

GEOLOGICAL STRUCTURES CONTROLLING SLIP BEHAVIORS IN THE CONVERGENT PLATE MARGIN

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<https://hdl.handle.net/2324/7182448>

出版情報 : Kyushu University, 2023, 博士 (工学) , 課程博士
バージョン :
権利関係 :



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FAHRUDIN

2024

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THE CONVERGENT PLATE MARGIN**

By

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A dissertation submitted in partial fulfillment of requirements

for the degree of

Doctoral Engineering



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January 2024

Abstract

The shear zone (or shallow plate interface) may contain high pore fluid pressure, which can cause tectonic tremors. Minor pore pressure increase at the shear zone can cause fault rupture along the shallow plate interface if the fault zones are critically stressed. Thus, continuous monitoring of the shallow plate interface with high pore pressure is important in plate convergent margins such as the Nankai Trough and Japan Trench. In this study, I evaluated the mechanism of tectonic tremor and the factors that influence its occurrence and fluctuations in the Nankai Trough. First, seismic reflection data was used to map in detail a shear zone associated with the plate boundary fault in the Nankai accretionary prism. The shear zone can be characterized based on chaotic or weak reflection and low seismic velocity anomaly hosted high pore fluid pressure. High pore fluid pressure can be accumulated in the shear zone due to fluid expulsion from the inner wedge – transition zones to the outer wedge in Nankai Trough, southwest Japan.

In Chapter 1, I described an introductory section about slow slip events at the shallow plate interface. This chapter reviews previous studies regarding geological structures controlling slip behaviors in the convergent plate margin. Furthermore, I described that the shear zone thickness and slow slip distribution are not straightforwardly related.

In Chapter 2, I described the geological setting and previous research studies in this region. An episodic style and the migrating activity of slow earthquakes can promote tectonic tremors in the plate subduction zone. Particularly in the Nankai Trough, slow slip events are caused by the subduction between the Philippine Sea Plate and the Japanese Islands plate. The previous study interpreted that these slow slips are caused by the reactivation of strike-slip faults accommodating shear strain in the plate interface. Some studies have proposed that the

low-velocity zone (LVZ) or the shear zone near the plate boundary may be the location of slow earthquakes including tectonic tremors. Slow earthquakes are associated with thickness fluctuations of the shear zone hosted by high pore pressure around the plate interface. Such heterogenous geological structure around this plate interface is also influenced by the collision of accretionary prism with the igneous dome operated as a static backstop around the Shionomisaki.

In Chapter 3, I described seafloor bathymetric and seismic reflection datasets. A high-resolution bathymetric map of the study area is used to map fault distribution based on the basis of its elevation and topography with a spatial resolution of ~ 100 m. The shear zone is accurately defined on the 3D seismic volume. To estimate the shear zone distribution in the extensive area, the shear zones were mapped using several 2D seismic reflection profiles by comparing with the 3D seismic volume.

In Chapter 4, I investigated how shear zone thickness and faulting can influence the tectonic tremor distribution change in spatiotemporal. I proposed three-steps investigation. Step one, I calculated seismic attributes from the 2D and 3D seismic reflection data. Then, step two is mapping fault distribution from seafloor bathymetric and seismic reflection data. After mapping fault distribution, I identified the shear zone based on the seismic facies and physical characteristics. From the last step, a detailed comparison of the mapped shear zone and the distribution of tectonic tremors was obtained. The results demonstrate the tectonic tremors seem to have relationship with the shear zone.

In Chapter 5, I showed the results and geological interpretation of this study, which contain interpretation of structures around the shear zone and estimation of the shear zone thickness. I interpret the strike-slip faults around the shear zone based on features and pattern structures from bathymetric, seismic profiles, and coherence attributes. The fault geometry

indicates the positive flower structures that tended to occur in the thick part of the shear zone. In addition, high pore fluid pressure in the thick shear zone may weaken faults and thereby trigger slow earthquakes.

In Chapter 6, I discussed about slow earthquakes or tremor distribution controlled by shear zone thickness, relationship between the tremor number and the shear zone, and tectonic structures controlling shear zone distribution. Tectonic tremors line up on the oblique strike-slip faults, which changed and varied the shear zone thickness as well as fluctuation of stress and pore pressure. The thickness of the shear zone and tremor distributions are controlled by the oblique strike-slip faults, which are theorized to have originated as a result of the collision of an accretionary prism with a landward tectonic backstop around the Cape Shionomisaki.

In Chapter 7, I described the key findings and summaries of this study. The thickness of shear zone and the strike-slip fault influence the spatiotemporal tectonic tremors in Nankai Trough. This study contributes to evaluating the fluctuations of slow earthquakes. The number of tremors are influenced by variations in shear zone thickness and faults. Strike-slip faults that influences upon the variation of shear zone thickness may be elucidated by the collision of accretionary prism with a tectonic backstop. This study will certainly help to evaluate earthquake monitoring, and long-term and short-term crustal deformation monitoring associated with slow earthquakes. In addition, this study helps to understand the geological structures controlling slip behaviors in the convergent plate margin.

List of paper

This thesis is based on the following paper.

Fahrudin, Chhun, C., Tsuji, T., (2022). Influence of shear zone thickness and strike-slip faulting on tectonic tremor in the Nankai Trough, southwest Japan. *Tectonophysics* 838 (2022) 229519. <https://doi.org/10.1016/j.tecto.2022.229519>.

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1. Introduction

1.1. Research background

Slow and fast earthquakes generally occur in the plate interface due to pair of stress accumulation in critical state with different spatiotemporal them (Obara, 2020). Geological structures control slip behaviors of slow and fast earthquakes in the convergent plate margin. In the outer wedge of the Nankai Trough (Fig. 1.1a), the source of slow earthquakes has been attributed to gently dipping thrust faults at a depth of 6–9 km (Sugioka et al., 2012; Nakano et al., 2018), normal faults within the accretionary prism (To et al., 2015), or strike-slip faults at greater depths (Nakano et al., 2018). Their locations are bounded by mega-splay faults (Toh et al., 2018) and imbricate thrust faults (Shiraishi et al., 2020) landward of the basement high of the oceanic crust (Tsuji et al., 2009). However, the normal faults of the Cretaceous basement located in the forearc basin beneath hypocenter of the Tohoku-oki earthquake that may be related with rupture expansion to south and north of hypocenter due to extensional continental crust (Fig. 1.1b) (Tsuji et al., 2013, 2011; Wang et al., 2016) in Japan Trench. In other subduction zones, the earliest stages of seamount collision controlled the fault distribution associated with slow slip events in the Hikurangi margin, New Zealand (Todd et al., 2018; Davidson et al., 2020).

In Japan Trench, slow slip events may progressively increase shear strain, triggering of large interplate earthquakes such as the 2011 Tohoku-Oki earthquake (Ito et al., 2013). Slow slips together with tectonic tremors are hosted in the shallow plate interface (depth < 15 km), where the forearc boundary has districthed them from fast seismic slip (i.e., $M_w \sim 7$ earthquakes) (Abe, 1977; Bassett et al., 2016; Kubo and Nishikawa, 2020; Yamanaka and Kikuchi, 2004).

Therefore, during the Tohoku-oki earthquake, the co-seismic ruptures extended to the trench with slip fault around ~60 m in plate interface (Figure 1.1b; Ide et al., 2011; Lee et al., 2011; Ozawa et al., 2011; Linuma et al., 2012; Yue & Lay 2013; Tsuji et al., 2013; Wang et al., 2016).

In Nankai Trough, slow earthquakes occur on the shallow plate boundary near the trench (i.e., shallower than the coseismic plate boundary; Barnes et al., 2020), which is thought to be a conditionally stable (Scholz and Campos, 2012) or aseismic zone (Wang and Hu, 2006). Slip behaviors at a plate interface (e.g., Tsuji et al., 2007; Park et al., 2010; Bangs et al., 2009) showed that the low-velocity zone (LVZ) seismic facies is characterized as weakly reflective, an antiformal stacking of underplated materials, and a low density as underthrust materials. This interface close to the outer ridge (i.e., transition from outer to inner wedge) has apparently hosted slow earthquakes (Tsuji et al., 2014a; Tonegawa et al., 2017; Hendriyana and Tsuji, 2021; Fahrudin et al. 2022). Tsuji et al., (2014b) further defined LVZ as a shear zone (Fig. 1.1a) based on high pore pressure as well as discontinuous reflectors located between the basal decollements and underthrust of the outer accretionary wedge. Because this shear zone unit has elevated pore pressure (Kitajima and Saffer, 2012; Tsuji et al., 2014b), it appears to be connected to the frequency of tremor occurrences (Tonegawa et al., 2017). Furthermore, slip behaviors around the plate interface is related to thickness variations of the shear zone hosted high pore pressure. Thus, the shear zones are correlated with the occurrence of VLFE and tremor, which is consistent with stress buildup in locked areas (Obana and Kodaira, 2009; Takemura et al., 2019; Hendriyana and Tsuji, 2021) and stress changes following frequent earthquakes (To et al., 2015; Annoura et al., 2017).

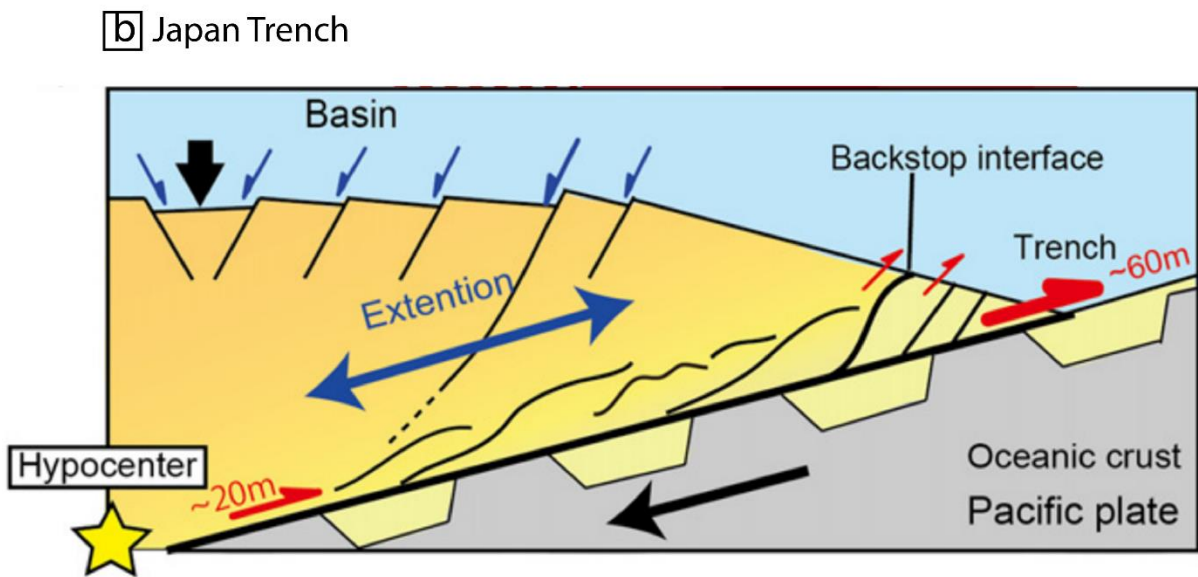
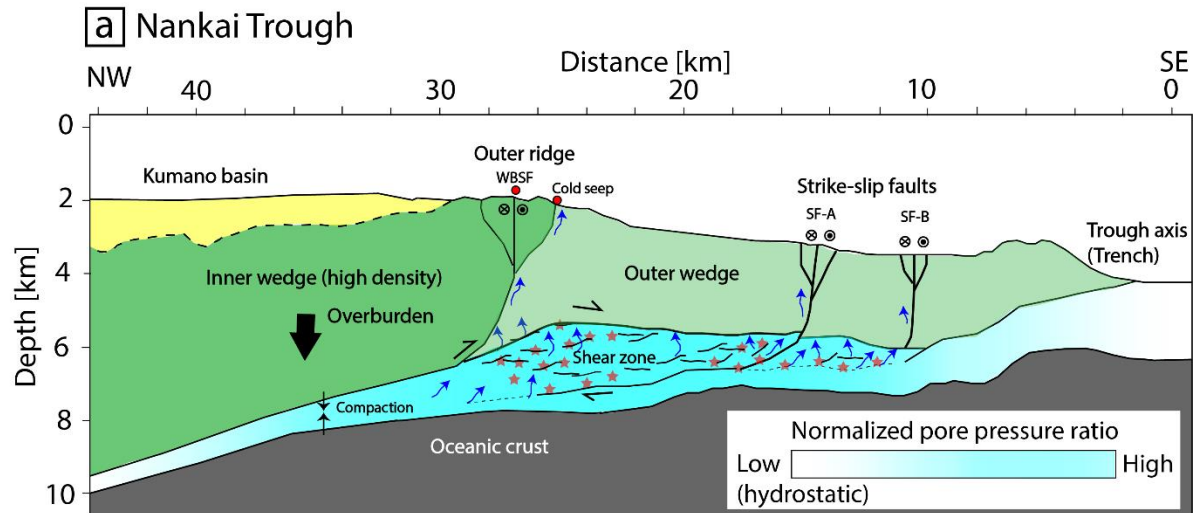


Figure 1.1. Schematic image of geological structures controls slip behaviors in the convergent plate margin in (a) Nankai Trough, southwest Japan (Fahrudin et al., 2022) and (b) Japan Trench, northeast Japan (Tsuji et al., 2013).

The slip behaviors are related to the inherited geological structures and the historical tectonic event. Both the convergent plate margin had the Cretaceous age of intrusion or sedimentary rock that subdue and collide today as common geological features in the subduction zone of the world (Park et al., 2003; Isozaki et al., 2010). Furthermore, the geological provinces contrast provides an interpretation of the orogenic mechanism at the Quaternary deformation (~2 Ma) in the Japan Trench and Nankai Trough. In Nankai Trough,

the presence of the Kumano Basin near the east coast of the Kii Peninsula (Fig. 1.1a) suggested that the accretionary prism formerly served as a tectonic backstop (Tsuji et al., 2015). Since ~6 Ma, the collision occurred between the tectonic backstop and the younger accretionary prism, then at ~2 Ma subducted beneath the inner wedge southeast of the Kii Peninsula creating a set of strike-slip faults (Tsuji et al., 2013).

1.2. Research questions and motivation

Slow earthquakes occur in the depth of the plate interface. It can be measured using seismometers that are installed offshore. In Japan, the Dense Ocean floor Network system for Earthquakes and Tsunamis (DONET), comprising approximately 50 observation points, was constructed by the Japan Agency for Marine-Earth Science and Technology (JAMSTEC) offshore to the southeast of the Kii Peninsula, to evaluate the potential future great earthquakes in Nankai Trough (Obara et al., 2020; Hendriyana and Tsuji, 2021). Installing seismometers made it possible to monitor crustal and shallow plate interface activities in the Japanese Islands using slow earthquakes (e.g., tectonic tremors). Various studies have been associated with the tectonic tremors and shear zone or low-velocity zone (LVZ) with high pore-pressure in southeast, Japan using DONET data. Several examples are slow earthquakes related to fast earthquakes (e.g., Annoura et al., 2017; Araki et al., 2017; Nakano et al., 2018; Toh et al., 2018; Takemura et al., 2019; Hendriyana and Tsuji, 2021). In Nankai Trough, Japan, the shear zone thickness may relate to area strain induced by strike-slip partition due to the collision of the tectonic backstop (Fahrudin et al., 2022), as well as strain accumulation from plate subduction (Ikeda & Tsuji, 2018).

Regarding the impact of fluid migration to on the shear zone (shallow plate interface), spatial factors such as fracture network (geological structures) play an important role in slip behaviors. According to the researches (Tonegawa et al., 2017; Hendriyana and Tsuji, 2021),

tremor frequently happens near a plate boundary fault. I observe a shear zone with high pore pressure conditions in the Nankai accretionary prism off Kumano (Tsuji et al., 2014b), and the tremors appear to occur near the shear zone (Hendriyana and Tsuji, 2021). Nevertheless, the shear zone and tremor distribution may not necessarily have a clear link. Thus, based mainly on 2D and 3D seismic reflection data, this study examines the shear zone thickness and geological features that regulate the spread of tectonic tremors in southeast Japan. By applying my approach in the Nankai Trough, I hope to address the following questions:

1. Which geological structure features allow a shear zone to influence the tectonic tremors change?
2. How does the shear zone thickness spread to influence the tectonic tremors?

The shear zone thickness and geological structures contribute to the impact of tectonic tremor distribution on the shallow plate interface. This study can further evaluate the areas where the tectonic tremor occurred in spatiotemporal of tremors. Furthermore, the benefit of the study is significant not only for earthquake monitoring but also for exploration in the methane gas, and long-term and short-term crustal deformation monitoring associated with slow earthquakes.

1.3. Sections description

Figure 1.2 shows the outline of this dissertation. Chapter “1. Introduction” describes the research background and the motivation for this work. In this dissertation, I considered the geological structures controlling slip behaviors in the convergent plate margin, where slips occur in the shear zone (shallow plate interface) in southwest Japan. Chapter “2. Geological Setting” to “4. Methods” focuses on geological setting, data and methods for investigating the shear zone thickness and the geological structures in the research with title “Influence of shear

zone thickness and strike-slip faulting on tectonic tremor in Nankai Trough, southwest Japan”. This study introduced three-steps analysis to investigate these source mechanisms and controlling factors. Step one is processing seismic data and seismic attributes using 2D and 3D seismic reflection data. After step one, I identified structures to map fault distribution from seafloor bathymetric and seismic reflection data. Later, step three is mapping shear zones using several 2D seismic reflection profiles where the shear zone is source of the tectonic tremors.

In Chapter “5. Result and interpretation” and “6. Discussion”, I analysed structures around the shear zone and shear zone thickness to know the source mechanisms and controlling factors of tectonic tremors. To understand their relationships with interplate earthquakes at 2D and 3D seismic reflection in the study area, I discuss the tremor numbers from the shear zone that are controlled by their thickness. The shear zone thickness varies depending on the pore-pressure condition. Moreover, strike-slip faulting can influence tremors changes through direct overburden and diffusion of pore-pressure of the shear zone during the collision with the tectonic backstop. Spatial variation of the shear zone and the strike-slip faults control tectonic tremor distribution. The results of this research can be used to evaluate the area where the shear zone (as indicated by the pore-pressure change) is susceptible to slow earthquakes and related to interplate earthquakes. The content of this chapter was published in the journal *Tectonophysics* (Fahrudin et al., 2022).

Chapter 7 summaries the primary finding from work in this study. In this section, I also went over the research implications and suggested more research.

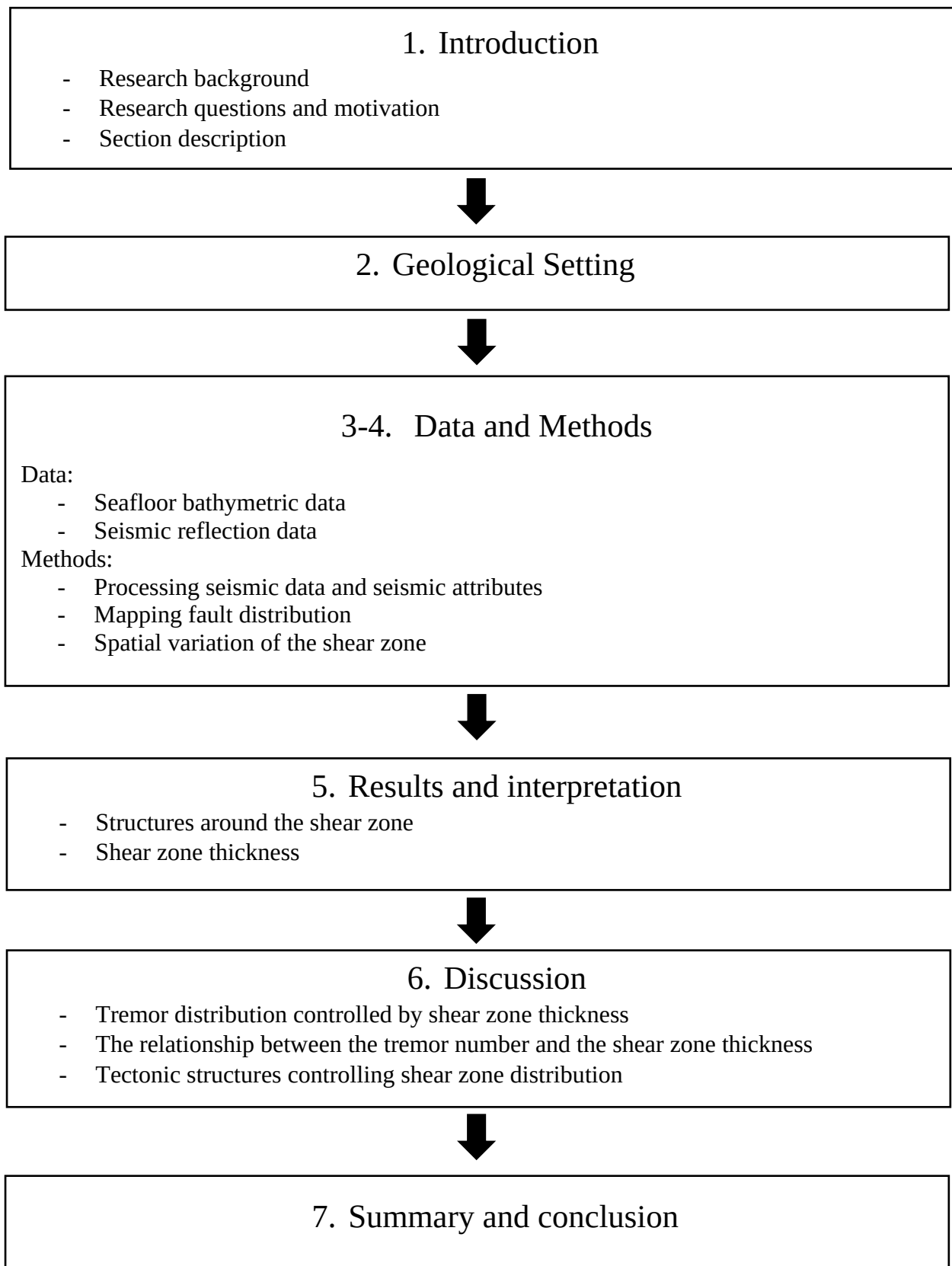


Figure 1.2 Outline of the dissertation

2. Geological Setting

Slow earthquakes are energetic events in subduction zones that are characterized by an episodic style and migrating activity (Annoura et al., 2017; Araki et al., 2017; Nakano et al., 2018; Toh et al., 2018; Takemura et al., 2019; Wallace et al., 2016). They are widely distributed and migrate due to the sensitivity of stress perturbation or pore pressure diffusion at the source points to nearby areas (Annoura et al., 2017). They may be classified on the basis of geodetic observations (i.e., short-term or long-term slow slip events) or seismic data (i.e., very low frequency events (VLFs) and low-frequency tectonic tremor) (Obara and Kato, 2016). Tectonic tremor has a dominant frequency of about 2 to 10 Hz (e.g., Obara and Kato, 2016) and is often associated with VLFs, which have dominant frequencies < 2 Hz (Ide et al., 2007). In subduction zones (i.e., Nankai Trough), the Philippine Sea Plate currently subducts obliquely beneath the accretionary prism (Seno et al. 1993; Miyazaki and Heki 2001), it had subducted perpendicular to the trench until approximately 2 to 4 Ma (Kimura et al. 2005; Takahashi 2006; Ikeda et al. 2009). The change in plate motion reactivated some faults within the accretionary wedge as strike-slip faults that could accommodate trench-parallel shear strain due to the oblique plate subduction.

One common location for slow earthquakes is on the shallow plate boundary near the trench (i.e., shallower than the coseismic plate boundary; Barnes et al., 2020), which is believed as a conditionally stable (Scholz and Campos, 2012) or aseismic zone (Wang and Hu, 2006). A plate interface close to the outer ridge (i.e., transition from outer to inner wedge) has apparently hosted slow earthquakes in the Nankai Trough, southwest Japan (Tsuji et al., 2014a;

Tonegawa et al., 2017; Hendriyana and Tsuji, 2021). Such slow earthquakes were believed to occur in low-velocity zone (LVZ) around the plate boundary. Previous studies (e.g., Tsuji et al., 2007; Park et al., 2010) demonstrated that the LVZ seismic facies are characterized as weak reflectivity and are formed by antiformal stacking of underplated materials. Bangs et al., (2009) also characterized LVZ as weak reflectivity and low density as underthrust materials. Tsuji et al. (2014b) further interpreted LVZ as a shear zone based on high pore pressure as well as discontinuous reflectors. The shear zone is located between basal decollements and underthrust of outer accretionary wedge, and it includes the interpreted underplating sequence. Because this shear zone unit has elevated pore pressure (Kitajima and Saffer, 2012; Tsuji et al., 2014b), it seems to be related to the number of tremor occurrences (Tonegawa et al., 2017). Furthermore, some studies have suggested that thickness variations of the shear zone hosted high pore pressure around the plate interface are correlated with the occurrence of VLFE and tremor, which is consistent with stress accumulation in locked areas (Obana and Kodaira, 2009; Takemura et al., 2019; Hendriyana and Tsuji, 2021) and stress adjustments in the aftermath of regular earthquakes (To et al., 2015; Annoura et al., 2017).

In the Nankai Trough, the source of VLFs from 2004 to 2016 has been attributed to gently dipping thrust faults at a depth of 6–9 km (Sugioka et al., 2012; Nakano et al., 2018), normal faults within the accretionary prism (To et al., 2015), or strike-slip faults at greater depths (Nakano et al., 2018). Their locations are bounded by mega-splay faults (Toh et al., 2018) and imbricate thrust faults (Shiraishi et al., 2020) landward of the basement high of oceanic crust (Tsuji et al., 2009). Geological structures around Kii Peninsula observed the tectonic backstops with volcanic materials. Gabbroic rock is observed at Cape Shionomisaki that has been dated to approximately 15–13 Ma (Hoshi et al., 2003). An igneous dome is clearly observed on the seismic profiles from off Cape Shionomisaki (Tsuji et al., 2013b). Because the structures at Cape Shionomisaki coincides with a gravity anomaly and high seismic velocity

area (Kodaira et al., 2006; Kimura et al., 2014), that reflector could be interpreted as the surface of the igneous dome with high density. Since the thick accretionary prism is developed on its trench ward side of the igneous dome (Tsuji et al., 2013b), the igneous dome could also work as static backstop (i.e., igneous backstop) (Figure 2.1).

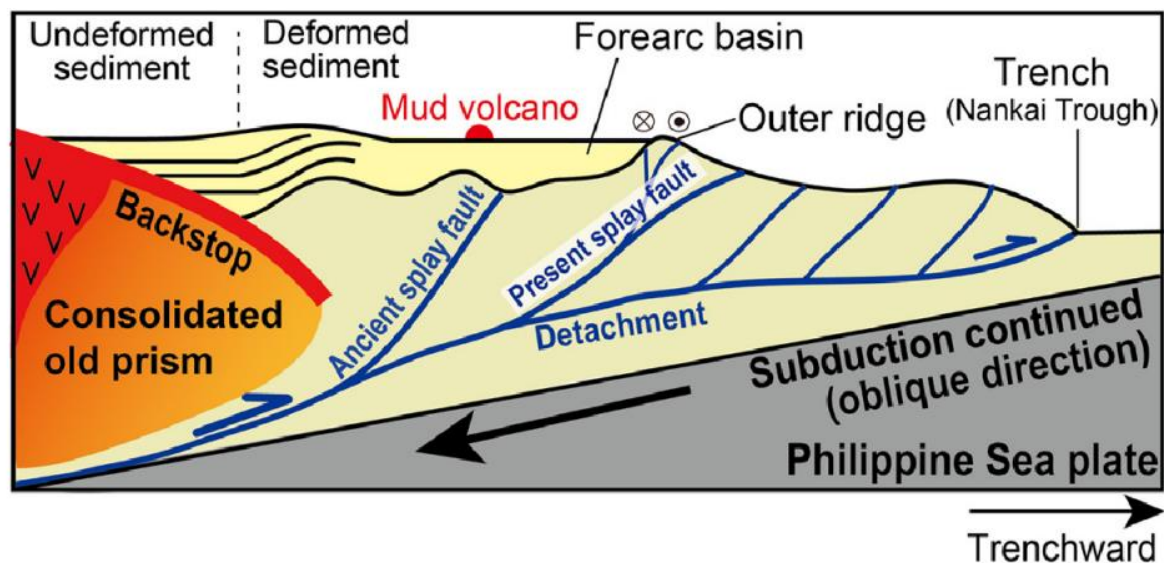


Figure 2.1. Schematic diagrams showing the collision tectonic backstop and the accretionary prism off Kumano at 2 Ma – present (Tsuji et. al., 2015).

3. Data

This study identified the shear zone and relevant fault systems by using bathymetric and seismic reflection datasets. The bathymetric data revealed the traces of oblique strike-slip faults (Yamada et al., 2013; Azevedo et al., 2018), and 2D and 3D seismic reflection data were used to characterize the seismic facies, calculate thickness of the shear zone (Tsuji et al., 2014b), and interpret the relevant geological structures, such as strike-slip faults. Tremor source locations obtained from Hendriyana and Tsuji (2021) were then overlaid on the resulting map of shear zone thickness.

3.1. Seafloor bathymetric data

A high-resolution bathymetric map of the study area off the Kii Peninsula (Figs. 3.1.) was produced from multibeam echo sounder data acquired during cruises YK01-04, YK02-02, and YK03-03 by the Japan Agency for Marine-Earth Science and Technology (JAMSTEC; Tsuji et al., 2014a). I mapped structures and their boundaries between the outer accretionary complex and the trench slope basin on the basis of their elevation and topography with the spatial resolution of ~100 m (Ashi et al., 2007). Dextral strike-slip faults and anticlines were visible as seafloor lineaments. The lineaments were mainly interpreted as anticlines and thrusts, and as slump scars in the eastern region. Two prominent strike-slip faults (SF-A and SF-B; Yamada et al., 2013; Azevedo et al., 2018) have displaced these anticlines and thrusts in the outer wedge area.

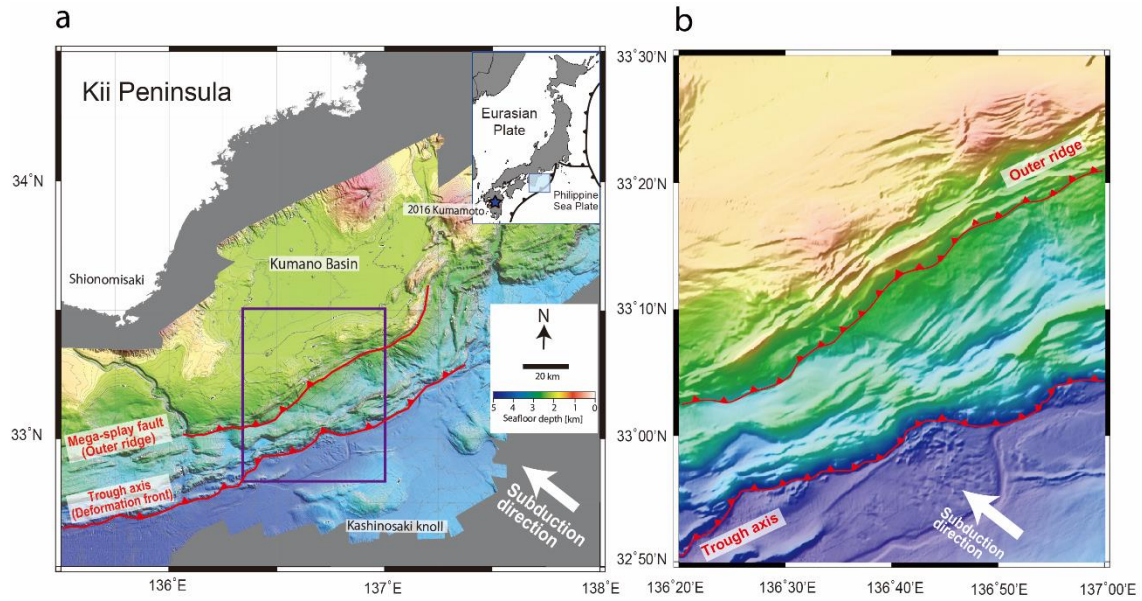


Figure 3.1. (a) Bathymetric features of the Nankai Trough off the Kii Peninsula without the detailed interpretation. Purple box shows the location of panel (b). (b) Detail of the study area showing bathymetric without the detailed interpretation.

3.2 Seismic reflection data

This study used 27 profiles of 2D and 3D multichannel seismic reflection data acquired by the JAMSTEC (Figs. 3.2, 3.3). The ODKM-03 survey, conducted in 2003 by R/V Polar Princess, employed a tuned airgun array with a total volume of 70 L (4240 inch³), fired at 50 m intervals, and a streamer 6 km long with 480 receivers. The KR0211 survey, conducted in 2002 by R/V Kairei, used an airgun array with a volume of 200 L (12,000 inch³), fired at 50 m intervals, and a streamer about 5 km long with 204 receivers (Tsuji et al., 2014a).

Here I mainly showed the seismic profiles in lines A, 16, and I of the ODKM-03 survey (Figs. 3.2a-3.2c) and line D7 of the KR0211 survey (Fig 3.2d). We applied a multi-term higher-order normal move-out correction (Castle, 1994) in addition to the conventional techniques of noise trace editing, 3–60 Hz bandpass filtering, deconvolution with a 24-ms predictive distance,

amplitude compensation for spherical divergence, velocity analysis, multiple suppression, dip move-out, muting, stacking, and post-stack time migration.

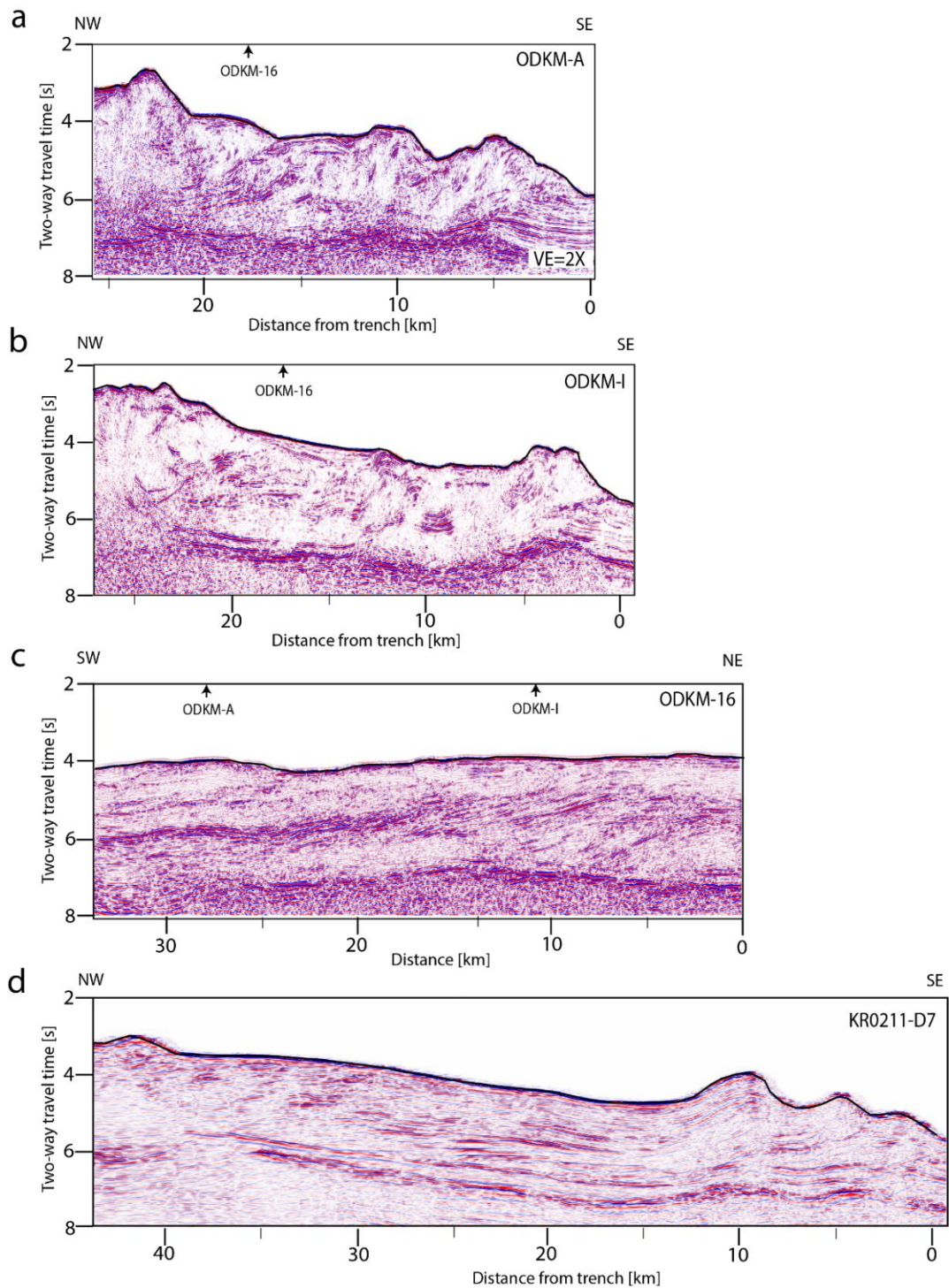


Figure 3.2. 2D seismic profiles of the outer accretionary wedge without interpretation; (a) ODKM-A, (b) ODKM-I, (c) ODKM-16, and (d) KR0211-D7. These profiles are same as in Figure 5.2

I also used 3D seismic data acquired in 2006 by M/V Nordic Explorer from a 12×56 km survey area (Moore et al., 2009; blue outline in Fig. 5.1a; 3.3). Data processing included bandpass filtering, spherical divergence corrections, FK filtering, velocity analysis, and 3D anisotropic Kirchhoff pre-stack time migration. We also applied 3D pre-stack depth migration to the data using a velocity model obtained by a tomography-based approach. The details of processing method are explained in chapter 4. Then, the 3D survey area includes a plate boundary décollement, mega-splay faults within the accretionary prism, and a shear zone. The detailed structure of the dextral strike-slip fault was identified on cross sections of the 3D seismic section.

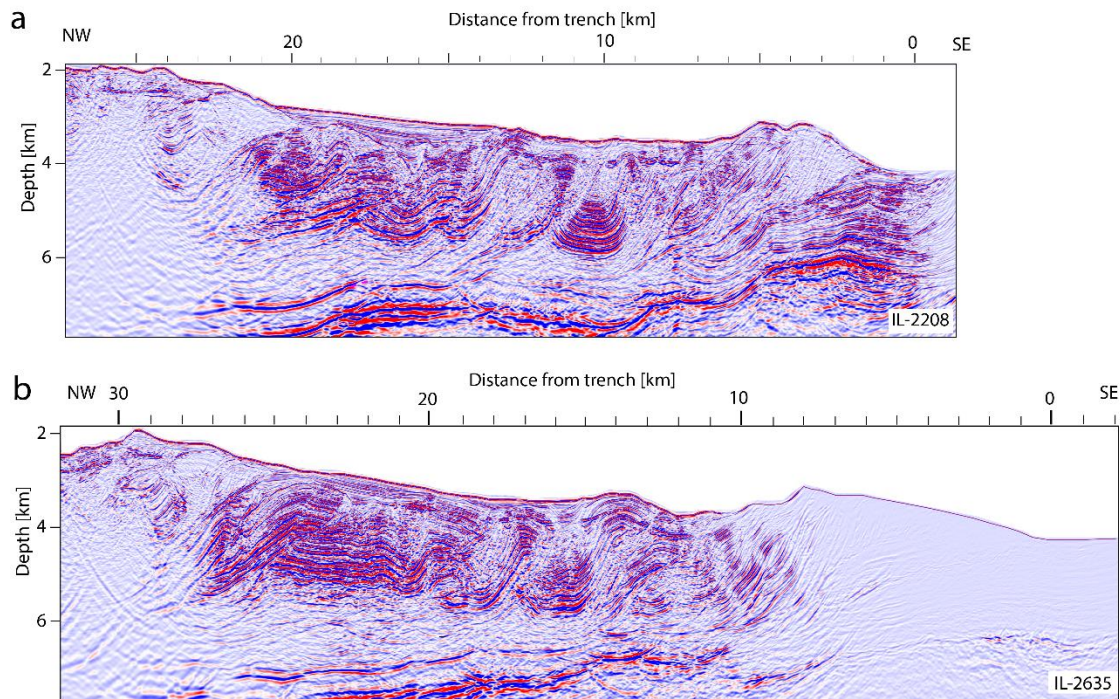


Figure 3.3. The reflection seismic profiles without interpretation derived from 3D seismic data. These profiles are same as in Figure 5.3

4. Methods

4.1. Processing seismic data and seismic attributes

The 2D seismic data was acquired in the Nankai accretionary prism off Kumano, and its survey lines extend close to the coast of the Kii peninsula that located between the outer ridge and the trench. Data processing for the 2D seismic data involved filtering, velocity analysis, stacking, deconvolution, and post-stack migration. To extract horizons of top and bottom of the shear zone from intersecting seismic reflection lines, we did not apply depth conversion to the time-domain seismic profiles, because the seismic velocity was not precisely ascertained in the deep oceanic crust due to small moveout in the velocity model. Indeed, the paucity of reflectors within the shear zone prevented us from estimating seismic velocities for a depth conversion. However, I use the assumption $V_p = 2.8$ km/s to estimate the top and bottom shear zones from the 3D seismic data.

In addition to processing the 2D seismic data, I used the processing of the 3D seismic data. We applied 3D pre-stack depth migration to the 3D seismic data using a tomography-based approach to obtain the high-resolution P-wave velocity distribution. In principle, the stages in seismic data processing can be grouped into 3 stages: pre-processing, main processing, and post-processing. Furthermore, the process includes demultiplexing, trace gathering, editing, and muting, gain recovery, static correction, filtering, deconvolution, normal move out, stacking, velocity analysis, and migration (Yilmaz, 2001; Moore et al., 2009).

In the first stage, Petroleum Geo-Services (PGS) supplied a 3-D stack and post-stack migration, which proved valuable in understanding the regional seismic reflection characteristics and selecting parameters. PGS performed a resample from 2 ms to 4 ms and implemented a low-cut filter at 5 Hz with an 18 dB/octave slope during the demultiplexing process, as illustrated in Fig. 4.1. This step involved converting the data into a user-friendly format that is consistently employed throughout the processing. The specific format is determined by the processing system utilized by Petroleum Geo-Services (PGS). An industry-standard format commonly employed for data exchange in the seismic field is SEG-Y.

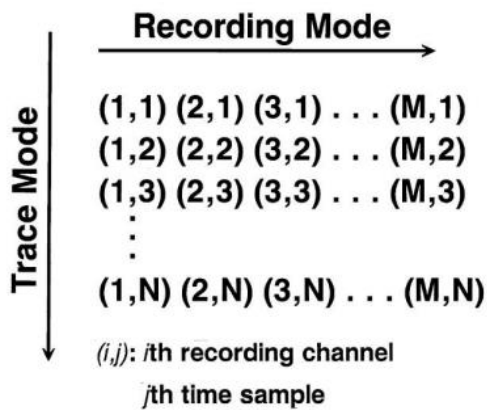


Figure 4.1. Demultiplexing is the process of organizing the data into separate columns based on samples, where all the time samples from one channel are grouped together before moving on to the samples in the subsequent channels (Yilmaz, 2001).

Pre-processing also eliminates noise that occurs during seismic recording. The noise present in seismic data comes in the form of multiples. Multiples are reverberations between zones with low velocity or low frequency but high amplitude, such as the ocean surface and seafloor. Strong multiples originating from the seafloor were mitigated prior to the migration processing stage. We implement swell noise attenuation and linear noise attenuation using FK

filtering to remove or reduce this noise from the seafloor or other features. Then, we discard adjacent traces after applying the FK spatial anti-alias filter (receiver interval = 12.5 m).

Seismic waves originating from seismic sources undergo convolution as they travel through the medium until the geophone detects them. Consequently, the medium (earth) acts as a filter, causing the originally sharp and high-amplitude seismic wavelets from the seismic sources to become wider and have reduced amplitudes over time. Deconvolution is a process aimed at correcting the filtering effects of the earth, allowing the recorded wavelets to be restored to their original sharpness and high amplitudes (as shown in Fig. 4.2).

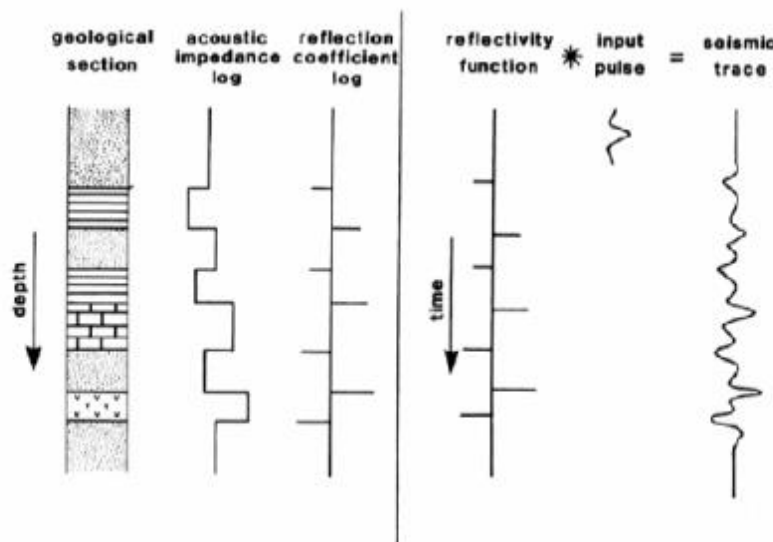


Figure 4.2. The model that illustrate the concept of convolution (Yilmaz, 2001).

In addition to providing an improved signal-to-noise ratio, we applied 3D pre-stack depth migration to the data using a velocity model. Pre-Stack Depth Migration (PSDM) velocity analysis was performed with horizon and grid-based tomography within 31 iterations (Moore et al., 2009). The reflected signals present in seismic traces carry information about the

subsurface layer velocities. The purpose of velocity analysis is to obtain sufficiently accurate velocity values in determining the depth, thickness, and dip of a reflector. Also, we employed P-wave velocity structures derived from the wide-angle ocean-bottom seismograph study to characterize seismic velocities within the deep oceanic crust in pre-stack depth migration.

The study utilized four receiver cables positioned at intervals of 150 meters, each with a length of 4500 meters, and 360 receiver groups spaced 12.5 meters apart. This configuration of a 4-streamer/2-source array resulted in 8 common midpoint (CMP) lines per sail line, spaced at 37.5 meters and providing nominal 30-fold data. Because of the offset distance between the seismic wave source and the geophone on a trace originating from one CMP (Common Mid-Point) or CDP (Common Depth Point), there are any effects. Furthermore, Normal Move Out (NMO) correction is needed to reduce the effects and to get the maximum results (as shown in Fig. 4.3). NMO corrected with residual move out (RMO) and time of arrival (ETA) values every 50 m x 37.5 m.

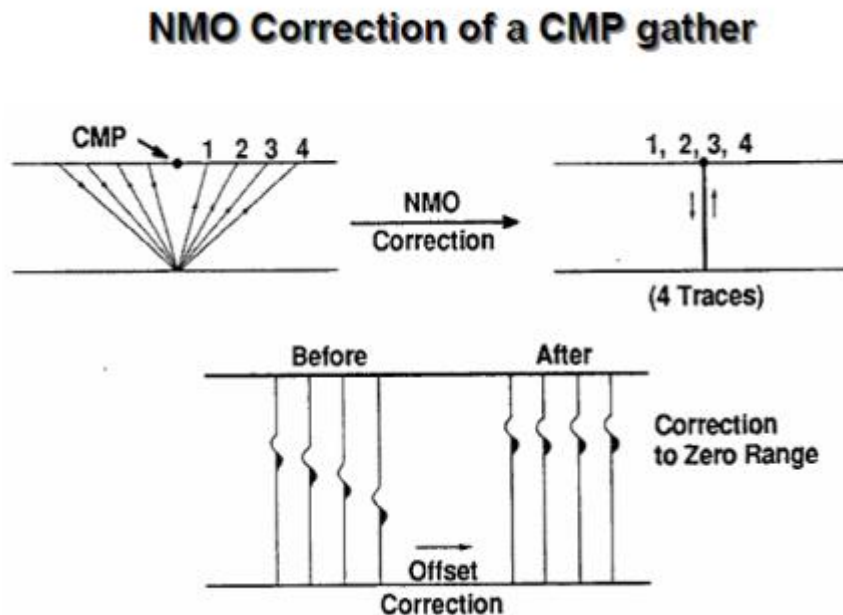


Figure 4.3. The concept of performing NMO correction on a CMP gather (Yilmaz, 2001).

Stacking involves summing seismic traces within a single data gather to enhance the signal-to-noise ratio (Fig. 4.4). When all the traces have been corrected, each reflection in the CDP gather format becomes horizontal, and if the horizontally transformed traces are stacked in each CDP, the S/N ratio can be raised. By stacking the horizontally aligned traces within each CDP, the signal-to-noise ratio can be further improved. To obtain a reflection profile aligned with the subduction direction, we conducted stacking of the imaging result along the axis of the trough. The resulting 2D pseudo profile is oriented in the NNW-SSE direction.

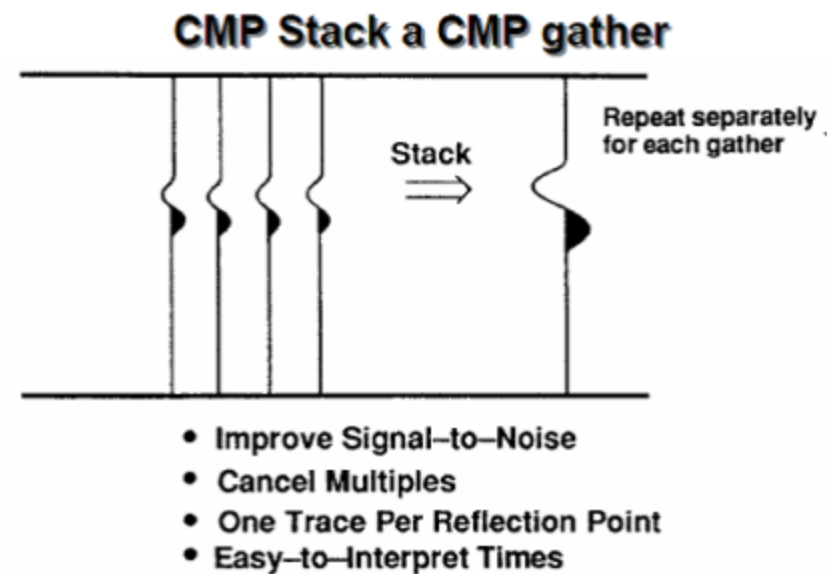


Figure 4.4. Stacking processing at CMP gathers (Yilmaz, 2001).

Wave propagation direction was determined based on the configuration of Ocean Bottom Seismometers (OBS), and unnecessary ray paths were eliminated by assuming that northward wavefield propagation was predominant. Using cross-correlation, pseudo-active source seismic data was generated and then migrated using a velocity model derived from a P-wave tomographic velocity structure. The velocity structure was smoothed and extended along

the axis of the trough to create a comprehensive 3D velocity model. The resulting migrated section, displayed with scaled amplitudes (Fig. 4.5), aims to represent the geological cross-section along the line traverse. Although the migration output is typically displayed in time, similar to the input stacked section, it effectively reduces the diffraction effects caused by structural elements like faults and folds, allowing for a clearer and more detailed representation of the subsurface structure.

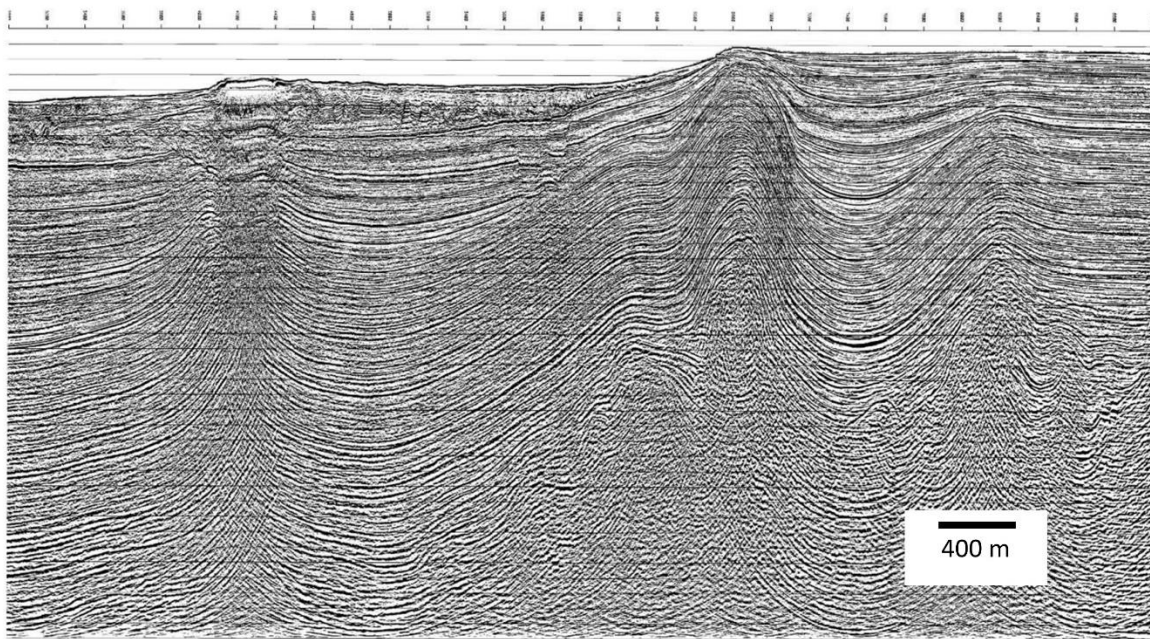


Figure 4.5. The seismic profile shows structures and layers of rocks as the result of migration processing (Yilmaz, 2001).

In this study, the displacement along strike of strike-slip faults is calculated at the depth of 3600 m using seismic attribute (coherence; Fig. 4.6). Through the measurement of these faults, we compared a single fault (i.e., SF A and SF B) with lineaments pattern using the seismic attribute (coherence) from the 3D seismic reflection data (e.g., Taner et al., 1979; Marfurt et al., 1998) in depth-domain to know associating fault system. Coherency defined discontinuity attributes as the normalized energy ratio within the measurement window by the

total trace energy (Marfurt et al., 1998; Ofuyah et al., 2014). The coherency defined the discontinuity attributes as the energy ratio normalized within the measurement window by the total trace energy. This attribute effectively maps the fault in the shallow geological formation (e.g., pop-up structures), where coherent reflectors are identified. Here, I characterize the strike-slip faults and their associated faults based on this seismic attribute (time-slice domain) because extracting the strike-slip faults based on the original profiles. Therefore, the geometry of strike-slip faults as representing of displacements can indicate difference morphologies of seafloor.

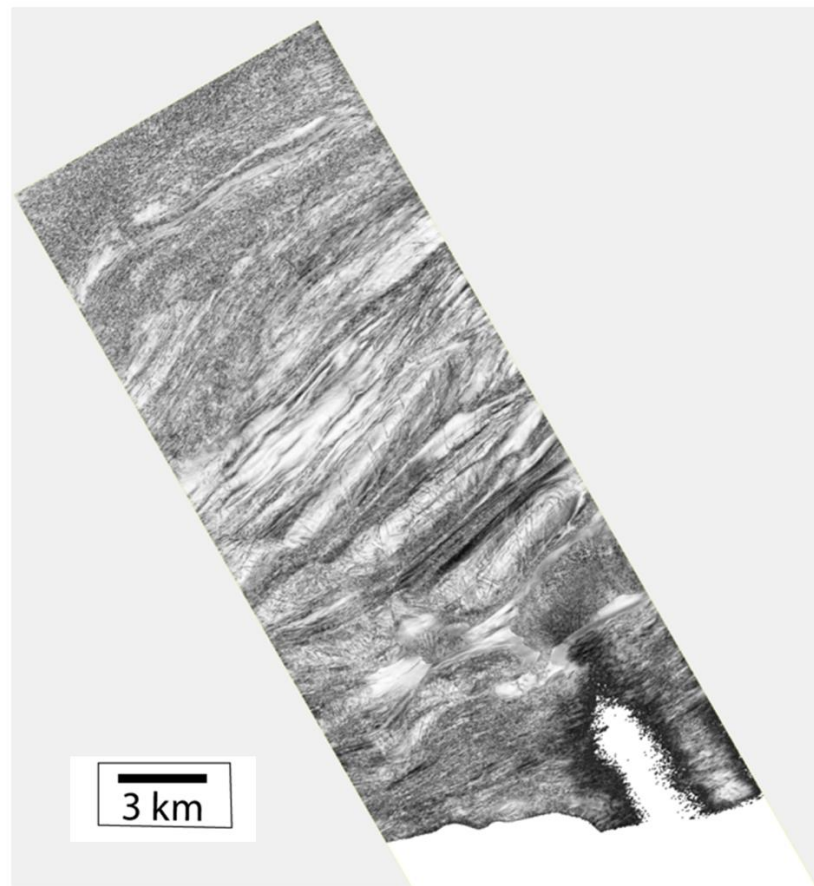


Figure 4.6. Depth slice of the coherence attribute generated from 3D seismic reflection data (3600 m below sea surface).

The coherence attribute in 3D seismic data is very effective for mapping faults and other geological structures, such as folding, using the algorithm formulated by Marfurt et al., (1998). The 3-D seismic coherency cube can be highly effective in identifying seismic faults when there is a significant lateral variation in acoustic impedance (Fig. 4.6). The coherency algorithm relies on the conventional normalized cross-correlation method. Initially, the in-line (l -lag) cross-correlation (ρ_x) is defined at time (t) between data traces (u) at positions (x_i, y_i) and (x_{i+1}, y_i), to be

$$\rho_x(t, l, x_i, y_i) = \frac{\sum_{\tau=-w}^{+w} u(t-\tau, x_i, y_i) u(t-\tau-l, x_{i+1}, y_i)}{\sqrt{\sum_{\tau=-w}^{+w} u^2(t-\tau, x_i, y_i) \sum_{\tau=-w}^{+w} u^2(t-\tau-l, x_{i+1}, y_i)}} \quad (\text{Eq. 4.1})$$

where w represents the duration of the correlation window in time.

Next, the formula establishes the cross-line (m -lag) cross-correlation (ρ_y) at time (t) between data traces (u) at (x_i, y_i) and (x_i, y_{i+1}) to be

$$\rho_y(t, m, x_i, y_i) = \frac{\sum_{\tau=-w}^{+w} u(t-\tau, x_i, y_i) u(t-\tau-m, x_i, y_{i+1})}{\sqrt{\sum_{\tau=-w}^{+w} u^2(t-\tau, x_i, y_i) \sum_{\tau=-w}^{+w} u^2(t-\tau-m, x_i, y_{i+1})}} \quad (\text{Eq. 4.2})$$

These in-line (l -lag) and cross-line (m -lag) correlation coefficients can be merged to produce a 3-D coherency estimate, ρ_{xy} .

$$\rho_{xy} = \sqrt{[\max \rho_x(t, l, x_i, y_i)] [\max \rho_y(t, m, x_i, y_i)]} \quad (\text{Eq. 4.3})$$

where, the lags l and m for which (ρ_x) and (ρ_y) reach their maximum values are represented by $\max \rho_x(t, l, x_i, y_i)$ and $\max \rho_y(t, m, x_i, y_i)$. The algorithm utilizes a semblance analysis across a variable number of traces to estimate coherency.

4.2. Mapping fault distribution

To map fault distribution in the study area, I use seafloor bathymetric, coherence attributes, and seismic profiles. I mapped structures (dextral strike-slip faults and anticlines by lineaments in bathymetric data. Also, coherence attributes were calculated from the 3D seismic reflection data (Fig. 4.6). In addition, I interpret the fault using the profiles of the 2D and 3D reflection seismic data. By using lineaments in bathymetric data, I identified an anticlinal ridge structure trending ENE–WSW related to imbricate thrusts. The structures are cut by the strike-slip faults. While the coherence attributes are measured at a depth of 3600 m. Measurement at that depth are in order to obtain a relevant structure which links the surface structures (bathymetric data) and the appearance of deeper structures using 3D reflection seismic data.

4.3. Spatial variation of the shear zone

The shear zone was mapped in the outer wedge area of the Nankai accretionary prism by using several 2D seismic reflection profiles. Firstly, I define the shear zone based on the structures as well as seismic attributes. The bottom of shear zone can be defined as a basal décollement that was clearly imaged on seismic profiles (approximately 1 km above the oceanic crust). The polarity of the basal décollement is positive (i.e., same polarity with seafloor reflector); thus, it could work as the lower boundary of the low-velocity overpressured shear zone as well as lower boundary of plate boundary fault in this area. On the other hand, the upper horizon of the shear zone cannot be defined as a clear reflector. Therefore, I defined the seismic facies and physical characteristic of shear zone on the basis of (1) relative low P-wave velocity (around 2.7–3.2 km/s; Park et al., 2010; Kamei et al., 2012) hosted high pore pressure, (2) chaotic, discontinuous and weak reflectors (Tsuji et al., 2014b) and (3) base of the imbricate thrusts sequence (i.e., the boundary between the coherent imbricated thrust unit

and the discontinuous reflector unit) from the 3D seismic data. Indeed, the upper boundary of low reflectivity zone with discontinuous reflector is well consistent with that of low velocity zone.

Furthermore, the shear zone is recognized as the low velocity zone (LVZ) as this zone has the velocities contrast with depth between accretion sediment and basal decollement. Akuhara et al. (2020) estimated slower S-wave velocity (V_s) in the shear zone ($V_s = 1.0 - 1.6$ km/s) compared to that in the accretion sediment ($V_s = 1.5 - 1.8$ km/s) (Figure 4.7). Lower V_s in the shear zone indicates that fluid is confined in the volume between the décollement and the base of accreted sediment.

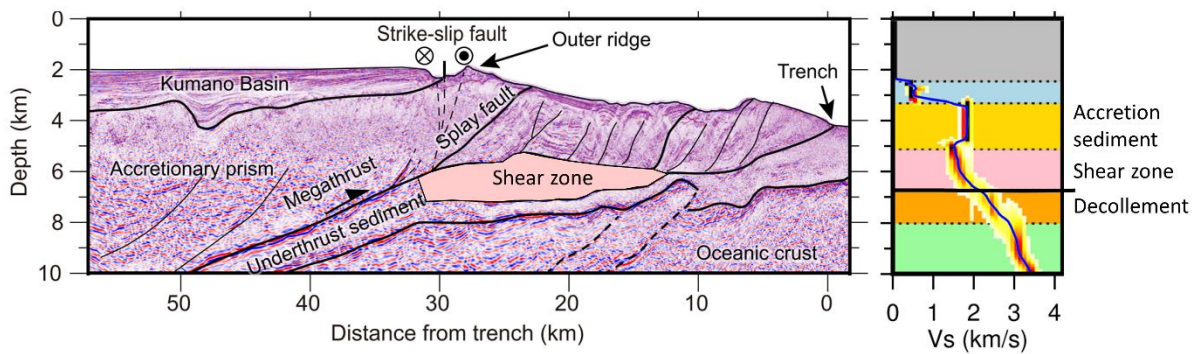


Figure 4.7. The interpreted profile showing the shear zone and structures. Posterior marginal probability distributions of the S-wave velocity (V_s) (Akuhara et al., 2020).

For all of the 2D regional seismic reflection profiles, shear zones were delineated in the time domain, because the poorly constrained seismic velocity precludes matching the depths of target reflectors across multiple profiles in fence diagrams. I manually selected the top and bottom horizons of the shear zone on the 2D profiles and validated the 2D results with a 3D seismic volume. The thickness of the shear zone can then be calculated by assuming a constant seismic velocity in the shear zone; For time-to-depth conversion of seismic profiles, I used a constant P-wave velocity derived from waveform inversion ($V_p = 2.8$ km/s; Tsuji et al. 2014b),

because I cannot obtain an accurate velocity model for each 2D seismic line. The P-wave velocity derived from 2D seismic data include a large error, thus I do not want to include such additional error in depth conversion.

Because of the poorly constrained seismic velocity of the 2D seismic data, the definition of the top of the shear zone has uncertainty. Its uncertainty includes the space between two points and an interval spanning of seismic data (Fig. 4.8). The space between two points of the top horizon of the shear zone is less than 100 m. Then, a suite of seismic attributes was extracted from an interval spanning from 10 ms below the base of the top horizon of the shear zone to 10 ms above the top of the shear zone. Thus, an uncertainty of top horizon of the shear zone is ~28 m.

After calculating the thickness of the shear zone on the 2D seismic profiles, I traced its spatial distribution. I then overlaid the resulting map with the locations of tectonic tremor from October 2015 to December 2016, which were inferred from polarization attributes using the broadband seismometers of the Dense Ocean-floor Network for Earthquake and Tsunamis (DONET; Hendriyana and Tsuji, 2021).

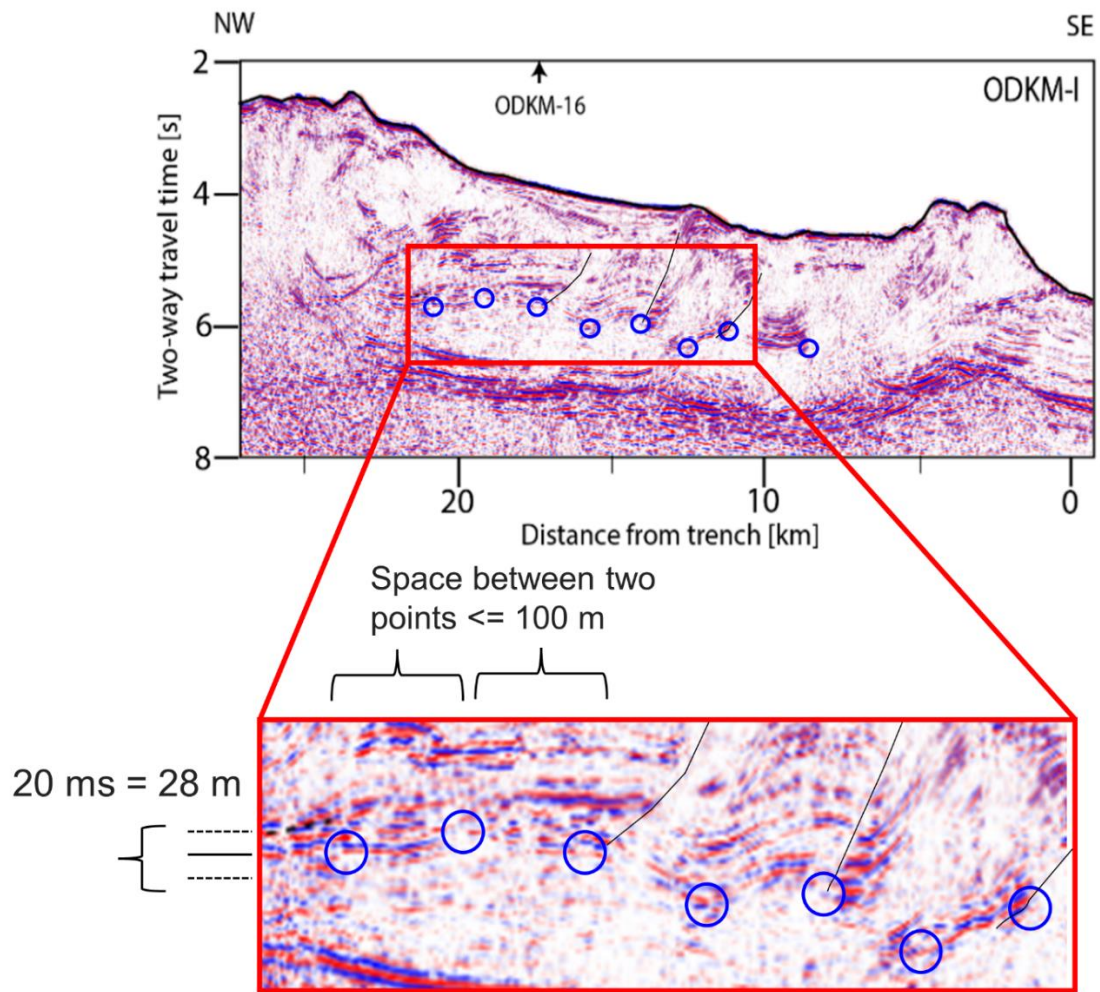


Figure 4.8. The 2D seismic profiles of the outer accretionary wedge (ODKM-A) shows the top of the shear zone. The top of the shear zone indicates the blue cycles. It has an *uncertainty*.

5. Results and interpretation

5.1. Structures around the shear zone

The Nankai Trough is where the Philippine Sea plate subducts beneath the Japanese Islands at an azimuth of 300° – 315° and a rate of 4.1–6.5 cm/year (Fig. 5.1a; Miyazaki and Heki, 2001). Because of the oblique angle of subduction, a number of mega-splay and strike-slip faults are distributed in the accretionary wedge (Figs. 5.1-5.5). Around the trough axis off the Kii Peninsula, a wedge of sand-rich trench sediments approximately 1.3 km thick overlies about 1.1 km of sediments of the Shikoku Basin, which lie in turn on igneous basement (Moore et al., 2009; Tilley et al., 2022; Fig. 5.3). To characterize the shear zone, I focused mainly on the outer wedge and the outer ridge (i.e., transition zone between outer and inner wedges), where tectonic tremors have intensively occurred (Hendriyana and Tsuji, 2021). As described above, the shear zone can be identified by its low reflection strength, and several discontinuous reflectors are distributed within it in seismic profiles. These features are interpreted as discontinuous structures with high pore pressure conditions (blue area in Fig. 5.3; Tsuji et al., 2014b). The top of the shear zone is shaped by the roots of a series of imbricate thrusts (Figs. 5.3, 5.4), where the root reflector could partially work as a plate interface. The bottom of the shear zone is a continuous reflector that was interpreted as basal décollement.

Based on the internal reflection characteristics, I roughly defined two types of the shear zone (Figs. 5.2c). In the northeast (NE) section, the shear zone is characterized as discontinuous southwest (SW) dipping reflectors (green lines in Fig. 5.2c). The SW dipping reflectors could indicate that some imbricated thrusts sole into the basal décollement or antiformal stacking occurs there. On the other hand, the profile at the SW area indicates that horizontal discontinuous reflections distribute within the shear zone. Their reflectors correlated with

heterogeneous rocks (chaotic and weak) such as mélangé, although it is difficult to estimate the formation types (Tsuji et al., 2014b).

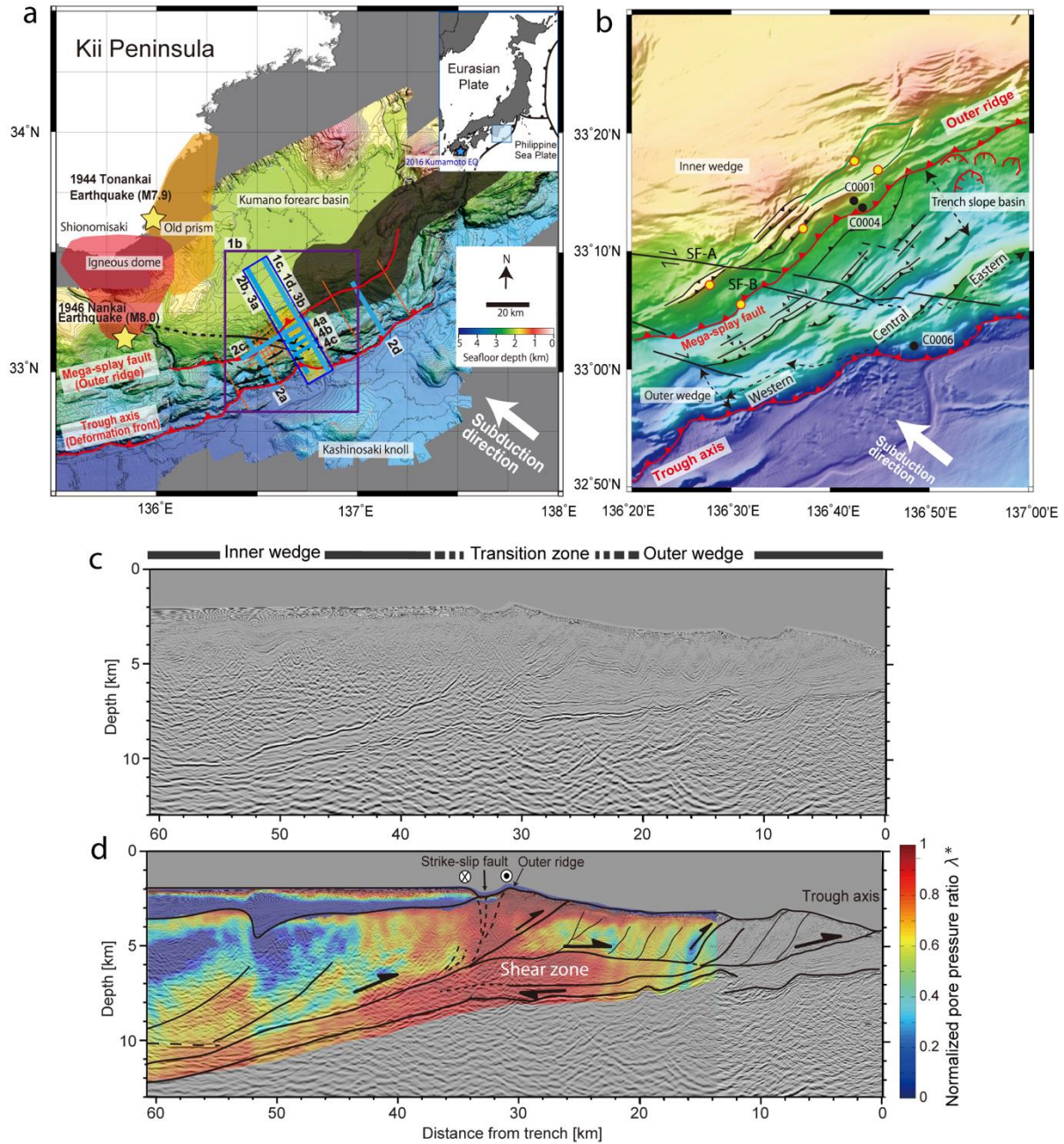


Figure 5.1. Tectonic, bathymetric, and geologic features of the Nankai Trough off the Kii Peninsula. (a) Regional setting and locations of significant features, including epicenters of the 1944 Tonankai and 1946 Nankai earthquakes (yellow stars), igneous intrusions (red shading) and volcaniclastic prism (orange shading) (Tsuji et al., 2015), the paleo-Zenisu ridge (gray shading; Park et al., 2003), 3D and 2D seismic reflection lines (orange and blue lines) and 3D seismic reflection data (blue box),

interpreted active strike-slip fault (black lines). The bold blue line indicates profiles shown in this study. Purple box shows the location of panel (b). Inset map shows the location, major plate boundaries, and epicenter of the 2016 Kumamoto earthquake. (b) Detail of the study area showing the extent of the trench slope basin and outer accretionary wedge (arrows), strike-slip faults SF-A and SF-B (black lines), thrust faults (black lines with triangle), anticline axes (black lines with across arrows), cold seep (yellow circles; Ashi et al., 2007), and location of wells determining the maximum horizontal stress (SHmax) (black dots; Lin et al., 2016). (c) Seismic reflection profile indicated in blue in panel (a). (d) The interpreted profile showing the normalized pore pressure ratio (Tsuji et al., 2014b), defined by $\lambda^* = (P_f - P_{hf}) / (P_c - P_{hf})$, where P_c is the confining pressure, P_f is the pore pressure, and P_{hf} is the hydrostatic pressure.

To map fault structures in the study area, I relied on the seafloor bathymetry (Fig. 5.1) and coherence attribute map (Fig. 5.5), and identified an anticlinal ridge structure trending ENE–WSW that is related to imbricate thrusts. I clearly identified two right-lateral strike-slip faults (SF-A and SF-B) from offsets of features in the coherence map (Fig. 5.5a) as well as seafloor bathymetry (Fig. 5.1b). Based on the coherence map, the displacement on the strike-slip fault (SF-A) is ~1 km and accounts for the offset of the anticlinal structures on the imbricate thrust faults. Furthermore, three conjugate faults (F1, F2, F3) can be clearly identified on the profile (Fig. 5.4) and coherence attribute map (Fig. 5.5). The high-angle shear of the conjugate faults (67° to 75°) is associated with displacements of SF-A and accommodates the internal counterclockwise rotation of the uplifted block (red shading in Fig. 5.5b). This uplifting could be partially related to the spatial variation of the shear zone thickness (Fig. 5.4, 5.6).

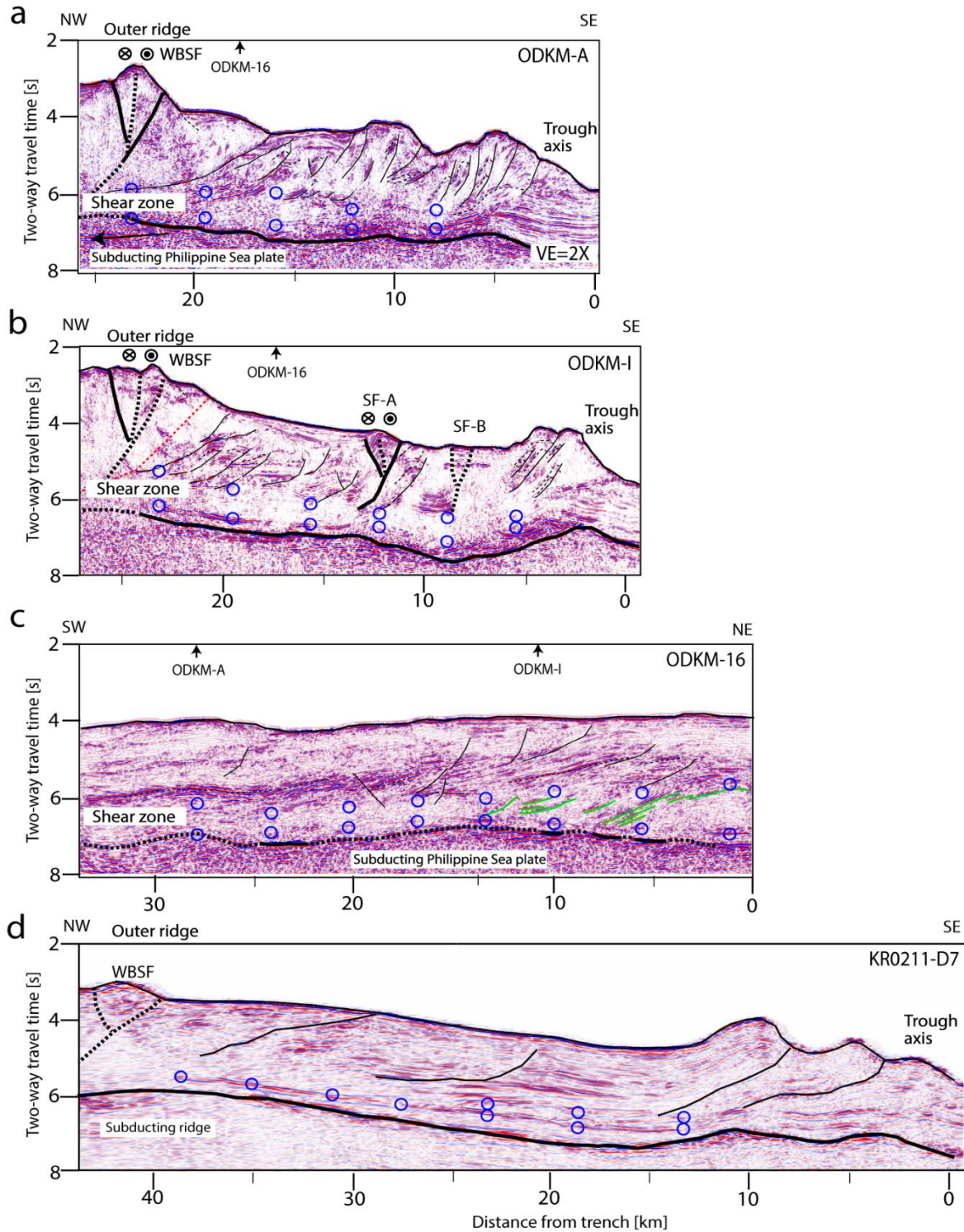


Figure 5.2. Interpreted 2D seismic profiles of the outer accretionary wedge; (a) ODKM-A, (b) ODKM-I, (c) ODKM-16, and (d) KR0211-D7. These locations are on Fig. 5.1a. The shear zone is the low-reflectivity zone between blue circles above the oceanic crust seaward of the outer wedge. The NE of profile (c) showing shear zone with the discontinuous SW dipping of reflection in northeast area (green lines). Other structures include strike-slip faults (bold black and dash black lines; SF-A and SF-B) and imbricate thrusts (thin black lines).

This fault geometry (i.e., strike-slip faults and their conjugate faults) is consistent with the maximum horizontal stress (SHmax) direction determined from borehole breakouts trended $\sim 330^\circ$ to $\sim 335^\circ$ (or parallel to the plate convergence vector) (Lin et al., 2016). Although Lin et al. (2016) interpreted SHmax as the intermediate principal stress direction (σ_2), Chang et al. (2010) have interpreted it as the maximum principal stress (σ_1), indicating a change with depth from normal faulting to strike-slip or possibly thrust faulting. Tsuji et al. (2011) have used shear wave splitting to estimate that SHmax had a similar direction in the outer wedge area.

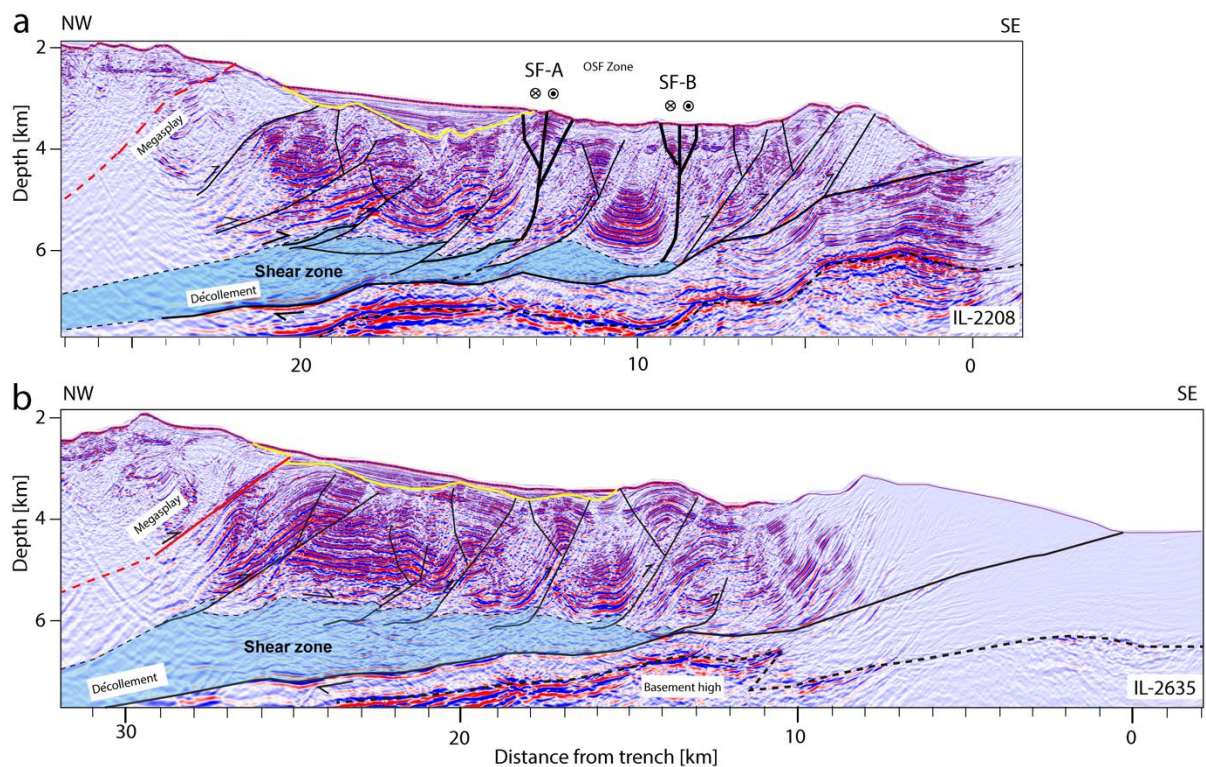


Figure 5.3. Interpreted reflection profiles derived from 3D seismic data (blue rectangle in Fig. 5.1a). (a) Profile of inline 2208 showing horizons and structures include bottom of quaternary sediment (yellow line), the shear zone (blue), the oceanic crust interface (bold dashed black line), imbricate thrust faults (black lines) and the oblique strike-slip faulting zone, defined by SF-A and SF-B (bold black lines). (b) Profile of inline 2635 showing the shear zone thickening landward and imbricate thrusts. Location map of seismic profiles shown in Figure 5.1a (blue lines).

In addition, SF-A and SF-B bound symmetrical pop-up structures (Fig. 5.4), evident as domal uplifts with positive structural and topographic relief (Mcclay and Bonora, 2001). For example, in pop-up 1 (Fig. 5.4), SF-A produced an area of uplift with ~3.8 to 5.4 km width and 150 to 200 m higher than its surroundings, bounded by steeply dipping reverse or strike-slip faults rooted in the shear zone. The fault geometry including the pop-up structures evidently indicates the positive flower structures (Fig. 5.2b), although negative flower structure could be also identified in other parts. Pop-up 1 was formed by SF-A, whereas SF-B generated pop-up 2 (Figs. 5.4c, 5.5c). These features indicate that lateral slip on these strike-slip faults may produce symmetrical pop-ups in zones of oblique strike-slip faulting (Fig. 5.4). As I discuss below, these strike-slip faults influence upon the shear zone distribution (Fig. 5.6).

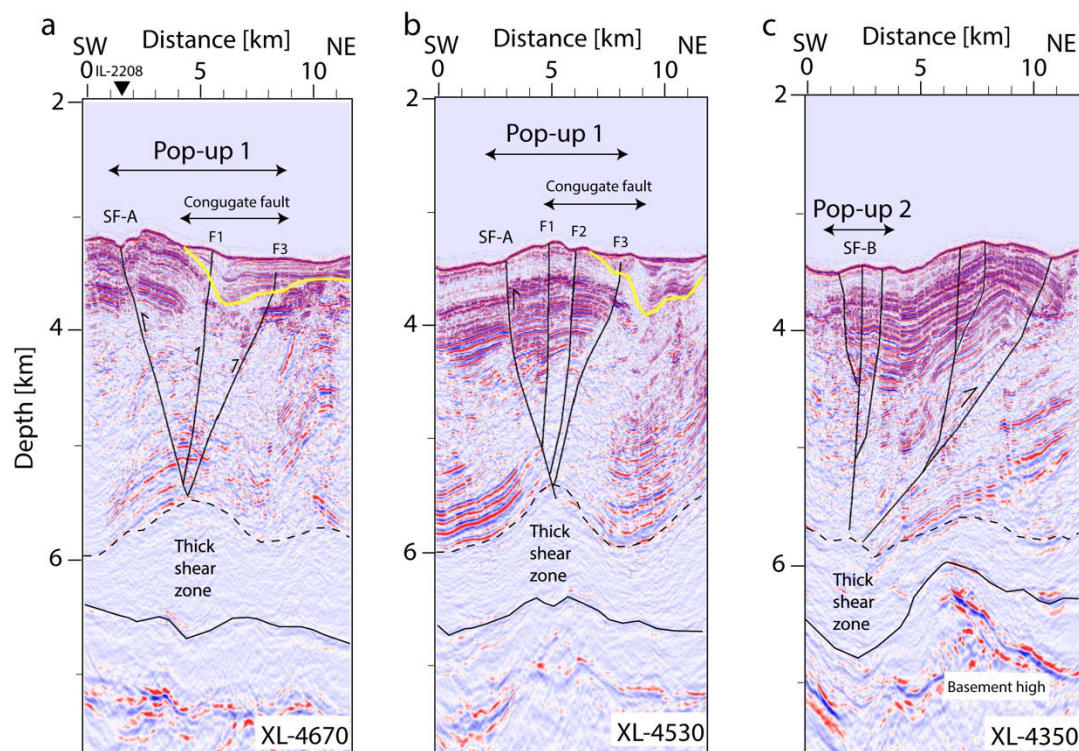


Figure 5.4. Interpreted reflection profiles derived from 3D seismic data (trench-parallel direction). These locations are shown as blue lines in Fig. 5.1a and arrows in Fig. 5.5a for detailed locations. (a, b) Pop-up structure related to SF-A. (c) Pop-up structure related to SF-B. Also shown are the top of

the shear zone (broken line), the décollement (solid line), and the base of Quaternary slope sediment (yellow line).

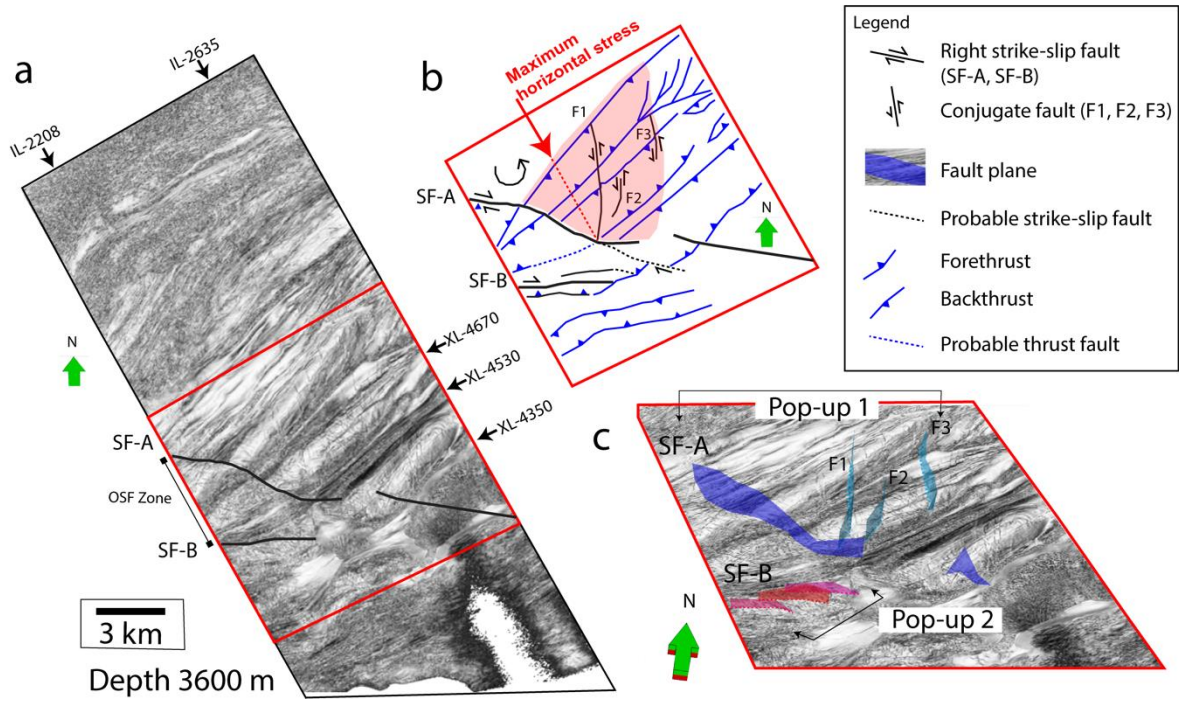


Figure 5.5. (a) Depth slice of the coherence attribute generated from 3D seismic reflection data (3600 m below sea surface). SF-A and SF-B are indicated by black lines. (b) Schematic interpretation of the area outlined in red rectangle in panel (a). Red arrow indicates the maximum horizontal stress. (c) 3D view of the area outlined in (a) showing the fault planes corresponding to the flower structures shown in Fig. 5.4.

5.2 Shear zone thickness

The shear zone covers an extensive area of the Nankai accretionary complex ($\sim 100 \times \sim 25 \text{ km}^2$; Fig. 5.6a) and ranges up to $\sim 2 \text{ km}$ in thickness. To evaluate the special distribution of the thickness, I divided this area into eastern, central, and western regions (see Fig. 5.6a). It is thickest in the outer wedge in the central part of the study area. In the eastern and western regions, the shear zones reach a maximum thickness of $\sim 1200 \text{ m}$ to $\sim 1600 \text{ m}$, and their minimum thickness is less than $\sim 200 \text{ m}$. They become thinner toward their seaward edge,

decreasing from ~1200 m to ~600 m in the western region and from ~1600 m to ~200 m in the eastern region (Fig. 5.6a).

In the central region near SF-A and SF-B, the shear zone varies greatly in thickness toward the landward side (~300 m to ~2000 m) (Fig. 5.3). In particular, the top of the shear zone is deformed by offsets on SF-A and the imbricate thrust faults (Figs. 5.3, 5.4). Moreover, the shear zone has been apparently thickened by interactions of the strike-slip fault system with reverse faults (Figs. 5.2, 5.3). Here I should note that the seafloor traces of strike-slip faults (SF-A and SF-B) and thick shear zone are not consistent on the thickness distribution (top view; Fig. 5.6a), because of the dip of strike-slip faults (Fig. 5.4). The shear zone is very thick (~2000 m) near the outer ridge and thins in the seaward direction as its bounding décollement rises (Tsuji et al., 2014b) to a depth of about 4–5 km beneath the seafloor (Fig. 5.3). Thus, the presence or absence of an imbricate thrusts and strike-slip faults might result in local thinning or thickening of the shear zone, respectively. Additionally, roughness in the décollement may result in sediment being scraped off the subducting plate to form an intercalated layer at the base of the upper plate (Tsuji et al., 2009; Park et al., 2010; Kitajima and Saffer, 2012). This underthrust sediment might also account for the presence of the shear zone.

The thick shear zone extends more than ~80 km along the trench direction (Fig. 5.6b), and its edge appears to have thinned to less than 200 m in the eastern and western regions. The edge of thick shear zone may result from the strike-slip faults related to the tectonic collisions, namely interaction with a static tectonic backstop in the western region (Tsuji et al., 2015). The effects of tectonic collisions may account for the large-scale structure of the shear zone, including the NW-SE trending strike-slip faults (Fig. 5.6). In the eastern region, subduction of the paleo-Zenisu ridge (gray shading in Figs. 5.1a, 5.6a; Fig. 5.2d; Park et al., 2003) could influence the shear zone thickness; The western edge of the paleo-Zenisu ridge may define a boundary of thick shear zone (black dashed line in the eastern region in Fig. 5.6).

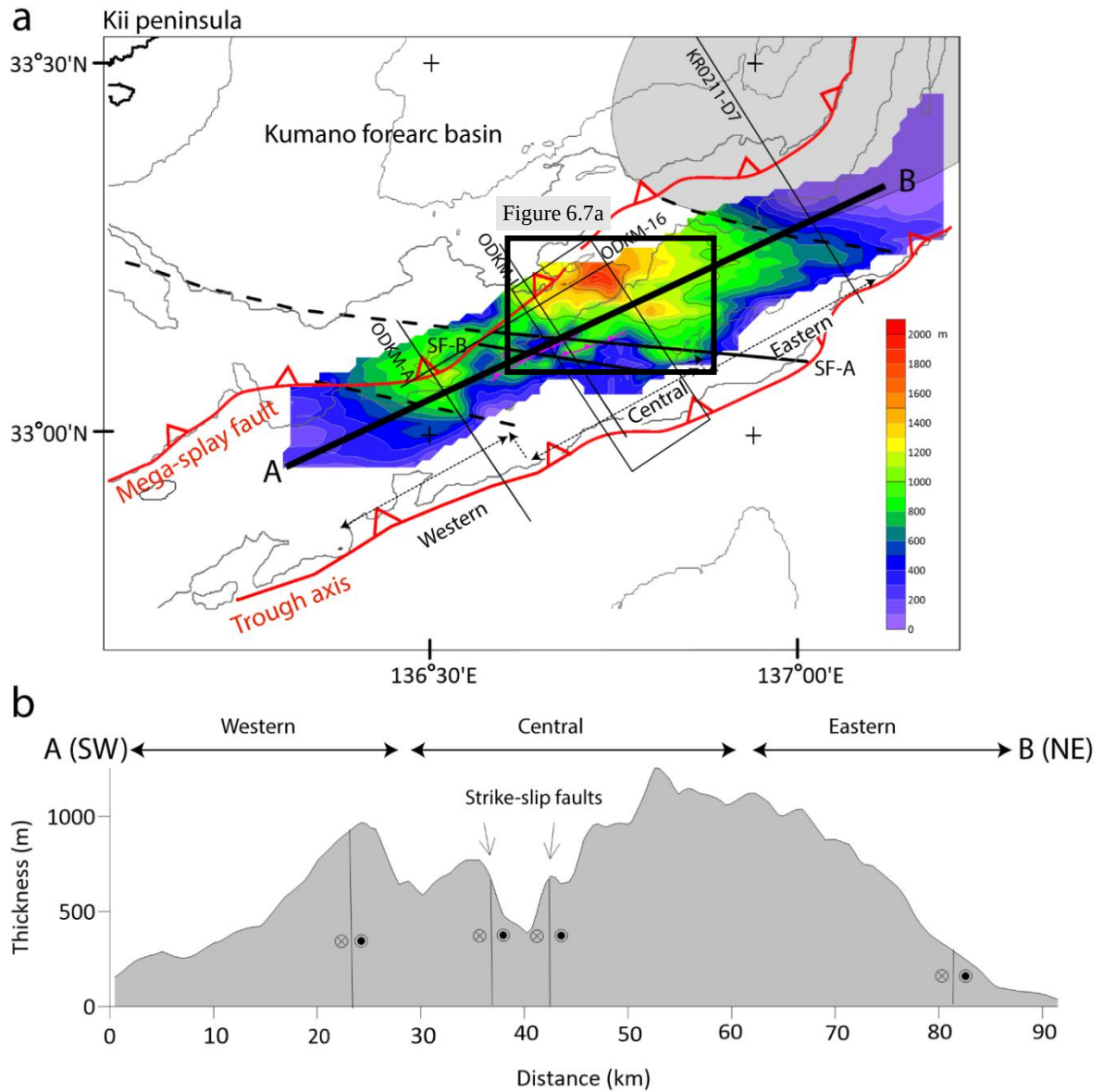


Figure 5.6. (a) Map of shear zone thickness, locations of profiles and strike-slip faults. Profiles ODKM-A, ODKM-16, ODKM-I, and KR0211-D7 are 2D seismic lines, the area of 3D seismic reflection data is outlined in black, reverse faults are shown in red, and the paleo-Zenisu ridge (Park et al., 2003) is indicated by gray shading. (b) The thickness of the shear zone and strike-slip fault locations along the profile A–B in panel (a).

6. Discussion

6.1 Tremor distribution controlled by shear zone thickness

I found that the most tremor sources were within the shear zone around the plate interface (Fig. 6.1a; Hendriyana and Tsuji, 2021), where near-lithostatic pore pressure has been inferred (Fig. 5.1d; Tsuji et al., 2014b). However, Hendriyana and Tsuji, (2021) estimated the uncertainty of tremor depths ~ 4 km, and the depth uncertainty could be caused by the less differential travel times and the less accurate P-wave, S-wave velocity and density model for source estimation. These tremors episodes are associated with the known slow slip events or slow earthquakes (Annoura et al., 2017; Araki et al., 2017; Nakano et al., 2018), and the previous studies intensively discussed slow slip events that accompany very low-frequency earthquakes in the shallow plate interface. For example, Araki et al. (2017) demonstrated that SSE events crept up at 1-4 cm in this study area (equivalent to $M_w1 - M_w5$). Therefore, the tremor I used in this study could have occurred mainly around the plate interface (i.e., shear zone). Furthermore, most tremor events are shallower than 10 km (black dots in Fig. 6.1a), indicating that their occurrence above or close to the plate interface.

Tremor episodes from October 2015 to March 2016 (Fig. 6.1b) were distributed over the whole study area, but they were concentrated in the thicker parts of the shear zone and in the central region near SF-A and SF-B. Another cluster of events was centered around SF-A in the landward inner wedge. This period has the normal distribution of tremors with total number around 446 (Fig. 6.2). During the 2–15 April episodes (Fig. 6.1c), tremor occurred throughout the central region, especially in the thick part of the shear zone with total number around 958 (Fig. 6.2). Few tremor episodes occurred west side of SF-A and SF-B. During the remaining monitoring term (16 April to 31 December 2016; Fig. 6.1d), tremor occurred in the eastern

region where the thickness of the shear zone is thinner. However, at the eastern edge, the tremor distribution seems to align for the subduction direction (i.e., edge of the paleo-Zenisu ridge or strike-slip fault orientation). These source distributions appear to be strongly related to the shear zone thickness as well as the presence of a strike-slip fault.

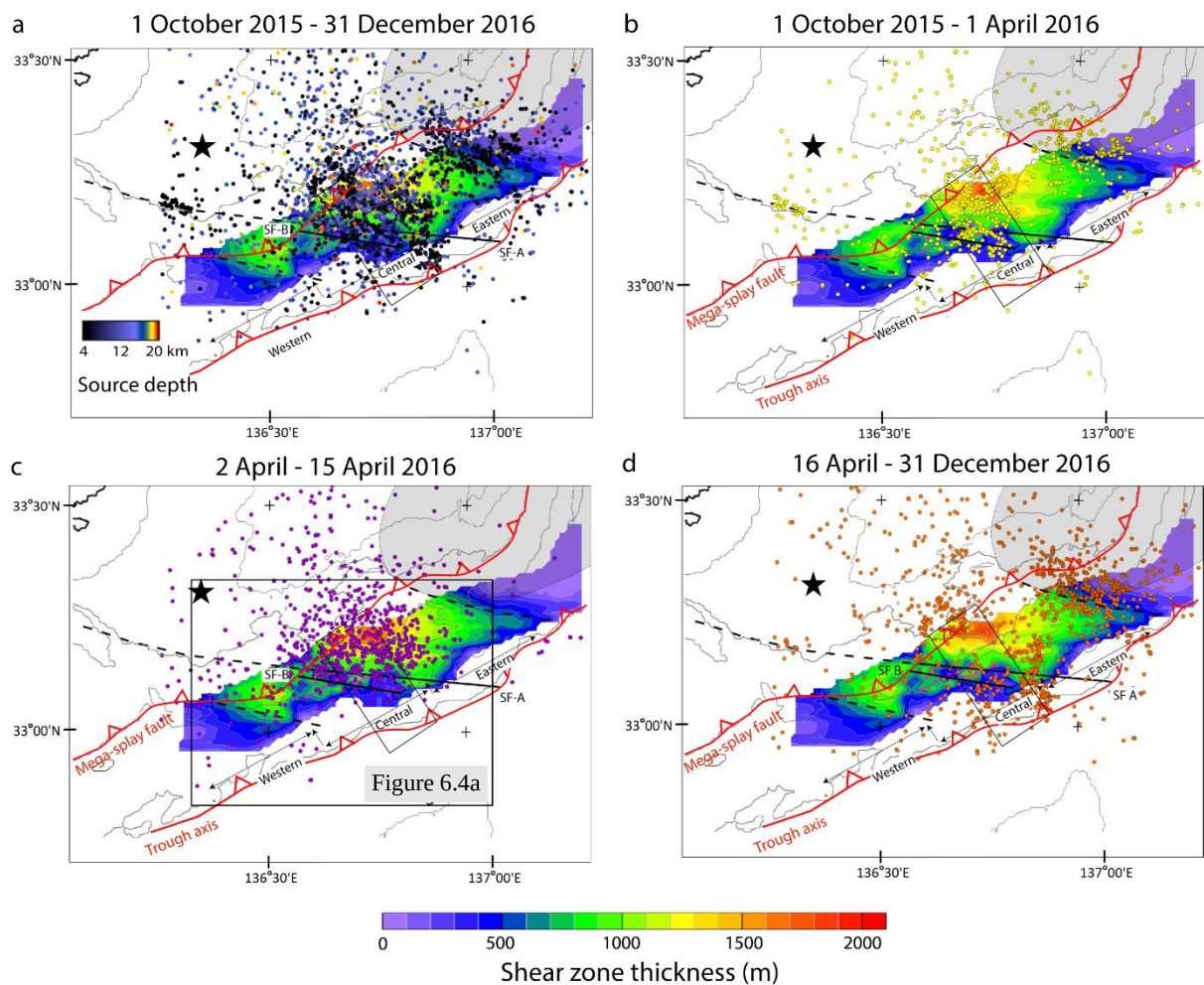


Figure 6.1. Distribution of tremor (dots) during selected time periods superimposed on the shear zone thickness map. The paleo-Zenisu ridge (Park et al., 2003) is shown by gray shading. (a) Results from October 2015 to 31 December 2016, with source depth indicated by color. (b) Results from 1 October 2015 to 1 April 2016 (October–March episode). (c) Results from 2 to 15 April 2016 (April episode). The epicenter of the 1 April 2016 off-Mie earthquake is shown (black star). (d) Results from 16 April to 31 December 2016 (April–December episode).

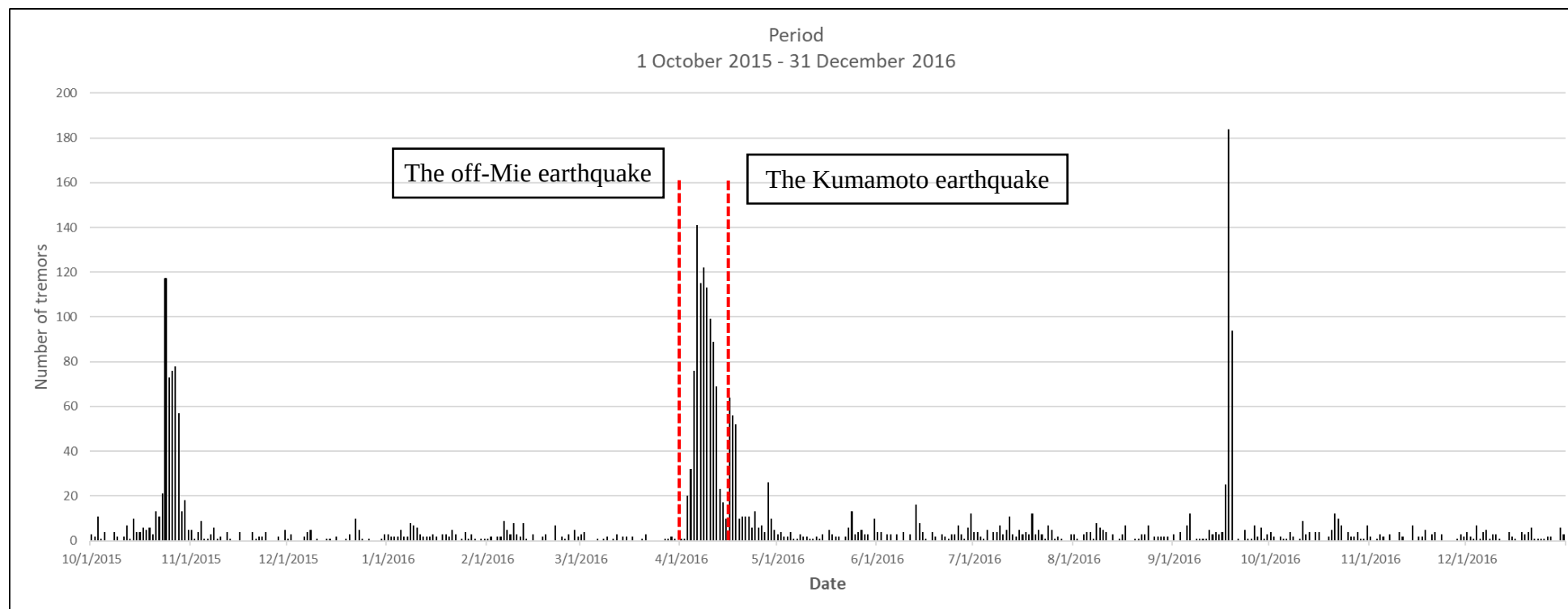


Figure 6.2. Histogram of tremor from 1 October 2015 to 31 December 2016.

Here I focus on the episode of 2-15 April 2016 (Figs. 6.1c, 6.4), because the tremors occurred as a single sequence (short term). Because the most tremor locations were on the eastern side of the SF-B during the April episode, the tremor could not propagate across the SF-B (Fig. 6.4a). Furthermore, some tremors occur along the strike-slip faults. Therefore, the tremors could be related to the geometry of the strike-slip fault or stress/pore pressure difference across the strike-slip faults. The strain partitioning due to strike-slip displacement could affect the stress state across the fault. In addition, since the intensive tremors occur in the thick and shallow shear zone associated with the strike-slip faults (Fig. 6.4), the fluid upward migration and its accumulation may relate to the tremor occurrence around the strike-slip faults (Figure 6.3).

Furthermore, the tremor during the 2–15 April shifted to the central area (Fig. 6.4a), where SF-A and the outer ridge are located (Fig. 5.1b). These areas have high pore fluid pressures (Fig. 5.1d) owing to fluid release from clay dehydration at temperatures from $\sim 80^{\circ}\text{C}$ to $\sim 250^{\circ}\text{C}$ (e.g., Saffer and Wallace, 2015) and sediment compaction due to overburden (Tsuji et al., 2014b). The presence of clay minerals in the oceanic crust below the shear zone, Ikari et al., (2020) suggested that altered basalt may allow the possibility the tectonic tremors at $\sim 4\text{--}7$ km depth, and the occurrence of tremors relates to the fault friction (Ujiie and Kimura, 2014) in the plate interface. Despite the fact that friction data indicates velocity-strengthening behavior, it is still possible for tremors to initiate in the altered Nankai basalt during stable slip. In particular, the overburden may increase substantially at the landward of the outer ridge (i.e., inner wedge), because the higher density of the inner wedge overlays the plate interface (Fig. 6.9). Furthermore, the underthrust sequence in the inner wedge is thinner than the shear zone in the outer ridge and transition zone (Figs. 5.1, 6.9). In sum, large amounts of fluid might be expelled from sediments at the WBSF or outer ridge, causing increasing pore pressure there

(Fig. 6.9). Indeed, cold seeps were observed at the seafloor around the outer ridge (Figs. 5.1b, 6.9).

Because the seismogenic faults around the shear zone could be close to a critical state (Tsuji et al., 2014b), minor increases in pore pressure might trigger slow slip events. The external forces that influence pore pressure, such as sea level change, may modulate underground pore pressure change and increase the diffusion of pore pressure changes (Andajani et al., 2020). Their role in slow earthquake occurrence should be investigated further.

Tectonic tremor may contribute to the accumulation of stress around its source area and possibly serve as a trigger to interplate earthquakes (Ito et al., 2013; Tsuji et al., 2017). For instance, during the first period of tremor episodes, tremor was concentrated for a period of 8 days near the thick shear zone in the central region, which is influenced by the reverse faults and the oblique strike-slip faults SF-A and SF-B (Figs. 6.1b, 6.2; Moore et al., 2009). This cluster occurred approximately five months before the off-Mie earthquake of 1 April 2016. Moreover, the off-Mie earthquake (~30 km from the outer ridge; black star in Fig. 6.1) and the more distant 16 April 2016 Kumamoto earthquake (Fig. 5.1a) may have influenced the occurrence and migration of tremor in subsequent days (Figs. 6.1c, 6.1d, 6.9; Annoura et al., 2017; Hendriyana and Tsuji, 2021). Tremors likely tended to migrate along the strike and updip direction of shear zone (Figs. 5.6; 6.9). Therefore, monitoring tremor episodes and mapping their source mechanisms appears to be essential for better understanding of large interplate earthquakes.

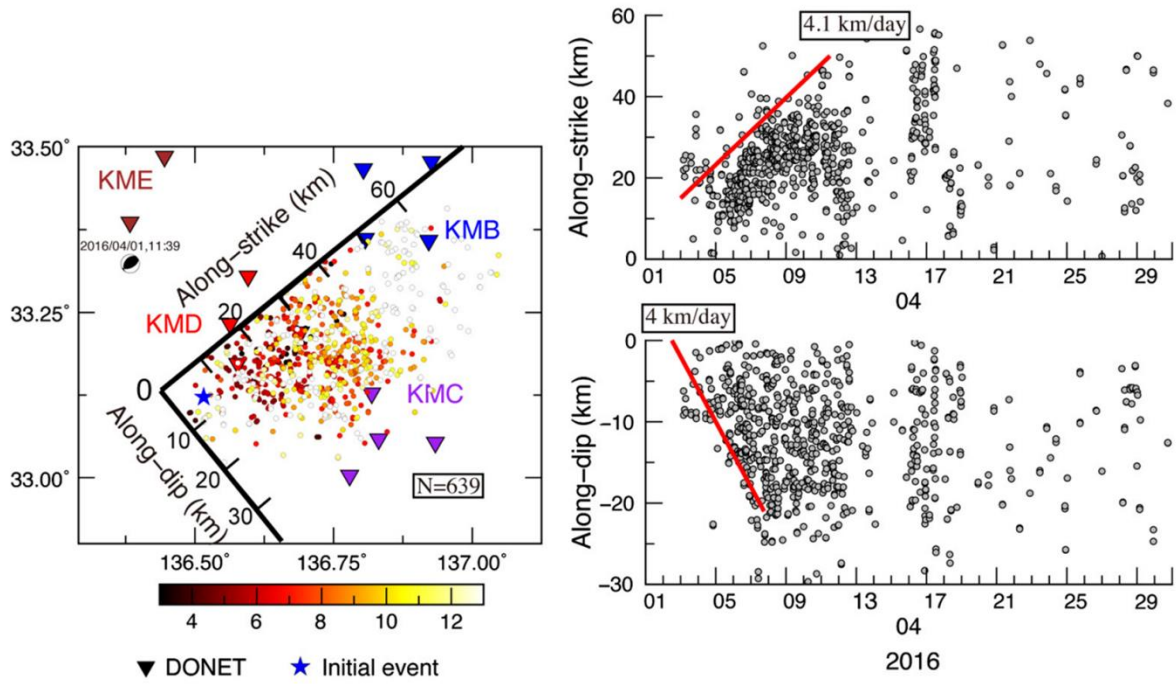


Figure 6.3. Plots showing migration of tremors along dip (up-dip) and along strike during the April 2016 period (2 April – 15 April 2016) (Hendriyana and Tsuji, 2021).

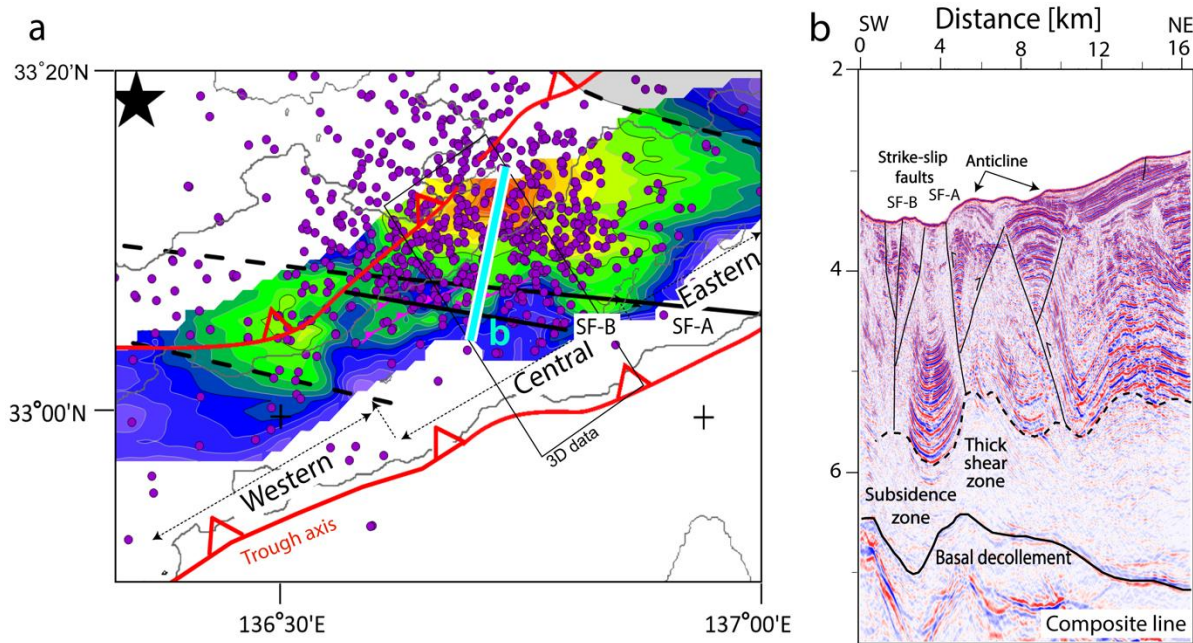
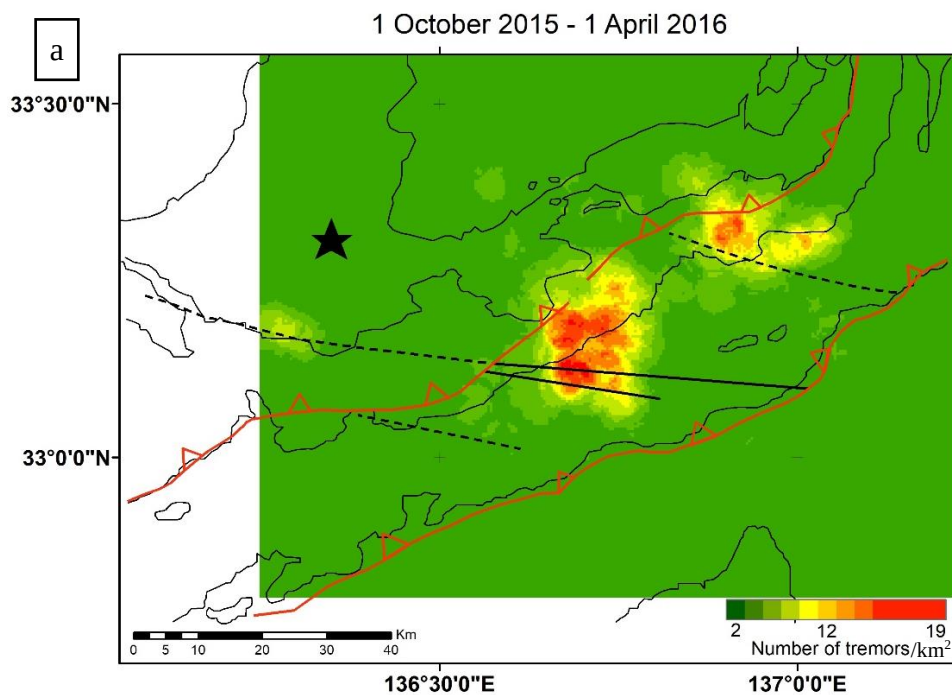


Figure 6.4. (a). Enlarge relationship between tremors during 2 to 15 April 2016 (April episode) and shear zone thickness. (b) Interpreted reflection profiles derived from 3D seismic data (blue line in panel a). The profile perpendicular to the strike-slip faults demonstrates that the strike-slip faults induce thicker shear zone.

6.2 The relationship between the tremor number and the shear zone thickness

I computed a magnitude of the tectonic tremor per unit area (km^2) by analyzing point features located within the vicinity of each cell. The data presented in Figure 6.1, depicted spatially, underwent processing and reclassification to generate a tremor density map as illustrated in Figures 6.5a, 6.5b, 6.5c. These figures indicate that red areas have the high number of tremors. A consistently high number of tremors occurs around thick shear zone (Figure 5.6a) in episodes from October 2015 to March 2016 and episodes of the 2–15 April (Figures 6.5a, 6.5b). However, in episode from 16 April to 31 December 2016, the high number of tremors located randomly in three areas (See in Figure 6.5c).



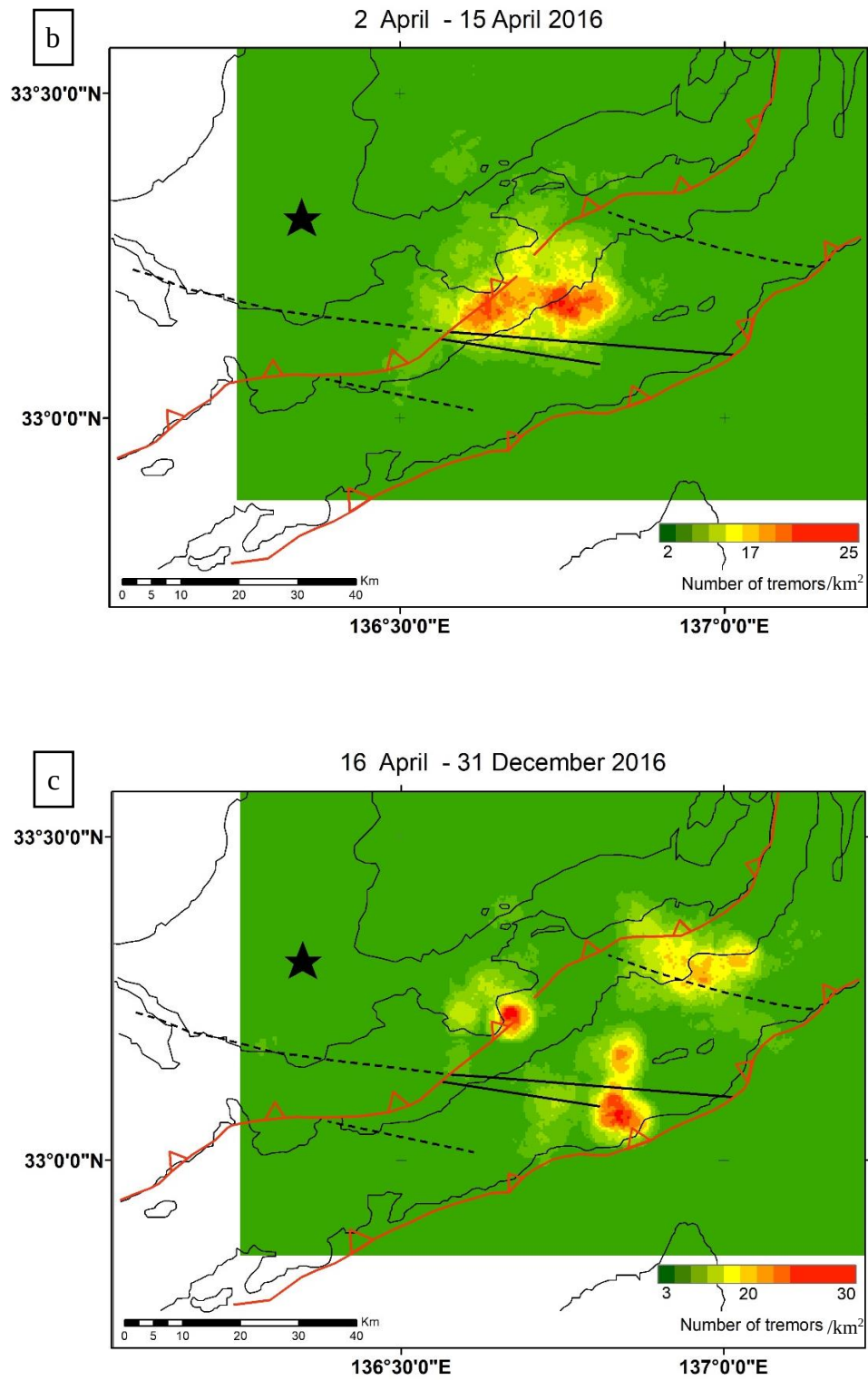
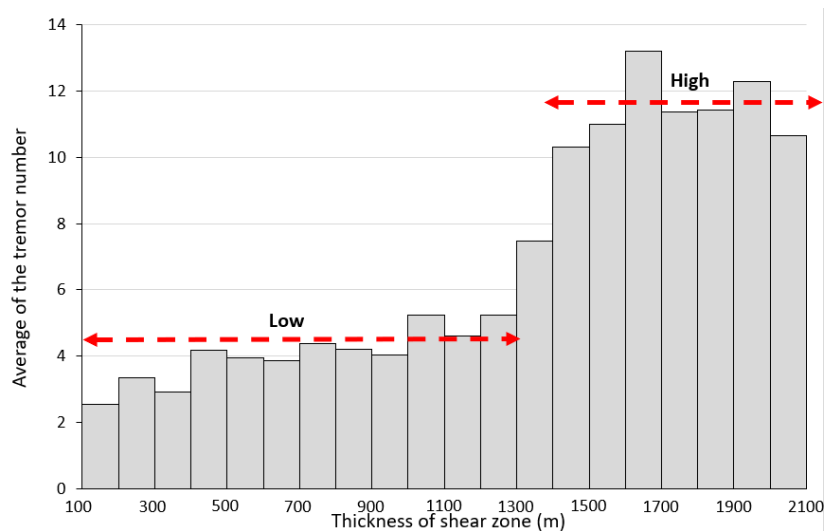


Figure 6.5. Map of tremor density from 1 October 2015 to 31 December 2016. (a) Results from 1 October 2015 to 1 April 2016 (October–March episode). (b) Results from 2 to 15 April 2016 (April episode). (c) Results from 16 April to 31 December 2016 (April–December episode).

To reveal the correlation between the thickness of shear zone and the tremor events, I computed the mean tremor count across various ranges of shear zone thickness as given in Tables (Refer to Tables 2, 3, 4 and the method description to obtain the correlation in Appendix A). The graph of the relationship between the average of the tremor number and shear zone thickness for each period is described in Figure 6.6. An interval thickness of shear zone thickness from 1400 m to 2100 m experienced a high number of tremors before the off-Mie earthquake (Figure 6.6a). In addition, the graphs have the same pattern of tremor numbers after the off-Mie of earthquake and the Kumamoto earthquake (i.e., higher and lower tremor numbers; see in Figs. 6.6b and 6.6c). The number of tremors is higher, correlating with the thickness of the shear zone from 1300 m to 1700 m, while a lower number of tremors occur in the shear zone thickness range from 1700 m to 2100 m.

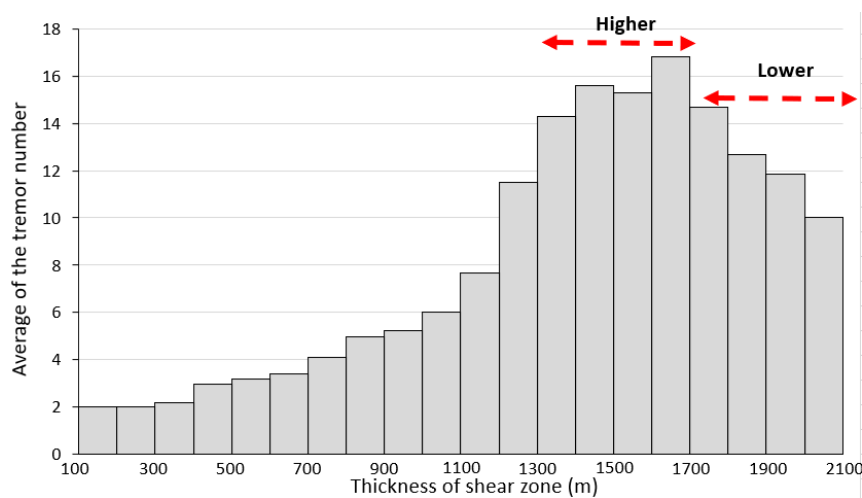
a

1 October 2015 – 1 April 2016



b

2 April – 15 April 2016



c

16 April – 31 December 2016

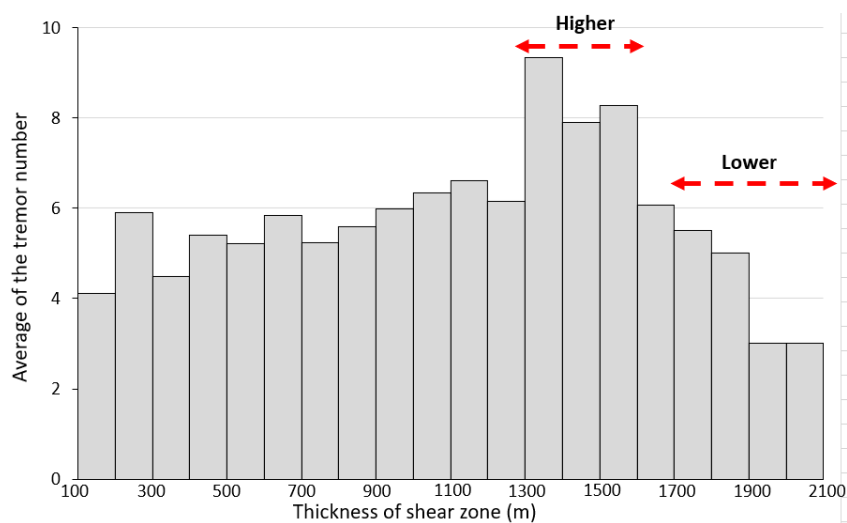


Figure 6.6. Graph of the relationship between average of the tremor number and shear zone thickness (m). (a) Results from 1 October 2015 to 1 April 2016 (October–March episode). (b) Results from 2 to 15 April 2016 (April episode). (c) Results from 16 April to 31 December 2016 (April–December episode).

The shear zone is divided into three areas according to thickness (Figs. 6.7a and b). The mid area (orange color zone) has a shear zone thickness ranging from 1300 m to 1700 m. Bedding or foliation within the shear zone (See in Figs. 6.7c and d) may undergo the activation of two deformation mechanisms (compaction and opening; see deformation mechanisms in Maillot and Leroy, 2006). This activation could be controlled by a rotational motion (vertical and lateral displacement) due to tectonic activities. Furthermore, these activities result in the higher stress (Figures 6.7a and b), where the accumulation of the higher number of tremors occurs (Figures 6.6b, 6.6c). On the other hand, some inactive parts are present in areas with a shear zone thickness from 1700 m to 2100 m (red color zone in Figs. 6.7a and b). A lower tremor numbers (Figures 6.6b and 6.6c) of the red zone (Figure 6.7a) could be mentioned as inactive parts of shear zone. An inactive part may be controlled by the material in the shear zone only be displaced vertically due to the tectonic forces.

The two phenomena (higher and lower tremor numbers) of tectonic tremor after earthquakes (i.e., the off-Mie earthquake and the Kumamoto earthquake) mentioned above was influenced by the geometry of the shear zone. Figure 6.7b illustrates that the top of shear zone has an anticline (red color zone; Fig. 6.7c) and limb shape (orange color zone; Fig. 6.7c). The number of tremors is more significant in the limb area of the shear zone compared to those of the anticline area (Figure 6.7b). Maillot and Leroy (2006) illustrated a kink fold model, where high stress is generalized to folding in the presence of limbs of an anticline, i.e., rotated bedding between faults (Figure 6.8). The mid area with high stress can be interpreted as the initiation of tremors, followed by the migration of tremors here, as indicated by the high tremor number (Figs. 6.6b, 6.7a-b). Hendriyana and Tsuji, (2021) suggested that initial tremors occur close to the WBSF and strike-slip fault (see in Figure 6.7a; see the initial tremor in Fig. 6.3) several days after the off-Mie earthquake. Then, tremors migrated both up-dip and parallel to the trough.

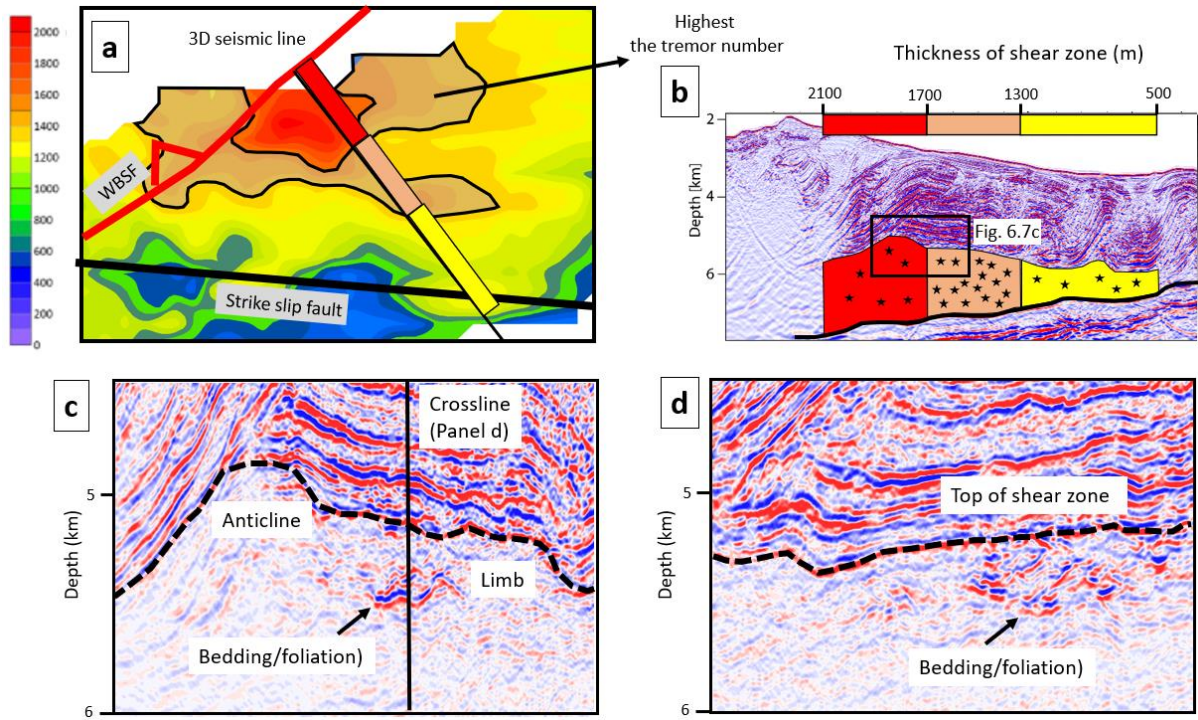


Figure 6.7. (a). Enlarge of shear zone thickness map divided into three areas due to the thickness. (b) Showing the tremor numbers in the interval thickness of shear zone in the reflection profiles (coloured blocks in panel a). Black star is the tremor. Black box is panel c. (c) (d) Enlarge of 3D seismic profile of inline (c) and crossline (d) showing bedding or foliation into the shear zone. Panel c and d derived in Appendix A in Figures A7, A8 and A9.

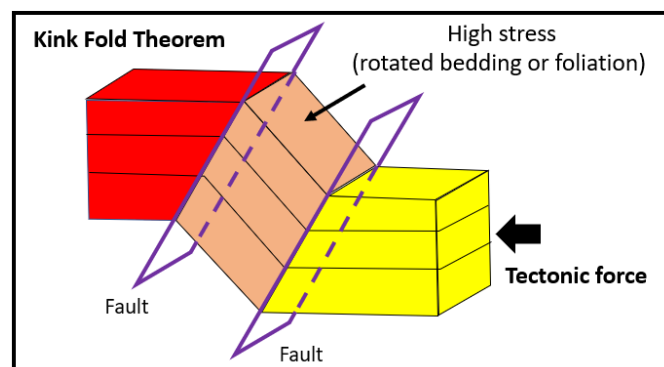


Figure 6.8. The Kink fold model from Maillot and Leroy, (2006) showing a high stress in the limb.

6.3 Tectonic structures controlling shear zone distribution

The continuous compression between the obliquely subducting Philippine Sea plate and the tectonic backstop of the Kii Peninsula influences the evolution of the accretionary prism in the Nankai Trough. The presence of undeformed forearc basin sediment overlying the accretionary wedge near the east coast of the Kii Peninsula suggests that the accretionary prism formerly served as a tectonic backstop (Tsuji et al., 2015). The collision at ~6 Ma between the tectonic backstop and the younger accretionary prism (Fig. 5.1) and the resumption of subduction since ~2 Ma have produced a set of structural features oriented NW-SE on the inner wedge southeast of the Kii Peninsula (Fig. 5.1a) (Tsuji et al., 2013). One of these may have propagated northwest and connected with SF-A (black line in Figs. 5.1a, 5.6a). Thus, the collision of an accretionary prism with the tectonic backstop (i.e., strain partitioning) appears to have initiated the oblique strike-slip faults (Azevedo et al., 2018).

The strike-slip faults and shear zone could be further influenced by the geometry of basal decollement or oceanic crust surface (Azevedo et al., 2018). The decollement and oceanic crust surface are strongly deformed by the thrusts within the oceanic crust in this region (Tsuji et al., 2009; Shiraishi et al., 2020). The thrusts within the oceanic crust seem to be also related to the tectonic backstop (Tsuji et al., 2013). Therefore, the backstop off the Kii peninsula induces strike-slip faults and largely influences shear zone thickness. Indeed, Shiraishi et al. (2020) found the relationship between the slow slip and roughness of oceanic crust surface.

The strike-slip faults caused variation of the shear zone thickness with an order of magnitude from 400 m to 1200 m resulting in higher strain accumulation at the fault that generates the pop-up structures with folds as well as anticline structure at the roof of the shear zone (Figs. 5.4, 5.6, 6.9). The fold structure associated with the strike-slip fault (pop-up structure or anticlines in Figs. 5.4, 6.9) is more dominant compared with the fold associated with imbricated thrusts. Because the sources and migrations of tremor appear to favor the thick

parts of the shear zone and the oblique strike-slip faults, I suggest that the spatial variation of tremor are controlled by oblique strike-slip faults that formed during the collision with the tectonic backstop in the Nankai Trough off the Kii peninsula.

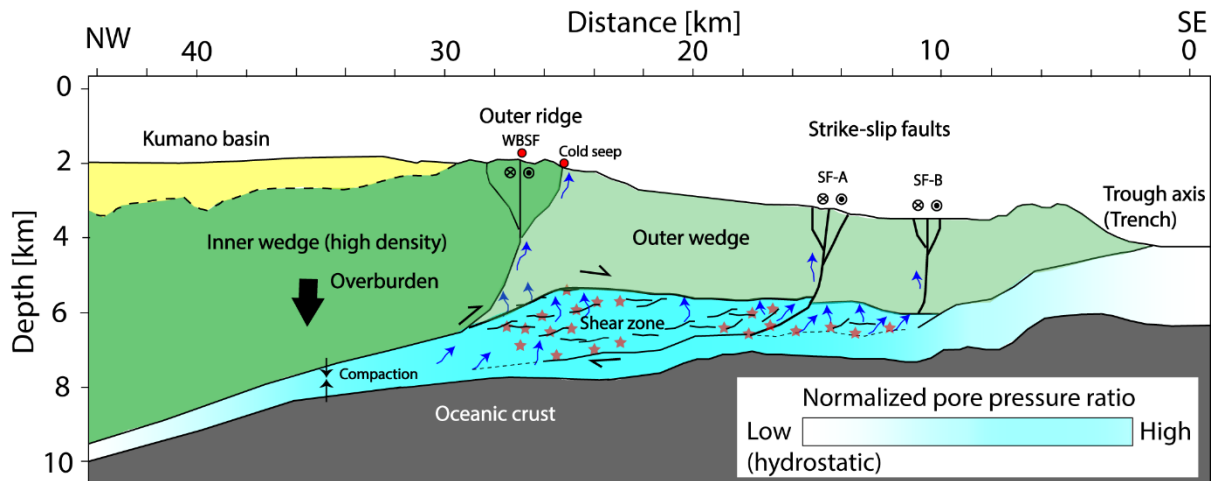


Figure 6.9. Schematic image of the relationship between shear zone, pore pressure, and tremor distribution (stars). Black arrows show the fluid migration direction.

7. Summary and Conclusion

7.1. Research findings

I investigated the geological structures that controls slip behaviors in the convergent plate margin in the Nankai Trough. This work is summarized as follows:

Research in the Nankai Trough, I applied three steps to investigate the influence of shear zone thickness and strike-slip faulting on tectonic tremor in the Nankai Trough, southwest Japan. From step one, I obtained the processing seismic data and calculate seismic attributes. The outcome of step one revealed the seismic data that can interpret the structures and determine the shear zones. The possibility of tremors aligning on the oblique strike-slip faults was further investigated in step two by mapping fault distribution and comparing a single sequence (short term) of tectonic tremors that resulted in stress accumulation. From step two, I obtained that the oblique strike-slip faults control the thickness of the shear zone and tremor distributions. Then, step three obtained the mapping shear zone thickness and the seismic facies and the physical characteristics of the shear zone. The outcome of step three revealed the thick part of the shear zone, where high pore fluid pressure may weaken faults and thereby facilitate slow earthquakes. The varied shear zone is controlled by the oblique strike-slip faults that result from the collision of an accretionary prism and the tectonic backstop on landward.

Based on the summary of sections, the key finding of this study is summarized as follows:

1. Tremor episodes were localized in oblique strike-slip fault zones and thick sections of the shear zone, where these areas can be interpreted as the areas with a high pore pressure.
2. There is a same pattern of tremor numbers in the shear zone thickness in both events (i.e, after the off-Mie earthquake and after the Kumamoto earthquake). The number of tremors is higher, correlating with the thickness of the shear zone from 1300 m to 1700 m, while a lower number of tremors occur in the shear zone thickness range from 1700 m to 2100 m.
3. The oblique strike-slip faults and the shear zone thickness with the tremor number may be controlled by the collision of an accretionary prism with a tectonic backstop.
4. Slow slip occurrences near the plate interface might be made possible by the physical diversity of the shear zone (i.e., the shear zone is divided into three areas according to thickness) and the prominent geological features (i.e., strike slip faults) in the shallow plate interface of Nankai Trough, southeast Japan.

7.2. Research implication

The proposed method in this study is useful for evaluating slip behaviors controlled by oblique strike-slip faults and imbricate thrust faults in the outer wedge of the Nankai Trough. The spatial-temporal interpretation method for monitoring tectonic tremors hosted in thick parts of the shear zone appears essential for better understanding the fast earthquakes. Because minor increases in pore pressure could trigger slow slip events, the study is needed to monitor the long-term and short-term crustal deformation of slow slip events. With the increasing interest in the slow slip studies for predicting the huge earthquakes in the future, this approach

can be applied to monitor and evaluate the area prone to failures due to expelled fluid from the shear zone.

7.3. Recommendation for future work

The research in this dissertation opens opportunities to understand the geological structures controlling slip behaviors in the convergent plate margin. The following are recommendations for future research on the topic of this dissertation:

The discussion in sections emphasizes the influence of shear zone thickness and strike-slip faulting changes in the spatial-temporal of the tectonic tremor distribution because of the fluctuation of stress and pore pressure. Nevertheless, the current approach of the pore pressure studies in the shear zone is still applied only to the two-dimensional (2D) parameters in the Nankai Trough region. A better interpretation of slip behaviors in the convergent plate margin results in the shear zone thickness within the 2D model of pore pressure in each profile. For further validation, the result derived from this approach should be compared with the slow earthquake data over a long period. Additionally, the current approach is limited to the case where shear zones with strike-slip faults can be assumed to be a 2-dimensional model of pore pressure. In reality, shear zones with strike-slip faults are more complex as they could vary with depth and dimension. Hence, it is recommended for the next following study consider the hydraulic diffusion of fluid on the shear zone.

8. References

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Acknowledgements

This thesis was produced to fulfil the requirements of the doctoral program at the Earth Resources Engineering Department of the Graduate School of Engineering. In this acknowledgment, I sincerely thank everyone who has encouraged and inspired me to finish this dissertation. First and foremost, I express my sincere gratitude to Prof. Takeshi Tsuji (Tokyo University), who served as my supervisor and who provided me with essential guidance throughout my research. I am grateful to him for all the possibilities I received while pursuing my PhD. I also want to express my profound appreciation to Associate Prof. Tatsunori Ikeda (Kyushu University) for his guidance and unwavering emotional support. I further also like to extend my heartfelt gratitude to Dr. Channaly Chhun (Kyushu University) for her constant support, fruitful discussion, and helpful criticism. Additionally, I would like to express my profound gratitude and great appreciation to Prof. Akira Imai (Kyushu University) and Prof. Yamada Yasuhiro (Kyushu University) for their time and feedback to improving this dissertation. Their vast expertise and wealth of experience have continuously motivated me in both my daily life and academic studies.

I gratefully acknowledge the support for my 4-year Ph.D. studies at Kyushu University provided by the Grants-in-Aid for Scientific Research (KAKENHI) from the Japan Society for the Promotion of Science (JSPS) (No. JP20H01997 and JP21H05202). These studies were also funded by the Diponegoro University scholarship (Indonesia).

For this study, I would like to thank the Japan Agency for Marine-Earth Science and Technology (JAMSTEC) and for providing the 3D and 2D seismic reflection data in Nankai Trough, Southwest Japan. I appreciate Andri Hendriyana (Kyushu University) for the discussions and Fernando Lawrens Hutapea (Kyushu University) for his technical support.

I am thankful to the research support staff Ms. Tomomi Yamaguchi for her support during my Ph.D. time. I am also thankful to the research support staff of the Engineering Faculty and the Department of Earth Resources Engineering, Ms. Yasura Oiwa, Ms. Noda, and Ms. Ito. I thank Ms. Maki Jagawa for providing me constructive feedback on my writing. My immense appreciation to my Japanese colleagues for their support during my stay at Kyushu University ever since I was a doctoral student. Special thanks to all Doctoral students (Bakhtiyar Daulet, Ahmad Bahaa, Tarek Samir A.A, Nthaba Bokani) for good talks and a stimulating discussion with me. They have known me since I was a doctoral student. I extend my special thanks to my close friends at Tokyo University, Dr. Rezkia Dewi Andajani, for consistently giving me suggestions to improving my study and mind.

Last but not least, I would like to thank my family for their support and encouragement during my academic career. I could not have made it this far without their great support and understanding over the previous few years. I would especially like to thank my wife, Dyah Ika Rinawati (student in Kyushu University), for her insightful advice and persistent encouragement to continue my study. Also, all my children (Farid Ahmad A, Haniya Tazkiya A, Muhammad Ziyad A, and Kaisya Tsamara A) for being happy and cute and accompanied me during the Ph.D program.

Fukuoka, January 2024

Fahrudin

Appendix A

Steps to calculate the average of the tremor number for an interval of the shear zone thickness

In the initial step, I calculated a magnitude (number of tremors) per unit area (square kilometer) from point features that fall within a neighborhood around each cell. I used the Arc-Gis Program to get a tremor density map utilizing a menu of point density in the spatial analysis tools. Data (location of tremor events in Figure A1) as a point in the spatial was processed and reclassified to get a tremor density map as shown in Figure A2. Table 1 shows the reclassification of an interval of the tremor number (raster data) to obtain a new value (raster data with a layer). The new value is the maximum of the tremor number in each interval, for example, a range of value 2 - 4 was reclassified into value 4.

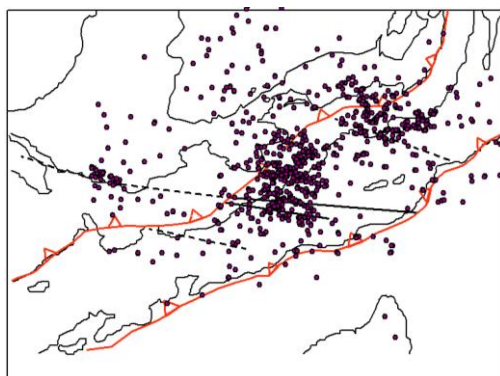


Figure A1. Distribution of tremor (dots) during selected time periods (1 October 2015 to 1 April 2016)

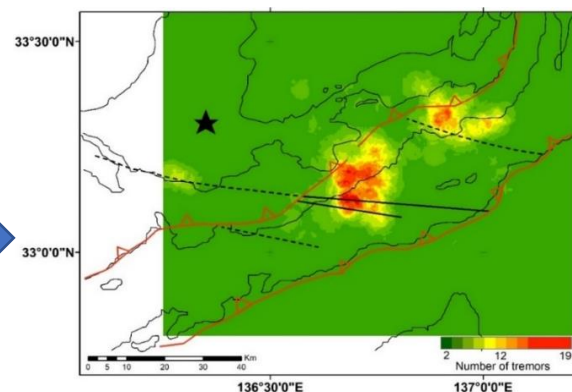


Figure A2. Map of tremor density (1 October 2015 to 1 April 2016)

Table 1. The reclassification results

an interval of the tremor number (raster data)	maximum of the tremor number (raster data with a layer)
0 - 2	2
2 - 4	4
4 - 6	6
6 - 8	8
10 - 12	12
12 - 15	15
15 - 17	17
17 - 19	19

Furthermore, in order to get the tremor number from the reclassified value of density map, the Fishnet method was applied. The Fishnet method is a feature class to contain a net of rectangular cells, which has a point in the center of cells (Figure A3). Based on the map of shear zone thickness (as shown in Figure 5.6a in chapter 5), the grid was created by Fishnet method (Figures A4). Then, map of tremor density (Figure A2) and map of shear zone thickness (Figure 5.6a in chapter 5) were extracted from multi value to points using the grid (Figure A4).

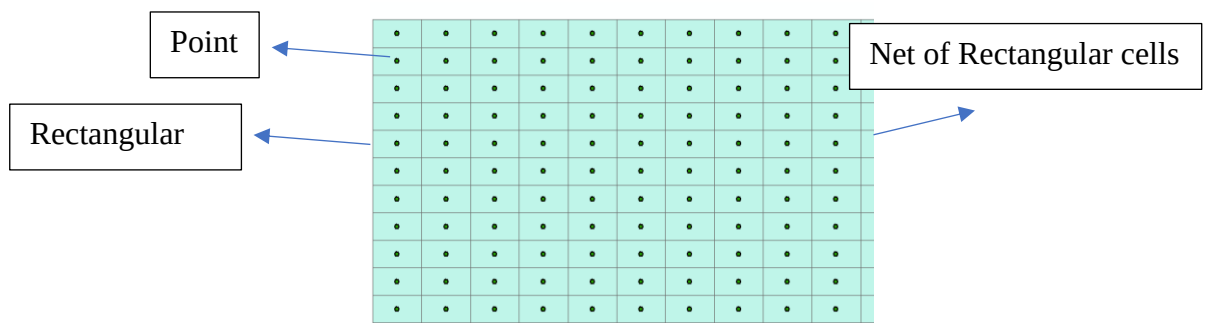


Figure A3. Illustration of Fishnet method

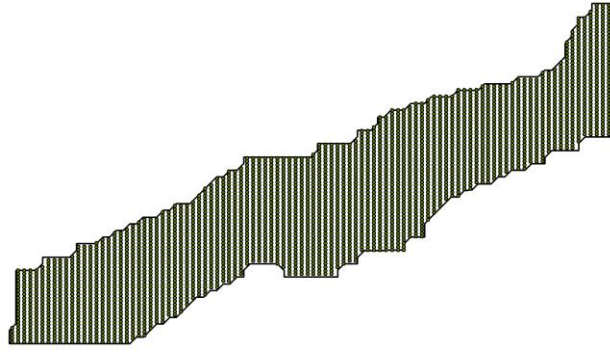


Figure A4. A grid created by Fishnet with the shear zone boundary

The results of this extraction process are both maps, newly map of the shear zone thickness and the tremor number (See in Figures A5a and A6a). To know the value of the central point of the grid, both maps (newly map of the shear zone thickness and the tremor number) were enlarged, as described in Figures A5c and A5c. Figure A5c shows that the values of the central point at uniform intervals of shear zone thickness are the same (Figure A5a); for example, the area with an interval of shear zone thickness 600 – 700 m has a similar central point value i.e., 700 m (See in Figure A5c). A similar step was applied to the tremor number map as illustrated in Figure A6c. For instance, the area with an interval of tremor number 6 – 8 has a central point value of 8.

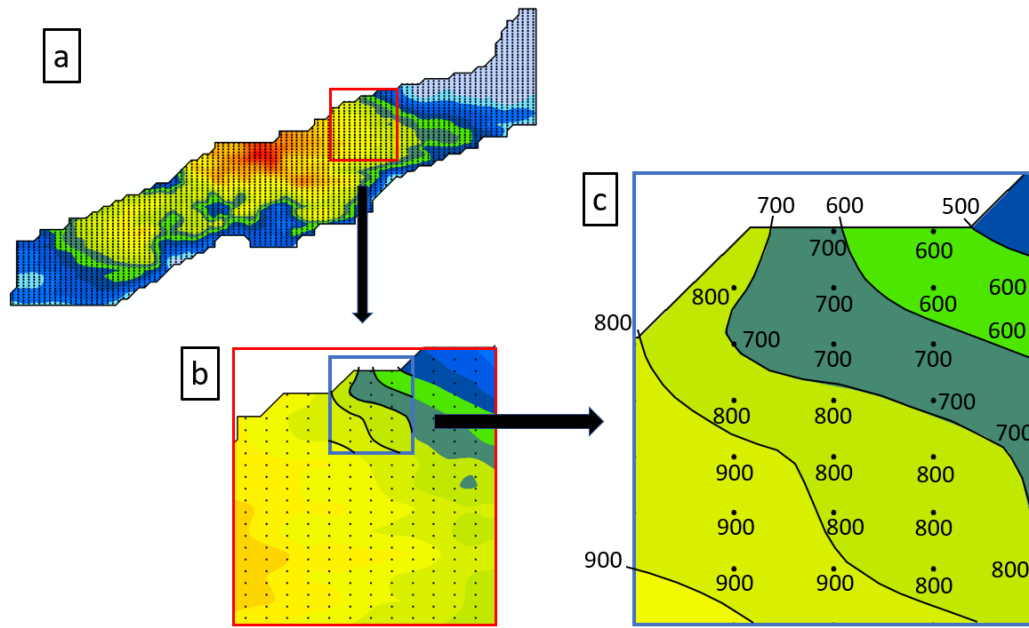


Figure A5. Newly map of the shear zone thickness with the grid (a) and enlarging areas (b and c) showing values point at different intervals of the shear zone thickness.

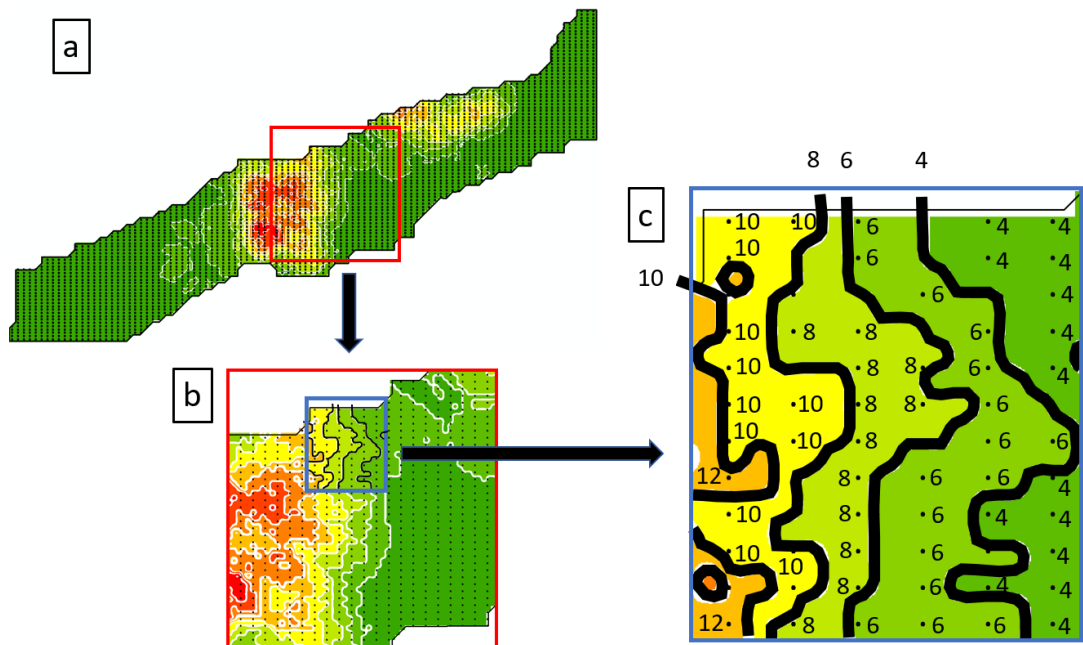


Figure A6. Newly map of the tremor number with the grid (a) and enlarging areas (b and c) showing values point at different intervals of the tremor number.

To create a graph of the relationship between shear zone thickness and the number of tremors, I calculated the average of the tremor number for different intervals of shear zone thickness as given in Table 2. The graph of the relationship between the average of the tremor number and shear zone thickness is described in Figure 6.6a in chapter 6.

Table 2. Average of the tremor number from 1 October 2015 to 1 April 2016

Thickness of shear zone (m)	Average of tremor number
100-200	2.6
200-300	3.4
300-400	2.9
400-500	4.2
500-600	4.0
600-700	3.9
700-800	4.4
800-900	4.2
900-1000	4.0
1000-1100	5.2
1100-1200	4.6
1200-1300	5.2
1300-1400	7.5
1400-1500	10.3
1500-1600	11.0
1600-1700	13.2
1700-1800	11.4
1800-1900	11.4
1900-2000	12.3
2000-2100	10.7

With the same steps, I calculated tremor density maps for tremor events from 2 to 15 April 2016 (Figure 6.5b in chapter 6) and from 16 April to 31 December 2016 (Figure 6.5c in chapter 6). Then, I calculated the average of the tremor number for different intervals of shear zone thickness in other episodes as shown in Table 3 and Table 4, respectively. The graph of the

relationship between the average of the tremor number and shear zone thickness is described in Figures 6.6b and 6.6c in chapter 6, respectively.

Table 3. Average of the tremor number from 2 to 15 April 2016.

Thickness of shear zone (m)	Average of tremor number
100-200	2.0
200-300	2.0
300-400	2.2
400-500	2.9
500-600	3.2
600-700	3.4
700-800	4.1
800-900	5.0
900-1000	5.2
1000-1100	6.0
1100-1200	7.7
1200-1300	11.5
1300-1400	14.3
1400-1500	15.6
1500-1600	15.3
1600-1700	16.8
1700-1800	14.7
1800-1900	12.7
1900-2000	11.9
2000-2100	10.0

Table 4. Average of the tremor number from 16 April to 31 December 2016.

Thickness of shear zone (m)	Average of tremor number
100-200	4.1
200-300	5.9
300-400	4.5
400-500	5.4
500-600	5.2
600-700	5.8
700-800	5.2
800-900	5.6
900-1000	6.0
1000-1100	6.3
1100-1200	6.6
1200-1300	6.1
1300-1400	9.3
1400-1500	7.9
1500-1600	8.3
1600-1700	6.1
1700-1800	5.5
1800-1900	5.0
1900-2000	3.0
2000-2100	3.0

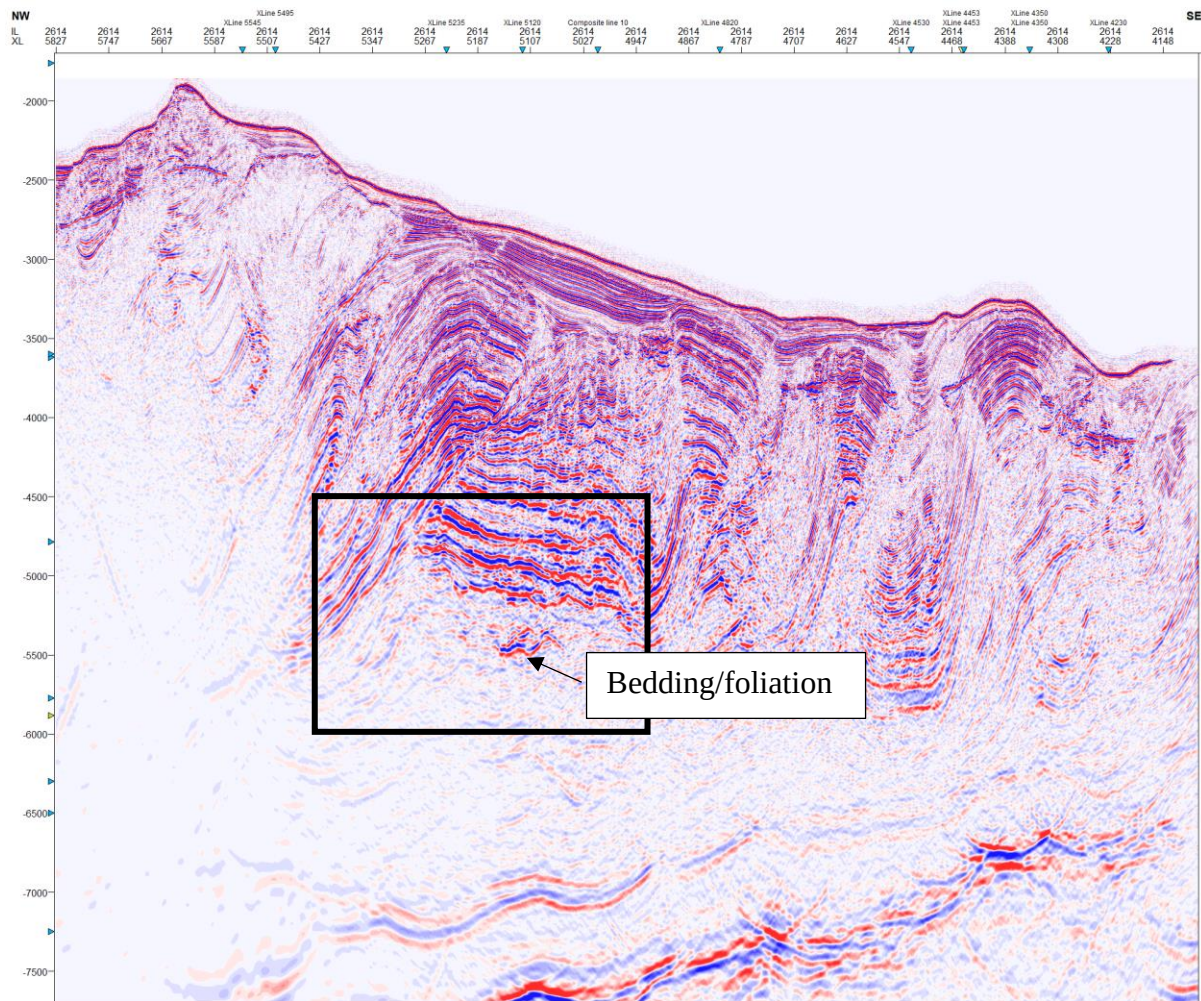


Figure A7. Seismic reflection profile (inline) showing the bedding or foliation into the shear zone in black box.

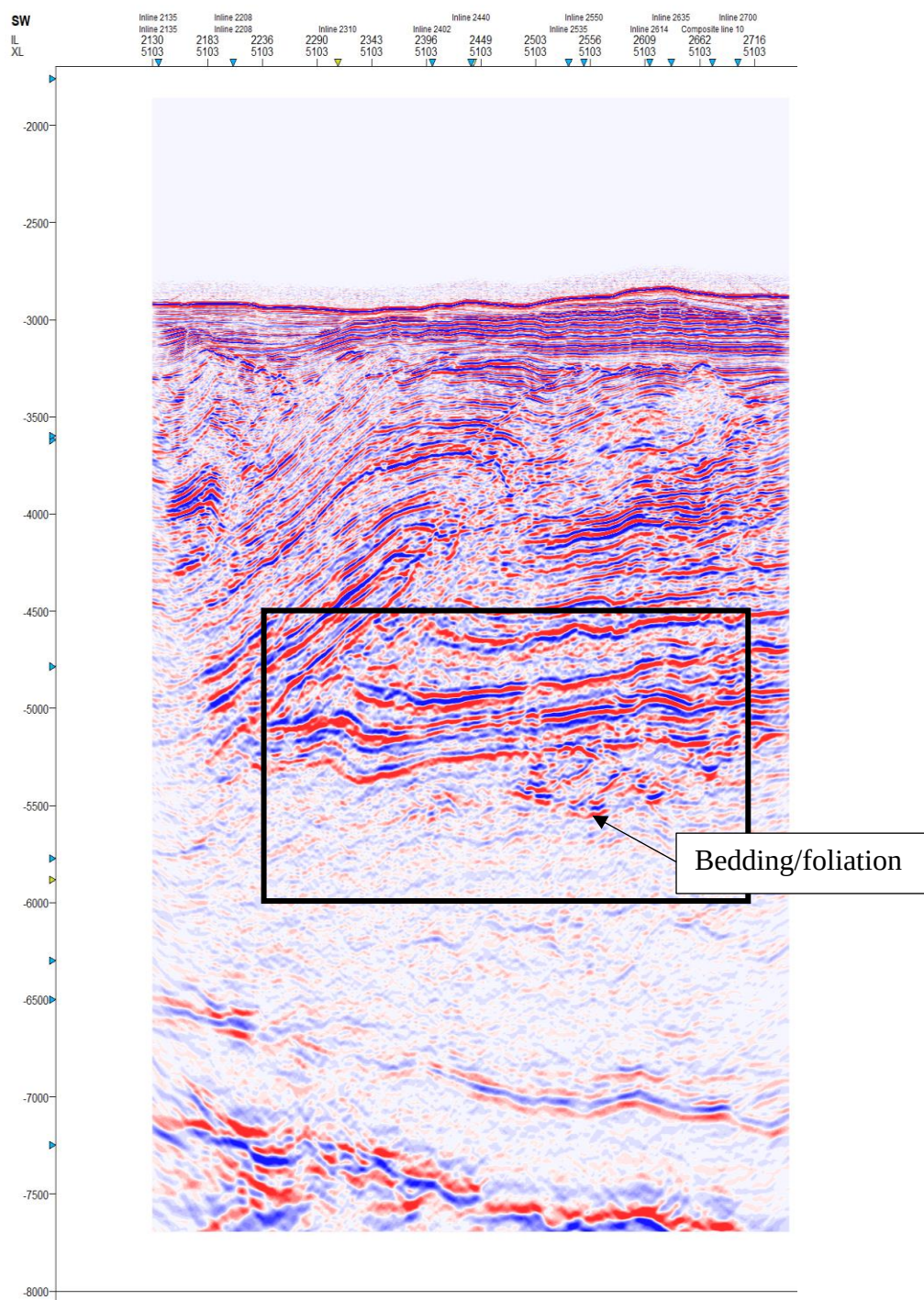


Figure A8. Seismic reflection profile (crossline) showing the bedding or foliation into the shear zone in black box.

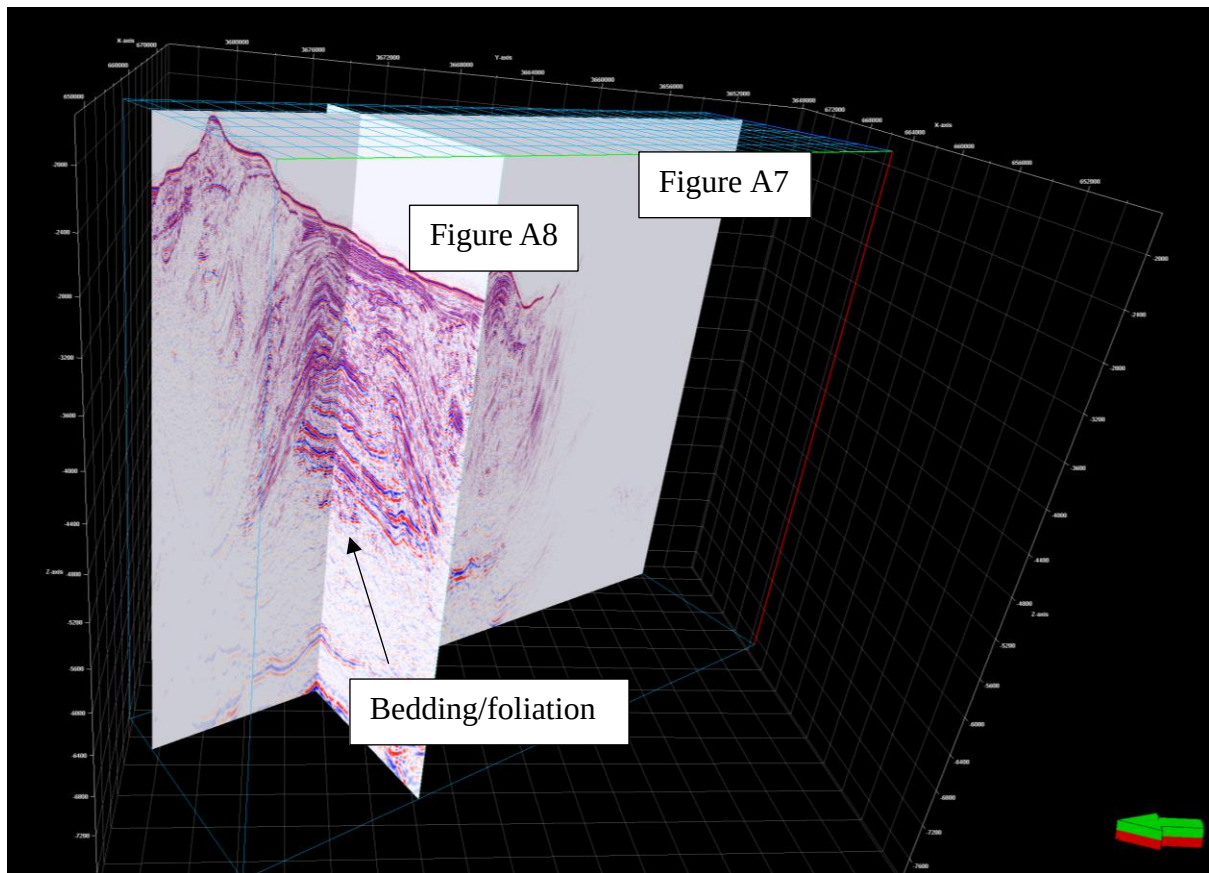


Figure A9. 3 Dimension of seismic data showing the bedding or foliation into the shear zone in Figure A7 and Figure A8.