Tidal triggering of earthquakes in the subducting Philippine Sea plate beneath the locked zone of the plate interface in the Tokai region, Japan

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Abstract

We found a characteristic space–time pattern of the tidal triggering effect on earthquake occurrence in the subducting Philippine Sea plate beneath the locked zone of the plate interface in the Tokai region, central Japan, where a large interplate earthquake may be impending. We measured the correlation between the Earth tide and earthquake occurrence using microearthquakes that took place in the Philippine Sea plate for about two decades. For each event, we assigned the tidal phase angle at the origin time by theoretically calculating the tidal shear stress on the fault plane. Based on the distribution of the tidal phase angles, we statistically tested whether they concentrate near some particular angle or not by using Schuster’s test. In this test, the result is evaluated by p-value, which represents the significance level to reject the null hypothesis that earthquakes occur randomly irrespective of the tidal phase angle. As a result of analysis, no correlation was found for the data set including all the earthquakes. However, we found a systematic pattern in the temporal variation of the tidal effect; the p-value significantly decreased preceding the occurrence of Mz 4.5 earthquakes, and it recovered a high level afterwards. We note that those Mz ≥ 4.5 earthquakes were considerably larger than the normal background seismicity in the study area. The frequency distribution of tidal phase angles in the pre-event period exhibited a peak at the phase angle where the tidal shear stress is at its maximum to accelerate the fault slip. This indicates that the observed small p-value is a physical consequence of the tidal effect. We also found a distinctive feature in the spatial distribution of p-values. The small p-values appeared just beneath the strongly coupled portion of the plate interface, as inferred from the seismicity rate change in the past few years.

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1. Introduction

The Earth tide produces periodic stress variations of the order of 10^3 Pa in the Earth. Although these stress variations are much smaller than typical stress drops of earthquakes, their rates are generally much larger than those of tectonic stress accumulation. Therefore, it has been postulated that the tidal stress might act as a triggering factor for earthquakes. On the basis of this standpoint, many searches have been carried out for the correlation between the Earth tide and earthquake occurrence, but the results of these
investigations have been controversial (Emter, 1997); various studies indicated positive correlations (Heaton, 1975; Klein, 1976; Young and Zurn, 1979; Souriau et al., 1982; Ulbrich et al., 1987; Rydelek et al., 1988; Shirley, 1988; Tsuruoka et al., 1995; Wilcock, 2001; Tolstoy et al., 2002), while many others reported negative conclusions (Schuster, 1897; Knopoff, 1964; Simpson, 1967; Shudde and Barr, 1977; Heaton, 1982; Curchin and Pennington, 1987; Hartzell and Heaton, 1989; Rydelek et al., 1992; Vidale et al., 1998).

Recently, however, we have successfully verified a clear correlation between the Earth tide and earthquake occurrence by carefully examining global earthquake data (Tanaka et al., 2002b). In our analysis, a weakness in the previous studies was improved; the indirect term of the Earth tide due to the ocean tide loading was properly incorporated into the theoretical calculation of the Earth tide by using an advanced ocean tide model. Applying these new tidal synthetics, we further found characteristic space–time pattern of the earthquake–tide correlation that is closely related to the occurrence of the 1982 $M_w=7.5$ South Tonga earthquake (Tanaka et al., 2002a); a high correlation appeared in and around the focal region only for several years preceding the large earthquake, and the correlation vanished after that. Similar space–time patterns were observed for several large earthquakes in Japan (Tanaka et al., 2004). These results suggest that the tidal triggering effect may appear when the stress in the focal region is close to a critical condition to release a large earthquake; in other words, the degree of the correlation between the Earth tide and earthquake occurrence may be spatially and temporally related to the tectonic stress condition. In this paper, we further explore the relationship between the tidal effect and the stress condition by focusing on the microearthquake activity in the Tokai region, central Japan, where the space–time pattern of the seismic activity has been precisely examined in connection with an impending great earthquake.

Great interplate earthquakes with magnitude 8 or larger repeatedly took place in the Tokai region, where the Philippine Sea plate subducts beneath the Eurasian plate at the Suruga trough (see Fig. 1a). On the basis of recurrence intervals of the past events and strain accumulation detected from geodetic measurements, the possibility of a next earthquake in the near future has been pointed out (e.g., Mogi, 1970; Ishibashi, 1981). A dense seismic network covering this area was deployed by the National Research Institute for Earth Science and Disaster Prevention (NIED), which has provided high-quality data of microearthquake activity in and around the Tokai region for about two decades (Okada, 1984; Hamada et al., 1985). Careful investigations of these data greatly extended our knowledge of the mechanism of the plate subduction in this area. For example, the geometrical configuration of the subducting Philippine Sea plate was delineated from the hypocentral distribution of these microearthquakes (Ishida, 1992). The precise hypocenter locations revealed two distinct seismically active layers as shown in Fig. 1b, where earthquakes are plotted on the vertical cross section along the direction of relative plate motion; a narrow gap zone shown by a double dotted line clearly separates an upper layer with the depth of 0–25 km and a lower layer dipping to the northwest. Both the two layers are characterized by predominantly strike-slip faulting, not by thrust faulting, although there is a marked difference in the P-axis direction between them (E–W in the upper layer and N–S in the lower one). Based on these observations, the earthquakes in the upper and lower layers are interpreted to occur within the overriding and subducting plates, respectively (Ishida, 1992). Thus, the boundary between those two plates may pass through the aseismic narrow gap zone between the seismically active layers, which is shown by a double dotted line in Fig. 1b for the depth range from 10 to 30 km. In the zones just above and below this narrow gap zone, the intraplate seismic activity is relatively high compared to that in other portions in the seismically active layers. Evaluating this observation jointly with geodetic data, Ishida (1995) suggested that the two plates might be strongly coupled within this narrow gap zone. Characteristic pattern closely related to the narrow gap zone was also found in the focal mechanism solutions observed in the lower seismically active layer (Matsumura, 1997). In the intra-slab seismic layer just beneath the narrow gap zone, the dip angle of P-axes systematically rotates along the down-dip direction of the plate boundary; shallow angles at the trench side and near vertical angles at the deep part. Supposing that such a pattern might come from the stress concentration due to the locking within a limited zone of the plate boundary, Matsumura (1997) identified the locked portion between the two plates that is to be ruptured by the anticipated Tokai earthquake. The identified locked zone is shown in Fig. 1a by a shaded area. The total area is about 4000 km$^2$, which corresponds to the focal area of an earthquake with magnitude 8. Around this inferred locked zone, furthermore, the space–time pattern of the microearthquake activity was examined...
in detail (Matsumura, 2003, 2006—this issue). For over 10 years, the activity had been characterized by surprisingly stable and constant rate. However, in August 1999, a clear decrease of seismicity was detected inside the subducting slab. Detailed analysis of Matsumura (2003, 2006—this issue) revealed that the seismicity rate increased in some sub-regions and it decreased in some others in spite of total decrease of seismicity rate. This activated/quiescent pattern of the microearthquake activity is considered to reflect the spatial pattern of the coupling degree between the two plates (Wyss, 1986; Wiemer and Wyss, 1994).

These aforementioned studies, revealing detailed features of seismicity in the Tokai region, provide a strong background for attempting a search of a tidal effect on earthquake occurrence associated with space–time variation of seismic activity. In this paper, we measure the correlation between the Earth tide and earthquake occurrence with a high resolution both in space and time in order to compare it with the characteristic behavior of microearthquake activity in this area. Based on the result, we discuss the possibility that an intensified tidal triggering effect may indicate the final stage of tectonic stress increase prior to the occurrence of a mainshock.

Fig. 1. Hypocenter distribution of microearthquakes ($M \geq 1.5$) in the Tokai region: (a) map view and (b) vertical cross section. Solid rectangle in the map view indicates the location of the vertical cross section. Shaded area indicates the locked zone of the plate interface after Matsumura (1997). Arrow indicates the convergence direction of the Philippine Sea plate. Abbreviation of tectonic plates in the inserted map is EU, Eurasian plate; NA, North American plate; PAC, Pacific plate; PHS, Philippine Sea plate. Double dotted line in the vertical cross section indicates the boundary between the two distinct seismically active layers.
We also propose that the asperity of a pending earthquake may be discriminated by carefully examining the spatial distribution of intensity of the tidal triggering effect.

2. Data

In the Tokai region, a high-sensitivity dense seismic net of NIED has provided accurate hypocenter locations with location errors smaller than 1.0 km in latitude and longitude and 1.5 km in depth for the past nearly two decades. We use the origin times, hypocenter locations, and focal mechanism solutions of $M \geq 1.5$ earthquakes that were observed by the NIED net for the period from June 1986 to February 2003, extending the data set used by Matsumura (2003). We focus our analysis on intraslab earthquakes, which occur randomly in time and distribute uniformly in space (Matsumura, 1997). As for the study area, we set the rectangular area of 80 km $\times$ 70 km that is marked by a solid rectangle in Fig. 2; this is the same area that Matsumura (2003, 2006—this issue) selected for monitoring microearthquake activity to reveal the state of interplate locking in this region. The study area covers almost the locked zone of the plate interface after Matsumura (1997).

Fig. 2. Hypocenter distribution of 808 earthquakes used in this study (black dots, $M \geq 1.5$, 1986/06–2003/02): (a) map view and (b) vertical cross section. Dot-dashed rectangle in the map view indicates the location of the vertical cross section. Solid rectangle indicates the study area of this study. Stars are the hypocenters of $M \geq 4.5$ earthquakes. Focal mechanism solutions are shown for those $M \geq 4.5$ earthquakes. Shaded area indicates the locked zone of the plate interface after Matsumura (1997).
plate interface inferred by Matsumura (1997), which is shown by a shaded polygon in Fig. 2. The southwestern edge of the locked zone, which is located beneath Lake Hamana, is excluded from the study area since the extremely high seismicity beneath Lake Hamana may reflect a local special condition other than simple coupling of the plate interface (Matsumura, 1997).

From the data set, we remove clustered events such as aftershocks and earthquake swarms that may lead to erroneous conclusions. For the declustering, we use a simple algorithm of Matsumura (2003) based on a space–time distance between events. In this algorithm, events occurring within a specified space–time window are linked into one group. When events belonging to different groups are linked, the respective groups are redefined as one group. Completing this procedure for all the events, we pick up only the first event of each group and remove the others. For the space–time window, we select 3 km for latitude and longitude and 7 days for time after Matsumura (2003, 2006—this issue) (see Matsumura (2006—this issue) for the details). Applying the declustering algorithm to the 919 earthquakes that occurred within the study area, we eliminated 111 events (12%) from the original catalog. We use the remaining 808 earthquakes for the analysis. The hypocentral distribution of these earthquakes is shown in Fig. 2 by black dots.

3. Method of analysis

We statistically investigate the correlation between the Earth tide and earthquake occurrence following the method of Tanaka et al. (2002a,b). At first, we theoretically calculate the Earth tide at the hypocenter of each earthquake for the Preliminary Reference Earth Model (Dziewonski and Anderson, 1981). This calculation includes both the direct solid Earth tide and the indirect term due to the ocean tide loading for a reliable ocean tide model NAO.99b (Matsumoto et al., 2000; Takanezawa et al., 2001). To relate the Earth tide with fault rupture, we focus on the stress components on the fault plane by using the focal mechanism solution of each earthquake. The important components to affect the fault rupture are the normal and shear stresses on the fault plane. In the area of this study, those two components of the Earth tide have different amplitudes (1–5 kPa for the normal stress and 0.5–1.5 kPa for the shear stress) and are not in phase, so we should consider both. However, we do not consider the normal stress in the present study because it is difficult to identify the fault plane from the two nodal planes. On the other hand, the shear stress does not depend on the selection of the fault plane due to the symmetry of the stress tensor. As the tidal component, therefore, we only consider the shear stress along the slip direction on the fault plane (positive in the same sense as the fault slip).

From the calculated tidal shear stress, we next assign a tidal phase angle at the occurrence time for each earthquake as illustrated in Fig. 3. A tidal phase angle, taking a value between $-180^\circ$ and $180^\circ$, is assigned by using a linear scale with time from $-180^\circ$ to $0^\circ$ or from $0^\circ$ to $180^\circ$, where $0^\circ$ and $\pm 180^\circ$ correspond to the maximum and the minimum of the tidal shear stress in the slip direction, respectively. For example, an earthquake occurring at a time that is exactly centered between the times of the maximum tidal stress and the subsequent minimum stress (a cross-marked event in Fig. 3) is assigned a tidal phase angle $\theta_i = 90^\circ$.

Determining the phase angles for all the earthquakes, we statistically test whether they concentrate near some particular angle or not by using Schuster’s test (e.g., Emter, 1997). In this test, we represent each earthquake as a unit length step in the direction defined by its tidal phase angle in two-dimensional space. For a data set including $N$ events, we obtain a vector sum with the length $D$ as written by

$$D^2 = \left( \sum_{i=1}^{N} \cos \theta_i \right)^2 + \left( \sum_{i=1}^{N} \sin \theta_i \right)^2. \quad (1)$$

When the phase angles of $N$ events distribute randomly, the steps become a random walk. The probability that the length of vector sum is equal to or larger than $D$ is then approximately given by

$$p = \exp \left( - \frac{D^2}{N} \right), \quad (2)$$

where the approximation is sufficient when $N$ is larger than 10 (Heaton, 1975, 1982). This probability there-
fore represents the significance level to reject the null hypothesis that the earthquakes occur randomly, irrespective of the tidal phase angle. A smaller $p$-value indicates a higher correlation between the Earth tide and earthquake occurrence. To judge a significant correlation between them, a threshold of $p \leq 5\%$ was tentatively adopted in some previous studies (Emter, 1997; Tanaka et al., 2002b; Tolstoy et al., 2002).

4. Result

Fig. 4 shows the result of analysis for all the 808 earthquakes used in this study. The histogram is the frequency distribution of tidal phase angles, in which earthquakes are gathered into phase angle bins of $30^\circ$ width. In this case, the phase distribution is nearly flat, and the $p$-value is as large as 29%. This indicates that a significant correlation is not seen between the Earth tide and earthquake occurrence for the data set that includes all the earthquakes.

However, we find a characteristic tidal effect by precisely examining the space–time pattern of the $p$-value. Fig. 5 shows the temporal variation of the $p$-value (Fig. 5a) together with the magnitude–time plot of the earthquakes (Fig. 5b). We apply a moving window technique to the data set. In this analysis, a time window of 300 days is used to secure enough number of earthquakes for the statistical analysis (at least 20 events are included in each window), and is shifted by 50 days to resolve the variation of the $p$-value. For each time window, we also estimate the standard deviation of the $p$-value by using 100 bootstrap samples (Efron and Tibshirani, 1998). As a result of analysis, we find that the $p$-value was larger than 20% for the early years of
the investigation period. However, a clear decrease appeared for the period from July 1989 to August 1991. After that, the $p$-value returned to a high level. Throughout the entire period, the $p$-value exhibited such a decrease–recovery pattern repeatedly.

In Fig. 5, the magnitude–time plot of the earthquakes is also shown (Fig. 5b). In this figure, we note that $M \geq 4.5$ earthquakes were considerably larger than the normal background seismicity in the study area. There were four $M \geq 4.5$ earthquakes within the investigation period. These events are listed in Table 1 and their locations are shown in Fig. 2 by star marks. The tidal phase angles of these events are 135°, 0°, −171°, and 74° for Events A, B, C, and D, respectively. The occurrence times of these events are indicated in Fig. 5a by dotted lines. We find that the $p$-value pattern is closely correlated with them. A clear drop of the $p$-value appeared preceding the occurrence of the $M \geq 4.5$ earthquakes. The minimum value attained prior to each event was very variable; $p=0.59%$ 13%, 6.1%, and 4.2% for Event A, B, C, and D, respectively. Some of those values are not small enough to judge a significant correlation between the Earth tide and earthquake occurrence. However, the temporal change of the $p$-value is very systematic and exactly synchronized with the occurrence of the $M \geq 4.5$ earthquakes.

As an example, Fig. 6 shows the tidal phase distribution for the pre-event period of Event A in which the $p$-value exhibited the most drastic decrease. We pick up a time period of 700 days preceding the occurrence of Event A during which the $p$-value was less than 5% (shaded portion in Fig. 5a). Using all the earthquakes occurring in this period, we obtain a very small $p$-value of 0.37%. The phase distribution in this period had a peak near the angle 0°, where the tidal shear stress is at its maximum to accelerate the fault slip. Similar patterns in the phase distributions were observed for the pre-event periods of the other moderate-sized earthquakes (Event B–D). This suggests that the observed small $p$-value is not a stochastic chance but is a physical consequence of the tidal effect.

We also investigate the spatial distribution of $p$-values. In this analysis, we again use the earthquakes occurring within the entire investigation period to secure enough number of earthquakes. We use a spatial moving window of 11 km × 11 km that is moved by 3 km both in the north–south and east–west directions (we have confirmed that we obtain essentially the same pattern when using a larger window). The result of analysis is shown in Fig. 7. The circle size is scaled to the $p$-value for each window. Although small $p$-values are found in several locations, the most dominant low-$p$ appears in the central part of the study area.

We replot the locations of small $p$-values ($p<5\%$) in Fig. 8. In this figure, a spatial window of 11 km × 11 km is moved by 1 km, which is approximately equal to the epicenter accuracy. The resultant $p$-value for each window is plotted in the 1 × 1 km square at the center of the window. We clearly see that the most remarkable small-$p$ area is located at the center part of the study area. In this figure, we also show the spatial pattern of the seismicity rate change in the past few years. This pattern is obtained by using the same spatial moving window as the $p$-value analysis. The activated and quiescent areas are defined as the areas where the seismicity rate increased (the rate ratio is greater than 150%) and decreased (the rate ratio is less than 100%) in recent years from August 1999 to December 2002 compared to that in the reference period from June 1986 to May 1996 (Matsumura, 2003). The dominant low-$p$ region at the center of the study area is found to be located within an activated area of the recent seismic activity.

To statistically compare the spatial distribution of small $p$-values with that of the seismicity rate change, we perform a simple test using synthesized random earthquake data. To do this, we synthesize 1000 data

![Fig. 6. Frequency distribution of tidal phase angles for the 700 days preceding Event A (shaded portion in Fig. 5a). Solid curve represents a sinusoidal function fitted to the distribution.](image)

### Table 1

<table>
<thead>
<tr>
<th>Event</th>
<th>Date</th>
<th>Longitude (E)</th>
<th>Latitude (N)</th>
<th>Depth (km)</th>
<th>M</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>1991/04/25</td>
<td>138.2092</td>
<td>35.0480</td>
<td>27.3</td>
<td>4.8</td>
</tr>
<tr>
<td>B</td>
<td>1995/04/18</td>
<td>138.6042</td>
<td>35.0721</td>
<td>17.7</td>
<td>4.9</td>
</tr>
<tr>
<td>C</td>
<td>1997/10/21</td>
<td>138.2313</td>
<td>35.1220</td>
<td>26.4</td>
<td>4.5</td>
</tr>
<tr>
<td>D</td>
<td>2001/04/03</td>
<td>138.1006</td>
<td>34.9987</td>
<td>26.5</td>
<td>4.9</td>
</tr>
</tbody>
</table>
sets of random earthquake occurrence, in which the origin time is randomly assigned for each hypocenter. For each of the 1000 random data sets, we draw the spatial distribution of $p$-values and compare it with the distribution of the seismicity rate change. To quantify the correlation of the low-$p$ region with the seismic activated area, we count the number of $p < 5\%$ windows that are located within the activated area both for the observed real earthquake data ($n_{\text{obs}}$) and each of the synthesized random data ($n_{\text{syn}}$). Among the 1000 data sets we examined, 121 cases (12\%) indicate $n_{\text{obs}} < n_{\text{syn}}$, and 879 cases (88\%) $n_{\text{obs}} > n_{\text{syn}}$. Thus, it is confirmed that the low-$p$ regions tend to concentrate in the activated areas although the statistical significance is not high enough.

We observe a clear contrast of the tidal effect between the seismically activated and quiescent regions. Fig. 9 shows the frequency distributions of tidal phase angles for two data subsets; one includes only the earthquakes occurring in the activated area, and the other includes only those in the quiescent area. For the activated area, the $p$-value is small ($p = 2.2\%$), and the phase angles again tend to concentrate near $0^\circ$. For the quiescent area, on the other hand, the $p$-value is large ($p = 17\%$), and the distribution shows little phase selectivity.

5. Discussion

We observed a characteristic temporal variation of the $p$-value that was strongly coincident with the occurrence of the $M \geq 4.5$ earthquakes. The $p$-value exhibited a clear decrease prior to these moderate-sized earthquakes, and recovered a high value afterwards. Such a systematic change of the $p$-value indicates that a strong tidal effect appears preceding the occurrence of the $M \geq 4.5$ earthquakes in this region. It is most probable that a significant decrease of the $p$-value indicates that the stress state is at a critical level to release a main rupture. However, there might be other possible causes of the $p$-value change. For example, a careful analysis of leveling data revealed that the rate of subsidence of the coast side compared to the inland side anomalously increased in the Tokai region during the period from 1987 to 1990 (Takayama and Yoshida, 2002). Just in the same period, furthermore, the seismic
quiescence was detected for $M \geq 2.5$ earthquakes in this region (Yoshida and Maeda, 1990), although such quiescence was not seen in the earthquake data used in this study that includes more smaller events ($M \geq 1.5$). Those two anomalies are interpreted to be related to a change in the coupling state of the plate interface due to the occurrence of a slow slip event (Takayama and Yoshida, 2002), which might have caused a drastic $p$-value decrease in the period from 1989 to 1991. On the other hand, a recent analysis of GPS data showed another slow slip event started in 2000 in the western Tokai region centered beneath Lake Hamana, which is adjacent to our present study area (Ozawa et al., 2002). A small $p$-value observed in almost the same period might be related to this slow slip event also.

To make the relationship between the $p$-value change and the occurrence of the 2000 slow slip event clearer, we performed an additional analysis for...
the 20 km × 20 km area centered at Lake Hamana by using 171 in-slab earthquakes in the investigation period. The analysis does not show any p-value decrease around 2000, but a high p-value larger than 20% continued throughout the entire investigation period. This result implies that the change of the p-value we observed was not caused by the occurrence of the slow slip event. We conclude that the systematic temporal change of the p-value reflects the pre-seismic anomaly of the moderate-sized earthquakes.

Similar temporal changes of the p-value have been reported for several large earthquakes in the Tonga–Kermadec ($M = 7.5$) and Japan ($M \geq 6.0$) regions (Tanaka et al., 2002a, 2004). Those observations suggest that the anomalous tidal effect in the pre-seismic stage may be a common nature of earthquake occurrence in variable range of the mainshock magnitude. Here, we note that the pre-seismic low-p period observed in this study is much shorter than that reported in our past studies; a significant decrease of the p-value was found for several hundred days preceding the mainshocks of $M = 4.5–4.9$ in this study, while the leading time was several thousand days for the $M = 7.5$ South Tonga earthquake (Tanaka et al., 2002a). This suggests that the duration of anomalous low-p may be scaled with the mainshock magnitude. However, this scaling relation is not yet established due to limited number of observations. Further progress based on case studies is needed to contribute to predicting the magnitude of a pending earthquake.

The pre-seismic low-p anomalies shown in Fig. 5a were detected by using all the earthquakes occurring in the study area of roughly 80 km × 70 km, which is much larger than the focal dimension of the moderate-sized mainshocks (~1 km). It is of particular interest to resolve whether the anomalous low-p was limited to the vicinity of corresponding mainshocks or not. However, it is difficult to apply a moving window analysis to the short-term anomalous periods due to limited number of earthquakes. Detailed analysis may be possible when a larger mainshock with a longer anomalous period takes place in the future.

In the present study, we also revealed a high correlation between the Earth tide and earthquake occurrence was restricted in an activated area, by using the events occurring in the entire investigation period. In this area, the seismicity rate had increased in the past few years compared to the background rate. Based on the recent observations on the microearthquake activity change before several large earthquakes (Wyss, 1986; Wiemer and Wyss, 1994), Matsumura (2003) argued that the activated/quiescent pattern of the microearthquake activity in the subducting plate might reflect the spatial distribution of the degree of coupling between the two plates. In this interpretation, the activated area of in-slab seismicity is regarded as a stress concentration zone just beneath the strongly coupled portion of the plate interface, which is to be an asperity of an impending great earthquake. If this is the case, our result may predict that the tidal triggering effect may appear in the close vicinity of an asperity. A systematic monitoring of the tidal triggering effect is expected to contribute to effectively detecting asperities of future earthquakes.

Throughout this study, we focused on the locked zone of the plate interface that is inferred from the microearthquake activity by Matsumura (1997). In the Tokai region, back-slip analyses using geodetic data have also been performed to estimate the strongly coupled zone of the plate boundary (Yoshioka et al., 1993; Sagiya, 1999). These studies resulted in essentially a similar pattern of the strongly coupled zone; however, there exists a clear difference in the results between the microearthquake analysis and the back-slip inversion analysis. Compared with the locked zone of Matsumura (1997) that is located under land area, the large back-slip zone that is considered to be strongly coupled portion of the plate interface is shifted south-eastward to the ocean area. It would be interesting to extend our analysis to the inferred large back-slip zone. However, since the seismicity of this zone is very low and sparse, it is difficult to get a data set that is feasible to examine the space–time pattern of the tidal effect.

As for the stress component of the Earth tide, we only focused on the shear stress on the fault plane. To reveal the effect of the normal stress, we need to discriminate the fault plane from the two nodal planes. This may be possible for some larger events by analyzing the distribution of active faults and/or aftershock distributions. When the effect of the normal stress is clarified, more quantitative discussions on the physical mechanism of the tidal triggering effect will be possible in view of the Coulomb failure criterion (e.g., Jaeger and Cook, 1969) and laboratory-driven friction laws (e.g., Dieterich, 1979). Such discussions may provide a new tool to get insight into the nature of fault rupture generation.

6. Conclusion

We measured the correlation between the Earth tide and earthquake occurrence in and around the locked zone of the plate interface in the Tokai region, central Japan, by using microearthquakes that took place in the subducting Philippine Sea plate for about two decades.
As a result of the statistical test, a significant correlation was not found for the data set including all the earthquakes. However, we found a systematic pattern in the temporal change of the earthquake–tide correlation. Remarkable phase selectivity appeared 1–2 years preceding the occurrence of the $M \geq 4.5$ earthquakes, and vanished afterwards. We note that those $M \geq 4.5$ earthquakes were considerably larger than the normal background seismicity in the study area. The frequency distribution of tidal phase angles in the pre-event period exhibited a peak at the phase angle where the tidal shear stress is at its maximum to accelerate the fault slip. This indicates that the observed significant correlation is a physical consequence of the tidal stress change. The strong correlation was also observed in particular just beneath the strongly coupled portion of the plate interface, which was inferred from the seismicity rate change in the past few years.

This paper shows the new evidence of the tidal effect on earthquake occurrence that was spatially and temporally related to the characteristic behavior of seismic activity in the Tokai region. The result we obtained suggests that a systematic monitoring of tidal effect may be feasible for detecting the final stage of tectonic stress prior to the occurrence of a main rupture. Our observation further suggests that the studies on tidal triggering may provide a new insight into asperities of pending earthquakes.

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References


