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Deep low-frequency tremor that correlates with passing surface waves
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[1] The large surface waves from the 2004 Sumatra-Andaman earthquake dynamically perturbed the upper mantle structure in Japan and triggered periodic deep low-frequency seismic tremor in eastern and western Shikoku, western and central Tokai, and the Kii peninsula. We use the relationship between the amplitude of the triggered tremor and the stresses of the seismic waves to investigate the mechanism of deep low-frequency seismic tremor. Volumetric strain changes from the 15–30 s Rayleigh waves play an important role in the strong triggering, likely via Coulomb failure stress changes. Building on previous results that the tremor signals become increasingly strong with increasing dilatation, we observe a clear increase in the triggered tremor with an increase in the dilatation due to the Rayleigh waves at the 30 km depth source regions. We also observe a correlation with the Coulomb failure stress change resolved on an appropriate plane. There is an exponential relationship between the signal amplitude from triggered tremor and both the dilatation and the Coulomb shear stress at the source region. This combined with the shape of the tremor packets implies that the tremor amplitude is predictable based on the amplitude of the incoming waves. The amplitude variations can be explained by a distribution of sources in the tremor source region.


1. Introduction

[2] Deep low-frequency tremor on subducting slabs is a recently discovered and novel seismic signal [Obara, 2002]. The waves are unusually long-period for their amplitude and the signal is often continuous with little punctuation over times ranging from tens of minutes to a few days. These seismic features are often seen in volcanic regions and usually interpreted as signs of fluid-filled resonant conduits in that setting. Therefore early work on deep, slab tremor suggested a fluid source by analogy. Geochemical and petrological constraints supported the inference by suggesting that fluids are released from the subducting slab in a series of dehydration reactions at the depth of the tremor [e.g., Toriumi and Inui, 2001; Omori et al., 2004]. The coincidence of the tremor with slow slip also suggests that there is some physical connection with the large-scale slab processes at the depth of 30–40 km.

[3] More recent work has suggested that low-frequency earthquakes and tremor can be generated by simple shear failure in both the subduction and the volcanic settings. A low rupture velocity or low-stress drop can result in the low-frequency waves and overlapping earthquakes can generate the apparently continuous signal [Harrington and Brodsky, 2007; Shelly et al., 2006]. Still, the occurrence of the tremor and slow slip at the depth of dehydration suggests that fluids may be an important component.

[4] Here we study the origin of slab tremor by using a constraint posed by special cases of slab tremor that are triggered by the seismic waves from distant earthquakes. The 2004 Sumatra-Andaman earthquake (Mw 9.2) was enormous and the surface wave amplitudes measured in Japan were comparable to or a few times as large as those from the Denali earthquake (Mw 7.9) in 2002 that contributed to the well-studied earthquake triggering around North America [e.g., Prejean et al., 2004]. Miyazawa and Mori [2006] showed that periodic triggering of deep low-frequency events in western Japan was due to the Rayleigh waves from the Sumatra earthquake, and suggested that the triggering is well correlated with the large tensile dilatation at the source regions (Figures 1 and 2). Similar triggered tremor occurred after small local earthquakes and large teleseisms [Obara, 2003], and during the surface waves of the 2003 Tokachi-oki earthquake (Mw 8.1) [Miyazawa and Mori, 2005]. These initial observations were thought to distinguish the tremor from the ordinary earthquakes triggered in Alaska that were promoted by shear failure [West et al., 2005]. Recently, Rubinstein et al. [2007] found that the bursts of similar nonvolcanic tremor in Cascadia subduction zone were triggered by Love waves from the 2002 Denali earthquake.

This paper begins with a review of prior observations of triggered tremor combined with a discussion of the most salient first-order features found here. We then derive the strain changes in the tremor source regions due to the Sumatra seismic waves. By comparing the spectra of the triggered and input signals, we determine which frequencies of the surface waves are capable of significantly triggering the tremor. We proceed to more carefully investigate the functional dependence of the amplitude of the tremor given various assumptions about the spatial distribution of sources. We also examine the data’s ability to resolve the orientation of the input stresses. Finally, we discuss the physical implications of our results for both an individual failure event and a distribution of sources.

2. Observations

2.1. Overview of Triggered Tremor

Miyazawa and Mori [2006] used the High Sensitivity Seismograph Network (Hi-net) to uncover nonvolcanic deep low-frequency tremor triggered by the Sumatra earthquake. By filtering the continuous waveforms, they found discrete episodes of low-frequency tremor that correlated well with the packets of the Rayleigh waves (Figure 1). They established that volumetric expansion in the source region correlated well with the triggering (Figure 2). When the dilatation had peak values of about $10^{-5}$, strong triggering was observed. (Throughout this paper, we use the convention of extension and expansion as positive strain and the term “dilatation” strictly for volumetric strain.)

Miyazawa and Mori [2006] located the tremor using a modified envelope correlation method [Obara, 2002; Miyazawa and Mori, 2005], which measures traveltime differences between the envelopes by using cross correlations of the envelope time series for a range of time lags. We relocated the hypocenters using a double-difference method [Waldhauser and Ellsworth, 2000] and a velocity model JMA2001. The source locations are shown by red circles in Figure 3 and the sources located at depths of $\sim$30 to $\sim$40 km, which correspond to the regions where the deep low-frequency earthquakes have episodically occurred (yellow circles in Figure 3) above the subducting Philippine Sea plate. The triggered tremor seems to have occurred in five

Figure 1. Observed velocity waveforms from the 2004 Sumatra-Andaman earthquake (a) filtered with a passband from 2 to 16 Hz at a borehole high sensitive station KWBH and (b) filtered with a passband from 0.01 to 1 Hz at a broadband station TSA. Zero is the origin time of the Sumatra earthquake (26 December 2004, 0058:53 UT). Three traces indicate the vertical, radial, and transverse components from the top to the bottom. The two station locations are shown in Figure 3. The epicentral distance is about 5000 km.
clusters: western and eastern Shikoku, western and central Tokai and the Kii peninsula.

The observations below show that the triggered tremor-like events are deep low-frequency signals that are similar to other reports of slab tremor or low-frequency earthquakes. In order to avoid a premature identification of the physical mechanism of the tremor, we continue to use the term “event” for discrete episodes of tremor with well-defined arrivals. Ultimately, we will identify a continuous episode of tremor with overlapping earthquakes in a model similar to that suggested by Shelly et al. [2006].

The spectrum of the triggered events is consistent with other tremor observation and is distinct from any of the features of the seismic waves originating at the main shock. Figure 4 shows an example of the Fourier spectra of vertical and horizontal components from 1400 to 2400 s (see Figure 1) at stations with almost the same epicentral distance from the Sumatra earthquake, KWBH and IKNH, where the tremor has been observed and has not, respectively. The ratios of the same components for two stations are shown in the bottom. The spectra lower than 1 Hz almost correspond to each other, while there are large differences in ranges higher than 1 Hz, which are especially significant in the horizontal component. The isolated tremor spectrum is consistent with other measures of the spectra of episodic deep low-frequency earthquakes showing unusually low frequencies for its amplitude and a roughly $1/f$ frequency decay [Ide et al., 2007b]. The 20 Hz peak is known contamination due to a borehole resonance.

Like other examples of low-frequency earthquake and tremor, the triggered events generated predominately shear waves as inferred by the apparent velocity and large amplitude in horizontal components (Figures 1 and 5), and had large amplitudes for waves from 1 to 15 Hz (Figure 4). Figure 5 shows the largest tremor observed in western Shikoku as an example. The observed 2–16 Hz waveforms at each epicentral distance are shown for each component. Two dotted curves roughly show $P$ and $S$ wave arrivals, assuming the origin time is 0 on the scale shown. The large wave packets travel at $S$ wave velocity, while we can find arrivals of subtle proceeding signal in vertical components which seem to be on the $P$ wave arrivals. The proceeding signals may be $P$ waves and/or $S$ waves from another...
triggered tremor. If the source model of the triggered tremor is equivalent to a double couple model, we can observe $P$ waves at the stations located in the compressional/tensional quadrant of the focal mechanism, which are along northwest and southeast directions at some distances at western Shikoku. We do not always observe them for other triggered tremor. As the early phases are still unclear at other stations and are not common to the other smaller triggered tremor, the events may occur on or above the plate boundary. Higher-frequency waves do not show as clear an orientation from the particle motion. The results are consistent with the source mechanism of a deep low-frequency earthquake by Ide et al. [2007a].

[13] An interesting feature of the envelope waveform is that the large packets of tremor are almost symmetrical at each station. To see if this is caused by path effects, we looked at large earthquakes ($M \sim 4$) that occurred in a similar but slightly deeper location (depth $\sim 40$ km) near the plate boundary or in the subducting slab. The earthquakes radiated waves with similar frequencies to the tremor, but amplitudes several thousand times larger. For the earthquakes, the onsets of $P$ and $S$ waves are clear. In contrast, the waveforms of the tremor in Figure 5 show emergent arrivals even at the distant stations and on high-frequency components. This feature distinguishes the tremor from ordinary tectonic earthquake and implies that the triggered tremor likely produces mostly $S$ waves in some continuous or repeated process. Since the envelope shape apparently reflects a source process, it suggests that we should analyze the correlation of the tremor amplitude as a continuous function of the incoming seismic wave strains/stresses.

[14] Larger amplitude surface waves generate more abundant and stronger tremor. Figure 7 shows short-period and long-period root-mean-square (RMS) envelope waveforms from representative examples of each of the clusters in Figure 3. The short-period waveforms are constructed from waveforms filtered with a passband of 2–16 Hz and show the activity of triggered events, and the long-period waveforms are the envelope waveforms of Love and Rayleigh waves filtered with a passband of 0.01–1 Hz. In western Shikoku and the central Tokai regions, the tremor is more actively triggered than other three regions. In all areas, the amplitude of Rayleigh waves (gray line) rather than Love waves (dotted line) appears to correlate with the magnitude of large triggered event. In sections 2.2, 2.3, and 2.4, we will attempt to quantify and expand upon this apparent correlation.

[15] An additional source information on triggered tremor comes from the $M_{w}8.1$ Tokachi-oki earthquake. Although less spectacular than the larger Sumatra earthquake, it also triggered tremor in the same regions of the subduction zone. Figure 8 shows observed waveforms at western Shikoku from the Tokachi-oki earthquake, where the epicentral distance is about 1400 km. Three peaks of significant tremor indicated by arrows evidently correspond to the peaks of Rayleigh wave envelope (Figure 8). By using the tremor source location [Miyazawa and Mori, 2005] we observed the large dilatation at source region for the large signal amplitudes, like shown in Figure 2b.

[16] Clear triggering of tremor by Rayleigh waves occurred in Japan for at least one other distant earthquake, the Solomon Islands earthquake ($M_{w}8.1$) of 1 April 2007. The general features are similar to those reported here.

2.2. Strains at Depth

[17] Miyazawa and Mori [2006] examined the strains at the triggered tremor source regions estimated by continuing the wave field observed on the surface to depth using a kernel for one cycle of surface waves, and found that triggered tremors were synchronized with large dilatations.
Here we investigate the strains at depth for the full spectrum using appropriate kernels for continuous waveforms.

[18] To extrapolate the observed dilatational strains to depth, we use solutions for Rayleigh wave equation for a simple half-space structure [e.g., Lay and Wallace, 1995] and calculate the displacements and then the resulting strains. We only use the fundamental modes as they comprise the components of the strain field that is less

![Figure 5. Three-component velocity waveforms (2–16 Hz) of a sample event (western Shikoku event 14 in Figure 7) for epicentral distance. The scale of amplitude is the same for every waveform. Time 0 indicates the origin time of the source that produced the largest peak. Two dotted curves of earlier and later arrivals roughly show P and S wave traveltime curves, respectively. The station azimuth is indicated in the right margin.](image)

![Figure 6. Horizontal particle motions (0.5–2 Hz) of three large triggered events. A gray bold line indicates the direction of the 40 km depth contour of subducting Philippine Sea plate boundary estimated from the seismicity. Events are as numbered in Figure 7: event 14 in western Shikoku, event 9 in western Tokai, and event 10 in central Tokai observed at KWBH, URSH, and NUKH, respectively.](image)
Figure 7. High-frequency (2–16 Hz) root-mean-square (RMS) envelope waveforms (top panels) and magnitude-time graphs with low-frequency (0.01–1 Hz) RMS envelopes of Love and Rayleigh waves observed at the 5 stations KWBH, IKWH, HRKH, URSH, and NUKH (bottom panels). The stations are indicated in Figure 3. The high-frequency envelopes in the solid black lines at the top of each panel are constructed from filtered waveforms from the three components (UD, NS, and EW). The bottom panels show the Love wave envelopes in dashed line from the transverse component and the Rayleigh wave envelopes in gray lines from the vertical and radial components with calibration (see section 2.4). The vertical scale of amplitude is shown at the left portion for each. The bottom panels also include the timing and magnitude of each of the triggered tremor events with magnitudes >1 in black circles with the origin time based on the largest amplitude signal. The scale of magnitude is labeled in the right vertical axis.
sensitive to local structure. We take positive for radial and vertical directions. The normal strain changes across radial and vertical directions are given by

\[ e_{rr} = -A_0 k^2 \cos(kr - \omega t) \]
\[ \cdot \left[ \exp(-\omega \tilde{\eta}_a d) + \frac{1}{2} \left( \frac{c^2}{\beta^2} - 2 \right) \exp(-\omega \tilde{\eta}_b d) \right] \]
\[ e_{zz} = A_0 k \cos(kr - \omega t) \]
\[ \cdot \left[ c \omega \tilde{\eta}_a \exp(-\omega \tilde{\eta}_a d) \right. \]
\[ \left. + \frac{\omega}{2c} \left( \frac{c^2}{\beta^2} - 2 \right) \exp(-\omega \tilde{\eta}_b d) \right] \]

respectively, where

\[ \tilde{\eta}_a = \sqrt{\frac{1}{c^2} - \frac{1}{\alpha^2}} \quad \tilde{\eta}_b = \sqrt{\frac{1}{c^2} - \frac{1}{\beta^2}} \]

\( A_0 \) is a constant coefficient, \( k \) is the wave number, \( \omega \) is the angular frequency, \( r \) is the radial distance, \( d \) is the depth \((z = -d)\), \( \alpha \) is \( P \) wave velocity, \( \beta \) is \( S \) wave velocity and \( c \) is the phase velocity of the Rayleigh wave. We assume \( \alpha = 8.7 \text{ km/s}, \beta = 5.0 \text{ km/s}, \text{ and } c = 3.5 \text{ km/s} \). When we change these velocities within reasonable ranges, the results show little differences. The volumetric strain change \( \Delta V/V \) is approximately given by \( e_{rr} + e_{zz} \). Ideally, there is no contribution to \( e_{rr} \) and \( e_{zz} \) from Love waves.

[19] To calculate strains beneath a station, we directly estimate the phases and amplitudes from the observed vertical component \( u_{v,obs} \) at the surface, while Miyazawa and Mori [2006] use only equations (1) and (2) for each cycle of the surface waves. Because the vertical particle motion at the surface is

\[ u_v|_z=0 = A_0 k \cos(kr - \omega t) \cdot \left[ c \tilde{\eta}_a + \frac{1}{2c\tilde{\eta}_b} \left( \frac{c^2}{\beta^2} - 2 \right) \right], \]

Figure 8. The 2003 Tokachi-oki earthquake (Mw, 8.1) observed at western Shikoku. The lower three waveforms are observed at TSA and are filtered with a passband of 0.01–1 Hz. (top) Three root-mean-square envelope waveforms. The envelope in gray is constructed by the three-component high-frequency (4–16 Hz) waves observed at KWBH, where the epicentral distance is about 1400 km. The scale of the amplitude is labeled in the left vertical axis. The envelopes in black solid and dashed lines are constructed from the vertical and radial components and transverse component from the Tokachi-oki earthquake, respectively, shown in Figure 8 (bottom). These envelopes in black solid and dashed lines with large amplitudes indicate Rayleigh wave envelope and Love wave envelope, respectively. The scale of amplitude is labeled in the right vertical axis. Zero is the origin time of the Tokachi-oki earthquake (25 September 2003, 1950:08 UT). Three arrows indicate the significant triggering of tremor, which is consistent with the peaks of Rayleigh wave envelope. (bottom) Vertical and radial components and transverse component from the Tokachi-oki earthquake.
$e_{rr}(u_{r}|z=0)$ and $e_{zz}(u_{z}|z=0)$ include neither $A_0$ nor the sinusoidal function. We obtain the predominant period of the surface wave at arbitrary time to give $k$ and $\omega$, assuming that the period is unique and the surface wave is not dispersed during the cycle. Hence we much more accurately have the computed normal strain changes beneath the station at arbitrary depth as

$$e_{rr} = \frac{e_{\text{ref}}}{u_{r}|z=0} u_{r}^{\text{obs}},$$

$$e_{zz} = \frac{e_{\text{ref}}}{u_{z}|z=0} u_{z}^{\text{obs}}$$

where $e_{rr}^{\text{ref}}$ and $e_{zz}^{\text{ref}}$ are reference strains $e_{rr}$ and $e_{zz}$ in equations (1) and (2), respectively. Similarly, for the radial particle motion $u_{r}$ and the vertical particle motion $u_{z}$, we obtain the normal strain changes at the deep low-frequency source beneath the station at arbitrary depth as

$$\frac{u_{r}|z=0}{u_{r}|z=0} = \frac{\omega \eta_{0}^{2} \exp(-\omega \eta_{0} d)}{c^{2}} + \frac{2}{\omega \eta_{0}} \left( \frac{\omega \eta_{b}}{c^{2}} - 2 \right) \exp(-\omega \eta_{b} d),$$

$$\frac{u_{z}|z=0}{u_{z}|z=0} = \frac{2 k \eta_{0}^{2}}{c^{2}} \left[ \left( \frac{\omega \eta_{0}}{c^{2}} - 2 \right) \exp(-\omega \eta_{0} d) \right],$$

and $u_{r}^{\text{obs}}$ is the observed radial particle motion at the surface. Finally, shear strain change at depth is given by

$$e_{rr} = e_{zz} = \frac{1}{2} \left( \frac{u_{r}|z=0}{u_{r}|z=0} + \frac{u_{z}|z=0}{u_{z}|z=0} \right) u_{r}^{\text{obs}}.$$
the spectrum of individual components at depth than the dilatation.

2.4. Strain Amplitude and Triggered Amplitude

[25] One way to investigate the quantitative relationship between tremor triggering and surface waves, is to compare the amplitude of the strain components in the Rayleigh waves to the local magnitude. The event magnitude calibration includes attenuation of waves propagating through the structure, and may therefore be an improvement on simply comparing the peak amplitude. Figure 10 shows the relationship between tremor magnitude and the largest volumetric strain change observed during the tremor excitation. The larger amplitude waves seem to be capable of triggering tremor with large magnitude, but small tremor also occurs during the large amplitude surface waves. This is also seen in Figure 7, where some tremor magnitudes are plotted lower than the envelope waveforms. As a result, we could not find a clear relationship with magnitude using this method.

[26] We therefore proceed to investigate the relationship between signals from triggered sources and strain changes. We study the correlation of the tremor with the dilatational strain change ($\Delta V/V$), the shear strain resolved on the fault plane, and the change in the Coulomb failure function change

\[ \Delta \text{CFF} = \Delta \tau + \mu(\Delta \sigma_n + \Delta p), \]

where $\mu$ is the friction coefficient, $\Delta \tau$ and $\Delta \sigma_n$ are shear stress change resolved in slip direction and normal stress change on the fault plane, respectively, and $\Delta p$ is pore pressure change. Pore pressure changes are anticorrelated with the normal stress changes with $\Delta p = -B\Delta \sigma_n$, where $B$ is the Skempton coefficient. Therefore the second term of Coulomb failure function change can also be written as $\mu' \Delta \sigma_n$, where $\mu' = (1 - B)\mu$.

[27] We obtain the strain changes ($\Delta V/V$, $e_{rr}$, and $e_{zz}$) during the surface waves by using equations (5) and (6) with the appropriate periods and the observed vertical displacement waveforms.

[28] Strain changes at depth from the Love wave are also obtained from Lay and Wallace [1995]. We give a slow velocity layer above half-space, where we assume the differences of shear moduli and velocities between the top layer and the lower half-space is small so that the structure is similar to that used for Rayleigh wave. We use the fundamental mode. Then the dispersion equation is approximately given by

\[ \frac{\mu_2 \eta_{i_2}}{\mu_1 \eta_{i_1}} = 1, \]

Figure 9. (a) Fourier spectra for the vertical displacement, velocity, and acceleration of Rayleigh waves recorded at TSA. Time window is selected from 1300 to 4000 s. The vertical scale shown at top right portion corresponds to the values indicated in parentheses in each case. The horizontal axis is the period of Rayleigh wave. (b) Largest volumetric strain changes $\Delta V/V$ and corresponding two normal strain changes $e_{rr}$ and $e_{zz}$ (in vertical and radial directions, respectively) at the depth of 30 km, for the function of the period of Rayleigh wave. Amplitudes are normalized by the largest vertical displacement at the surface to demonstrate the depth correction factor (equations (5) and (6)). (c) Relationship between the period of Rayleigh wave and absolute strain changes (top) $\Delta V/V$, (middle) $e_{rr}$, and (bottom) $e_{zz}$, from the Rayleigh waves. The curves of TSA are made by combining Figures 9a and 9b. The similar relationships at UMJ, KMT, WTR, and NAA are shown in different colors. Station locations are in Figure 3. (d) Peak amplitudes of triggered deep low-frequency signals at five regions during arrivals of Rayleigh waves, with respect to indicated predominant period. The colors indicate the regions same as shown in Figure 9c (bottom).
The normal strain change in transverse direction, and

where \( \mu \) is shear modulus, \( \eta \) refers to equation (3), \( \eta = \frac{2}{3} \), \( \beta_1 < c < \beta_2 \) for Love wave velocity \( c \), \( \mu_2 = 1.5 \mu_1 \), and suffixes 1 and 2 denote the top layer and the lower half-space, respectively. For the transverse particle motion \( u_t \) and the depth \( d \geq H \),

\[
\frac{u_{t,2}}{u_{t,1=0}} = \frac{\pi \eta \beta_3}{4\sqrt{2}\eta \beta_1 H} \exp \left( \frac{-\pi \eta \beta_3}{4\eta \beta_1} d - H \right) \quad (12)
\]

\[
\frac{u_{r,2}}{v_{r,1=0}} = -\frac{k}{\sqrt{2}V} \exp \left( \frac{-\pi \eta \beta_3}{4\eta \beta_1} \frac{d - H}{H} \right) \quad (13)
\]

where \( H \) is the thickness of the top layer, \( u_{t,1=0} \) and \( v_{r,1=0} \) are transverse particle motions in displacement and velocity, respectively, at surface. We use \( H = 30 \) km and calculate the shear strain changes at the same depth. Shear strain changes at depth are given by

\[
e_{y} = e_{r} = \frac{1}{2} \left( \frac{u_{t,1=0}}{u_{r,1=0}} \right) r_{\text{obs}} \quad (14)
\]

and

\[
e_{y} = e_{r} = \frac{1}{2} \left( \frac{u_{r,1=0}}{v_{r,1=0}} \right) r_{\text{obs}} \quad (15)
\]

The normal strain change in transverse direction, \( e_{r} \), is equal to zero.

We then resolve the stresses on the fault plane and compare the tremor timing with each of the strain components. As shown by Ide et al. [2007a], the focal mechanisms of low-frequency earthquakes are consistent with the fault plane being aligned with the subducting slab at western Shikoku. We use the Ide et al. [2007a] fault plane solution to resolve the shear stresses and total Coulomb failure stress on the fault plane only in western Shikoku. Seismic waveforms observed at a station above the triggered tremor region would be ideal for this purpose. We use the Hi-net data rather than the F-net data because the denser Hi-net network has more stations very near the triggered deep low-frequency source regions (Figure 3). The Hi-net seismometers have natural periods of 1 s, and they may not be suitable to get the actual particle motions for long-period waves such as from the Sumatra earthquake without additional calibration. By matching the Hi-net records with nearby F-net records, we calculate that the Hi-net amplitudes should be about 4.5 times as large as specified by the standard calibration. We also apply this calibration to make surface wave envelopes in Figure 7 but not in Figure 9 because we cannot calibrate the amplitude for a wide frequency range.

From the determination of the tremor locations shown in Figure 3, the hypocenters appear to locate at almost the same region in each of five clusters. The relocated hypocenters still have significant errors in depth. In sections 2.4.1 and 2.4.2 we consider two possibilities: Either the hypocenters are colocated in each cluster or they are not. We will show that the results are not sensitive to either assumption. As discussed above, we use the horizontal components for the high-frequency observations in order to focus on the clearest, largest packets of triggered tremor for each.

2.4.1. Assuming Tightly Clustered Events

Here we assume that in each cluster of events, individual sources are separated by much less than the wavelength of the incoming surface waves. We use waveforms observed at 5 Hi-net stations shown in Figure 3 (diamonds). The source location is assumed to be the mean location of each cluster.

We then compare the tremor amplitudes with the triggering strain/stress changes by shifting the observed surface waves at the station closest to the epicenter (corrected for depth) to the source location. The time shift is computed by combining the traveltime for the tremor and the phase shift of the surface waves resulting from the propagation from the tremor epicenter to the observation station.

Miyazawa and Mori [2006] found that triggering correlated with peak amplitude of volumetric strain changes. When we relocated the hypocenters, the result is almost the same. Figure 11 shows the relationships between the strain changes at the source region and the deep low-frequency signals at the 5 regions. The strain changes span \( \pm 3 \times 10^{-7} \). In 4 regions except for the eastern Shikoku region (IKWH), we find a clear relationship that large dilatation excites both large and small tremor signals while the contraction only excites small tremor. At eastern Shikoku, the trends seem to be opposite to the other regions. However, the relocated hypocenters are not robust and include large location errors, because the signal-to-noise ratios are very poor among the 5 regions (Figure 7), and the result may not be resolved. As a result, the mean tremor amplitude generally increases with volumetric strain change (solid line) as does the variance (shaded region).

In Figure 11 (right), the averages of signal amplitudes are fit by two- and three-dimensional polynomials and exponential functions for volumetric strain changes. An \( n \)-dimensional polynomial for \( x \) is given by \( \sum_{i=0}^{n} a_i x^i \), where larger \( n \) fits the data better. An exponential function for \( x \) is given by \( a \exp(bx) + c \). The standard deviations of the
residuals between the functions and observations are indicated in parentheses. In every region, though the polynomial functions match the observations better, the differences are small. To address the small signal under the compression, the exponential model is also reasonable as the differences between models are small.

[35] Motivated by the correlation between acceleration and tremor spectra in Figure 9, we also obtain the relationship between the vertical acceleration at source regions and the deep low-frequency envelopes at the same 5 stations (Figure 12). We apply the same time shift procedure as above. The acceleration at depth is given by

\[
\dot{u}_z = \frac{\mu_{\text{eff}}}{\mu_s} \dot{u}_{\text{obs}}.
\]

The values are within ±1.3 mm/s² which is considerably smaller than gravitational acceleration. Because of the phase difference by \( \pi \) from \( u_z \) or \( e_{zz} \), the negatively large acceleration corresponds to the large signal amplitude except for IKWH. The standard deviations are large for large mean signal amplitudes and small for small ones. At URSH the peak value does not appear at the negatively largest acceleration as well as in Figure 11. From Figure 9a acceleration and Figure 9d KWBH, the peak values seem to correspond to each other and the acceleration has much better relationship with the event amplitude than dilatation \( \Delta V/V \). Nevertheless we find large variances as we did for the dilatation (Figure 11). In Figure 9d we draw the largest amplitude of triggering for period and exclude other small ones, which can cause the amplitude to be variable when we take all the signals into account (Figures 11 and 12). The acceleration may apparently and coincidentally explain the tremor amplitudes well.

[36] We now examine the relationship between the seismic wave stresses and the triggered tremor amplitude (Figures 13 and 14). For this exercise we use examples from western Shikoku where we assume the triggered tremor has the same mechanism as that by Ide et al. [2007a]. The Ide et al. mechanism is also consistent with the particle motion of these particular events (Figure 6). We compare the shear stress changes from the Rayleigh and Love waves separately to the triggered signal as well as the composite shear stress, normal stress, and Coulomb failure stress changes. For the purpose of this figure, the Coulomb failure stress change (\( \Delta \text{CFF} \)) is calculated from equation (10) with \( \mu = 0.2 \).

[37] There is an obvious and clear correlation between the tremor amplitude and the incoming stresses from the seismic waves. Large, positive stresses that promote shear failure result in large amplitude tremors. This is true for all three of the possible composite stresses (\( \Delta \text{CFF}, \Delta \sigma_s \), and \( \Delta r \)). For the Sumatra earthquake (Figure 13) it is also true for each of the surface waves viewed in isolation. However, for the Tokachi-oki (Figure 14), the Love waves by themselves do not give a strong correlation. This observation motivated some of the earlier work on triggered tremor [Miyazawa and Mori, 2006]. The normal stress variations cover a larger range than the shear stresses and thus provide a better resolved function in all cases.

[38] The correlation with all of the components of stress results in an inherent ambiguity in the stress field driving...
the tremor. To clarify this point, we measure the correlation between the tremor amplitude and the seismic wave stresses as a function of $m_0$ (Figure 15). If the normal stress changes alone are driving the tremor, as originally proposed by Miyazawa and Mori [2006], the correlation between the signals should increase with increasing values of $m_0$. For the Sumatra earthquake, the correlation is high for all physical values of $m_0$ with a slight preference for a very low apparent coefficient of friction (<0.2). Negative values of $m_0$ are also included here to illustrate that the statistical correlation reflects a physical effect that can be only modeled with realistic values of friction.

[39] For the Tokachi-oki earthquake, the total data set shows a relatively weak correlation. However, if the data are limited to the data after the first 450 s of Figure 8, which is after the passage of a long-period Rayleigh wave that did not trigger tremor, the correlation becomes as high as the Sumatra earthquake for any positive value of $m_0$. In this case, the data weakly prefer the larger values of friction that correspond to a normal stress-controlled triggering. It is worth noting that the Sumatra earthquake also triggered the largest tremor events during the Rayleigh waves late in the wave train rather than during the early Love waves (see Figure 1).

[40] Another complication is that the $S$ wave from the Tokachi-oki earthquake also generated large shear stress change along the slip direction with peak amplitudes as much as ~20 kPa. Even though shear stress changes are larger than that from the surface waves, there was no significant triggering at the arrival of $S$ waves (Figure 8). Therefore either normal stress control is important or the triggering simply failed to begin into relatively late in the wave train: after both the $S$ wave and the long-period Rayleigh waves. For the Sumatra earthquake, there was also no significant triggering by $S$ wave (Figure 1) because the generated peak shear stress change along the slip direction is smaller than ~1 kPa because of the fault geometry.

2.4.2. Assuming Multiple Locations

[41] The hypocenters obtained above (Figure 3) still have errors (especially in depth) and it is possible that their separation in each cluster is significant compared with the wavelength of the surface waves. To deal with this possibility, we pursue a different approach to the analysis. Instead of assuming that the pulses colocate, we consider each pulse separately as an individual tremor event that must be shifted in time relative to the others in order to derive the correct phase relationship with the incoming wave. To investigate the relationship as above, we should deterministically give appropriate time shifts for each tremor.

[42] We select well-relocated events with small location errors to minimize the time shift errors. We compare the beginning of the wave packet with a calculated strain change at the source. The symmetric nature of the wave packets and the slowly arising amplitude of the arrivals suggest that the tremor envelopes reflect source time functions (see section 2.1). However, we cannot discount the possibility that the envelope shape after the peak is independent from the effect of the structure during the propagation. Then it is reasonable to correlate the first half packet before the peak with the dilatation in this way. We neglect other factors including path and site effects and radiation patterns, which are unknown. Figure 16 shows the relationships, in which we use the 17 clear deep low-frequency tremor events (9 in western Shikoku, 1 in the Kii peninsula, 2 in western Tokai, and 5 in central Tokai) with large signal-to-noise ratios (for example, see the numbered events in Figure 7).

Figure 12. Relationships between the vertical acceleration at the source region and the deep low-frequency signals at five stations. The vertical axis indicates the signal amplitude of triggered deep low-frequency tremor and the horizontal axis indicates the vertical acceleration from Rayleigh wave. The shaded area shows the standard deviation of residuals between the observation and the curve. Time window used for the analysis is from 1300 to 2600 s.

[43] The tremor amplitudes are large for the positive volumetric dilatation and the vertical expansion, and for the radial compression. The general trend is the same as in
Figures 11 and 13. Since the location errors in depth still include about 5 km or more, even though we use the double-difference method for the relocation, the error in the applied time shift may be the cause of the more scattered results. As a result, it is more difficult to discern the exact relationship between triggering strain changes and tremor amplitudes than it was in section 2.4.1. Given the greater error in individual locations, we take the results here to

Figure 13. Relationships between triggered signal amplitude and stress changes on the fault plane due to surface waves from the 2004 Sumatra earthquake. $\Delta CFF$ is the Coulomb failure stress change, $\Delta \tau$ is shear stress change resolved in the slip direction, and $\Delta \sigma_n$ is normal stress change. $\Delta \tau$ consists of shear stress change from Love wave, $\Delta \tau_L$, and that from Rayleigh wave, $\Delta \tau_R$. The fault mechanism is the same as that of Ide et al. [2007a]. The relationships are in black solid lines with errors in shades and some are fit by exponential curves in white dashed lines. Time window is the same as in Figure 11.

Figure 14. Relationships between triggered signal amplitude and stress changes on the fault plane due to surface waves from the 2003 Tokachi-oki earthquake. Time window used for the analysis is from 350 to 600 s (Figure 8).
confirm our earlier interpretation with the caveat that the precise relationship between triggering strain changes and tremor amplitude may be much less well constrained if the tremors are not well located.

[44] The stress changes on the fault plane are also calculated as shown in section 2.4.1. We only verify that event 14 at western Shikoku (Figures 2, 5, 6, and 7) has the same source mechanism similar as shown by Ide et al. [2007a], although other well-relocated events may also have the similar mechanism. Figure 17 shows the stress changes on the fault plane and the amplitude of the event 14 observed at three stations with the epicentral distances less than 20 km (see Figure 5). (This data set includes KWBH that was featured in Figure 13.) The triggering is correlated with all of the stress changes including DCFF. Thus the results of section 2.4.1 are robust to the assumptions about the degree of clustering.

3. Discussion

[45] The key observations of triggered low-frequency tremor are the following:

[46] 1. The low-frequency tremor is triggered by dilatation. This dilatation with the vertical expansion and the radial compression can act directly or via increasing the Coulomb failure stress.

[47] 2. The amplitude of the tremor is an exponential function of the strength of the triggering stresses. Unlike earthquakes, the strength of the tremor appears to be size-predictable. We now pursue the mechanistic implications of these observations.

3.1. Triggering Mechanisms

3.1.1. Coulomb Failure

[48] A simple Coulomb failure model is the most obvious and easiest explanation for the data. The correlation between the tremor amplitude and the resolved Coulomb shear stress change is over 70% for a large range of parameters (Figure 15). Frustratingly, the data are unable to resolve the friction and is highly correlated with the DCFF for any positive value of $\mu$. 

[49] The strong correlation with dilatation that was noted in earlier work on the subject is completely consistent with the Coulomb failure model [Miyazawa and Mori, 2006]. The dilatation is an important part of the process as it unlocks the fault. The geometry of the fault and the resolved stresses along with the more extreme amplitudes of the Rayleigh waves make them a more direct predictor of failure than the S and Love waves in isolation. In particular, the normal stress resolved on the fault plane covers a much larger range of stresses than the shear. The range of normal stresses is about three times larger than that of the shear stresses.

[50] That said, Coulomb failure is not required by the data. A very large coefficient of friction, i.e., normal stress control, is also consistent with the data and even preferred by the Tokachi-oki earthquake (Figure 15). The lack of S wave correlation also suggests dilatational control. Given that petrological arguments suggest abundant fluids in the source zone, it is worth investigating whether the normal
stress might be able to trigger the tremor directly by interacting with the fluid. We pursue this possibility below.

[51] Simple, single process models like Coulomb failure cannot directly address the second major observation of this paper. The models predict failure of an indeterminate size as a result of a strain change, but we observe a correlation between amplitude of the triggered large event and the amplitude of dilatational strain and normal stress changes is demonstrated in Figures 11, 13, 14, and 15. As discussed below, this problem can be solved by introducing a distribution of triggered events.

3.1.2. Tensile Failure

[52] Dilatation exciting tremor might be explained if the source model of the deep low-frequency tremor is a tensile (mode I) crack. However the deep low-frequency tremor mainly produces shear waves. The opening crack is capable of producing shear waves but the amplitude is smaller than or comparable to $P$ waves. We could not find strong evidence that the observed waves included dominant $P$ waves. Therefore a shear stress in the source seems to be required.

3.1.3. Permeability Pumping

[53] Miyazawa and Mori [2006] suggest that the correlation of extensional stresses to the triggered tremor is a fingerprint of fluid flow, which weakens the friction of fractures, where the fluid is dehydrated from subducting slab. Our more thorough statistical analysis does not require such a model anymore, nor does it rule it out (see Figure 15). The normal stresses or dilatations still provide an adequate prediction of the occurrence of tremor. Therefore we flesh out the potential physics of such a mechanism below to provide a basis for testing the scenario that the geometry of future earthquakes allows us to resolve the distinction in stresses.

[54] Compressional normal strain change tends to close cracks and prevents fluid flow; expansion tends to open cracks and increases flow. In fracture-dominated systems, fluid flow rates are very sensitive to crack aperture. For instance, for infinite planar cracks in an otherwise impermeable medium, the permeability is proportional to $b^s$, where $b$ is the aperture [Snow, 1969]. Therefore the dilatation of the seismic waves can generate an oscillation of permeability and thus pump an oscillatory flow. The flow of fluid into a dry fault zone can then weaken the surface by increasing the pore pressure (equation (10)) and/or by suddenly reducing the friction coefficient on the fault, and promote shear failure (Figure 18). Also the increasing normal stress on the fault plays an important role. The dilatation determines the volume of the fluid injected into the fault region. Thus the dilatation as well as increasing normal stress may control the area of failure and hence the magnitude of the resulting slip event.

[55] This fluid-related process can explain why large shear stress change from $S$ wave cannot independently trigger the tremor even though this can also enlarge $\Delta CFF$ for the Tokachi-oki earthquake. During the arrival of the $S$ wave, the amplitudes of the normal strain changes are very small.

[56] This model is similar some ways to that proposed for long-range earthquake triggering by Brodsky and Prejean [2005]. Interestingly, the stresses inferred in Figures 13 and 14 are similar to the stresses necessary to trigger crustal earthquakes in Long Valley.

3.2. Distribution of Sources

[57] The problem of increasing stress increasing the amplitude of the tremor can also be addressed by considering a distribution of faults. Some faults might be closer to failure than others. Small strain changes will only affect the faults near failure while larger strain changes will affect both the ones far and near failure. Here we consider the failure as a function of the normal stress changes as they are the largest perturbing stresses in Figures 13 and 14. An analogous calculation can be done for the shear stresses.

[58] The nearly exponential dependence of signal amplitude on the input stresses implies an exponential distribution of failure stress, i.e., if the shear stresses on faults are distributed such that for stresses $\sigma$ less than the critical stress $\sigma_c$

$$n(\sigma) = C_1 \exp \left( C_2 \frac{\sigma_c - \sigma}{\sigma_c} \right),$$

where $n(\sigma)$ is the number of faults with stress $\sigma$ and $C_1$ and $C_2$ are constants, then the number of faults triggered by a change in effective stress $\Delta \sigma$ is found by integrating equation (17) from $\sigma_c - \Delta \sigma$ to $\sigma_c$. The result is

$$N(\Delta \sigma) = C_1 \frac{\sigma_c}{C_2} \exp \left( C_2 \frac{\Delta \sigma}{\sigma_c} \right) - 1,$$

where $N(\Delta \sigma)$ represents the relative fault area in a particular stress state. The observed tremor is the constructive sum of all the failure signals, so the more area that fails, the higher

![Figure 17](image-url)
the amplitude of the resulting tremor. The parameters $C_2 / \sigma_c = 3.40 \times 10^{-5}$ and $3.61 \times 10^{-5}$ Pa$^{-1}$ for normal stress changes at KWBH in Figures 13 and 14, respectively. At other three stations (HRKH, URSH, and NUKH), we also have the same ordered value inferred from the exponential curves in Figure 11, while at IKWH the average of triggering amplitudes against background noise is too poor to obtain some relationships. This distribution approach is similar to that taken by Dieterich [1994] and results in a similar distribution of failure strengths.

[59] To relate equation (18) to the rate state model, we can identify $\sigma_c / C_2$ with $A \sigma$, where $A$ is an empirically derived constant and $\sigma$ is the background stress. Dieterich [1994] derives seismicity rate change as a function of shear stress change divided by $A \sigma$ and the effective stress changes described here produce the similar effect. The combined value of $A \sigma$ of about $10^5$ Pa is similar to that inferred from crustal triggering studies [e.g., Toda and Stein, 2003], but the much greater depth of the triggering poses a problem. If $A$ is a nearly constant material property of order $10^{-5}$, then the background effective stress must be nearly 2 orders of magnitude below the lithostatic pressure at 30 km. Such a scenario would require extraordinarily high fluid pressures, so fluids would still be a key to the triggering process.

[60] Once the strain energy has been released, the triggering sequence is not expected at the same region unless some strain energies remain. The variable strength of the tremor as a result of identical strains/stresses in Figures 10, 11, 13, and 14 may indicate the possibility that each deep low-frequency tremor has slightly different hypocenter. There may be many strong and weak (or large and small) asperity regions, where the large and small strain energies, respectively, are to be released relatively slowly. The smaller dilatation is capable of moving smaller amount of fluid and triggering tremor, while the larger dilatation is capable of moving larger amount of fluid, providing high and various opportunities for the fluid to contact with both high- and low-strain preexisting regions to cause much larger tremor. This possibility is one of the reasons to explain large variance for large dilatations in Figure 11.

3.3. Nature of Slip

[61] In all of the proposed mechanisms, the low-frequency nature of the seismic signals is controlled by the slip process rather than the fluid flow. The frequency of shear failure events with a given seismic moment is determined by the stress drop and rupture velocity. If either quantity is unusually low, the seismograms will appear low-frequency for their amplitude. From spectra in Figure 4, it is difficult to estimate the corner frequency, but original waveforms show the tremor of $\sim 1$ Hz on which high frequency ($<15$ Hz) waves are superimposed. Using these observations, we could not conclude which factor contributes to the generation of low-frequency waves.

[62] Another clue to the origin of the low-frequency signals is that deep low-frequency events in western Japan and the similar episodic tremor-and-slip (ETS) event found in Cascadia subduction, where the feature of tremor is similar to the deep low-frequency event, have been observed accompanied with the slow slip events in each region [Obara et al., 2004; Rogers and Dragert, 2003], suggesting that the phenomenon occurs at the aseismic-seismic transition. In the rate state framework, the transition is where the slip is stable, $A \approx B$. If rupture velocities or stress drops are low in this transitional region, the slip events would radiate unusually long-period waves for their magnitudes.

[63] There are observations of deep low-frequency earthquakes related to slower strain changes [e.g., Obara et al., 2004; Miyazawa and Mori, 2005; Obara and Hirose, 2006]. Kao et al. [2006] propose that episodic tremor-and-slip events are excited by the procedure that dilatational strain field changes due to slow slip on the plate boundary cause fluid migration and excite the events. This observation relating quasi-static deformation to low-frequency tremor reinforces our connection of the triggered tremor to strain, rather than acceleration. We observed clear triggering when the normal strain changes are at least on the order of $10^{-8}$.

4. Conclusions

[64] Pulsating triggering of deep low-frequency tremor was observed in western Japan during arrivals of the
Rayleigh waves from the 2004 Sumatra-Andaman earthquake. The significant triggering is coincident with the large dilatation and thus large Coulomb failure stresses at the source region ~30 km deep. Short-period (15–30 s) Rayleigh waves triggered the events much more strongly than the first arrival long-period (>40 s) ones. Larger input stresses result in larger amplitude tremor. The shape of the tremor packets alone indicates that the tremor amplitude is predictable based on the strength of the incoming waves. More detailed analysis suggests that the mean amplitudes of triggered events increase exponentially with the strain change.

[65] Mechanically, the data are entirely consistent with tremor being generated by a series shear failures triggered by Coulomb failure. However, nothing in this particular data set lets us rule out more direct dilatation control of shear slip via permeability pumping where the expansion increases permeability and thus fluid flow to the fault resulting in slip. In either case, a distribution of sources can reproduce the correlation between the input stresses and the resulting amplitudes of the waves. In the case of a permeability pumping, the amplitude correlation may also happen directly for a single fracture due to the strong dependence of permeability on the dilatation. We are left with an array of possibilities and several useful tests for future triggering episodes to ultimately determine the source of the deep low-frequency tremor.

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References


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