

Evolution of the Mio–Pleistocene forearc basin
induced by the plate subduction
in the Boso Peninsula, central Japan

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Abstract

Subduction processes have controlled the geological evolution of the islands of Japan, and generate earthquakes, tsunamis, and volcanic eruptions. Forearc basins that formed between trench–slope breaks and volcanic arcs record the development of plate subduction in the region. The Boso Peninsula in central Japan is located at a plate boundary and comprises an accretionary prism, a trench–slope break, and a forearc basin, which indicate the peninsula has formed as a result of subduction. The forearc basin in the Boso Peninsula contains the large-scale Kurotaki Unconformity. Although the geological structure and lithologies associated with the Kurotaki Unconformity have been investigated, the details of its evolution, such as its thermal history and maximum burial depth, are unclear.

In this study, geological and engineering approaches were used to investigate the evolution of the Miocene–Pleistocene forearc basin in the Boso Peninsula. Given that forearc basins are also sites of natural gas energy resources, investigation of the forearc basin evolution is important from a resource perspective. In this study, geological investigations, vitrinite reflectance measurements, and consolidation tests were undertaken to investigate the geological structure, thermal history, and burial depth of the Boso forearc basin.

Geological observations suggest that deformation is more intense near the southern part of the forearc basin, close to the trench–slope break. Mesoscale faults are distributed in this region, and most of the normal and reverse faults in the forearc basin have obvious slip displacements that are now healed, which suggest these faults formed soon after sedimentation. A structural analysis shows that the fault attitudes are different above and below the Kurotaki Unconformity, whereby both E–W- and NE–SW-trending faults are present in the Miura Group (below the unconformity), but only NE–SW-trending faults are present in the Kazusa Group (above the unconformity). Fabric analysis using anisotropy of magnetic susceptibility measurements suggests that both the Miura and Kazusa groups have a generally consolidated fabric. However, the fabric of the Miura Group is weaker than that of the Kazusa Group.

The maximum paleo-temperature of the Miura Group was ca. 70–90°C and essentially uniform across the western and eastern parts of the basin. In the central area of the basin, part of the Kinone Formation records the highest paleo-temperature (>100°C). Given the location and temperature of the Mineoka ophiolite complex, it is suggested that advective heating occurred during emplacement of the ophiolite, which heated the Miura Group. The paleo-temperature of the Kazusa Group was ca. 50°C in the western part of the basin, whereas it was 75–90°C in the eastern part. Therefore, the paleo-temperatures obviously differ between the Kazusa and Miura groups (i.e., above and below the Kurotaki Unconformity in the western part of the basin). The paleo-temperature distribution in the eastern part of the Boso forearc basin implies that rapid, continuous, and thick basin-fill sediment accumulation has occurred.

The results of consolidation tests indicate the western, central, and eastern parts of the basin have different characteristics. A difference in the relationship between the consolidation yield stress (p_c) and formation thickness was confirmed in the Miura and Kazusa groups in the western part of the basin, which suggests that the eroded thickness in this region is larger than the expected thickness. The normal consolidation trend in the central and eastern parts of the basin is ca. 4 and 8 MPa lower than the calculated overburden pressure using the sediment thicknesses, respectively, whereas there is no difference in the consolidation trends between the two groups. In addition, strain softening was identified in the eastern part of the basin. Given that macro- and/or micro-cracks such as brittle fractures were not found in the specimens after testing, the strain softening may be due to actual softening of the siltstone lithology. In the eastern part of the basin, the consolidation yield stress for some formations was lower than the normal consolidation trend. Given the distribution of mass transport deposits, it is suggested that the pore fluid pressure was higher than the hydrostatic pressure (i.e., the formations were excess-pressured).

The geological, paleo-thermal, and consolidation test data were integrated to develop an evolutionary model for the Boso forearc basin. In the western part of the basin, major

differences above and below the Kurotaki Unconformity were identified in the distribution of geological units, and structural, paleo-temperature, and consolidation data. These variations probably relate to a change in the convergence direction of the Philippine Sea Plate at ca. 3 Ma. Uplift and erosion of the Miura Group during the formation of the unconformity and compression parallel to the direction of subduction caused the variations in the attitudes of the strata. The forearc basin, particularly near the trench–slope break, has structural features and physical properties that reflect plate tectonic processes and development of the trench–slope. In the eastern part of the basin, the consolidation yield stress indicates that burial in the forearc basin occurred despite the development of the Kurotaki Unconformity, as evidenced from the p_c data for the Kazusa and Miura groups above and below the unconformity, respectively. Vitrinite reflectance data that correspond to paleo-temperatures also show no abrupt or gradual changes across the Kurotaki Unconformity. Given this observation is similar to the p_c trend, any differences in vitrinite reflectance across the unconformity are considered to have been overprinted by the burial process.

There are no abrupt changes in the vitrinite reflectance and consolidation yield stress data across the Kurotaki Unconformity in the eastern part of the basin, which suggest that any such changes were overprinted during the progressive deposition and burial of the Kazusa Group in this area. The continuous trends in vitrinite reflectance and consolidation yield stress data in the eastern part of the basin between the Miura and Kazusa groups can be explained by the greater subsidence in this area, as compared with the western part of the basin, since ca. 2.3 Ma. Therefore, the contrasting evolutions of the western and eastern parts of the Boso forearc basin can be used to investigate the mechanisms of subduction.

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Chapter 1

Introduction

The islands of Japan have formed in a tectonic setting of plate subduction and related dynamic processes. At a subduction zone, unique structures are formed between the trench and volcanic arc, including accretionary prisms, trench–slope breaks, and forearc basins (Fig. 1.1). In addition, subduction can generate large earthquakes and tsunamis, meaning that studies of subduction zones are important for understanding how earthquakes and tsunamis are generated.

A forearc basin occurs between the volcanic arc and trench–slope break (e.g., Dickinson, 1995). The trench–slope break is a topographic high dividing the forearc basin from the trench–slope area, which marks the arc-ward limit of deformation related to plate subduction (Moore & Karig, 1976). The inner wedge corresponds to the forearc basin area and is generally stable throughout earthquake cycles, thereby acting as a

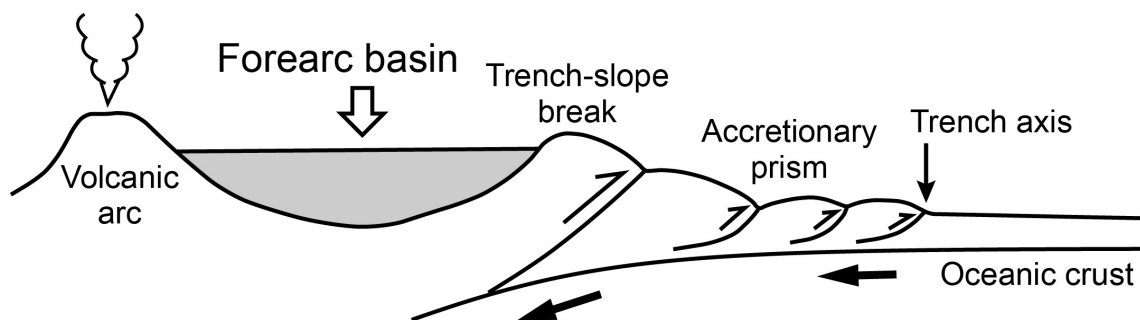


Figure 1.1 Schematic diagram of a subduction zone (modified after Dickinson, 1995).

backstop and providing a stable environment for the formation of forearc basins (Wang & Hu, 2006). Consequently, trench slopes preserve kinematic and mechanical information on subduction-related tectonics (e.g., Strasser et al., 2009; Chiyonobu et al., 2017). However, Moore et al. (2013) suggested that the formation of typical fault systems in the Kumano Basin, SW Japan, occurred in association with plate subduction and the formation of an accretionary prism along the Nankai Trough. The orientations of older fault populations were controlled by uplift of the underlying accretionary prism, implying that the forearc basin region is not as stable as previously thought. In contrast, the occurrence of active normal faults throughout the basin supports the idea that the horizontal compression oriented parallel to the direction of plate convergence does not reach the critical stress necessary to activate or form thrust faults, or produce horizontal shortening in the shallow part of the inner wedge (Moore et al., 2013). Therefore, geological structures such as folds and faults in forearc basins may yield unique information regarding tectonic processes related to plate subduction, as such structures form in association with the development of the accretionary prism and trench–slope break.

The Kumano Basin has only recently been examined in detail (e.g., Expedition 319 Scientists, 2010; Ramirez et al., 2015), and the sensitivity of the structural and mechanical properties of the forearc basin to regional tectonics remains unclear. In comparison with the offshore Nankai forearc basin, a non-metamorphosed post-middle Miocene to Pleistocene forearc basin is exposed onland in the northern parts of the Miura and Boso peninsulas, central Japan. As such, the Boso Peninsula is an easily accessible forearc basin that can be considered a geological analogue of the Nankai Trough. Although two short sedimentary hiatuses are evident in the subduction zone along the Nankai Trough, the forearc basin preserves a continuous geological record of the past 15 Ma and is an ideal setting for examining how forearc basins record subduction-related tectonism. In this study, I undertook structural and paleo-temperature analyses of the forearc basin on the Boso Peninsula to examine the relationship between

basin evolution and subduction processes. Geotechnical consolidation tests were also conducted to examine variations in the consolidation characteristics of the basin.

Investigations of forearc basins are also relevant to natural resources, as they can be exploited for natural gas energy. In the eastern part of the forearc basin in the Boso Peninsula, natural gas and iodine are dissolved in formation brines. The distribution of natural gas and geological features of the natural gas field have been reported previously (e.g., Kawai, 1960, 1961). The structure of the gas field has been investigated using logging data and the relationships between the distribution of natural gas and sedimentary facies and groundwater (Kunisue et al., 2002). These studies suggested that the evolution of the forearc basin, such as the depositional environment and burial processes, controlled the generation of natural gas. As such, investigating the evolution of the Boso forearc basin is important for understanding the generation and/or reservation system of a range of natural resources.

Large-scale geological structures on the Pacific side of central to southwestern Japan formed coevally during the Cenozoic (Fig. 1.2). The Miyazaki area in and around Kyushu is characterized by a widespread unconformity that formed at ca. 2.5–2.0 Ma (Kitahori & Sakaki, 2013). The Kumano Basin, located offshore from the Kii Peninsula, and a forearc basin in the Nankai Trough record a rapid change in sedimentation rate at ca. 2.0 Ma (Expedition 319 Scientists, 2010), and a marginal uplift zone formed at Kakegawa in Shizuoka Prefecture, central Japan (Oda et al., 2011). These tectonic events along the Nankai Trough occurred at the same time, and likely resulted from a change in the movement direction of the Philippine Sea Plate (e.g., Nakajima, 2018). The Boso Peninsula of central Japan, which developed as a result of subduction of the Philippine Sea Plate beneath the North American Plate, contains a late Miocene–Pleistocene forearc basin. One of the large-scale geological structures that formed ca. 20 km inland and parallel to the northern boundary of the Philippine Sea Plate (Sagami Trough) is the ca. 3 Ma Kurotaki Unconformity, which is only observed on the Boso Peninsula.

The Kurotaki Unconformity represents a depositional hiatus and marks the boundary between the underlying Miura Group and the overlying Kazusa Group in the Boso forearc basin. The orientations of geological structures (e.g., bedding and faults) differ between the Kazusa and Miura groups (Koike, 1951), and the stratigraphy is different in the eastern and western parts of the basin. The Miura Group has been substantially eroded in the eastern part of the basin, and thus the uppermost part of the group in the western part of the basin is absent in the east. The Kazusa Group is thickest in the eastern part of the basin, and the lower part of this group does not occur in the western part of the basin. The Boso Peninsula is located at the triple junction between the Philippine Sea, North American, and Pacific plates; consequently, the tectonics of the Boso Peninsula is complex and is a subject of ongoing research.

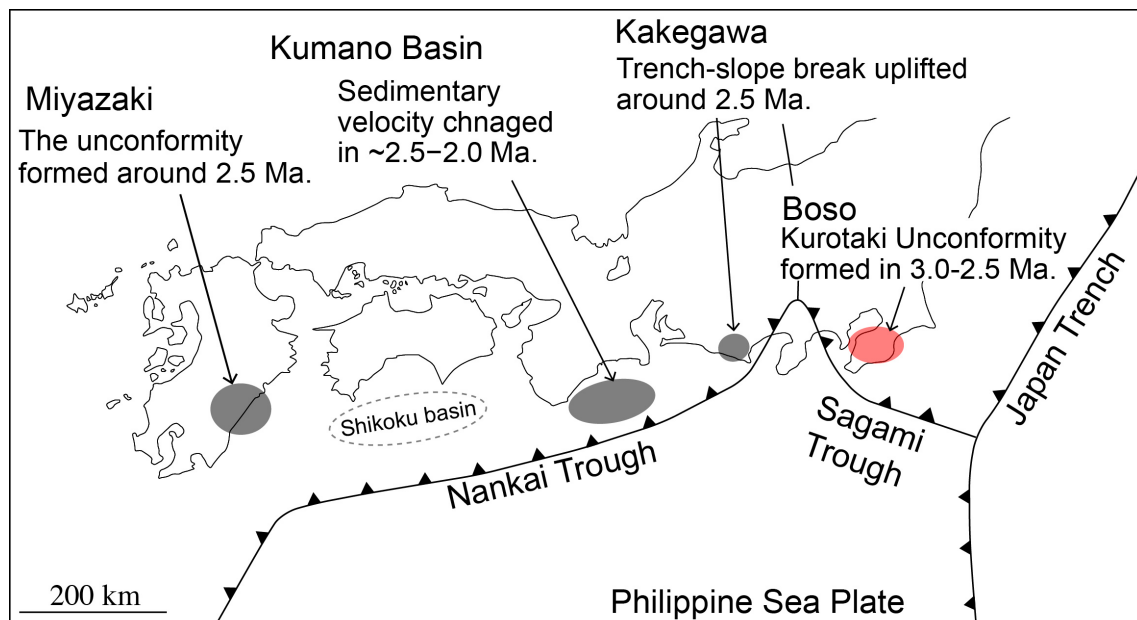


Figure 1.2 Correlation of Plio–Pleistocene tectonic events in central and southwest Japan.

The Boso forearc basin also contains sedimentary rocks that are categorized as soft rocks. Given that soft unconsolidated sediments change to hard rocks during diagenesis and consolidation, the mechanical behavior of soft rocks is categorized as being intermediate between that of soils and hard sedimentary rocks. In general, the physical properties of sedimentary rocks are affected by diagenesis and metamorphism, including heating and compaction. However, the sedimentary rocks of the Boso Peninsula, especially in the Boso forearc basin, have not been metamorphosed. Therefore, re-mineralization and cementation is limited, and the rocks have retained their mechanical properties during consolidation.

The structure of this thesis is shown in Fig. 1.3, and an outline of each chapter is provided below.

In Chapter One, the background and objectives of the study are described. An overview of subduction zone tectonism is presented, and important research questions about the simultaneous geological event occurred along the Philippine Sea plate outlined.

In Chapter Two, the geology and tectonic setting of the Boso Peninsula and forearc basin are summarized based on previous studies. The location of the study area and its geological structures are described. The structures, lithologies, sediment ages, and other noteworthy features of the Boso forearc basin are also provided.

In Chapter Three, the geological features of the Boso forearc basin determined from field surveys are presented, including outcrop descriptions, spatial trends in the deformation, thin-section observations, and anisotropy of magnetic susceptibility data for mudstones. The Boso forearc basin is deformed, with the deformation being stronger toward the south. Observations of outcrops and thin sections suggest that meso-scale faults in the Boso forearc basin formed soon after deposition of the sediments (Kamiya et al., 2017). Fault orientation data indicate that the Boso forearc basin was formed under compressive stress oriented parallel to the direction of plate convergence. In

addition, analysis of the anisotropy of magnetic susceptibility suggests that mudstones in the Boso forearc basin have a consolidation fabric (Kamiya et al., 2018).

In Chapter Four, paleo-temperatures obtained from vitrinite reflectance measurements are presented, along with their spatial distribution. Below the Kurotaki Unconformity, the maximum paleo-temperatures are similar in all formations, whereas above the unconformity the paleo-temperatures differ between the western and eastern parts of the Boso forearc basin. There is a clear difference in paleo-temperatures across the unconformity in the western part of the basin, whereas this is not seen in the eastern part of the basin (Kamiya et al., in press a). Given the presence of a high-temperature ophiolite complex south of the forearc basin, the paleo-temperature distribution suggests the maximum temperature attained by formations below the Kurotaki Unconformity reflects advective heating of the trench–slope break by the ophiolite (Kamiya et al., 2017).

In Chapter Five, I present yield stress data from consolidation tests. These data are used to constrain the maximum overburden pressure of the Boso forearc basin. The consolidation tests show different results in the western and eastern parts of the basin (Kamiya et al., in press a), and strain softening related to consolidation processes was identified (Kamiya et al., in press b). In addition, the eastern part of the Boso forearc basin contains a unit in which the consolidation yield stress is smaller than the consolidation trend, which suggests the unit has been in a zone of excess pore pressure (Kamiya et al., 2018).

In Chapter Six, the burial history of the Boso forearc basin is discussed and an evolutionary model of the basin is proposed, based on the results presented in Chapters Three to Five. The relationship between plate tectonics and the evolution of the Boso forearc basin is also discussed. The geological features, paleo-temperatures, and consolidation yield stresses differ across the Kurotaki Unconformity in the western part of the basin. However, no differences in paleo-temperature, and only a gradual change in consolidation trend, were identified across the unconformity in the eastern part of the

basin. These differences between the western and eastern parts of the Boso forearc basin reflect variations in burial depth after ca. 2.3 Ma (Kamiya et al., in press a).

In Chapter Seven, the conclusions of this research are presented.

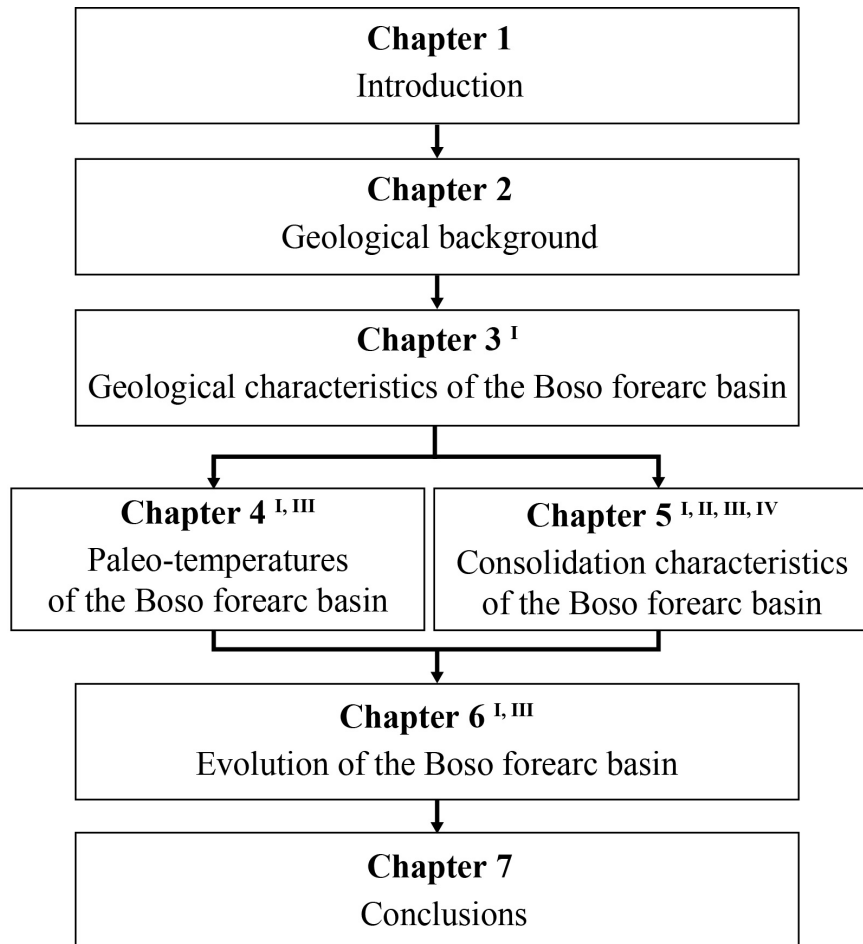


Figure 1.3 Thesis structure. Roman numbers indicate published and submitted papers on these topics mentioned in each chapter (I: Kamiya et al., 2017; II: Kamiya et al., 2018; III: Kamiya et al., in press a; IV: Kamiya et al., in press b).

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Chapter 2

Geological background

2.1 Tectonic setting of the Boso Peninsula

The Boso Peninsula in central Japan is located north of the Sagami Trough, which marks the boundary between the North American and Philippine Sea plates (Fig. 2.1). Given that the Pacific plate also subducts under the Philippine Sea plate, the Boso Peninsula is located at a triple junction between these three plates, which means the subduction processes around the Boso Peninsula are complex. Beneath the Boso Peninsula, the depth to the top of the Philippine Sea plate is 20–30 km, whereas the depth to the top of the Pacific Plate is 60–80 km (Ishida, 1990; Matsubara et al., 2005).

Tokuyama et al. (2001) analyzed the submarine fault distribution around Japan, based on high-resolution bathymetric maps, and identified spatial and temporal changes in tectonism due to subduction. Near the Sagami Trough, the Pacific Plate has been subducting to the WNW since ca. 37 Ma (Maruyama & Seno, 1986). Paleostress analysis based on the fault distribution in the Boso Peninsula indicates that the tectonic stress field around the peninsula changed from major NNE–SSW compression prior to 2 or 3 Ma to more recent major NNW–SSE compression in the west and WNW–ESE extension in the east, since 2–1 Ma (Angelier & Huchon, 1987). These authors also suggested that the direction of convergence of the Philippine Sea Plate changed from NNW–SSE to NW–SE. Based on geological data from the Shizuoka district, the subduction direction of the Philippine Sea Plate was towards the NNW prior to 5 Ma and towards the NE after ca. 2 Ma; the tectonics were affected by the collision of the Izu-Bonin Arc to Honshu arc at ca. 1 Ma (Sugiyama, 1992). The Euler pole of the Philippine Sea Plate migrated to its present location at 3 Ma, coincident with an

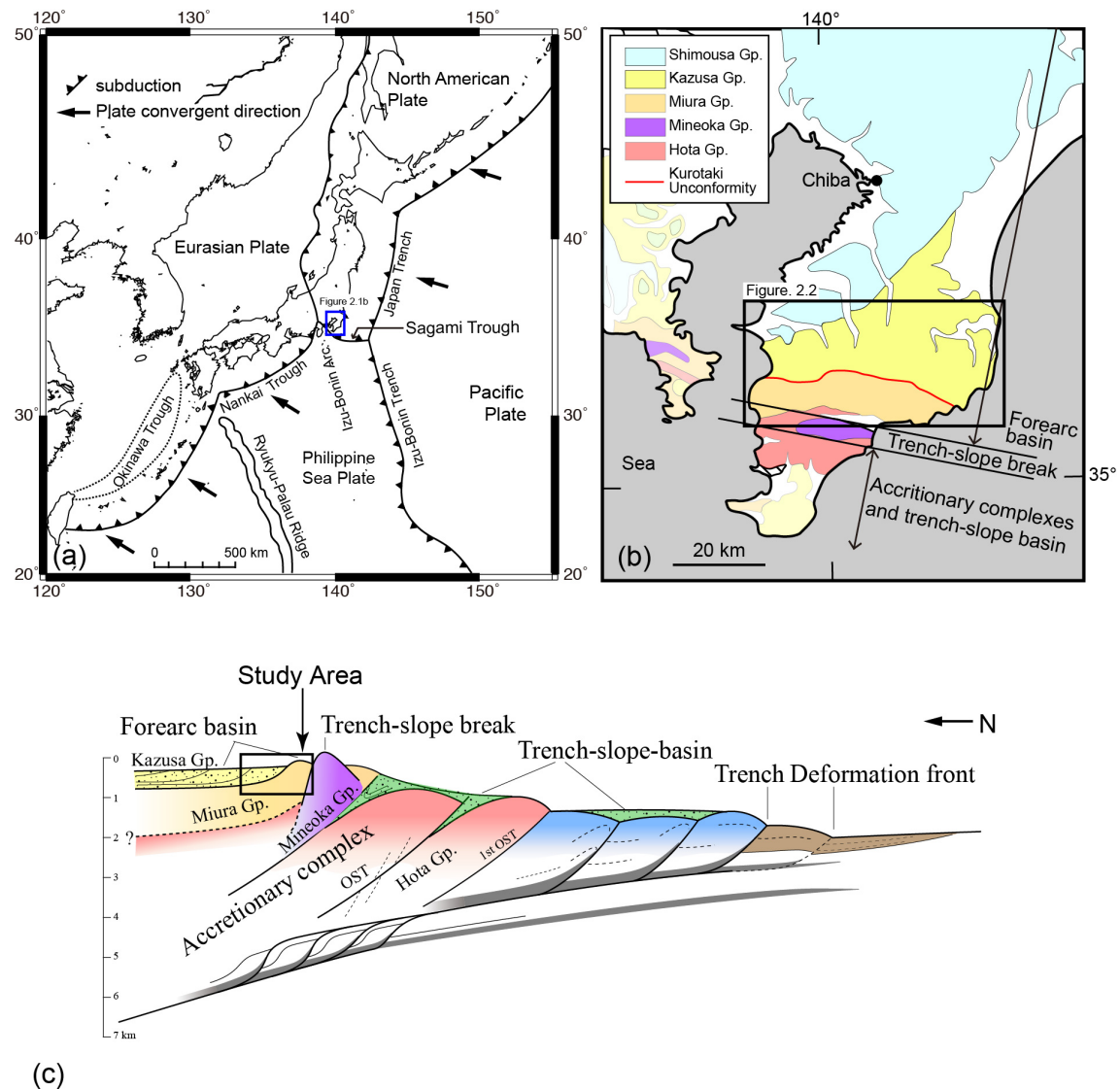


Figure 2.1. Tectonic setting and geological structure of the Boso Peninsula. (a) Plate configuration of Japan. The Boso Peninsula is located in the northern part of the Sagami Trough, which is the boundary between the Philippine Sea and North American plates. (b) Geological map of the Miura and Boso peninsulas. The geology of the Boso Peninsula is subdivided into three parts: early Miocene and late Miocene–Pliocene accretionary complexes and middle Miocene–Pleistocene trench–slope basins in the south; a post-middle Miocene forearc basin in the north; and an intervening trench–slope break that consists of an ophiolite complex. (c) Schematic cross-section of the Boso Peninsula (modified after Yamamoto et al., 2017).

interruption to forearc basin deposition in southwest Japan (i.e., the Kurotaki Unconformity; Takahashi, 2006).

The geology of the Boso Peninsula is dominated by strata that formed during subduction, and is divided into three parts from north to south (i.e., the subduction direction of the Philippine Sea Plate). The southern part comprises early Miocene to Pliocene accretionary complexes and trench–slope basin deposits (Saito, 1992; Yamamoto & Kawakami, 2005; Kawakami & Shishikura, 2006; Yamamoto et al., 2006; Chiyonobu et al., 2017). The middle part comprises the Hayama–Mineoka tectonic belt, which is dominated by ophiolites that mark a trench–slope break (e.g., Ogawa & Taniguchi, 1987), and the northern part is a post-middle Miocene forearc basin (Mitsunashi et al., 1979; Nakajima et al., 1981; Nakajima & Watanabe, 2005).

The accretionary complexes located in the southern part of the Boso Peninsula are divided into two types according to sediment age. The trench–slope basins cover these accretionary complexes and contain unconformities (e.g., Kawakami & Shishikura, 2006). The accretionary complexes and trench–slope basin formed in response to subduction of the Philippine Sea Plate and Izu–Bonin collision (Yamamoto & Kawakami, 2005).

The Hayama–Mineoka tectonic belt is located between the accretionary complexes and forearc basin. The ophiolitic rocks are mainly tholeiitic pillow basalts and diabases, alkali basalt lava flows, and calc-alkaline dioritic to gabbroic rocks (e.g., Sato et al., 1999). Hirano et al. (2003) inferred that the Mineoka ophiolitic assemblage was part of another Cenozoic oceanic plate (i.e., the Mineoka Plate), which was the site of island arc volcanism during the Miocene as a result of subduction initiation at a fracture zone or transform fault system, due to a change in the position of the Euler rotation pole of the Pacific Plate at ca. 43 Ma. However, the Mineoka ophiolite may represent intrusions that formed along a transform fault (e.g., Arai & Okada, 1991). Upper mantle to crustal rocks of the paleo-Izu arc were preferentially incorporated into serpentinite mélanges along the transform fault at the northern end of the Shikoku Basin during opening of the

basin in the Early Miocene. The basement rocks of the paleo-Izu arc and rocks of the Shikoku Basin were finally incorporated into the serpentinite mélanges (Takahashi et al., 2012). A pull-apart basin associated with a transform fault in the Boso Peninsula formed at ca. 15 Ma, related to opening of the Shikoku Basin (Saito, 1991). Subsequently, the Mineoka ophiolite has been located in the Hayama–Mineoka tectonic belt at the trench–slope break in the Miura and Boso regions since ca. 15 Ma, which led to the formation of the Boso forearc basin.

2.2 Geology of the Boso forearc basin

The Boso forearc basin contains the 15–3 Ma Miura and 2.5–0.3 Ma Kazusa groups (Fig. 2.2). The Miura Group is divided into the Neogene Kinone, Amatsu, Kiyosumi, and Anno formations (from bottom to top). The Kinone and Amatsu formations are predominantly hemipelagic mudstones, and the Kiyosumi and Anno formations are mostly turbidite successions. The total thickness of the Miura Group is >2000–3000 m (Suzuki, 2002). In the Miura Group, the bedding strikes essentially E–W and exhibits mesoscale folds. In contrast, the Kazusa Group was deposited in a marine environment during the Pliocene to middle Pleistocene and occurs mainly in the central and northeastern Boso Peninsula. This group is divided into the Kurotaki, Katsuura, Namihana, Ohara, Tomiya, Kiwada, Otadai, Umegase, Kokumoto, Ichijuku, Chonan, and Kasamori formations (from bottom to top; Mitsunashi & Suda, 1980; Nakajima & Watanabe, 2005), and has a total thickness of >2000 m (Suzuki, 2002). Bedding in the Kazusa Group strikes NE–SW in the eastern part of the basin and E–W in the western part (Fig. 2.2).

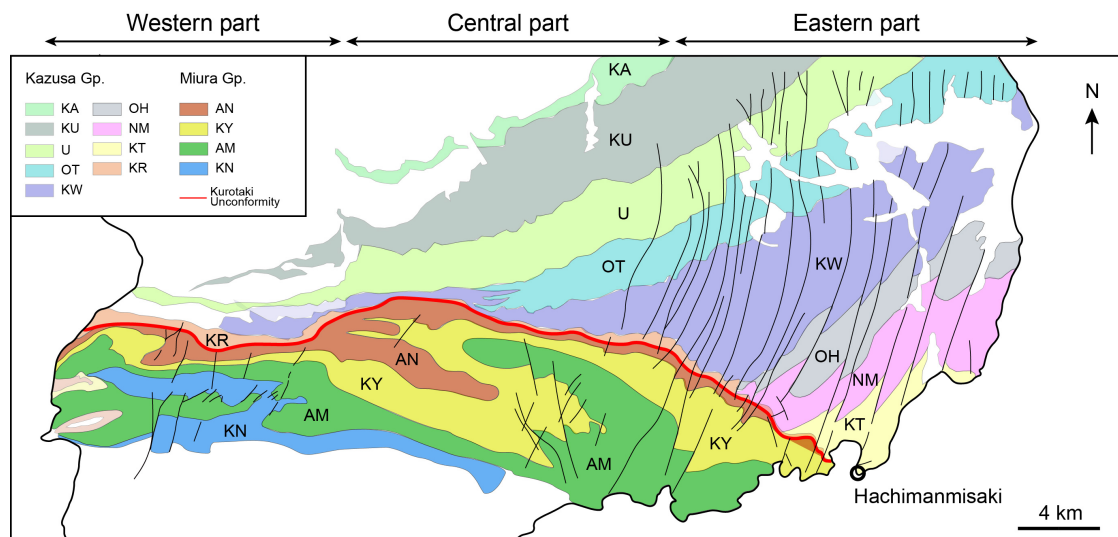


Figure 2.2 Geological map of the Boso forearc basin (modified after Mitsunashi et al., 1979; Nakajima and Watanabe, 2005). The abbreviations for the formations in the legend are based on Fig. 2.3.

The boundary between the Miura and Kazusa groups is marked by the Kurotaki Unconformity (Koike, 1951), which represents a depositional hiatus and submarine erosion at ca. 3 Ma when the convergence direction of the Philippine Sea Plate changed (Takahashi, 2006). The lithostratigraphy differs between the eastern and western Boso Peninsula (Fig. 2.3). The Kurotaki Unconformity is present in the upper Anno Formation in the west and center of the peninsula, and in the upper Kiyosumi Formation in the east. The entire Anno Formation and the upper part of the Kiyosumi Formation have been removed by erosion in the east. The lower Kazusa Group includes the Katsuura, Namihana, and Ohara formations, which overlie the Miura Group in the east of the peninsula, but these formations were not deposited in the west. The Kazusa Group overlies the Kurotaki Unconformity with a maximum thickness in the eastern Boso Peninsula. The time gap represented by the Kurotaki Unconformity is >1 Myr in the western part of the peninsula and >2.0 Myr in the east (Kameo & Sekine, 2013; Tamura et al., 2019). Given these characteristics, the Boso forearc basin is divided into three parts (western, central, and eastern) in this study.

The Miura Group in the southern part of the Boso forearc basin is folded along E–W-trending fold axes. Fujita & Kodama (1990) and Ban et al. (1990) suggested that the folding resulted from two-dimensional, thick-skinned deformation, as inferred from a vertical cross-section oriented perpendicular to the fold axes. Kokado et al. (2015) tested this hypothesis and proposed that the strata were subjected to thin-skinned rather than thick-skinned deformation. They found that lateral variations in strata thickness record three-dimensional, syn-depositional, map-scale fold structures and outcrop-scale faults that reflect the pattern of three-dimensional deformation. The Anno Formation in the upper part of the Miura Group comprises a turbidite succession (e.g., Ishihara et al., 1997). Most paleocurrent directions for the turbidites, as determined from current ripple marks, sole marks, and the fabric of mud clasts, show southward flow from the central part of the peninsula, although some show northward flow in the southern part and variable directions in the western part (Ishihara & Tokuhashi, 2005). Yamaji et al.

(2003) identified NNE–SSW-trending normal faults in the Kazusa Group in the eastern part of the basin and suggested that this area was subjected to WNW–ESE-directed tensile stress at ca. 1.0 Ma.

The depocenter of the Kazusa Group has shifted intermittently towards the northwest with time. The upper part of the Kazusa Group was deposited in a neritic sea setting, and the depocenter was in the northeastern part of the peninsula and northern part of Tokyo Bay (Suzuki et al., 1983). The Kazusa Group is also a natural gas reservoir and shows an increasing gas-to-water ratio with increasing gas production. After the deposition of the Kazusa Group, the central part of the peninsula was strongly uplifted, whereas the northern part was uplifted by a small amount (Suzuki et al., 1983).

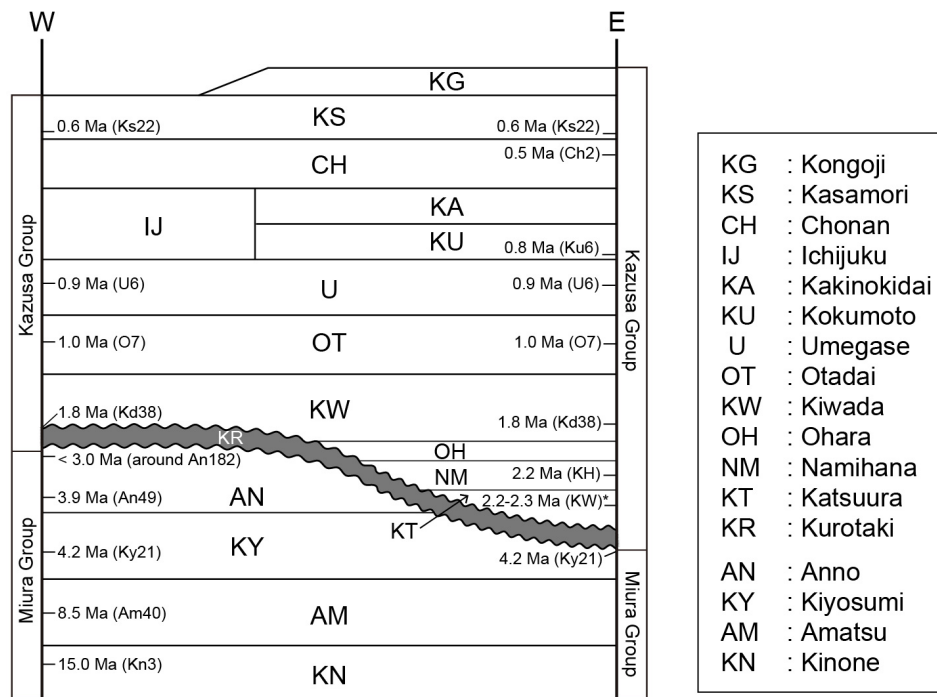


Figure 2.3 Schematic diagram showing the stratigraphy of the Boso forearc basin. The age data are constrained by key tephra layers (shown in parentheses following the age data) in each formation (Tokuhashi & Endo, 1984; Nakajima & Watanabe, 2005). The age data marked with an asterisk are based on calcareous nanofossils, and the other data are zircon fission-track ages (Tamura et al., 2019).

Numerous tuff layers on Boso Peninsula serve as useful marker beds for correlations and inferring the orientation of the basin and structures. The tuff layers are distinguished based on characteristics such as facies, thickness, grain size, and sedimentary conditions. Refractive index values and major and minor element data for the volcanic glasses have been used to characterize the tuff layers. Based on these data, key marker beds have been identified in each area.

Figure 2.4 shows the tephra marker bed Am78, which is characterized by five “salt and pepper” layers and a thin scoria layer. This marker bed is widely distributed in the Miura and Boso peninsulas. In the Boso Peninsula, the distribution and correlation of these marker beds have been investigated in detail (e.g., Urabe et al., 1990; Satoguchi et al., 2000; Takahashi et al., 2005). Given that the thickness of the tuff layer changes with location, the thickness between the top of the group and each sampling point was calculated for each part. The thickness of each tuff marker bed varies across the study area due to changes in the depositional environment (Fig. 2.5).

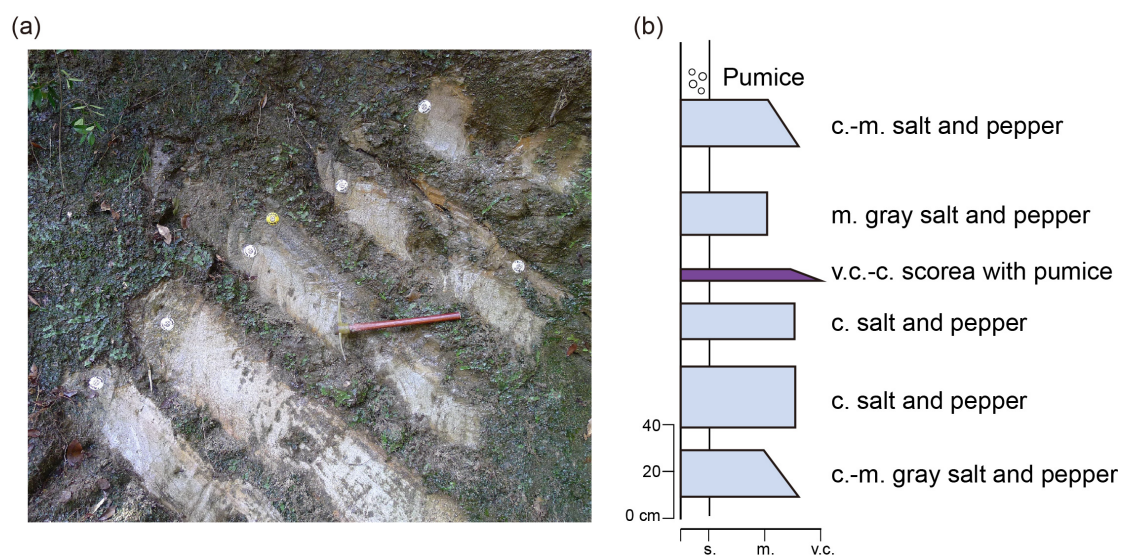


Figure 2.4 The Am78 (OK) marker bed at Minato River in the western part of the Boso forearc basin. (a) Photograph. (b) Schematic stratigraphy. Horizontal axis shows grain size and is divided into silty sand (s.), medium sand (m.), coarse sand (c.), and very coarse sand (v.c.).

In the Boso forearc basin, numerous tephra marker beds have been identified. In the Miura Group, the Kinone, Amatsu, Kiyosumi, and Anno formations contain 2, 101, 33, and 182 numbered major marker beds, respectively (Natural History Museum and Institute, Chiba, 1993, 1994, 1995, 1996, 1997, 1998). In the lower Kazusa Group, the Katsuura, Namihana, Ohara, Kiwada, Otadai, Umegase, and Kokumoto formations contain 8, 20, 19, 48, 27, 12, and 6 numbered major marker beds, respectively (Utsunomiya et al., 2019). In the upper part of the Kazusa Group, the Kakinokidai, Chonan, Kasamori formations contain 2, 3, and 17 numbered major marker beds, respectively (Nanayama et al., 2016).

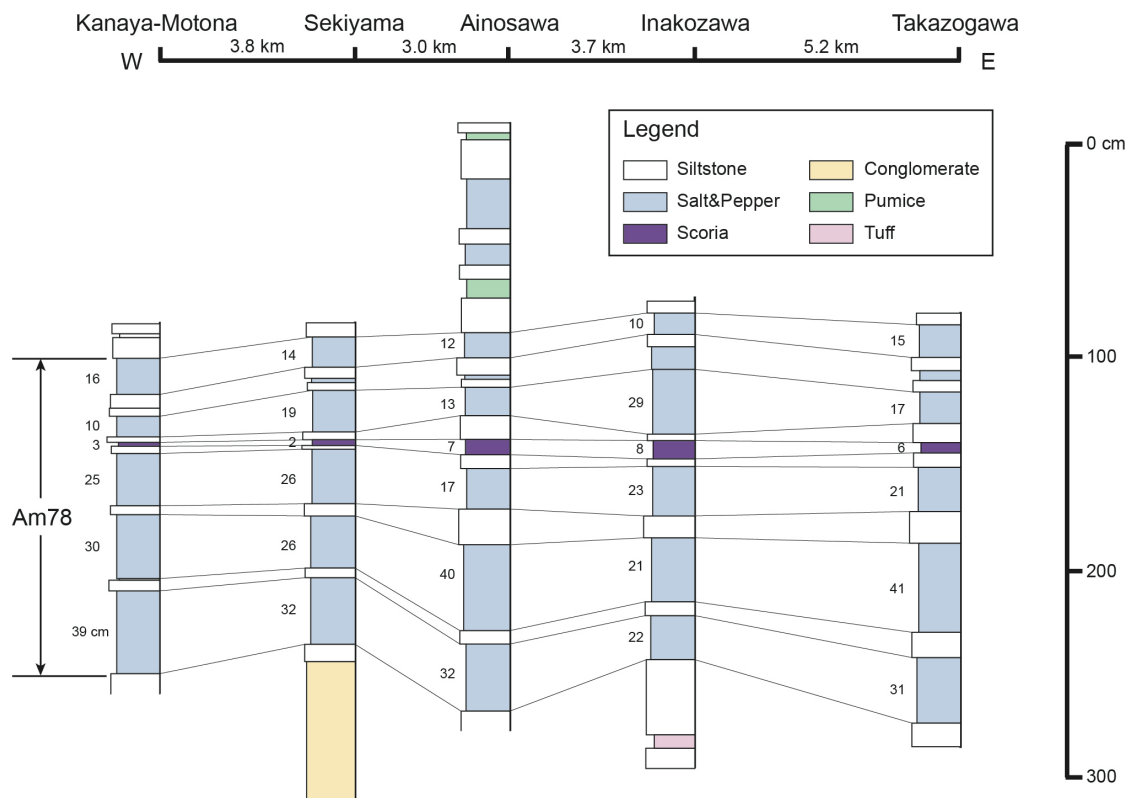


Figure 2.5 Schematic stratigraphy of the Am78 marker bed in the western part of the Boso forearc basin (modified from Nakajima & Watanabe, 2005). Numbers are thickness of each layer.

Mass transport deposits (MTDs) are abundant in the lower Kazusa Group. To understand the generation of these deposits, it is necessary to understand the physical and rock mechanical characteristics of the basin fill. Excess fluid pressure can be generated in geological bodies by fluid migration from deeper or lateral sections, rapid sedimentation, dehydration of clay minerals, and liquefaction of sandy layers. This excess fluid pressure can trigger slope failure that results in MTDs (Kvalstad et al., 2005). In the deep waters of the Gulf of Mexico, overpressure exists within low-permeability mudstones just above the aquifer, where multiple large-scale MTDs have developed (Flemings et al., 2008). In this case, the overpressure reduces the slope stability, leading to the formation of large-scale MTDs. The consolidation state also has a significant impact on the generation of MTDs. Shear-induced densification occurs in the MTD interval in the Ursa region of the northern Gulf of Mexico (e.g., Yamamoto & Sawyer, 2012). The densification acts as a fluid source, and the low permeability of the dense MTDs maintains the overpressure of the sediments (Dugan, 2012).

The MTDs in the Kiwada Formation comprise blocks of folded mudstone and chaotic muddy deposits, which commonly coexist with injected sandy deposits (Ogiwara & Ito, 2011). The Ohara, Namihana, and Katsuura formations also contain MTDs that are several meters thick. One example is exposed at Cape Hachiman-misaki (see Fig. 2.2 for the location) on Boso Peninsula. The MTDs in the Katsuura Formation consist of silty blocks in a sandy matrix (Fig. 2.6a and b). The matrix comprises coarse- to medium-grained volcanic sands and contains fossil shells, pumice grains, and rounded gravel clasts of chert, shale, and granite. The thickness of the MTDs is 10–20 m over an areal width of 12 km (Koike, 1955).

Fluidization and minor liquefaction features, such as dish structures, sand dikes, and mud injections, were identified near the MTDs (Fig. 2.6c and d). Dish structures are characterized by thin mud layers in sandstone, which laboratory experiments suggest form by the fluidization of sand layers (Tsuji & Miyata, 1987). The sand dikes formed

by liquefaction. Both structures occur in unconsolidated layers, which implies their formation was associated with excess pore pressure (Anegawa & Miyata, 2001).

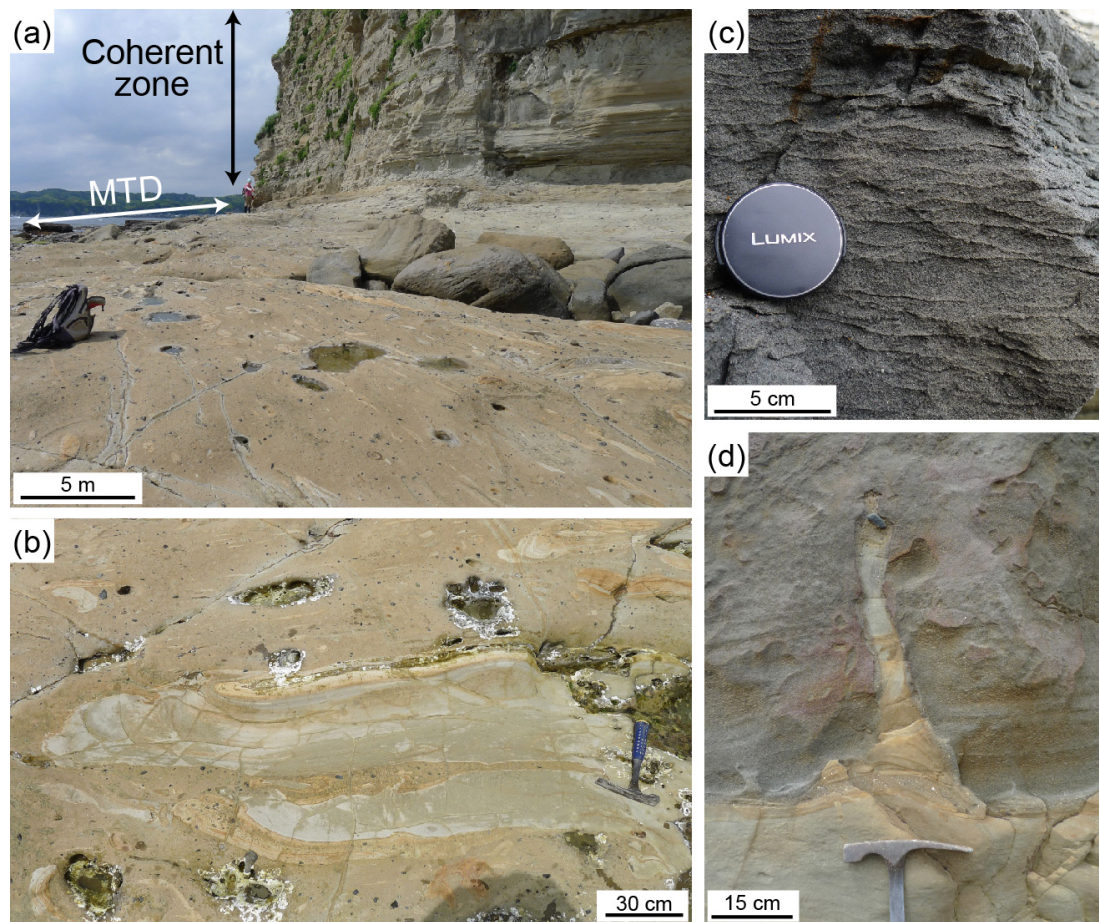


Figure 2.6 Photographs of outcrops of mass transport deposits (MTD). (a) A MTD exposed in the Hachimanmisaki region on the eastern coast of Boso Peninsula (Fig. 2.2). The MTD comprises a sandy matrix, silty blocks, and pumice grains. (b) A silty block in a sandy matrix. (c) Dish structure. (d) Sand dike.

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Chapter 3

Geological characteristics of the Boso forearc basin

3.1 Introduction

Geological structures yield basic information that can constrain the deformation of a rock mass. Deformation structures such as folds and faults reflect both the stress state and lithology, with the latter related to the environment of deposition. Published seismic cross-sections and geological maps only show megascale structures, and cannot depict mesoscale and microscale features. Therefore, geological fieldwork or surveys are critical for understanding the details of geological structures.

I mainly conducted geological fieldwork in the western part of the Boso Peninsula (Fig. 3.1). The Kanaya coast, Shikoma River area (hereafter called Shikoma River), and Minato River area were the survey routes. To document the detailed structure of the Boso forearc basin, I used the distribution of volcanoclastic marker beds for correlation and then constructed a cross-section. The naming scheme for key marker beds is from Nakajima and Watanabe (2005). Outcrop- and micro-scale structural analyses of the faults were also conducted to determine the fault characteristics.

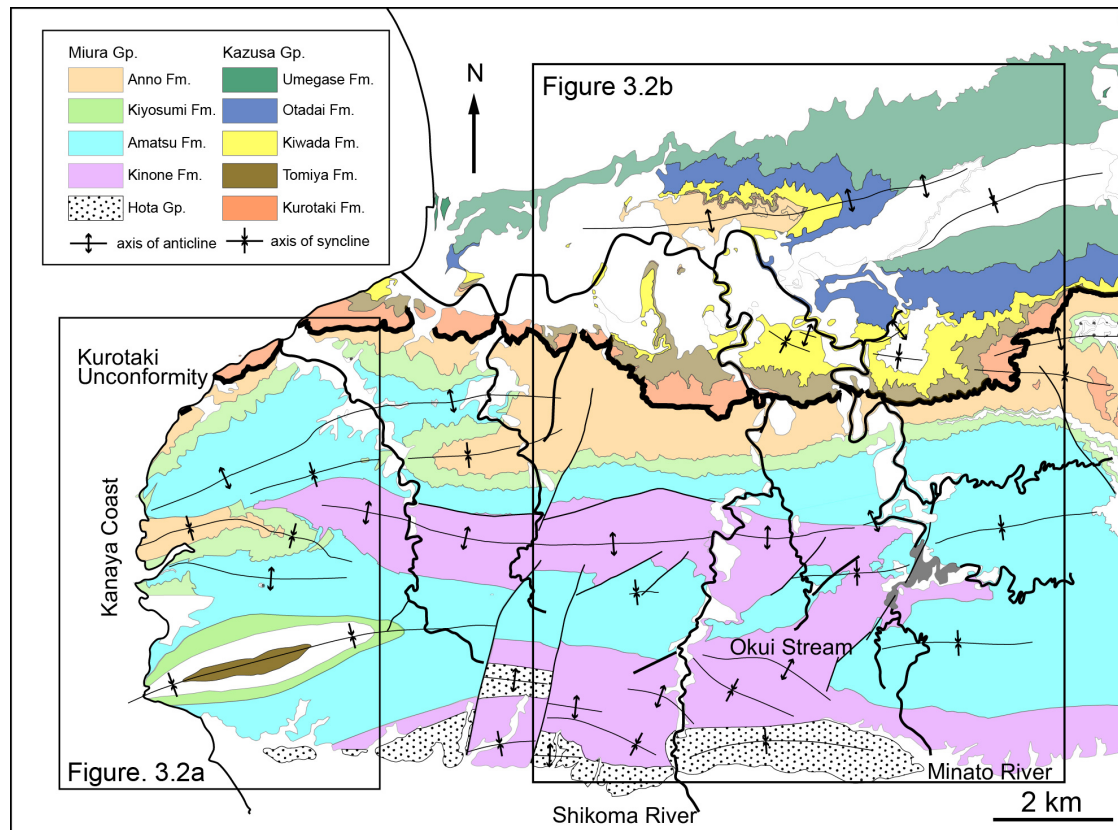


Figure 3.1 Geological map of the forearc basin on the western part of the Boso Peninsula (after Nakajima and Watanabe, 2005; Suzuki et al., 1990). The rock sequence of the Boso forearc basin consists of the 15–3 Ma Miura Group and 2.5–0.6 Ma Kazusa Group, and many E–W-striking folds occur in the central part of the peninsula. The Miura Group is divided into the Neogene Kinone, Amatsu, Kiyosumi, and Anno formations, and the Kazusa Group is divided into the Kurotaki, Tomiya, Kiwada, Otadai, Umegase, and Kokumoto formations in the study area.

3.2 Geological structure and lithology

The lithologies in the study area include conglomerate, sandstone, alternating sandstone–mudstone, mudstone, and numerous tuff marker beds. Along the Kanaya coast, the same key marker beds are repeated due to folding (Fig. 3.2a). The southern part of the Shikoma and Minato rivers are barely deformed, and some strata are overturned (Fig. 3.2b). Figure 3.3 is a cross-section along the Kanaya coast and Fig. 3.3 is a cross-section along the Minato River, based on Nakajima & Watanabe (2005). These cross-sections reveal that the deformation is more intense in the southern region. In the Shikoma and Minato rivers, outcrops of the Kurotaki Unconformity were observed. The lithology changes abruptly from alternating sandstone–mudstone to conglomerate at the unconformity, which is marked by erosion of the underlying layers in the Shikoma River (Fig. 3.4a). Mudstone blocks (10–30 cm in diameter) occur immediately below the unconformity (Fig. 3.4b).

Stereoplots of poles to bedding show a N–S girdle distribution, indicating that the bedding orientations are controlled by E–W folds (Fig. 3.5). Several fold axes were identified on the geological map and along the N–S-striking cross-section. Relatively intense deformation (i.e., faulting and folding) was observed in the southern part of the forearc basin, close to the trench–slope break (i.e., the Hayama–Mineoka Tectonic Belt), whereas deformation is weaker in the northern region. In particular, the dip of bedding is generally $\sim 30^\circ$ in the southern and $>80^\circ$ in the northern region. The Kurotaki Unconformity is exposed between the Anno and Kurotaki formations, and is characterized by basal conglomerates of the Kurotaki Formation. Kokado et al. (2015) suggested that large-scale folds in the Miura Group in the northwestern region formed by thin-skinned deformation, which did not involve basement deformation, and also that an anticline had formed due to tectonic compression.

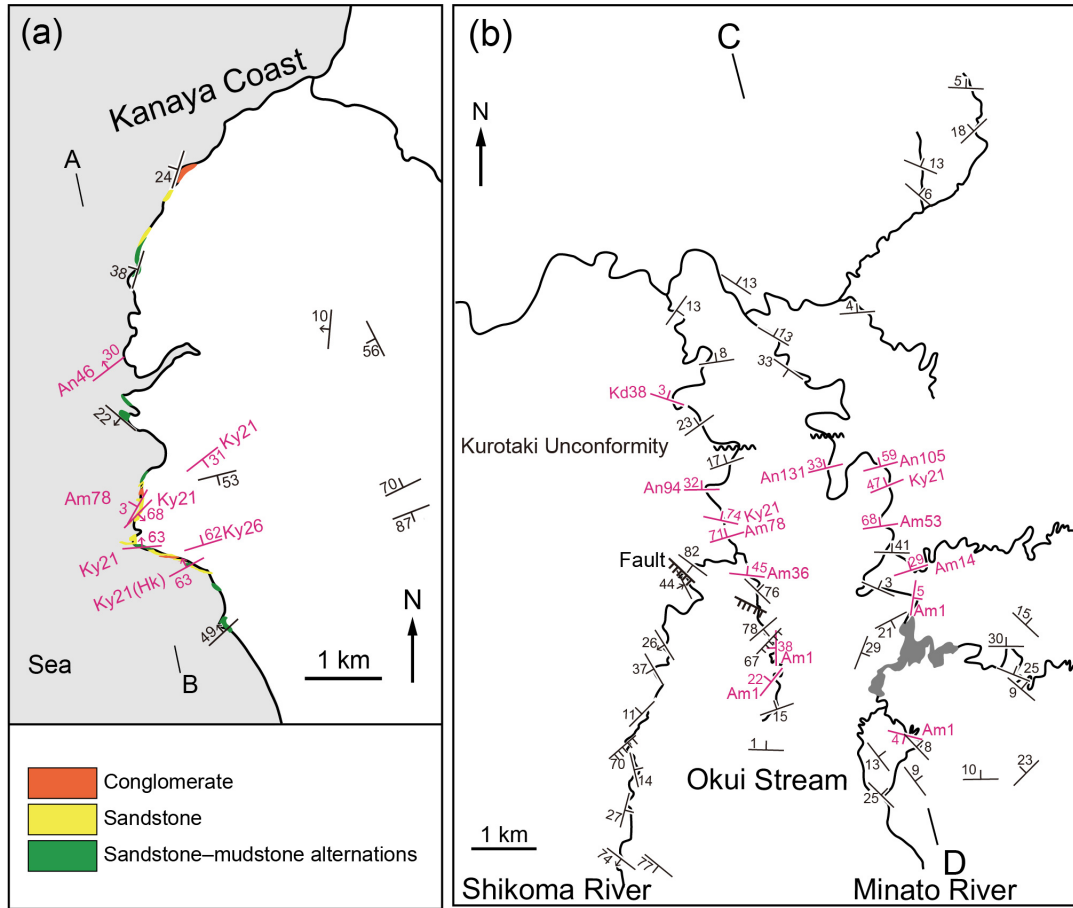


Figure 3.2 Geological traverses of the western part of the Boso forearc basin. The pink data are for tuff marker beds. (a) Kanaya coast. A–B shows the location of the cross-section in Fig. 3.3. (b) Shikoma and Minato rivers. C–D shows the location of the cross-section in Fig. 3.3.

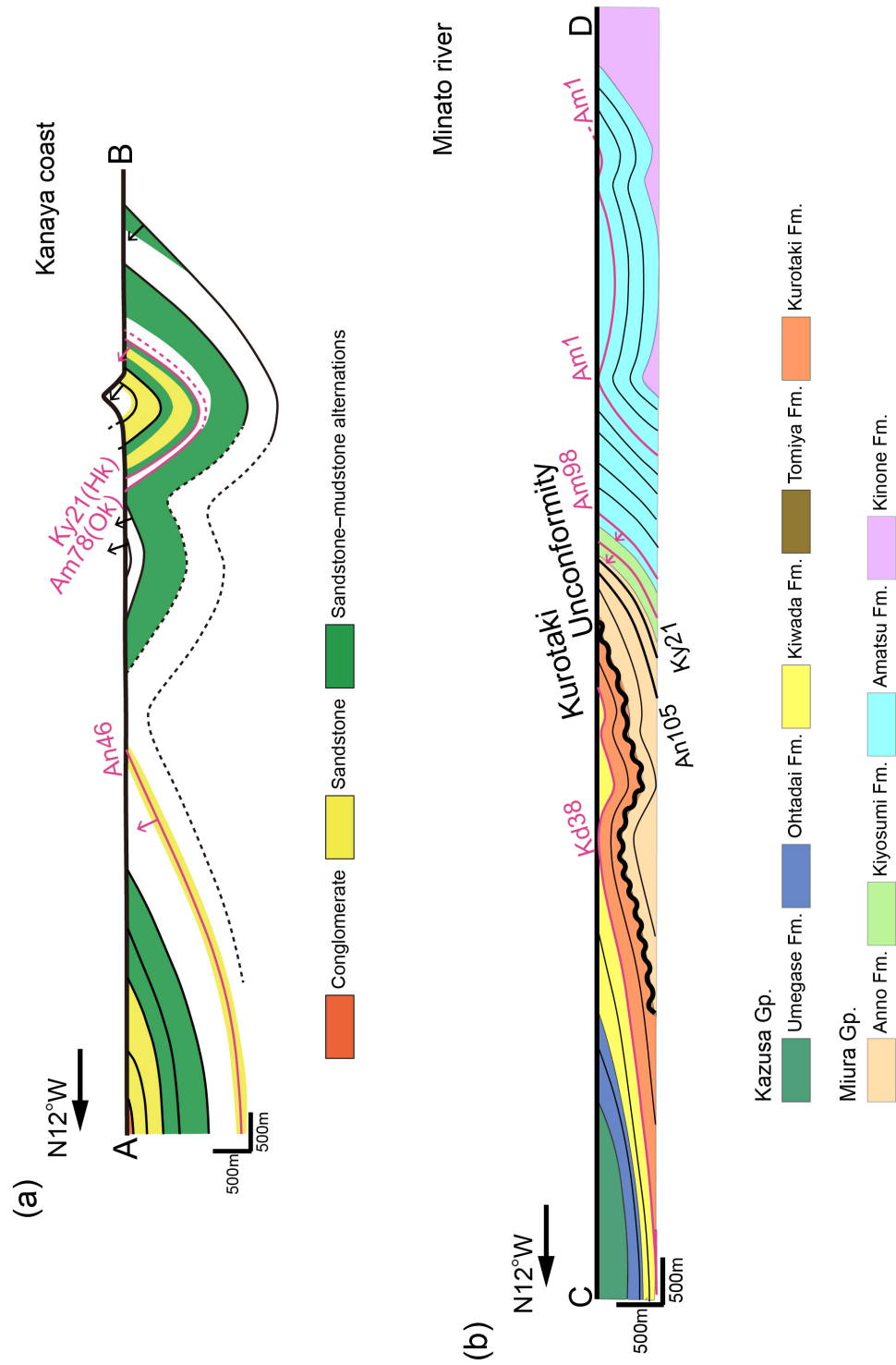
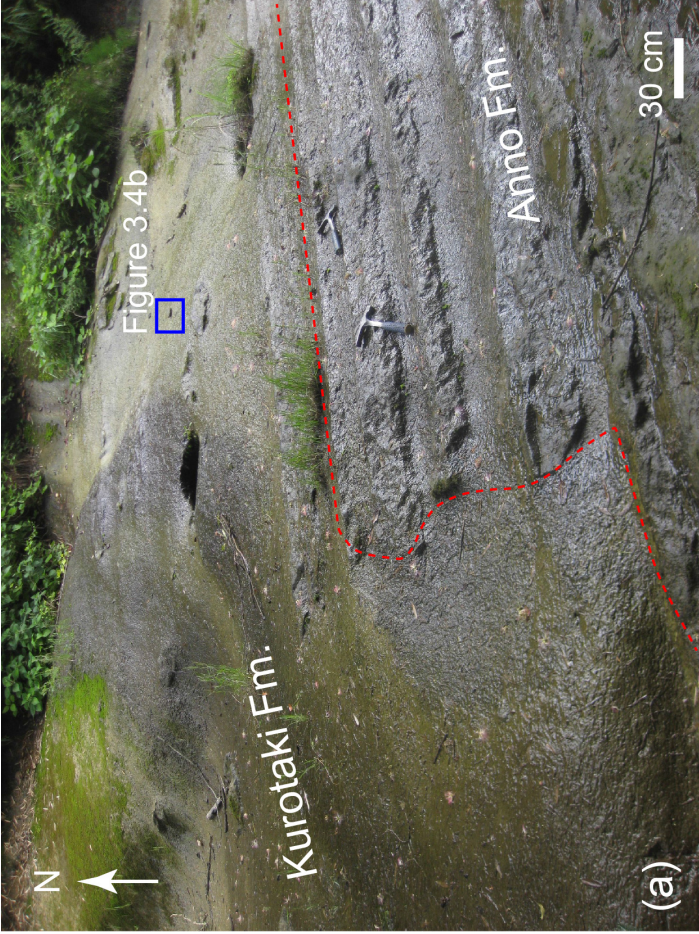


Figure 3.3 Cross-sections along the Kanaya coast (a) and the Shikoma River (b). Location is indicated in Fig. 3.2. The forearc basin sediments are folded.



Kurotaki Unconformity



Figure 3.4 Outcrop of the Kurotaki Unconformity in Shikoma River. (a) Photograph of the Kurotaki Unconformity (red broken line), which is the boundary between the Kazusa and Miura groups. (b) Mudstone block at the base of the Kurotaki Formation.

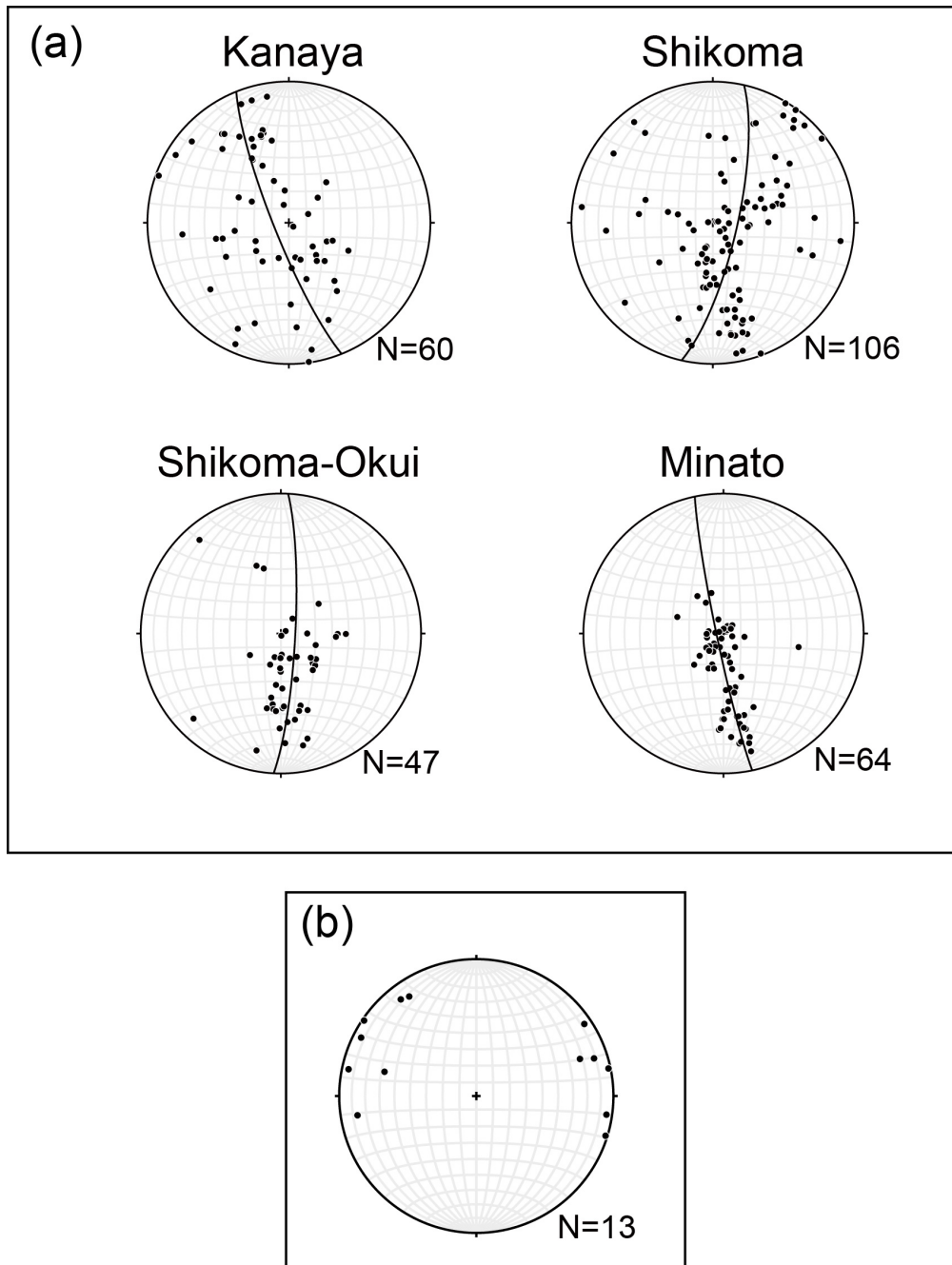


Figure 3.5 (a) Stereoplots showing poles to bedding planes and best-fit great circle for the study area (lower hemisphere equal-area projections). Data are separated into the Kanaya coast, Shikoma River, Shikoma River–Okui, and Minato River areas (Fig. 3.1). The orientation of bedding is controlled by approximately E–W-striking folds. (b) Stereoplots showing the orientations of fold axes in the study area.

3.3 Fault distribution

Mesoscale faults are widely distributed in the western part of the Boso forearc basin. Figure 3.6 shows photographs and sketches of outcrops of these faults. The displacement along the faults is 30–100 cm, and some of the faults have produced drag folds. Attitudes of the faults are shown in stereoplots as black lines representing the measured orientation and red lines representing the restored orientation once bedding has been rotated to horizontal. Given that the bedding is relatively steeply dipping in the southern part of the Boso forearc basin, the attitudes of the fault planes were corrected to obtain their initial orientations when the bedding was horizontal.

Figures 3.6.1 and 3.6.2 show outcrops of faults along the Kanaya coast. These are located around the fold axis in the southern part of the A–B cross-section (Fig. 3.3) and cut the Anno Formation. The faults are branched and associated with drag folding. The fault planes are unclear and the fault planes are healed. The drag folds and healed faults suggest these faults formed shortly after sediment deposition. The restored fault attitudes are mainly E–W-striking (reverse faults) and N–S-striking (normal faults). Figure 3.6.3 shows an outcrop of a E–W-striking reverse fault (Fig. 3.6.3c) in the Anno Formation in the Shikoma River. The faults diverge and cut bedding planes. The total displacement on the fault is ca. 30 cm. Figure 3.6.4 shows faults cutting bedding in the Amatsu Formation at an outcrop in the Shikoma River. A duplex-like structure has formed, and cross-cutting relationships indicate different generations of faults (Fig. 3.6.4b). The fault planes are diffuse and are healed, which suggests these faults formed shortly after sediment deposition. The original strikes of the faults were E–W, NE–SW, and N–S.

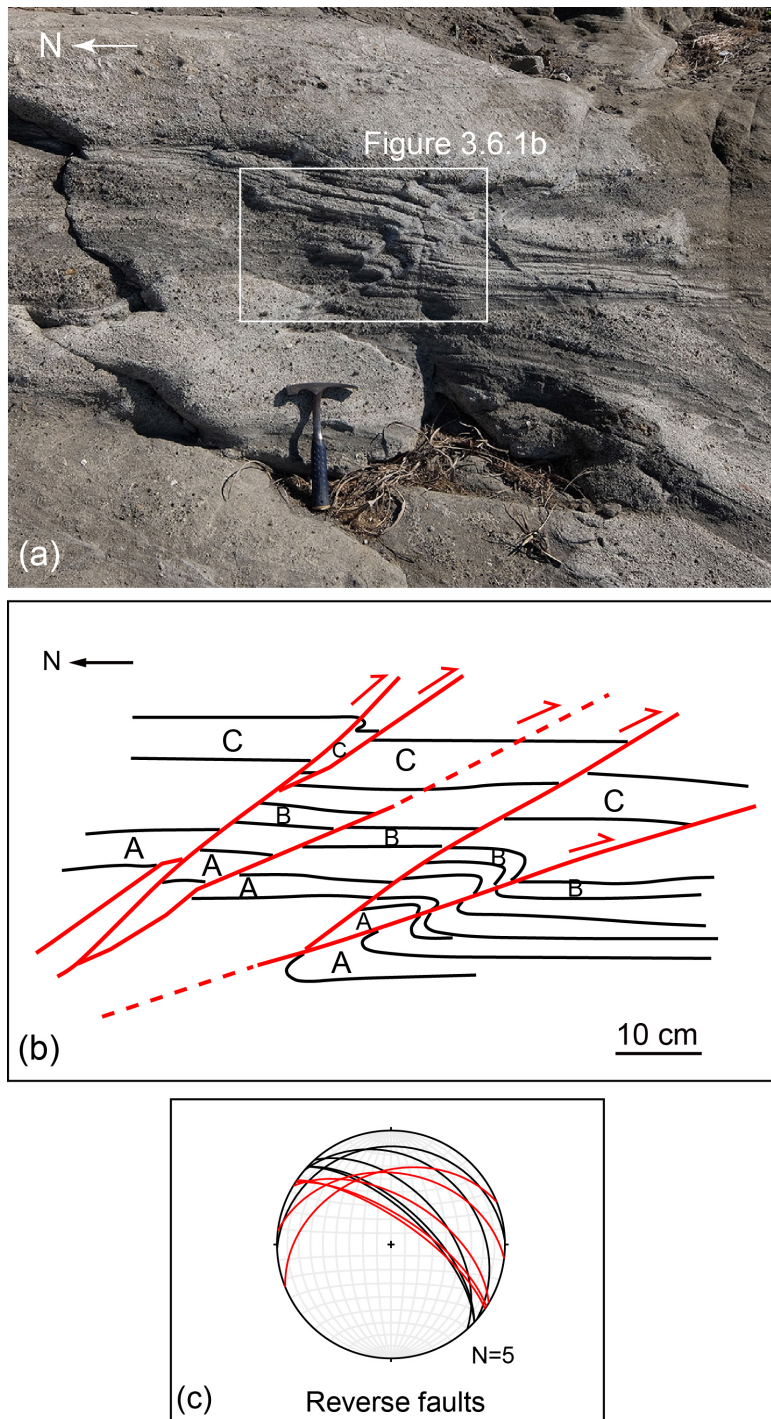


Figure 3.6.1 Outcrop of mesoscale faults in the Anno Formation along the Kanaya coast. (a) Photograph. (b) Sketch of the outcrop. Red lines are faults. A, B, and C indicate distinct layers in the outcrop. (c) Stereograph showing fault attitudes (lower hemisphere equal-area projections). Black lines show the measured fault attitudes at the site. Red lines show the modified fault attitudes after tilt correction and restoration of bedding planes to horizontal.

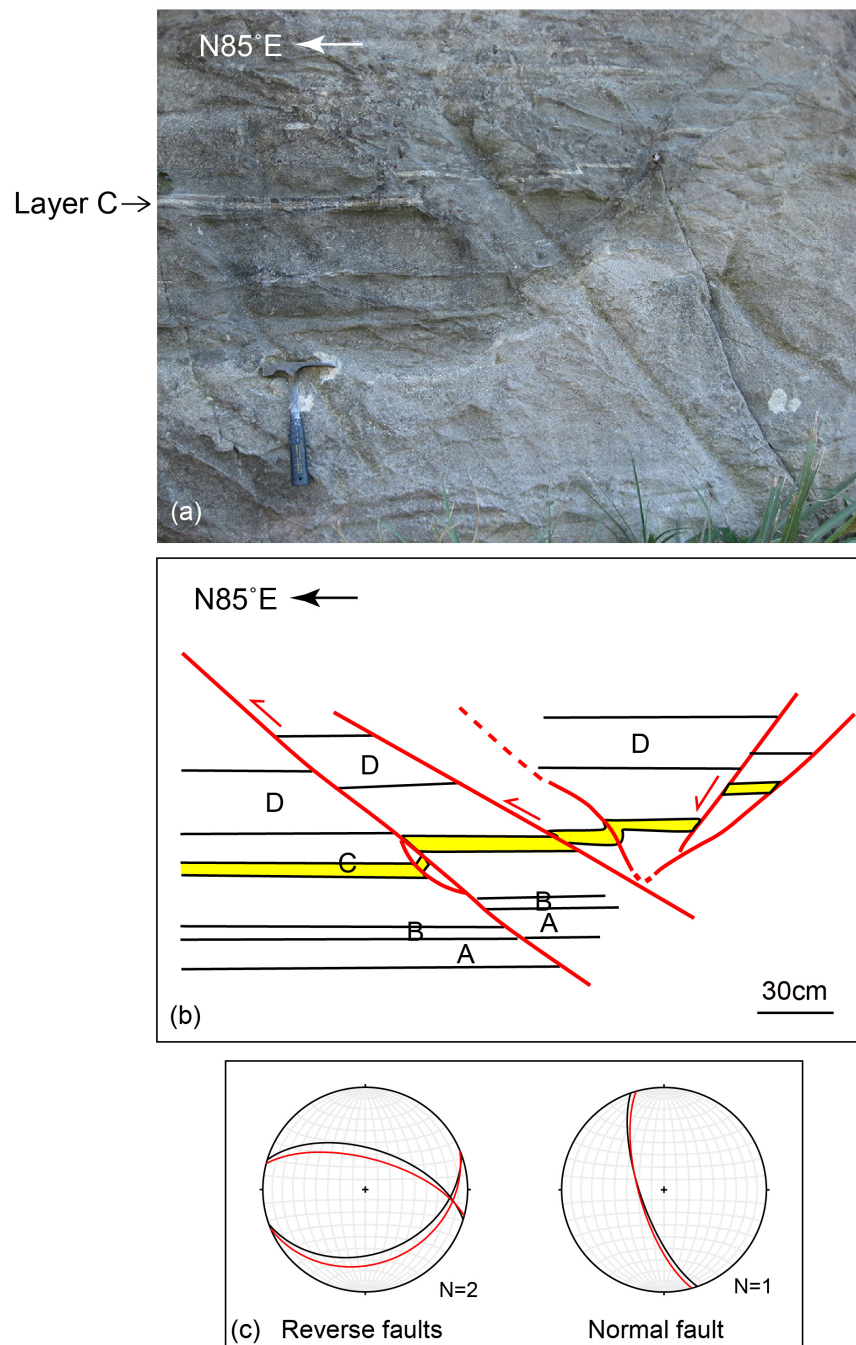


Figure 3.6.2 Outcrop of mesoscale faults in the Anno Formation along the Kanaya coast. (a) Photograph. (b) Sketch of the outcrop. Red lines are faults. A, B, C, and D indicate distinct layers in the outcrop. (c) Stereograph showing fault attitudes (lower hemisphere equal-area projections). Black lines show the measured fault attitudes at the site. Red lines show the modified fault attitudes after tilt correction and restoration of bedding planes to horizontal.

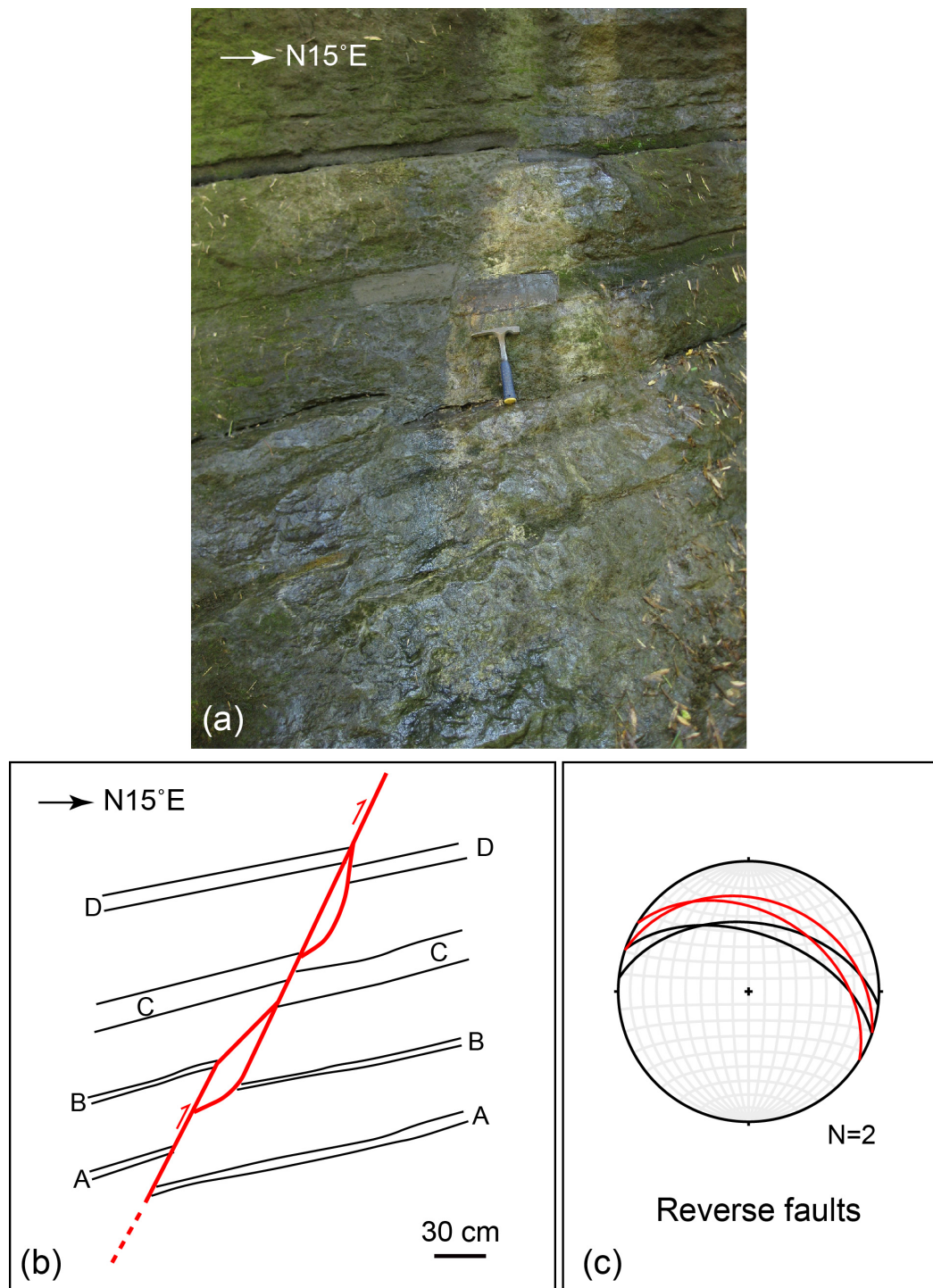


Figure 3.6.3 Outcrop of mesoscale faults in the Anno Formation along the Shikoma River. (a) Photograph. (b) Sketch of the outcrop. Red lines are the faults. A, B, C, and D indicate distinct layers in the outcrop. (c) Stereograph showing fault attitudes (lower hemisphere equal-area projections). Black lines show the measured fault attitudes at the site. Red lines show the modified fault attitudes after tilt correction and restoration of bedding planes to horizontal.

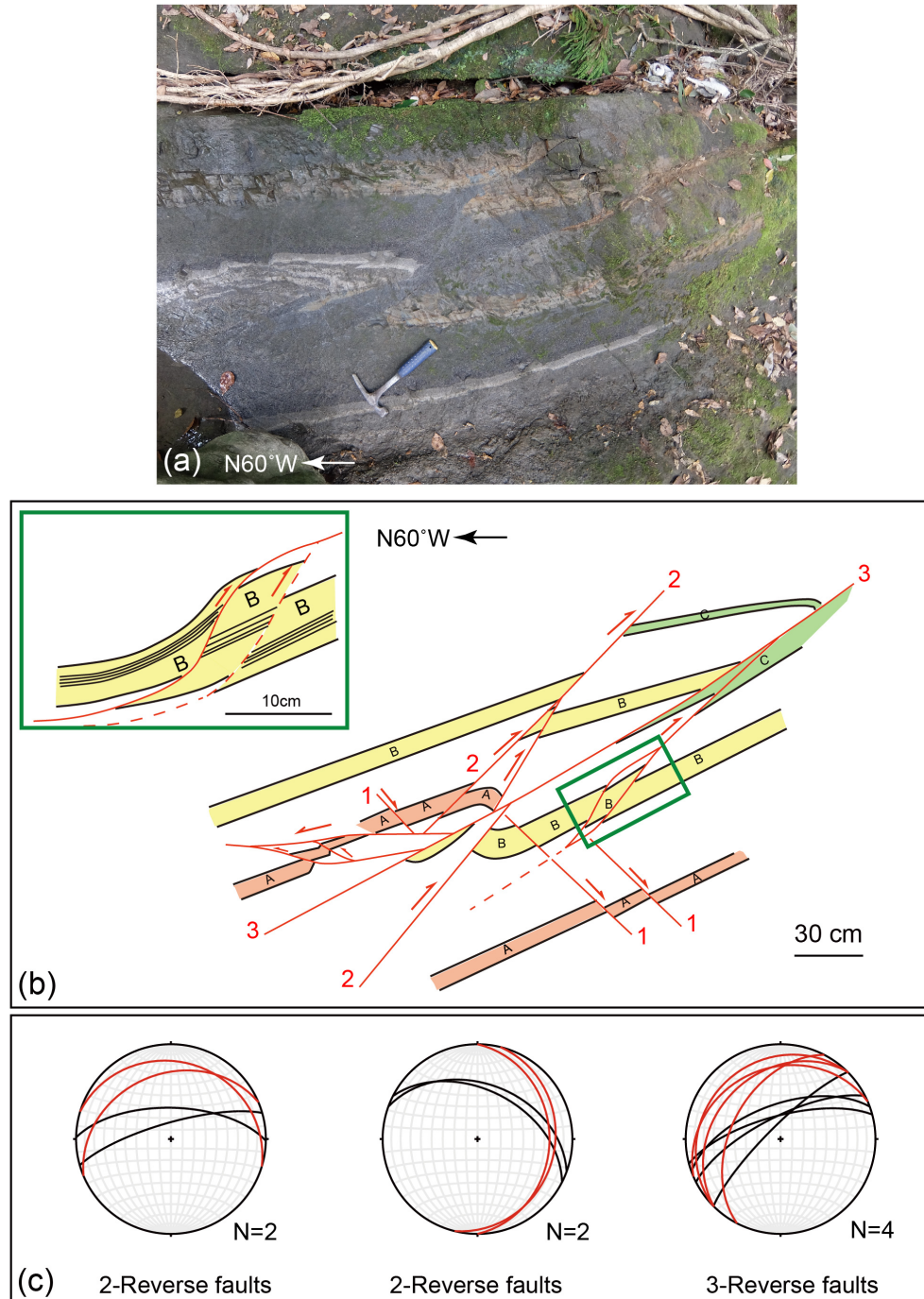


Figure 3.6.4 Outcrop of mesoscale faults in the Amatsu Formation along the Shikoma River. (a) Photograph. (b) Sketch of the outcrop. Red lines are the faults. A, B, and C indicate distinct layers in the outcrop. (c) Stereograph showing fault attitudes (lower hemisphere equal-area projections). Black lines show the measured fault attitudes at the site. Red lines show the modified fault attitudes after tilt correction and restoration of bedding planes to horizontal.

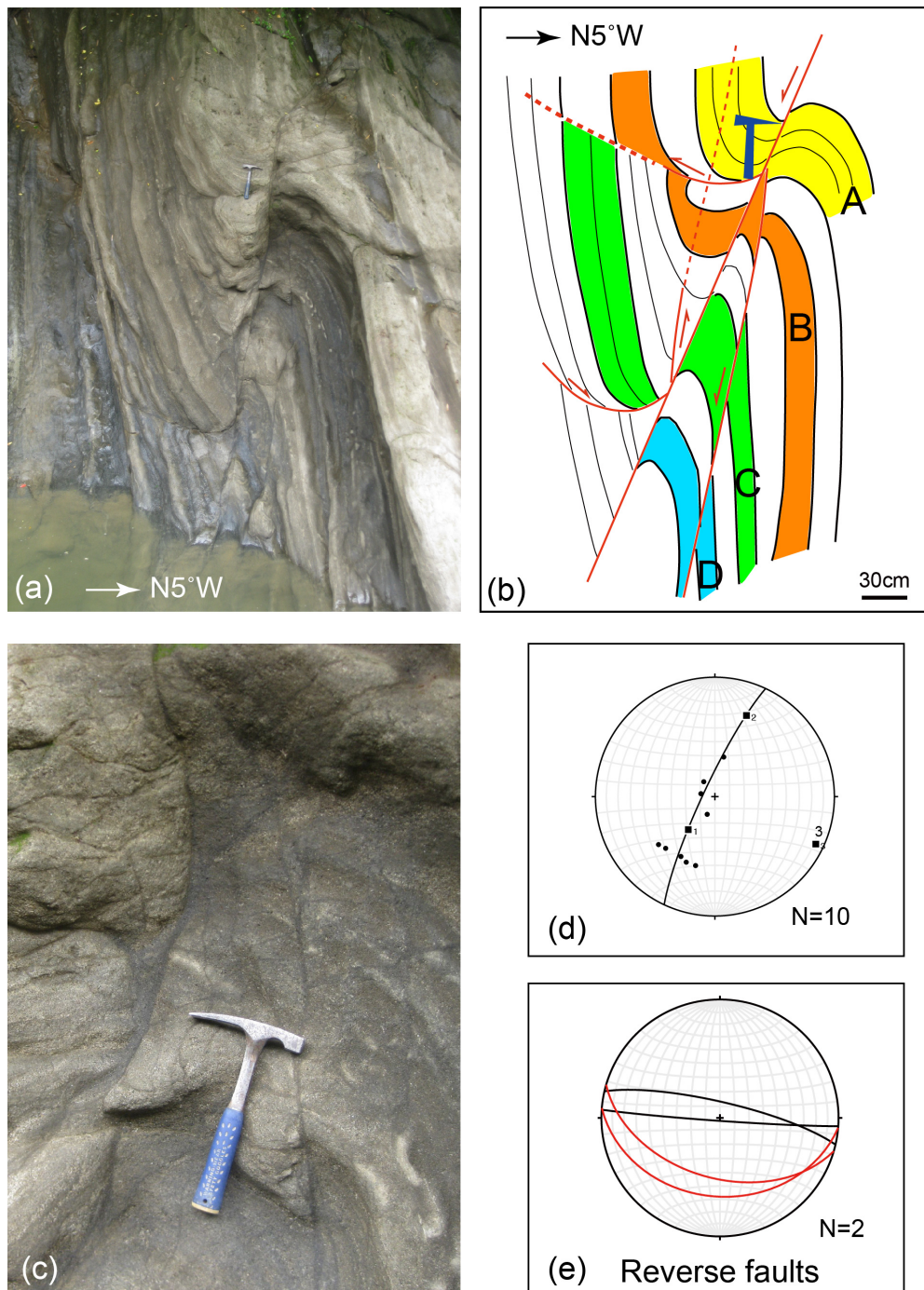


Figure 3.6.5 Outcrop of mesoscale faults in the Amatsu Formation along the Shikoma River. (a) Photograph. (b) Sketch of the outcrop. Red lines are the faults. A, B, and C indicate distinct layers in the outcrop. (c) Photograph of the faults. (d) Stereograph showing attitudes of fold plane, shows fold axis (black square) of this outcrop. (e) Stereograph showing fault attitudes (lower hemisphere equal-area projections). Black lines show the measured fault attitudes at the site. Red lines show the modified fault attitudes after tilt correction and restoration of bedding planes to horizontal.

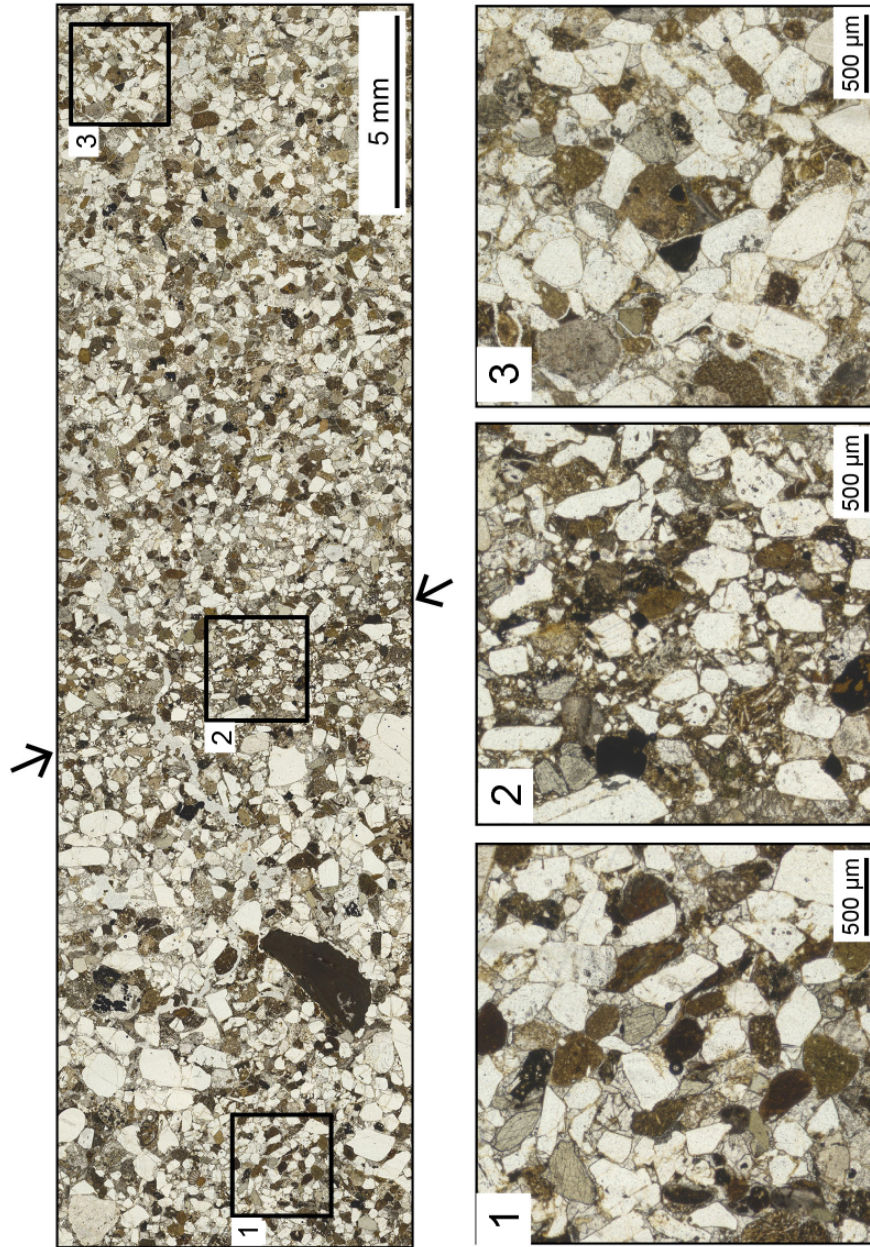


Figure 3.7 Photomicrographs of the fault. The sample was collected from the outcrop shown in Fig. 3.6.5. This sandstone consists of quartz and feldspar. "1", "3" Part of the host rock shows a grain-supported texture. "2" The fault zone exhibits loose packing and contains smaller sized rounded sand grains compared with the host rock. These structures are similar to those observed in syn-sedimentary disaggregation bands (Fossen, 2007).

Most of the normal and reverse faults in the study area show displacement and are now healed. The beds have adhered across the fault, and thus the fault plane is difficult to identify. This means that traditional kinematic indicators, such as slickenlines and slickensteps, are difficult to observe. However, given that the strikes of the faults and strikes of the fault-related fold axes are parallel to each other in outcrops of the Amatsu Formation in the Shikoma River (Fig. 3.6.5), the fault is inferred to be a dip-slip reverse fault with top-to-the-south displacement. Although the sediment grains in the host rock are grain-supported, within the fault zone the sand particles are not grain-supported (Fig. 3.7). In addition, although the fault zones do not contain cracked or broken grains, they are characterized by rounded sand grains that are smaller in size than those in the host rock. This texture is characteristic of disaggregation bands with a dilation component (Du Bernard et al., 2002; Fossen, 2007). There are many more faults and folds in the Miura Group (below the Kurotaki Unconformity) than in the Kazusa Group (above the unconformity). The healed reverse and normal fault planes were tilt-corrected by restoration of bedding planes to horizontal. This revealed two groups of reverse faults (E–W and NE–SW striking) in the Miura Group, but only NE–SW-striking faults in the Kazusa Group (Fig. 3.8). E–W- and NE–SW-striking reverse faults in the Miura Group have also been reported in the central Boso forearc basin (Ishihara & Tokuhashi, 2001; Sato et al., 2014). The normal faults have a wide range of strikes. In contrast, younger strike-slip faults are NNE–SSW- and NNW–SSE-striking.

The healed reverse faults that mostly exhibit a dip-slip sense of displacement and drag folds are similar to syn-sedimentary disaggregation bands (Fossen, 2007). This suggests that these faults formed shortly after sedimentation. The fault attitudes are different above and below the Kurotaki Unconformity (Fig. 3.8). The unconformity reflects a change in the plate convergence direction from northward to northwestward that occurred at ca. 3 Ma (Takahashi, 2006). Bedding in the Miura Group strikes E–W, whereas in the central and eastern part of the basin, the bedding in the Kazusa Group strike NE–SW. Bedding planes dip to the N and NW in the Miura and Kazusa groups,

respectively, which is parallel to the convergence direction of the Philippine Sea Plate. The dips define a monoclinial structure, with dips being steep in the southern part of the basin and gentle in the northern part.

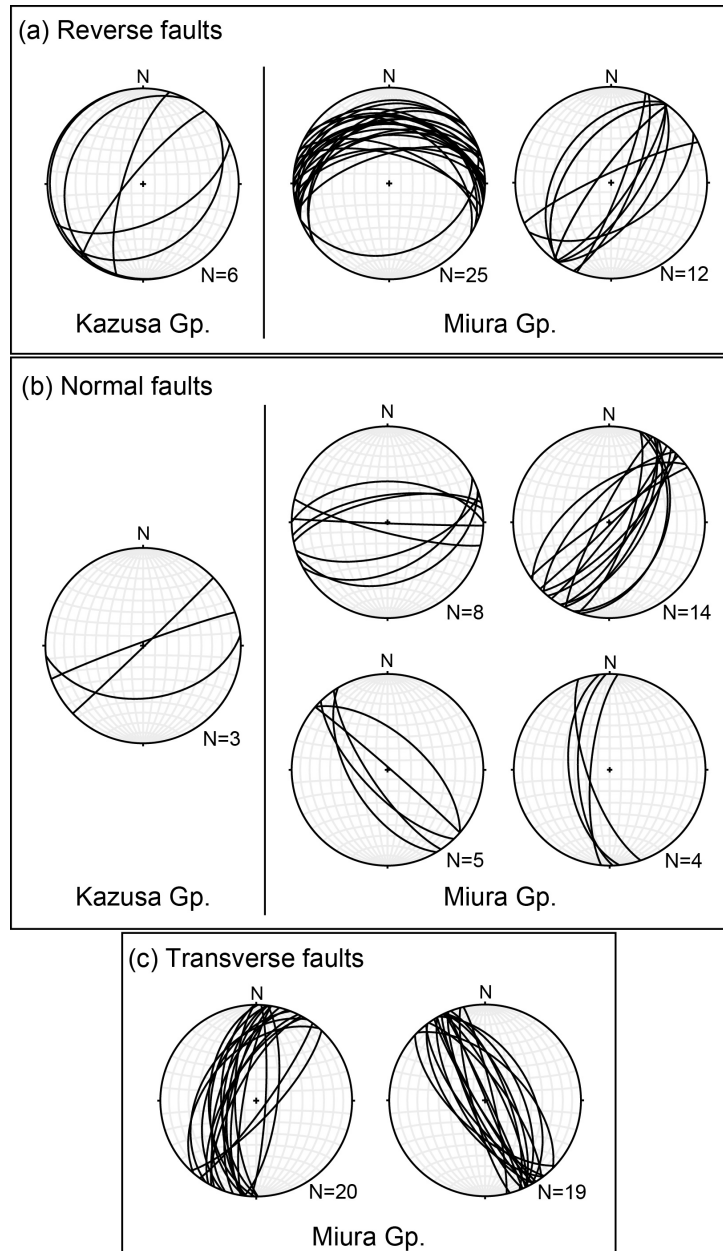


Figure 3.8 Stereographs showing fault attitudes in the study area (lower hemisphere equal-area projections). Tilt correction involved restoration of bedding planes to horizontal, which was applied to the reverse and normal faults. No correction was applied to the strike-slip faults.

3.4 Fabric of mudstones in the Boso forearc basin

Rock fabrics can reveal the depositional and deformational history of a geological unit. In this study, the fabrics of *in situ* siltstone layers were investigated using anisotropy of magnetic susceptibility (AMS) measurements. AMS data can be used to investigate deformation and paleocurrent directions (e.g., Kanamatsu et al., 1996; Kawamura et al., 2002).

The samples for the AMS and porosity measurements were collected from coherent siltstone layers using a portable, engine-powered drill with a 25 mm inner diameter coring bit. The cores were oriented using a magnetic compass and corrected for the declination of the present geomagnetic field. The cores were then sliced into 21–22-mm-thick sections. The samples were collected from the Miura Group and lower Kazusa Group (Kiwada and Otadai formations) in the western part of the basin, and from the lower Kazusa Group (Katsura, Namihana, Ohara, Kiwada, and Otadai formations) in the eastern part of the basin. AMS analysis was performed using the KLY-4 and -3 Kappabridge instruments at JAMSTEC and Kyoto University, Japan, respectively. Microfabric observations using a scanning electron microscope (SEM) were also undertaken at Kyoto University, Japan.

Given that the magnetic carrier in the siltstones from the study area is dominated by magnetite (Yamamoto, 2006), magnetic susceptibility was converted to particle length. Directions of three axes, $K_{\max}$, K_{int} , and $K_{\min}$, were plotted on lower hemisphere stereographs to investigate the sediment fabric (Fig. 3.9). Figure 3.10 shows stereoplots of the magnetic susceptibility ellipsoids, and Fig. 3.11 shows Flinn diagrams of the magnetic foliation ($F = K_{\text{int}}/K_{\min}$) versus lineation ($L = K_{\max}/K_{\text{int}}$) for each formation.

According to the stereographic plots, the orientations of $K_{\min}$ are essentially vertical and the $K_{\max}$ and K_{int} axes are horizontal, suggesting these sediments in the western part of the basin have a fabric formed by consolidation (Fig. 3.10a). Flynn diagrams for the Otadai, Kiwada, and Anno formations show a foliation fabric, whereas those of the Kiyosumi, Amatsu, and Kinone formations have a random fabric (Fig. 3.11a). Both the

Miura and Kazusa groups generally have a consolidation fabric. However, this fabric in the Miura Group is weaker than that in the Kazusa Group.

In the eastern part of the basin, the siltstone layers in the Kazusa Group have a consolidated fabric, rather than a random or shear-induced fabric, as $K_{\min}$ is vertically oriented and $K_{\max}$ and K_{int} are parallel to the bedding planes (Fig. 3.10b). The Flinn diagrams also show a consolidation fabric characterized by low L (<1.02) and high F (1.02–1.08; Fig. 3.11b). All the samples from the lower Kazusa Group have a similar fabric. The magnetic ellipsoids of samples taken from level Kd38 exhibit a similar orientation to the samples from the lower section. However, several points on the Kd38 stereograph are scattered, and the lineation is stronger than in the lower Kazusa Group. This variation is due to the presence of coarse volcanoclastic material within the samples taken from Kd38.

A preferred orientation of clay minerals parallel to bedding was identified by SEM observations (Fig. 3.12). In some cases, clay aggregates are preserved in the porous fabric. The clay aggregates bind sediment particles together and maintain the pore space, representing cementation by intergranular bonding. Given that such clay aggregates characterize cemented marine sediments (Ujiie et al., 2003), sediment of the lower Kazusa Group in the eastern part of the basin is inferred to be weakly cemented. The intergranular cementation is particularly developed in the Ohara and Katsura formations.

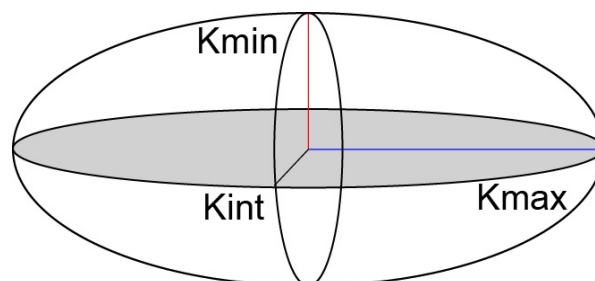
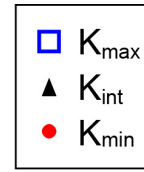
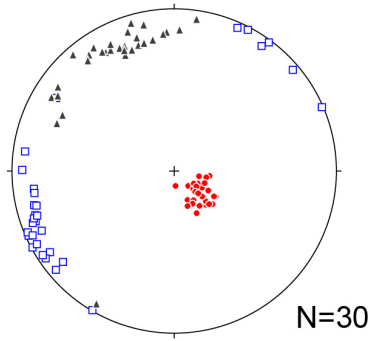


Figure 3.9 Schematic figure showing the relationship between an AMS measurement result and the directions of the maximum, intermediate, and minimum axes of a particle.

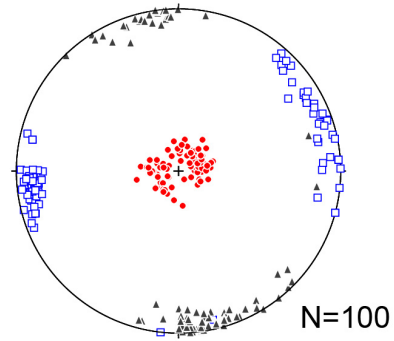
(a)



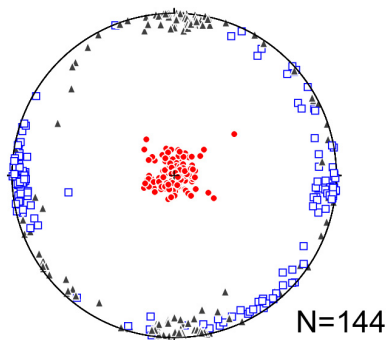
Otadai Fm.



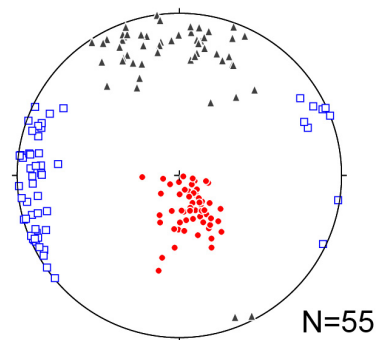
Kiwada Fm.



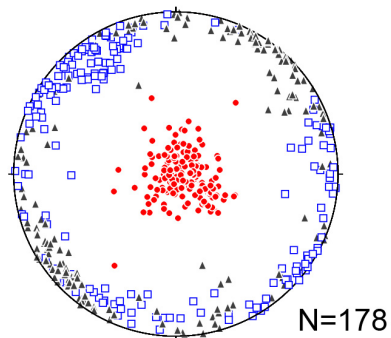
Anno Fm.



Kiyosumi Fm.



Amatsu Fm.



Kinone Fm.

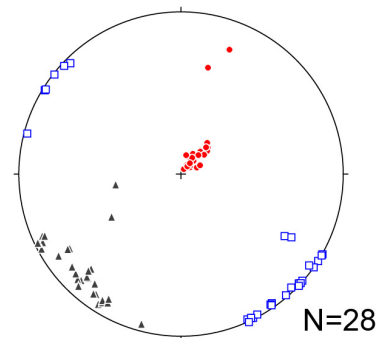


Figure 3.10 Stereoplots of AMS results (lower hemisphere) for the Boso forearc basin. Since the orientation of $K_{\min}$ is basically vertical and $K_{\max}$ and K_{int} axes show a girdle distribution, these sediments have a fabric formed by consolidation. (a) Western part of the basin. (b) Eastern part of the basin.

(b)

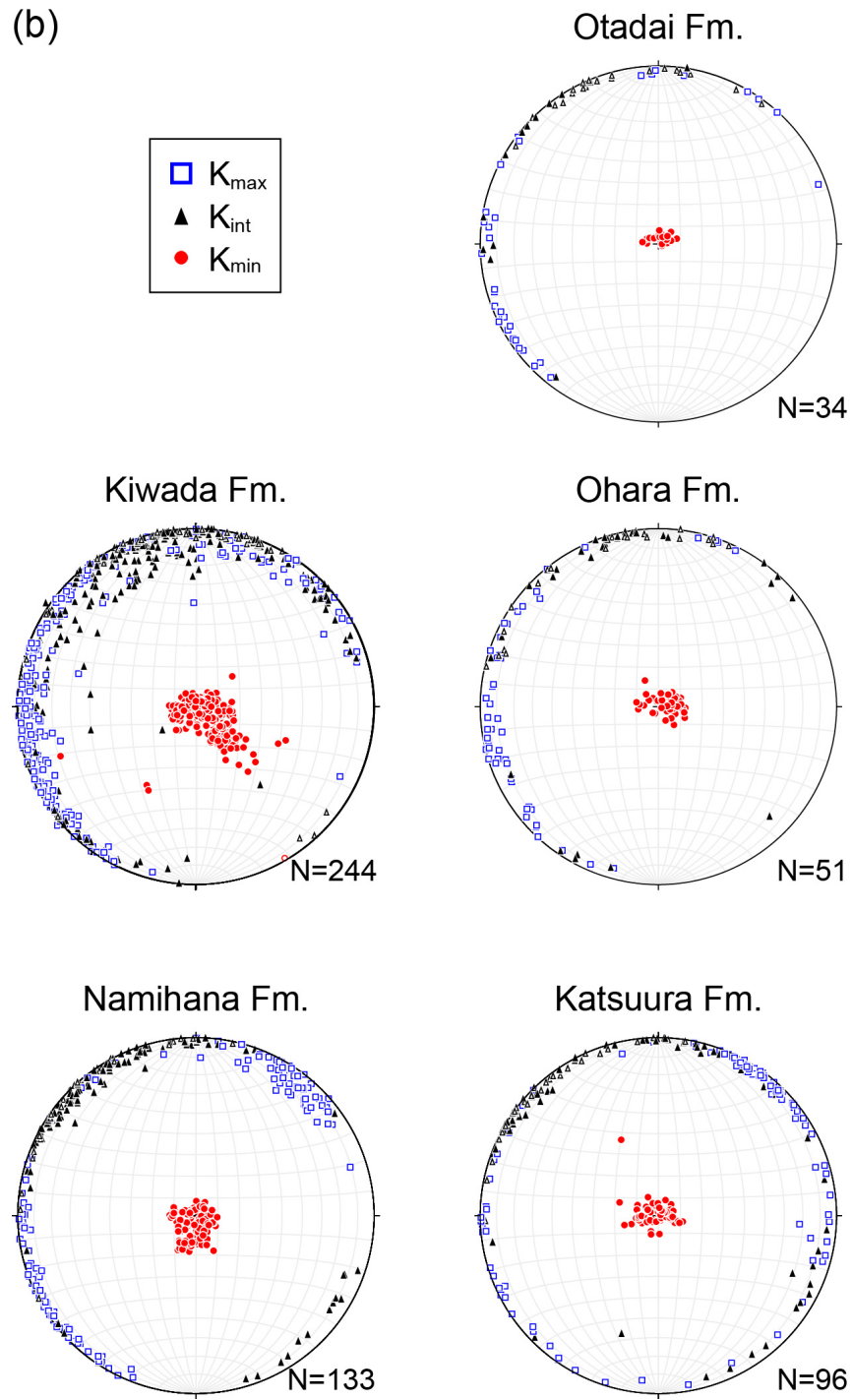


Figure 3.10 Continued.

(a)

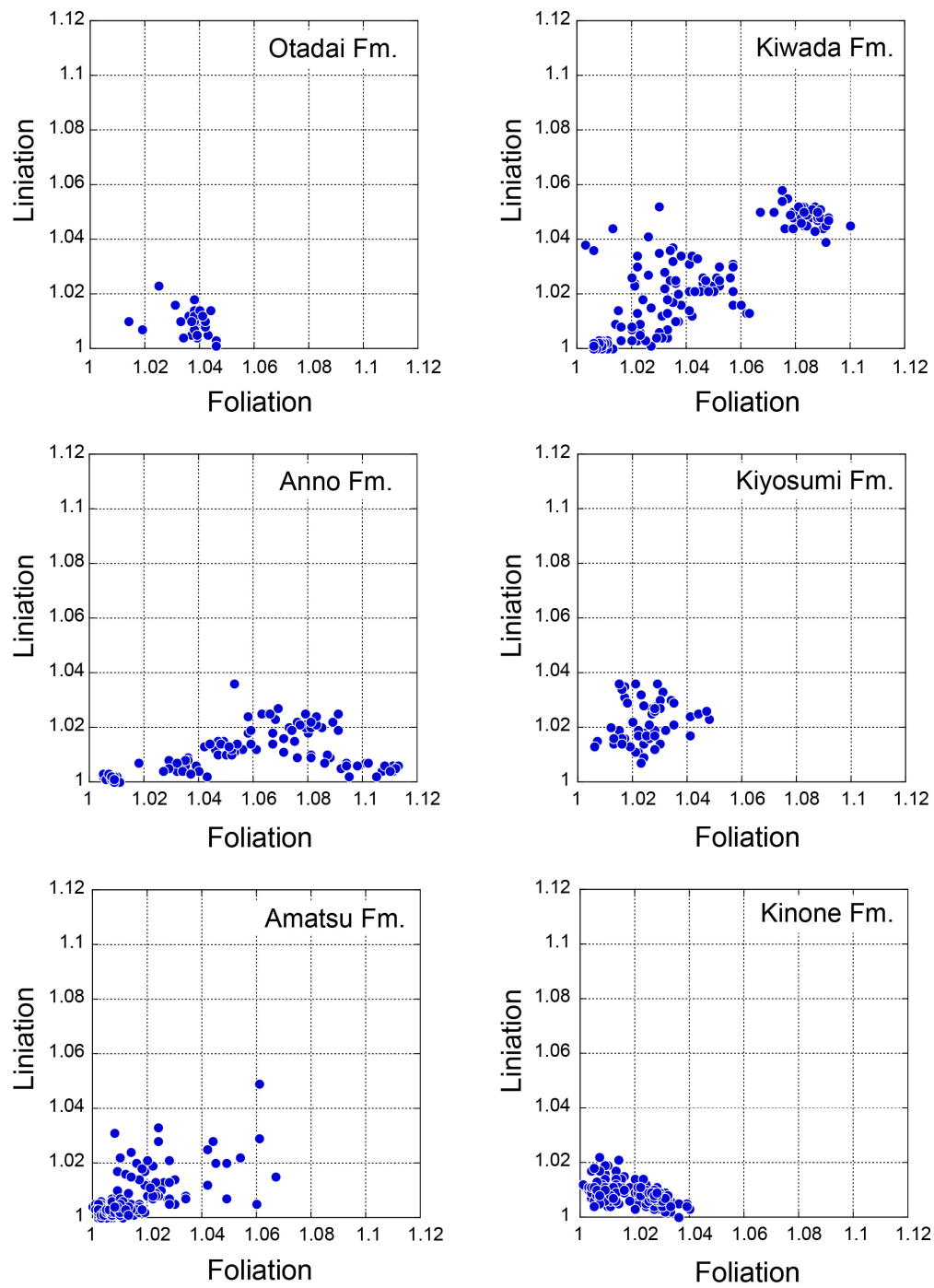


Figure 3.11 Flynn diagrams of the magnetic susceptibility ellipsoids. (a) Western part of the basin. (b) Eastern part of the basin.

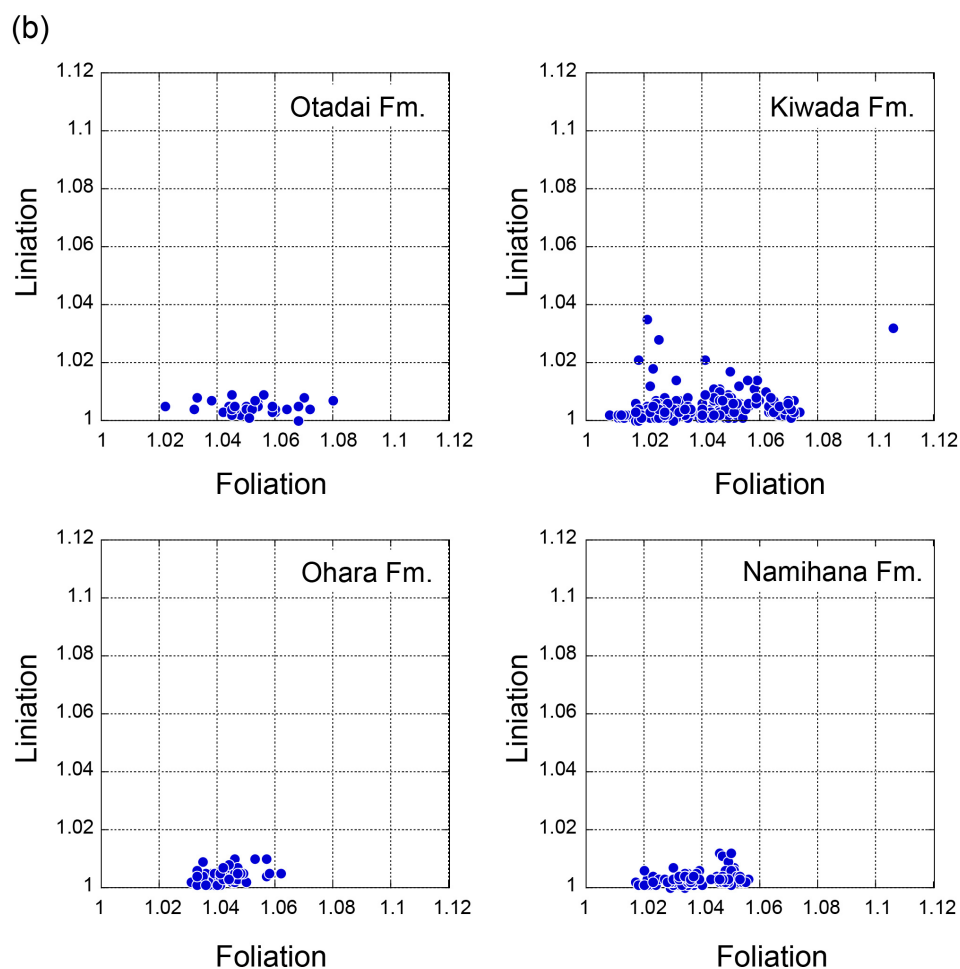


Figure 3.11 Continued.

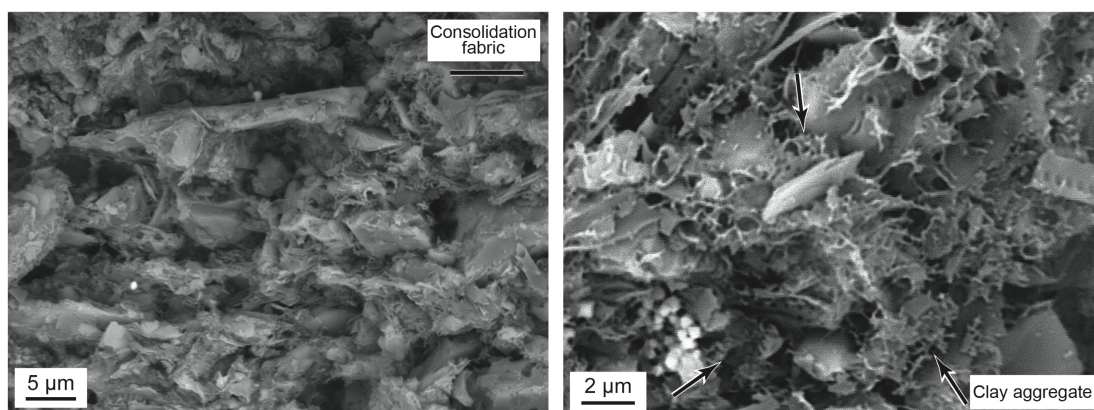


Figure 3.12 SEM images of siltstone samples from the Ohara Formation. Consolidation structures are characterized by a preferred orientation of clay minerals parallel to the bedding plane (left), and clay aggregates in pore spaces indicate cementation by intergranular bonding (right).

3.5 Conclusions

The bedding orientations in the study area are controlled by E–W-striking folds. Several fold axes were identified by geological mapping and from the N–S-striking cross-section. Relatively intense deformation, such as faulting and folding, is observed in the southern part of the forearc basin, close to the trench–slope break (Hayama–Mineoka Tectonic Belt), whereas deformation is weaker in the northern part of the basin. In particular, bedding planes generally dip steeply in the southern part of the basin, whereas they are gentler in the north. The Kurotaki Unconformity occurs between the Anno and Kurotaki formations, and is characterized by a basal conglomerate of the Kurotaki Formation. The mesoscale faults in the Boso forearc basin are healed, which indicates they formed soon after sediment deposition. Two groups of reverse faults, E–W- and NE–SW-striking, were identified in the Miura Group, but only a NE–SW-striking population is present in the Kazusa Group. The normal faults have a wide range of orientations. In contrast, younger strike-slip faults strike NNE–SSW and NNW–SSE. Given these features, it is suggested that the Boso forearc basin formed under compressive stress. AMS analysis suggests that the fabrics of sedimentary rocks in the Boso forearc basin were controlled mainly by consolidation, whereas in the Kiyosumi, Amatsu, and Kinone formations in the Miura Group, the sedimentary rocks have a random fabric related to compression.

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Chapter 4

Paleo-temperatures of the Boso forearc basin

4.1 Introduction

Basin temperatures are generally controlled by the geothermal gradient and increase with depth. However, the geothermal gradient changes with the geological setting and through time, and advective heating can also change basin temperatures. Therefore, by comparing the geothermal gradient and paleo-temperatures, thermal anomalies related to basin evolution can be identified. The present-day geothermal gradient of the Boso Peninsula measured in a well at Kimitsu is ca. 20°C/km (Sakai et al., 2011). However, the paleo-temperatures of the Boso forearc basin were complex, because the Miura Group is folded and an ophiolite complex is present in the southern part of the basin (e.g., Ogawa & Taniguchi, 1987). To investigate the thermal history of the Boso forearc basin, I obtained paleo-temperature data by vitrinite reflectance analysis.

The mechanism of organic matter diagenesis can be attributed to chemical reactions involving the removal of H, C, and O in the form of CH_n , H_2O , and CO_2 . Vitrinite is an organic material derived from buried wood, and its reflectance changes with the degree of thermal maturation (Fig. 4.1). The resulting increase in vitrinite reflectance depends on the maximum temperature and duration of heating. In general, vitrinite reflectance is not affected by retrograde reactions (except weathering), and therefore records the peak temperature (e.g., Burnham and Sweeney, 1989). Vitrinite reflectance is widely used in integrated basin analysis and oil–gas exploration (Hunt, 1979; Waples, 1981; Tissot et al., 1987), and has also been used to estimate the maximum temperature of sedimentary rocks in geothermal areas (Barker, 1989) and frictional heating on seismogenic faults (Sakaguchi et al., 2011). The maximum paleo-temperature of the Boso forearc

sediments and the paleo-geothermal structure were estimated based on vitrinite reflectance data in this study.

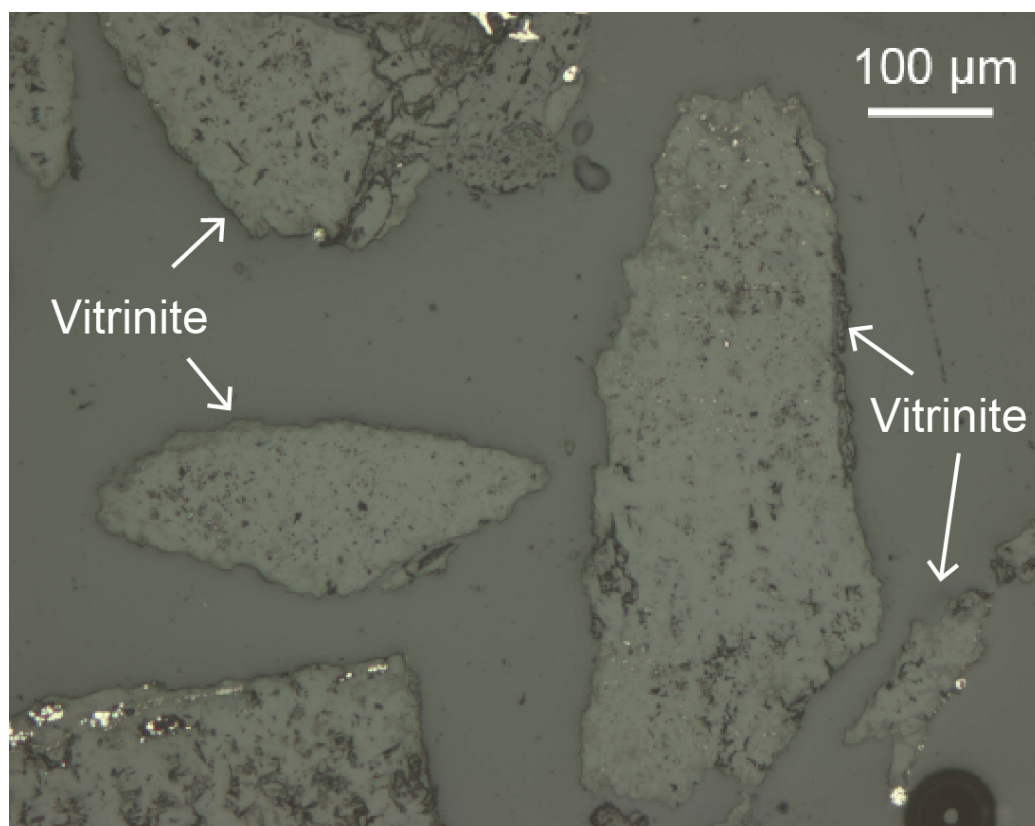


Figure 4.1 Photomicrograph of vitrinite grains under the microscope.

4.2 Methodology

A total of 23 samples from the forearc basin (16 from the Miura Group and 7 from the Kazusa Group) were analyzed. BosV-66, -67, -68, -69, -72, -73, and -85 were taken from the western part of the basin; BosV-77, -78, -79, -80Kn, -86, -87, -88, -89, -90, and -92 were taken from the central part of the basin; and BosV-80, -81, -82, -83, -93, and -94 were taken from the eastern part of the basin (Fig. 4.2). Samples for vitrinite reflectance analysis were collected from turbiditic sand layers. Sandstone or siltstone samples containing carbonaceous grains (sample volume = 80–160 cm³) were collected using a chisel. The samples were crushed in a mortar and sieved to obtain the 75–180 and 180–780 μm fractions under running water. Carbonaceous grains were removed by heavy liquid separation using a sodium polytungstate (SPT) solution. Separated carbonaceous grains (Fig. 4.3a) were mounted in epoxy resin and ground using progressively finer sandpaper (up to #2400) and polished using a 0.06 μm alumina suspension (Fig. 4.3b). Vitrinite reflectance measurements were performed using a dual illumination system (Sakaguchi & Mukoyoshi, 2012) at Japan Agency for Marine-Earth

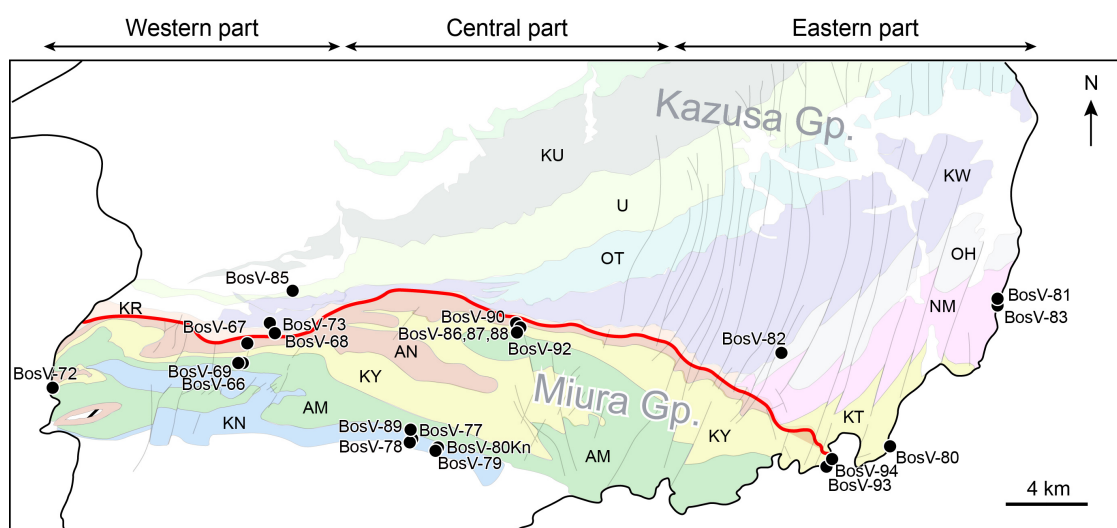


Figure 4.2 Geological map showing the sampling points for the vitrinite reflectance measurements.

Science and Technology (Figs 4.3c and 4.4). This approach was used previously to estimate the frictional heating on the megasplay fault in Nankai Trough seismogenic zone (Sakaguchi et al., 2011). Using this system, a homogeneous beam illuminates a wide field of view to allow identification of vitrinite, and a concentrated beam of 1.6 μm in diameter is used to measure reflectance. The vitrinite reflectance is referred to as R_o , $R_{\min}$, or $R_{\max}$, corresponding to mean, minimum, and maximum values, respectively. The measurements were conducted in immersion oil. Up to 100 grains were measured per sample, depending on whether a sufficient number of polished grains were exposed on the sample surface. The measurement was performed one point per one particle. R_o was calculated from a frequency histogram (Fig. 4.5). I measured the mean random reflectance.

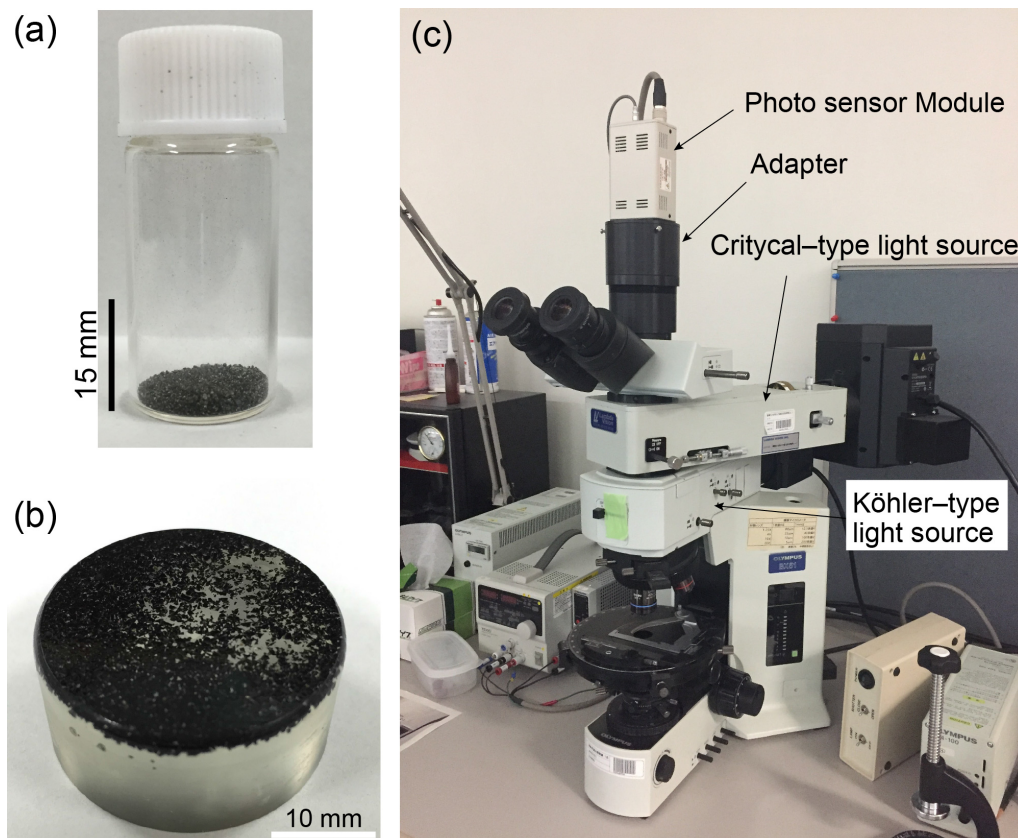


Figure 4.3 (a) Organic material picked from the samples. (b) Polished organic material mounted on resin. (c) The dual illumination system used for measurement of vitrinite reflectance.

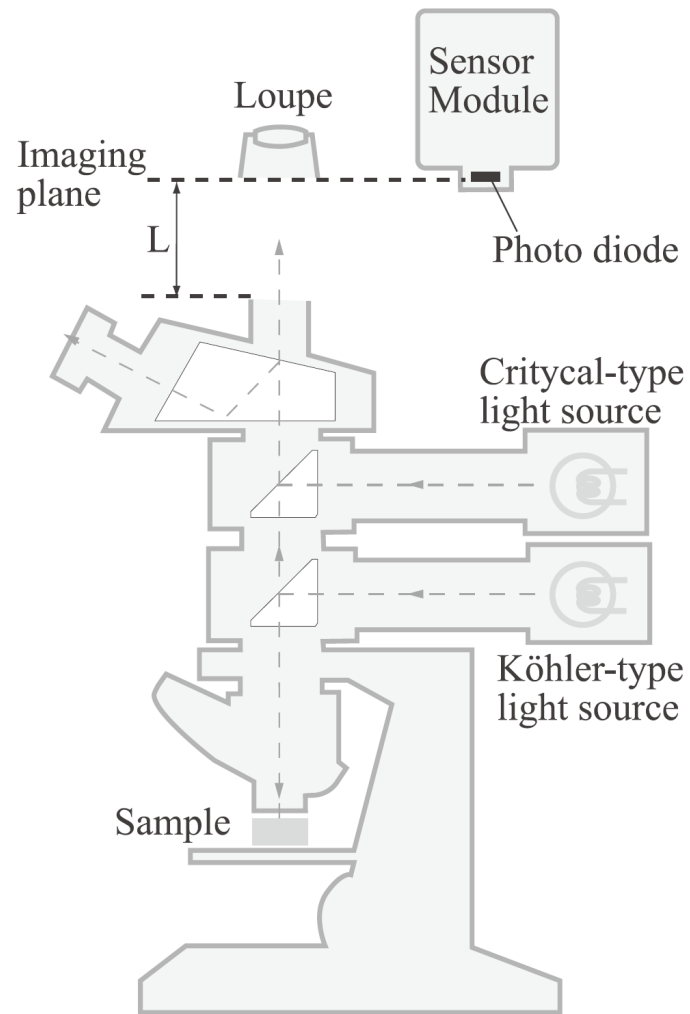


Figure 4.4 Schematic representation of the structure of the apparatus. The photodiode sensor measures reflectance at the imaging plane. The distance between the focal plane and body of the microscope (L) was measured using a loupe, and is the baseline length of the photodiode sensor (Sakaguchi & Mukoyoshi, 2012).

I estimated the maximum paleo-temperatures from R_o . Sweeney & Burnham (1990) proposed a representative model for calculating the maximum paleo-temperature from R_o using the H/C and O/C atomic ratios of coal. The heating time is defined as the total time that a sample remains at the maximum temperature (Sekiguchi & Hirai, 1980). Given this time is typically 1–10 Myr at active margins (Laughland & Underwood, 1993; Ohmori et al., 1997), two values were calculated to provide lower and upper limits corresponding to 1 and 10 Myr of heating, respectively. The maximum paleo-temperatures was calculated from the follow equations formulated by Mukoyoshi et al. (2006) according to the Sweeney & Burnham (1990) model:

$$T_1 = 174 + [93 \times \ln(R_o)] \quad (4-1)$$

$$T_{10} = 158 + [90 \times \ln(R_o)] \quad (4-2)$$

where T_1 and T_{10} are the (paleo)temperatures (°C) for a 1 and 10 Myr heating duration, respectively, and R_o is the mean vitrinite reflectance (%).

4.3 Results

4.3.1 Western part of the basin

The frequency histograms for vitrinite reflectance for the samples of BosV-66, BosV-67, BosV-69, BosV-72, BosV-73, and BosV-85 show a single mode, whereas the histogram for BosV-68 has widely scattered R_o values between 0.1% and 0.5% (Fig. 4.5.1). Several grains in the Kazusa Group (BosV-68, BosV-73, and BosV-85) have high R_o values ($>0.4\%$) and are considered to represent reworked vitrinite that experienced a greater degree of thermal maturity in the past (Bostick, 1979; Dow and O'Connor, 1982). Such data were therefore excluded and R_o was recalculated (R_{oII} in Table 4.1). In the Miura Group, sample BosV69 contains a high- R_o particle. R_o values of $>0.6\%$ are considered to be reworked vitrinite in the Miura Group and were excluded in the calculation of R_{oII} . In the western part of the basin, vitrinite reflectance values (R_{oII}) of the Kinone, Kiyosumi, Anno, Kiwada, Otadai, and Umegase formations are 0.38%–0.41%, 0.44%, 0.40%, 0.25%, 0.23%, and 0.31%, respectively (Table 4.1). Therefore, the vitrinite reflectance differs between the Miura and Kazusa groups.

Previous studies have suggested that scattered R_o values reflect heterogeneity of the thermal maturity of vitrinite in a sample, and are unreliable and do not reflect depth ($R_o < 0.5\%$; Hirai, 1979) or reasonable maximum heating temperatures. However, as the R_o values in the Boso forearc basin are relatively constant, except for sample BosV-68, the level of thermal maturity appears to be similar and the R_o values can be regarded as being reliable. Samples BosV-66 and BosV-69 were obtained from the Kinone Formation that was deposited at 15–13 Ma, and the equations for both 1 and 10 Myr heating durations were applied to these samples to estimate the maximum paleo-temperatures. In contrast, only the equation for a heating duration of 1 Myr was used for the other samples, because these samples were collected from younger strata (<5 Ma). The R_{oII} values suggest the predominant maximum paleo-temperatures were 71–78°C for the Kinone Formation, 98°C for the Kiyosumi Formation, and 89°C for the Anno

Formation. In contrast, in the lower part of the Kazusa Group, above the Kurotaki Unconformity, the maximum paleo-temperature was $<45^{\circ}\text{C}$ for the Kiwada and Otadai formations and ca. 68°C for the Umegase Formation.

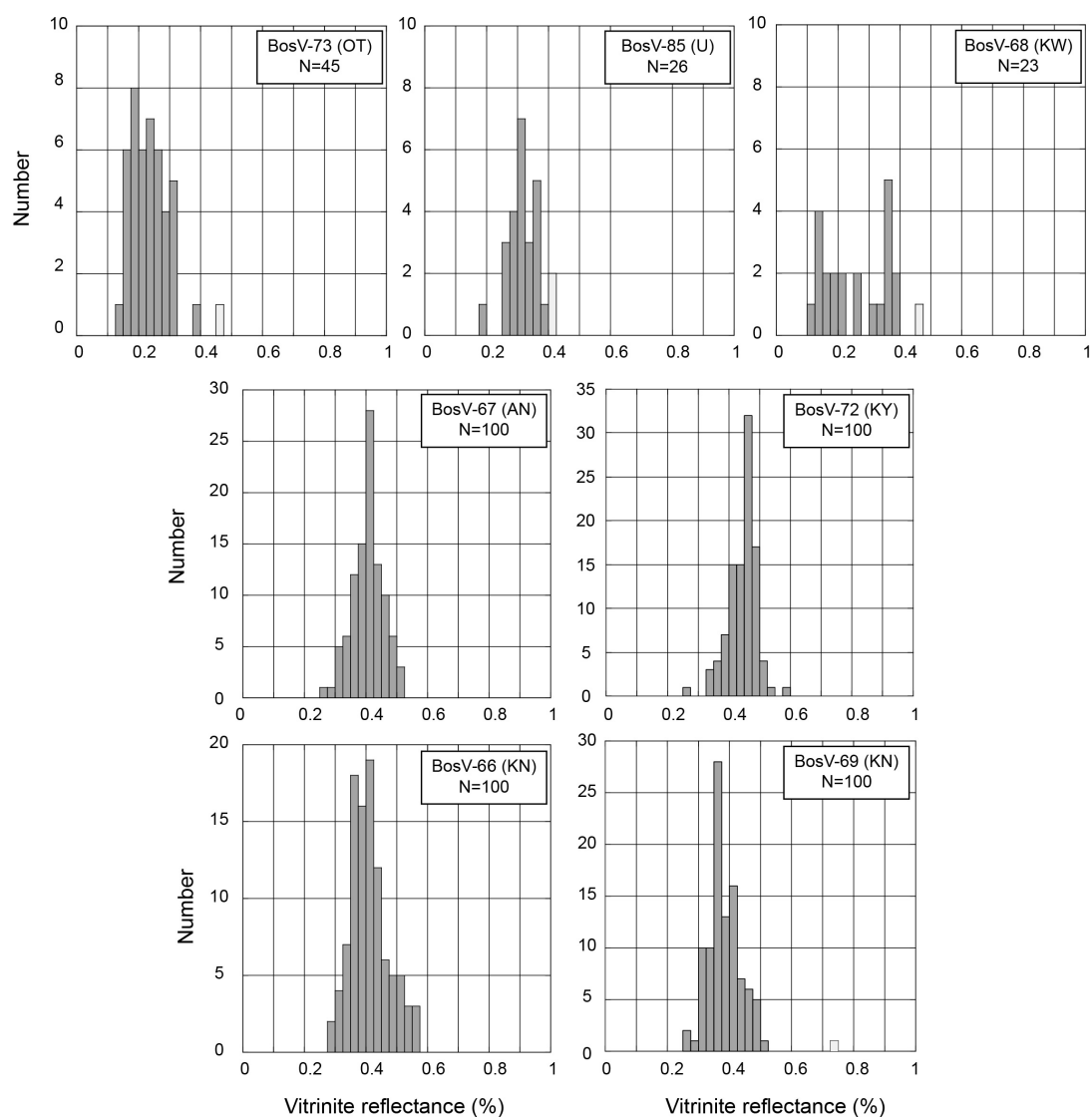


Figure 4.5.1 Histograms of the vitrinite reflectance in the western part of the basin. Samples from the Miura and Kazusa groups. Sampling sites are marked in Fig. 4.2. The number “N” indicates the quantity of measured particles for each sample. White histogram bars show data for reworked particles removed in the calculation of $R_0\text{II}$.

4.3.2 Central part of the basin

Frequency histograms of the vitrinite reflectance results for the central part of the basin have a single mode, except for sample BosV-77 (Fig. 4.5.2). Several grains in samples BosV-87, BosV-78, and BosV-80Kn have high R_o values ($>0.65\%$) and are considered to represent reworked vitrinite that experienced a higher thermal maturity in the past (Bostick, 1979; Dow & O'Connor, 1982). Such data were excluded, and R_o was recalculated as R_{oII} (Table 4.1). The histogram for BosV-77 shows a bimodal distribution, which indicates mixing of reworked vitrinite particles. In such a case, the higher peak is considered to be the reworked particles. Given that vitrinite reflectance increases with temperature but does not retrogress, the vitrinite reflectance indicates the maximum paleo-temperature. If the lower vitrinite reflectance values corresponded to reworked particles, then the lower peak would be overprinted during increased thermal maturation because of the higher post-deposition temperature.

In the central part of the basin, vitrinite reflectance values for the Anno, Kiyosumi, Amatsu, and Kinone formations are 0.34%–0.40%, 0.39%–0.42%, 0.33%, and 0.33%–0.54%, respectively. Assuming a heating duration of 1 Myr, the maximum paleo-temperatures calculated from R_{oII} values are 74–89°C for the Anno Formation, 86–93°C for the Kiyosumi Formation, 71°C for the Amatsu Formation, and 71–117°C for the Kinone Formation in the central part of the basin. Given the age of the Kinone Formation is 15–10 Ma, I assumed two heating durations of 1 and 10 Myr (Table 4.1), which resulted in paleo-temperatures for this formation of 58–103°C after 10 Myr of heating.

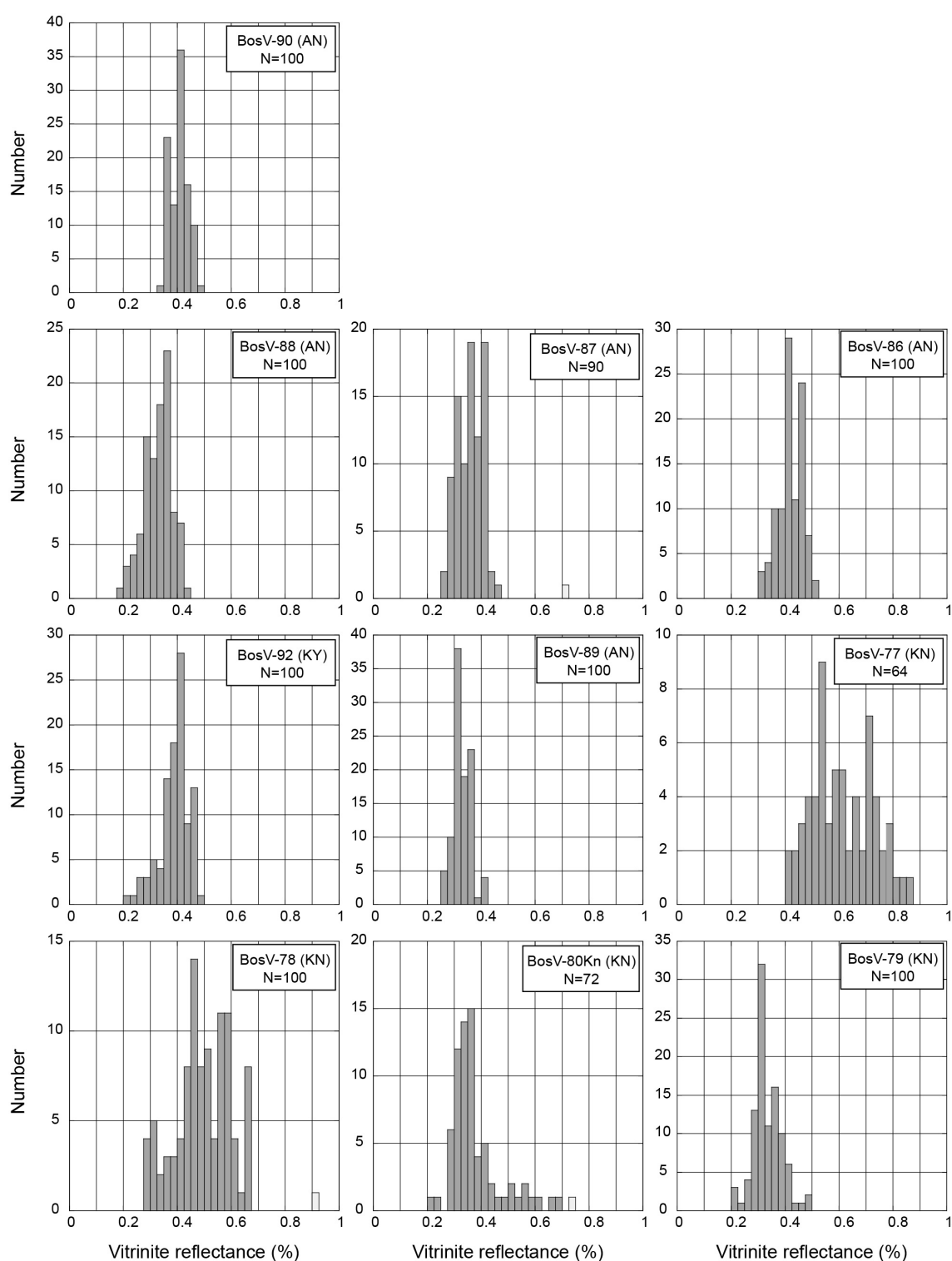


Figure 4.5.2 Histograms of the vitrinite reflectance in the central part of the basin. Samples from the Miura Group. Sampling sites are marked in Fig. 4.2. The number “N” indicates the quantity of measured particles for each sample. White histogram bars show data for reworked particles removed in the calculation of R_0II . (b) (c) The eastern part of the basin. Samples from the Kazusa and Miura groups.

4.3.3 Eastern part of the basin

Considering the eastern part of the basin, frequency histograms of vitrinite reflectance results show a single mode (Fig. 4.5.3). Therefore, R_o was calculated using all the data and R_o and R_{oII} values are the same for these samples. Vitrinite reflectance values for the Ohara, Namihana, Katsuura, and Kiyosumi formations are 0.36%, 0.34%–0.37%, 0.41%, and 0.35%–0.41%, respectively (Table 4.1). Considering the age of these formations, the heating duration was assumed to be 1 Myr. The calculated paleo-temperatures are 79°C for the Ohara Formation, 74–82°C for the Namihana Formation, 91°C for the Katsuura Formation, and 76–91°C for the Kiyosumi Formation (Table 4.1).

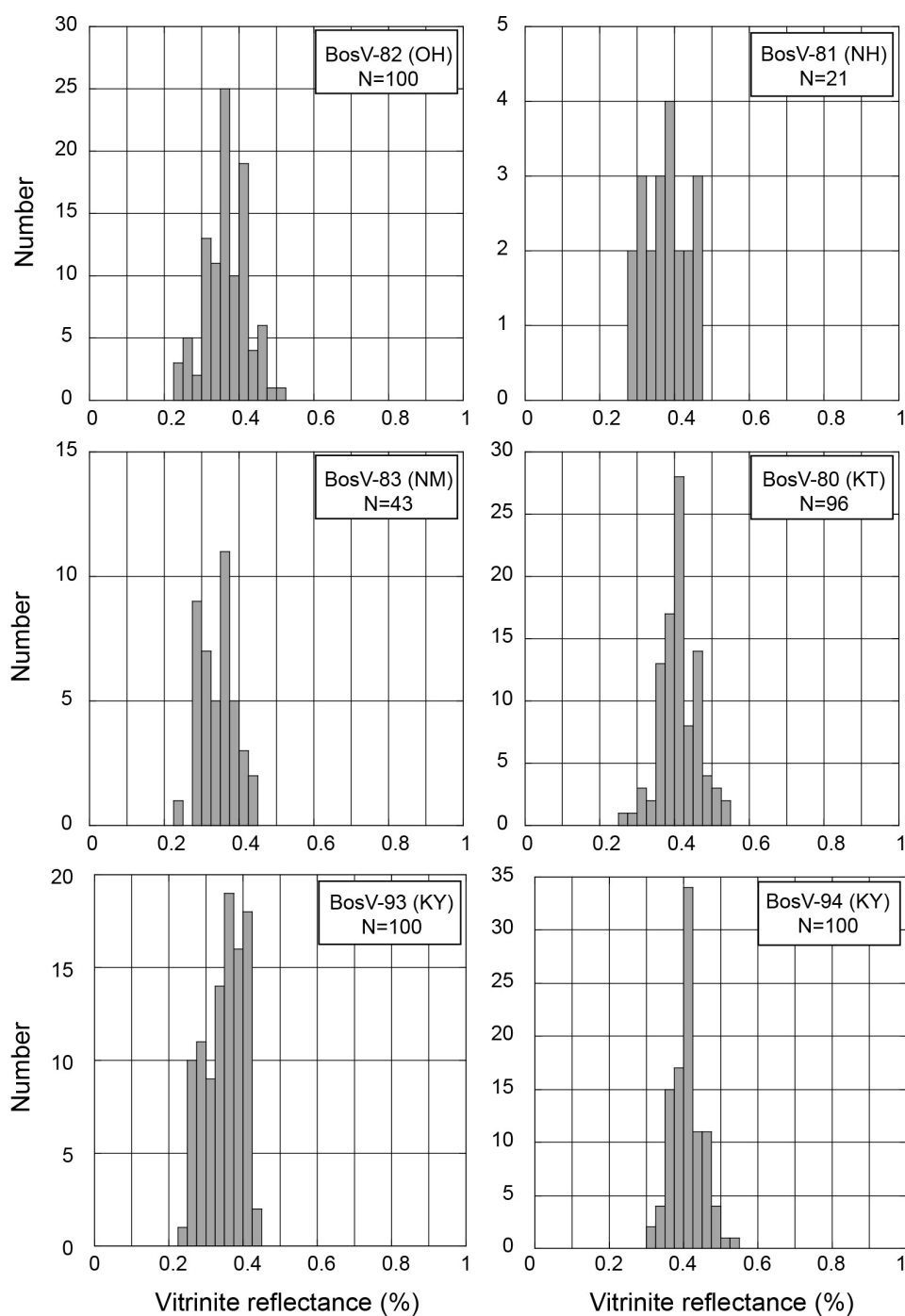


Figure 4.5.3 Histograms of the vitrine reflectance in the eastern part of the basin. Samples from the Kazusa and Miura groups. Sampling sites are marked in Fig. 4.2. The number “N” indicates the quantity of measured particles for each sample. White histogram bars show data for reworked particles removed in the calculation of R_oII .

Table 4.1. Vitrinite reflectance, paleo-temperature, and lithofacies data. R_0 II indicates R_0 data after removal of reworked vitrinite data. Paleo-temperatures were estimated using the equations of Mukoyoshi et al. (2006).

Sample ID	Sampling area	Group	Formation	Key bed	A number of measured particles for R_0	R_v , %	A number of measured particles R_0 II	R_0 II, %	Paleotemperature, °C		Host rock
									10 Myr	1 Myr	
BosV-85	West	Kazusa	Umegase		26	0.32 ± 0.05	24	0.31 ± 0.05	-	68 ± 15	sandstone
BosV-73	West	Kazusa	Otadai	Kd8	45	0.23 ± 0.07	44	0.23 ± 0.06	-	37 ± 25	mudstone
BosV-68	West	Kazusa	Kiwada	Kd38	23	0.26 ± 0.11	22	0.25 ± 0.10	-	45 ± 39	sandstone
BosV-67	West	Miura	Anno	An70-80	100	0.40 ± 0.05	100	0.40 ± 0.05	-	89 ± 12	mudstone
BosV-72	West	Miura	Kiyosumi	Ky21	100	0.44 ± 0.05	100	0.44 ± 0.05	-	98 ± 11	sandstone
BosV-66	West	Miura	Kinone		100	0.41 ± 0.06	100	0.41 ± 0.06	78 ± 13	91 ± 14	mudstone
BosV-69	West	Miura	Kinone		100	0.38 ± 0.06	99	0.38 ± 0.05	71 ± 12	84 ± 12	mudstone
BosV-90	Central	Miura	Anno	AN65	100	0.40 ± 0.03	100	0.40 ± 0.03	-	89 ± 8	mudstone
BosV-88	Central	Miura	Anno		100	0.34 ± 0.10	100	0.34 ± 0.10	-	74 ± 20	mudstone
BosV-87	Central	Miura	Anno		90	0.36 ± 0.06	89	0.36 ± 0.05	-	79 ± 13	mudstone
BosV-86	Central	Miura	Kiyosumi		100	0.42 ± 0.05	100	0.42 ± 0.05	-	93 ± 10	mudstone
BosV-92	Central	Miura	Kiyosumi	Ky21	100	0.39 ± 0.05	100	0.39 ± 0.05	-	86 ± 15	mudstone
BosV-89	Central	Miura	Amatsu	AM5	100	0.33 ± 0.03	100	0.33 ± 0.03	-	71 ± 9	mudstone
BosV-77	Central	Miura	Kinone		64	0.61 ± 0.11	64	0.61 ± 0.11	114 ± 16	128 ± 17	mudstone
BosV-78	Central	Miura	Kinone		100	0.50 ± 0.11	99	0.49 ± 0.10	94 ± 16	108 ± 19	mudstone
BosV-79	Central	Miura	Kinone	KN2	100	0.33 ± 0.05	100	0.33 ± 0.05	58 ± 14	71 ± 14	mudstone
BosV-80Kn	Central	Miura	Kinone	KN3	72	0.38 ± 0.10	71	0.38 ± 0.09	71 ± 22	84 ± 22	mudstone
BosV-82	East	Kazusa	Ohara		100	0.36 ± 0.06	100	0.36 ± 0.06	-	79 ± 15	mudstone
BosV-81	East	Kazusa	Namihana		21	0.37 ± 0.06	21	0.37 ± 0.06	-	82 ± 14	mudstone
BosV-83	East	Kazusa	Namihana		43	0.34 ± 0.05	43	0.34 ± 0.05	-	74 ± 13	mudstone
BosV-80	East	Kazusa	Katsuura		96	0.41 ± 0.05	96	0.41 ± 0.05	-	91 ± 12	mudstone
BosV-93	East	Miura	Kiyosumi		100	0.35 ± 0.05	100	0.35 ± 0.05	-	76 ± 14	mudstone
BosV-94	East	Miura	Kiyosumi		100	0.41 ± 0.04	100	0.41 ± 0.04	-	91 ± 9	mudstone

4.4 Paleo-temperature distribution in the Boso forearc basin

In the western part of the basin, vitrinite reflectance data show clear differences above and below the Kurotaki Unconformity (0.40% versus 0.25%, respectively), although the vitrinite reflectance values obtained for the Miura and Kazusa groups are both relatively low. Considering that the stratigraphic thickness between the two sampling points (i.e., the Kiwada and Anno formations) is only 335 m, the vitrinite reflectance gap is obvious. There are three possible reasons for the variations in thermal maturity: (1) there was advective heating when the Miura Group was deposited; (2) the geothermal gradient was higher before the unconformity formed; and (3) the burial depth of the Miura Group was much greater than that of the Kazusa Group and a large amount of the former was eroded prior to deposition of the latter.

Vitrinite reflectance values for the Miura Group are 0.33%–0.42% in the central and eastern parts of the Boso forearc basin. A high geothermal gradient appears unlikely, given the near-uniform vitrinite reflectance values in the Miura Group regardless of stratigraphic position. Advective heating prior to the formation of the Kurotaki Unconformity would appear to be the likely cause of the observed variation in thermal maturity. Assuming advective heating, 1 Myr is an appropriate heating duration for the Kinone Formation. In this case, part of the Kinone Formation has the highest paleo-temperature of >100°C (Table 4.1).

In contrast, paleo-temperatures for the Kazusa Group in the eastern part of the basin vary from 74 to 91°C, with the highest temperature being for the Katsuura Formation, which is the lowermost unit. This suggests a vertical geothermal gradient in the Kazusa Group in the eastern part of the basin. The Kazusa Group shows spatial variations in vitrinite reflectance, with higher values in the eastern part of the basin than in the west. There is also an increase in paleo-temperature with depth in the Kazusa Group in the eastern part of the basin. The results do not show an abrupt change in paleo-temperatures across the Kurotaki Unconformity in the eastern part of the basin. The distribution of vitrinite reflectance values and paleo-temperatures based on a heating

duration of 1 Myr in the Boso forearc basin are shown in Fig. 4.6. The paleo-temperatures for the Miura Group are 60–90°C and fairly uniform from east to west, despite the stratigraphic differences and large-scale folding. The Kinone Formation in the central and western areas of the basin shows the highest paleo-temperatures (>100°C). Given the maximum paleo-temperature of the ophiolite complex is >130°C (Yamamoto et al., 2017), the high temperatures reflect the present location of the complex, which is immediately below and to the south of the Miura Group in the central Boso Peninsula (Figs 2.1 and 4.6). This suggests that advective heating occurred during ophiolite emplacement.

Vitrinite reflectance values in the Miura Group in the forearc basin are obviously higher than those in the accretionary prism (<0.27%; Yamamoto et al., 2005) and trench–slope basin (<0.28%; Yamamoto et al., 2017) located in the southernmost Boso and Miura peninsulas, despite these strata having the same age. Rapid sedimentation and continuous and thick accumulation of basin-fill sediments are expected to have occurred in the forearc basin. However, a smaller amount of sedimentation and rapid uplift due to accretion would be expected in the accretionary prism and trench–slope basin.

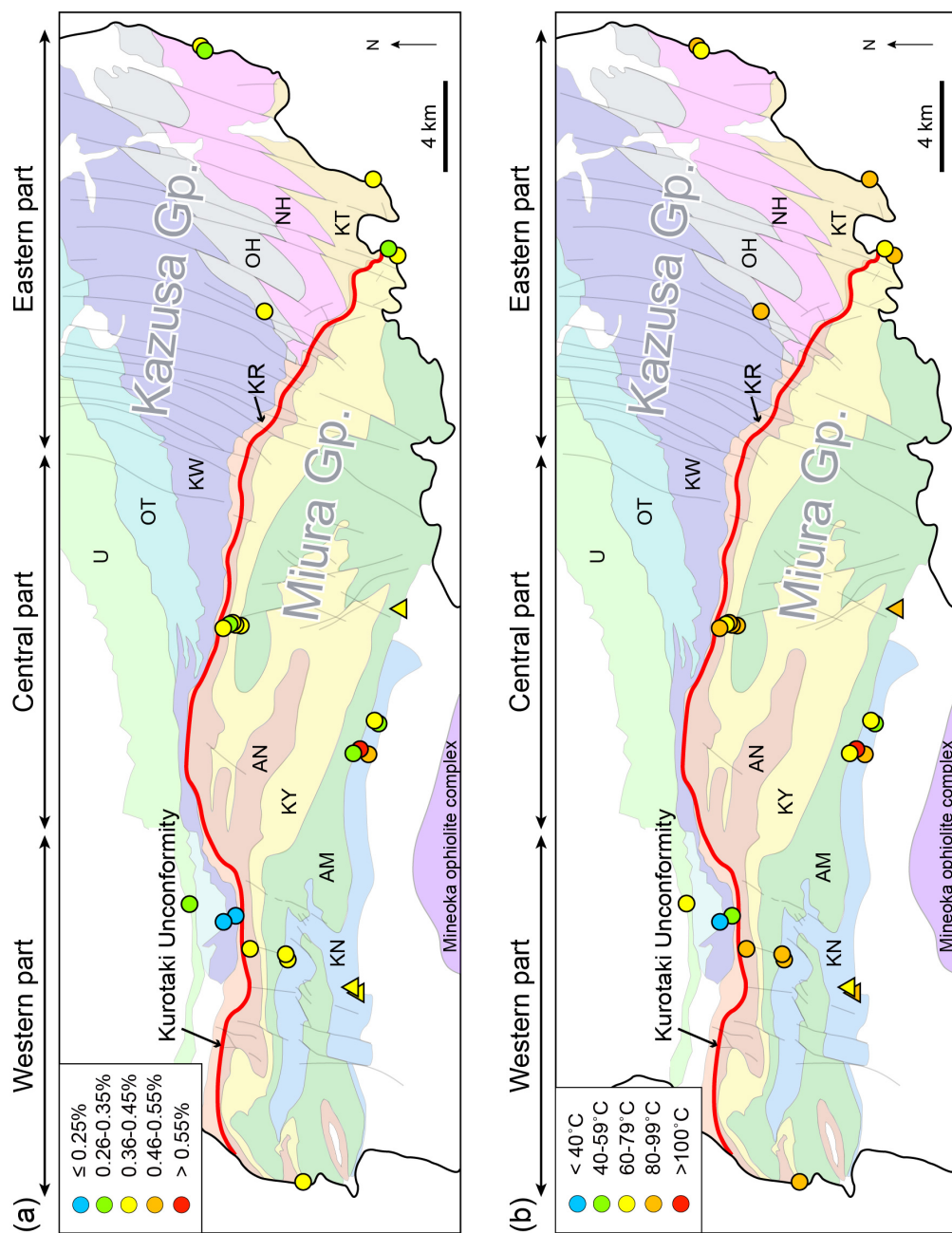


Figure 4.6 Distribution map of vitrinite reflectance (a) and paleo-temperature (b) in the Boso forearc basin. Circles were measured and calculated in this study. Triangles are reported by Yamamoto et al. (2017).

4.5 Conclusions

Vitrinite reflectance data were used to estimate the maximum paleo-temperatures in the Boso forearc basin and constrain the thermal history of the basin. Vitrinite reflectance in the western part of the basin is ca. 0.40% in the Miura Group and ca. 0.25% in the Kazusa Group. The vitrinite reflectance in the central part of the basin is ca. 0.33%–0.61%, and in the Miura and Kazusa groups in the eastern part of the basin is 0.34%–0.41%. The paleo-temperatures of the Miura Group are ca. 70–90°C and essentially uniform across the western and eastern parts of the basin. In the central part of the basin, part of the Kinone Formation has the highest paleo-temperature ($>100^{\circ}\text{C}$). Considering the location and paleo-temperature of the Mineoka ophiolite complex, it is suggested that advective heating occurred during emplacement of the ophiolite, which heated the Miura Group. The paleo-temperature of the Kazusa Group is ca. 50°C in the western part of the basin, whereas it is 75–90°C in the eastern part of the basin. Therefore, the paleo-temperatures differ above and below the Kurotaki Unconformity in the western part of the basin. The contrasting paleo-temperatures between the western and eastern parts of the basin imply that rapid sedimentation and continuous and thick accumulation of basin-fill sediments occurred in the forearc basin.

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Chapter 5

Consolidation characteristics of the Boso forearc basin

5.1 Introduction

The mechanical behavior of rock masses is controlled by weak planes, such as joints, fractures, and faults. However, the mechanical behavior of soft rocks is dominated by the mechanical properties of the rock itself. Given that soft rocks have different physical properties to hard rocks, it is essential to understand the mechanical behavior of soft rocks when tunnels and large structures are constructed in such rocks. Neogene and Quaternary sedimentary rocks that are generally considered to be soft rocks are distributed throughout Japan. Soft sedimentary rocks have physical properties intermediate between those of soil and hard rocks (Akai, 1993).

Sediment is compacted by the overburden pressure and both the rock's bulk volume and pore volume decreases with dehydration. This process is referred to as consolidation, which progresses with increasing overburden pressure. Given that the maximum burial depth of basins increases with sediment thickness, consolidation increases with burial depth. Consolidation tests involve a uniaxial compression experiment that is conducted until the specimen yields. These geotechnical experiments can be used to investigate subsidence and the maximum sediment overburden pressure. The behavior of the consolidation curve differs for over-consolidated versus normally consolidated regions, and the transition between these two states is defined as the consolidation yield stress, which corresponds to the maximum overburden pressure that a sample has experienced. Consolidation tests are generally applied to soil samples, but have been adapted for testing of compacted soft sedimentary rocks (e.g., Hosono et al., 1993; Kitajima & Saffer, 2014).

As such, consolidation tests are effective for understanding the mechanical characteristics of soft rocks. In general, strain increases as stress increases with loading. However, strain softening is a phenomenon where the stress does not increase with loading after the stress reaches a certain value, or the strain increases significantly for a small stress increase. The phenomenon is commonly observed during shearing of dense sands, over-consolidated clays, and soft rocks, and is closely related to the progressive failure observed in many geotechnical studies. Mechanical models for strain softening using numerical analysis have been constructed (Adachi & Ogawa, 1980; Adachi et al., 2000). Triaxial compression tests conducted on soft sedimentary rocks, such as porous tuffs (Akai et al., 1978; Adachi et al., 1998) and diatomaceous soft rocks (Maekawa & Miyakita, 1983), have constrained their mechanical properties. Akai et al. (1978) and Adachi et al. (1998) conducted triaxial compression tests on Oya stone, which is a porous tuff, and showed that the elastoplastic behaviour, dilatancy, and time dependence were characterized by the intermediate between soft soil and hard rock. Adachi & Oka (1995) developed constitutive equations for the elastoplastic behavior of strain softening in soft rocks, and proposed that strain softening is caused by both localization of deformation and softening of the rock material itself. They showed that the models and experimental results were consistent. Furthermore, Zhang et al. (2005) and Iwata et al. (2011) improved these constitutive equations and compared them with actual phenomena. However, there are currently no reports of strain softening in consolidation tests. There are a number of ways to view the term “strain softening.” When the softening occurs during loading of the rock material, shear bands simultaneously form in the loaded specimens. Therefore, this softening is not “strain softening”, but is recognized as a phenomenon caused by plastic hardening of the material and localization of deformation, such as shear bands. This topic was a subject of debate in the 1980s and 1990s, and remains an unresolved issue.

The siltstone and mudstone of the Boso forearc basin are both classified as soft rocks, and have a consolidation fabric characterized by clay minerals arranged parallel to the

bedding. Therefore, consolidation tests on siltstones of the Boso forearc basin can constrain the basin behavior and features of these soft rocks. I performed consolidation tests on siltstone samples using a one-dimensional consolidation testing method under controlled strain loading (American Society for Testing and Materials, 2004; Japanese Standards Association (JSA), 2009a), and then estimated the maximum overburden pressure from the consolidation yield stress (Fig. 5.1).

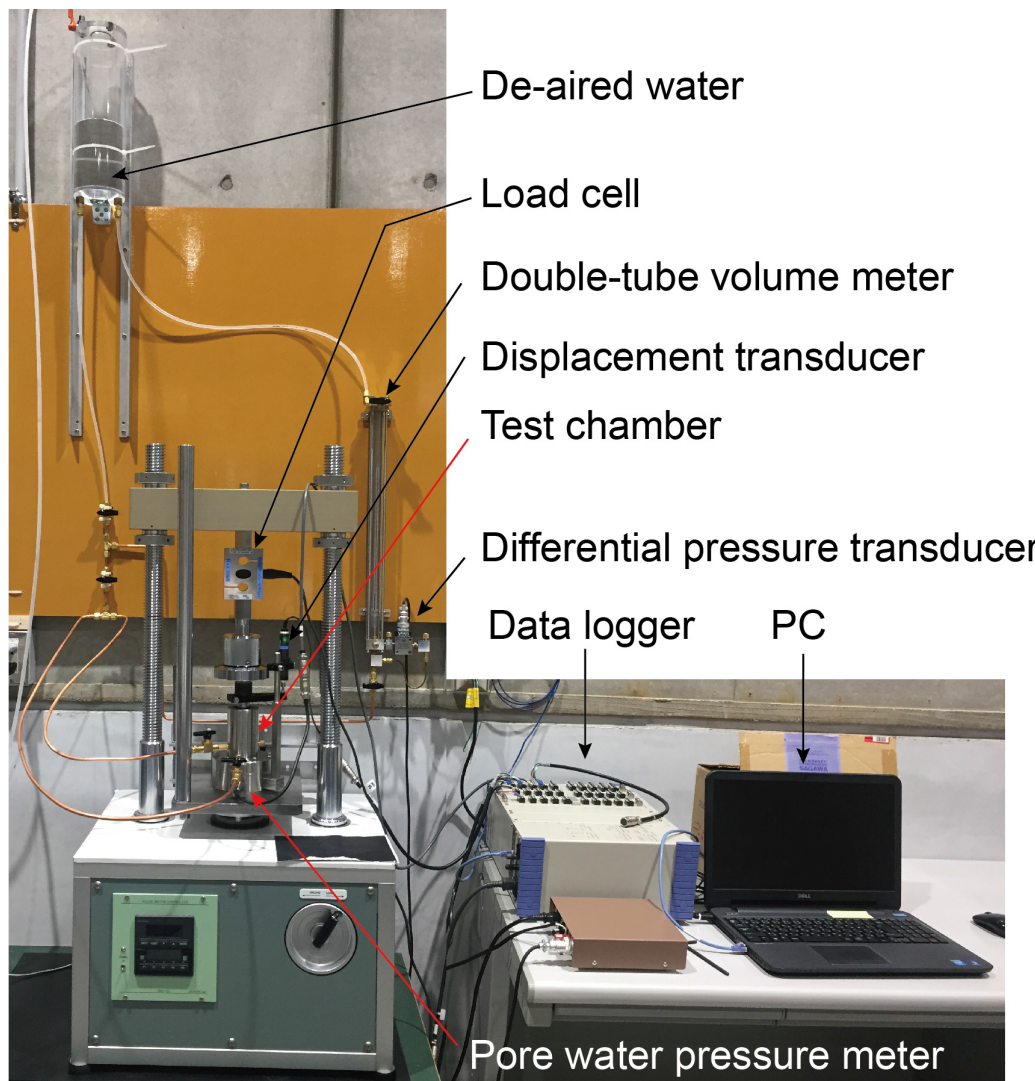


Figure 5.1 Photograph of the consolidation test apparatus for the consolidation tests under K_0 -conditions. Specimen is in the test chamber.

5.2 Methodology

5.2.1 Physical and mechanical principles of consolidation

Here I provide a brief review of how sediments respond to burial, unloading, and consolidation. I also introduce the mechanical terminology that will be used throughout the following sections.

Morgan & Ask (2004) described the consolidation state of underthrust sediments and discussed the evolution of the décollement along the Nankai accretionary margin. They also summarized the physical and mechanical responses to reconsolidation of sediments.

During burial, marine sediments are subjected to increasing vertical stresses imposed by the thickening overburden. The total vertical stress acting on the sediment is:

$$\begin{aligned}\sigma_v &= (\rho_b - \rho_w) \cdot g \cdot z + \rho_w \cdot g \cdot H \\ &= (\rho_b - \rho_w) \cdot g \cdot z + P_w\end{aligned}\quad (5-1)$$

where ρ_b and ρ_w are the sediment bulk density (kg/m^3) and pore water density (1035 kg/m^3), respectively, g is the gravitational acceleration (9.8 m/s^2), z is the burial depth (m below seafloor; mbsf), and H is the depth of the water column (m). The first term represents the buoyant weight of the sediment, and P_w is the hydrostatic water pressure imposed by the weight of the oceanic water column (MPa). Pore fluid pressures, u , in excess of the hydrostatic pressure support some of the vertical load acting on the sediment, so a more meaningful stress variable is the effective vertical stress, which is defined as follows:

$$\sigma'_v = \sigma_v - u = (\rho_b - \rho_w) \cdot g \cdot z - P_* \quad (5-2)$$

$$P_* = u - P_w \quad (5-3)$$

where P_* is the excess pore pressure. In the case that $u = P_w$ and there is no excess pore pressure, I refer to the hydrostatic effective vertical stress:

$$\sigma'_{vh} = \sigma_v - P_w = (\rho_b - \rho_w) \cdot g \cdot z \quad (5-4)$$

Finally, to distinguish the actual stresses acting on the sediment *in situ* from the stresses measured in the laboratory, I refer to the *in situ* effective vertical stress, σ_{v*}' . The maximum excess pore fluid pressure acting on the sediment is then given as follows:

$$P_*^{max} = \sigma'_{vh} - \sigma_{v*}' \quad (5-5)$$

Sediment porosities are assumed to be in equilibrium with the effective vertical stress; the increase in stress occurs at a rate such that the internal pore pressure is being dissipated. This relationship is obtained by plotting $\log(\sigma'_v)$ versus porosity (η) (Fig. 5.2), which defines a line with a slope that is defined as the modified compression index:

$$C_c = \Delta\eta / \Delta\log(\sigma'_v) \quad (5-6)$$

If P_* is greater than zero, then the pore pressure supports some of the vertical load, leading to what is referred to as an “underconsolidated” state in geological studies (e.g., Jones, 1994), although this term has little physical meaning, because it simply denotes $\sigma'_v < \sigma'_{vh}$, and is not a unique consolidation state. In contrast, sediment unloading due to a reduction in σ'_v or increase in P_* , causes the sediment to return to an elastic state (Wood, 1990), accompanied by modest elastic rebound and a porosity increase (e.g., Karing and Hou, 1992), and is referred to as “overconsolidated” (Fig. 5.2).

Laboratory uniaxial consolidation experiments are conducted by imposing steadily increasing σ'_v while maintaining the lateral dimensions of the sample (e.g., Lambe &

Whitman, 1979). Initially, sediments follow an elastic reloading trend, recovering the elastic volume strain they experienced upon unloading (Fig. 5.2); this leads to lower elastic $C_{c\eta}$ than unloading normal consolidation. This phase is followed by sediment yield, at which point the sample experiences renewed plastic deformation, increased volume strain, and a return to the normal consolidation trend (Fig. 5.2). The stress at sediment yield defined on the η – $\log(\sigma'_v)$ plot reflects the maximum effective vertical stress that the sample has previously experienced, which is known as the pre-consolidation stress, σ'_c . For normally consolidated sediments, σ'_c generally coincides with the effective vertical stress acting on the sediment in place, which I denote as σ'_{v*} (Lambe & Whitman, 1979). Consequently, the consolidation tests provide a means for determining *in situ* stress and pore pressure conditions.

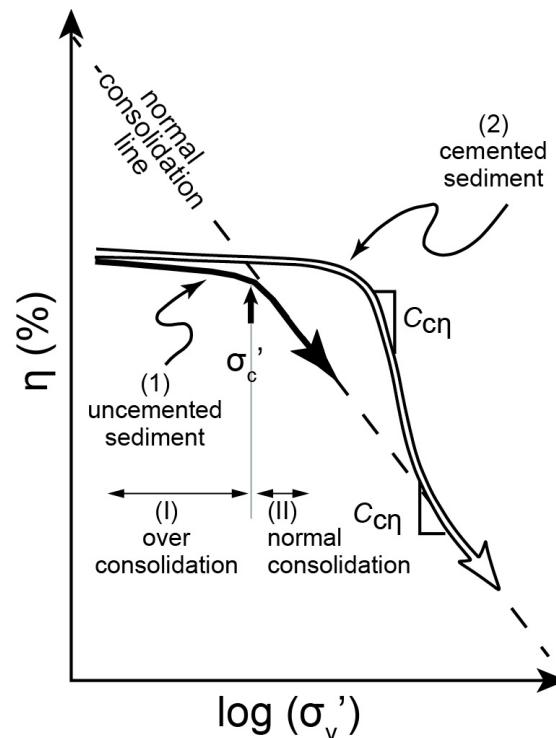


Figure 5.2 Schematic diagram of consolidation experiments for uncemented consolidated (1) and cemented (2) sediment (modified after Morgan & Ask, 2004).

5.2.2 Methods of consolidation test and samples

Consolidation tests were conducted using a uniaxial compression system following JIS (Japanese Industrial Standard) A1227 (JSA, 2009a) with a constant strain rate (Figs 5.1 and 5.3a). The experimental apparatus can continuously increase axial load under K_0 conditions, which is one-dimensional consolidation representing *in situ* conditions. Lateral expansion was restricted using a rigid consolidation ring (Fig. 5.3b). Silicon grease sandwiched by two Teflon sheets was wrapped around the specimen to reduce friction between the consolidation ring and the rock specimen. Porous metal plates were placed on the top and bottom of the specimen to allow pore water to drain. A slower strain rate of 0.05%/min was applied, in order to avoid the generation of pore pressure. The maximum vertical loading capacity was 50 kN, which was applied to generate a maximum compressive stress of ca. 80 MPa for specimens that were 25 mm in diameter. The vertical load and displacement, pore water pressure, and drainage water volume were measured and recorded using a data logger.

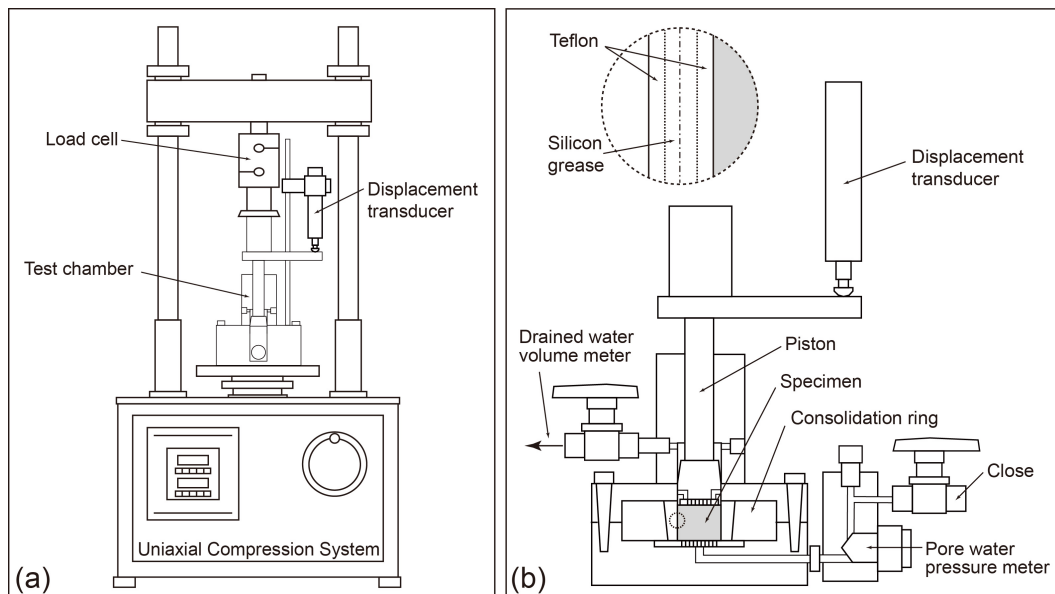


Figure 5.3 (a) Schematic illustrations of the compression system. (b) Schematic illustrations of the test chamber.

The compression index was measured as follows (JSA, 2009a). A consolidation curve that shows the effective stress, p , versus the void ratio, e , on a semi-logarithmic axis was plotted (Fig. 5.4). In a constant strain rate consolidation test, the effective stress, p , is expressed as

$$p = \sigma - \frac{2}{3}u \quad (5-7)$$

where σ is the axial compressive stress and u is the pore water pressure. The void ratio was calculated as follows:

$$e = \frac{V}{V_g} - 1 \quad (5-8)$$

where V and V_g are the total volume of the specimen and the grain volume of the specimen, respectively, calculated using the measured dry, wet, and submerged weights. Two points on the normal consolidation line were selected, and the compression

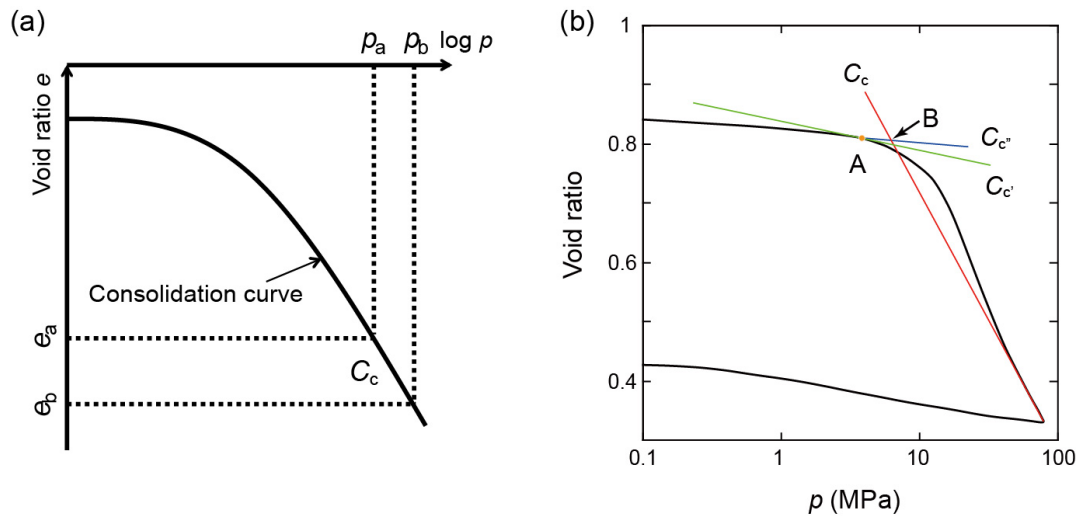


Figure 5.4 Void ratio versus log pressure diagram. (a) The estimation of compression index. (b) The determination of consolidation yield stress.

index, C_c , is calculated as the following equation essentially same as Eq.5–6

$$C_c = \frac{e_a - e_b}{\log p_b - \log p_a} \quad (5-9)$$

where e_a and e_b are the void ratios under effective stresses p_a and p_b , respectively (Fig. 5.4a). The consolidation yield stress, p_c , is measured using the graphical method of Mikasa (JSA, 2009b) as follows (Fig. 5.4b). A line is drawn with gradient:

$$C_c' = 0.1 + 0.25C_c \quad (5-10)$$

which is tangential to the consolidation curve at point A. A line (shown in green in Fig. 4b) is then drawn from point A with the following gradient:

$$C_c'' = 0.5C_c' \quad (5-11)$$

The lateral axis value (i.e., effective pressure) of point B, where the line intersects the normal consolidation line, is the consolidation yield stress.

Block samples without cracks and weathered parts were collected from outcrops (Fig. 5.5). The effects of surface weathering were minimized by excluding any specimens exhibiting a weathered rusty color, and excluding the uppermost part of each core. Only blue–gray fresh siltstones were cut into blocks. The block samples were cut into rectangular shapes, both parallel and perpendicular to the bedding plane, using an electric-powered rock saw. After cutting, the samples were cored using an electric-powered coring machine with a 27 mm inner diameter coring bit. The core diameters were modified with an electric-powered cylindrical grinder so that they would fit inside the rigid consolidation ring with double Teflon sheets. The samples were cut to a height of ca. 21 mm using a rock cutter and the top and bottom surfaces were polished using a surface grinder, such that both ends were parallel and the cores were 20 mm high. These

cores were immersed in tap water in a vacuum desiccator in order to saturate the samples.

The density and void ratio of the specimens were measured using the buoyancy method. In detail, the wet and submerged weights of the specimens were measured before the tests. The specimens were dried in an oven at 60°C for 4 d after the tests and the dry weight was then measured. The equations for calculating soil particle density, ρ_g , void ratio, e , and porosity, ϕ , of each specimen are as follows:

$$\rho_g = \frac{M_{dry}}{(M_{wet} - M_{water}) - (M_{wet} - M_{dry})}, \quad (5-12)$$

$$e = \frac{(M_{wet} - M_{water}) - M_{dry}/\rho_g}{M_{dry}/\rho_g} \text{ and} \quad (5-13)$$

$$\phi = \frac{M_{wet} - M_{dry}}{M_{wet} - M_{water}}, \quad (5-14)$$

where M is the weight, ρ is the density, and the subscripts *dry* and *wet* for M are the weights of the dry and wet specimen measured in the atmosphere, respectively, and *water* is the weight of the wet specimen whilst submerged.

I performed the consolidation tests on mudstone and siltstone samples taken from the Boso forearc basin, and determined the consolidation yield stress to estimate the maximum overburden pressure of each formation. Apparently homogeneous siltstone samples were collected from the Miura and Kazusa groups. A total of 28 samples were obtained from the Kinone, Amatsu, Kiyosumi, Anno, Kiwada, Otadai, and Umegase formations in the western part of the basin; the Anno and Kiwada formations in the central part of the basin; and the Amatsu, Kiyosumi, Katsuura, Namihana, Ohara, Kiwada, Otadai, Umegase, and Kokumoto formations in the eastern part of the basin (Fig. 5.6). I estimated the thickness from the top of the Kazusa Group to each sampling

point using a key tuff layer (Table 5.1). Figure 5.7 shows the specimen before and after the consolidation test.

I used two testing apparatuses at Kyoto University and Nagoya Institute of Technology, Japan, to assess the reproducibility of the consolidation behavior. Both apparatuses are almost the same system and have similar specifications. These consolidation tests were performed using the sample taken from the Otadai (BosC10)

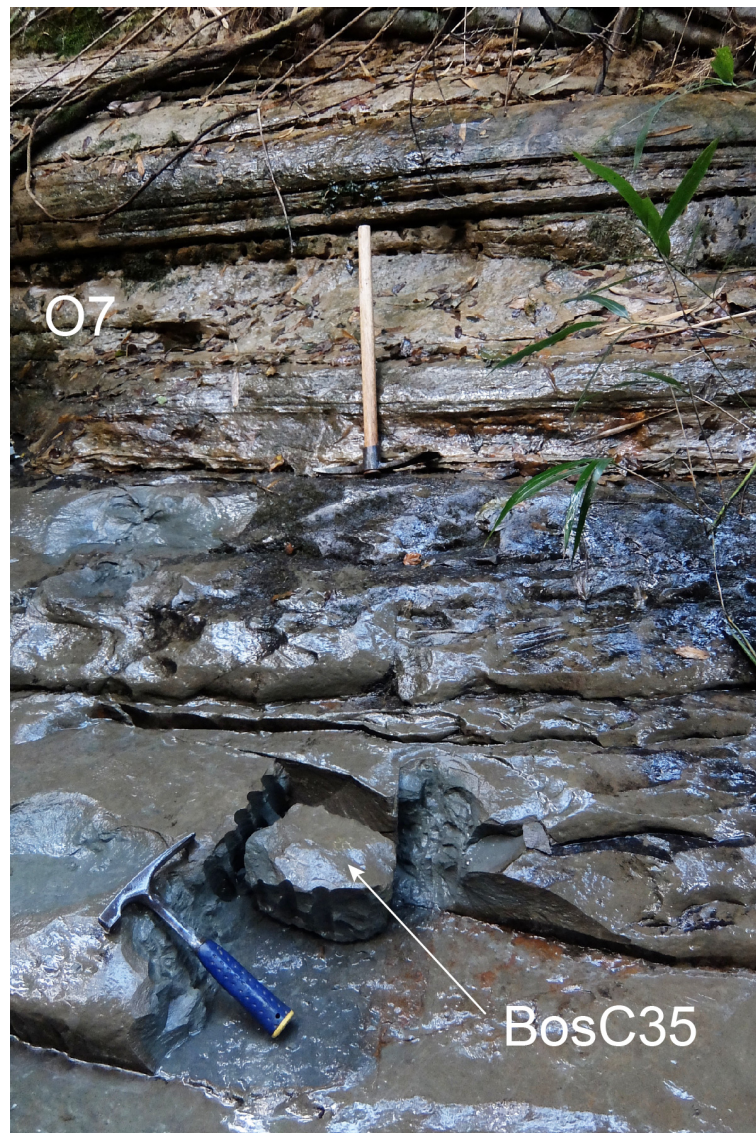


Figure 5.5 Photograph of an example of the sampling point and block sample.

and Kiwada (BosC26) formations. Two specimens were formed from the same block sample and each test was performed using the same procedure on each apparatus. By comparing the consolidation curves and consolidation yield stress obtained for each specimen, the reproducibility of the tests was assessed.

All the specimens used in this study were also subjected to micro-focus X-ray CT imaging before and after the consolidation tests, and the internal structures of the specimens were observed. CT images were acquired using a HMX225-ACTIS+5 (TESCO Corp.) before the test and an Xradia (ZEISS Corp.) after the test. This equipment is housed at Kochi University, Japan. The imaging conditions involved a tube voltage of 90 kV, tube current of 50 μ A, and spatial resolution of ca. 35 μ m.

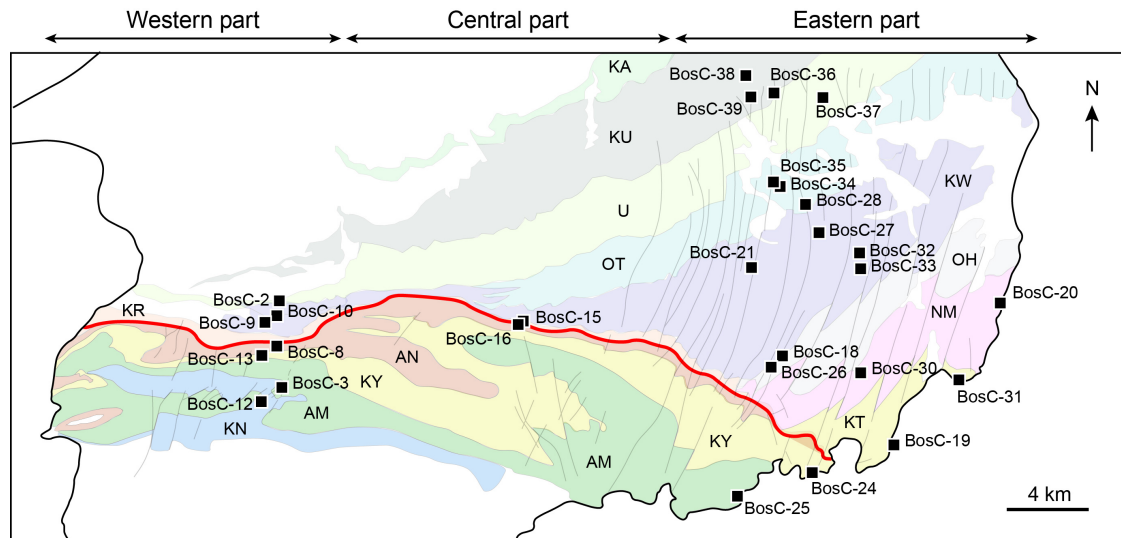


Figure 5.6 Geological map of the Boso forearc basin showing the sampling locations for the consolidation tests.

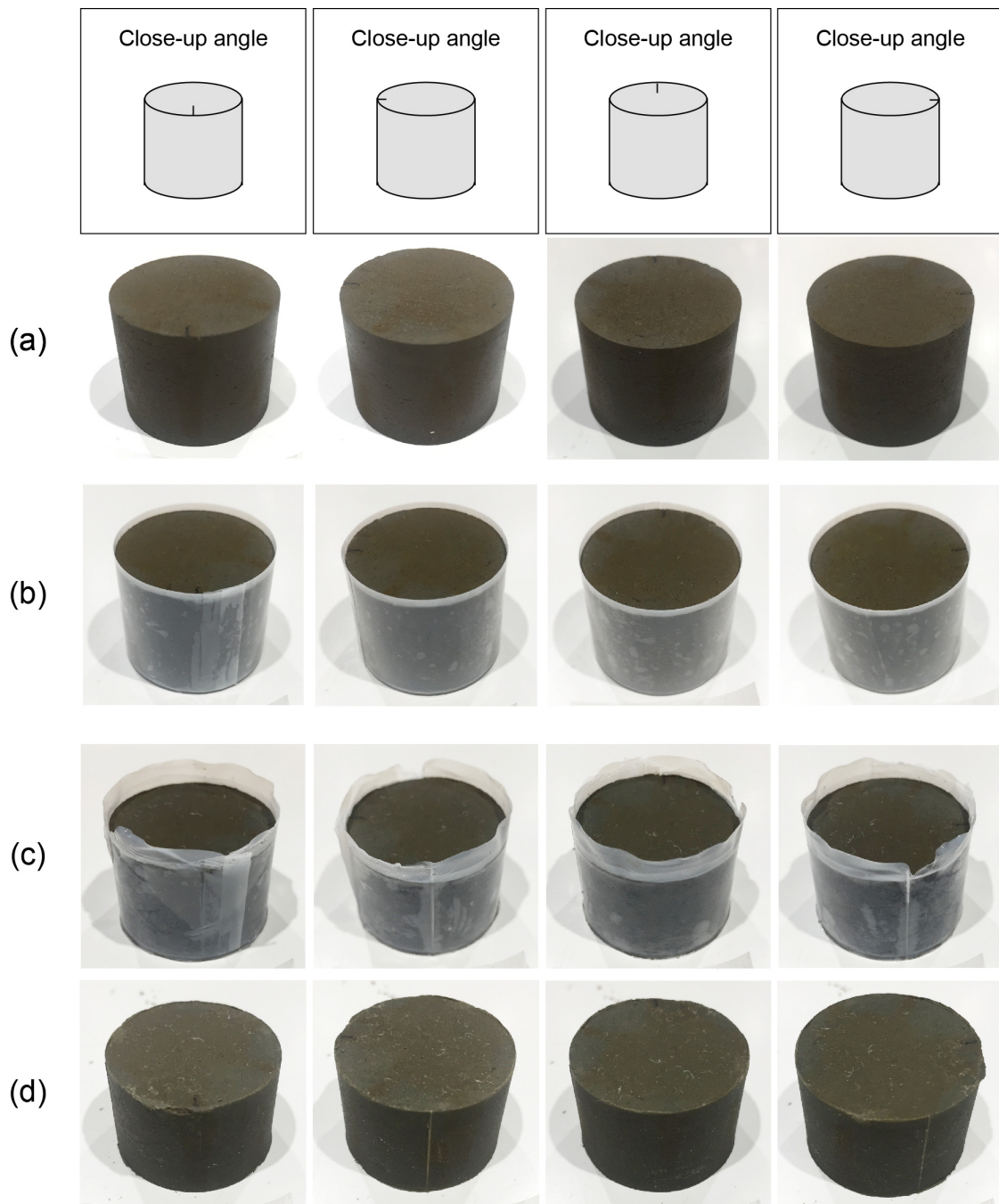


Figure 5.7 Photograph of a specimen used for the consolidation tests (BosC26-K). The diameter is 25 mm. Schematic figures at top show the viewing direction in each column. (a) Before the tests. (b) Before the tests and wrapped with a Teflon sheet. (c) After the tests. (d) After the tests and removal of the Teflon sheet.

5.3 Results

5.3.1 Western part of the basin

Initial porosities (the porosity before consolidation test) of the specimens were 33%–43% and the grain densities were 2.6–2.8 g/cm³ (Table 5.1). Figure 5.8 shows the consolidation curves for each specimen. All specimens yielded and the consolidation curves showed over- and normal-consolidation areas (Fig. 5.8a). After yielding, the weakly cemented specimens from the Kazusa Group (Bjerrum, 1967) exhibited a normal consolidation curve over 20 MPa. In contrast, below the unconformity, the samples from the Miura Group showed normal consolidation curves after yielding.

The specimens have similar C_c values of 0.35–0.39 (Table 5.1). The specimens from the Miura Group do not show a cementation pattern in the consolidation curves, and are characterized by normal consolidation curves after yielding. Similar C_c values were identified of 0.43 for the Anno Formation and 0.38 for the Amatsu Formation. The consolidation yield stress, p_c , of specimens was calculated using C_c values obtained from the normal consolidation curves under a high loading stress. The estimated p_c values are 7.5 MPa for the Umegase Formation, 7.6 MPa for the Otadai Formation, 9.6 MPa for the Kiwada Formation, 23.7 MPa for the Anno Formation, 15.5 MPa for the Kiyosumi Formation, 31.0 MPa for the Amatsu Formation, and 24.6 MPa for the Kinone Formation. The p_c values generally increase with depth, however, those of the Kiyosumi and Kinone formations are smaller than those of the overlying formation. The p_c values of the Kiyosumi Formation in the western part of the basin are lower than the normal consolidation trend in this part of the basin.

The consolidation curve of BosC10-K shows a decreasing trend in terms of void ratio, similar to that of BosC10-N (Fig. 5.8a). The p_c was 7.6 MPa for BosC10-N and 8.0 MPa for BosC10-K. Given that the initial void ratio was 0.74 for BosC10-N and 0.69 for BosC10-K, there is a discrepancy in the consolidation curves for the two samples. Basic physical properties, such as porosity and soil particle density, vary

among specimens even from the same geological unit, because of the inhomogeneity of natural rocks. A void ratio difference of ± 0.05 can be considered a natural variation. The consolidation curves for specimens BosC10-N and BosC10-K were drawn smoothly from the beginning of loading to the completion of unloading.

(a)

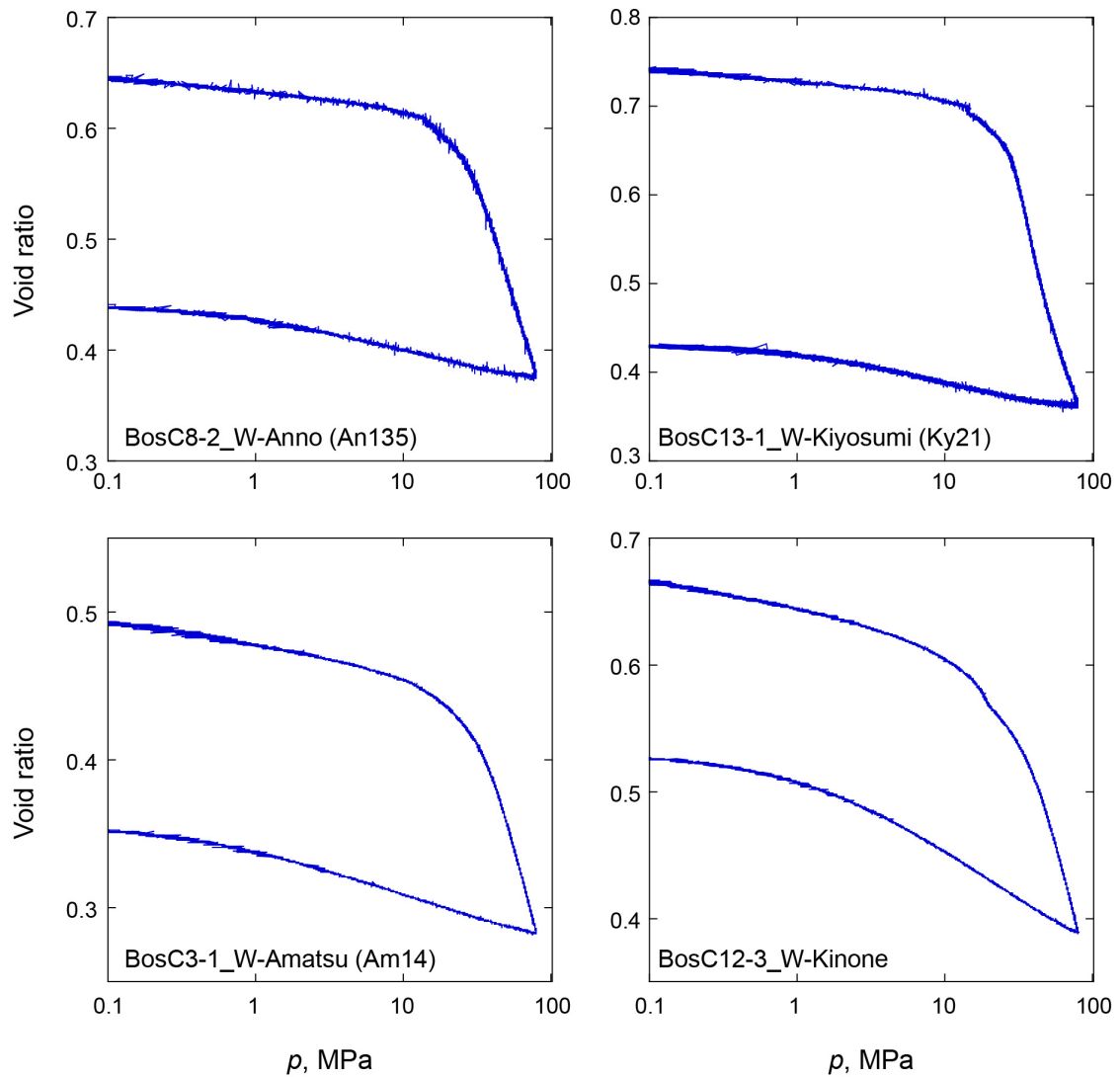


Figure 5.8 Consolidation curves of the samples collected from the western part of the Boso forearc basin. Blue curves show the results obtained by the apparatus at Nagoya Institute of Technology. Red curves show the results obtained by the apparatus at Kyoto University. (a) Western part of the Miura Group. (b) Western part of the Kazusa Group.

(b)

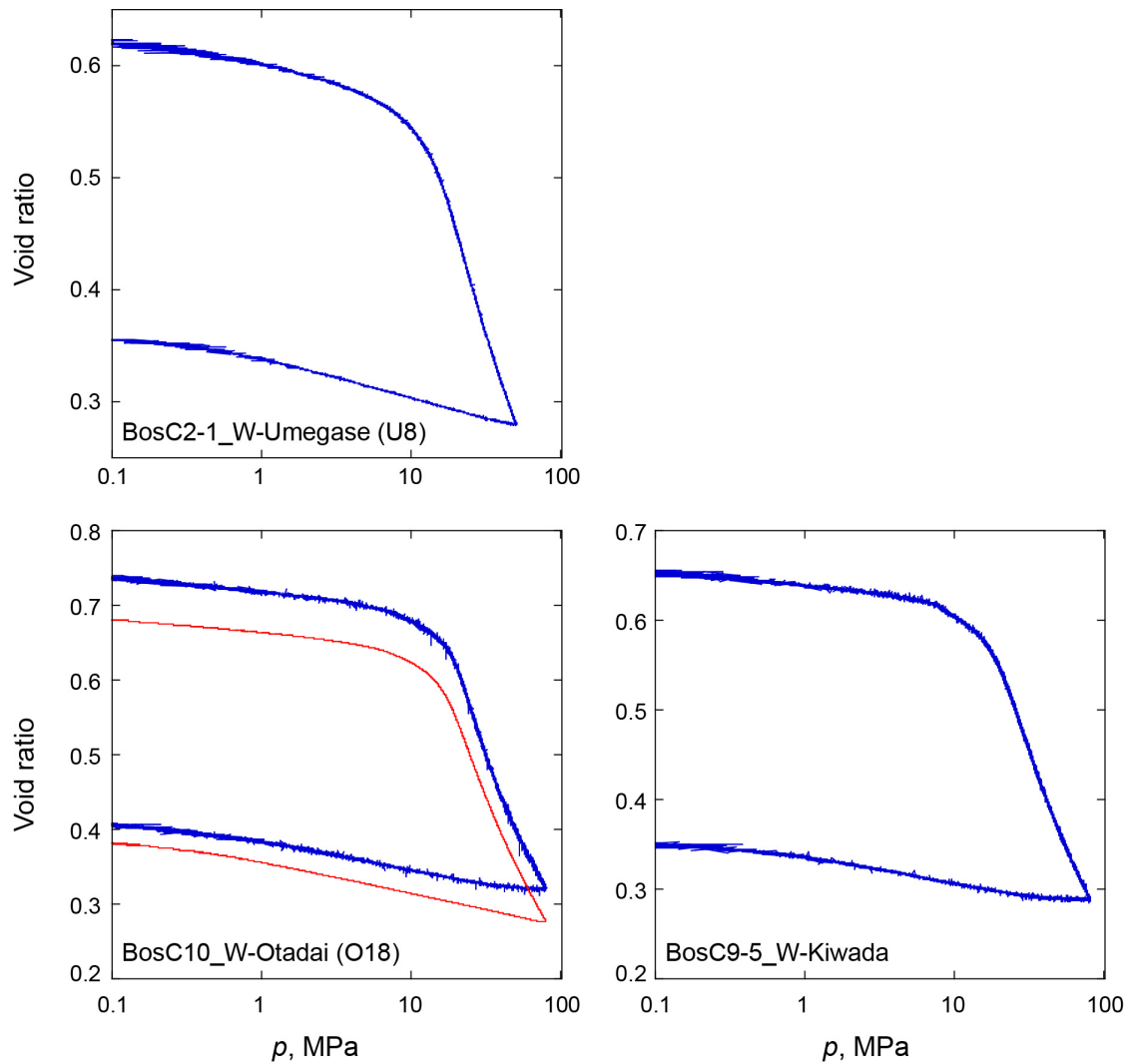


Figure 5.8 Continued.

5.3.2 Central part of the basin

Initial porosities and grain densities were 41.1% and 2.64 g/cm^3 for the Kiwada Formation and 43.8% and 2.64 g/cm^3 for the Anno Formation, respectively (Table 5.1). The consolidation curves show normal- and over-consolidated areas in both formations (Fig. 5.9). These curves show continuous and typical consolidation behavior, but the curves slightly overhang to a high-pressure area after yielding. The compression index, C_c , which is the inclination of the normal consolidation line, is 0.38 for both the Kiwada and Anno formations, and p_c is 15.8 MPa for the Kiwada Formation and 11.7 MPa for

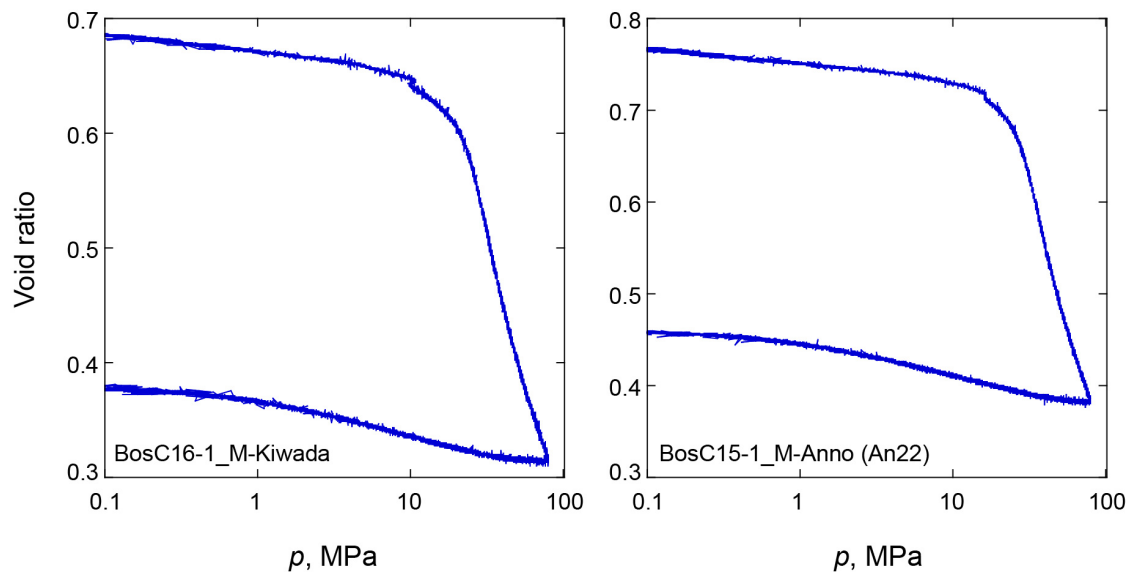


Figure 5.9 Consolidation curves for the samples collected from the central part of the Boso forearc basin. These tests were performed using the apparatus at Nagoya Institute of Technology.

the Anno Formation. In the central part of the basin, p_c for the Anno Formation is larger than that of the Kiwada Formation.

5.3.3 Eastern part of the basin

In general, porosity decreases from 50% to 35% with depth, although there are slightly higher-porosity zones between 2250–2000 and 2800–2500 m in which the porosity is ca. 50%. The sampling intervals in the Kiwada, Ohara, Namihana, and upper Katsuura formations correspond to the higher-porosity zone. The grain densities of the specimens are 2.5–2.7 g/cm³ (Table 5.1). All specimens yielded and the consolidation curves showed over- and normal-consolidation areas, with the stress of some specimens decreasing abruptly before yielding (Fig. 5.10). Just after the yield stress and up to 50

MPa, some curves slightly overhang in the high stress area, reflecting the effects of cementation (e.g., Spinelli et al., 2007); almost all specimens from the Kazusa Group exhibit this cementation effect.

The consolidation yield stresses are listed in Table 5.1. When the consolidation curves show a rapid stress drop, the tangents were drawn using the consolidation curves after this stress drop (Fig. 5.10a). The curves for the Kazusa Group between 50 and 80 MPa are considered to be compaction curves.

The consolidation curves for BosC26-N and BosC26-K are similar, and both show rapid stress drops just before yielding (Fig. 5.10b). Considering the strain–stress relationship around the stress drop point, the values of stress (p) at which the rapid stress drop occurred were ca. 12.3 MPa for BosC26-N and ca. 15.0 MPa for BosC26-K (Fig. 5.11). The p_c of BosC26-N was 9.4 MPa and that of BosC26-K was 10.3 MPa, a difference of ca. 10%. The initial void ratio was 0.99 for BosC26-N and 0.97 for BosC26-K. The specimens had no cracks on the surface both of before and after the test (Figure 5.7). Figure 5.12 shows micro-focus X-ray CT images of BosC26-N taken before and after the test. Although small white spots were observed inside the specimen before the test, there were no large features such as macrofossils and cracks. This means that this specimen is a relatively homogeneous siltstone. The small white spots are microfossils and/or micro-particles of volcanic ash. The specimen was deformed during the test and had a convex shape after the test. This deformation likely occurred because the porous central part of the upper face was softer than the surrounding stainless steel plate. However, CT images of a specimen that experienced high stress during the test indicated that cracks formed by deformation were not distributed throughout the specimen. Although strain softening of soft rocks due to a constant strain rate consolidation test has not been reported in previous studies, this phenomenon was observed in the present study.

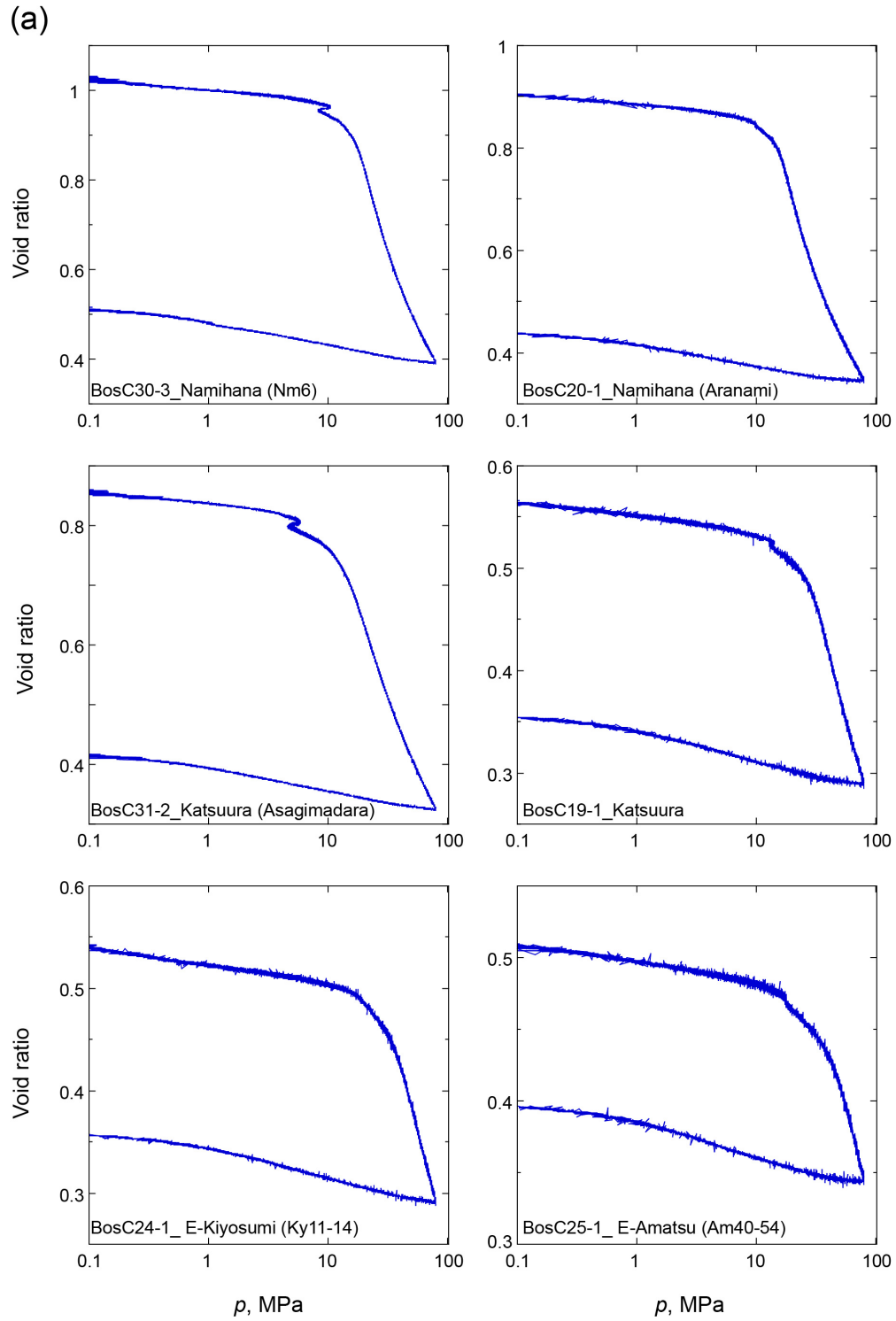


Figure 5.10 Consolidation curves for the samples collected from the eastern part of Boso forearc basin. Blue lines show the results obtained by the apparatus at Nagoya Institute of Technology. Red lines show the results obtained by the apparatus at Kyoto University. (a) Eastern part of the Miura and Kazusa groups. (b–d) Western part of the Kazusa Group.

(b)

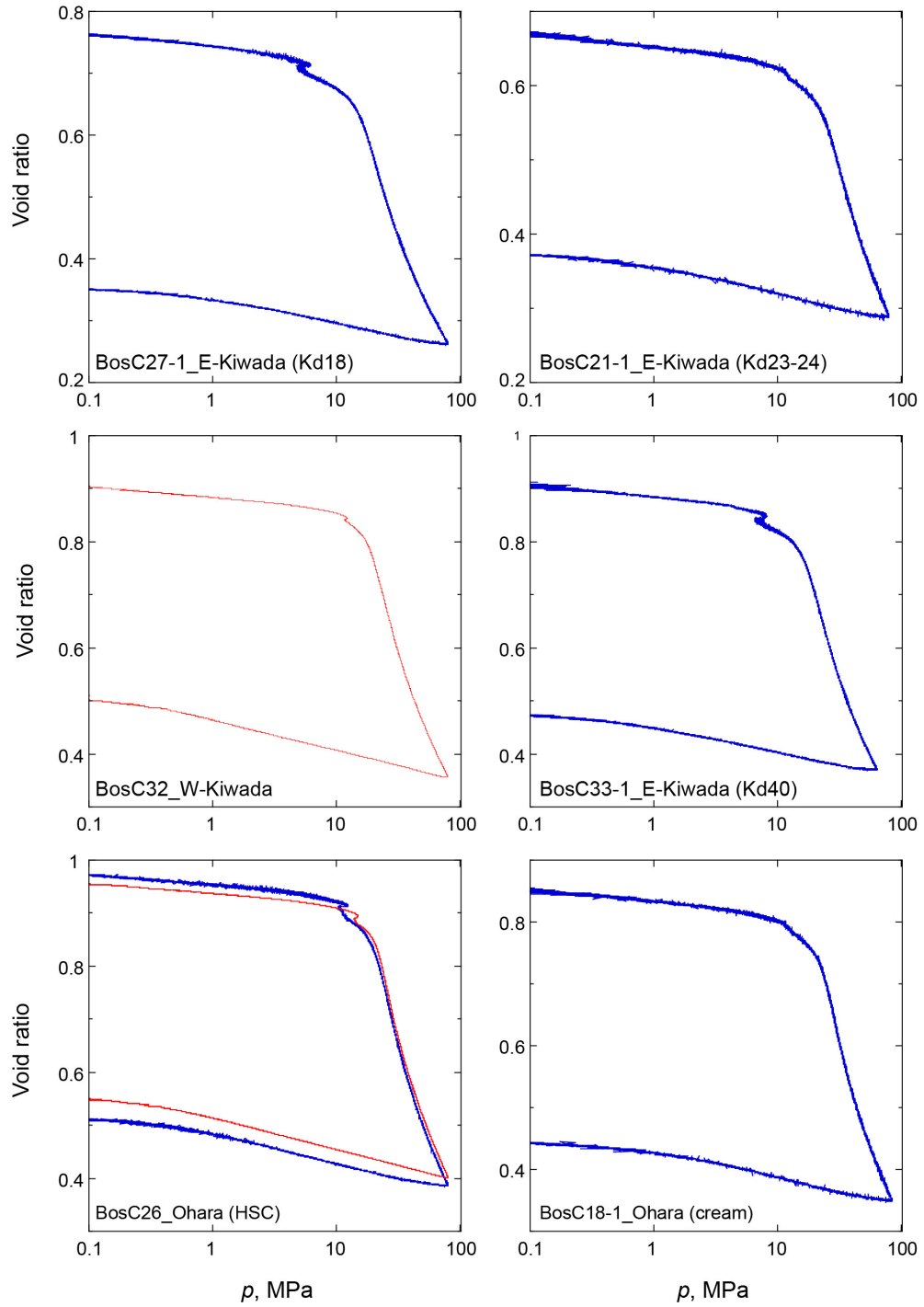


Figure 5.10 Continued.

(c)

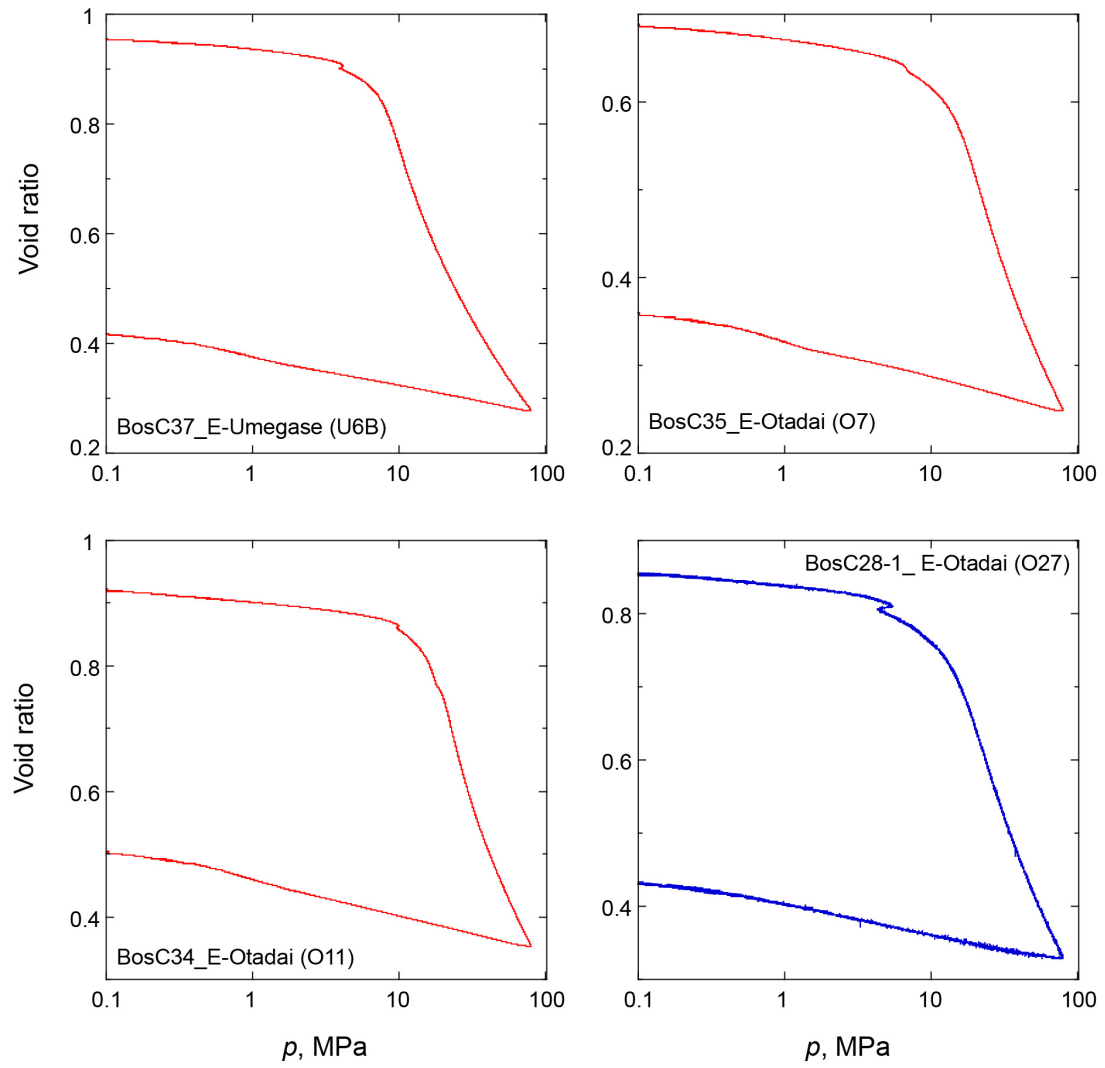


Figure 5.10 Continued.

(d)

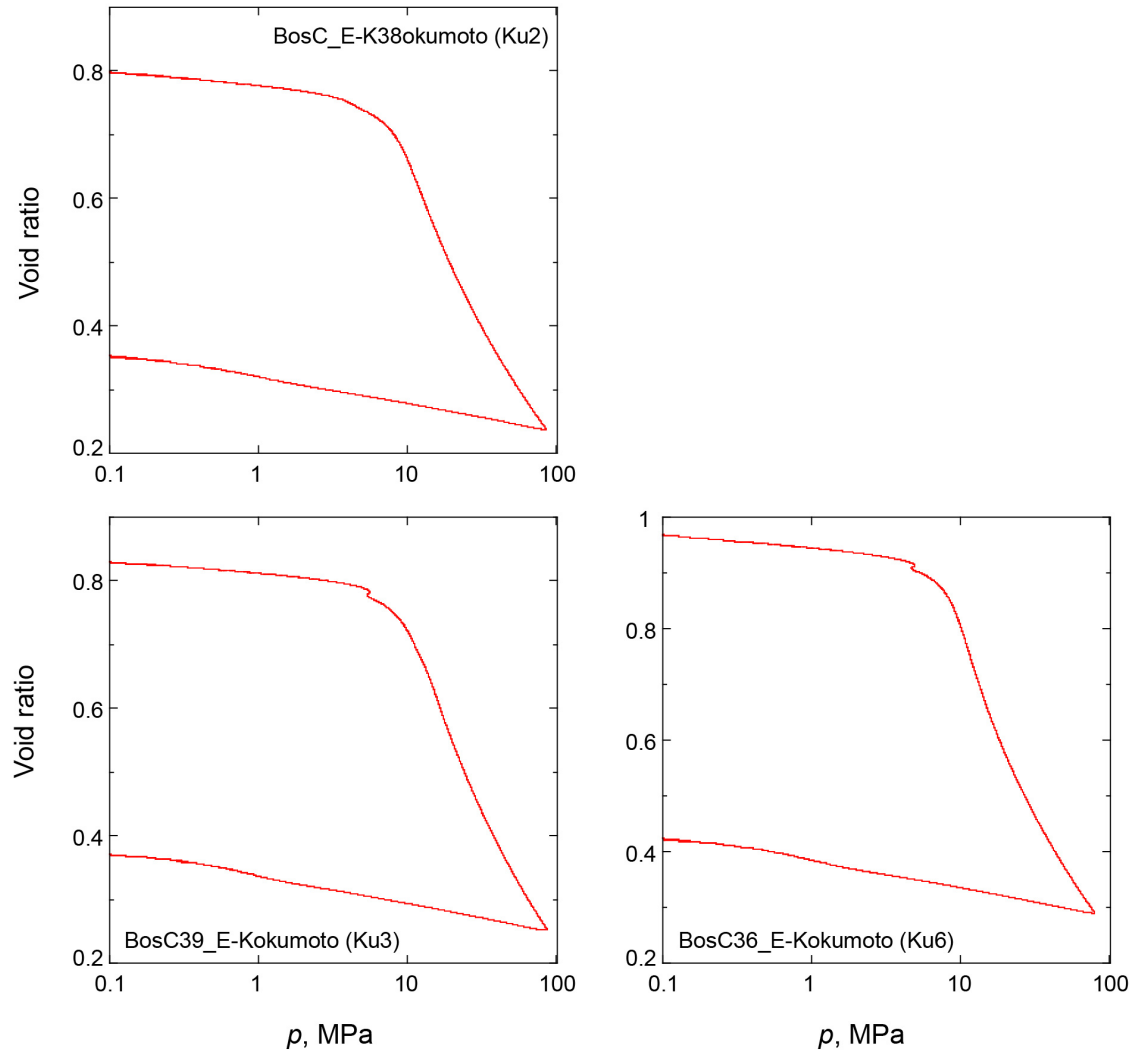


Figure 5.10 Continued.

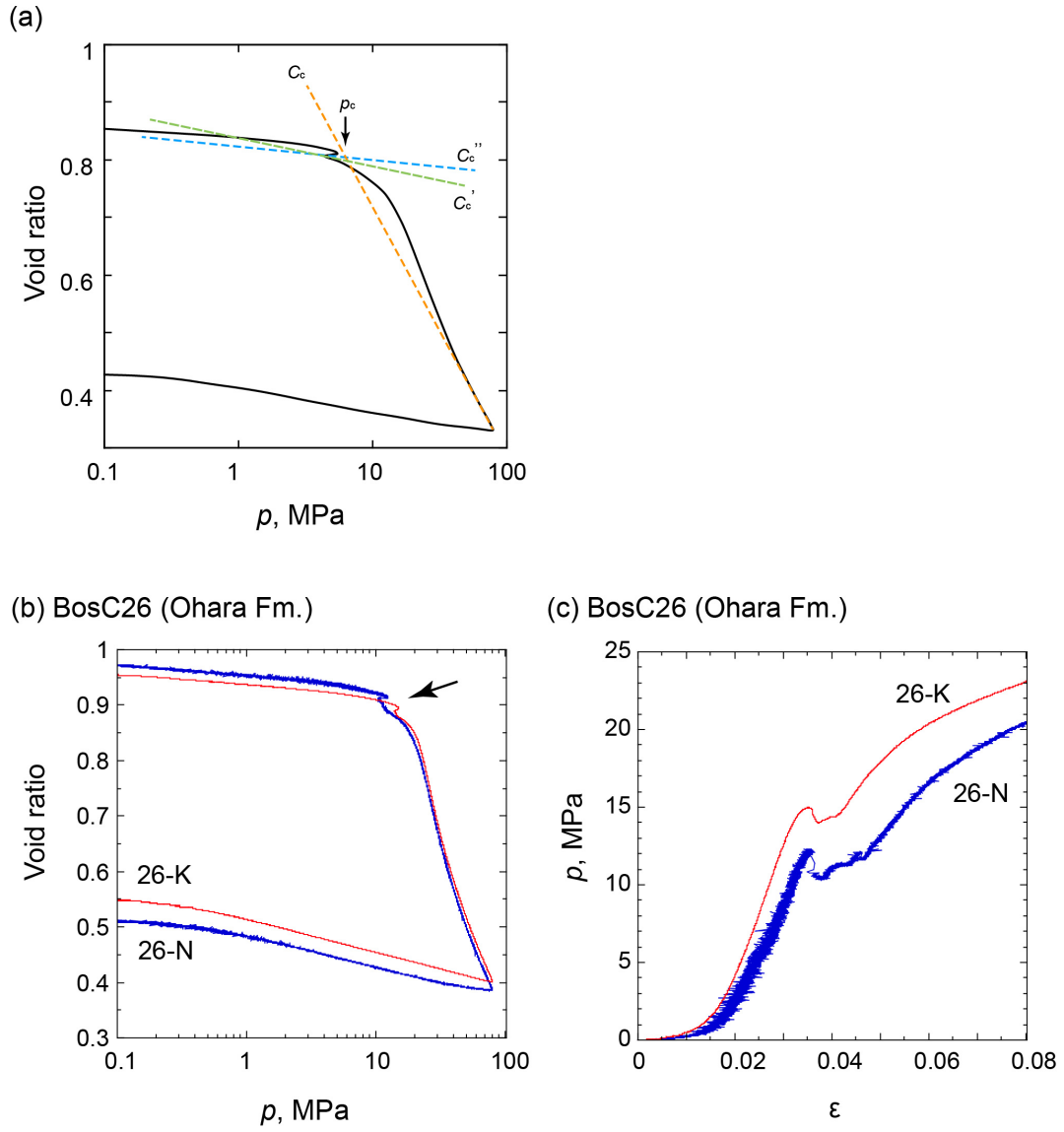


Figure 5.11 (a) The measurement of consolidation yield stress for the specimen with strain softening. (b) Consolidation curves for BosC26-N and BosC26-K. Red curves show the test performed at Kyoto University and the blue curves show the test performed at Nagoya Institute of Technology. (c) Stress versus strain diagram for BosC26 focused on the strain softening.

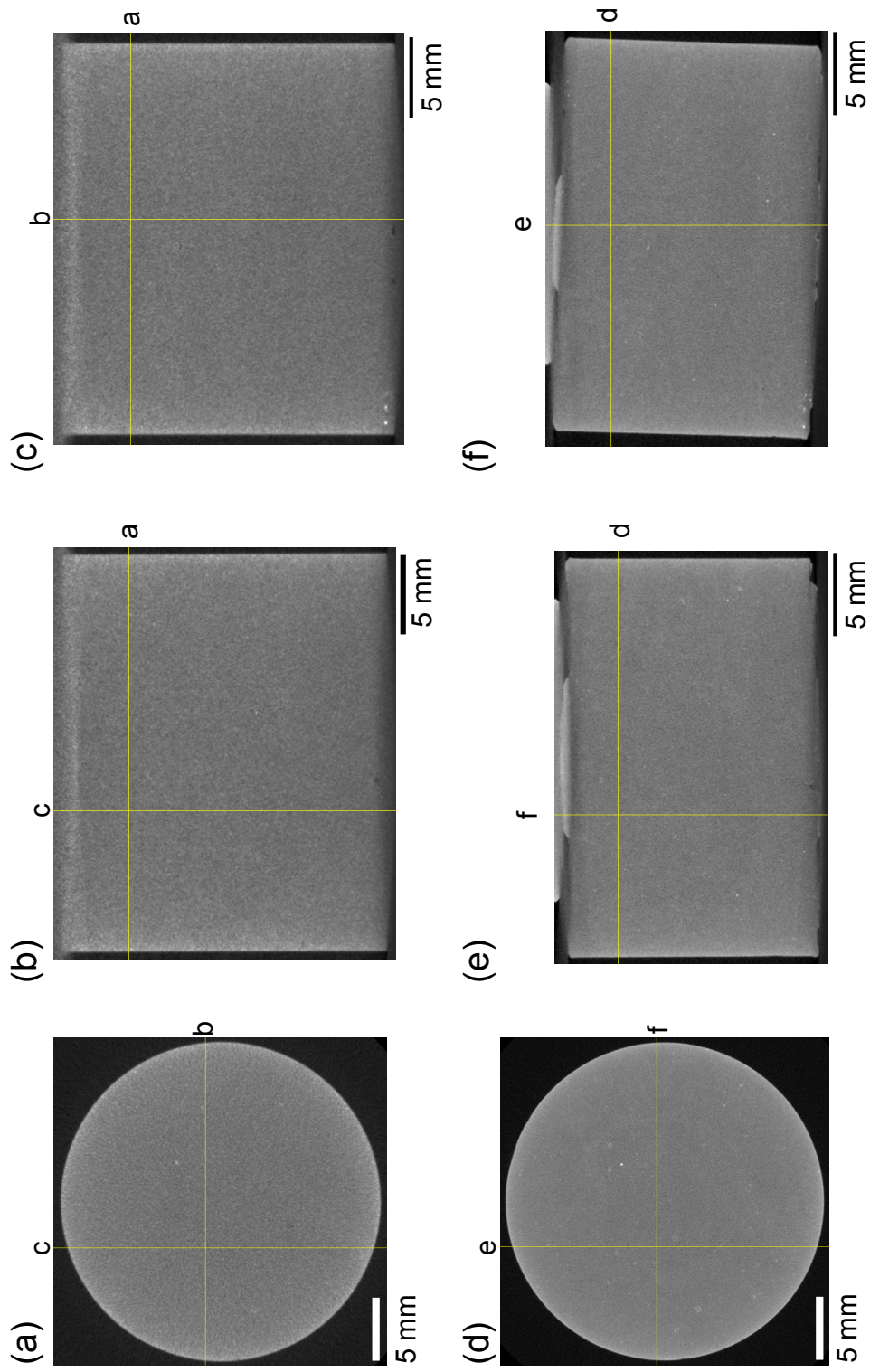


Figure 5.12 X-ray CT images taken before (a, b, c) and after (d, e, f) the consolidation test. The compressive axis is vertical in (b) (c) (e) and (f). The lines in each image show the position of section images.

Table 5.1 Physical properties of the specimens and results of the consolidation tests.

Sampling area	Sample ID	Formation name	Key bed	Thickness from top of the Kazasa Gp. (m)	Equipment	Maximum Compression Stress (MPa)	c_c	p_c (MPa)	Porosity (%)	Bulk density (g/cm ³)	Grain density (g/cm ³)	Softening
West	BosC-2-1	Umegase	U8	654	NIT	50	0.39	7.5	39.5	2.01	2.67	-
West	BosC-10-1	Otadai	O18	749	NIT	80	0.39	7.6	42.7	1.94	2.64	-
West	BosC-10-K	Otadai	O18	749	KU	80	0.37	8	40.9	1.92	2.62	-
West	BosC-9-5	Kiwada		893	NIT	80	0.35	9.6	40.1	2.00	2.67	-
West	BosC-8-1	Anno	An135	1080	NIT	80	0.43	23.7	39.7	2.06	2.76	-
West	BosC-13-1	Kiyosumi	Ky21	1462	NIT	80	0.50	15.5	43	1.93	2.64	-
West	BosC-3-1	Amatsu	Am14	2257	NIT	80	0.38	31.0	33.5	2.13	2.70	-
West	BosC-12-3	Kinone		2809	NIT	80	0.48	24.6	40.3	1.91	2.52	-
Central	BosC-16-1	Kiwada	Kd8	1615	NIT	80	0.38	11.7	41.1	1.96	2.64	-
Central	BosC-15-1	Anno	An22	1799	NIT	80	0.38	15.8	43.8	1.92	2.64	-
East	BosC-38	Kokumoto	Ku2	646	KU	80	0.32	2.8	44.6	1.90	2.63	-
East	BosC-39	Kokumoto	Ku3	730	KU	80	0.40	3.5	45.6	1.87	2.60	5.6
East	BosC-36	Kokumoto	Ku6	792	KU	80	0.41	2.0	49.4	1.80	2.59	5.0
East	BosC-37	Umegase	U6B	949	KU	80	0.37	1.5	49.0	1.81	2.59	4.1
East	BosC-35	Otadai	O7	1319	KU	80	0.35	5.9	41.0	1.95	2.62	(6.7)
East	BosC-34	Otadai	O11	1346	KU	80	0.42	6.4	48.1	1.82	2.58	10.0
East	BosC-28-1	Otadai	O27	1568	NIT	80	0.42	6.6	46.5	1.81	2.52	5.5
East	BosC-27-1	Kiwada	Kd18	1813	NIT	80	0.40	6.4	43.7	1.90	2.60	6.1
East	BosC-21	Kiwada	Kd24	1931	NIT	80	0.40	12.0	40.5	1.93	2.57	-
East	BosC-32_K4	Kiwada	Kd38	2076	KU	80	0.49	7.0	47.7	1.83	2.58	12.3
East	BosC-33-1	Kiwada	Kd40	2129	KU	64	0.20	7.7	47.8	1.84	2.61	8.1
East	BosC-26-1	Ohara	HSC	2260	NIT	80	0.54	9.4	49.6	1.79	2.56	12.3
East	BosC-26-K	Ohara	HSC	2260	KU	80	0.53	10.3	49.3	1.79	2.55	15.0
East	BosC-18	Ohara	Cream	2297	NIT	80	0.49	9.3	46.4	1.85	2.58	-
East	BosC-30-3	Namihana	Nm6	2522	NIT	80	0.53	6.9	50.8	1.77	2.57	10.4
East	BosC-20	Namihana	Aranami	2580	NIT	80	0.38	5.5	47.8	1.83	2.60	-
East	BosC-31-2	Katsuura	Asagimadara	2708	NIT	80	0.49	8.4	48.7	1.83	2.61	12.2
East	BosC-19	Katsuura		3072	NIT	80	0.42	20.0	36.5	2.05	2.66	14.1
East	BosC-24	Kiyosumi	Ky11-Ky14	3410	NIT	80	0.45	27.5	35.7	2.03	2.60	-
East	BosC-25	Amatsu	AM40-AM54	3990	NIT	80	0.30	32.2	34.3	2.08	2.64	-

5.4 Discussion

5.4.1 Strain softening

Consolidation tests were conducted on mudstone and siltstone samples taken from the Boso forearc basin, which are categorized as soft rocks. The consolidation curves of some specimens taken from the Kazusa Group in the eastern part of the basin showed a rapid stress drop just before yielding. Consolidation tests for two specimens (BosC26-N and BosC26-K) from the Otadai Formation were performed using the apparatus at Kyoto University and Nagoya Institute of Technology, respectively (Fig. 5.11). Similar stress reductions were observed in both tests, which shows that this rapid stress drop is reproducible.

The rapid stress drop revealed in this study can be regarded as strain softening. Strain softening has also been reproduced in the elastoplastic constitutive equation for soft rocks, and is generally considered to be related to a decrease in stiffness and reduction in yield surface (Adachi et al., 1998). In soil materials, unsaturated soils exhibit significant strain softening. Decreased strength due to a reduction in meniscus water is a cause of strain softening (Kido et al., 2017). However, in rock materials, strain softening is attributed to localization of deformation and softening of the rock material itself (i.e., deterioration due to destruction of the framework structure; Adachi & Oka, 1995). X-ray CT images of BosC26-N before and after the test showed that particularly large particles such as pumice and fossils were not present in the specimen. In addition, micro-cracks such as brittle fractures were not found in the specimen after the test (Fig. 5.12). Therefore, it is suggested that the strain softening is attributable to softening of the siltstone itself.

In general, siltstone has a relatively high clay mineral content. Because the friction coefficients of some clay minerals such as smectite are small (Ikari et al., 2009; Ujiie et al., 2013), the mineral component of siltstone and its mechanical behavior might be related to strain softening. However, the Otadai and Kiwada formations contain

insignificant amounts of montmorillonite, which is a type of smectite (Kashiwagi & Shikazono, 2005). The samples used in this study are siltstones taken from the forearc basin, whose development is related not only to consolidation but also tectonic effects, such as horizontal compaction associated with plate subduction. Therefore, it is possible that the strain softening in this study reflects the micro-cracks and internal structures that developed during siltstone formation. In future research, the micro-structures of the specimens immediately after strain softening should be observed, which would enable elucidation of the softening mechanism.

5.4.2 Contrasting consolidation trends in the western and eastern parts of the basin

Depth profiles of p_c , porosity, bulk density, and grain density are shown in Figs 5.13, 5.14, 5.15, and 5.16, respectively. Depth is the “equivalent depth” assumed to be equal to the thickness from the top of the Kazusa Group to each sample. In Fig. 5.13, the solid line shows the effective overburden pressure calculated from the bulk density of the rocks (Table 5.1) and water density (1.0 g/cm^3). The dashed line is the assumed “normal” consolidation trend in each study area. The slope of the dashed line is the same as the solid line. These profiles show that the distribution of consolidation yield stress differs between the eastern, central, and western parts of the basin.

In the western part of the basin, p_c values for the Kazusa Group above the Kurotaki unconformity are almost identical to the calculated effective overburden pressure, whereas those for the Miura Group are higher than expected from the overburden (Fig. 5.13a). The p_c of the Kiyosumi and Kinone formations are lower than the assumed normal consolidation trend of the Miura Group. Considering an increasing of the overburden pressure by the thickness, it is implied that consolidation was provided in these formations. The difference of the p_c values between two groups suggests that the thickness of the Miura Group prior to the erosion that formed the Kurotaki Unconformity was thicker than at present, and that the maximum burial depth of the Miura Group was greater than the depth calculated using the present stratigraphy. Based on the consolidation tests, if this variation in p_c was only due to different burial depths, then submarine erosion of a ca. 2000 m thickness of the upper Miura Group is expected to have occurred prior to deposition of the Kazusa Group.

In the central part of the basin, the consolidation yield stress of the Kiwada Formation is smaller than that of the Anno Formation (Fig. 5.13b). The slope of the dashed line is the same as the solid line, which suggests that the difference between the two consolidation yield stresses is due to the overburden pressure at each sampling

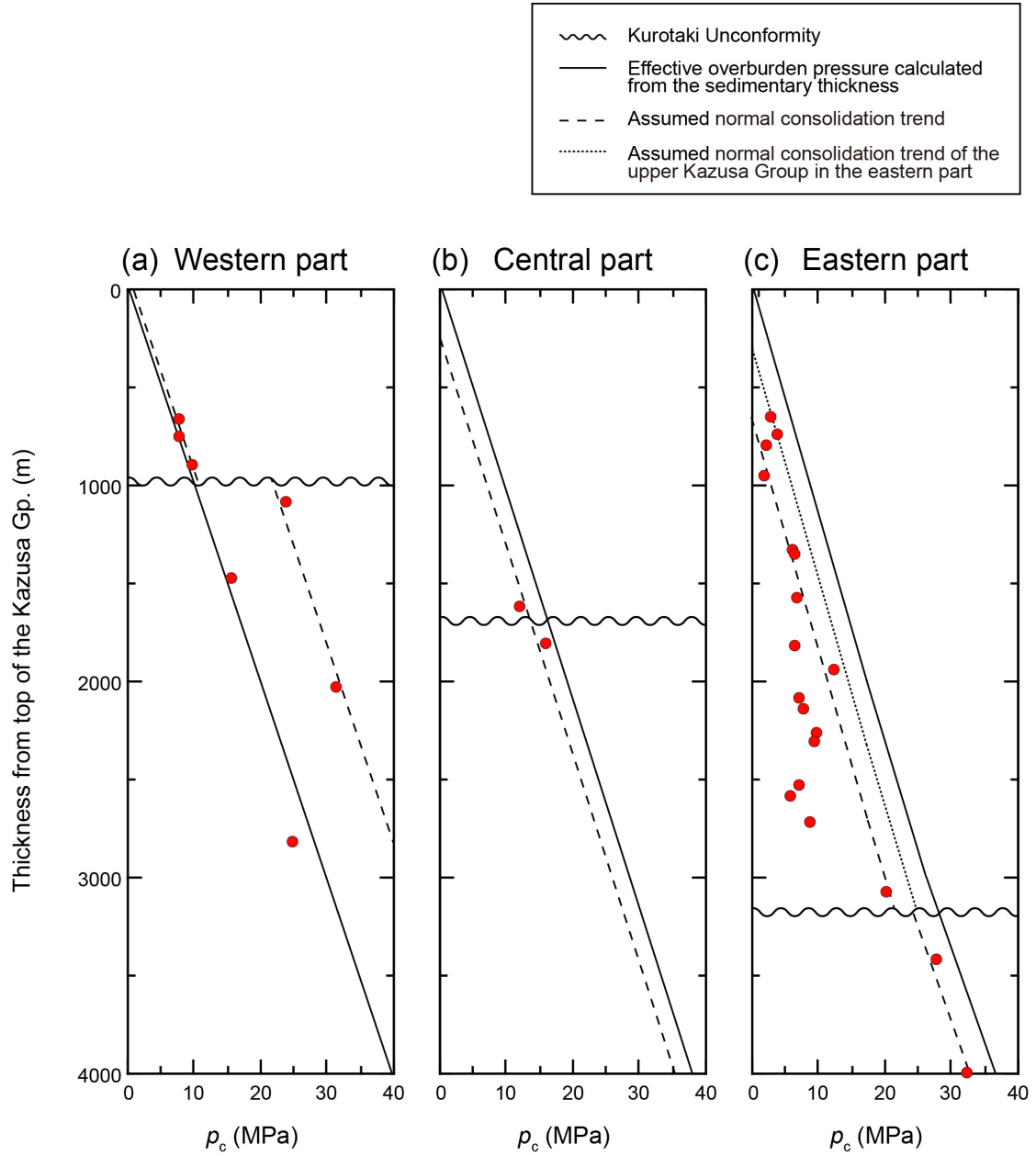


Figure 5.13 Depth profile of consolidation yield stress (p_c). The depth is the thickness from the top of the Kazusa Group to each sample. The western part (a), the central part (b), and the eastern part (c) of the basin.

location. The dashed line is ca. 4 MPa lower than the solid line, which implies that the upper part of the Kazusa Group was not deposited in the central part of the basin.

The “normal” consolidation trend in the eastern part of the basin is lower than the effective overburden pressure estimated from the assumed thickness, and differs by ca. 8 MPa (Fig. 5.13c). Although the Kurotaki Unconformity is located between the two groups, the consolidation trend is consistent across the unconformity. The difference between the effective overburden pressure and normal consolidation trends is equivalent to ca. 800 m of overburden thickness. This disparity can be explained by the reduced overburden thickness of the upper Kazusa Group in this area, as the upper Kazusa Group was not deposited here due to tilting of the basin. The p_c values reach the effective overburden pressure in the Kokumoto Formation, however, the p_c values for the upper part of the Kokumoto Formation are 4 MPa smaller than the calculated overburden pressure. This implies that the eastern part of the basin had started tilting and the depocenter changed, which meant that the uppermost part of the Kazusa Group (i.e., that part above the Umegase Formation) was not deposited in this region at the time the Kokumoto Formation was being deposited in the western part of the basin.

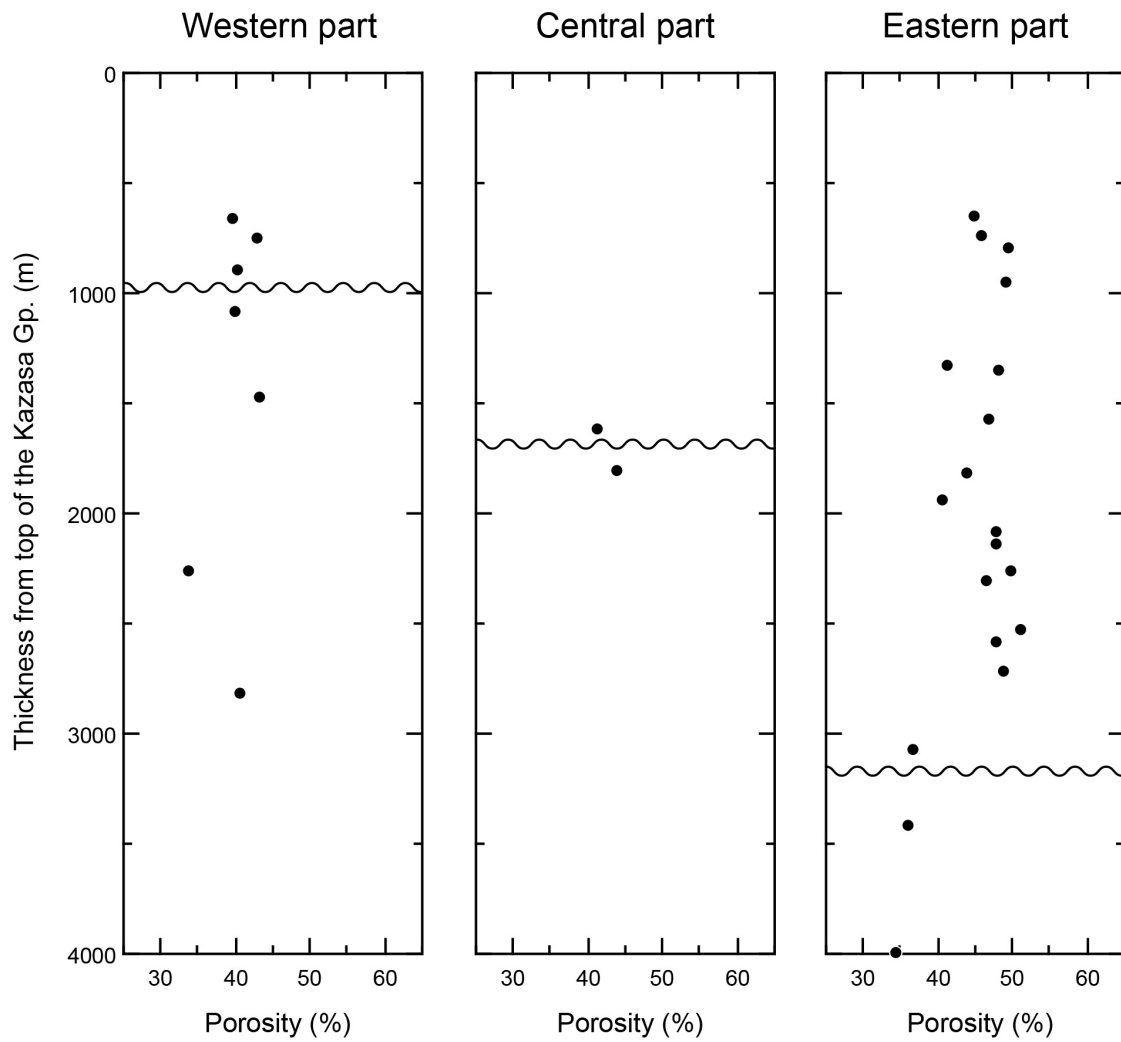


Figure 5.14 Depth profile of initial porosity of the specimens used for the consolidation tests. The “depth” is assumed to be equal to the thickness from the top of the Kazusa Group to each sample.

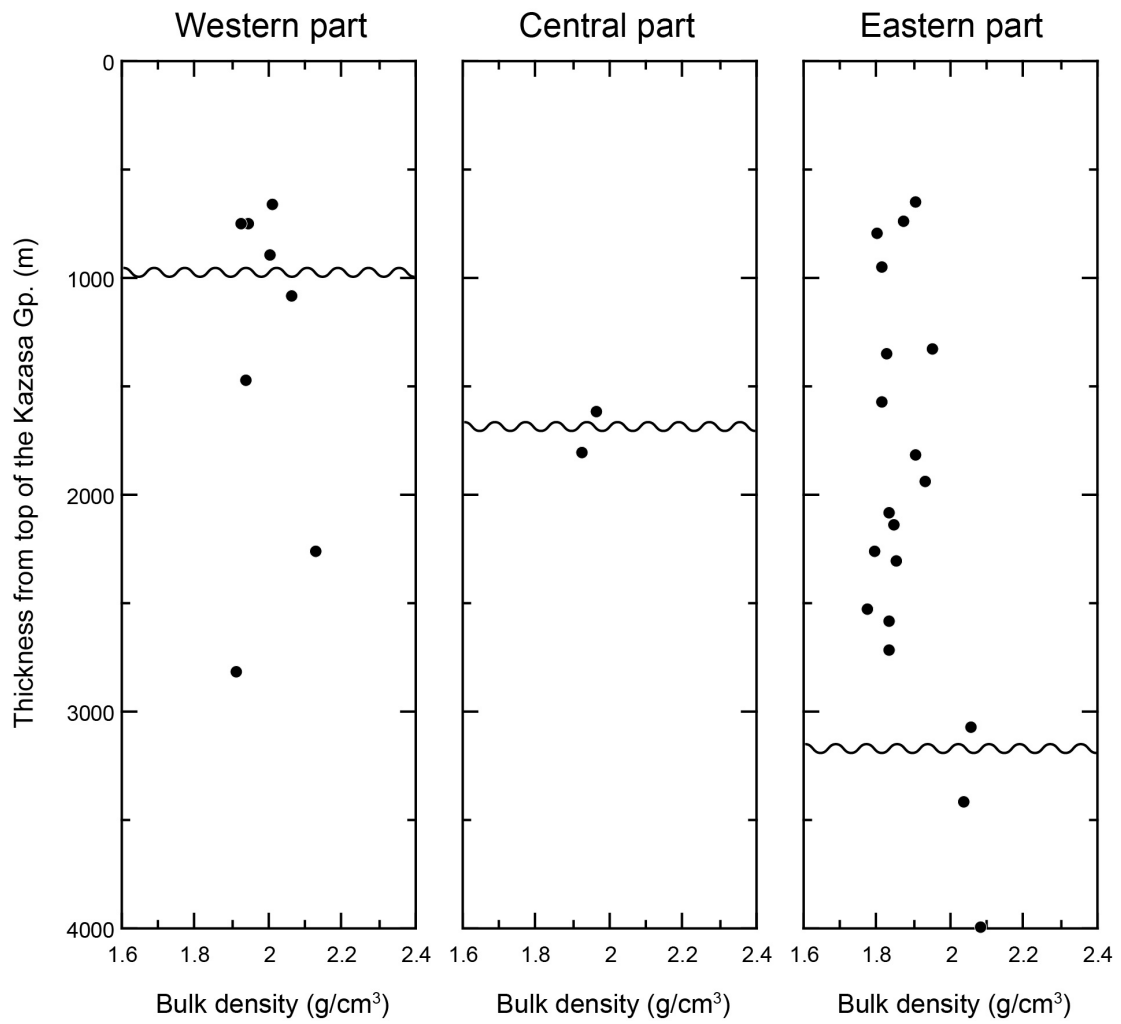


Figure 5.15 Depth profile of initial bulk density of the specimens used for the consolidation tests.

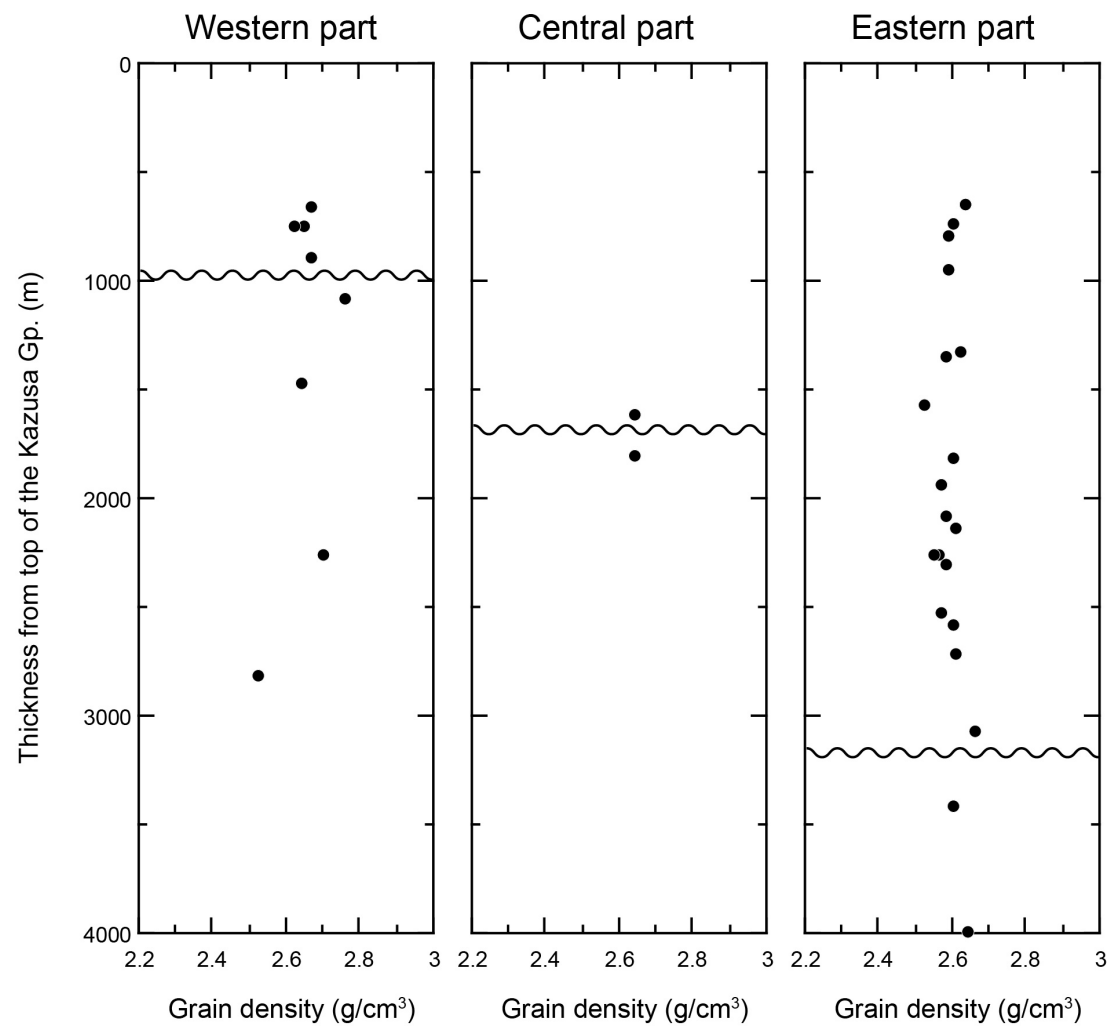


Figure 5.16 Depth profile of grain density of the specimens used for the consolidation tests.

5.4.3 Paleo-excess pore pressure zone in the eastern part of the basin

In the eastern part of the basin, the consolidation yield stresses of the Kiwada, Ohara, Namihana, and Katsuura formations are lower than the consolidation trend of this part of the basin (Fig. 5.13c). This discrepancy suggests that these formations have been prevented from undergoing consolidation. The zones of low p_c values correspond with excess-porosity zones, and excess pore fluid pressure can decrease the effective stress and then hinder consolidation. Geological features indicative of liquefaction, fluidization, and dewatering (Fig. 2.6) identified in these zones also support this interpretation. The excess pore fluid pressure is estimated to be 5–13 MPa, as calculated from the difference between the assumed normal p_c trend and the p_c values of the Kiwada, Ohara, Naminaha, and Katsuura formations (Fig. 5.13c).

High slope angles, earthquakes, high sedimentation rates, and dissociation of gas hydrates are factors that control the generation of submarine landslides. However, large submarine landslides also occur in deepwater environments with a low-gradient seafloor, low sedimentation rates, and a lack of gas hydrates (Van Der Merwe et al., 2011; Urlaub et al., 2012). The generation of excess fluid pressure is, therefore, key to understanding slope instability. Overpressure generation on a stable slope is controlled by the sedimentation rate and configuration, sediment consolidation, and stratigraphic features (Urlaub et al., 2015). Flemings et al. (2008) also demonstrated in the offshore Mississippi delta that overpressure near the seabed reduces stability and provides the trigger for submarine landslides on a slope with a low regional gradient (2°). In the Kazusa Group, submarine landslides also occurred on a low-gradient slope (0.31° – 0.46°), although the safety factor based on a nomogram which has been constructed for analyzing slope stability in fine-grained sedimentary environments (Booth et al., 1985) requires the critical slope angle to be $>1.2^\circ$ (Fukuda et al., 2015). Although relatively rapid sedimentation rates (ca. 0.5 and 2.0–3.0 m/kyr in the lower and upper Kazusa Group, respectively; Kazaoka et al., 2015) were previously reported, no evidence of

excess fluid pressure reducing the effective normal stress had been recognized. The excess fluid pressure identified in this study prevented the consolidation of sediments in the MTD-rich interval in the Pliocene–Pleistocene Boso forearc basin. This interval possibly formed due to the presence of an impermeable layer, rapid sedimentation, and minor fluidization and liquefaction of the sediments. These characteristics resulted in low effective stresses in the Boso forearc basin, and enabled small tectonic events to generate MTDs.

The Otadai and Umegase formations are the major reservoirs of natural gas dissolved in water in the Boso Peninsula (Kunimatsu et al., 2002). Assumed porosity trend along the low porosity data, the porosities of the lower Kazusa Group in the eastern part is lower than the trend and the reservoir formations also correspond to the low p_c and high porosity zone (Fig. 5.17). Inami (1983) suggested that the high porosity zone formed due to the anomalously high pressure, natural gas generation, and consolidation and deformation of strata. This study suggested the generation of the excess pore fluid pressure, which also supported the generation of the higher pressure of the pore space than the hydrostatic pressure.

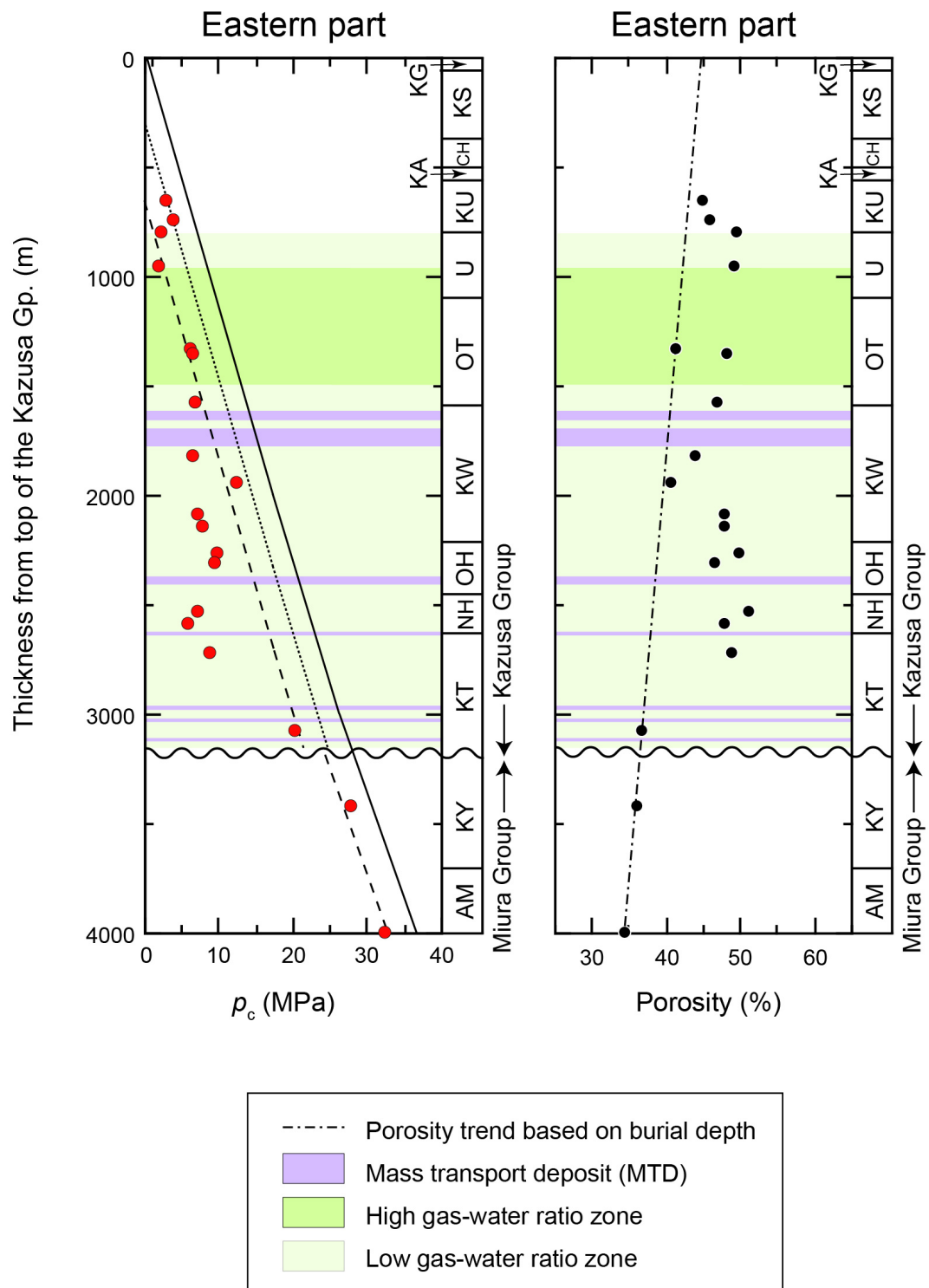


Figure 5.17 Distribution of mass transport deposits (MTDs) and natural gas reservoirs in the eastern part of the Boso forearc basin.

5.5 Conclusions

I performed consolidation tests on soft sedimentary rocks from the Boso forearc basin to estimate the maximum overburden pressure. The normal consolidation trends for the western, central, and eastern parts of the basin have different characteristics. A difference in the trend was confirmed between the Miura and Kazusa groups in the western part of the basin, which suggests that the eroded thickness in the western part of the basin is larger than the expected thickness. The normal consolidation trends in the central and eastern parts of the basin are ca. 4 and 8 MPa lower than the calculated overburden pressure using the sediment thickness, respectively. There is no difference in the consolidation trend between the two groups. In addition, strain softening was identified in the eastern part of the basin. Considering that micro-cracks such as brittle fractures were not found in the specimens after the tests, the strain softening observed in this study is attributed to softening of the siltstone itself. In the eastern part of the basin, consolidation yield stresses for the Kiwada, Ohara, Namihana, and Katsuura formations are lower than the normal consolidation trend. This, and the distribution of mass transport deposits, suggest the pore pressure was higher than the hydrostatic pressure in these formations.

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Chapter 6

Evolution of the Boso forearc basin

6.1 Basin evolution

6.1.1 Evolution of the western part of the basin

The normal consolidation trend in the western part of the Miura Group differs from that of the Kazusa Group (Fig. 5.13a). If this variation in p_c is only due to different burial depths, submarine erosion of a ca. 2000 m thickness of the upper Miura Group is expected to have occurred prior to deposition of the Kazusa Group. However, given the sedimentation rate in the western part of the basin changed from 0.15 to 0.33 m/kyr at ca. 4 Ma and became 0.50 m/kyr after the Kurotaki Unconformity formed (Fig. 6.1), an erosion thickness is estimated 500–800 m. Compared with the Kazusa Group, a denser population of map-scale folds has been identified in the Miura Group (Ishihara & Tokuhashi, 2001; Nakajima & Watanabe, 2005). Given the variations in p_c values, the Miura Group apparently experienced tectonic compression before sedimentation of the Kazusa Group (Fig. 6.2). In addition, given the present thickness of the Kazusa Group (ca. 1000 m) and the 500–800 m of erosion of the Miura Group, advective heating prior to the unconformity forming appears to have been responsible for the variations in thermal maturity. Given the almost constant vitrinite reflectance values in the Miura Group (0.38%–0.44%), regardless of stratigraphic position, a high geothermal gradient is unlikely. The advective heating was potentially associated with emplacement of the ophiolite complex that is now located just below the Miura Group. Therefore, both tectonic compression and advective heating apparently affected the Miura Group. Part of the Miura Group was eroded during the interval of time represented by the Kurotaki

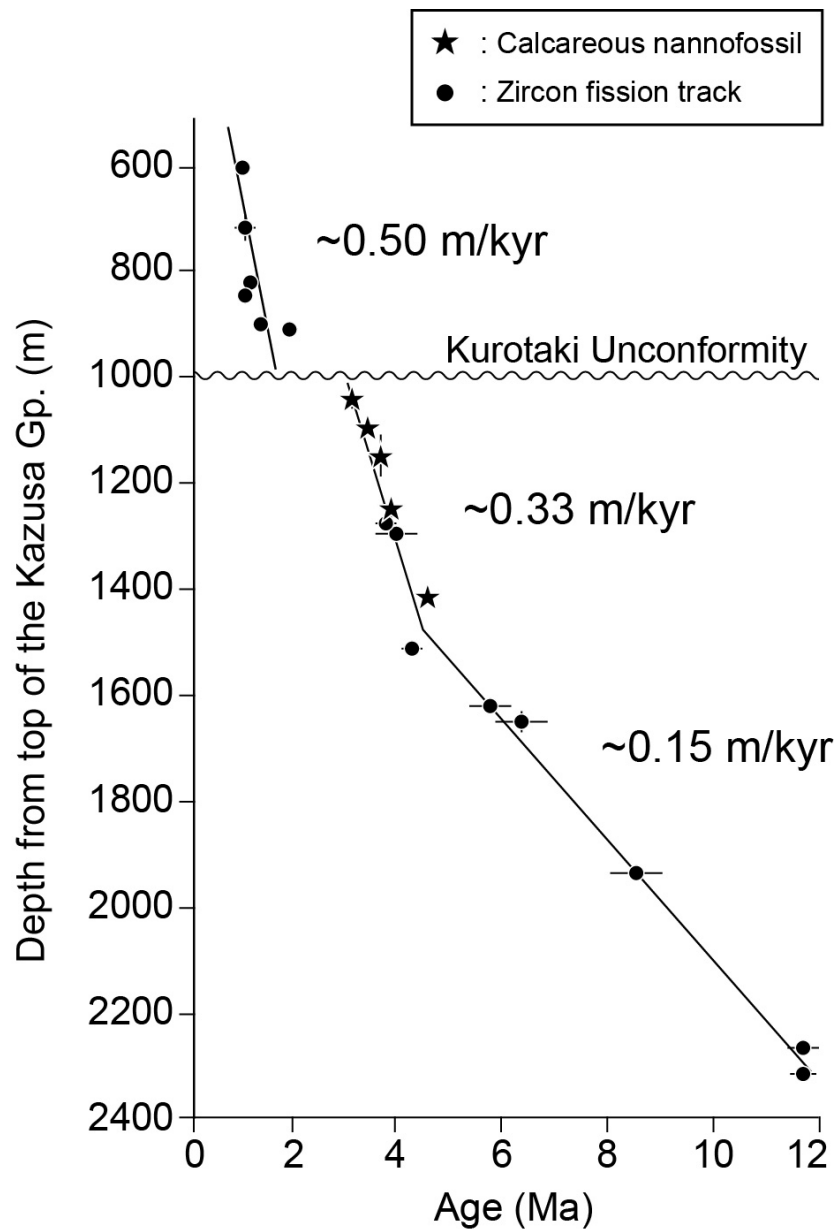


Figure 6.1 Age–depth diagram showing the sedimentation rate in the western part of the basin. Thickness of strata (Nakajima & Watanabe, 2005), calcareous nannofossils (Kameo & Sekine, 2013), and zircon fission-track (Watanabe & Danhara, 1996; Takahashi & Danhara, 1997; Tokuhashi et al., 2000; Tsuji et al., 2005; Tamura et al., 2019) data were compiled for this figure.

Unconformity, particularly in the eastern part of the Boso Peninsula, where the Kurotaki Formation overlies the Kiyosumi Formation.

The sedimentation rate increased at ca. 4 Ma when the Anno Formation was deposited (Fig. 6.1). The sedimentation rate was related to consolidation, which in turn is associated with dehydration, which makes the sediment volume decrease. However, consolidation does not occur when dehydration is prevented. When the sedimentation rate increases, dehydration is generally prevented. Given that the sedimentation rate of the Anno Formation increased at ca. 4 Ma, excess pore pressure was generated in the Kiyosumi Formation in the western part of the basin.

Major differences above and below the Kurotaki Unconformity were identified in terms of the distribution of geological units, structures, maximum paleo-temperatures, and consolidation characteristics. These differences probably relate to a change in the convergence direction of the Philippine Sea Plate at ca. 3 Ma. Uplift and erosion of the Miura Group during the formation of the unconformity, and compression parallel to the direction of plate subduction apparently caused the variations in the attitudes of the strata. The forearc basin, particularly near the trench–slope break, records structural and physical properties reflecting plate tectonic processes and the development of the trench–slope.

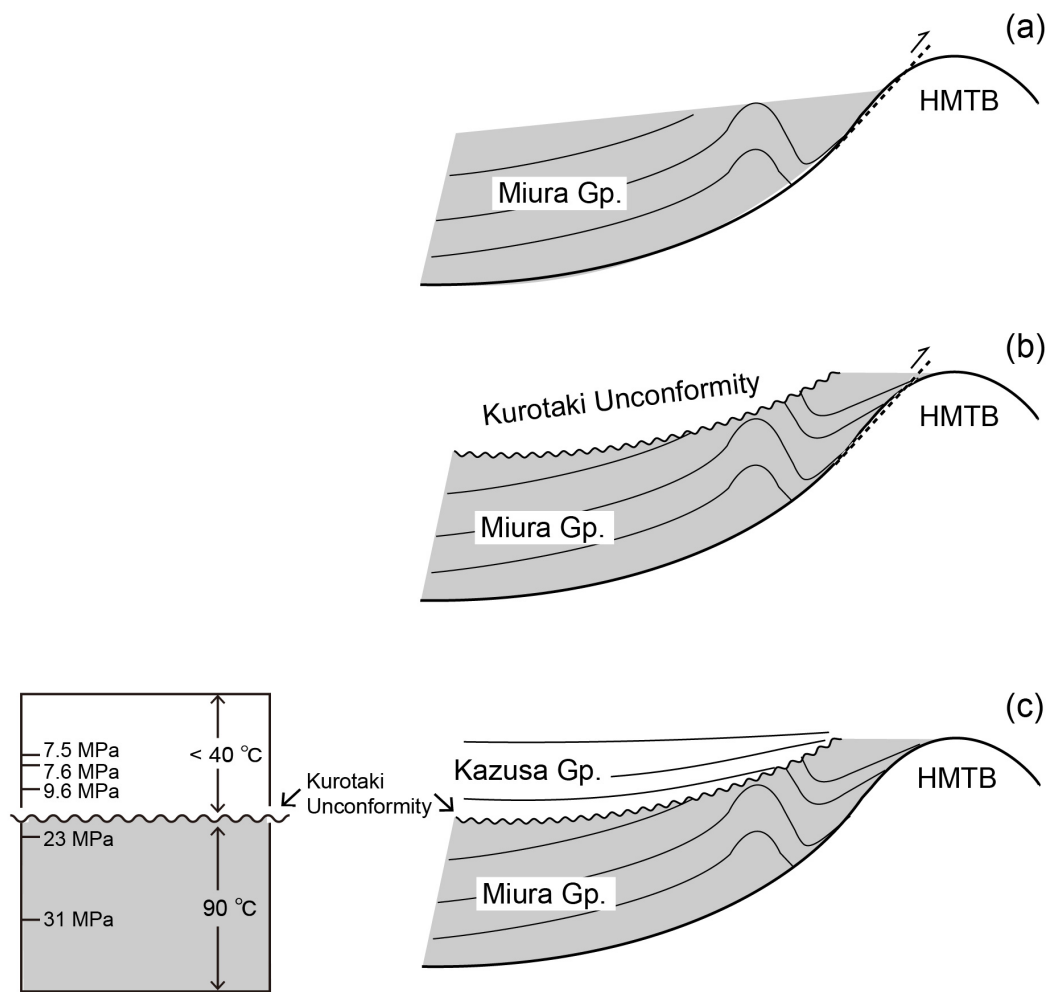


Figure 6.2 Schematic figure showing the evolution of the western part of the Boso forearc basin. (a) Sedimentation of the Miura Group. Possible advective heating from the ophiolite complex (HMTB; Hayama–Mineoka Tectonic Belt) occurred. (b) The Miura Group was tectonically compacted, uplifted, and eroded (i.e., the Kurotaki Unconformity formed). (c) The Kazusa Group was deposited above the unconformity. The maximum paleo-temperature and burial depth deduced from vitrinite reflectance data and the consolidation tests are also shown.

6.1.2 Evolution of the eastern part of the basin

The distribution of the consolidation yield stress in the eastern part of the basin indicates that the burial of the forearc basin occurred regardless of the development of the Kurotaki Unconformity, as ascertained from the patterns of p_c for the Kazusa and Miura groups above and below the unconformity (Fig. 5.13c). The Miura Group was deposited until 3 Ma, and the overburden pressure increased with sediment thickness, which led to progressive consolidation and an increase in consolidation yield stress with burial depth (Fig. 6.3a). The Kurotaki Unconformity then formed at ca. 3.0 Ma. At that time, the upper Miura Group was eroded, and the middle Kiyosumi Formation was exposed on the ocean floor (Fig. 6.3b). After ca. 2.3 Ma, the Kazusa Group began to be deposited on top of the Miura Group. A break in the trend of the consolidation yield stress was formed during the early depositional stage, following the consolidation of the Kazusa and Miura groups, due to the erosion at 3.0 Ma that produced the Kurotaki Unconformity (Fig. 6.3c-1).

As the thickness and burial depth of the Kazusa Group increased, the p_c of the group also increased, whereas that of the Miura Group did not increase because it was within the over-consolidation area (Fig. 6.3c-2). When the burial depth of the Miura Group reached the maximum depth, the difference in p_c between the Kazusa and Miura groups was overprinted (Fig. 6.3c-3). If burial had continued, the p_c of the Miura and Kazusa groups would have increased together (Fig. 6.3d). However, considering that p_c values in the Miura Group in the western and eastern parts of the basin are almost identical, it is suggested that burial ceased at approximately this time. In the eastern part of the basin, vitrinite reflectance also shows no break in trend or values across the Kurotaki Unconformity. Given this feature is similar to the p_c trend, any differences in vitrinite reflectance across the unconformity are thought to have been overprinted by the burial process.

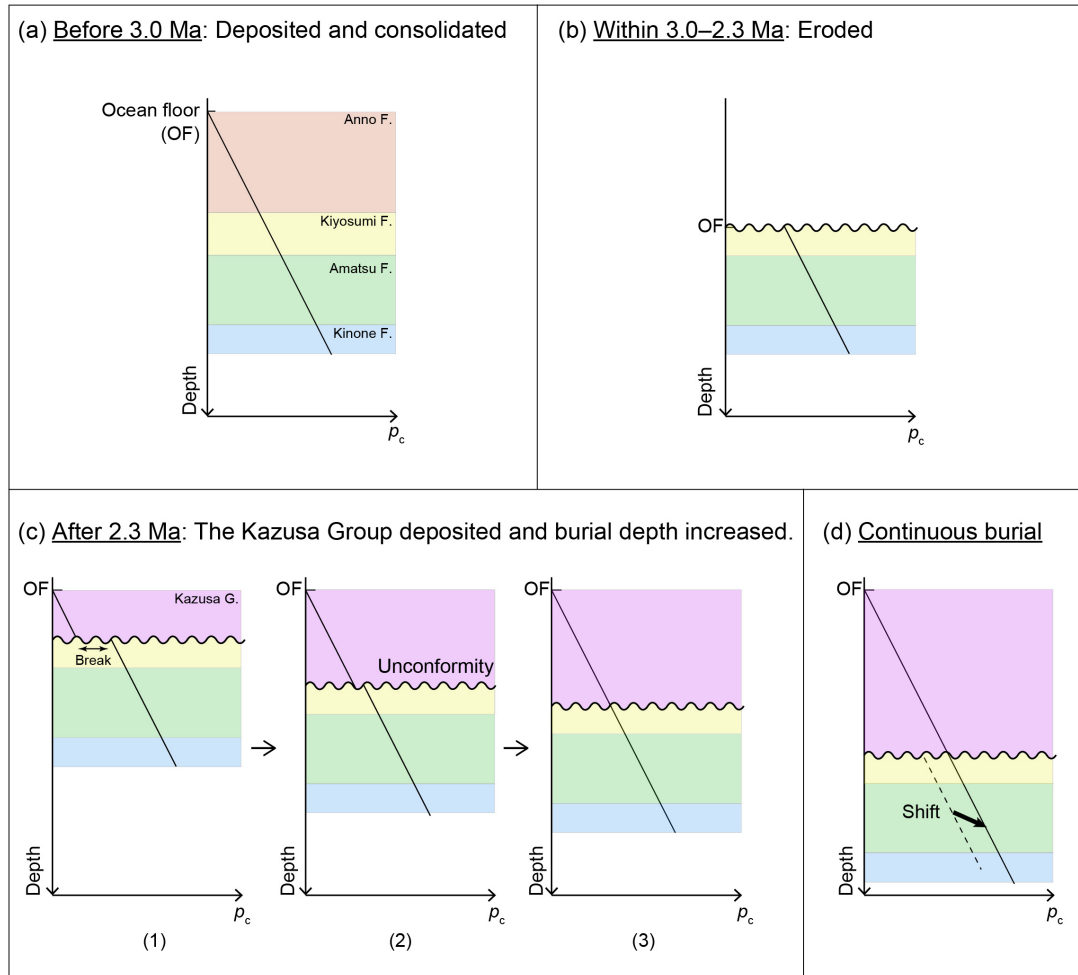


Figure 6.3 Schematic figure showing the evolution of p_c values in the eastern part of the Boso forearc basin. (a) Before 3.0 Ma: regular deposition and consolidation of the Miura Group. (b) 3.0–2.3 Ma: the Kurotaki Unconformity formed due to erosion. (c) After 2.3 Ma: the Kazusa Group was deposited and the burial depth increased. (d) Continued burial. The dashed line is the previous p_c trend shown in Fig. 6.3c-3.

The development of the Boso forearc basin has been previously investigated in terms of tephra distribution and paleocurrents (e.g., Ito, 1994). Analysis of the paleobathymetry using benthic foraminifera indicates that the paleo-depth of the ocean floor during deposition of the Kazusa Group was constant (ca. 1500 m) until ca. 1.3 Ma, when the eastern part of the Boso forearc basin began to be uplifted (Kitazato, 1997). Given that the paleo-depth of the ocean floor remained constant during deposition, which suggests the basin was subsiding, the eastern part of the Boso forearc basin is inferred to have undergone subsidence until 1.3 Ma. The absence of any difference in p_c across the Kurotaki Unconformity in the eastern part of the basin and the difference in p_c trend between the eastern and western parts of the basin implies that subsidence of the eastern part was greater than that of the western part.

When the Kokumoto Formation was deposited, the normal consolidation trend changed (Fig. 5.13), which suggests the basin was uplifted or tilted, and the depositional environment changed. Considering the sedimentary age of the Kokumoto Formation, this occurred at ca. 0.8 Ma.

6.2 Plate subduction and evolution of the Boso forearc basin

The migration of the Euler pole of the Philippine Sea Plate to its present location should have occurred at 3 Ma, because deposition in the forearc basin in southwest Japan was interrupted, as represented by the Kurotaki Unconformity, at 3 Ma (Takahashi, 2006). In addition, Yamaji (2003) reported NNE–SSW-trending normal faults in the Kazusa Group in the eastern part of the basin, and suggested that the area was subjected to a WNW–ESE-directed tensile stress at ca. 1 Ma. These authors proposed that changes in the movement direction of the Philippine Sea Plate at ca. 3 and 1 Ma controlled the erosional and tensile faulting events.

At ca. 1 Ma, the tectonism was related to collision of the Izu-Bonin Arc (Sugiyama, 1992). The trench–slope basin deposition was occurring in the northern part of the Boso Peninsula, similar to the forearc basin sedimentation (Kawakami & Shishikura, 2006). The Toyofusa Group, which is part of the trench–slope basin distributed on accretionary complexes, contains an unconformity that formed at ca. 0.78 Ma (e.g., Kotake et al., 1995). The deposition in the trench–slope basin was controlled by plate tectonics and basin evolution related to the formation of the accretionary complexes.

Bedding in both the Miura and Kazusa groups in the western part of the Boso forearc basin is E–W-striking, however, the attitude of the mesoscale reverse faults is different in the two groups. Paleo-temperatures are also different in the Miura and Kazusa groups; the Miura Group experienced advective heating from the Mineoka ophiolite, whereas the paleo-temperature of the Kazusa Group was controlled by the geothermal gradient and burial depth. The consolidation trend shows a distinct boundary at ca. 3 Ma, and there is a difference in the consolidation trend between the Miura and Kazusa groups in the western part of the Boso forearc basin, which implies erosion related to the formation of the Kurotaki Unconformity. However, the consolidation trends show no differences between the two groups in the eastern part of the basin, which is related to the burial progress of the eastern Kazusa Group.

Considering that the maximum burial depth before ca. 3 Ma was similar for the eastern and western parts of the Boso forearc basin, the inferred difference in subsidence after ca. 3 Ma suggests that tectonism caused this change. Numerical modeling has indicated that the distribution of sedimentation and uplift differs depending on the shape and angle of the subducting plate (Hashima et al., 2016), which suggests the basin evolution would vary depending on the location.

In addition, the consolidation yield stress increase and porosity decrease in the Kokumoto Formation (age of ca. 1 Ma) implies the depositional environment changed at ca. 1 Ma. These coeval tectonic changes in the eastern Boso forearc and trench–slope basins imply that the Boso forearc basin developed in associations with an accretionary prism due to plate subduction.

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Chapter 7

Conclusions

Forearc basins are formed by plate subduction and control the generation of a range of natural resources. This thesis investigated the geological structure, paleo-thermal history, and consolidation state of a Miocene–Pleistocene forearc basin in the Boso Peninsula, central Japan, and developed an evolutionary model for this basin. The Boso forearc basin contains the Kurotaki Unconformity, which occurs widely along the Nankai Trough in southwest Japan. The sediment age and thickness differ between the Miura and Kazusa groups in the eastern and western parts of the basin, respectively. Considering these geological features, I focussed on sedimentary processes and the burial rate of the basin at the time of formation of the Kurotaki Unconformity. Geological fieldwork, vitrinite reflectance measurements, and consolidation tests were performed to investigate the evolution of the Boso forearc basin. The main conclusions obtained from each chapter are described below.

In Chapter Three, the results of geological fieldwork suggest that bedding in the western part of the Boso forearc basin is controlled by E–W trending folds and that deformation was more intense in the southern part of the basin, close to the trench–slope break (i.e., the Hayama–Mineoka Tectonic Belt). Mesoscale faults are widely distributed and most of the normal and reverse faults in the forearc basin are now healed, which suggests these faults formed shortly after sedimentation. The fault attitudes are different above and below the Kurotaki Unconformity; i.e., both E–W- and NE–SW-trending faults are present in the Miura Group below the unconformity, but only NE–SW-trending faults are present in the Kazusa Group above the unconformity. Fabric analysis using anisotropy of magnetic susceptibility data suggests that both the Miura and Kazusa groups have a generally consolidated fabric. However, the fabric of the Miura Group is weaker than that of the Kazusa Group.

In Chapter Four, the paleo-temperatures of the Miura Group were shown to be ca. 70–90°C and essentially uniform across the western and eastern parts of the basin. In the central part of the basin, part of the Kinone Formation has the highest paleo-temperature of >100°C. Considering the location and temperature of the Mineoka ophiolite complex, it is suggested that advective heating occurred during emplacement of the ophiolite, which heated the Miura Group. The paleo-temperature of the Kazusa Group was ca. 50°C in the western part of the basin, whereas it was 75–90°C in the eastern part of the basin. Therefore, the paleo-temperatures differ above and below the Kurotaki Unconformity in the western part of the basin. The contrast in the paleo-temperature distribution between the western and eastern parts of the basin implies rapid sedimentation and the continuous and thick accumulation of basin-fill sediments in the forearc basin.

In Chapter Five, consolidation tests were presented that indicate the normal consolidation trends of the western, central, and eastern parts of the basin have different characteristics. A difference in the normal consolidation trend was identified between the Miura and Kazusa groups in the western part of the basin, which suggests the eroded thickness in the western part of the basin is larger than expected. The normal consolidation trends in the central and eastern parts of the basin are ca. 4 and 8 MPa lower than the calculated overburden pressure using the sediment thickness, respectively, whereas there is no difference in the consolidation trend between the two groups. In addition, strain softening was identified in the eastern part of the basin. Considering that micro-cracks such as brittle fractures were not found in the specimens after testing, it is suggested that the strain softening is attributable to softening of the siltstone itself. In the eastern part of the basin, the consolidation yield stresses for the Kiwada, Ohara, Namihana, and Katsuura formations are lower than the normal consolidation trend. Considering the distribution of mass transport deposits (MTDs), it is suggested that the pore fluid pressure was higher than the hydrostatic pressure (i.e., an over-pressured state).

In Chapter Six, an evolutionary model for the Boso forearc basin was presented, based on the results of Chapters Three, Four, and Five. In the western part of the basin, major differences above and below the Kurotaki Unconformity were identified in the distributions of geologic bodies and their structures, paleo-maximum temperatures, and consolidation characteristics. Uplift and erosion of the Miura Group during the formation of the unconformity, and compression parallel to the direction of plate-subduction apparently caused the variations in the attitudes of geologic bodies. The forearc basin, especially near the trench-slope break, records structural and physical properties reflecting the plate tectonics and development of the trench-slope. The absence of breaks in trends of vitrinite reflectance and consolidation yield stress across the Kurotaki Unconformity in the eastern part of the basin suggest that any such breaks were overprinted during the progressive deposition and burial of the Kazusa Group. The continuous variation in vitrinite reflectance corresponding the paleo-temperature and the consolidation trend in the eastern part of the basin between the Miura Group (below the unconformity) and Kazusa Group (above the unconformity) can be explained by the greater subsidence in the eastern part of the Boso forearc basin since ca. 2.3 Ma. If subsidence and uplift around the Boso Peninsula were controlled by subduction, then the variable evolution of the western and eastern parts of the Boso forearc basin can be used to investigate subduction processes.

This integrated study of field observations, paleo-temperature determinations, and consolidation tests has constrained the evolution of the Boso forearc basin. The estimates of paleo-burial depths allowed evaluation of the paleo-excess pore pressure. For numerical modelling purposes, vitrinite reflectance and consolidation yield stress data are vital for investigating the basin evolution. Basin modelling using such data can reveal the details of basin development, and constrain the timing of each deformation event that in turn is related to subduction. Given that basin evolution is also associated with natural resources, this is also important for understanding the formation mechanisms of natural reservoirs.

PhD. Dissertation

Evolution of the Mio–Pleistocene forearc basin
induced by the plate subduction
in the Boso Peninsula, central Japan

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