 2005 (e.g., Shimizu et al. 2006; Uehira et al. 2006; This study reveals...
The hypocenter was located at 130.1616°E, 33.7434°N, and 9.5 km depth, with a best-fit focal mechanism solution showing left-lateral strike slip with a tension axis aligned N23°W–S23°E (Uehira et al. 2006). Many aftershocks occurred at 1–16 km depth, and were distributed along a 25-km-long NW–SE-trending linear feature (Shimizu et al. 2006; Uehira et al. 2006). The largest aftershock (Mj 5.8) occurred near the southeastern end of the aftershock region on 20 April 2005, followed by additional aftershock activity to the southeast (e.g., Uehira et al. 2006). Numerous research studies have also investigated the coseismic slip distribution of the mainshock, spatial distribution of static stress drops, postseismic deformation, and attenuation structure in the focal region of the 2005 West Off Fukuoka Prefecture Earthquake (e.g., Asano and Iwata 2006; Horikawa 2006; Iio et al. 2006; Matsumoto, S. et al. 2009; Nakao et al. 2006; Nishimura et al. 2006). Iio et al. (2006) have provided important details for hazard assessments in the focal region via an investigation of the static stress drop distribution of the aftershocks using the waveforms from temporary seismic stations; they highlighted the possibility of a large stress concentration around the southeastern end of the aftershock area, which includes the largest aftershock.

Many active faults exist in northern Kyushu district (The Research Group for Active Faults of Japan 1991). The southeastern extension of the aftershock region is believed to correspond to the northwestern edge of Kego Fault, which is known to be active fault (e.g., Okamura et al. 2009; Uehira et al. 2006). Kego Fault runs beneath the Fukuoka metropolitan area, with a population of about 1.5 million. Okamura et al. (2009) used acoustic exploration and sediment cores to investigate the paleoseismicity of the Kego Fault and estimated that the latest two major earthquakes occurred 4,500–4,000 yBP and 8,500–6,500 yBP, respectively. These short recurrence intervals are a stark contrast to the 15,500-year recurrence interval estimated by Shimoyama et al. (2005). These geophysical and geological results
suggest that the next major earthquake might impact segments of Kego Fault that have not previously ruptured, which run through the central part of Fukuoka City. The National Institute of Advanced Industrial Science and Technology (AIST) (2005) calculated the Coulomb stress change induced by the 2005 mainshock, and found a positive stress change of up to $0.1-0.5$ MPa around the southeastern and northwestern ends of the aftershock area. The probability of a $M = 7.0$ earthquake in the next 30 years on the Kego Fault (counted from 2005) increased about 7% due to the 2005 mainshock (AIST 2005). The metropolitan area would be devastated if such a major earthquake were to occur. However, details of the stress state have not been evaluated in and around the focal region, although the area remains seismically active.

Here the $b$-values of the Gutenberg–Richter Law (Gutenberg and Richter 1944), which are indicators of the stress state (e.g., Scholz 1968, 2015), are investigated in the aftershock region of the 2005 West Off Fukuoka Prefecture Earthquake using a seismic catalog that was compiled from data recorded after the 2005 mainshock. Recently, the stress-dependent characteristics of the $b$-values were used for real-time seismic hazard assessment. One of the main concerns after a large earthquake is whether or not a stronger subsequent event will occur. Gulia and Wiemer (2019) examined time series of $b$-values in many aftershock sequences, and found that a larger subsequent event is more likely to occur if the $b$-value decreases substantially. Real-time monitoring of the $b$-value is thus considered useful to estimate whether a larger subsequent event is likely. Furthermore, the focal mechanisms are inverted to obtain the stress tensors and assess the present stress states in the study area. This inversion does not determine the magnitude of the deviatoric stress at seismogenic depths (e.g. Gephart and Forsyth 1984; Michael 1984, 1987). However, the orientation of the maximum horizontal stress axis relative to the fault strike direction is useful to infer how the stress field evolves with time (e.g., Hardebeck and Hauksson...
The style of faulting, which depends on the directions of the principal stress axes, also indicates the absolute stress level from the Coulomb failure criterion (e.g., Jaeger and Cook 1979). This study aims to clarify the spatiotemporal stress states as well as resolve highly-stressed areas with high likelihoods of nucleating large earthquakes in the focal region, through unified analysis of $b$-values and stress tensor inversions.

2. Methods and data

2.1. $b$-values

The empirical relationship between the earthquake frequency and magnitude distribution, which is known as the Gutenberg–Richter Law, is a power law of the form:

$$\log N = a - bM,$$

where $N(M)$ is the cumulative number of earthquakes with magnitude equal to or greater than $M$, and $a$ and $b$ are constants (Ishimoto and Iida 1939, Gutenberg and Richter 1944). The constant $b$ is termed the $b$-value, and represents the ratio of small to large earthquakes. The $b$-value depends on various factors, including the stress state, strength, material heterogeneity, and thermal gradient (e.g., Mogi 1962; Scholz 1968, 2015; Warren and Latham 1970; Wyss 1973; Urbancic et al. 1992), and varies between 0.5 and 1.5 for different tectonic settings, with an average value of 1.0 (Frohlich and Davis 1993). The stress-dependent characteristics of the $b$-values are among the most widely studied earthquake parameters in the literature (e.g., Scholz 1968, 2015; Schorlemmer et al. 2004; Schorlemmer and Wiemer 2005).
differential stress generally increases as the $b$-value decreases, and vice versa. Many studies have examined the stress states in various tectonic settings using spatiotemporal $b$-value distributions (e.g., Chiba 2019, 2020; Ghosh et al. 2008; Nanjo and Yoshida 2018; Nanjo et al. 2016, 2019; Schorlemmer et al. 2004; Schorlemmer and Wiemer 2005; Tormann et al. 2015). The highly stressed areas in any tectonic setting, such as regions with a large coseismic slip and large slip deficit rate, are usually characterized by low $b$-values. For example, Schorlemmer and Wiemer (2005) examined the $b$-value distribution in the focal area prior to the 2004 M$_{w}$ 6.0 Parkfield Earthquake on the San Andreas Fault (California, USA), and found a close spatial correspondence between low $b$-values and large coseismic slip during the mainshock. Ghosh et al. (2008) found a negative correlation between the $b$-value distribution and plate locking inferred from geodetic estimates of interface locking along the Cocos Plate near Nicoya Peninsula, Costa Rica. Nanjo et al. (2016) analyzed the $b$-value distribution in the focal region of the 2016 Kumamoto Earthquake, Japan, prior to the mainshock, and discovered that a zone near the mainshock corresponds to a low $b$-value region. These examples verify that spatiotemporal $b$-value analysis is an important and useful tool for seismic hazard assessment in regions with a high likelihood of large earthquakes.

2.2. $b$-value data and computational procedure

The unified earthquake catalog maintained by the Japan Meteorological Agency (JMA catalog) was used for the $b$-value calculations in this study, which included 15,381 events that were recorded between 12:20 on 20 April 2005 (2005.3 in decimal year) to 23:59 on 31 January 2020 (JST) in the region bounded by 130.0–130.5°E, 33.55–33.85°N, and 0.0–20.0 km depth. Data within one month of
the mainshock were excluded since the events recorded during this period were highly heterogeneous and the JMA catalog was incomplete. Fig. 1 shows hypocenters of the $M \geq 0.5$ events; this choice of magnitude threshold is explained below.

ZMAP (Wiemer 2001) was employed for the seismicity analysis of this dataset, including the $b$-value computations. An estimation of the magnitude of completeness $M_c$, which is the minimum magnitude in the earthquake catalog that satisfies the power law in Eq. (1), is crucial for the $b$-value analysis. The detection limits of the JMA catalog have improved significantly since 2000; $M_c$ is typically 0.0–1.0 in Kyushu district (Nanjo et al. 2010). The maximum curvature method (Wiemer and Wyss 2000) was employed to investigate temporal changes in $M_c$ in the study area using overlapping 500-event windows and a 20-event step. Generally, $M_c < 0.5$ throughout the analysis period (Fig. 2a).

Gridding methods were further applied by calculating the spatial distribution of $M_c$ using the maximum curvature method (Wiemer and Wyss 2000). The $M_c$ cross-section for events within 5 km of line A-B in Fig. 1 was calculated using 0.5 km grid spacing and a constant sampling radius of 2.5 km, which represents the optimal radius value for $b$-value computation of cross-sections, as explained below. $M_c$ varies across the study area, with $M_c \geq 0.5$ in the northwestern subregion (Fig. 2b). Two-step screenings were therefore employed to choose $M_c$ for the $b$-value calculations. First, a total of 9,324 $M \geq 0.5$ events were used for further analysis, selected based on spatiotemporal $M_c$ distributions. Next, local $M_c$ values were recalculated at each grid point for $M \geq 0.5$ events, using this catalog and the maximum curvature method (Wiemer and Wyss 2000); this was a precursory step to $b$-value calculations due to observed spatial variations in $M_c$. Events with magnitudes below each new $M_c$ value were discarded. In addition, the maximum curvature method is known to often underestimate $M_c$ by 0.2 on average (Woessner and Wiemer 2005); therefore, an $M_c$ correction of +0.2 was applied. The analysis
window was divided into two periods to investigate the temporal changes in $b$-values: the approximately 8-month period after the mainshock (period 1: 4335 events, 12:20 JST on 20 April 2005–31 December 2005) and the 2006–2020 period (period 2: 4989 events, 1 January 2006–31 January 2020). However, it is insufficient to evaluate temporal changes in $b$-values using only arbitrarily-selected time windows. An additional analysis, which divided the analysis window into three periods, was performed to investigate the temporal changes in $b$-values in detail: period a1 was 12:20 JST on 20 April 2005–23:59 on 31 August 2005, and had 3298 events; period a2 was 1 September 2005–31 December 2007, with 3017 events; period a3 was 1 January 2008–31 January 2020, with 3009 events. This analysis was only used for $b$-value calculations because there are too few available focal mechanisms to constrain stress tensor inversions in three temporal subwindows.

Gridding methods were applied by calculating the maximum-likelihood $b$-values. Spatial $b$-value distributions were constructed to produce plan-view maps and cross-sections using gridding intervals of 0.005° and 0.5 km for the plan-view maps and cross-sections, respectively. Determining the optimal sampling radii for the $b$-value calculations at each grid point is both necessary and non-trivial. Fewer grid points match the required minimum number of events to constrain the $b$-values when the sampling radii are small. Conversely, the resolution of the regional $b$-value heterogeneities is reduced when the sampling radii are too large. The optimal sampling radius is the largest radius that still resolves $b$-value heterogeneities in detail (Schorlemmer et al. 2004). The optimal sampling radius was determined to be 2.5 km for the plan-view maps and cross-sections in this study after exploring a broad range of sampling radii. The $b$-value was then calculated at each grid point via the maximum likelihood method (Aki 1965):

$$b = \log e / (M_{\text{mean}} - M_0).$$

(2)
where \( M_{\text{mean}} \) is the mean magnitude and \( M_0 = M_c - 0.05 \) for uniform 0.1 magnitude bins. The corresponding \( b \)-values were not calculated if a grid node had <50 events within a 2.5-km radius.

The heterogeneities in the \( b \)-value distributions were quantitatively evaluated using the \( p \)-test (Utsu 1992), which is defined as:

\[
p = e^{(-dA/2-2)},
\]

where \( dA = -2N\ln(N) + 2N_1\ln(N_1 + N_2b_1/b_2) + 2N_2\ln(N_1b_2/b_1 + N_2) - 2 \), \( N_1 \) and \( N_2 \) are the numbers of events within the volumes to be compared, and \( N = N_1 + N_2 \). Smaller \( p \)-values indicate more significant \( b \)-value heterogeneities. Previous studies have found that \( b \)-value heterogeneity is considered statistically significant if \( \log p \leq -1.3 \) (Schorlemmer et al. 2004; Utsu 1992, 1999).

2.3. Stress tensor inversion data and computational procedure

The focal mechanisms that were used for the stress tensor inversion were estimated from F-net data, which were recorded by the permanent seismic stations operated by the National Research Institute for Earth Science and Disaster Resilience (NIED). Twenty-three focal mechanisms (13 events during period 1 and 10 events during period 2), each with a variance reduction of >60% and trace data from >3 stations, were inverted to determine the best-fitting normalized stress tensor for each period. The stress tensor inversion was performed using the MSATSI software (Martinez-Garzon et al. 2014), a MATLAB package based on Hardebeck and Michael (2006). This method uses the linearized inversion scheme of Michael (1987), which is based on the Wallace-Bott hypothesis (Wallace 1951; Bott 1959), whereby an earthquake slip’s vector is parallel to the resolved shear stress on the fault plane, and a
Damped inversion method is used to avoid the apparent spatial variability resulting from dividing the analytical region into small subregions. The stress tensor inversion has four unknown parameters: the three principal stress axes, $\sigma_1, \sigma_2, \sigma_3$, and the stress ratio:

$$R = (\sigma_1 - \sigma_2) / (\sigma_1 - \sigma_3).$$

which represents the relative stress magnitude. The uncertainties in the results were evaluated via 500 bootstrap resamplings (Michael 1987). The number of events used to estimate the uncertainty in each stress parameter was $\geq 20$ in each trial. As mentioned in Section 2.2, stress tensor inversions were not performed for each of the three temporal subwindows because there were too few available focal mechanisms.

3. Results

The hypocenter distribution consists of a roughly NW–SE-oriented linear feature (Fig. 1). There is no clear migration of events between periods 1 and 2. The $b$-value cross-sections are first calculated for the events within 5.0 km of transect A–B in Fig. 1 during both analysis periods (Fig. 3) based on the NW–SE alignment of the seismicity; this resolves depth-dependent characteristics with a transition at the mainshock depth (9.5 km). The shallower region has relatively high $b$-values ($b = 0.7$–1.4) during both periods, whereas the $b$-values ($b = 0.5$–1.0) are lower in the deeper region (Fig. 3). The reduction in $b$-value ($b = 0.5$–0.7) is especially significant in the northwestern and southeastern parts of the deep region. Conversely, the characteristics of the $b$-values near the mainshock hypocenter differ slightly from the overall feature. The $b$-values ($b \sim 1.0$) at 12.0–16.0 km depth (e.g., regions 1 and 5 in Fig. 3) are higher than those ($b = 0.5$–0.7) in the surrounding deeper region (e.g., regions 2 and 6) during
both periods; the $b$-values ($b \sim 0.7$) are relatively low at 6.0–10.0 km depth and near the mainshock hypocenter (region 3) during period 1 compared to those ($b = 1.0–1.4$) in the surrounding shallow areas (e.g., region 4); and the $b$-values increase ($b \sim 1.0$) in the areas near the mainshock (e.g., regions 5 and 7) during period 2 (Fig. 3). Namely, low $b$-value areas at shallow depths near the mainshock are present during period 1 but disappear during period 2. The observed spatial $b$-value heterogeneities for each period are statistically significant at the 99% confidence level based on $p$-value tests (Utsu 1992), as shown in Fig. 4.

Plan-view maps of the depth-dependent $b$-value characteristics are shown in Fig. 5 for the shallow (0.0–9.5 km depth) and deep (9.5–20.0 km) regions during both periods, with the mainshock depth serving as the transition between these two regions. The basic features are the same as those in the cross-sections: the shallow region possesses relatively high $b$-values (Fig. 5a, c), whereas the $b$-values in the deep region are low (Fig. 5b, d). This depth dependence is especially noticeable during period 2. The $b$-values in the northwestern and southeastern parts of the aftershock region in the deep region are lower than those near the mainshock (Fig. 5b, d). The $b$-values in the shallow region near the mainshock are lower than those in the surrounding shallow area during period 1 (Fig. 5a). Noticeable temporal increases in the $b$-values are also observed in the shallow region (Fig. 5a, c), whereas significant temporal variations are not found at greater depths (Fig. 5b, d).

The differences in $b$-values between the two analyzed periods ($\Delta b = b_{period2} - b_{period1}$) were calculated separately for the shallow and deep regions (Fig. 6a, c) to qualitatively evaluate the temporal changes in $b$-values; the corresponding $p$-values indicate that these changes are statistically significant. The $b$-values also exhibit statistically significant temporal increases ($p \leq 0.0498$) in many parts of the shallow region, including the area near the mainshock (Fig. 6b). Areas with significant
temporal $b$-value changes are not widely distributed at greater depths (Fig. 6d).

Results for the three temporal subwindows indicate no clear migration of events between them (Fig. S1). The general patterns in spatial and temporal distributions of $b$-values are the same as when the analysis window is divided into two periods (Figs. S2–S4). However, using three subwindows captures some temporal evolution of the $b$-values that was not clear when only two periods were used. The $b$-value distributions in periods a1 and a3 are roughly similar to those of both subwindows in the two-period analysis (Figs. 3, 5, S2, and S3); the $b$-value distribution for period a2 appears to show a transition between periods 1 and 2. Most of the low $b$-values in the shallow region near the mainshock disappear during period a2 (Figs. S2b and S3c). The observed low $b$-value anomaly may be characteristic of an early stage during period 1. In addition, significant temporal increases in $b$-values between periods a1 and a2 were found in many parts of the shallow region (Fig. S4a, b). $b$-values remained low at greater depths during all three periods; areas with significant temporal $b$-value changes were not widely distributed (Fig. S4c, d, g, h).

The relationship between the $b$-values and focal mechanisms was then investigated (Figs. 5 and S3). The focal mechanisms for the $M \geq 3.5$ events, which are relatively large earthquakes for the analyzed region, were located in the low $b$-value ($b = 0.5–0.7$) area of the deep region; this finding is consistent with the definition of the $b$-value that moderate to large earthquakes are likely to occur in regions with low $b$-value. A stress tensor inversion was performed for each period using these focal mechanisms. The stress state obtained for period 1 was dominated by strike-slip fault type (Fig. 7), with $R = 0.51$ and a E–W-oriented maximum principal stress. A counterclockwise rotation of the maximum principal stress and changes in the stress ratio were determined for period 2, with a decrease in the stress ratio to $R = 0.29$ and a ENE–WSW-oriented maximum principal stress. The rotation of the maximum
principal stress was statistically significant at the 95% confidence level. This relatively low stress ratio represents a stress state that is indicative of strike-slip/normal faulting (Fig. 7). This observation implies that the stress state changed at greater depths, even though there were no significant temporal changes in $b$-values.

4. Discussion

The spatiotemporal $b$-value distributions are highly heterogeneous in the aftershock region of the 2005 West Off Fukuoka Prefecture Earthquake, with the stress-tensor inversion results indicating that the stress state below 9.5 km depth changed from a strike-slip-dominated to strike-slip/normal faulting regime. These results imply that the stress state is heterogeneous across the study area. Possible causes of the observed stress heterogeneity in the analyzed region are discussed in this section by comparing the results obtained in this study with those from previous studies. The areas with a high likelihood of nucleating large earthquakes are also evaluated.

The key seismic characteristics found in this study include the overall depth-dependence of the $b$-value distributions. The shallow region possesses relatively high $b$-values, whereas the $b$-values in the deep region are low (Figs. 3 and 5). Uehira et al. (2006) reported that the aftershock distribution in the central part of the study area was aligned on different fault planes that split at the mainshock depth, with a difference of ~10° between their respective strikes and dips. Furthermore, they noticed that the mainshock focal mechanism estimated from $P$-wave first-motion polarities was consistent with the deeper fault plane, whereas the moment tensor solution estimated from broadband seismic waveform data (Matsumoto, T. et al. 2006) was consistent with the shallower fault plane. The large coseismic slip of the
mainshock was mainly observed in the southeastern part of the aftershock region, near the mainshock, at shallower depths (e.g., Asano and Iwata 2006; Horikawa 2006). Uehira et al. (2006) suggested that the mainshock rupture initiated at the hypocenter and propagated downward, then ruptured the lower plane and propagated upward, with this last rupture segment releasing most of the associated seismic moment. These findings suggest that more accumulated stress was released at shallower depths. The high $b$-values in the shallow region can therefore be interpreted as a low-stress state associated with significant stress release due to large coseismic slip. Nakao et al. (2006) found notable postseismic deformation on the coseismic fault at <3.0 km depth, with the postseismic slip on the fault accounting for up to ~13% of the coseismic slip. The $b$-values in the shallow region are especially high in the uppermost layers of Fig. 3. These findings are all consistent with a low-stress state in the shallow region.

A comparison of the presented results with a $b$-value analysis in the focal region of the 2016 Kumamoto Earthquake is useful for understanding the potential causes of the $b$-value heterogeneities found in this study. The 16 April 2016 M$\text{$_J$}$ 7.3 Kumamoto Earthquake occurred in the Futagawa-Hinagu fault zone, Kumamoto Prefecture, central Kyusyu, Japan. Nanjo et al. (2019) analyzed the $b$-value distribution after the mainshock in the focal region, and found a highly stressed area with low $b$-values at the southern end of the causative faults. Their interpretation was that the heterogeneous $b$-values in the focal region were due to postseismic deformation. The focal region was located at the boundary between an afterslip-dominated region and viscoelastic deformation-dominated region (Nanjo et al. 2019). They suggested that there was a local increase in the stresses at the boundary between regimes dominated by different postseismic deformation processes based on the aftershock decay law of Utsu (1961) and a seismicity rate model inferred from the stressing history (Dieterich 1994). Compared with the Kumamoto earthquake, the postseismic deformation associated with the Fukuoka earthquake was restricted to very
shallow depths in the focal region (e.g., Nakao et al. 2006). The significant temporal changes in \( b \)-values in the present study are mainly found in the shallow region, where afterslip occurred. Nakao et al. (2006) detected postseismic deformation using GPS time-series data from 21 March to 27 June 2005. Significant postseismic deformation was clearly seen around the aftershock area during the analysis period of the present study (after 2005-04-20). In addition, the Geospatial Information Authority of Japan (GSI) (2006) reported that postseismic deformation ended in November 2005. These observations imply that significant postseismic deformation occurred during the early stage of period 1, which approximately corresponds to period a1 (2005-04-20–2005-08-31) in the three-subwindow analysis of the present study. Evidently, significant temporal increases in \( b \)-values between periods a1 and a2 were observable in many parts of the shallow region (Fig. S4a, b). Therefore, the postseismic deformation following a mainshock may play an important role in the temporal evolution of the \( b \)-values. Furthermore, the increase in \( b \)-values near the mainshock in the shallow region, which approximately corresponds to regions 3 and 7 in Fig. 3, may also be ascribed to a decrease in the effective normal stress due to increased pore fluid pressure, which will be discussed below. These findings indicate that it is unlikely that the shallow region is the potential location of another large earthquake in the near future since a significant amount of stress was released by coseismic and postseismic slip.

The heterogeneity of the \( b \)-value distribution in the deeper region is then discussed. The southeastern extension of the aftershock region appears to correspond to the northwestern edge of Kego Fault, which extends beneath the Fukuoka metropolitan area (e.g., Okamura et al. 2009; Uehira et al. 2006). An evaluation of the stress state in the deeper region is therefore considered crucial for the assessment of possible large earthquakes in the future. Matsumoto, S. et al. (2009) investigated the seismic-wave attenuation \( (Q^{-1}) \) structure in and around the aftershock region of the 2005 mainshock, and
discovered a highly heterogeneous $Q^{-1}$ structure in the focal region. Areas with large slip and high
aftershock activity were located in low $Q^{-1}$ regions, whereas high a $Q^{-1}$ region was found along the
southeastern edge of the aftershock region, near the largest aftershock. Matsumoto, S. et al. (2009)
suggested that the low and high $Q^{-1}$ regions corresponded to relatively high and low fault strengths,
respectively. The largest aftershock has the same strike angle as Kego Fault (Uehira et al. 2006).

Matsumoto, S. et al. (2009) posited a segment boundary between the mainshock fault and Kego Fault,
where the fault strength is weak, based on the difference between the strike angles of the mainshock and
largest aftershock. Iio et al. (2006) investigated the spatial distribution of the static stress drops of the
Fukuoka earthquake aftershocks from 23 March to 31 May 2005, and suggested the possibility of more
stress being concentrated along the southeastern edge of the aftershock region. The $b$-values along the
southeastern edge of the deeper region are consistent with the results of Iio et al. (2006), and were low
throughout the entire analysis period (Figs. 3 and 5). These findings suggest that the southeastern extent
of the aftershock region, including the largest aftershock, comprises a highly stressed area with a
relatively low fault strength. This area may therefore have a high probability of experiencing a large
earthquake in the future compared to northwestern part of the aftershock region, even though both
subregions have low $b$-values. The potential causes of the relatively high $b$-values below the mainshock
are considered based on the findings of other geophysical studies. Wang and Zhao (2006) investigated the
3-D seismic velocity and Poisson’s ratio structures in the epicentral area of the 2005 mainshock and
found that the mainshock was located at the boundary between high- and low-velocity regions in the
upper crust, with the lower crust side corresponding to a low-velocity, high Poisson’s ratio region. They
interpreted the low-velocity region beneath the mainshock to possess fluids associated with mantle
upwelling due to the opening of Okinawa Trough. This finding suggests that fluids in and around a
mainshock source area may play an important role in earthquake generation (Wang and Zhao 2006). The relatively high $b$-values in the region beneath the mainshock are therefore considered to be generated by a reduction in the effective normal stress due to abundant fluids in the focal region. The geophysical interpretation of Wang and Zhao (2006) is helpful for inferring the cause of the temporal increase in $b$-values in the shallow region near the mainshock (regions 3 and 7 in Fig. 3). Fluid injection from the deeper region may continue if this increase in $b$-values occurs due to an increase in pore pressure related to the fluids beneath the focal region.

Most of the focal mechanisms included in the stress tensor inversion were located in the northwestern and southeastern parts of the aftershock region, which possess low $b$-value areas in the deeper region (Fig. 5). Unfortunately, the data are too sparse to invert for a stress tensor at each edge since only 13 and 10 events were available for periods 1 and 2, respectively. It should therefore be noted that the results of the stress tensor inversion show average stress states in both the deep northwestern and southeastern subregions during each period; despite this limitation, the inversion results revealed a change in the stress field that was characterized by counterclockwise rotation of the maximum principal stress axis and a change in $R$ from strike-slip to strike-slip/normal faulting. This change in $R$ is useful to infer the differential stress near both edges of the aftershock region. The Coulomb failure criterion states that normal faulting events sustain the lowest differential stresses, strike-slip events are intermediate, and reverse faulting events have the highest stresses, assuming that the vertical stress is equal to the overburden pressure (e.g., Jaeger and Cook 1979). The observed change in $R$ implies that the average differential stress decreased slightly in these subregions during the study period. Areas with the low $b$-values appeared to decrease further during period 2 based on the $b$-value cross-sections and plan-view maps (Figs. 3b and 6c, respectively), although the decreases in the low $b$-value areas were not necessarily
statistically significant at many nodes (Fig. 6d). The angle between the strike of the fault plane and maximum principal stress axis direction is used to infer the shear strength in the focal region (e.g., Hardebeck and Hauksson 2001). The maximum principal stress axis direction should be aligned \( \sim 30^\circ \) relative to the fault plane for a strong fault, which is controlled by the frictional force (Byerlee 1978). On the other hand, if the fault is weak, then the direction is predicted to be at a higher angle. The hypocenter distribution consists of a roughly NW–SE-oriented linear feature (Fig. 1). Therefore, the angle of \( \sigma_1 \) relative to the mainshock fault strike was rotated, and increased, from period 1 to period 2 (Figs. 7a, c).

This finding also supports the observation of a slight decrease in average differential stress. The potential cause of the slight decrease in the differential stress in the region with low \( b \)-values is unknown because no significant postseismic deformation was observed in the deeper region (Nakao et al. 2006). However, a region with significant low \( b \)-values does not necessarily lead to a future large earthquake. For example, Tormann et al. (2015) analyzed the \( b \)-values along the subducting Pacific Plate off Japan over a period that included the 2003 \( M_w \) 8.3 Tokachi-oki and 2011 \( M_w \) 9.0 Tohoku-oki earthquakes, and found that the former occurred within an area of persistent low \( b \)-values off Hokkaido, but not in a region of significant low \( b \)-values. On the other hand, the latter occurred within a subregion of distinct low \( b \)-values in the subsequent high-slip area of the mainshock (Tormann et al. 2015). The causes of these \( b \)-value differences remain unknown. Whether a significant decrease in \( b \)-value prior to a major earthquake is an ubiquitous phenomenon is a question that should be approached with caution, as some previous studies of major earthquakes found precursory increases in \( b \)-values (e.g., Smith 1981).

This point will be addressed in future research, where it can be considered in the context of other geophysical studies. It is noteworthy, however, that El-Isa and Eaton (2014) reviewed previous research on spatial and temporal variations in \( b \)-values and found that the vast majority of previous papers support
a decrease in $b$-value before a major earthquake. A significant and/or moderate decrease in $b$-value is therefore useful information when drawing inferences about stress accumulation. Here the deeper region along the southeastern edge of the aftershock region is considered to have a high likelihood of triggering future large earthquakes due to the low $b$-values; however, there is a possibility that the differential stress decreased slightly in the focal region. Temporal changes in the $b$-values that were related to the presence of fluids were also found near the mainshock hypocenter. An increase in $b$-values related to a reduction in effective stress, which is due to the presence of abundant fluids, may also be an important precursor to the next large earthquake. Wiemer et al. (1998) reported that increased pore pressure lowered effective stress and increased $b$-value, a phenomenon associated with the onset of an earthquake swarm at Long Valley Caldera, California. The possibility that seismic activity increases due to abundant fluids in the focal region cannot be ruled out.

5. Conclusions

This study examined the spatiotemporal $b$-value distributions in the aftershock region of the 2005 West Off Fukuoka Prefecture Earthquake. The $b$-value distributions possess distinct depth-dependent characteristics, with the mainshock depth forming the transition zone. Relatively high $b$-values ($b = 0.7–1.4$) are distributed throughout the shallow region, whereas the $b$-values ($b = 0.5–1.0$) in the deep region are low. The high $b$-values in the shallow region are ascribed to significant stress release due to large coseismic and postseismic slip. Although the $b$-values are generally low in the deep region, the area below the mainshock possesses relatively high $b$-values ($b \sim 1.0$). This anomaly may reflect a decrease in the effective stress due to an increase in pore fluid pressure that is caused by abundant fluids.
in the focal region. The $b$-values ($b = 0.5–0.7$) below 9.5 km depth in the northwestern and southeastern parts of the aftershock region remained low throughout the analysis period, although a stress tensor inversion using focal mechanisms revealed a slight decrease in the average differential stress in these areas. It is possible that the southeastern part of the aftershock region possesses a high likelihood of hosting future large earthquakes.

Acknowledgments

The author used data from the Japan Meteorological Agency (JMA) unified earthquake catalog (https://www.data.jma.go.jp/svd/eqev/data/bulletin/index.html). The ZMAP MATLAB software package (Wiemer 2001) (http://www.seismo.ethz.ch/static/stat_2010_website/stat-website-pre2010/www.earthquake.ethz.ch/software/zmap.html) was used for the $b$-value analysis. The figures were prepared using the Generic Mapping Tools software package (Wessel et al. 2013). This study was partially funded by Tokio Marine Kagami Memorial Foundation, Japan (EAKF320500). The author thanks the editor Carla Braitenberg and two anonymous reviewers for helping to improve the manuscript.

REFERENCES

Aki, K. (1965). Maximum likelihood estimate of $b$ in the formula $\log N = a – bM$ and its confidence limits. *Bulletin of Earthquake Research Institute of the University of Tokyo, 43*, 237–239.

Asano, K., & Iwata, T. (2006). Source process and near-source ground motions of the 2005 West Off Fukuoka Prefecture Earthquake. *Earth, Planets and Space, 58*(1), 93-98.
Bott, M. H. P. (1959). The mechanics of oblique slip faulting. *Geological Magazine*, 96, 109-117.

Byerlee, J. (1978). Friction of rock. *Pure and Applied Geophysics, 116*, 615–4626.

Chiba, K. (2019). Spatial and temporal distributions of b-values related to long-term slow-slip and low-frequency earthquakes in the Bungo Channel and Hyuga-nada regions, Japan. *Tectonophysics, 757*, 1-9. https://doi.org/10.1016/j.tecto.2019.02.021.

Chiba, K. (2020). Stress state along the western Nankai Trough subduction zone inferred from b-values, long-term slow slip events, and low frequency earthquakes. *Earth, Planets and Space, 72*(3).

https://doi.org/10.1186/s40623-020-1130-7.

Dieterich, J. (1994). A constitutive law for rate of earthquake production and its application to earthquake clustering. *Journal of Geophysical Research, 99*(B2), 2601-2618. https://doi.org/10.1029/93JB02581.

El-Isa, Z.H. & Eaton, D.W. (2014). Spatiotemporal variations in the b-value of earthquake magnitude-frequency distributions: Classification and causes. *Tectonophysics, 615-616*, 1-11.

https://doi.org/10.1016/j.tecto.2013.12.001.

Frohlich, C., & Davis, S. (1993). Teleseismic b values; or, much ado about 1.0. *Journal of Geophysical Research, 98*(B1), 631-644.

Geospatial Information Authority of Japan. (2006). Crustal deformation and a fault mode of the West–off Fukuoka Prefecture Earthquake in 2005. https://www.gsi.go.jp/common/000025591.pdf (in Japanese).

Gephart, J. W., & Forsyth, D. W. (1984). An improved method for determining the regional stress tensor using earthquake focal mechanism data: Application to the San Fernando earthquake sequence. *Journal of Geophysical Research, 89*, 9,305-9,320.

Ghosh, A., Newman, A.V., Thomas, A.M., & Farmer, G.T. (2008). Interface locking along the
subduction megathrust from $b$-value mapping near Nicoya Peninsula, Costa Rica. *Geophysical Research Letters*, 35, L01301. https://doi.org/10.1029/2007GL031617.

Gulia, L., & Wiemer, S. (2019). Real-time discrimination of earthquake foreshocks and aftershocks. *Nature*, 574, 193-199. https://doi.org/10.1038/s41586-019-1606-4.

Gutenberg, B., & Richter, C. F. (1944). Frequency of earthquakes in California. *Bulletin of the Seismological Society of America*, 34, 185-188.

Hardebeck, J.L., & Hauksson, E. (2001). Crustal stress field in southern California and its implications for fault mechanics. *Journal of Geophysical Research*, 106(B10), 21,859-21.882.

Hardebeck, J. L., & Michael, A. J. (2006). Damped regional-scale stress inversions: Methodology and examples for southern California and the Coalinga aftershock sequence. *Journal of Geophysical Research*, 111, B1310. https://doi.org/10.1029/2005JB004144.

Horikawa, H. (2006). Rupture process of the 2005 West Off Fukuoka Prefecture, Japan, earthquake. *Earth, Planets and Space*, 58, 87-92.

Iio, Y., Katao, H., Ueno, T., Enescu, B., Hirano, N., Okada, T., Uchida, N., Matsumoto, S., Matsuhima, T., Uehira, K., & Shimizu, H. (2006). Spatial distribution of seismic stress drops for aftershocks of the 2005 West Off Fukuoka Prefecture Earthquake. *Earth, Planets and Space*, 58, 1611-1615.

Ishimoto, M., & Iida, K. (1939). Observations sur les seisms enregistré par le microseismograph construit demiement. *Bulletin of the Earthquake Research Institute, Tokyo Imperial University*, 17, 443–478.

Jaeger, J., & Cook, N.G.W. (1979). Fundamental of Rock Mechanics, 3rd ed, Chapman and Hall, London.

Martinez-Garzon, P., Kwiatek, G., Ickrath, M., & Bohnhoff, M. (2014). MSATSI: A MATLAB package for stress inversion combining solid classic methodology, a new simplified user-handling and a
visualization tool. Seismological Research Letters, 85(4). https://doi.org/10.1785/0220130189.

Matsumoto, S., Uehira, K., Watanabe, A., Goto, K., Iio, Y., Hirata, N., Okada, T., Takahashi, H., Shimizu, H., Shinohara, M., & Kanazawa, T. (2009). High resolution $Q^{-1}$ estimation based on extension of coda normalization method and its application to $P$-wave attenuation structure in the aftershock area of the 2005 West Off Fukuoka Prefecture Earthquake (M 7.0). Geophysical Journal International, 179, 1039-1054. https://doi.org/10.1111/j.1365-246X.2009.04313.x.

Matsumoto, T., Ito, Y., Matsubayashi, H., & Sekiguchi, S. (2006). Spatial distribution of $F$-net moment tensors of the 2005 West Off Fukuoka Prefecture Earthquake determined by the extended method of the NIED F-Net routine. Earth Planets Space, 58, 63-67.

Michael, A. J. (1984). Determination of stress from slip data: Faults and folds. Journal of Geophysical Research, 89, 11,517-11,526.

Michael, A.J. (1987). Stress rotation during the Coalinga aftershock sequence. Journal of Geophysical Research, 92, 7,963-7,979.

Mogi, K. (1962). Magnitude-Frequency relation for elastic shocks accompanying fractures of various materials and some related problems in earthquakes. Bulletin of the Earthquake Research Institute, University of Tokyo, 40, 831–853.

Nakao, S., Takahashi, H., Matsushima, T., Kohno, Y., & Ichiyanagi, M. (2006). Postseismic deformation following the 2005 West Off Fukuoka Prefecture Earthquake (M7.0) derived by GPS observation. Earth Planets Space, 58, 1617-1620.

Nanjo, K. Z., Ishibe, T., Tsuruoka, D., Schorlemmer, D., Ishigaki, Y., & Hirata, N. (2010). Analysis of the Completeness Magnitude and Seismic Network Coverage of Japan. Bulletin of the Seismological Society of America, 100(6), 3261-3268. https://doi.org/10.1785/0120100077.
Nanjo, K. Z., Izutsu, J., Orihara, Y., Furuse, N., Togo, S., Nitta, H., Okada, T., Tanaka, R., Kamogawa, M., & Nagao, T. (2016). Seismicity prior to the 2016 Kumamoto earthquakes. *Earth, Planets and Space*, 68, 187. https://doi.org/10.1186/s40623-016-0558-2.

Nanjo, K.Z., & Yoshida, A. (2018). A b map implying the first eastern rupture of the Nankai Trough earthquakes. *Nature Communications*, 9, 1117. https://doi.org/10.1038/s41467-018-03514-3.

Nanjo, K.Z., Izutsu, J., Orihara, Y., Kamogawa, Y., & Nagano, T. (2019). Changes in seismicity pattern due to the 2016 Kumamoto Earthquakes identify a highly stressed area on the Hinagu fault zone. *Geophysical Research Letters*, 46(16). https://doi.org/10.1029/2019GL083463.

Scholz, C. H. (1968). The frequency-magnitude relation of microfracturing in rock and its relation to earthquakes. *Bulletin of the Seismological Society of America*, 58, 399–415.

Scholz, C. H. (2015). On the stress dependence of the earthquake b value. *Geophysical Research Letters*, 42, 1399–1402. https://doi.org/10.1002/2014GL062863.

Schorlemmer, D., Wiemer, S., & Wyss, M. (2004). Earthquake statistics at Parkfield: 1. Stationarity of b
values. *Journal of Geophysical Research, 109*, B12307. https://doi.org/10.1029/2004JB003234.

Schorlemmer, D., & Wiemer, S. (2005). Microseismicity data forecast rupture area. *Nature, 434*, 1086.

https://doi.org/10.1038/4341086a.

Shimizu, H., Takahashi, T., Okada, T., Kanazawa, T., Iio, Y., Miyamachi, H., Matsushima, T., Ichiyanagi, N., Uchida, N., Iwasaki, T., Katao, H., Goto, K., Matsumoto, S., Hirata, N., Nakao, S., Uehira, K., Shinohara, M., Yakiwara, H., Kame, T., Urabe, T., Matsuwo, N., Yamada, T., Watanabe, A., Nakahigashi, K., Enescu, B., Uchida, K., Hashimoto, S., Hirano, S., Yagi, T., Kohno, Y., Ueno, T., Saito, M., & Hori, M. (2006). Aftershock seismicity and fault structure of the 2005 West Off Fukuoka Prefecture Earthquake (MJMA 7.0) derived from urgent joint observations. *Earth, Planets and Space, 58*, 1599-1604.

Shimoyama, S., Iso, N., Matsuda, T., Ichihara, T., Chida, N., Okamura, M., Mogi, T., Suzuki, S., Ochiai, H., Nagasawa, S., Imanishi, H., Kawabata, F., Yakabe, H., Ooteki, M., & Matsuura, K. (2005). Trenching study at Yaukuin site across the Kego fault, Fukuoka City, West Japan. *Active Fault Research, 25*, 117-128.

Smith, W.D. (1981). The b-value as an earthquake precursor, *Nature, 289*, 136-139.

https://doi.org/10.1038/289136a0.

The Research Group for Active Faults of Japan. (1991). Active Faults in Japan (revised edition), 488 pp, *University of Tokyo Press*, (in Japanese).

Tormann, T., Enescu, B., Woessener, J., & Wiemer, S. (2015). Randomness of megathrust earthquakes implied by rapid stress recovery after the Japan earthquake. *Nature Geoscience, 8*(2), 152-158.

https://doi.org/10.1038/ngeo2343.

Uehira et al. (2006). Precise aftershock distribution of the 2005 West Off Fukuoka Prefecture Earthquake
(Mj = 7.0) by using a dense onshore and offshore seismic network. Earth, Planets and Space, 58, 1605-1610.

Urbancic, T.I., Trifu, C. I., Long, J. M., & Young, R. P. (1992). Space-time correlations of b values with stress release. Pure and Applied Geophysics, 139(3), 449–462.

Utsu, T. (1961). A statistical study on the occurrence of aftershocks. Geophysics, 30, 521-605.

Utsu, T. (1992). On seismicity, in: Report of Cooperative Research of the Institute of Statistical Mathematics. The Institute of Statistical Mathematics Tokyo, 34, 139–157.

Utsu, T. (1999). Representation and analysis of the earthquake size distribution: A historical review and some approaches. Pure and Applied Geophysics, 155, 509-535.

Wallace, R. E. (1951). Geometry of shearing stress and relationship to faulting. The Journal of Geology, 59, 111-130.

Wang, Z., & Zhao, D. (2006). Seismic evidence for the influence of fluids on the 2005 west off Fukuoka prefecture earthquake in southwest Japan. Physics of the Earth and Planetary Interiors, 155, 313-324.

Warren, N. W., & Latham, G. V. (1970). An experimental study of thermally induced microfracturing and its relation to volcanic seismicity. Journal of Geophysical Research, 75, 4455–4464.

Wiemer, S. (2001). A software package to analyze seismicity: ZMAP. Seismological Research Letters, 72(3), 373–382.

Wiemer, S., & Wyss, M. (2000). Minimum magnitude of completeness in earthquake catalogs: examples
from Alaska, the Western United States, and Japan. Bulletin of the Seismological Society of America, 90, 859–869.

Wiemer, S., McNutt, S.R., & Wyss, M. (1998). Temporal and three-dimensional spatial analyses of the frequency-magnitude distribution near Long Valley Caldera, California. Geophysical Journal International, 134, 409-421. https://doi.org/10.1406/j.1365-246x.1998.00561.x.

Woessner, J., & Wiemer, S. (2005). Assessing the quality of earthquake catalogues: estimating the magnitude of completeness and its uncertainty. Bulletin of the Seismological Society of America, 95(2), 684-498. https://doi.org/10.1785/0120040007.

Wyss, M. (1973). Towards a physical understanding of the earthquake frequency distribution. Geophysical Journal of the Royal Astronomical Society, 31, 341–359. https://doi.org/10.1111/j.1365-246X.1973.tb06506.x.
Figure Captions

Figure 1

Left: Epicentral distribution of the \( M \geq 0.5 \) earthquakes. The red and green circles represent the period 1, (12:20 JST on 20 April 2005 to 23:59 on 31 December 2005) and period 2 (00:00 JST on 1 January 2006 to 23:59 on 31 January 2020) events, respectively. The events are scaled by their magnitudes. The yellow and blue stars represent the hypocenters of the mainshock and largest aftershock (\( M_j 5.8, 06:11 \) JST on 20 April 2005), respectively. The yellow and blue focal mechanisms are those for the mainshock and largest aftershock, respectively. The red lines indicate active faults (The Research Group for Active Faults of Japan 1991). Line A–B shows the location of the \( b \)-value cross-section in Fig. 3. The red rectangle represents the region used for \( b \)-value analysis in Figs. 5 and 6. The inset map shows the analyzed region with respect to western Japan. Right: Cross-section of the hypocenters within 5 km of A–B.

Figure 2

(a) \( M_c \) as a function of time in the area bordered by the red rectangle in Fig. 1. The dashed gray lines represent the standard deviation of the calculated \( M_c \) values. The maximum curvature method (Wiemer and Wyss 2000) was employed to investigate the \( M_c \) distribution in the study area using overlapping 500-event windows and a 20-event step. (b) Spatial distribution of \( M_c \) estimated using the maximum curvature method (Wiemer and Wyss 2000) with 0.5 km grid spacing and a constant sampling radius of 2.5 km. The yellow and blue stars represent the mainshock and the largest aftershock, respectively.
Figure 3

Cross-sections of the $b$-values in (a) period 1 and (b) period 2. The yellow and blue stars represent the mainshock and the largest aftershock, respectively. The black circles show the regions where the $p$-value test was performed. The small black circles represent the events that occurred in each region.

Figure 4

$p$-values for the regions with different $b$-values: (a–d) period 1; (e, f) period 2. $N$ represents the number of events that occurred each region.

Figure 5

(a, b) Plan-view $b$-value maps of the shallow (0.0–9.5 km depth) and deep (9.5–20.0 km depth) regions during period 1. (c, d) Plan-view $b$-value maps of the shallow and deep regions during period 2. The focal mechanisms from the F-net data are superimposed on the $b$-value maps for the appropriate depth range and time window. The sizes of the focal mechanisms are proportional to their moment magnitudes. The other symbols and meanings are the same as in Fig. 1.

Figure 6

(a) Difference in $b$-values between the first and second periods ($\Delta b = b_{\text{period2}} - b_{\text{period1}}$) in the shallow region. (b) $p$-values estimated from $\Delta b$ in (a). (c) Difference in $b$-values in the deep region. (d) $p$-values estimated from $\Delta b$ in (c). The other symbols and meanings are the same as in Fig. 1.
Stress tensor inversion results for (a) period 1 and (b) period 2. The large square, triangle, and circle show the best orientations of the maximum, intermediate, and minimum principal stresses, respectively. The small red, green, and blue circles represent the 95% confidence limits of the $\sigma_1$, $\sigma_2$, and $\sigma_3$ axis orientations, respectively. The right figures show frequencies of the stress ratios $R$ within the 95% confidence limits.