Detection of “Rapid” Aseismic Slip at the Izu-Bonin Trench

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Abstract No great earthquakes have been historically documented at the Izu-Bonin Trench, where subduction is believed to occur largely by aseismic slip, although the details are poorly understood. We deployed an array of ocean bottom pressure gauges here for a year from May 2015. The array recorded the coseismic seafloor uplift/subsidence and tsunamis generated by the nearby Mw6.0 thrust earthquake. In association with this event, we detected two much larger aseismic slip events with rise times around 1 h. The total moment of these two aseismic events was 17 times larger than that of the mainshock. Such aseismic, yet still rapid, slip can be interpreted as one amid the transitional regime. This regime is expected to host slow slip events near its boundary with the stable sliding regime and, possibly, tsunami earthquakes and very low frequency earthquakes near the boundary with the unstable seismic slip regime. Slip in the transitional regime may be a prevalent mode of subduction in the Izu-Bonin trench, where effective normal force on the frictional plate interface is tectonically reduced.

Plain Language Summary Movement of tectonic plates occurs mainly by relative slip between two plates in frictional contact. Slip can be fast enough to radiate seismic waves as earthquake. Slip can also be slow enough to allow stable sliding at a speed expected from the plate tectonic theory. Relatively recent discovery of slow slip events with source durations ranging from days to years motivated seismologists to think about slip mode in the transitional regime between the unstable seismic regime and the stable sliding regime. Here, we report detection of slip events amid the transitional regime by using an ocean bottom array of absolute pressure gauges. The detected events were four orders of magnitude faster than previously reported slow slip events but four orders of magnitude slower than ordinary earthquakes. Our finding has thus provided a missing link in understanding the whole spectrum of slip modes.

1. Introduction

Subduction zones are often classified as Chilean-type or Mariana-type (Uyeda & Kamamori, 1979). The Izu-Bonin Trench is a typical example of the Mariana-type (Bilek & Lay, 2018; Scholz & Campos, 1995), where the two plates in contact are thought to be only weakly coupled, making their relative slip largely aseismic. The degree of seismic coupling is often measured using the seismic coupling coefficient, defined as the ratio of the long-term seismic slip rate to the relative plate velocity (Kato & Hirase, 1997; Pacheco et al., 1993; Peterson & Seno, 1984; Scholz & Campos, 1995). For example, the 80 years-seismicity-based estimate of the coupling coefficient is 0.01 for Izu-Bonin and Mariana, in contrast to the estimates of 1.57 for south Chile and 0.84 for east Aleutians (Peterson & Seno, 1984), although there might still be a possibility of future M ≥ 9 earthquakes at trenches classified as Mariana-type (Ikuta et al., 2015; McCaffrey, 1997).

Aseismic slip includes stationary sliding at the relative plate velocity and transient slip at a faster speed. The transient slip may occur spontaneously as slow slip events (SSEs) or result from postseismic relaxation as aftserslip (Avouac, 2015; Helmstetter & Shaw, 2009; Jolivet & Frank, 2020; Obara & Kato, 2016). Both phenomena involve time scales of days to years. For example, Schwartz and Rokosky (2007) listed 21 subduction-zone SSEs with displacements of 1–50 cm and with durations in a range from 3 to 2,000 days (median = 40 days). They also listed 10 subduction-zone aftserslip events with displacements of 30–200 cm, for which the durations ranged from 1 to 1,400 days (median = 83 days). The temporal growth of aftserslip is in general well described by a logarithmic curve (Heck et al., 1997; Marone et al., 1991). Our objective here...
is to address what aseismic process occurs at the Izu-Bonin Trench and how different this process is from previously reported slow slip or afterslip.

2. Observations

In May 2015, we installed an array of stations (B01–B10) on the relatively flat seafloor at depths of around 5,000 m to the west of the steep slope of the northern Bonin Trench (Figure 1 and Table 1). The array consisted of equilateral triangles with the minimum and maximum side-lengths of 10 and 30 km, respectively. Each station deployed a free-fall/pop-up ocean bottom pressure-meter, equipped with an absolute pressure gauge (APG) (Absolute depth sensor PARO-8B7000-I-005, Paroscientific, Inc., Redmond, WA, US), which nominally had an accuracy of 0.01% and a resolution of $1 \times 10^{-8}$ for absolute pressure at water depths of 7,000 m (Fukao et al., 2018; see also Wallace et al., 2016 for use of the APG array to detect seafloor deformation). The APG records were sampled at 4 Hz with a cutoff frequency of 0.7 Hz. All data were successfully recovered except for two. A malfunction was found for the APG at B05, where a broadband ocean bottom seismometer and a differential pressure gauge with a sampling rate of 100 Hz were installed instead. Data were poorly recovered from B04 because of the logger system malfunction.

The array recorded a nearby thrust earthquake (M6.0) of 01 September 2015, the foreshock (M5.6) about 1 min before the mainshock, and numerous aftershocks, most of which were too small to be reported by the U.S. Geological Survey (USGS). Among these aftershocks, the event 3.5 days
after the mainshock is of particular interest as it accompanied a significant aseismic event. We determined the hypocenters of this aftershock and the foreshock based on the first arrival data from the nine stations (all but B04) (Table 2), using a one-dimensional P-wave velocity model constructed by referring to several nearby seismic profiles (Kamimura et al., 2002; Takahashi et al., 2015). The first arrivals of the mainshock were difficult to identify in conjunction with the foreshock disturbance, so the parameters of the mainshock were taken from the global-centroid-moment-tensor solution (GCMT, http://www.globalcmt.org/CMTsearch.html). The foreshock was located near the northern boundary of the estimated slip plane of the mainshock. The aftershock was located slightly further north (Figure 1). The magnitude of the aftershock was estimated to be around M3.2 by comparing the maximum amplitude on the broadband seismogram at B05 to those of three nearby larger events with known magnitudes (M4.8, 4.6, and 4.8 according to the USGS).

### 3. Data Processing and Analysis

#### 3.1. Tide-Removal

We first identified several anomalous events by visual inspection of the 1-year records. They included the targeted aseismic signals on the order of a few hPa in amplitude and on the order of 1 h in time scale. Such

| Information     | Date     | Origin time h:m:s GMT | Mag.   | Lat., N Deg. | Lon., E Deg. | Depth km | Travel time residual s |
|-----------------|----------|------------------------|--------|--------------|--------------|----------|------------------------|
| Foreshock       | 2015/9/1 | 15:24:09.7             | M5.6   | 31.34° ± 5.2 km | 141.84° ± 3.0 km | 9.7 ± 6.0 | 0.22                    |
| Mainshock       | GCMT     | 2015/9/1               | Mw5.8  | 31.16°         | 141.80°       | 12       |                         |
| Aftershock      | 2015/9/5 | 01:51:34.0             | M3.2   | 31.40° ± 7.5 km | 141.78° ± 2.3 km | 9.6 ± 6.9 | 0.87                    |

Table 2

Hypocentral Parameters

Note. Foreshock: The hypocenter is determined in this study. The magnitude is referred to USGS. Mainshock: The centroid parameters are referred to GCMT. The Mw value reported by USGS is 6.0. Aftershock: The hypocenter and magnitude are determined in this study.

![Figure 2](image)

Figure 2. Original APG record at B10, the closest station to the M6.0 mainshock. (a) 1-day record. (b) 120-s record. The mainshock was preceded by about 1 min by the M5.6 foreshock.
signals are largely masked by the greater amplitudes of tidal oscillations and high-frequency seismic waves (Figure 2). Tidal oscillations were removed in a time window of 40,000 s that consisted of three sections (Figure 3). The central section of a time interval of 10,000 s includes the targeted signal in it. The former and latter sections of intervals of 15,000 s were assumed to be free of any signals other than the tide. The tidal signal was approximated by a set of harmonic oscillations with an equal frequency interval. Their amplitudes were determined by least-squares-fitting using the records outside the central section. Offsetting of the baseline across the central section was allowed, which provides a measure of the permanent sea-

Figure 3. Removal of tide to retrieve aseismic signal from the record at station B10. A time window of 40,000 s consists of three sections as indicated by two vertical green lines. The central section includes the targeted signal. Tidal fitting was made to the former and latter sections, where different baselines were allowed as shown by two horizontal gray lines. (a) The 50s lowpass-filtered version of the original record (black), the estimated tidal curve (red), and the residual trace (blue) for the first aseismic event. Lowpass-filtering was made here for ease of illustration. Time 0 corresponds to 40,000 s of 01 September 2015. The arrow indicates the M6.0 mainshock. (b) The original record (black), the estimated tidal curve (red), and the residual trace (blue) for the second aseismic event. Time 0 corresponds to 74,000 s of 04 September 2015. The arrow indicates the M3.2 aftershock.
The impact of the foreshock was approximately accounted for. The reported focal mechanism was simplified as a pure dip-slip fault, striking in the N-S direction (approximately parallel to the trench axis) and dipping westward at 16° along the seismologically inferred plate boundary (Iwasaki et al., 2015). We also assumed a uniform source time function with different amplitudes among different fault segments with an equal area of 4 × 4 km². Figure 5 shows the slip-distribution model we obtained. The centroid was located roughly under station B10. The agreement between the observed waveforms leaving the cumulative offsets and those calculated for this slip model is remarkable (Figure 4).

The sum of the distributed slips gives a seismic moment of 0.95 × 10¹⁸ Nm, assuming a crustal rigidity of 30 GPa (Lay et al., 2012). This moment is in between the GCMT value of 0.73 × 10¹⁸ Nm and the UGGS value of 1.25 × 10¹⁸ Nm and corresponds to Mw6.0. The effective fault area was defined as the assembly of fault segments whose displacements were more than 20% of their maximum. This area, 210 km², was used to define the average displacement consistent with the seismic moment. It was also used to estimate the stress drop assuming a circular fault having the same boundary (Iwasaki et al., 2015). We also assumed a uniform source time function with different amplitudes among different fault segments with an equal area of 4 × 4 km². Figure 5 shows the slip-distribution model we obtained. The centroid was located roughly under station B10. The agreement between the observed waveforms leaving the cumulative offsets and those calculated for this slip model is remarkable (Figure 4).

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### 3.3. Aseismic Events Analysis

Figure 6 shows the 600 s lowpass filtered tide-removed traces exhibiting an aseismic event in the central section of the time window. In Figure 6a, slow seafloor uplifting (the first aseismic event) started with the occurrence of the mainshock. To extract the aseismic contribution, we removed the impact of the mainshock by using the slip distribution model of the

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**Table 3**

| Station | 1st event | 2nd event |
|---------|-----------|-----------|
|         | Central offset | Former RMS | Latter RMS | Central offset | Former RMS | Latter RMS |
| B01     | 296       | 28        | 31        | 162        | 13        | 11        |
| B02     | 92        | 29        | 32        | 205        | 10        | 10        |
| B03     | 113       | 32        | 37        | 249        | 12        | 12        |
| B06     | 270       | 32        | 33        | 389        | 15        | 12        |
| B07     | 477       | 31        | 30        | 230        | 16        | 15        |
| B08     | 201       | 30        | 31        | 384        | 14        | 11        |
| B09     | 75        | 28        | 33        | 503        | 10        | 18        |
| B10     | 464       | 30        | 30        | 421        | 12        | 13        |

*Note. RMS: Root-mean-square of residuals of observed record from its fitting curve. Fitting was made using data in the former and latter sections. Data in the central section were not used in tidal fitting.*

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**Figure 4.** Observed (black) and calculated (red), 30-s lowpass filtered, tide-removed records of the mainshock. The records show the dynamic and static responses of seawater to coseismic seafloor uplift/subsidence, including the tsunami generation and propagation. The synthetic records were calculated based on the slip distribution model of the mainshock (Figure 5). The impact of the foreshock was approximately accounted for.
mainshock (Figure 5). In Figure 6b, slow seafloor uplifting (the second asismic event) started about 3,000 s before the M3.2 aftershock but was abruptly accelerated upon its occurrence. We inverted these traces to obtain the slip-growth models of the first and second asismic events. Slip was assumed to be slow enough to allow a static field treatment (Oka-da, 1992) so that the inversion could be made independently at each time step of pressure records. The slip, dip, and strike directions were taken to be the same as those of the mainshock slip model. No branching of the fault was assumed. A total of 60 km of fault length, with displacement varying in the dip direction of the plate boundary, was a required feature to explain the observation. Accordingly, the fault plane was divided into three segments with an along-dip length of 20 km. An along-strike width of 20 km was chosen among trial values of 10, 20, 30, and 40 km as the one with which the model gave the largest variance reduction for the observed records. The center of the upper fault segment was placed at a depth of 10 km, beneath station B10. Those of the middle and lower segments were then at depths of about 15 and 20 km, respectively (Figure 7).

Let $\mathbf{P}_j(t)$ be the displacement of the $k$th fault segment at time $t$, where $k = 1, 2$ or 3 for the upper, middle or lower segment. Our objective was to find the set of $\mathbf{U}_k(t)$ that minimizes the difference between $\mathbf{P}_j(t)$ and $\mathbf{P}_j(t)$ in a least squares sense. The size of the difference between $\mathbf{P}_j(t)$ and $\mathbf{P}_j(t)$ were measured by the variance reduction, $VR(t)$, defined as:

$$VR(t) = 1 - \frac{\sum [P_{j}^\text{obs}(t) - P_{j}^\text{cal}(t)]^2}{\sum [P_{j}^\text{obs}(t)]^2}.$$  \hspace{1cm} (1)

Variance reduction in this definition was expected to remain unstable until asismic slip starts. To avoid such instability, $VR(t)$ at $t < t_o$ was defined as

$$VR(t) = 1 - \frac{\sum [P_{j}^\text{obs}(t) - \hat{P}_{j}^\text{obs}(t)] - [P_{j}^\text{cal}(t) - \hat{P}_{j}^\text{cal}(t)]^2}{\sum [P_{j}^\text{obs}(t) - \hat{P}_{j}^\text{obs}(t)]^2},$$  \hspace{1cm} (1')

Table 4

|               | $A_k$ cm | $B_k$ cm | $T_k$ s | Final slip cm | Duration s | Fault area km$^2$ | Moment 10$^{18}$ Nm | Stress drop MPa | Rupture velocity km/s |
|---------------|----------|----------|---------|---------------|------------|-------------------|----------------------|------------------|----------------------|
| Mainshock     |          |          |         |               |            |                   |                      |                  |                      |
| 1st asismic slip | k = 1  | 0.20     | 17      | 4,600         | <30        | 210               | 0.95                 | 0.77             | 3.6                  |
|               | k = 2  | 2.8      | 3.6     | 1,400         | 17         | 9,200             | 20 x 20              | 2.0              | 0.62                 |
|               | k = 3  | 0.0      | 44      | 3,800         | 6.4        | 2,900             | 20 x 20              | 0.77             | 0.23                 |
| 2nd asismic slip | k = 1  | 1.7      | 10      | 1,500         | 44         | 7,600             | 20 x 20              | 5.3              | 1.6                  |
|               | k = 2  | 5.3      | 19      | 2,200         | 12         | 3,000             | 20 x 20              | 3.0              | 0.90                 |
|               | k = 3  | 3.1      | 30      | 2,200         | 25         | 4,300             | 20 x 20              | 4.0              | 1.2                  |

Note. $k = 1, 2, 3$ for the upper, middle, and lower fault segments (Figure 7).

*A default option given by Denolle and Shearer (2016).*
where \( t_o \) was chosen to be the origin time of the mainshock for the first aseismic event and the origin time of the mainshock for the second aseismic event. The quantities with hat represent averages throughout the time window. Figures 8a and 8b show the estimated \( U_k(t) \) for the first and second aseismic events, respectively. As expected, the estimated \( U_k(t) \) was close to zero in the former 15,000-s interval of the time window and close to a nonzero constant in the latter 15,000-s interval. For the first aseismic event, the largest slip occurred in the lower segment \((k = 3)\) and the smallest slip in the middle segment \((k = 2)\). The VR(t) for this slip model was about 90% throughout the time window (Figure 8c). For the second aseismic event, the slip was largest in the lower segment \((k = 3)\) and smallest in the upper segment \((k = 1)\). The VR(t) was as high as 95% generally through the time window (Figure 8d). In Figures 6a and 6b, the pressure changes calculated for these slip models were compared to the observed changes. The agreement was in general very good. A noticeable exception was the first aseismic event at B01, where the observed uplift was significantly larger than the calculated uplift, implying an additional local source at a shallow depth beneath this station, possibly within the outer wedge of the overriding plate.

The temporal variation of \( U_k(t) \) was fitted by the two types of model slip function (Shen et al., 1994). The first was of exponential-decay type as used for modeling seismic slip process (Brune, 1970):

\[
U_k^I(t) = A_k + B_k \left[ 1 - \exp \left( -\frac{t - t_o}{T_k^I} \right) \right],
\]

(2)

where \( T_k^I \) is the rise time for the \( k \)th fault segment. Fitting was made for \( t \geq t_o \), where \( t_o \) was taken to be the origin time of either the mainshock or the M3.2 aftershock. The initial slip velocity is given as

\[
\dot{U}_k^I(t_o) = \frac{B_k}{T_k^I}.
\]

(3)

The second is of logarithmic type as used for modeling afterslip process (Marone et al., 1991):

\[
U_k^H(t) = C_k + D_k \cdot \ln \left( 1 + \frac{t - t_o}{T_k^H} \right).
\]

(4)

The initial slip velocity is given by
The short-tailed nature of slip function \( U_k^I (t) \) is in sharp contrast to the long-tailed nature of \( U_k^II (t) \). The values of \( A_k, B_k \) and \( T_k^I \) were determined by minimizing the difference between \( U_k^I (t) \) and \( U_k^I (t \geq t_0) \) in a least squares sense. Similarly, the values of \( C_k, D_k \) and \( T_k^II \) were determined by minimizing the difference between \( U_k^II (t) \) and \( U_k^II (t \geq t_0) \). Figure 9 demonstrates how better the \( U_k^I (t) \) fits to the \( U_k^I (t) \) than the \( U_k^II (t) \) for any \( k \) (\( =1, 2 \) or \( 3 \)) and for either of the first and second aseismic events. In what follows we used the results with \( U_k^I (t) \) rather than \( U_k^II (t) \). For example, the final displacement of the \( kth \) fault segment was given as \( U_k^I (\infty) = A_k + B_k \). Figure 10 schematically summarizes the slip distributions along the plate boundary for the mainshock and the first and second aseismic slip events.

4. Results and Implications

Table 4 summarizes the fault parameters estimated for the mainshock and the first and second aseismic events. The parameters for the mainshock were better constrained because they were based on both the dynamic and static responses of seawater to the seafloor disturbances. The listed parameters include fault area \( S \), final slip \( U_k^I (\infty) \), seismic moment \( M_w \), stress drop \( \Delta \sigma \), rise time \( T_k^I \), source duration \( T_k \) and rupture velocity \( V_r \) defined as the ratio of fault length to source duration (Gao et al., 2012). Source duration in our study was defined as \( 2T_k^I \) at which slip \( U_k^I (t) \) attains 87\% of its final value. The total seismic moment of the three fault segments for either of the two aseismic events was more than eight times as large as that of the mainshock, whereas the slip-weighted average of stress drop (Noda et al., 2013) was on the order of 1 MPa, comparable to that of the mainshock. The stress drops of the mainshock and two aseismic events were lower than the global average of seismic stress drop but were not unusually low (Allmann & Shearer, 2009; Denolle & Shearer, 2016). The slip-weighted average of rise time was on the order of 1 h, much longer than the reported half duration of 2 s of the mainshock (GCMT).

The aseismic events we detected were not as rapid as seismic slip but nor were they as slow as geodetically detected slow slip or afterslip. Their slip histories were better fitted by the short-tailed exponential-decay functions than the long-tailed logarithmic functions (Figure 9). This is opposite to the cases of observed afterslip where the temporal behavior of slip is better represented by the form \( \log(t) \) than the form \( \exp(\text{–}t/T) \) (Savage & Svarc, 2005; see also Avouac, 2015; Fukuda et al., 2013; Heki et al., 1997; Hsu et al., 2006; Hutton et al., 2001; Marone et al., 1991). Avouac (2015) pointed out that for \( M_w > 7.5 \) events, the moment of after-slip is in general 0.1–0.4 of the coseismic moment. In our case, to the contrast, the total moment of the two aseismic events was more than 17 times as large as that of the mainshock (Table 4). The above line of argument implies that events we detected were spontaneously generated slips releasing tectonic stresses accumulated in the seismogenic zone rather than afterslip relaxing stresses produced by the mainshock within velocity-strengthening areas of the plate interface. They might have been triggered by either the afterslip of the mainshock or the mainshock itself as in the case of SSEs triggered by a nearby great earthquake (Kato et al., 2014). Figure 11 demonstrates how unique these aseismic events were among other stress drop events. In Figure 11a, previously reported source durations, \( T_p \), were plotted against seismic moments, \( M_w \) (Denolle & Shearer, 2016; Gao et al., 2012; Gomberg et al., 2016), where our observations were included. The slip-weighted average of \( T_p \) was about 7,600 s for the first aseismic event \( (M_w \sim 8.1 \times 10^{18} \text{Nm}) \) and about 4,200 s for the second aseismic event \( (M_w \sim 8.4 \times 10^{18} \text{Nm}) \). The plots fell neatly into the observational gap in source durations between earthquakes (typically shorter than a minute) and slow slips (typically longer than a day). In Figure 11b, previously reported values of rupture velocity, \( V_r \), were plotted against reported
stress drop values, $\Delta \sigma$ (Denolle & Shearer, 2016; Gao et al., 2012; Gomberg et al., 2016). The plots include our data, where $V_r$ was taken to be 3.6 km/s for the mainshock (Denolle & Shearer, 2016). The slip-weighted averages of $V_r$ and $\Delta \sigma$ were 2.5 m/s and 1.2 MPa for the first aseismic event, and 4.9 m/s and 1.0 MPa for the second aseismic event. Our data, again, fell well into the observational gap in the $V_r - \Delta \sigma$ relationship, between ordinary earthquakes (Denolle & Shearer, 2016; Kanamori & Anderson, 1975) and slow slips (Gao et al., 2012; Gomberg et al., 2016).

**Figure 8.** Slip histories of the three fault segments and the associated variance reductions. The slip values at each time step were obtained by inverting the data only at that time step. The upper, middle and lower segments are marked by different colors by referring to Figure 7. (a) Slip histories for the first aseismic event. The synthetic records for this model are compared with the observed records in Figure 6a, taking the mainshock contribution into account. (b) Slip histories for the second aseismic event. The synthetic records for this model are compared with the observed records in Figure 6b. (c) Variance reductions by the model of the first aseismic event. (d) Variance reductions by the model of the second aseismic event. In panels (a and c), the vertical dashed line indicates the origin time of the mainshock. In panels (b and d), the vertical dashed line indicates the origin time of the aftershock. The definition of variance reduction is different to the left (gray) and right (black) sides of the vertical dashed line.
In the framework of rate- and state-dependent friction law, the stable sliding (creeping) regime and the unstable stick-slip (earthquake) regime can be separated by the $W/h^*$ ratio (Liu & Rice, 2005; Rubin, 2008; Wu & Chen, 2014), which is almost equivalent to the $R_u$ number (Dieterich-Ruina-Rice number) defined as the ratio of the critical stiffness, $k_c$, to the stiffness, $k$, of the fault system (Barbot, 2019). More specifically,

$$R_u = \frac{k_c}{k} = \frac{(b-a)\sigma_n}{(G/W)} \sim \left( W/h^* \right).$$

(6)
where \(b-a\) is the friction parameter, \(\sigma_w\) the effective normal stress (actual normal stress less pore pressure), and \(d\), the characteristic weakening distance, whereas \(G\) is the rigidity and \(W\) the fault asperity size (Ruina, 1983; Scholz, 2002). The nucleation size, \(h^*\), is a quantity introduced by Rice (1993) to define a critical size for an instability to nucleate in the rate-weakening region, where \(b-a > 0\). If \(h^*\) is sufficiently small compared to \(W\), slip would have enough room to expand unstably as earthquake (Liu & Rice, 2005). If \(h^*\) is sufficiently large relative to \(W\), slip would have little room to be accelerated, so it would remain as creeping. Thus, the \(W/h^*\) ratio (or the \(R_u\) number) divides slip mode into the unstable seismic slip regime and the stable sliding regime with the transitional regime in between (Barbot, 2019). Figure 12 schematically illustrates how slip velocity changes when \(R_u\) varies across the transitional regime while the ratio of the frictional parameters, \((b-a)/b\), is kept constant. If, for example, this ratio is reduced toward zero, the range of \(R_u\) prone to generating slow slip events increases (Barbot, 2019). As shown in this figure, the transitional regime hosts SSEs on the lower-\(R_u\) side near its boundary with the stable sliding regime. The figure also suggests that the transitional regime hosts very low frequency earthquakes (VLFEs) (Fukao & Kanjo, 2007; Sugioka et al., 2012) and tsunami earthquakes (Bilek & Lay, 2002; Kanamori, 1972) on the higher-\(R_u\) side near its boundary with the unstable seismic slip regime. The events we detected are identified as those amid the transitional regime, where slip is accelerated to an intermediate level between SSEs and VLFEs. One of the three off Sanriku ultraslow earthquakes identified by Kawasaki et al. (1995, 2001) had a time constant of 1 day for a seismic moment of \((1-4) \times 10^{20} \text{Nm}. This event may be categorized as one in the transitional regime (see Figure 11a). Events in this regime are rare mainly because the transition from the stable to unstable mode occurs sharply (Figure 12). Existence of the transitional regime implies the absence of an intrinsic gap of slip mode (Peng & Gomberg, 2010; Romanet et al., 2018). Ide et al. (2007, 2008) and Ide (2014), on the other hand, argued for the real gap between the fast and slow slip modes which follow mutually distinct scaling laws. See also Nanjundiah et al. (2020) for the variety of source durations of slow slip events. Clearly, more work is needed to resolve the issue.

The \(W/h^*\) ratio (or the \(R_u\) number) includes the term \(\sigma_w W\) (Equation 6) which is understood as the effective normal force on the plate interface per unit length along the trench axis. The tectonic contribution to this term can be estimated by examining the force balance among plates associated with subduction (Scholz & Campos, 1995). The estimated contribution for the Izu-Bonin trench is a significant reduction of \(\sigma_w W\) (Scholz & Campos, 1995) by which the slip mode may well be shifted from the unstable seismic regime to the transitional regime (Figure 12). Slip in the transitional regime may be a prevalent mode of subduction in the Izu-Bonin Trench and other Mariana-type trenches.

5. Conclusions

Slip mode changes sharply from the unstable seismic regime to stable sliding regime across the transitional regime when the \(R_u\) number (or \(W/h^*\) ratio) (Equation 6) is reduced (Barbot, 2019). Detection of slip events amid the transitional regime has been challenging partly because this regime exists only in a narrowly limited range of \(R_u\). There has also been an observational challenge such that the involved characteristic time is too long to detect by ordinary seismological means but too short to monitor by conventional GPS networks. Our finding implies that there is no intrinsic gap in slip mode across the transitional regime when \(R_u\) is changed from the unstable seismic regime to the stable frictional sliding regime (Peng & Gomberg, 2010).
We suggest that the transitional regime hosts very low-frequency earthquakes (VLFEs) and tsunami earthquakes near its boundary on the higher $R_u$ side and slow slip events (SSEs) near the boundary on the lower $R_u$ side. The transitional regime is a key to understand the whole spectrum of slip mode.

Figure 11. Source parameters of the mainshock (black circle), the first aseismic event (three red circles for the three fault segments) and the second aseismic event (three hollowed red circle for the three fault segments). Comparisons are made with ordinary earthquakes (Denolle & Shearer, 2016) and SSEs. The plots for SSEs are based on Table S1 of Gao et al. (2012) but limited to samples from the three subduction zones, Cascadia, Japan, and Mexico, where the most abundant data were available. For the Japanese data, only those re-examined in Table S1 of Gomberg et al. (2016) are plotted. (a) Source duration ($T_d$) versus seismic moment ($M_o$). (b) Rupture velocity ($V_r$) versus stress drop ($\Delta \sigma$). Either $M_o$ in panel (a) or $\Delta \sigma$ in panel (b) can be linearly scaled with the final slip of each segment.

Figure 12. Schematic illustration for how slip mode changes when the $R_u$ number is varied while the frictional parameter ratio, $(b-a)/b$, is kept constant (Barbot, 2019).
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Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Data Availability Statement

The pressure data used in the analyses is available on Zenodo (https://doi.org/10.5281/zenodo.5139450).

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