Heterogeneous Sediment Input at the Nankai Trough Subduction Zone: Implications for Shallow Slow Earthquake Localization

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Abstract Subducted sediment plays a key role in modulating pore pressure and seismic behavior at subduction zones. We investigated the seismic character of incoming sediments to test how sediment and basement variations relate to the along-strike changes within the accretionary prism and plate boundary conditions at the Nankai Trough. High-resolution seismic data reveal for the first time the presence of contourite mounded drifts in the Shikoku Basin. These features have probably introduced permeability heterogeneities into an otherwise homogenous mud-dominant unit. Additionally, we found that normal faults in this unit are more extensive than previously documented, which probably enhances along-strike fluid transport. The wedge taper is more correlative with the thickness of the mud-dominant facies than the turbidite thickness. This may be due to the permeability heterogeneities associated with contourite deposits and normal faults, or due to the absence of thick turbidite deposits. Turbidite deposits can either aid the drainage of the margin where they are not confined by basement topography, or contribute to high pore fluid pressures where they are confined by less permeable mudstone or basement topography. When confined turbidite deposits and contourite mounded drifts are subducted, they may contribute to localized compartments of excess pore pressure which provide the necessary conditions for slow slip behavior. We determined that along-strike variations in seismic behavior are likely related to the subducting basement topographic and sediment characteristics, which vary on a local (<10 km) scale.

Plain Language Summary Subduction zone faults, such as those off the coast of southwest Japan, have been shown to slip at a range of speeds. The ruptures range from typical earthquakes, which occur within seconds or minutes, to slow earthquakes, which can rupture over the course of days or weeks. The sediment characteristics of the incoming plate play a key role in determining the type of slip behavior along a margin. When sediments are subducted, they are compressed and expel the fluid trapped in the sediment. If the fluid is unable to escape, the fluid pressure builds up, and can reduce the strength of the plate boundary fault. The type of sediment, as well as whether it has been faulted or modified by the ocean currents, can control how well fluid can escape. The sediment characteristics and corresponding plate boundary strength vary greatly on a sub-10 km scale. Slow earthquakes in our study area are located with subducted seamounts. We suggest that this is because the sediments commonly deposited on the sides of large seamounts are conducive to forming high pore fluid pressures upon subduction. This result in a locally weaker plate boundary fault, which may be a necessary condition for slow earthquakes.

1. Introduction

Along-strike variations in the structural architecture and accretionary prism taper angle of subduction margins correlate spatially with changes in the seismogenic behavior of the plate interface (Bassett et al., 2014; Fujie et al., 2013; McNeill & Henstock, 2014). Lateral variations in the roughness of the incoming igneous crust and the overlying sediment types and cumulative thickness are thought to modulate the along-strike variations in accretionary wedge mechanics (Saffer & Bekins, 2006; Sun, Ellis, & Saffer, 2020; Sun, Saffer, & Ellis, 2020). The taper of the subduction wedge is used as a proxy to indicate fault strength (Cubas et al., 2013; Dahlen, 1990; D. Davis et al., 1983; Koge et al., 2014). Pore pressure within the subducting...
sedimentary section influences both the wedge mechanics and taper, as well as various fault-slip behaviors, ranging from stable sliding to slow slip events (SSEs), large earthquakes, and tsunamis (e.g., Kameda et al., 2015; Saffer & Bekins, 2006; Saffer & Tobin, 2011; Saffer & Wallace, 2015; Seno, 2009; Tobin & Saffer, 2009). Field studies, laboratory experiments, and conceptual models suggest that heterogeneous subducting sediments may be responsible for transitional slip behavior (e.g., Barnes et al., 2020). If that concept is valid, then detailed knowledge of margin-wide variations in lithostratigraphy and facies architecture is needed to test site-specific cause and effect.

Basement topography on the subducting plate is one of the key variables that control the type and thickness of sediment entering subduction zones (Barnes et al., 2020; Spinelli & Underwood, 2004; Underwood, 2007; Underwood et al., 2005). Other important variables include rates of biogenic productivity and proximity to major sources of terrigenous sediment (i.e., large rivers, submarine canyons, and associated deep-water fans). Large basement relief can block flow paths or deflect turbidity currents seaward of the trench, resulting in along-strike variations in facies (e.g., Macdonald, 1993). The variations in sediment thickness and composition, as well as the stratigraphic position of the décollement and the thermal state of the subduction lithosphere, govern the three-dimensional compaction, consolidation, and diagenesis patterns (Spinelli & Underwood, 2004; Underwood et al., 2005). These factors also control the hydraulic connectivity between the basement, the plate boundary, and the seafloor, thereby modulating the pore pressure and the shear strength of the accreted sediment and the décollement (Saffer & Tobin, 2011).

The Nankai Trough offshore Japan (Figure 1) has been the focus of decades of geological and geophysical research, making it an ideal laboratory to study margin-scale variations in lithostratigraphy (J. C. Moore & Karig, 1976; G. F. Moore et al., 2001a; Pickering et al., 1993; Underwood & Moore, 2012). The sedimentology and basement topography of Shikoku Basin are more complicated than for most comparable incoming plates (e.g., McNeill et al., 2017; Spinelli & Underwood, 2004; Underwood et al., 2005). In most previous studies, the Nankai Trough was broadly subdivided into three geographical regions (Figure 2), with the maximum basement relief occurring in the central (Muroto) area in the vicinity of an extinct spreading ridge and remnant seamounts (Ike et al., 2008b). Significant topographic and sedimentary differences also occur on more localized scales within each of these regions, which makes stratigraphic correlation between the three Deep Sea Drilling Project (DSDP), Ocean Drilling Program (ODP), and Integrated Ocean Drilling Program (IODP) drilling transects challenging (Underwood & Pickering, 2018). To improve the documentation of sediment and basement heterogeneities, we conducted a more detailed margin-wide study by integrating the ground truth provided by drilling results with both legacy and recent seismic-reflection surveys.

This study is relevant to other subduction zones because it addresses the fundamental role of subducting sediment and basement topography in producing variations in deformation processes and discontinuities within the accretionary prism. High-resolution seismic reflection profiles display amplitudes and dimensions that are indicative of previously unrecognized contourite drifts. We show how basement topography influenced the distribution of facies types, including potentially pore-fluid-rich contourite deposits. We measure the frontal wedge taper to infer the plate boundary frictional conditions and compare the results with known variations in seismic behavior. Our results can be applied to test generic hypotheses that link topographic and sediment characteristics with megathrust behavior (e.g., Bilek & Lay, 2018; Hüpers et al., 2018; Saffer & Tobin, 2011; Sun, Ellis, & Saffer, 2020; Sun, Saffer, & Ellis, 2020; Wang & Hu, 2006).

2. Geological Setting

The Nankai Trough is host to a variety of fault-slip behaviors. The margin has a history of recurring and typically tsunamigenic great earthquakes, including the 1944 Tonankai (Mw = 8.1) and 1946 (Mw = 8.4) Nankaido earthquakes (Ando, 1982). SSEs, very-low-frequency earthquakes (VLFEs), and nonvolcanic tremors are also well-displayed in the Nankai region (e.g., Nakano et al., 2018; Obara et al., 2004). These events are spatially correlated with areas that have high pore fluid pressure (Kodaira et al., 2004; Nakajima & Uchida, 2018; Shelly et al., 2006), seamount subduction (Ide, 2010), and a low slip-deficit rate (Yokota et al., 2016). Additionally seamounts and local concentrations of high fluid pressures may act as barriers to large tsunamigenic slip (Kodaira, 2000; Tobin & Saffer, 2009). Accordingly, this particular margin is an ideal place to test ideas regarding lithologic controls on fault-slip phenomena.
The Shikoku Basin, a back-arc basin in the northeastern Philippine Sea plate, is currently being subducted beneath the Japanese Islands (Figure 1; Underwood & Pickering, 2018). The basin formed between 30 and 15 Ma during back-arc rifting and seafloor spreading behind the Izu-Bonin arc system (Kobayashi & Nakada, 1978; Okino et al., 1994; Sdrolias et al., 2004; Taylor & Fujioka, 1992; Watts & Weisell, 1975). The KSC (Figure 1) formed along the extinct spreading axis and continued erupting until 8–7 Ma (Ishii, 2000; Ishizuka et al., 2009; Sato et al., 2002). The seamount chain is still a bathymetric high that influences the locations of different types of sediment deposition (Ike et al., 2008b).

The Philippine Sea plate began subducting beneath the proto-Nankai Trough at ~15 Ma, before the Shikoku Basin spreading center became extinct (Hibbard & Karig, 1990; J.-I. Kimura et al., 2005). Subduction apparently ceased around ~12 Ma when relative motion along the proto-Nankai margin became transcurrent (Underwood & Pickering, 2018). Approximately trench-normal subduction resumed at an increased rate at ~6 Ma (G. Kimura et al., 2014, 2018; Wu et al., 2016). This plate reorganization led to the formation of the bathymetric trench that now defines the Nankai Trough and blocked many continuous sediment pathways from the landward edge into the center of the basin. Coarse terrigenous sediment however has been routed to the trench through several large submarine canyon systems (e.g., Suruga Trough, Tenryu Canyon), with

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*Figure 1.* Tectonic map of southwest Japan and the present day surface, intermediate, and deep water masses. The compilation of water masses is based on Kaneko et al. (2001), Kawabe and Fujio (2010), Talley (2011), Kubota et al. (2015), and Kender et al. (2018). The inset shows the location of the Nankai Trough in East Asia.

**2.1. Tectonic Setting**

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accelerated delivery since ∼2 Ma (Marsaglia et al., 1992; Underwood & Fergusson, 2005; Underwood & Pickering, 2018; Usman et al., 2014). Sediment delivery accelerated as a result of the Izu-Bonin Arc collision with Honshu, combined with the uplift of eastern Honshu as a byproduct of the convergence between southwestern and northeastern Japan (G. Kimura et al., 2018; Underwood & Guo, 2018).

2.2. Sedimentation History

Sedimentation within back-arc basins is usually influenced by proximity to volcanic eruptions, thermal decay and subsidence of the basement, prevailing winds and currents, and biogenic productivity (Karig & Moore, 1975; Macdonald, 1993; Marsaglia et al., 1995). Unlike most such basins, the Shikoku Basin is bounded on one side by a continental landmass, which provides a potentially large source of terrigenous sediment. The basin history involves volcanic and volcaniclastic activity on three sides, plate boundary reorganization, collision of two active arcs (Izu-Bonin and Honshu), collision of a remnant arc and an active arc (Kyushu-Palau Ridge and Kyushu), and uplift-denudation of the Japanese hinterland (Underwood & Fergusson, 2005). Consequently, the basin hosts sediments that include volcaniclastic, pyroclastic, and hemipelagic deposits, as well as siliciclastic sandstone bodies (Ike et al., 2008a, 2008b; Underwood & Pickering, 2018) (Figures 3 and 4). This diverse assemblage of lithologies has been modified by varying degrees of consolidation and diagenesis, resulting in substantial margin-wide heterogeneities in physical properties outboard of the subduction front (Hüpers et al., 2018). The sediments are thick enough to almost completely bury the basement topography (Figure 3). Many of the facies are localized, however, and the results of scientific ocean drilling show that most facies boundaries are diachronous, which has led to inconsistent facies designations and provisional correlations that proved to be incorrect when extended across the entire basin (G. F. Moore et al., 2001a; Taira et al., 1991; Underwood & Guo, 2018). In this study, we use the facies terminology defined by Underwood and Pickering (2018).

The basement characteristics of Shikoku Basin are largely controlled by the seafloor spreading associated with volcanic extrusions and extensional faults from the spreading axis (Ike et al., 2008b). The basement
Topography is roughest and youngest in the center of the Shikoku Basin (basement relief of 1–2 km between basement highs and lows), moderately rough in the east (relief predominantly <600 m, except for Kashi-nosaki Knoll which is ∼1,500 m) and relatively smooth in the west (relief <200–400 m) (Ike et al., 2008b).

Figure 3. (a) Uninterpreted and (b) interpreted seismic line 134 extending from the accretionary prism to the basin. Red dotted line indicates the seaward projection of the décollement. Sediment beneath this line is likely to be subducted in the future. Ocean Drilling Program drill Site 1173 is projected perpendicular to the local geology. The sedimentary section drilled at 1173 is different from the section seaward of the basement high.

Figure 4. Seismic stratigraphy of Ocean Drilling Program cores at Sites 1177 (western Nankai) and 1173 (central Nankai) and Integrated Ocean Drilling Program Site C0011 (eastern Nankai). The locations of the cores are shown in Figure 2.
The oldest sediment facies in the western (Ashizuri) and central (Muroto) portions of the margin is a volcaniclastic-rich facies which was deposited directly on top of the basaltic basement (Figure 4). Along the eastern (Kumano) transect, the basal facies is pelagic claystone, which is overlain by compositionally diverse sandstones that have a comparable age to the volcaniclastic-rich facies in the western Shikoku Basin (Site 1177). Overlying the basal volcaniclastic unit at Site 1177 is a package of siliciclastic turbidites with abundant organic matter (Moore et al., 2001c). These turbidites are largely confined to basement lows (Underwood et al., 2007), because the elevation of the spreading center’s axial ridge blocked most of these sediment flows from traveling to the eastern part of the basin. A superficially similar lithofacies was recovered at the eastern (Kumano) transect, the basal facies is pelagic claystone, which is overlain by compositionally diverse sandstones that have a comparable age to the volcaniclastic-rich facies in the western Shikoku Basin (Site 1177). Overlying the basal volcaniclastic unit at Site 1177 is a package of siliciclastic turbidites with abundant organic matter (Moore et al., 2001c). These turbidites are largely confined to basement lows (Underwood et al., 2007), because the elevation of the spreading center’s axial ridge blocked most of these sediment flows from traveling to the eastern part of the basin. A superficially similar lithofacies was recovered at the eastern transect. However, compositional correlations to the coeval deposits at Sites 297 and 1177 have never been established in detail. Thus, two separate fan systems may have existed on either side of the KSC.

A hemipelagic facies extends across the entire basin. In the central and western portions of the basin it transitions upwards into the youngest Shikoku Basin unit; a hemipelagic/pyroclastic unit. This unit and was de-
Along the northern edge of the subducting Shikoku Basin, a thick wedge of axial trench sediment has prograded over the top of the hemipelagic-pyroclastic facies (Mountnay & Westbrook, 1996). The Quaternary trench wedge thickens landward and results from a combination of margin-parallel transport of lithic sediment derived from the Izu-Honshu collision zone (e.g., Underwood & Fergusson, 2005; Underwood & Pickering, 2018) and transverse delivery through submarine canyons that are incised into the accretionary prism (e.g., Kawamura et al., 2009; Soh & Tokuyama, 2002). The seaward limit of the trench wedge is irregular in plan view, with extensive spill-over onto the northeast flank of Kashinosaki Knoll (Ike et al., 2008a).

3. Oceanographic Setting

Surface water masses in the North West Pacific are located between the sea surface and 500 m water depth. The regional circulation of these water masses is dominated by the vigorous Kuroshio Current flowing to the northeast, but is also affected by recirculation, large meanders, and the subtropical countercurrent (Figure 1; Talley, 2011). The North Pacific Intermediate Water (NPIW) circulates northeastward at water depths of 500–800 m in this area (Figure 1; Kaneko et al., 2001; Kawabe & Fujio, 2010). Beneath the NPIW, the Pacific Deep Water (PDW) circulates at >1,500 m water depth, due to the upwelling and recirculation in the Pacific of the modified Circumpolar Deep Water and Antarctic Bottom water (Kaneko et al., 2001; Kawabe & Fujio, 2010). In this study area, the PDW is circulating toward the northeast (Figure 1). The deepest water mass corresponds to the Lower Circumpolar Deep Water (LCDW), which is located at depths >3,500 m (Kawabe & Fujio, 2010). In this study area, the LCDW circulation is constrained by the seafloor irregularities, local highs and ridges, which results in a southwest flow along the Nankai Trough (Figure 1; Kaneko et al., 1998, 2001; Kender et al., 2018; Talley, 2011).

4. Methods

4.1. Seismic Data

We used 15 2D seismic reflection data sets, totaling >70 lines collected by the Japan Agency for Marine-Earth Science and Technology (Figure 2). These data were acquired over 20 years using a variety of source and receiver parameters (see Table S1 in Supporting Information S1). Standard data processing was done including noisy trace editing, common mid-point (CMP) sorting, bandpass filtering, deconvolution, velocity analysis, normal moveout (NMO correction), muting, CMP stacking, post-stack time migration, and depth conversion. Additionally, post-stack depth migration was done on the 1997–2002 and 2011 R/V Kairei lines. The velocity data used were acquired from Ocean Bottom Seismometer surveys (Nakanishi et al., 2018), constant velocity models (Ike et al., 2008b) and NMO correction and migration. We also used an 8 × 80 km 3D seismic volume collected by a US-Japan collaboration off Cape Muroto (Costa Pisani, 2005).

The interpretation was initially done in two-way travel time to ensure the correct ties of stratigraphic boundaries at trackline intersections, which may differ slightly in depth due to different velocity models applied. Most of the later interpretation was then done in depth, although the sediment distribution was mapped using the time-migrated data so that the older seismic lines with less accurate velocity data models could be included. We used average interval velocities based on pre-stack depth migration velocity analyses of the new (high resolution) lines to convert key values in the isochron maps from two-way travel time to depth (1.8 km/s for hemipelagic-hemipelagic/pyroclastic facies, 2.2 km/s for volcanioclastic turbidite facies, 2.6 km/s for siliciclastic and mixed turbidite facies, and 3.0 km/s for more deeply buried volcanioclastic-rich facies). These velocities were accurate when the depths were compared with drilling data (Figure 4).

4.2. Stratigraphic Analysis

A seismic stratigraphic analysis was executed according to the methods of Mitchum et al. (1977). Stratigraphic units were distinguished by seismic facies and boundaries (Table 1). We defined six seismic units (SUs) (1–6), including the acoustic basement (SU1). We correlated the SUs with the seven lithological units of Underwood and Pickering (2018). SU5 corresponds to a composite of two lithofacies units (hemipelagic and hemipelagic/pyroclastic) because it was not possible to clearly distinguish the boundary between the two lithofacies units in the seismic data unless they were separated by turbidites. Additionally, the oldest
SU that lies on top of the igneous oceanic crust (SU2), is lithologically complex, and encompasses at least two lithofacies units.

Depositional features associated with bottom currents were identified by their external shape and internal reflection geometry. Drape deposits were identified as parallel reflections with uniform deposition independent of bottom relief, whereas mounded drift deposits were primarily identified by their lenticular and upward convex geometry (Figure 5; Faugères et al., 1999; Rebesco et al., 2014).

4.3. Structural Analysis

Extensive networks of normal faults were observed throughout the basin. These faults were only measurable in the high resolution seismic lines that extend into the basin. These lines are 120, 122, 133, 134, and Sb01. We measured the length of the fault, the apparent dip magnitude, and the dip direction in each line. These properties were measured along the sections of the lines that were within 5 km of the intersection of the strike line and the dip lines to determine if there were any spatial variations. This corresponded to between ~50 and 160 faults along each line.

The dip of a fault measured in a seismic section is an apparent dip and depends on the angle between the seismic line and the trend (strike) of the faults. Where two seismic lines intersect and image the same fault,
the true dip can be calculated from the two apparent dips. Four such intersections of the high-resolution seismic lines allowed us to calculate the true dip of normal faults using basic trigonometry (see Supporting Information S1). We tested the calculation by measuring the apparent dips of a single fault in both the inline and crossline directions of the Muroto 3D volume and compared the calculated dip with the true dip measured in 3D. The measurements were within 1° of each other, validating the trigonometric method used.

The Coulomb wedge taper is the sum of the surface slope and the décollement dip (D. Davis et al., 1983). We calculated the surface slope of the wedge taper using seafloor bathymetry transects perpendicular to the margin (see Supporting Information S1). The transects intersect with seismic lines that have good resolution and depth conversion. The wedge taper varies greatly with distance landward, partially because it was influenced by previously subducted sediment; however, because we are interested in relating the wedge taper to the subducting sediment we only measured the taper of the frontal ~5 km landward of the frontal thrust. The décollement dips were measured in the seismic data between the frontal thrust and 5 km landward of the frontal thrust. Uncertainties are introduced when measuring both the décollement dip and the wedge taper angle; additionally the toe of the wedge may be affected by a collapse at the toe. Therefore, where there are few thrusts, or complex topography in the frontal 5 km we extrapolate our measurement landward and use the topography further landward as a guide (see Supporting Information S1). There are slight (<1°) differences between the wedge taper values in this study and those recorded along some of the same lines by Tilley et al. (2021) due to differences in the distance over which the taper was measured, and the angle of obliquity to the margin that was measured. The previous study (Tilley et al., 2021) measured the taper solely in seismic data, which are at varying degrees of obliquity to the margin, whereas we used both bathymetry and seismic data to measure the taper angle perpendicular to the accretionary prism. We calculated the Spearman correlation coefficient (\( \rho \)) and \( p \)-value to test for statistically significant relationships between the sediment thicknesses and wedge taper angle. \( \rho \) can vary between −1 and +1, where +1 is a strong positive relationship. The lower the \( p \)-value is, the less likely it is that the relation is not correlated. We use a \( p \)-value ≤0.05 as statistically significant.

5. Results

We identified 6 stratigraphic units (SU1-6), including the oceanic basement (SU1) and correlated the SUs with coring results from the three ocean drilling transects to date the unit boundaries and assess along-strike lithological and chronological heterogeneities (Table 1, Figure 4).

5.1. Basement Topography (SU1) and Total Sediment Thickness

SU1 corresponds to the acoustic basement, which is the igneous oceanic crust. The upper boundary is a strong continuous to discontinuous reflection. The internal reflections are chaotic and the amplitude of the reflections decreases significantly with depth. The upper surface is irregular with localized relief of >2 km between basement lows and adjacent basement highs. The basement is relatively smooth in the western portion of the basin, with little topographic relief (<600 m; Figure 6). The largest basement relief is in the center of the basin, and has a large horst and graben east of the Muroto transect (<2,300 m relief). The basement topography in the east of the basin is relatively rough (<1,400 m relief at the eastern edge of the study site), and is dominated by the Kashinosaki Knoll basement high (~800 m relief). The oceanic crust is youngest (>15.2 Ma) along the Muroto (central) drill transect (Table 1, Figure 4), >20 Ma in the western (Ashizuri) transect, and >18.9 Ma in the eastern (Kumano) transect.

The map of geographic variations in the total sediment thickness has a similar pattern to the map of the basement topography (Figure 6). The sediment thickness is generally less, and more uniform along the western portion of the margin (Table 2). Both the minimum (~200 m) and maximum (~2,000 m) sediment thicknesses occur in the center of the margin, near the extinct spreading ridge, but the greatest variability on average occurs along the eastern portion of the margin. The area of maximum thickness extends from the frontal thrust to the basin and is directly adjacent to, and to the northeast of, the area of minimum thickness (Figure 6). This area of thin sediment is directly seaward of the indentation in the margin that may be related to subduction of a large basement high (Kodaira, 2000).
5.2. SU2

SU2 was deposited directly on top of the oceanic basement (Table 1). The upper boundary of the unit is mostly sub-parallel to the basement topography, forming a thin drape (Figures 4, 7 and 9). However, a few localized thick (up to ~825 m) mounded deposits occur where the boundary is not sub-parallel to the basement topography. The upper boundary of the unit is most distinct where it occurs beneath SU3 due to the vast contrast in their seismic characteristics; SU2 exhibits low amplitudes, with few coherent reflections, whereas SU3 consists of high amplitude, continuous reflections (Figure 7). Where SU2 occurs directly beneath SU5 the boundary is more subtle and defined by onlap of the younger reflections.

SU2 includes at least two different lithofacies (Table 1); a pelagic clay facies and a volcaniclastic-rich facies. The oldest sediment recovered is a pelagic clay facies, but this was only recovered at IODP Site C0012 (Figure 4). A volcaniclastic rich facies was recorded along all three drill transects, and likely makes up the largest portion of SU2. None of the large mounded deposits in SU2 (e.g., Figure 7) were drilled, so it is not possible to verify if they are volcaniclastic or pelagic clay facies.

Mounded deposits in SU2 occur in basement lows and on the east side of basement highs, resulting in an easterly dip of reflections (Figure 7b). The mounded deposits are thickest in the center (up to ~500 ms two way travel time [TWTT]~825 m) and thin steeply at the edges. These deposits occur within 75 km west of the extinct spreading center and are up to 25 km wide in the along-strike direction. These deposits may also occur east of the spreading center. However, because they are acoustically very similar to SU5, which
typically directly overlies SU2 in this area, and there are few diagnostic onlapping reflections, it is not possible to conclusively map these deposits east of the extinct spreading center.

### 5.3. SU3

SU3 fills basement lows and onlaps the younger deposits at basement highs. The upper boundary is sub-horizontal and approximately parallel to the seafloor and generally occurs as a continuous strong reflection.

SU3 is interpreted as the siliciclastic or mixed siliciclastic/volcaniclastic turbidite facies recovered at the Ashizuri and Kumano transects (Table 1). Although SU3 on the east (Kumano) side of the KSC appears to correlate in stratigraphic position with the lower portion SU3 to the west (Ashizuri) of the seamount chain.

| Line          | Taper (°) | Depth to décollement (km) | Vertical distance between décollement and oceanic crust (km) | Vertical distance between décollement and diagenetic front (m) | Turbidite thickness (km) | Total sediment thickness (km) |
|---------------|-----------|---------------------------|-------------------------------------------------------------|----------------------------------------------------------------|--------------------------|-------------------------------|
| kr9810_1      | 3         | 1.1                       | 0.3                                                         | 0                                                              | 0.2                      | 1.4                           |
| hdn107        | 6         | 1                         | 0.5                                                         | 90                                                             | 0.32                     | 1.5                           |
| hdn109        | 6         | 0.9                       | 0.5                                                         | 0                                                              | 0.12                     | 1.4                           |
| kr0114-5      | 6         | 1.1                       | 0.4                                                         | 40                                                             | 0.25                     | 1.5                           |
| hdn111        | 5         | 1.1                       | 0.4                                                         | 50                                                             | 0.36                     | 1.5                           |
| hdn113        | 6         | 0.8                       | 0.7                                                         | 0                                                              | 0.4                      | 1.5                           |
| hdn115        | 5         | 1                         | 0.4                                                         | 20                                                             | 1.4                      |                               |
| hdn117        | 9         | 1                         | 0.5                                                         | 50                                                             | 0.17                     | 1.5                           |
| hdn119        | 3         | 0.6                       | 0.3                                                         | 0                                                              | 0.9                      |                               |
| hdn120        | 4         | 0.5                       | 0.3                                                         | 0                                                              | 0.8                      |                               |
| hdn121        | 2         | 1                         | 0.3                                                         | 40                                                             | 0.24                     | 1.3                           |
| hdn122        | 2         | 1.1                       | 0.4                                                         | 90                                                             | 0.19                     | 1.5                           |
| hdn123        | 1         | 1.1                       | 0.3                                                         | 70                                                             | 0.18                     | 1.4                           |
| hdn125        | 1         | 1                         | 0.3                                                         | 60                                                             | 0.05                     | 1.3                           |
| KR9704-MS104  | 5         | 0.8                       | 0.1                                                         | 20                                                             | 0.9                      |                               |
| hdn127        | 4         | 0.8                       | 0.3                                                         | 80                                                             | 1.1                      |                               |
| hdn129        | 2         | 0.8                       | 0.3                                                         | 100                                                            | 1.1                      |                               |
| hdn131        | 3         | 0.7                       | 0.4                                                         | 30                                                             | 1.1                      |                               |
| hdn133        | 5         | 0.8                       | 0.4                                                         | 0                                                              | 1.2                      |                               |
| hdn134        | 4         | 0.8                       | 0.4                                                         | 90                                                             | 1.2                      |                               |
| KR9806_08     | 7         | 1.4                       | 0.7                                                         | 100                                                            | 0.6                      | 2.1                           |
| KRO114-4      | 3         | 1                         | 0.3                                                         | 90                                                             | 0.14                     | 1.3                           |
| KR9806_06     | 12        | 1                         | 0.5                                                         | 0                                                              | 1.5                      |                               |
| KRO114-2      | 5         | 0.9                       | 0.3                                                         | 30                                                             | 1.2                      |                               |
| KRO114-1      | 9         | 1.3                       | 0.3                                                         | 100                                                            | 1.6                      |                               |
| hdn222        | 7         | 1                         | 0.6                                                         | 50                                                             | 0.23                     | 1.6                           |
| KR9806_04     | 4         | 0.7                       | 0.3                                                         | 0                                                              | 1                        |                               |
| kr108-5       | 4         | 1.3                       | 0.4                                                         | 0                                                              | 1.7                      |                               |
| ODKM-b        | 2         | 1.3                       | 0.7                                                         | 0                                                              | 0.55                     | 2                             |
| KR9806_02     | 13        | 1.3                       | 0.6                                                         | 0                                                              | 1.9                      |                               |
| KR9806_01     | 11        | 1                         | 0.5                                                         | 0                                                              | 0.13                     | 1.5                           |
(Figure 7), the deposits have distinct differences on either side of the basin (Table 1). Both deposits consist of mudstones with sandy layers, but the deposits along the Kumano transect have a tuffaceous component, which is absent along the Ashizuri transect. The deposits at Ashizuri also span a greater time interval (6.9–15.6 Ma) than the deposits along the Kumano transect (12.3–13.9 Ma) (Figure 4). Therefore, we divide the unit into two sub-units; SU3a for the siliciclastic (Ashizuri) deposits and SU3b for the mixed (Kumano) deposits. SU3 was not recovered at the Muroto transect.

SU3 varies in thickness up to 350 ms (TWTT) (up to 455 m) on the east side of the KSC (SU3b), and up to ~700 ms (TWTT) (up to ~900 m) on the west of the KSC (SU3a) (Figure 6). In the isochron map of the
deposits on the east of the KSC (Figure 6), the axis of the maximum thickness has an approximately north-south trend, which is roughly trench-perpendicular. West of the spreading center, the axis of the maximum sediment fill has a slightly more westerly orientation.

5.4. SU4

SU4 is a lens-shaped unit within SU5, and has markedly different seismic characteristics than SU5 (Figure 7, Table 1). The upper boundary is a strongly reflective horizon, approximately parallel to the seafloor, whereas the lower boundary is more irregular, resulting in a unit with thickness variations independent of basement topography. The upper interval of SU4 has high-amplitude, continuous reflections, whereas the lower interval has more chaotic reflections, with a variable amplitude.

SU4 is interpreted as a volcanlastic turbidite facies (Table 1). It onlaps onto topographic highs within SU5 and is the thinnest mapped unit, generally <150 ms (<∼165 m) (Figure 6). SU4 deposits are approximately spatially co-located with SU3b deposits (Figure 6). The thickest deposit occurs along the Kumano transect, immediately seaward of the frontal thrust (Figure 6). This deposit has a margin-parallel isopach trend, which contrasts with all of the other turbidite units.

5.5. SU5

SU5 occurs above either SU3 or SU2 and commonly downlaps onto the other units. The upper boundary of SU5 is the seafloor, which is predominantly sub-horizontal. SU5 consists of low to high amplitude, chaotic to linear reflections, which generally decrease in amplitude and continuity with depth (Figure 4, Table 1).

SU5 includes two mud-dominant lithofacies: an older hemipelagic facies, and a younger hemipelagic/pyroclastic facies (Table 1). The boundary between the lithofacies is defined by the onset of ash beds, which is not associated with a distinct change in seismic characteristics (Figures 4 and 7). A strong band of reflections in SU5 was previously interpreted as the boundary between these two lithofacies along the Muroto transect (Ike et al., 2008b). However, these reflections are approximately horizontal and cut across stratigraphic horizons (Figure 5), most notably where there are mounds and dipping strata in SU5. Therefore, we interpret the feature as a diagenetic front rather than a stratigraphic surface. This is consistent with a diagenetic front associated with amorphous silica (Spinelli et al., 2007; White et al., 2011). SU5 also hosts an extensive population of normal faults as well as the décollement. The lithofacies boundary is dated at 3.3 and 3.9 Ma at Muroto and Ashizuri respectively, but much older at 7.6–7.8 Ma at Kumano, where it is marked by a distinctive facies change to SU4.

SU5 makes up the greatest proportion of the sediment in the Shikoku Basin. It is ubiquitous throughout the basin and varies between ∼400 and 1,100 m thick. The thickness of this unit increases to the northeast, which is largely due to the absence of thick SU3 deposits. The depth to the diagenetic front ranges from ∼200 to 600 mbsf. The diagenetic front is shallowest (relative to the seafloor) over basement highs, and near to the extinct spreading ridge (Figure 8), but the magnitude of variations in the depth of the horizon is much less than that of the basement (Figure 7). The diagenetic front is preserved beneath the frontal portion of the accretionary prism.

5.6. Depositional Features

The localized mounded deposits of SU2 (up to ∼800 m thick and ∼25 km wide) have few coherent internal characteristics (Figure 7). The upper boundary of the unit is wavy and is not sub-parallel to the basement or present seafloor. The waves are approximately symmetrical, with wavelengths of 2–4 km and heights of >100 m. The upper boundary has an apparent dip of 2° to the north east in trench parallel seismic lines, and <1° in both directions in the trench perpendicular lines.

Some of the SU3 deposits in the west of the study area, above relatively flat topography have approximately symmetric, wavy internal reflections (Figure 5e). The wave amplitude is generally <100 m and the wavelength <1,500 m. This wavy unit is ∼95 km wide (along-strike) and ∼200 m thick. This is in contrast to the predominantly sub-parallel, sub-horizontal reflections of this unit throughout most of the study area. Near
basement highs there are some asymmetric mounds, which downlap onto mounded SU2 deposits and the basement topography. Additionally, in some narrow sub-basins between basement highs the deposits have an upward concave shape and dip toward the center of the sub-basin. These wavy deposits predominantly occur near the extinct spreading ridge (near mounded deposits in SU2 and SU5) in units up to 25 km wide (along-strike) and ~900 m thick (Figure 5g).

SU4 has an irregular base that is characteristic of erosive depositional processes (Figure 7, Table 1). The lowermost horizons are discontinuous and onlap onto SU5. The upper horizons are continuous and sub-parallel to the seafloor, with no significant wave features.

The oldest sediment in SU5 has extensive sediment waves, mounds and onlap boundaries (Figure 5). Converging normal faults are common at the top of these mounds. The younger sediment consists of approximately parallel linear reflections, which fill in the depressions between the mounds, and onlaps the older portion of the unit (Figure 7). However, the onlap boundaries are only visible in the new high-resolution data and mostly landward of the trench wedge. Therefore, it is not possible to map this horizon throughout the entire basin, using the lower-resolution seismic data. The dominant sediment waves have a height of up to 500 m and wavelengths of 2.5–10 km. Above basement highs, these waves are asymmetric, but above smoother basement topography are more symmetric. Where these waves are asymmetric, the steeper side is on the north-east in trench-parallel lines, but there is no discernable asymmetry in the trench-perpendicular lines. Some basement highs are flanked by erosive channels parallel to the relief (moats), where the oldest sediment of SU5 is absent. These moats are filled in with younger sediment, which have sub-horizontal reflections which onlap the basement high and a large sediment mound on either side.

5.7. Structural Features

5.7.1. Décollement

The décollement is recognized in the new high resolution data as narrow band of discontinuous reflections that has a higher amplitude than the surrounding strata (e.g., Figure 3), although the amplitude is generally lower than the deformation front reflections. The décollement can only be traced <10 km landward of the frontal thrust, because of significant attenuation beneath the accretionary prism.

Along almost all of the margin the décollement propagates into SU5. The depth to the décollement varies from 0.5 to 1.4 km beneath the seafloor (Table 2). The décollement is generally shallower relative to the seafloor where there is rough topography and no subducting SU3. The thickness of the sediment subducted...
beneath the décollement ranges between 0.1 and 0.7 km. The lines with the thickest subducting sediment are the lines that have the thickest SU3 deposits. Along the entire margin the décollement is within 100 m of the diagenetic front. Significantly, the décollement is beneath the diagenetic front on all the lines.

5.7.2. Normal Faults

The normal faults are similar in characteristics to the polygonal faults previously identified in this area (Heffernan et al., 2004), but are more spatially extensive and predominantly confined to SU5 (mud-dominant facies) (Figure 9). Therefore they introduce significant heterogeneities into an otherwise relatively homogenous unit. There appear to be two subsets of faults: short faults, which exist between the diagenetic front and the underlying sediment, and taller faults which can extend from the underlying sediment to the seafloor. The taller faults commonly converge at the same depth as the diagenetic front, and at the top of mounded deposits (Figure 5). The vertical offset of the taller faults increases down dip, and is predominantly between 5 and 20 m. The offset of the shorter faults is mostly <10 m. There is no difference in mean dips between these two subsets of faults. There is also no significant spatial variation in the fault dips. The apparent dips are 68–70° in the strike lines, and 66–72° in the dip lines (Table 3). The true dips of the faults range from 74 to 87° (see Supporting Information S1 for method). Three of the intersections between the dip and strike lines (120, 122, and 133) have a similar dip (72–78°) and fault bearing (10–25°). The normal

Figure 9. Uninterpreted (a) and interpreted (b) section of seismic line 120 highlighting the normal faults (solid black lines) both above and below the diagenetic front (black dashed line). The green circles highlight the location of fault clusters converging at the diagenetic front. The blue box is the location of the zoomed-in sections in (c–e). Thicker black lines in (d) correspond to normal faults with measurable offset (above the seismic resolution) across arbitrary horizons (blue lines). The offset of each horizon (where it is above the seismic resolution) is shown in (e).
faults at the intersection with line 134 have a bearing of 65° and dip of 87°. In contrast to the other three intersections, this intersection is located in a basement low.

The average fault density over 10 km segments near the intersections of the dip and strike lines is greater in the strike line (15–16 faults/km) than in the dip lines (7–13 faults/km). However, the individual 1 km segments vary from 5 to 25 faults/km, implying that variations are on a sub-10 km scale. These variations are largely due to localized clusters of converging normal faults (Figure 9).

### 5.7.3. Wedge Taper

We measure the wedge taper to assess how the subducting sediment influences the pore fluid pressure along the décollement. We compare the taper values with variations in margin and sediment characteristics using $\rho$ (Spearman's correlation coefficient) and $p$-value (how likely the data would have occurred by random chance) to see which characteristics correlate with the wedge taper. The frontal wedge taper varies between 1° and 13° and has a mean value of 5° (Table 2, Figure 10). The wedge taper does not have a straightforward along-strike increase or decrease in taper. Rather, the taper varies locally on a sub-10-km scale. Large changes in taper (4°) occur between lines only 10 km apart. These large localized changes are more common in the central and eastern lines (Figure 10). The highest taper is at line KR9806_02, which is near the Kumano 3D survey. Along this line the décollement propagates in a different sediment package than the rest of the margin, resulting in well-drained sand beneath the frontal thrust (Screaton et al., 2009). The lowest taper in this study is at lines hdnt123 and hdnt125. These lines are southwest of Muroto and have thick underthrust sediment and a very wide trench wedge and protothrust zone (Tilley et al., 2021).

The characteristics that are most strongly correlated with the wedge taper are mud-dominant sediment thickness, and the total sediment thickness (Figure 10). However, there are a few lines that reveal a more complex relationship. These include lines 123 and 125, where there is an approaching basement high and ODKM-b, near the Kumano 3D transect and recently subducted basement high (G. F. Moore et al., 2009). Within the mud-dominant unit, there is a stronger correlation between the wedge taper and the sediment thickness below the décollement, than above the décollement ($\rho = 0.42$, $p$-value = 0.017 and $\rho = 0.11$, $p$-value = 0.56 respectively). There is also a positive correlation between the distance along strike and the wedge taper angle, although this is a weaker relationship than sediment thickness and mud-dominant sediment thickness, and may reflect the sedimentation patterns along the margin. There is a relatively strong correlation between the distance along-strike and the mud-dominant sediment thickness, but no significant correlation between the distance along strike and total sediment thickness, or turbidite thickness. This highlights that the mud-dominant sediment thickness broadly increases to the northeast. There is no statistically significant correlation between turbidite thickness and wedge taper, between the accreting trench wedge turbidites and the wedge taper angle, or between the depth to the diagenetic front and the wedge taper angle.

| Fault density (fault/km) | Mean apparent dip (°) | Proportion of NW/SW dipping faults (%) | Proportion of NE/SE dipping faults (%) | True fault strike (°) | True fault bearing (°) | True dip (°) |
|-------------------------|----------------------|--------------------------------------|--------------------------------------|----------------------|-----------------------|----------------|
| Sb01 × 120              | 16                   | 70                                   | 57                                   | 43                   | 107                   | 17             | 74             |
| Sb01 × 122              | 15                   | 68                                   | 37                                   | 63                   | 100                   | 10             | 78             |
| Sb01 × 133              | 16                   | 69                                   | 35                                   | 65                   | 115                   | 25             | 72             |
| Sb01 × 134              | 15                   | 68                                   | 34                                   | 66                   | 155                   | 65             | 87             |
| 120 × Sb01              | 6                    | 70                                   | 64                                   | 36                   |                       |                |                |
| 122 × Sb01              | 11                   | 66                                   | 47                                   | 53                   |                       |                |                |
| 133 × Sb01              | 11                   | 70                                   | 25                                   | 75                   |                       |                |                |
| 134 × Sb01              | 13                   | 72                                   | 34                                   | 66                   |                       |                |                |

Note. The method used for calculating the true fault dip and strike from the apparent dip and orientation of the lines is found in Supporting Information S1.
6. Discussion

The key results presented in this study relate to sediment deposition, post-depositional sediment modification, and subducting sediment pore pressure. The porosity of sediment beneath and surrounding the décollement is important because it can influence the pore fluid pressure along the plate boundary and thus the frictional and seismic characteristics of the margin. The primary factors controlling the pore pressure in accretionary complexes are sediment permeability, sediment thickness, and plate convergence rate (Saffer & Bekins, 2006; Sun, Ellis, & Saffer, 2020; Sun, Saffer, & Ellis, 2020). The variations in plate convergence rate along the strike length of the Nankai Trough are negligible; therefore we focus on the competing influences of sediment thickness and permeability.

6.1. The Evolution of Sediment Deposition in the Nankai Trough

To determine the origin of sediment heterogeneities, it is important to establish a basin-wide understanding of the evolution of sedimentation at the Nankai Trough. As part of this, we must consider the interactions between paleoceanographic currents and basement topography, which influences the distribution and
The oldest sedimentary unit (SU2) includes at least two lithofacies (Table 1): a pelagic clay facies and a volcaniclastic-rich facies. The oldest drilled lithofacies is the pelagic clay facies, but this was only recovered at IODP Site C0012, whereas the volcaniclastic-rich facies was recovered along all three drill-transects. At ODP Site 808 SU2 consists of thick rhyolitic tuffs (Mikada et al., 2002), which differ compositionally from the claystone with ash-rich claystone/mudstone deposits at Site 1177 (Moore et al., 2001c). The compositional differences along the margin point to different sources and dispersal routes for this unit. The mound-ed drift deposits in SU2 are concentrated near the extinct spreading center, which may be due to a localized source, or related to the interaction of the bottom currents with the KSC. The oldest turbidite deposits (SU3), referred to as Kyushu Fan turbidites by Underwood and Pickering (2018), were interpreted by Clift et al. (2013) to be sourced from the Chinese mainland along a westerly supply route that traversed the East China Sea. Our study supports that interpretation for the deposits to the west of the axial basement high (SU3a) because the thickest deposits have north-west to south-east axial trend and are blocked by the axial basement high to the east (Figure 6). The seismically similar mixed turbidite deposits east of the basement high (SU3b) have different axes of thickest deposits, with a more north-south axial orientation, which suggests that they traveled into the basin from different directions. It is possible that during the initial stages of Kyushu Fan deposition the geometry of the basement high was such that thicker or higher-velocity turbidity currents were able to travel around the basement high and transport sediment into the eastern portion of the basin from a more northerly direction (Figure 11). It is equally plausible, however, that the eastern “Kyushu Fan” turbidites were derived from sources on the ancestral Japanese Islands instead of continental China. That alternative source is more in line with the quartz provenance indicators of Jaeger et al. (2019). Some of the Kyushu Fan turbidites have symmetric to assymetric wavy internal reflections (Figure 5), which are interpreted to be (sheeted) drift deposits formed by the bottom currents interacting with the turbidity currents.

The axial orientation of the thickest volcaniclastic turbidite deposits (SU4) is parallel to the margin, with an approximately north-east to south-west trend. This suggests an entirely different source location than the older turbidites and is consistent with the proposed east or north-east provenance (Pickering et al., 2013), which was associated by geochemical fingerprinting with the collision zone between Honshu and Izu-Bonin arcs (Kutterolf et al., 2014). The volcaniclastic turbidity currents were unable to travel over or around the KSC basement high, so the associated turbidites are confined to the eastern side of the Shikoku Basin.

The most prevalent mound ed drift features are in the lower portion of the mud-dominant (SU5) facies, and to the west of the spreading ridge. These are interpreted to be mound contourite drifts (e.g., Faugères et al., 1999). East of the spreading ridge there is evidence of bottom-current interaction with the sediment, but to a much smaller degree than to the west. Distinct moats around seamounts in the western portion of the basin (Figures 5 and 7) are evidence of a strong, erosive bottom current. The external geometry of contourite drifts can reveal the paleoceanographic conditions at the time of sediment remobilization and deposition (e.g., Rebesco et al., 2014). The paleoflow direction is difficult to discern in our study because we can only recognize the contourite-like drifts in the 18 new 2D high resolution seismic lines. Additionally, only a few of those lines extend far enough into the basin to observe the contourites prior to being compacted and distorted by the trench wedge. The contourite drifts in crosslines are approximately symmetric, making it difficult to determine the paleoflow. However, we can use the asymmetric drifts in the strike line to gain some idea of the past bottom current direction (Figure 11). Local erosion can occur upstream on both sides of the obstacle, and contourite accumulation can occur downstream (Chen et al., 2014). Additionally, in the northern hemisphere, where a bottom current encounters an obstacle its flow is accelerated to the left and decelerated to the right (Hernández-Molina et al., 2006), resulting in greater erosion on the left of an obstacle. In the crossline, the greatest erosive features are on the southwest of the large basement high (Figures 5g and 7). This leads us to interpret a bottom current that was moving from the south or east, which is approximatley parallel to the spreading ridge (Figure 11). This interpretation is consistent with the modern day bottom current direction (Figure 1; Kaneko et al., 1998, 2001; Kawabe & Fujio, 2010; Kender et al., 2018; Lee & Ogawa, 1998). On-land contourite analogs in Japan were originally deposited on the Pacific-facing slope of the Izu-Bonin Arc (Hüneke & Stow, 2008). The paleocurrent directions interpreted from these
deposits also correlate with the present-day bottom current direction, and are related to the North Pacific Deepwater current and the Antarctic Bottom Water (Lee & Ogawa, 1998).

Mounded contourite drift deposits are not hosted in the youngest (∼4 Ma) deposits in Shikoku Basin, which points to a change to weak deep-water circulation of the LCDW in the area. This weakening was probably caused by adjacent basement highs isolating the basin from the rest of the Pacific, resulting in recirculation within the basin. This change in the paleocirculation coincides with an increase in sedimentation rate at Sites 1173 and 1177 and an increase in the proportion of ash in the sediment (Underwood &

Figure 11. Simplified reconstruction of the northern Shikoku Basin and surrounding regions during the deposition of different sediment packages: (a–c) Sediment pathways and tectonic reconstructions (modified from Underwood & Pickering, 2018). Areas shown in (d–f) are highlighted with black boxes. (d–f) Schematic diagram of the interaction of the sediment with hypothesized seamounts (numbered ovals: 1Kodaira, 2000; 2Bangs et al., 2006; 3Park et al., 1999). The largest seamount is estimated to have begun subducting ∼3 Ma, given an approximate convergence rate of 4 cm/yr. The cross section shown in (g–i) is highlighted by a red line. (g–i) Approximate sequence of sediment deposition between the largest seamount (1) and a graben associated with seafloor spreading. The seamount geometry is estimated from seamounts in the Shikoku Basin (Figure 7), and the graben geometry is estimated from images of the graben seaward of the deformation front (G. Kimura et al., 2021). The black line in g indicates the mounded drift in the lower portion of SU5.
Pickering, 2018). It is possible that a reduction in bottom current velocity contributed to less winnowing and resuspension of fine particles, with enhanced vertical settling.

6.2. The Relationship Between Basement Topography and Subducting Sediment

The irregular basement topography of the Shikoku Basin is important because it plays a key role in controlling the sediment thickness and distribution throughout the basin. This sediment is eventually subducted and therefore influences the plate boundary conditions and subduction dynamics. Despite the basement irregularities having amplitudes of up to 2.3 km, the basement topography is largely buried by thick sediments, so the seafloor topography is minimal. This points to vast variations in sediment thickness between sites that are above versus between basement highs.

The axial basement high, which is related to the extinct spreading ridge, also restricted the sediment transport across the basin, resulting in considerable differences between sediment facies on either side of the extinct spreading ridge (Figure 11). The ridge separates the deposits of SU3a (siliciclastic turbidites) in the west from SU3b (mixed turbidites) and SU4 (volcaniclastic-rich turbidites) in the east. SU3a and SU3b were both deposited on top of SU2 within basement lows. However, SU3a is much thicker and has spans a greater age range than SU3b (Table 1). SU4 are lens-shaped turbidites hosted within SU5 and are only observed east of the basement high, commonly directly above SU3b deposits.

The perturbation of ocean bottom currents around basement highs during sediment deposition has also influenced the sediment characteristics. There are depositional features in the SU2, SU3, and the lower portion of SU5 that are consistent with reworking of sediment by bottom paleocurrents (Figure 5). Thick contourite mounded drifts in SU2 are hosted close (within 75 km) to the extinct spreading ridge (e.g., Figure 7), and localized sheeted drifts are observed in SU3 in the western portion of the basin and near basement highs (Figures 5e and 6a). Mounded drifts and basement controlled drifts are both hosted in SU5 and are typically related to basement topography (Figures 5f, 5g and 7). These drifts have a more distinctive geometry than sheeted drifts and may contribute to significant local heterogeneities in the sediment properties near basement highs.

6.3. The Relationship Between Depositional and Diagenetic Processes and Subducting Sediment Properties

The modification of sediment in the Shikoku Basin by depositional and diagenetic processes changes the porosity and permeability of the sediment. These characteristics determine how well sediment can drain upon subduction, which influences the plate boundary conditions.

Bottom-current processes can modify the grain size distribution of the sediment (Brackenridge et al., 2018; de Castro, Hernández-Molina, de Weger, et al., 2020; de Castro, Hernández-Molina, Rodríguez-Tovar, et al., 2020; Hüneke et al., 2021; Rebesco et al., 2014), and there is a strong link between grain size and porosity-permeability characteristics (Gluyas & Swarbrick, 2021; Selley, 2000). Contourite deposits are likely to have different porosity and permeability characteristics than comparable gravity-flow or hemipelagic deposits because of the reworking and winnowing of finer particles by the bottom currents (de Castro, Hernández-Molina, de Weger, et al., 2020; de Castro, Hernández-Molina, Rodríguez-Tovar, et al., 2020; Shanmugam, 2006, 2018, 2020; Shanmugam et al., 1993; Viana, 2008). Sandy contourites could have good sorting by bottom-currents and which would result in a higher porosity and permeability than turbidites with similar textures (e.g., Sanders & Friedman, 1997; Shanmugam, 2012; Viana et al., 2007). We identified prominent sediment waves in SU3 in the west of the study area (~95 km wide and 200 m thick), and near the extinct spreading ridge (~25 km wide and 900 m thick), which are indicative of the turbidite-dominant facies being reworked by bottom currents (Figure 5). Sandy turbidites, such as those in SU4 and SU3, are typically poorly sorted due to sudden freezing of gravity flows (Shanmugam, 2012). Contourites and hybrid deposits within turbidite-dominant units may result in localized sites of higher porosity and permeability. We also observed large mound contourite drifts in SU2 (up to ~25 km wide and ~800 m thick) and SU5 (up to ~10 km wide and 500 m thick), which largely consist of fine-grained silty mudstone and claystone. Fine-grained (muddy and silty) contourites are often relatively poorly sorted because of the bioturbation (Stow & Faugères, 2008) and with values of permeability lower than turbidites of comparable porosity (e.g.,
Moraes et al., 2007) but higher than normal hemipelagic deposits. IODP drill sites 1173 and 1177 both intersect mounded contourite drifts in SU5 (Figure 12). Contourites were not previously recognized in the cores from these drill sites; we believe this is because the features of fine-grained contourites are subtle, and there are not unambiguous diagnostic criteria for contourites at a core scale (de Castro, Hernández-Molina, de Weger, et al., 2020; Rebesco et al., 2014). However, both drill sites have surprisingly high porosities at for silty clays (G. F. Moore et al., 2001a). The diagenesis of ash deposits is thought to contribute to this
anomalous porosity (Hüpers et al., 2015), but we suggest that reworking of the muddy sediment by bottom currents may also have been a contributing factor.

All of the Shikoku Basin deposits undergo compaction beneath trench wedge turbidites prior to subduction. However, although some fluid is expelled during rapid loading by trench sedimentation, the permeability may be insufficient to keep pace (Screaton et al., 2002). The drainage of fluid from compacting and subducting sediment also depends of the length of the fluid pathway and the sealing of high permeability units by juxtaposed low permeability deposits (Saffer & Bekins, 2006). For example, although turbidite packages may be efficient drainage pathways where they are laterally continuous (Hüpers et al., 2018), many turbidite packages in our study are bound laterally by basement highs and sealed above by hemipelagic sediment (Figure 7). We identified confined turbidites up to ~25 km wide and ~900 m thick. In these cases, large volumes of pore fluid may be trapped and subducted, resulting in localized compartments of high pore fluid pressure. Equally, localized compartments of higher pore pressure might be expected where mound drifts pinch out against basement highs, or are flanked by erosive moats which were subsequently filled with thick, less-permeable hemipelagic sediment (e.g., Figure 5g). Given that the décollement localizes within ~100 m beneath the diagenetic front, we can assume that the proto-décollement will be at or beneath the diagenetic front. Therefore, whilst the mound drifts in SU2 and SU3 are subducted and may be confined (depending on their association with basement topography), the mound drifts in SU5 are likely to be intersected by the décollement, and therefore will be unconfined. However, this unit has a low permeability, so the degree of sealing is less significant.

A strong “diagenetic front” reflection, such as we observe, is a typical response an amorphous silica transformation boundary (Hein et al., 1978; Sarkar et al., 2019). The diagenetic front marks a change in porosity of the sediment. Above the diagenetic front in Shikoku Basin, although the sediment is weakly cemented, the amount of silica cement is enough to retain anomalously high and nearly constant porosity (Spinelli et al., 2007; White et al., 2011). Beneath this boundary, where the system shifts from net silica precipitation to dissolution, the porosity collapses abruptly (Spinelli et al., 2007). The diagenetic front in the Shikoku Basin cuts across stratigraphic horizons. Therefore, although it is restricted to the mud-dominant (SU5) lithofacies, its formation must be controlled by burial pressure and/or temperature as well as lithology. The depth to the diagenetic front varies from ~200 to 600 mbsf and it is shallowest near the spreading ridge and above basement highs (Figure 8). Globally, variations in heat flow are attributed to lithospheric cooling, with advective effects of hydrothermal circulation (e.g., E. Davis & Lister, 1974; Harris & Chapman, 2004; Parsons & Sclater, 1977; Stein & Stein, 1992, 1994). Additionally, at the Nankai Trough, the heat flow is anomalously high close to (<30 km seaward of) the trench (Harris et al., 2013, 2017; Spinelli & Wang, 2008). The highest heat flow values in our study area are concentrated along the Muroto transect, which is closest to the fossil spreading ridge (Harris et al., 2013). The younger, hotter basaltic crust facilitates shallower dissolution of silica (Hüpers et al., 2015) and a shallower deformation front in the central portion of the basin (Figure 8). However, the deformation front is shallower than would be expected in the Kumano area if the depth to the diagenetic front were solely controlled by the crustal age. This could be due to a greater initial ash content, which undergoes alteration to silica gel (Hüpers et al., 2015), or because the sediment is unusually thin in this area (Figure 6) as a result of the KSC basement high. Where there is vigorous fluid circulation in the basaltic basement aquifer homogenizing the temperature at the top of the basement, heat flow is higher where the sediment is thinner (Fisher & Harris, 2010). Therefore, the diagenetic front may be shallower where the sediment is thinner. Conversely, where there is thicker sediment, such as above a graben seaward of the the trench off Cape Muroto (Figure 6) the diagenetic front is depressed. The significant differences in sediment thickness and distribution documented herein help explain why heat flow values seaward of the deformation front are so variable (Harris et al., 2013), rather than changing exclusively by proximity to the extinct spreading ridge.

The diagenetic spatial patterns may also influence the location and properties of the décollement. The depth to the décollement's stratigraphic position varies from 0.5 to 1.4 km. Yet, the décollement is located within 100 m of the diagenetic front's depth on every line and is almost exclusively beneath the diagenetic front. From this, we infer that the location of the décollement is likely to be related to the diagenetic boundary and the associated change in bulk sediment properties. The décollement propagates seaward along a stratigraphic surface that marks a contrast in strength. In some subduction systems, that preferred interval
coincides with an obvious compositional change, such as an unusually high content of smectite (Deng & Underwood, 2001; Kameda et al., 2014). However, Underwood (2007) showed that the stratigraphic position of the décollement at the Nankai Trough does not coincide with unusually high concentrations of smectite and the frictional properties remain weak throughout the hosting lithologic unit. Therefore, it is widely believed that décollement localization may be related to excess pore pressure, which causes low shear strength along the plate boundary (e.g., Tobin & Saffer, 2009). The reduction in porosity and permeability beneath the diagenetic front results in a pressure cap, retarding drainage from below (G. F. Moore et al., 1991). This results in excess pore pressure forming beneath this pressure cap, resulting in low shear strength and the propagation of the décollement. When the décollement becomes established, it introduces a secondary porosity anomaly. This hypothesis is supported by the borehole data along the Muroto and Ashizuri transects (G. F. Moore et al., 2001a), as well as our observations of the close association of the deformation front at the décollement.

Normal faults in the mud-dominant unit also appear to be spatially related to the diagenetic front; tall faults converge at the diagenetic boundary and short faults are confined beneath the boundary. This is consistent with patterns seen at other basins (Sarkar et al., 2019). Heffernan et al. (2004) interpreted these faults as polygonal faults imaged in the Muroto 3D volume, and inferred that they formed due to differential compaction of sediments over an irregular basement combined with compactional dewatering at the toe of the accretionary prism. However, our data show that these faults form far seaward of the accretionary prism toe. A recent study noted that some of these normal faults penetrate the igneous oceanic crust and are related to the reactivated of old graben bounding faults (G. Kimura et al., 2021). However, this interpretation does not account for the majority of the normal faults, which we believe are more likely to be related to differential compaction. Praeger (2009) showed that the maximum throw on the polygonal faults occurs at the same depth as the diagenetic front. Our data support this observation, and we infer that polygonal fault formation is likely related to silica diagenesis. During the opal-A to opal-CT transformation the porosity and volume of the rock are reduced. This results in differential compaction and faulting, especially where it occurs above an irregular basement. Small microfaults were observed along the Muroto and Kumano drill transects (Expedition 322 Scientists 2010a, 2010b; Expedition 333 Scientists 2012a, 2012b; Moore et al., 2001b, 2001c). These high-angle normal faults are on a much smaller scale than we observe seismically, but their dips and stratigraphic concentrations are consistent with what we observe and are attributed to vertical compaction of the sediments during burial. Regardless of how they form, extensive normal faulting will either increase or decrease the bulk permeability of the sediment and will introduce a permeability anisotropy (Ferrill et al., 2000).

The influence of normal faults on the bulk permeability of the sediment is likely to be most significant where dense clusters of faults converge (Figure 9). The fault density measured within 1 km segments near the intersection of seismic line varies from 6-25 faults/km, highlighting localized variability (Figure 6). This is in agreement with the previous 3D study, which identified a greater fault density above the basement highs (Heffernan et al., 2004). The 3D study reported that the normal faults were confined to the upper and middle portion of the Shikoku Basin sediment section. However, we identify faults both above and below the diagenetic front and the décollement, which may be significant in providing hydraulic connectivity between the subducted sediments, the décollement, and the seafloor. The predominant calculated true strike direction is approximately perpendicular to the margin (Table 3). This is consistent with the values previously recorded for basement highs (Heffernan et al., 2004). However, four faults is a very small sample size, and if we consider the apparent dip direction of the faults, we can determine the quadrant that has the dominant fault strike. Except for line 120, the majority of faults in both the dip lines are SE-dipping and strike lines are NE-dipping. This corresponds to a dip direction that is between 90° and 180° and therefore a dominant strike direction that is between 0 and 90°, which is approximately parallel to the margin. The previous study also reported far more faults parallel to than perpendicular to the margin. If the normal faults are fluid conduits and are well connected they may allow considerable margin-parallel fluid flow, which can be driven by temperature (Spinelli & Saffer, 2007) or pore pressure gradients (Saffer & Tobin, 2011). However, because faults can be both barriers to and conduits for fluid flow the actual hydraulic connectivity of the faults remains open to further study.
6.4. Sediment Heterogeneities and Plate Boundary Conditions

The critical taper model of accretionary wedges and thrust belts states that basal and internal forces are balanced against gravitational forces, which results in a critical wedge taper angle (D. Davis et al., 1983). Elevated décollement pore pressures reduce the effective basal stress, resulting in a decreased shear strength along the décollement. These conditions allow the development of a mechanically stable accretionary prisms with a low wedge taper (D. Davis et al., 1983). Previous studies suggest that sediment thickness and permeability are the two key controls over pore pressure variability at the Nankai Trough (Saffer & Bekins, 2006; Sun, Ellis, & Saffer, 2020; Sun, Saffer, & Ellis, 2020). However, the two influences are often competing. Thick, homogeneously low-permeability sediment can result in excess pore pressures that are near lithostatic. However, where there is a thick permeable subducting sediment package, such as a turbidite unit, the décollement may be better drained (Hüpers et al., 2018). We compared the along-strike variations in the sediment thickness and lithology with wedge-taper values to determine how the variables are related.

The wedge taper of the frontal Nankai accretionary prism is more variable in the northeast part of the basin (Figure 10) than in the southwest, which is consistent with the variability trend of the basement topography and sediment thickness (Ike et al., 2008b). The highest wedge tapers are associated with thick, mud-dominant sediment; which is the unit that also hosts the décollement. This is in contrast with models that predicted that thick, mud-dominant sediment would sustain high pore pressures and low wedge tapers (Saffer, 2010; Saffer & Bekins, 2006). Additionally, in previous studies an overpressured décollement was inferred at Muroto, where there is a comparatively low wedge taper (4°) and no turbidite package in the subduction inputs (G. F. Moore et al., 2001a). Conversely, the steeper wedge taper at Ashizuri (9°) was inferred to be a result of the more permeable subducting turbidite package enabling pore fluid to drain efficiently from subducting sediment. Previous studies attributed the large wedge taper in the Kumano area to the presence of well-drained sand beneath the frontal thrust, which results in a high fault strength (Ikari et al., 2013; Screaton et al., 2009). Our more extensive data set highlights the limitations of using only three transects in previous studies. We found no clear relationship between the turbidite thickness and wedge taper angle (Figure 10). Turbidites that are unconfined and laterally continuous may be efficient drainage pathways, and they can be associated with large wedge tapers, such as along the Ashizuri transect. However, where the thick turbidite-dominant units are confined by basement highs and sealed by impermeable sediment (e.g., Figure 7, ∼25 km wide and ∼900 m thick), large volumes of pore fluid are probably trapped, resulting in localized compartments of high pore pressure, and smaller wedge tapers. Our analysis includes a broader spectrum of transects, with both confined and unconfined turbidites, which is why there is no clear relationship between subducting turbidite thickness and wedge taper angle.

On the other hand, there is a strong statistical relationship between mud-dominant sediment thickness and the wedge taper angle (Figure 10h). This is the unit that also hosts the décollement, and therefore only the lower part is subducted, whilst the upper part is accreted. The mud-dominant sediment thickness beneath the décollement is more closely correlated with the wedge taper than the sediment thickness above the décollement. This highlights the significance of subducting sediment in influencing the plate boundary conditions. The relationship between mud-dominant sediment thickness and wedge taper may be because (a) the mud-dominant unit hosts smaller volumes of pore fluid when it undergoes subduction, (b) thick mud-dominant sediments occur in the absence of thick trench turbidites, which can trap large volumes of fluid, or (c) the post-depositional modification of the mud-dominant sediment has increased the permeability of the unit significantly. We cannot rule out any of the possible explanations for the relationship between mud-dominant sediment thickness and wedge taper, and it is more than likely to be a combination of all three. Previous studies determined that this unit was likely to be underconsolidated, and unable to drain effectively when loaded by thick trench turbidites (Saffer & Bekins, 2006). However, that may not always be the case; the permeability of this unit may be more heterogeneous than previously assumed (e.g., Hüpers et al., 2018) due to the presence of thick contourite deposits (up to ∼10 km wide and ∼500 m thick) and extensive normal faulting.

6.5. Sediment Heterogeneities and Seismicity

Lithological heterogeneity along the plate boundary may be an important condition for stable sliding to SSEs in other subduction systems (Barnes et al., 2020; Saffer & Wallace, 2015). In contrast to some margins (e.g.,
Bradley et al., 2019), the Nankai décollement propagates within a comparatively homogeneous mud-dominant lithofacies. Therefore, it is more likely that pore pressure in the subducting sediment plays a more important role in controlling plate boundary variations. Abnormally high pore pressures reduce the fault strength of the décollement and splay faults by reducing the effective normal stress along the fault planes. The lower effective normal stress allows for tremor behavior such as VLFE with low stress drops (0.1%–1% of ordinary earthquakes; Ito & Obara, 2006). If the regions with a low wedge taper correspond with sections of the plate boundary that have higher-than-hydrostatic pore fluid pressure, then these sites are also more likely to be associated with transitional slip behavior such as SSEs, LFEs and nonvolcanic tremor.

Steep wedge tapers (11–13°) occur immediately adjacent to very low wedge tapers (2–3°) at both the Muroto and Kumano VLFE sites (Figures 10 and 13). This observation highlights how localized these variations in plate boundary conditions can be. Both sites are co-located with seamount subduction, therefore the wedge taper may reflect both the changes in pore pressure due to subducted sediment, as well as the geometric adjustments of the décollement as it steps up over the seamount (Bangs et al., 2006). The VLFE events primarily occur landward of where we mapped the wedge taper and seismic stratigraphy. However, we can apply our understanding of the sediment deposition in the Shikoku Basin to the inferred subducted basement topography. The basement topography influences the deposition of turbidites and contourite deposits, and we have shown that these deposits may contribute to localized variations in the plate boundary conditions.

Two sub-clusters of VLFEs off Cape Muroto were recorded surrounding a large subducted seamount (Kodaira, 2000) and two smaller seamounts (Bangs et al., 2006; Park et al., 1999; Figure 13). The two sub-clusters of VLFEs are separated by a gap in seismicity (around 135°E). The eastern sub-cluster appears to align with a narrow area of thick subducting sediment, which includes a thick siliciclastic turbidite-dominant unit (SU3) with internal waves, and mounded drifts in SU2 and SU5 (Figures 5g and 6). The localization of this deposit was controlled by large (km-scale) horst and graben structures, which extend landward of our study area (G. Kimura et al., 2021). The turbidite-dominant unit, and the volcaniclastic-dominant mounded drift are both confined laterally by basement topography. Upon subduction, these units may contribute to high pore pressures which are likely to be responsible for the VLFEs. The western sub-cluster of VLFEs is in the
same area as the three subducted seamounts (Figure 13). Based on the relationship that we have established between basement topography, sediment deposition and bottom current interaction in this area (e.g., Figure 11), we speculate that the higher pore pressures responsible for the localized VLFEs may be the result of contourite mounds and confined turbidite deposits that are likely to be associated with these seamounts.

The cluster of VLFEs off the Kii Peninsula is also located near an inferred subducted basement high (G. F. Moore et al., 2009). At this location, the basement high is thought to be responsible for the décollement stepping upsection and a thick section of fluid-rich sediment being underthrust (Bangs et al., 2009). Therefore, the mechanism producing the anomalous pore pressure necessary for VLFEs is likely to be underthrust, under-consolidated muds on the northwest side of basement highs (Shiraishi et al., 2020) rather than large contourite deposits. This site is a good example of localized heterogeneities because <1 km seaward of the VLFE site the wedge taper steepens significantly and plate boundary is inferred to be well-drained and frictionally strong (Ikari et al., 2013).

The spatial relationship between transitional seismic behavior and basement highs has been documented at other margins (e.g., Wang & Bilek, 2014 and references therein). At the Hikurangi margin, SSEs are co-located with high-amplitude reflectivity zones that flank a seamount (Barker et al., 2018). These zones are interpreted as fluid-rich subducting sediment (Bell et al., 2010). At that margin, lithological heterogeneity associated with rough basement topography is thought to contribute to the variable seismic behavior (Barnes et al., 2020). Seismic reflection data show that contourite deposits are also widespread along the Hikurangi margin (Bailey et al., 2020), and it is likely that subducting biocalcareous deposits on the Hikurangi Plateau have been molded by the Deep Western Boundary Current (Carter et al., 2004; Davy et al., 2008; Joseph et al., 2004). However, contourite deposits have not been identified at the shallow SSE study site. We suggest that contourite deposits are more common than previously assumed at subduction margins, but the resolution and spatial coverage of seismic data and cores have impeded their identification, and thus, these deposits are underestimated and not widely considered in basin analysis. We infer that bottom currents and basement topography contribute on a more widespread basis to the lithological heterogeneities that promote transitional seismic behavior at subduction margins.

7. Conclusions

Our study provides the first evidence of contourite deposits at the Nankai Trough subduction margin. Basement topography related to an extinct spreading ridge in the subducting Shikoku Basin played an important role in the localization of both thick turbidite and contourite deposits. The spreading ridge largely prevented sediment from being transported by gravity flows across the basin in the along-strike direction. Additionally, the focusing of bottom currents around the spreading ridge and at the foot of other basement topographic highs (e.g., seamounts) resulted in extensive and previously unrecognized contourite deposits. The occurrence of both contourites and turbidites in the Nankai subduction inputs is likely to result in highly localized (sub-10 km) heterogeneities in the porosity and permeability. We used the wedge taper angle as an indicator of décollement pore pressure and found that the total sediment thickness exerts the strongest influence over the wedge taper angle. Surprisingly, the thickness of the mud-dominant sediment is more important than the turbidite thickness. Turbidites can either contribute to efficient drainage of subducting sediments, or they can trap large volumes of pore fluid, depending on whether the sand bodies are laterally unconfined or confined in 3D by less-permeable strata. Additionally, the mud-dominant unit, which was previously regarded as relatively homogenous, hosts large contourite mounded drifts (sometimes associated with contourite moats), as well as pervasive normal faults; both are likely to introduce localized permeability heterogeneities. The interactions between basement highs and paleocurrent directions in the Shikoku Basin can be applied to mapped subducted seamounts, which allows us to predict the locations of subducted contourite mounds. Upon subduction, these patchy sediment deposits provide a complex stratigraphic template for localized compartments of excess pore pressure, which, in turn, may pre-condition the plate interface for slow slip behavior.

Data Availability Statement

The seismic data can be obtained at http://www.jamstec.go.jp/obsmcs_db/e/policy.html.
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