Spatial Patterns in Frictional Behavior of Sediments Along the Kumano Transect in the Nankai Trough

Hanaya Okuda\(^1\),\(^2\), Matt J. Ikari\(^3\),\(^4\), Alexander Roesner\(^3\),\(^5\), Katja Stanislowski\(^3\),\(^6\), Andre Hüpers\(^3\), Asuka Yamaguchi\(^1\),\(^2\),\(^7\), and Achim J. Kopf\(^3\)

\(^1\)Department of Ocean Floor Geoscience, Atmosphere and Ocean Research Institute, University of Tokyo, Chiba, Japan, \(^2\)Department of Earth and Planetary Science, University of Tokyo, Tokyo, Japan, \(^3\)MARUM—Center for Marine Environmental Sciences, University of Bremen, Bremen, Germany

Abstract  Fault slip activity at subduction zones is governed by sediment frictional properties, which in turn are affected by diagenetic processes. To study the spatial patterns in frictional properties across the Nankai Trough, SW Japan, and their relations to fault slip activity, we used sediment samples (10%–59% clay mineral content) along the Kumano transect covering a large spatial range from the inputs via the outer prism to the inner prism, including the deepest sample ever recovered to date. We performed laboratory friction experiments under \textit{in situ} effective normal stresses and seawater-saturated conditions. Our results generally demonstrate that the friction coefficient inversely correlates with the clay mineral content. However, the outer prism sediments show higher friction coefficients than sediments from the other locations for a comparable clay mineral content. All samples show velocity-weakening behavior at low velocities, but the outer prism and the deep inner prism sediments show velocity strengthening at higher velocities. Based on the experimental results combined with a Coulomb wedge model, we propose that the lowest friction coefficient on the décollement occurs beneath the trenchward portion of the outer prism, whereas the minimum friction coefficient of the prism sediment occurs in the landward portion of the outer prism. In addition, the calculated critical nucleation length for slip instability suggests that the décollement beneath the outer prism area is more frictionally unstable than it is beneath the inner prism. This inference is consistent with the spatial distribution of very-low-frequency earthquakes and slow slip events along the shallow Nankai Trough.

Plain Language Summary  Subduction zones host the largest earthquakes on earth, and such earthquakes cause severe damage to regions nearby. Therefore, how and where an earthquake is triggered are the most important questions for the future mitigation of seismic hazards. An important factor controlling seismicity is the frictional behavior of the subduction zone fault. In this study, we study the Nankai subduction zone, SW Japan, which hosts both fast and slow earthquakes. Conducting laboratory friction experiments, we examine the spatial patterns in frictional properties of sub-seafloor sediments sampled by ocean drilling. Experimental results indicate that the friction coefficients of sediments vary with the position in the subduction system, which may reflect diagenetic processes or tectonic history. Combined with the topographical information, we estimate the friction coefficient of the plate boundary fault (décollement) and propose that the friction coefficient increases landward. Our results can provide useful information on the location of future fault slip activities at the Nankai Trough.

1. Introduction  Megathrust earthquakes at subduction margins induce hazardous ground motion and tsunami. The Nankai Trough, where the Philippine Sea plate subducts beneath the Amurian microplate offshore SW Japan at a convergence rate of 4.0–6.0 cm/yr (Miyazaki & Heki, 2001; Seno et al., 1993; Yasuda et al., 2017), has hosted large megathrust earthquakes such as the 1944 Tonankai (Mw 8.1) and 1946 Nankaido (Mw 8.3) earthquakes (Ando, 1975; Garrett et al., 2016). Two major domains characterize the geological structure of the accretionary prism at the Nankai Trough. The seaward domain is the outer prism, which is accreted sediment containing imbricate faults. The near-trench part of the outer prism with a steep seafloor slope is referred to as the “prism toe” region. The landward domain is the inner prism, which consists of older prism sediment overlain by the forearc basin sediment with a flat seafloor. The décollement develops beneath the outer and inner prisms as the plate boundary, and a megasplay fault branches from the décollement,
separating the two domains of the prism (Kimura et al., 2007). Both the megasplay fault and décollement have possibly hosted coseismic events as suggested by vitrinite reflectance geothermometry (Sakaguchi et al., 2011) and progressive illitization along the fault (Yamaguchi et al., 2011). Observed slow earthquakes such as very-low-frequency earthquakes (VLFE) and slow slip events (SSE) are also inferred to have occurred on the décollement (Araki et al., 2017; Ariyoshi et al., 2021; Ito & Obara, 2006; Nakano et al., 2018; Shiraishi et al., 2020; Sugioka et al., 2012). Because slow earthquakes may trigger large-magnitude fast earthquakes (Obara & Kato, 2016), the information on how those faults behave is important to understand fault slip activities.

To reveal the fault mechanics and explain fault slip activities at the Nankai Trough subduction zone system, the Nankai Trough Seismogenic Zone Experiment (NanTroSEIZE) project of the Integrated Ocean Drilling Program/International Ocean Discovery Program (IODP) drilled over 10 sites to measure physical properties and to collect sediment samples below seafloor along the Kumano transect offshore Kii Peninsula, Japan (Figure 1). Because friction directly controls the slip behavior of a fault (Marone, 1998; Marone & Saffer, 2007; Scholz, 2019), the frictional properties of recovered prism and input sediments have been studied through laboratory experiments (Ikari et al., 2009, 2013, 2018; Ikari & Hüpers, 2019; Ikari & Saffer, 2011; Roesner et al., 2020; Takahashi et al., 2013, 2014; Tsutsumi et al., 2011; Ujiie & Tsutsumi, 2010). Frictional properties are mostly controlled by lithological variations of sediments (Ikari et al., 2013, 2018;
In particular, the sediment’s clay mineral content is one of the important factors governing its friction coefficient. Experimental studies on synthetic gouge samples show an inverse relationship between friction coefficients and clay mineral contents (Logan & Rauenzahn, 1987; Saffer & Marone, 2003; Shimamoto & Logan, 1981; Takahashi et al., 2007; Tembe et al., 2010), which is also confirmed for sediments at various subduction zones including the Nankai Trough (Brown et al., 2003; Ikari et al., 2018; Kopf & Brown, 2003; Takahashi et al., 2013, 2014).

In addition to lithological composition, the variety of previous friction studies on samples from the Nankai Trough show that friction and its velocity dependent behavior, which determines whether the fault can nucleate an earthquake or not (Marone, 1998; Scholz, 1998), depend on experimental conditions such as the driving velocity, effective normal stress, and sample preparation. Experiments conducted at coseismic velocities (~1 m/s) show dynamic weakening behavior due to thermal pressurization (Tsutsumi et al., 2011; Ujiie & Tsutsumi, 2010). At lower velocities (~10^-4 m/s), fault materials generally exhibit low friction coefficients (Collettini et al., 2019; Ikari et al., 2009; Ikari & Saffer, 2011). Remolded sediments sheared at intermediate velocities (~10 μm/s) and above in situ stresses generally exhibit velocity-strengthening behavior (Ikari et al., 2009; Ikari & Saffer, 2011). Remolded sediments sheared at low velocity (as low as 0.002 μm/s) (Ikari & Kopf, 2017), or intact samples tested under in situ stress conditions (Roesner et al., 2020) reveal velocity-weakening behavior.

In a subduction zone system, spatial transitions in frictional properties are often postulated to explain the distribution of fault slip activities. As inferred from seismological observations and numerical studies of the thermal structure at plate boundaries, frictional properties may be affected by diagenetic processes and hence variations in physical properties during subduction and accretion (Hyndman et al., 1997; J. C. Moore & Saffer, 2001; Oleskevich et al., 1999; Vrolijk, 1990). This inference is supported by laboratory experiments, which suggest that the transition from velocity-strengthening to velocity-weakening behavior may be caused by processes related to diagenesis or metamorphism, such as porosity reduction due to lithification (Ikari & Hüpers, 2021; J. C. Moore & Saffer, 2001; Trütner et al., 2015), or to the temperature-dependent frictional behavior of illite-rich materials (den Hartog et al., 2012; den Hartog & Spiers, 2013). To corroborate these observational and experimental inferences, the spatial pattern of frictional properties in a subduction zone should be evaluated using measurements on natural sediment samples.

Since the NanTroSEIZE project has collected sediment samples from various locations along a transect of drill holes at the Nankai Trough, we can investigate the large-scale spatial patterns in frictional properties by using natural sediment samples. In this study, we perform friction experiments on sediment samples from three sites on the Kumano transect: the input sediment, the outer prism, and the inner prism, to reveal the variations in frictional properties of sediments across the shallow Nankai Trough. Such variations may represent the effect of past diagenetic processes during their accretion or subduction histories. Notably, we test the deepest possible intact sample from the inner prism (2,841 mbsf) collected by the NanTroSEIZE project, thus our samples cover the greatest possible range of in situ conditions in the Nankai accretionary prism. Based on the spatial pattern in frictional characteristics, we discuss the relation to the fault slip activities at the shallow Nankai Trough.

2. Sample Materials

Sediment samples were selected from three different sites: on the incoming plate seaward of the trench (IODP Site C0011), at the toe of the outer prism (IODP Site C0006), and at the inner accretionary prism (IODP Site C0002) (Figure 1). Samples were specifically selected (Table 1) to account for the variation of clay mineral content at the Nankai Trough. Mineral compositions of the samples were quantitatively analyzed by X-ray diffraction (XRD) techniques, using the method described in Vogt et al. (2002), which is a full-pattern method using a reference materials database. The “clay mineral content” in this study is the sum of the contents of smectite, illite, mica, and other types of phyllosilicates and mixed-layer clays (Table S1).

IODP Site C0011 is located on the incoming Philippine Sea plate (Figure 1) and five lithologic units are identified at this site (Expedition 322 Scientists, 2010a). Three lithostratigraphic units are often used to classify hemipelagic sediments at the Nankai Trough: Upper Shikoku Basin (USB) sediments containing air-fall pyroclastic deposits of Pleistocene to late Miocene age; Middle Shikoku Basin (MSB) sediments deposited...
as volcaniclastic turbidity currents in a submarine fan system in the late Miocene; and Lower Shikoku Basin (LSB) sediments deposited as terrigenous turbidity currents in a submarine fan system (Expedition 322 Scientists, 2010b). Three samples were selected from this site (Figure 1c): C0011B-55R-5 (green gray silty sandstone) with a clay mineral content of 10% from Unit IV (LSB sediment) at 782 mbsf (meters below sea-floor), C0011B-8R-3 (chaotic gray green volcaniclastic sandstone) with a clay mineral content of 18% from Unit IIB (MSB sediment) at 406 mbsf, and C0011B-25R-1 (green gray silty claystone) with a clay mineral content of 59% from Unit III (LSB sediment) at 540 mbsf. Smectite contents in the bulk sediment were 2% for C0011B-55R-5, 3% for C0011B-8R-3, and 10% for C0011B-25R-1.

IODP Site C0006 is located at the toe of the outer prism, and the boreholes penetrate both Pleistocene trench deposits and accreted Shikoku basin sediments (Expedition 316 Scientists, 2009, Figure 1). Three samples were selected (Figure 1c): C0006E-12H-3 (silty clay) with a clay mineral content of 25% from Unit II (accreted TF or USB sediment) at 73 mbsf, C0006E-36X-1 (greenish to greenish gray silty claystone) with a clay mineral content of 41% from Unit IIC (accreted TF sediment) at 278 mbsf, and C0006F-19R-1 (gray to greenish gray silty claystone) with a clay mineral content of 50% from Unit III (accreted USB sediment) at 563 mbsf. Smectite contents in bulk sediment were 2% for C0006E-12H-3, 3% for C0006E-36X-1, and 8% for C0006F-19R-1.

IODP Site C0002 is located at the inner accretionary prism and penetrates present forearc basin sediments, accreted prism sediments, and older sediments that originated as either accreted Shikoku basin sediments or middle Miocene trench fill sediments (Expedition 315 Scientists, 2009; Kitajima et al., 2020). Two samples were taken from this site (Figure 1c): sample C0002B-63R-1 (dark gray silty claystone to clayey sandstone) with a clay mineral content of 21% from Unit IV (accreted TF or USB sediment) at 1,034 mbsf, and sample C0002T-2K-1 (silty claystone) with a clay mineral content of 43% from Unit VB (underplated MSB or LSB sediment) at 2,841 mbsf. Smectite contents in bulk sediment were 4% for C0002B-63R-1, and 3% for C0002T-2K-1.

### 3. Methods

#### 3.1. Experimental Procedure

All experiments were conducted using a single-direct shear device at MARUM, University of Bremen, under room temperature conditions (Figure 2a) (Ikari & Kopf, 2011). Samples were trimmed from coherent core samples aligned with the core axis, hereafter called intact samples, to fit a cylindrical volume with a diameter of 25.4 and ~20 mm in height within the sample cell. The sample cell was flooded with simulated seawater with a salinity of 3.5%, and porous metal frits guaranteed sample drainage. We calculated the *in situ* effective normal stress ($\sigma'$) on the sample in the borehole as the difference between the vertical stress ($\sigma_v$) and hydrostatic pore pressure, based on shipboard bulk density measurements (Expedition 315 Scientists, 2009; Expedition 316 Scientists, 2009; Expedition 322 Scientists, 2010a; Expedition 333 Scientists, 2012; Kitajima et al., 2020; Tobin et al., 2015a, 2015b). Pore pressure cannot be applied in the device.
therefore we applied the calculated in situ effective stresses as the normal stress using an electric motor, and let the sample equilibrate to the applied stress overnight. Some samples from Sites C0006 and C0002 have experienced thrust fault or strike-slip fault stress regimes (Chang & Song, 2016; Kitajima et al., 2012, 2017; Kitajima & Saffer, 2014; Lin et al., 2016; Song & Chang, 2017). However, we tested these samples under their vertical effective stresses because the limited sample size did not allow reorientation of the sample. The applied stress conditions are listed in Table 1.

Shear force was horizontally applied to the lower cell plate by an electric motor (Figure 2a), and a shear zone was formed in the middle of the intact sample (Figure 2b). During shearing, the applied normal force is always aligned with the upper half of the sample. As the lower sample is sheared away, the upper sample becomes partially supported by, and shears against, the lower steel sample holder. Therefore, the normal stress on the sample is constant throughout the experiment and no correction is needed. Shear driving velocity was set to 10 μm/s for 5 mm as a “run-in” process to establish a residual friction value. The driving velocity was then reduced to 0.01 μm/s, and then increased stepwise by a factor of 3 up to 30 μm/s, with 700 μm of sample displacement at each step (Figure 2c). Note that the first velocity step from 10 μm/s (run-in) to 0.01 μm/s was not used in our analyses because we focus on accelerating fault slip. Two types of shear displacement were measured during experiments: the sample displacement was measured directly as the offset between the upper and lower sample cell plates, and the load point displacement was measured away from the sample at the load cell and represents the driving by the apparatus (Figure 2a). We used the sample

Figure 2. (a) Schematic view of the single-direct shear device (modified from Roesner et al., 2020). (b) Schematic views of the cylindrical intact and sheared samples. (c) Exemplary friction data for the run-in and velocity steps, with the in situ normal stress (σ’) applied. Inset shows an example of a velocity-step test. (d) Example data for a cohesion test, conducted under zero normal stress.
displacement for the representative shear displacement of our experiments, and the load point displacement was only used for fitting experimental data from velocity step tests. Shear stress $\tau$ and shear displacement were measured with sampling rates corresponding to two records per 1 $\mu$m of shear displacement.

Since clay-rich materials may show nonnegligible cohesion (Ikari & Kopf, 2011), we measured the cohesion in our samples before starting the run-in process. For this particular measurement, the intact samples were sheared without applying any normal load. We defined the cohesion $c$ as the maximum shear stress during the cohesion test (Figure 2d), and calculated the cohesion coefficient $\chi$ as outlined in Ikari and Kopf (2011):

$$\chi = \frac{c}{\sigma'},$$

which was used in this study to compare between samples which experienced different $\sigma'$, because previous studies have shown that cohesion is sensitive to maximum effective stress that the sample experienced (Ikari & Kopf, 2011). Here we assumed that the in-site effective vertical stress is the maximum overburden. However, as some samples would be overconsolidated, the $\chi$ value reported in this study might be the maximum estimation. Although we measured the cohesion of sediments directly, in order to compare with previous studies we defined an apparent sliding friction coefficient $\mu$ from the total shear stress as:

$$\mu = \frac{\tau}{\sigma'}$$

The steady-state friction coefficient $\mu_s$ was defined as the value of $\mu$ measured at the end of the run-in (Figure 2c).

For velocity step tests, the rate- and state-dependent friction (RSF) law (Dieterich, 1979; Ruina, 1983) was used to determine the velocity dependence of friction of samples. We used the one state variable version of the RSF law:

$$\mu = \mu_0 + a \ln \left(\frac{V}{V_0}\right) + b \ln \left(\frac{V_0 \theta}{d_c}\right),$$

where $\mu_0$ is the initial friction coefficient immediately before the velocity step, $a$ and $b$ are empirically determined constitutive parameters, $V_0$ and $V$ are the slip velocities before and after the change in velocity, $\theta$ is the state variable (units of time) which represents the memory of previous sliding, and $d_c$ is a critical slip distance over which $\theta$ evolves at the new sliding velocity. For the time evolution in $\theta$, the “Aging law” (Dieterich, 1979) was used in this study:

$$\frac{d\theta}{dt} = 1 - \frac{V\theta}{d_c}.$$

We used RSFit3000, a MATLAB code for fitting friction data with the RSF law (Skarbek & Savage, 2019), to obtain the friction parameters. Before fitting data with the equations above, a slip-dependent trend was removed from data using a linear function. The effect of the slip-dependent trend can also be described by additional state variables (Blanpied et al., 1998) and some studies proposed that the fabric development may be the cause of the slip-dependent trend (Haines et al., 2013; Marone, 1998). Here, we removed the slip-dependent trends in order to determine how the friction is affected by the velocity only (Ito & Ikari, 2015). Note that while the stiffness $k$ of the system was treated as a variable for fitting, $\mu_0$ was not, and no weighting function was applied. The velocity dependence of friction at steady state is represented by the parameter $a-b$; when the sign of $a-b$ is positive (known as velocity-strengthening friction), friction coefficient increases with velocity and accelerating slip is resisted, whereas slip may accelerate and nucleate a slip instability when the sign of $a-b$ is negative (velocity-weakening friction) and specific elastic conditions are met (Dieterich, 1986; Ruina, 1983).

To achieve dynamic slip leading to an earthquake, the shear stiffness $K$ of surrounding rocks should be lower than the critical stiffness $K_c$ defined as (Dieterich, 1986; Ruina, 1983):

$$K_c = \frac{-(a - b)\sigma'}{d_c}.$$
Assuming a circular crack with the radius of \( r \) for simplicity, the value of \( K \) at the center of the crack can be described as:

\[
K = \frac{\pi G}{24r},
\]

where \( G \) is the shear modulus of surrounding rock. Therefore, the minimum critical crack radius \( r_c \) for unstable slip can be described as (Dieterich, 1986):

\[
r_c = \frac{7\pi Gd_a}{24\sigma'(b-a)}.
\]

The critical nucleation length can be described as \( L_c = 2r_c \), which determines whether the rupture area slips unstably or not. When the dimension of slip area exceeds \( L_c \), unstable slip occurs.

### 3.2. Microstructural Observations

After the deformation experiments, samples were recovered from the experimental cell, dried in a low-humidity (<10% relative humidity) chamber at room temperature, and impregnated with epoxy resin to obtain polished sections for microstructural observation. An exception was the sample C0002T-2K-1, which was preserved for further testing. Samples were cut and polished parallel to the shear direction (Figure 2b) and observed with a Hitachi SU-3500 scanning electron microscope (SEM) at the Geological Survey of Japan. Backscattered electron images (BSEs) were obtained under the low evacuation mode with an accelerating voltage of 30 kV and a probe current of about 120 nA.

### 4. Results

#### 4.1. Frictional Properties

Steady-state friction coefficients \( \mu_s \) at all sites showed an inverse correlation with clay mineral content (Figure 3a). For samples from Site C0011, \( \mu_s \) for 10% clay mineral content was 0.59, for 18% clay mineral content 0.63, and for 59% clay mineral content 0.24; for samples from Site C0006, \( \mu_s \) for 25% clay mineral content was 0.83, for 41% clay mineral content 0.57, and for 50% clay mineral content 0.32; for Site C0002, samples with 21% clay mineral content showed \( \mu_s \) value of 0.55 and the sample with 43% clay mineral content showed 0.41. For lower clay mineral contents (10%–40%), \( \mu_s \) of samples from C0011 and C0002 is low compared to those from C0006. In contrast, we did not observe a relationship between clay mineral content and \( \chi \) (Figure 3c). Samples from Site C0002 showed low \( \chi \)-values of 0.02–0.03 compared to samples from Site C0011 with \( \chi \)-values of 0.04–0.05 and C0006 with those of 0.04–0.06 (Figure 3c). Samples with higher smectite content appeared to have lower \( \mu_s \) values, but do not show any correlation to \( \chi \) values (Figures 3b and 3d).

For the input samples from Site C0011, all samples exhibited velocity-weakening behavior with \( a-b \) ranging from −0.004 to 0.000 for almost all velocities (Figure 4). C0011B 25R-1 showed stick-slip at the velocities of 1 and 3 \( \mu m/s \). Stress drops were approximately 10 kPa at both velocities (Figure S1). The critical slip distance \( d_c \) ranged from 5 to 62 \( \mu m \). For all samples from Site C0006, both velocity-weakening and velocity-strengthening behaviors were observed, with a relatively wide range of \( a-b \) from −0.009 to 0.003. The velocity strengthening was mainly observed at higher velocities; 10 and 30 \( \mu m/s \) for C0006E-36X-1, and 30 \( \mu m/s \) for C0006F-19R-1. The \( d_c \) values ranged from 3 to 71 \( \mu m \). Note that the variation in \( d_c \) values for the sample C0006E-12H-3 may be large because of the relatively large fluctuations in the raw data of friction coefficient as a function of displacement, caused by low \( \sigma' \). For the samples from Site C0002, the velocity dependence of friction differed slightly between two samples. The shallower sample C0002B-63R-1 (21% clay mineral content) with \( \mu_s \) of 0.55 exhibited only velocity-weakening behavior \( (a-b = -0.003 \text{ to } 0.000) \). The deeper sample C0002T-2K-1 (43% clay mineral content) with \( \mu_s \) of 0.41 exhibited velocity-weakening \( (a-b = -0.001 \text{ to } 0.000) \) at driving velocities lower than 1 \( \mu m/s \), and velocity-strengthening \( (a-b = 0.000 \text{ to } 0.003) \) at higher driving velocities. The \( d_c \) values were 5–21 \( \mu m \) for C0002B-63R-1, and 3–30 \( \mu m \) for C0002T-2K-1.
4.2. Microstructure

When we compared samples with different clay mineral contents at Sites C0006 and C0011, samples with high clay mineral content (C0011B-25R-1 (silty claystone) with 59% clay mineral content, C0006F-19R-1 (silty claystone) with 50% clay mineral content) showed planar shear surfaces (Figures 5a, 5b, and 5g).

However, several cracks away from the shear surface were present, mostly oriented with the angle of Riedel shear or Y shear plane (Logan et al., 1979). Although some of them may be created when the samples were dried after the removal from the experimental plates, their angles with the shear direction suggest that the crack formation is related to the shear deformation. In contrast, relatively low clay mineral content samples (C0011B-55R-5 (silty sandstone) with 10% clay mineral content, C0011B-8R-3 (sandstone) with 18% clay mineral content) had wider shear zones of about 500 μm, which appeared to consist of finer grains compared to the undeformed surrounding area (Figures 5c, 5e, and 5h). Such a strain localization feature associated with shear deformation is often referred to as a “shear band” (Fossen et al., 2007; Maltman, 1998;
Sample C0006E-36X-1 (silty claystone) with 41% clay mineral content shows a narrower shear band with a thickness of \( \sim 50 \mu m \) (Figure 5d). In the case of sample C0006E-12H-3 (silty clay), the shear band was not clear (Figure 5f) although its clay mineral content was low (25%). Based on given microstructural observations, although low applied normal stress might make a narrower or unclear shear band, shear bands are associated with sediments which have low clay mineral content (<30%–40%) or the sediment contained larger grains, i.e., sandstone. Some Riedel shears outside of the shear band were observed, but the number of such cracks was less than in the high clay mineral content or clayey samples. In the case of the sediment from the inner prism (C0002B-63R-1, silty claystone to clayey sandstone, 21% clay mineral content), neither the shear band nor the planar shear surface was discernable, but rather a rough shear surface was observed (Figures 5i and 5j). Few instances of off-fault damage were observed for this sample.

5. Discussion

5.1. Possible Mechanisms for the Observed Patterns in Friction Coefficient

The observed decreasing trends in friction coefficient as a function of clay mineral content for sediments from all locations are consistent with previous studies (Brown et al., 2003; Ikari et al., 2018; Kopf & Brown, 2003; Logan & Rauenzahn, 1987; Saffer & Marone, 2003; Shimamoto & Logan, 1981; Takahashi et al., 2007, 2013, 2014; Tembe et al., 2010). However, the trends are different between the Site C0006 samples and those from Sites C0002 and C0011 (Figure 3a), possibly reflecting lithological differences in the samples. The \( \mu_c \) values of TF and USB sediments from Site C0006 showed a steeply decreasing trend over the clay mineral content, which is consistent with those of USB sediments from Site C0011 but different from...
the gently decreasing trend for MSB and LSB sediments from the Site C0011 (Ikari et al., 2013). We have not found any obvious microstructural difference between samples from Sites C0011 and C0006, which would explain a difference in $\mu_s$ for the low-clay samples (Figure 5). Sediment samples from Site C0002 showed similar $\mu_s$ values to the friction trend for MSB and LSB sediments from Site C0011, therefore the prism toe sediments from Site C0006 have exceptionally high $\mu_s$ values with respect to clay mineral content. The difference between the C0006 samples and MSB/LSB sediments from C0002 and C0011 was more significant for lower clay mineral content. The decrease in friction coefficient with increasing clay mineral content for Site C0006 are also higher than that for the Japan Trench and the Costa Rica subduction zones (Ikari et al., 2018).

High $\mu_s$ values for the prism toe sediments could be caused by multiple reasons. Friction coefficient sometimes inversely correlates to the applied normal stress under relatively low stress conditions for clay-rich

Figure 5. Microstructures for samples (a) C0011B-25R-1, (b) C0006F-19R-1, (c) C0011B-8R-3, (d) C0006E-36X-1, (e) C0011B-55R-5, and (f) C0006E-12H-3. Closer views of the planar shear surface for the sample with a high clay mineral content (g), and shear band for the sample with a low clay mineral content (h), whose locations are indicated in (a) and (c), respectively. Microstructure for sample C0002B-63R-1 (i) and its shear surface (j).
materials (Ikari et al., 2007; Saffer et al., 2001; Saffer & Marone, 2003), therefore the low effective normal stress condition for some prism toe sediments, such as 0.610 MPa for C0006E12H-3, may cause higher μs values for samples from Site C0006. Another possible cause of the difference in the friction coefficients may be the consolidation state of sediments (Guo et al., 2011; Stipp et al., 2013). Previous studies suggested that the prism toe sediments at Site C0006 are more overconsolidated (overconsolidation ratio–OCR: 2.61–3.59) compared to the input sediments at Site C0011 (OCR: 1.5–2.12), due to the lateral tectonic loading and the removal of overlying sediment (Kitajima et al., 2012; Kitajima & Saffer, 2014). Although the difference in OCR values may not be large enough to fully explain the difference in friction between two sites, overconsolidation also leads to higher χ values (Ikari & Kopf, 2015), which would explain relatively high χ values for Site C0006 (Figure 3c). Therefore, the difference in consolidation state could contribute to the difference in friction coefficient and cohesion coefficient.

Cementation of the TF or USB sediments may also cause the difference in mechanical properties. Sediments at the Nankai Trough are known to contain approximately 35% volcanic materials (Scudder et al., 2018). Dissolution-precipitation processes may cause volcanic glass to act as cement between grains and stiffen bulk sediment, preserving high porosity as observed in TF and USB facies at the Site C0006 and C0011 (Spinelli et al., 2007; White et al., 2011). Because the cementation of sediments increases the mechanical strength (Ikari & Hüpers, 2021; Schnaid et al., 2001), cementation sourced from volcanic glass may contribute to the observed difference in friction coefficient between TF/USB and MSB/LSB sediments in addition to the overconsolidation state. Cementation should increase the cohesion coefficient χ, and slightly larger χ-values were observed for prism toe sediments than input and inner prism sediments (Figure 3c). However, the difference in χ-values is ~0.02, which is still smaller than the difference in friction coefficients of ~0.2 between two regions (Figure 3a). Although the effects of low effective stress, overconsolidation state, and cementation individually are likely insufficient, the combined effects of these mechanisms and/or a yet unidentified mechanism may explain the differences in friction of our experiments using low-clay samples.

5.2. Spatial Patterns in Mechanical Strengths of Sediments: Inferences From Coulomb Wedge Modeling

We observe that the samples from the prism toe Site C0006 exhibit higher friction coefficients compared to samples from C0002 and C0011, despite similar clay mineral content. Although the obtained spatial variation in friction coefficients is generally only applicable at the locations where the samples were retrieved, we use the Coulomb wedge model, or critical taper model, to estimate the friction coefficients inside the prism and at the décollement, and their spatial patterns along the entire Kumano transect.

In the Coulomb wedge model (Dahlen, 1990; Dahlen et al., 1984; Davis et al., 1983), the wedge is assumed to be near failure which satisfies the following Coulomb criterion:

\[ \tau = \mu \sigma' \]

where \( \tau \) and \( \sigma' \) are shear and effective normal stresses, and \( \mu \) is the friction coefficient for the Coulomb criterion. Note that we assume a noncohesive wedge for simplicity (Dahlen, 1984). The angle \( \Psi_0 \), which is the angle of \( \sigma_1 \) from the surface slope for the critical state of the wedge, can be described as follows:

\[ \Psi_0 = \frac{1}{2} \arcsin \left( \frac{\sin \alpha'}{\sin \phi} \right) - \frac{1}{2} \alpha' \]

where \( \alpha' \) is the modified surface slope angle:

\[ \tan \alpha' = \left( \frac{1 - \rho_w}{1 - \rho} \right) \tan \alpha \]

where \( \rho \) and \( \rho_w \) are the density of wedge sediment (2.2 g/cm\(^3\)) and of seawater (1.0 g/cm\(^3\)), \( \alpha \) is the surface slope angle, and \( \lambda \) is the overpressure ratio within the prism (Davis et al., 1983; Saffer & Tobin, 2011) defined with seafloor-referenced pressure (\( P \)) as:
\begin{align}
\lambda &= \frac{P_{\text{fluid}}}{P_{\text{lithostatic}}},
\end{align}
\begin{align}
\lambda &= \frac{P_{\text{fluid}}}{P_{\text{lithostatic}}},
\end{align}

and \( \phi \) is the friction angle of prism sediment:
\begin{align}
\mu_w &= \tan \phi.
\end{align}
\begin{align}
\mu_w &= \tan \phi.
\end{align}

The angle \( \Psi_b \), which is the angle of \( \sigma_t \) to the décollement dip \( \beta \), is:
\begin{align}
\Psi_b &= \frac{1}{2} \arcsin \left( \frac{\sin \phi_b}{\sin \phi} \right) - \frac{1}{2} \phi_b = \alpha + \beta + \Psi_0,
\end{align}
\begin{align}
\Psi_b &= \frac{1}{2} \arcsin \left( \frac{\sin \phi_b}{\sin \phi} \right) - \frac{1}{2} \phi_b = \alpha + \beta + \Psi_0,
\end{align}

where \( \phi_b \) is the friction angle of the décollement sediment:
\begin{align}
\mu_b &= \frac{1 - \lambda}{1 - \lambda_b} \tan \phi_b,
\end{align}
\begin{align}
\mu_b &= \frac{1 - \lambda}{1 - \lambda_b} \tan \phi_b,
\end{align}

where \( \lambda_b \) is the overpressure ratio along the décollement.

Using these equations, we can obtain a possible relation between \( \mu_w \) and \( \mu_b \) for a given geometry (\( \alpha \) and \( \beta \)) and pore pressure conditions (\( \lambda \) for within the prism and \( \lambda_b \) for along the décollement). The geometry of the wedge at the Nankai accretionary prism has been determined by previous studies (G. F. Moore et al., 2009; Park et al., 2002), with \( \alpha \) and \( \beta \) values of 10.0° and 7.6° for the prism toe, 2.9° and 0.0° for the outer prism, and 0.1° and 10.1° for the inner prism. For the prism regions, previous studies reported slightly overpressured but near hydrostatic pore pressure conditions, therefore we fixed \( \lambda \) to 0.5 at all regions (Kitajima et al., 2017; Kitajima & Saffer, 2012; Screaton et al., 2009; Tsuji et al., 2014). For the décollement, the pore pressure condition varies spatially and may be highly overpressured, so that we tested three cases of \( \lambda_b \): 0.5, 0.7, and 0.9 (Akuhara et al., 2020; Kitajima & Saffer, 2012; Skarbek & Saffer, 2009; Tsuji et al., 2014). Using our friction measurements, we show possible relations between \( \mu_w \) and \( \mu_b \) for the prism toe, for the outer prism, and for the inner prism with \( \lambda = 0.5 \) and \( \lambda_b = 0.5–0.9 \) in Figure 6. Parameters used for the Coulomb wedge modeling are included in Table 2.

The experimental results are implemented to constrain the relation between \( \mu_w \) and \( \mu_b \). Because the TF and USB sediments are accreted and compose the prism, the experimentally obtained \( \mu_w \) values for TF and USB sediments from Site C0006 (C0006F-19R-1, C0006E-36X-1, and C0006E-12H-3) can be regarded as potential \( \mu_w \) values at the prism toe and the outer prism, and the \( \mu_s \) values for C0002B-63R-1 and C0002T-2K-1 can be used as potential \( \mu_s \) values for the inner prism (Figures 6b and 6c). Similarly, because the MSB and LSB sediments are underthrusted and host the décollement beneath the prism (Ikari et al., 2013; Screaton et al., 2009), the \( \mu_w \) values for MSB and LSB sediments from the inputs (C0011B-25R-1, C0011B-8R-3, and C0011B-55R-5) are candidate \( \mu_b \) values beneath the outer prism.

At the prism toe, because both the prism and décollement sediments are derived from TF or USB material (Ikari et al., 2013; Screaton et al., 2009), the \( \mu_w \) value at the prism toe should be similar to the \( \mu_b \) value. This condition can be achieved when the \( \lambda \) value is the same as the \( \lambda_b \) value for the given wedge geometry at the prism toe (Figure 6d). Because previous studies suggested that the prism toe area is unconsolidated and fluid can easily escape, the prism toe region is not considered to be overpressured (Akuhara et al., 2020; Screaton et al., 2009); therefore, \( \lambda = \lambda_b = 0.5 \) is the likely pore pressure condition at the prism toe. Of our three candidate samples from C0006, C0006E-36X-1 with a \( \mu_w \) value of 0.57 would be the most representative for the prism toe because its clay mineral content of 40.9% may be reasonable to represent the Nankai sediment near the trench, which possibly contains detrital-origin sediment in addition to hemipelagic sediment. In addition, the \( \mu_w \) value of 0.57 is consistent with the averaged friction coefficient of sediments from shallow parts of Site C0011 (\( \mu = 0.58 \)) which was also employed as the wedge strength at the prism toe in the previous study (Ikari et al., 2013). Assuming this value to be \( \mu_w \) indicates that the modeled \( \mu_b \) value would be 0.55 (Figure 6d).

Although we were not able to test a sample from the décollement itself, we measured a \( \mu_s \) value of 0.24 for the high-clay sample of the MSB or LSB input material from Site C0011 (C0011B-25R-1). This is consistent
Figure 6.
with \( \mu_b \) of about 0.20–0.28 obtained from previous experimental studies on décollement samples at the Muroto transect or LSB sediment at the Kumano transect (Ikari et al., 2013; Ikari & Saffer, 2011), suggesting that it may represent future décollement material beneath the outer prism. Applying the \( \mu_b \) value of 0.24 for the outer prism condition, the Coulomb wedge model indicates that the \( \mu_b \) is 0.54 for \( \lambda = 0.5 \) and \( \lambda_b = 0.7 \) (Figure 6c). This value is similar to the \( \mu_b \) value at the prism toe of 0.57, which supports the idea that both the sediments at the prism toe and the outer prism have the same origin (TF or USB sediments). The choice of \( \lambda_b = 0.7 \) at the trenchward part of the décollement beneath the outer prism may be reasonable because \( \lambda_b \) would gradually increase from the prism toe to the landward portion of the décollement beneath the outer prism where \( \lambda_b \) is close to 1 (Kitajima & Saffer, 2012; Tsuji et al., 2014). When \( \lambda_b \) is 0.9, the wedge geometry indicates that \( \mu_b \) has to be larger than 0.27 (Figure 6c). Therefore, the Coulomb wedge model implies that the \( \mu_b \) would increase at the landward part of the décollement beneath the outer prism, probably reflecting the influence of higher temperature, lithification, and higher effective stress at the décollement depth; all of which tend to increase the friction coefficient (den Hartog & Spiers, 2013; Morrow et al., 2017; Trütner et al., 2015). If we consider a decrease in \( \mu_b \) as suggested by the difference in trend of friction coefficients over clay mineral content between Sites C0002 and C0006 (Figure 3a), \( \mu_b \) would range from 0.3 to 0.4 and \( \mu_b \) would range from 0.5 to 0.6.

Beneath the inner prism, the décollement around Site C0002 is expected to be overpressured having \( \lambda_b \) of close to 1 (Kitajima & Saffer, 2012; Tsuji et al., 2014), but the décollement at the more landward part beneath the inner prism would be less overpressured (Saffer & Wallace, 2015). Because the temperature increases and lithification is more advanced at the inner prism compared to the trenchward area, both \( \mu_w \) and \( \mu_b \) would increase compared to the outer prism. Based on our measurements from Site C0002 samples, the Coulomb wedge model cannot be satisfied for \( \lambda_b = 0.9 \). If we consider \( \mu_w = 0.4–0.6 \) and \( \mu_b = 0.6–0.8 \), which are slightly higher values than for the outer prism accounting for the possible influence of increasing temperature, lithification, and effective stress at deeper depths (den Hartog & Spiers, 2013; Morrow et al., 2017; Trütner et al., 2015), the possible combination of \( \mu_w \) and \( \mu_b \) indicates \( \lambda_b \) between 0.7 and 0.9, satisfying the pore pressure condition along the décollement beneath the inner prism.

Based on the combination of the Coulomb wedge model, experimental results, and previously reported pore pressure conditions, we interpret spatially variable values for the friction coefficients in accordance with changes in wedge taper angle and pore pressure along the décollement: \( \mu_b = 0.55, \mu_w = 0.57, \) with \( \lambda_b = 0.5 \) at the prism toe; \( \mu_b = 0.24, \mu_w = 0.54, \) with \( \lambda_b = 0.7 \) at the trenchward of the outer prism; \( \mu_b = 0.5–0.6, \mu_w = 0.3–0.4, \) with \( \lambda_b = 0.9 \) at the landward of the outer prism; and \( \mu_b = 0.6–0.8, \mu_w = 0.4–0.6, \) with \( \lambda_b \sim 0.8 \) at the inner prism (Figures 6e–6h). The \( \lambda \) value was assumed to be 0.5 for all regions. Note that these results should be taken with caution, because the material at the décollement zone beneath the inner prism was not studied, since it was not recovered during the NanTroSEIZE project. In addition, thermal effects should be tested in future studies because the décollement zone beneath the inner prism would be over 100°C, which was not a tested condition in this study.

### 5.3. Relationship Between Spatial Distribution of Fault Slip Activities and Frictional Behavior at the Shallow Nankai Trough

At the trench, weak MSB and LSB sediments are underthrust along the plate boundary (Ikari et al., 2013; Screaton et al., 2009). Since those underthrust sediments likely compose the décollement and host plate-boundary fault slip there, we discuss the correlation between the obtained frictional behavior and fault slip activities. Beneath the outer prism, the décollement sediment would be similar to the MSB/LSB sediments at the Shallow Nankai Trough (Figure 14). Since the décollement beneath the outer prism would increase compared to the outer prism. Based on our measurements from Site C0002 samples, the Coulomb wedge model cannot be satisfied for

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**Figure 6.** (a) Geometry of the Nankai accretionary prism (modified from G. F. Moore et al., 2009), showing the surface slope angle (\( \alpha \)) and décollement dip (\( \beta \)). Solid lines in (b)-(g) represent the possible relations between wedge strength \( \mu_w \) (y-axis) and basal friction \( \mu_b \) (x-axis) calculated with the Coulomb wedge model for tested values of \( \lambda, \lambda_b, \alpha \), and \( \beta \) at the inner prism (b), (e), the outer prism (c), (f), and the prism toe (d), (g). We assumed \( \lambda = 0.5 \) in all regions. Left endpoint of each line is the critical point to keep the wedge geometry. In (b)-(d), we show the relation between the Coulomb wedge modeling and experimental results. Vertical dashed lines and filled squares for samples from C0011 represent the experimentally obtained \( \mu_b \) values, and horizontal dashed lines and open squares for samples from C0006 and C0002 are those for \( \mu_w \) for each sample. In (e)-(g), the preferred combination of \( \mu_w \) and \( \mu_b \) values for each location is represented by a colored shaded area. (h) Estimated, preferred spatial patterns in \( \mu_w \) and \( \mu_b \) values with possible \( \lambda \) and \( \lambda_b \) conditions. In every figure, colors of solid and dashed lines, open and filled markers, and shaded areas represent the locations: gray for the prism toe, orange for the outer prism, and red for the inner prism.
the sample C0002T-2K-1 has similar frictional properties of the décollement. It is because this sample is the underplated LSB intact sediment collected from the deepest possible depth of the inner prism (2,841 mbsf). Note that the location of the Philippine Sea plate prior to 10 Ma is still under debate (Kimura et al., 2014; Mahony et al., 2011; Underwood, 2018), therefore the inner prism sediment may have a different origin than the current input sediments if they are older than ∼10 Ma. However, the age of our C0002T-2K-1 sample is ∼9 Ma (Kitajima et al., 2020); therefore, based on the lithological interpretation of the collected samples we

| Symbol | Description | Value |
|--------|-------------|-------|
| $a$ | RSF parameter (direct effect) | Obtained by fitting Equations 3 and 4 to experimental data |
| $b$ | RSF parameter (evolutionary effect) | Obtained by fitting Equations 3 and 4 to experimental data |
| $c$ | Cohesion | Experimentally obtained |
| $d_c$ | RSF parameter (critical slip distance) | Obtained by fitting Equations 3 and 4 to experimental data |
| $G$ | Shear modulus of surrounding rock | 4.5 GPa (décollement beneath the outer prism) |
| $L_c$ | Critical nucleation length | 2r, (Equation 7) |
| $P_{\text{fluid}}$ | Seafloor-reference fluid pressure | Modeled via $\lambda$ and $\lambda_0$ |
| $P_{\text{lithostatic}}$ | Seafloor-referenced lithostatic pressure | Modeled via $\lambda$ and $\lambda_0$ |
| $r_c$ | Minimum critical circular crack radius for unstable slip | Modeled by $G$, $\sigma'$, $a$, $b$, and $d_c$ (Equation 7) |
| $\alpha$ | Seafloor surface slope angle | 10.0° (Prism toe) |
| $\beta$ | Décollement dip | 7.6° (Prism toe) |
| $\theta$ | RSF parameter (state variable) | Obtained by fitting Equations 3 and 4 to experimental data |
| $\lambda$ | Overpressure ratio within the wedge | 0.5 |
| $\lambda_0$ | Overpressure ratio at the décollement | From 0.5 to 0.9 |
| $\mu$ | Friction coefficient | $\tau/\sigma'$ (Equation 2) |
| $\mu_s$ | Steady-state friction coefficient | Value of $\mu$ at the end of run-in |
| $\mu_w$ | Friction coefficient of the sediment in the wedge | Experimental values $\mu_w$ for TF and USB sediments (Table 1), or estimated by $\mu_w$, $\alpha$, $\beta$, $\lambda$, and $\lambda_0$ |
| $\mu_b$ | Friction coefficient of the décollement sediment | Experimental values $\mu_b$ for MSB and LSB sediments (Table 1), or estimated by $\mu_b$, $\alpha$, $\beta$, $\lambda$, and $\lambda_0$ |
| $\rho$ | Density of wedge sediment | 2.2 g/cm$^3$ |
| $\rho_w$ | Density of seawater | 1.0 g/cm$^3$ |
| $\sigma'$ | Effective normal stress | Calculated based on shipboard bulk density measurements and hydrostatic pore pressure |
| $\tau$ | Shear stress | Experimentally obtained |
| $\chi$ | Cohesion coefficient | $c/\sigma'$ (Equation 1) |
| $\psi_b$ | Angle of the maximum principal stress axis from the seafloor surface slope | Modeled by $\alpha$, $\beta$, $\rho_w$, $\rho_w$, $\mu_w$, and $\lambda$ (Equations 9–11) |
| $\psi_b$ | Angle of the maximum principal stress axis from the décollement dip | Modeled by $\mu_w$, $\mu_b$, $\lambda$, and $\lambda_0$ (Equations 12 and 13), or $\alpha + \beta + \psi_b$ |

Table 2: List of Parameters Used in This Study
may assume that the inner prism sediment to at least to ∼3 km depth can be correlated to the LSB sediment at the input Site C0011 (Tobin et al., 2015b).

Hypocenters of VLFE locate beneath the landward part of the outer prism area and have not been observed beneath the inner prism (Nakano et al., 2018; Shiraishi et al., 2020; Sugioka et al., 2012; Takemura et al., 2019). On the other hand, SSEs have been observed to locate beneath the inner prism, and sometimes propagate seaward (Araki et al., 2017; Ariyoshi et al., 2021).

As for the velocity dependence of friction of underthrust/underplated MSB or USB sediments, velocity-weakening behavior was observed over the entire tested velocity range for samples from Site C0011. The sediment at the deep part of the inner prism (C0002T-2K-1) also showed mostly velocity weakening, but some velocity strengthening occurred at velocities >3 μm/s. Hence, differences in velocity dependence may influence the spatial variation in fault slip activities beneath the outer prism and inner prism. Although previous studies have often reported velocity-strengthening behavior for clay-rich gouges from the Nankai Trough (Ikari & Saffer, 2011), the velocity-weakening behavior in this study is consistent with the results of Roesner et al. (2020) who reported that intact samples tested under in situ effective normal stresses showed the velocity weakening, but their equivalent powdered samples under larger effective normal stresses showed velocity strengthening. Because we used intact samples, the observed velocity-weakening behavior could be related to the lithification state of the samples.

To test a possible correlation between velocity dependence and spatial distribution of fault slip activities, we calculated the critical nucleation length $L_c$ for slip instability for the underthrust sediment using the experimentally obtained $a$-b and $d_c$ values. It can be expected that the area with VLFEs has a smaller $L_c$ compared to the area with only SSEs, because VLFEs are faster than SSEs and release some energy as seismic waves (Ide et al., 2007), and therefore may be more likely to host unstable slip events compared to SSEs.

We considered two regions: (1) sediment at the décollement beneath the outer prism (3 kmbsf), and (2) sediment at the décollement beneath the inner prism (5 kmbsf). For simplicity, we assume bulk density of the overlying sediment is 2.0 g/cm$^3$; therefore, the total overburden stresses at the décollement are 60 MPa beneath the outer prism and 100 MPa beneath the inner prism. Shear moduli $G$ for the outer and inner prism cases are 4.5 and 12.5 GPa, based on $S$-wave velocities of 1.5 and 2.5 km/s (Akuhara et al., 2020) and sediment bulk density of 2.0 g/cm$^3$. We assume $a$-$b$ values are $-0.004$ for the outer prism case and $-0.001$ for the inner prism case, and $d_c$ value of 5 μm for both cases based on experimental result on samples from Sites C0011 and C0002 (C0002T-2K-1). Note that we used minimum values for $d_c$ and $a$-$b$ to test the most unstable possible case. The $L_c$ values for both cases calculated with the above conditions are displayed in Figure 7, which show inverse relations with the effective normal stress as described in Equation 7.

Based on previous studies and our Coulomb wedge modeling (Section 5.2), the $L_c$ value beneath the landward part of the outer prism where VLFEs occur is estimated to be 0.9, and 0.8 beneath the inner prism. The above conditions yield $L_c$ values of 1.7 m for the décollement beneath the outer prism and 5.7 m for the décollement beneath the inner prism, showing that the outer prism décollement is more prone to slip instability than the inner prism décollement. The smaller $L_c$ value at the outer prism is mostly due to the lower $G$ and larger velocity weakening. The low $G$ value is inferred from the low $S$-wave velocity, which is the result of high pore pressure along the shallow décollement.

This smaller $L_c$ value beneath the outer prism suggests that slip beneath the outer prism area is expected to be less stable than that beneath the inner prism area. Moreover, we observed the transition from negative to positive $a$-$b$ at $\sim$1 μm/s for C0002T-2K-1 from the deep inner prism (Figure 4c). This transition is used as a “cut-off” velocity in numerical models of slow earthquakes, where the velocity strengthening at higher slip velocities could resist the slip acceleration and result in SSEs (Shibazaki & Iio, 2003). Our results on input sediments do not show such a transition (Figure 4a), suggesting that such a mechanism for limiting slip velocity is absent for the décollement beneath the outer prism, which is therefore more unstable and may generate faster events. We speculate that the difference in $a$-$b$ values between the two locations might possibly originate from grain interactions during shear deformation (den Hartog & Spiers, 2013; Roesner et al., 2020) although we cannot verify this hypothesis without microstructural information for the sample C0002T-2K-1. Because we conducted our tests at room temperature, depth-dependent temperature effects could induce some uncertainty in our calculations as well. However, earlier work suggests that the
smectite-illite conversion process would not affect $a-b$ (Saffer et al., 2012). Higher temperatures at deeper depths may lower the $a-b$ value for illite-rich gouges (den Hartog & Spiers, 2013), meaning that the $L_c$ value for the inner prism might also be lower. The $a-b$ and $d_c$ values for the sediments in the Nankai Trough may depend on the effective normal stress and pore fluid pressure (Bedford et al., 2021); therefore, the estimated $L_c$ value may include some uncertainty because our experiments were conducted under hydrostatic conditions. Because our findings of a trench-perpendicular pattern in velocity dependence and strength of the décollement match the spatial distribution of SSE occurrence and VLFE locations, the spatial variations in frictional properties of the sediment as well as pore pressure state may be responsible for the slip behavior at the shallow Nankai Trough subduction zone.

6. Conclusions

Frictional characteristics of sediment samples from various locations along the Kumano transect at the Nankai Trough were experimentally measured to examine the spatial patterns in frictional behavior and their relation to fault slip activities. The samples were selected to cover a large spatial range along the NanT-roSEIZE transect, from the input site, the prism toe, and the inner prism, including the deepest intact sample ever collected thus far in the Nankai Trough. Friction coefficients were found to inversely correlate with the clay mineral content of sediments, but the prism toe sediments with low clay mineral content showed higher friction coefficients than the sediments with similar clay mineral content collected at the other locations. Velocity dependence of friction also varied among locations. The samples from the input site and the shallow inner prism showed velocity-weakening behavior at all the tested driving velocities, whereas the samples from the outer prism and the deep inner prism showed velocity weakening at low driving velocities and velocity strengthening at high driving velocities. For the input and the outer prism sediments, the deformation microstructure depends on the clay mineral content but no clear differences were observed between the two locations. A planar shear surface with significant off-fault damage is observed for high-clay samples, whereas a shear band and few instances of off-fault damage are characteristic for low-clay samples. Coulomb wedge modeling based on our experimental results, known wedge geometries, and estimated pore

![Figure 7. Critical nucleation length $L_c$ for the décollement sediment beneath the outer and inner prisms, as a function of the effective normal stress $\sigma'$ and the overpressure ratio $\lambda_b$ at the décollement.](image-url)
pressure conditions reveal that the décollement friction ($\mu_b$) and the wedge strength ($\mu_w$) at the prism toe are high, $\mu_b$ beneath the trenchward part of the outer prism is the lowest and increases landward and $\mu_w$ gradually decreases landward, and both $\mu_b$ and $\mu_w$ values would increase at the inner prism. These spatial patterns in friction may be results of diagenetic processes during the tectonic history and the sediments' lithological properties, but other possibilities exist. Critical nucleation lengths for slip instability were calculated for the candidate décollement sediments and showed that the décollement sediment beneath the outer prism has a smaller critical nucleation length than that beneath the inner prism, because of its larger negative $\alpha$-value and high pore pressure condition. This difference in the critical nucleation lengths suggests that the décollement beneath the outer prism may be less frictionally stable than beneath the inner prism. Combined with an observed transition from velocity-weakening to -strengthening behavior with increasing sliding velocity for the deep inner prism sediment, frictional properties for décollement zones beneath the outer and inner prisms can explain the observed spatial distribution of VLFE and SSEs at the shallow Nankai Trough.

Data Availability Statement

Additional data are available from the PANGAEA data publisher for earth and environmental science: https://doi.org/10.1594/PANGAEA.934272.

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