Deformation and Mechanical Strength of Sediments at the Nankai Subduction Zone: Implications for Prism Evolution and Décollement Initiation and Propagation.

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Abstract

As in many depositional settings, marine sediments at convergent margins undergo diagenetic changes before, during, and after mechanical consolidation and deformation. These changes influence mechanical behavior within and beneath the prism and along the décollement. To illustrate the interrelations between sediment diagenesis and deformation, we review physical properties and the types and distributions of deformation structures at several ODP and DSDP drill sites from the frontal regions of the Nankai accretionary prism. Both compactive and dilative deformation structures and fabrics are documented, denoting complicated stress paths during consolidation and tectonic deformation. Laboratory deformation experiments conducted on selected samples from these sites also demonstrate enhanced sediment strengths relative to their preconsolidation stress, both above and below the décollement horizon. This mechanical response indicates the presence of intergranular bonding or cementation that allows sediments to resist consolidation and deformation to relatively high stresses. Once their shear strengths are exceeded, however, cemented sediments can undergo rapid failure, leading to transient increases in pore pressures followed by consolidation. This deformation history may account for the localized compactive deformation bands within the prism. An analogous sequence may develop at depth within the underthrust sediments. Stress perturbations, e.g., near the up-dip limit of the seismogenic zone, may locally exceed the enhanced shear strengths of the underthrust sediments, leading to compactive failure and release of trapped pore fluids. Associated increases in pore fluid pressures may enable décollement downcutting and tectonic underplating. The resulting changes in structural and physical properties of the sediments may favor the onset of seismogenic slip along the décollement.

Keywords: Nankai Trough, accretionary prism, décollement, sediment mechanics, seismogenic zone.
1. Introduction

Active convergent margins are the locus of some of the largest earthquakes on Earth, a consequence of the combined effects of low-angle fault geometries, cold subducting lithosphere, and the compressional stress environment, that favor large seismic rupture areas of the fault. The extent of the seismically active portions depend on temperature and pressure distributions within the crust, and also, the compositional, elastic, and frictional properties of the fault and surrounding rocks [e.g., Scholz, 1990; Tichelaar and Ruff, 1993; Hyndman et al., 1995, 1997; Marone, 1998; Oleskevich et al., 1999]. Earthquake rupture, however, can extend up-dip along the fault into regions capable of supporting co-seismic slip. Co-seismic rupture may propagate onto splay faults, and finally to the seafloor, causing rapid ground motions capable of generating devastating tsunami [e.g., Ando, 1991; Park et al., 2000, 2002; Tanioka and Satake, 2001]. The frontal portions of the plate boundary fault may not sustain co-seismic sliding, but rather undergo aseismic slip as the overriding accretionary prism deforms during interseismic periods.

Significant changes must take place along the plate-boundary fault to account for the transition from seismic to aseismic slip. At the most frontal portions of accretionary prisms, porous sediments are partitioned by the décollement fault into accreted and underthrust packages that follow separate and distinct trajectories through the system [e.g., Karig and Morgan, 1994]. Such porous sediments cannot support high elastic shear strains and stress drops associated with earthquakes. At depth, progressive lithification, alteration, and probably pressurization, convert the marine sediments into rocks capable of unstable, seismogenic sliding. The detailed evolution in mechanical properties responsible for this change in behavior is not well understood; it is likely that many factors come into play, including evolving stress conditions, pore fluid pressures, physical properties, mineral fabric, and pressure- and temperature-induced diagenetic changes, and rheologies [e.g., Byerlee, 1990; Logan and Rauenszahn, 1987; Hyndman et al., 1995; Marone, 1998; Moore and Saffer, 2001].

The Nankai accretionary margin, located southeast of Japan (Figure 1), is a prime location to investigate transitions in physical and mechanical properties along the décollement and the surrounding wall-rocks, and their influence of sliding behavior. The margin has a long history of damaging earthquakes with a mean recurrence interval of ~180 yrs [Ando, 1975, 1991].
Ongoing microearthquakes along a well-instrumented transect offshore of Shikoku Island, cluster close to the up-dip limit of the 1946 Nankaido earthquake rupture [Obana, 2001], thereby constraining the position of the seismic-aseismic transition, or at least its lower bound [e.g., Norabuena et al., 2004]. A recent 3D seismic reflection survey along a transect parallel to Muroto Peninsula on Shikoku Island provides unprecedented imaging of the décollement in this area, as well as the frontal accretionary prism and incoming sedimentary section [Moore et al., 2001a, 2001b]. As discussed below, several key changes in deformation structure, seismic properties, and fluid pressures converge near the inferred up-dip limit of the seismogenic zone, which in combination, may enable the onset of unstable sliding along the fault. Additionally, seismogenic sliding may trigger rapid changes in physical properties and mechanical state that may account for the features described above.

To date, no samples have been acquired from the seismogenic zone along the Nankai margin, although this is the target of future deep drilling efforts [Kimura et al., 2003]. In the meantime, we must infer deep processes and properties based on materials and measurements collected from more accessible portions of the prism. Fortunately, the Nankai margin has been the site of many drilling ventures by the Ocean Drilling Program (ODP) and its predecessor, the Deep Sea Drilling Project (DSDP). Three ODP drill sites penetrate the frontal portion of the Nankai accretionary prism in line with the 3D seismic reflection survey, referred to here as the Muroto transect (Figure 1), and these sites can be compared with other sites that lie along the parallel Ashizuri transect, ~100 km to the southwest. In combination with the seismic data, the drill sites provide basic information about the physical properties, structure, and stratigraphy of the prism and the incoming sedimentary section [Shipboard Scientific Party (hereafter SSP), 1991, 2002a, 2001b]. These observations are augmented by results of laboratory experiments and analyses carried out to better constrain the physical and mechanical properties of the prism materials.

Here, we review the characteristics of the incoming sedimentary sections and synthesize observations and results relating to structural and mechanical evolution of sediments within and below the Nankai accretionary prism along the Muroto transect. We develop a conceptual understanding of how sediments evolve near the toe of the prism, and consider the consequences
of ongoing sediment diagenesis on deformation behavior within the prism and along the plate boundary décollement fault.

2. Large Scale Characteristics of the Nankai Margin

The Nankai accretionary prism lies at the boundary between the Eurasian and Philippine Sea plates, off the southwest coast of Japan. The inactive Shikoku back-arc basin and overlying sediments are presently being subducted northwestward at a relative convergence rate of ~2-4 cm/yr, normal to the margin [Seno et al., 1993]. A broad trench, the Nankai Trough, has formed above the downgoing plate, and gives way to a landward sequence of ridges and benches defining the imbricate stack of thrust sheets forming at the toe of the Nankai accretionary prism (Figure 1). The margin is the site of several Deep Sea Drilling Project (DSDP) and Ocean Drilling Program (ODP) surveys, including Legs 87, 131, 190, and 196 [e.g., Kagami et al., 1986; Hill et al., 1991; Moore et al., 2001b; Mikada et al., 2002]. A suite of six drill sites was cored during Legs 131 and 190 along a transect online with Muroto Point on Shikoku Island (Figure 1). We will focus on the drill sites penetrating the frontal portions of the prism, and the seaward Shikoku Basin: Site 1173 samples a reference site through the incoming Shikoku Basin sequence; Site 1174, penetrates the proto-thrust zone and an incipient thrust fault; and the older Site 808, which passes through the frontal thrust fault, active décollement zone, and the underthrust section below. Both Sites 1173 and 808 were revisited for logging and long-term fluid monitoring operations by ODP Leg 196 in 2001 [Mikada et al., 2002]. A comparative reference Site 1177 was drilled ~100 km to the southwest along the parallel Ashizuri transect (Figure 1), penetrating the deep incoming sedimentary section. Site 582 and 583, drilled during DSDP Leg 87, sampled shallower strata along this transect, within the Nankai Trough and the frontal accretionary prism, respectively.

The large-scale architecture of the Nankai accretionary margin along the Muroto Transect has been revealed by seismic reflection surveys [Moore et al., 2001a, 2001b; Gulick et al., 2004; Bangs et al., 2004]. The basal décollement lies within landward dipping Shikoku Basin strata at the deformation front, and is traced along a common acoustic reflection beneath the toe of the
prism. At ~30-40 km from the deformation front, the taper of the wedge increases, reflecting a large offset out-of-sequence thrust fault that rises from the décollement horizon; the décollement appears to step down to the top of the igneous crust in the landward direction; a large-offset out-of-sequence thrust fault splays from the décollement at this point (Figure 2). The décollement step-down approximately coincides with the up-dip limit of the seismogenic zone, based on the modeled rupture zone of the 1946 Nankaido earthquake [Ando, 1991; Tanioka and Satake, 2001], and locked region of the fault [e.g., Hyndman et al., 1995]. Increased microseismicity within 30-40 km from the deformation front falls close to the up-dip limit of seismogenic zone defined by thermal models [Obana et al., 2001]. The strong reflectivity of the décollement horizon beneath the frontal prism is attributed to a high impedance contrast, due to the downward increase in porosity, and possibly pore pressures [e.g., Moore et al., 1990, 1991; Moore and Shipley, 1993; SSP, 1991, 2001a, 2001b]. A decrease in reflection amplitude near the step-down has been interpreted to denote dewatering of the underthrust section [Bangs et al., 2004]. Hydrologic modeling of steady-state fluid flow indicates that the highest excess pore pressures should lie 20-30 km landward of the deformation front [Saffer and Bekins, 1998], which is close to the up-dip limit of the seismogenic zone [Moore and Saffer, 2001].

3. The Frontal Nankai Prism, Muroto Transect

The incoming sedimentary section on the Philippine Sea plate above the oceanic crust, is dominated by bioturbated hemipelagic sediments of the Shikoku Basin sequence, deposited between the early-mid Miocene until the Quaternary (i.e., ~20 – 0.26 ma) [SSP, 1991, 2001a,b; Moore et al., 2001b; Underwood, this volume]. The Upper Shikoku Basin (USB) facies is distinguished by interbedded ash layers, which are absent within the lower Shikoku Basin (LSB) facies. The broad Nankai Trough is filled by a northwestward thickening wedge of interbedded turbidites and hemipelagic sediments accumulating today (Figure 3) [Underwood et al., 1993; Underwood, this volume]. Onset of tectonic deformation is marked by a slight upward deflection of the flat-lying trench-fill strata, and stratal thickening and incipient thrust faulting within the proto-thrust zone [Moore et al., 1990, 1991]. An orderly stack of imbricate thrust
sheets [e.g., Gulick et al., 2004] marks the frontal portion of the prism. The basal décollement extends beneath the prism and proto-thrust zone, and lies entirely within the LSB facies; its seaward projection along the age-equivalent horizon is referred to as the proto-décollement. Three ODP drill sites intersect the trench and frontal portion of the prism (Figure 3). The full stratigraphic section was penetrated at each site.

There are four main structural domains at the toe of the prism, each exhibiting distinct characteristics indicative of their compactive, diagenetic, and tectonic histories. These domains are distinguished as: (a) reference section at Site 1173, presumed to have experienced no tectonic deformation, including units both above and below the proto-décollement horizon; (b) tectonically deformed strata found above the décollement zone within the proto-thrust zone and frontal thrust sheets; (c) the underthrust section found beneath the décollement, landward of the deformation front; and (d) the décollement zone itself, which separates the accreted and underthrust packages.

4. Physical Properties

Shipboard porosity measurements from the three sites [SSP, 1991, 2001a, 2001b; Screaton et al., 2002] provide a general overview of spatial changes in physical properties (Figure 4). The downward decreasing trend in porosity, $\eta$ (see Table 1 for all symbol definitions) at Site 1173 reverses at the top of the USB, at ~100 meters below sea floor (mbsf), and maintains a nearly constant porosity range of 65-68% to a depth of ~325 mbsf (Figure 4c). At this depth, porosities decrease to ~52 % over less than 100 m, and subsequently resume their steady downward decline. The sharp decrease in porosities coincides with the USB – LSB boundary, and accompanies the downward loss of interbedded ash layers [SSP, 2001a]. Acoustic compressional velocities, $V_p$, increase steadily across the USB and its lower boundary, rising to ~1800 m/s at the protodécollement horizon [SSP, 2001a], indicating steadily increasing particle coupling downhole. The porosity step in the absence of the velocity change at the USB-LSB boundary suggests some type of diagenetic transformation that enables renewed consolidation without loss of intergranular contact [e.g., SSP, 2001a]. Acoustic velocities step down to ~1750 m/s below
the protodécollement, while porosities show a very subtle step up, patterns again suggestive of an as yet undetermined diagenetic boundary.

Sites 1174 and 808 also show slight anomalies in porosities across the USB unit, but these are muted by comparison (Figure 4a and b). Slight downward increases in porosities occur across the frontal thrust fault at Site 808, ~410 mbsf, and across the protothrust fault at Site 1174, ~470 mbsf, probably reflecting incomplete consolidation of the footwalls following overthrusting [e.g., SSP, 1991]. More pronounced steps up in porosity occur across the basal décollement at both sites, ~964 mbsf at Site 808 and ~840 mbsf at Site 1174 (Figures 4a and b). Acoustic velocity profiles at these two sites are also complicated. Generally, velocities decrease in concert with porosity increases. Exceptions to this rule at Site 1174 include the lower trench wedge facies (i.e., Outer Trench Wedge) and the top of the USB facies, where velocities are anomalously high. The former anomaly terminates at the base of the trench wedge facies, where porewater profiles of both SiO₂ and Cl⁺ begin gradual downhole declines [SSP, 2001b]. Again, these distinctive patterns may reflect diagenesis within the sedimentary column.

Some of the physical properties distributions documented in the Shikoku Basin units and overlying strata along the Muroto transect are mirrored at comparable drill sites along the Ashizuri transect ~100 km to the southwest, although with slight variations. A downhole increase in porosity at the top of the USB facies was noted at DSDP Site 582 [Bray and Karig, 1986], which projects along strike between Site 1173 and the deformation front (Figure 3). Sampling at ODP Site 1177 began in the lower USB facies farther seaward of Site 582, but also documented anomalous acoustic velocities in that unit and a downhole decrease in porosity at the USB-LSB boundary [SSP, 2001c]. The high porosities observed in the USB unit at Site 582 were previously interpreted to result from overpressures [e.g., Bray and Karig, 1986, 1988], but reconsolidation experiments on sediment cores showed unusually high yield strengths, more consistent with cementation [e.g., Karig, 1993]. In sharp contrast to the monotonous hemipelagic sediments that make up the LSB facies along the Muroto transect, Site 1177 penetrated a ~330 m thick package of interbedded turbidites, which are characterized by porosities of 40% and higher [SSP, 2001c]. Although not the immediate focus of the present
study, the along strike variations in lithostratigraphy, and associated mineralogy and hydrogeology must play an important role in the regional evolution of the Nankai prism and underlying décollement zone [e.g., Moore et al., 1990; Underwood, this volume].

5. Deformation Structures and Fabrics

5.1. Ductile Deformation

Deformation within the Nankai prism sediments can take many forms, dependent on the stress history of the given domain [e.g., Karig and Morgan, 1994]. All units experience similar initial stages associated with deposition and subsequent burial at the seafloor. This normal compaction phase results in vertical shortening and consolidation, producing a downhole reduction in porosity, accompanied by the rotation of platy clay minerals into axisymmetric arrangements sub-parallel to bedding [e.g., Behrmann and Kopf, 1993; Morgan and Karig, 1993]. Both deformation modes are ductile, because they impart bulk changes in physical properties and microfabric. Tectonic loading can also cause ductile deformation of the sediment matrix, and potentially, further consolidation. This results from increased horizontal stresses that cause horizontal flattening of pore spaces and rotations of clay minerals away from the bedding planes. Using the method of X-ray pole figure goniometry, Morgan and Karig [1993, 1995b] quantified the clay mineral orientation distributions at Site 808, demonstrating axisymmetric arrangements of clay minerals parallel to bedding below the décollement consistent with uniaxial consolidation history (Figure 5). Up to 10% lateral anisotropy was evident above the décollement at Site 808, based on orthorhombic clay mineral distributions within the prism above the décollement (Figure 5b). If clay minerals are assumed to rotate passively within the sediment matrix [March, 1932], then the shape of the bulk strain ellipsoid can be calculated [Oertel, 1983, 1985], documenting the ductile components of horizontal and vertical shortening. All samples measured within the accreted section at Site 808 yielded lateral anisotropies ~10%, which equates to nominal ductile horizontal shortening of ~10% [Morgan and Karig, 1993].

More detailed information about sediment microfabrics has been extracted from SEM images [Ujiie et al., 2003; Sunderland and Morgan, 2004]. Above the proto-décollement zone at the
reference site, sediments exhibit a wide-range of grain sizes, and open, randomly oriented microstructures (Figure 6a). A fine-grained clay phase is dispersed throughout the samples, adhering to the surfaces of larger grains. The larger grains define sub-horizontal mineral preferred orientations, although local deviations up to ~30° were recognized [Sunderland and Morgan, 2004]. Below the proto-décollement horizon, clay minerals are typically larger than above, and more uniform in size and shape (Figure 6b). The distinctive fine-grained phase is absent. At these depths, clay mineral preferred orientations are dominantly sub-horizontal; distinctive large pores, some of which might derive from ash dissolution [Tribble and Wilkens, 1994], persist throughout the core, i.e., to depths of at least 700 mbsf. Ujiie et al. [2003] noted the presence of fine-grained clay aggregates above and within the décollement horizon at Site 1173, which they interpreted to represent a bonding phase; similarly, these aggregates are absent in the deeper sediments.

SEM images of sediment microfabrics at Site 1174 show evidence for primary vertical consolidation followed by a distinct tectonic overprint [Sunderland and Morgan, 2004]. Similar to Site 1173 above the proto-décollement, sediments in the shallowest units at Site 1174 exhibit open microstructures and the presence of the dispersed fine-grained clay phase (Figure 7a). This fine-clay phase, however, disappears about midway through the USB facies, by at least 690 mbsf (Figure 7b); this is more than 100 m above the décollement horizon, and therefore shallower than at the reference site. Below this depth, clay minerals exhibit more uniform shapes and sizes, reminiscent of the sub-décollement grains at the reference site (Figure 7c-d). Sediment microfabrics within the proto-thrust zone, particularly in the shallower reaches, display a subtle domainal structure, defined by packets of subparallel clay platelets separated by less well-oriented grains (Figure 7a). This fabric is most apparent in samples close to faults and other discrete deformation structures [Sunderland and Morgan, 2004]. Furthermore, the aligned grains are inclined up to 70 degrees to the horizontal plane, suggesting rotation in response to horizontal shortening (Figure 7a-b). These results are consistent with X-ray goniometry evidence for horizontal flattening, causing rotation of clay minerals away from the bedding plane (Figure 5); the SEM images demonstrate the heterogeneous, domainal nature of this grain rotation. Underthrust sediments at Sites 1174 and 808 (Figure 7d), lack the evidence for tectonic rotation.
observed above. Clay mineral preferred orientation is axisymmetric defining pre-tectonic bedding-parallel alignment. Both large and small pores exist throughout the Site 1174 cores, even below the décollement, despite the greater overburden.

5.2. Brittle Deformation

Evidence for brittle deformation is found throughout the prism toe [Taira et al., 1992; Moore et al., 2001b], although it is most intense at Site 808 [SSP, 1991], and quite minimal at the reference Site 1173 [SSP, 2001a]. The microfaults recognized at Site 1173, with few exceptions, are high angle normal faults with offsets of only a few centimeters [SSP, 2001a]. In one instance, the footwall of a small fault is also brecciated (Figure 8a), but all of the fractures are mineralized and healed, and clearly inactive. A thin zone of foliated breccia occurs below the proto-décollement horizon, but correlates with no other obvious features. Site 1173 also reveals rare dewatering features, one of which deflected a bedding horizon before apparently venting to the old seafloor (Figure 8b). Occasional high angle normal faults are also present within the Shikoku Basin units in the prism, at Sites 1173 and 808. In all cases, these structures reveal no preferred orientations, consistent with pre-tectonic extension of the sediments due to compaction or plate bending prior to subduction [SSP, 2001b].

Deformation bands occur at shallow depths in the drill holes at Sites 1174 and 808, preferentially in the trench wedge hemipelagites and the USB sediments [e.g., SSP, 1991, 2001b; Maltman et al., 1993; Ujiie et al., 2004]. These tabular structures are characterized by porosity reductions and anastamosing zones of rotated clay minerals (Figure 9a). The deformed zone is generally continuous with the adjacent wall rocks, such that the deformation bands are better classified as brittle-ductile, rather than purely brittle, structures [Karig and Lundberg, 1990]. Anastamosing shear planes and rotated clay minerals within the deformation bands most commonly reveal a reverse sense of slip, consistent with sub-horizontal maximum compressive tectonic stress. Deformation bands strike preferentially NE-SW, sub-parallel to the Nankai plate margin; they exhibit both northwest and southeast dips, and occasionally occur in conjugate pairs (Figure 9b) with a mean subvertical interband angle of ~62° [SSP, 1991, 2001b; Karig and Lundberg, 1990; Ujiie et al., 2004]. The high dips associated with the deformation bands differ
from those typically associated with conjugate brittle Coulomb shears, which form at \( \sim 30^\circ \) to the sub-horizontal maximum compressional stress field. To explain this angular discrepancy, it has been suggested that the deformation bands formed early during tectonic deformation, and were subsequently rotated to higher angles by bulk horizontal shortening [Karig and Lundberg, 1990; SSP, 2001b]. Alternatively, the deformation bands initiated as high angle reverse structures [e.g., Ujiie et al., 2004]. Karig and Lundberg [1990] described a temporal sequence of deformation preserved near the frontal thrust fault, where high angle deformation bands are overprinted by low-angle deformation bands, and finally, brittle microfaults that lie close to the 30° dips predicted for Coulomb shears.

Open fractures and microfaults, as observed in the recovered cores, become more common with depth at both Sites 1174 and 808; several show measurable offset. Most distinctive are zones of intersecting sets of sub-parallel inclined fractures (Figure 9c), interpreted to define conjugate shear fractures [SSP, 2001b]. Fracture orientations vary with position in the prism, but are generally consistent with sub-horizontal maximum compressive stress orientations. At Site 1174, several of the inclined fracture sets coincide with interpreted positions of seismic scale thrust faults, suggesting that the fractures represent deformation localized along or adjacent to discrete forethrusts or backthrusts within the proto-thrust zone [Moore et al., 2001a]. Commonly, the most fractured units are associated with foliated breccias and unusually poor core recovery, hinting at the loss of highly damaged fault rock during rotary drilling [Sunderland, 2003]. The most intense fracturing occurs just above the décollement zone at Site 1174 (Figure 10). This may reflect the presence of a deeper splay fault recognized on the seismic profiles through the drill site [Moore et al., 2001b]. As with the deformation bands, microfractures demonstrate a preferred strike subparallel to the plate margin, consistent with a tectonic origin for these features [SSP, 2001b].

Relatively few deformation structures were observed within the underthrust section. Those that do occur are almost exclusively healed, high-angle normal faults showing nominal offset [SSP, 1991, 2001b], all consistent with pre-tectonic origin. The difference in deformation
structures across the décollement shows that the fault is a zone of mechanical decoupling [e.g., Taira et al., 1992].

5.3. Décollement Zone

The décollement zone stands out in cores from both prism sites as a 20-30 m thick zone of intense fracturing and brecciation, defined by a web of anastamosing shear surfaces [Taira et al., 1992; Moore et al., 2001b]. The décollement shows the most intense shattering where penetrated at Site 1174 (Figure 10d-e), although the intensely deformed fault zone yielded nearly 100% core recovery [SSP, 2001b]. The intensity of deformation and fragmentation varies across the fault zone (Figure 10), but generally increases down-section, as measured by the spacing of fractures and the size of breccia fragments (Figure 10a). The base of the décollement at Site 1174 is very sharply defined; a thin layer of foliated breccia marks the base of the fault (Figure 10f). Immediately below, sediments are unfractured and porosities are noticeably higher, representative of the underthrust section (Figure 4a-b). The top of the décollement is less easily discerned: at Site 1174, closely spaced inclined fractures extend down to at least 824 mbsf (Figure 10c); this package was originally interpreted to lie within the upper décollement zone [SSP, 2001b], although the inclined fracture fabric suggests that, it may denote the base of an intersecting protothrust imaged on the seismic reflection profiles at this depth [Moore et al., 2001a].

Breccia fragments taken from within the décollement have porosities lower than sediments both above or below (Figure 4a-b), which could be interpreted as evidence for shear related dewatering [e.g., Moore et al., 1986]. X-ray goniometry measurements of clay mineral fabrics [Morgan and Karig, 1995b], however, show a gradation in lateral anisotropies across the décollement zone (Figure 5), indicating a downward decrease in horizontal tectonic strain with depth, independent of the low porosities [Morgan and Karig, 1995b]. SEM images corroborate this interpretation; the clay microstructure within the breccia fragments shows no evidence for enhanced preferred orientation, although the sediments are very dense, and lack the large pores that persist throughout the surrounding sediments [Ujiie et al., 2003; Sunderland and Morgan, 2004]. It appears that the breccia fragments within the décollement zone have experienced
enhanced consolidation under isotropic (hydrostatic) loading. Morgan and Karig [1995b] speculated that this consolidation was a result of the mechanical breakdown of intergranular cement within the décollement zone, which the surrounding wallrocks did not experience. Ujiie et al. [2003] recognized collapsed fine-grained clay-aggregates that they suggested may have served as the cementing agent. Despite the lack of internal shear fabrics within the décollement fragments, most of the fragments exhibit thin layers of highly oriented clay minerals at their surfaces (Figure 11a-b), indicative of sliding along shear planes within the fault. The thinness of these layers and lack of internal shear fabric argue for shear displacement occurring under relatively high pore fluid pressure conditions, reducing the effective normal stress and shear coupling across the slip surfaces, and preventing penetrative shear strain within the fragments [Morgan and Karig, 1995b; Ujiie et al., 2003].

6. Sediment Diagenesis

Drill core observations and physical properties measured at Site 1173, within the Shikoku Basin, are interpreted to define the reference state of the sediments that have entered the accretionary prism, at least at the very toe. Yet, measured physical properties at Site 1173 reveal distinct excursions from typical reference trends, even compared to the correlative units within the prism at Sites 1174 and 808 (Figure 4). These lateral variations in physical properties result in part from differing degrees of consolidation, and changing stress states across the prism [e.g., Bray and Karig, 1985; Morgan and Karig, 1995a], but also probably reflect diagenetic changes in sediment state. For example, the sharp downhole decrease in porosities at the USL-LSB boundary at Site 1173 coincides with a downhole decline in dissolved silica, suggesting is denotes a silica diagenetic boundary, e.g., opal A transforming to opal CT [e.g., SSP, 2001a]. At first appearance, the downhole densification and dissolved silica profile at Site 1173 resemble those correlated with progressive silica diagenesis, albeit in sediments initially much richer in biogenic silica [O’Brien et al., 1989; Nobes et al., 1992; Tribble et al., 1992]. For this reason, we have tentatively identify the USB – LSB boundary at Site 1173 as the opal-A to opal-CT
transition, a boundary that is likely to migrate with changing thermal state of the accreting sedimentary package.

Diagenetic changes in clay mineralogy have been better characterized along the Nankai margin. Semi-quantitative clay mineral analyses [Underwood et al., 1993; Underwood and Steurer, 2003; Steurer and Underwood, 2003] at all of the Nankai drill sites reveal downhole variations in smectite abundance, the combined result of variations in detrital contributions and ash alteration. At Site 1173, smectite abundance in the clay-size fraction generally increases with depth above the protodécollement, locally reaching values as high as 50-60 wt% of the clay-size fraction at ~390 mbsf, then drops markedly to ~40 wt% of the clay-size fraction below the protodécollement [Steurer and Underwood, 2003]; illite content increases slightly across the same boundary. Clay diagenesis is more advanced at the prism sites; at Site 1174, the onset of the smectite to illite transformation appears to have shifted ~100 m up-section relative to the age-equivalent décollement horizon, to a sub-bottom depth of ~700 m, where smectite abundance reaches ~40-50 wt% of the clay-size fraction. The smectite to illite transformation, however, has not gone to completion within the deeper sediments [Steurer and Underwood, 2003]. At both sites, the interpreted onset of the smectite diagenesis corresponds to the downhole loss of the dispersed fine-grained clay phase, which Sunderland and Morgan [2004] tentatively correlate with smectite (Figures 6 and 7). Along the Ashizuri transect, smectite diagenesis is not as far advanced due to the lower geothermal gradient; smectite occurs at concentrations of ~40-60 wt% of the clay size fraction throughout the base of the drill hole at reference Site 1177 [Steurer and Underwood et al., 2003].

Although additional analyses remain to be conducted, the limited evidence to date suggests that significant diagenetic changes, involving both clay and silica phases, accompany sediment burial, accretion, and presumably subduction along the Nankai margin. No doubt, these changes introduce complicating influences on sediment strength [e.g., Vrolijk, 1990; Saffer et al., 2001]. Of particular relevance to this study, opal and authigenic clays can serve as cementing agents that prevent sediment consolidation under ambient stress conditions [O’Brien et al., 1989; Dadey et al., 1991; Tribble et al., 1992; Tribble and Wikens, 1994]. The breakdown of these cements,
either by mechanical processes or diagenic alteration, can have dramatic effects on physical properties and deformation behavior, influencing the structural evolution of the Nankai prism.

7. Sediment Strength and In-situ Stress State

We can gain first order estimates of downhole effective stress conditions at the drill sites by comparing sediment porosities to the load imposed by the sediment overburden [e.g., Lambe and Whitman, 1979]. The dearth of deformation structures in cores recovered at Site 1173, and the downhole decrease in porosities in the LSB facies at Site 1173, suggest that the deeper sediments are normally consolidated [e.g., Screaton et al., 2002; Saffer, 2003]. Therefore, we can use the characteristics at this site to establish the reference state for sediments that will be incorporated into the accretionary prism. We estimate in-situ effective vertical stresses throughout the hole by integrating the densities of the overburden at a given depth [e.g., Morgan and Ask, 2004], assuming there are no excess pore pressures acting on the sediment, i.e., $P_\ast = 0$. We define this as hydrostatic vertical stress, $\sigma_{vh}'$. If pore pressures in excess of hydrostatic occur, the effective vertical stress is reduced, i.e., $\sigma_v' = \sigma_{vh}' - P_\ast$. These definitions are explained in further detail elsewhere [Morgan and Ask, 2004].

In the upper portion of Site 1173, i.e., over the lower range of $\sigma_{vh}'$, corrected porosities are very scattered, a result of the sandier lithologies in the Outer Trench Wedge unit (Figure 12). Between $\sigma_{vh}'$ of 0.5 and 2.0 MPa, corresponding to the USB facies, porosities are deflected to anomalously high values, probably due to the presence of some matrix cement which resists sediment consolidation [e.g., SSP, 2001a, Karig, 1993]. The rapid reduction in porosity across the USB-LSB facies boundary at Site 1173 (Figure 4c) reflects a downward decrease in mechanical strength of the sediment. In the deeper LSB facies at Site 1173, between $\sigma_{vh}'$ of 2.0 and 6.0 MPa, sediment porosities lie along a distinct curvilinear trend, which can be fit to a simple logarithmic relationship,

$$\eta = 65.4 - 0.409 \log(\sigma_{vh}') \cdot 100\%.$$  

(1)
This $\eta - \sigma_{vh}'$ relationship serves to predict for a given porosity, the maximum in-situ $\sigma_v'$ that the sediment has experienced in the past, referred to as the preconsolidation stress, $\sigma_c'$ (Figure 12). Porosities in the underthrust packages at the two prism Sites 1174 and 808 are clearly offset to higher values relative to the reference curve (Figure 12), suggesting a change in stress state or strength across the décollement. Several explanations for the offset in underthrust porosities have been proposed:

(1) Change from tectonic to uniaxial stress state across the décollement: Bray and Karig [1985] documented a slight decrease in porosities between the reference and prism sites to the southwest, which they interpreted to reflect the increase in mean stress due to tectonic loading. Similarly, the step-up in porosities observed below the décollement zone at Sites 1174 and 808 could reflect the reduced horizontal stress state in the underthrust package [Morgan and Karig, 1995b]. The general coincidence of porosities at Sites 1174 and 808 with the reference consolidation curve, however, suggests that the accreted sediments have not experienced significantly greater tectonic consolidation (Figure 4).

(2) Excess pore pressures within the underthrust sediments that prevent sediment consolidation: In this case, a lower effective stress state within the underthrust package would limit the amount of porosity reduction [e.g., Taira et al., 1992]. If so, $\sigma_v' < \sigma_{vh}'$, and the plotted porosities should be shifted upward to lower values of $\sigma_v'$ coincident with the reference line. The difference between $\sigma_{vh}'$ and $\sigma_v'$ serves as an estimate of $P_*$ [e.g., Screaton et al., 2002; Saffer, 2003]. Sediment yield strengths, therefore, could be predicted by the reference consolidation curve for a given value of $\eta$.

(3) Enhanced matrix strength, e.g., due to matrix cementation, allowing sediments to resist consolidation: The enhanced strengths would allow the sediments to maintain anomalously high porosities despite increased burial loads, as has been suggested for the USB sediments within the Shikoku Basin [Karig, 1993; SSP, 2001b]. In this case, the in-situ strength and associated yield stress during reconsolidation could be as high or higher than that predicted by either the in-situ consolidation curve, or hydrostatic vertical stress, $\sigma_{vh}'$. 


The last two possibilities are very different and have distinct implications for the deformation of sediments with increased burial and subduction. Higher $P_*$ and lower $\sigma_v'$ beneath the décollement would imply weak underthrust sediments, which might be susceptible to décollement downcutting and deformation. In contrast, if the underthrust sediments are stronger, they can resist such deformation to great depth, until subjected to increased tectonic stresses or pore fluid pressures.

Laboratory experiments provide a means to discriminate between overpressured sediments that have undergone normal consolidation (Option 2), and sediments that have been hardened by cementation or related diagenetic processes (Option 3) [e.g., Karig, 1993; Morgan and Ask, 2004]. During normal consolidation, sediment porosities, $\eta$ are assumed to be in equilibrium with the effective vertical stress, $\sigma_v'$ [Lambe and Whitman, 1979; Jones, 1994]. This relationship is represented by an essentially straight line in log($\sigma_v'$) - $\eta$ space (Figure 13a) over the higher stress ranges considered here [e.g., Karig and Ask, 2003]. The initial phase of uniaxial reconsolidation is characterized by elastic reloading with relatively little porosity change (Figure 13a), as indicated by the low value for the modified compression index, defined as $C_{c\eta} = -\Delta \eta / \Delta \log(\sigma_v')$ [Karig and Ask, 2003]. At the point of sediment yield, the sample experiences accelerated vertical strain and porosity loss, marked by a higher plastic $C_{c\eta}$ as the path returns to the normal consolidation trend. Uncemented sediments (Path 1) generally experience sediment yield near the intersection of the reload path with the normal consolidation line (Figure 13a). The sample’s preconsolidation stress, $\sigma_v'$, denotes the maximum value of $\sigma_v'$ that the sample has experienced in the past [e.g., Lambe and Whitman, 1979]. In contrast, if sediments have been hardened by secondary consolidation or cementation, they may support vertical loads in excess of their past maximum $\sigma_v'$ (Path 2) [e.g., Mitchell, 1993; Jones, 1994; Karig and Morgan, 1994].

A plot of differential stress, $\Delta \sigma = \sigma_v' - \sigma_h'$, against effective mean stress, $\sigma_m' = (\sigma_v' + 2\sigma_h')/3$ ($\sigma_h'$ is effective horizontal stress) may provide a more sensitive measure of sediment stress history [e.g., Jones, 1994; Karig, 1993]. As shown in Figure 13b, at low $\Delta \sigma$, uncemented sediments (Path 1) undergo elastic reloading marked by relatively low slopes (Phase
II); increased slopes denote initial yield as the stress path regains the normal (i.e., primary or $K_0$) consolidation line (Phase III). Due to their greater bulk moduli, cemented materials initially exhibit steeper reload paths (Path 2), and are characterized by three sub-phases following sediment yield (Figure 13b): Phase IIIA shows increasing $\Delta \sigma$ with $\sigma_m'$ but with decreasing slopes as sediment cohesion is lost; Phase IIIB initiates once the peak strength, $\Delta \sigma_{\text{peak}}$ is attained, and is marked by a decline in $\Delta \sigma$ with $\sigma_m'$ as intergranular cement rapidly breaks down. The sample eventually returns to $K_0$ consolidation path during Phase IIIC. The effective vertical stress at the point of $\Delta \sigma_{\text{peak}}$ is termed $\sigma_{\text{peak}}'$, and marks the onset of rapid porosity loss on the $\eta$- log($\sigma_v'$) plot (Figure 13a).

In order to determine the correct cause for the porosity step across the décollement horizon, we examine the results of several laboratory reconsolidation experiments carried out on samples collected from the frontal region of the Nankai Trough and prism.

8. Experimental Deformation Studies

Reconsolidation tests were carried out on whole-round cores from the USB facies at reference Site 582 to the southwest of the Muroto transect, and on the underthrust sediments at the three Muroto sites [Karig, 1993; Morgan and Ask, 2004], that is, both regions that show anomalously high porosities. The sample locations and intervals, and results of the experiments are shown in Figure 3 and listed in Table 2. (For simplicity, sample names are abbreviated by Site-Core throughout the text). Subcored samples had diameters of ~20 mm and heights of ~50-55 mm (i.e. ~2.5 times the diameter). Laboratory porosities were recalculated from the wet volume and weight of the sample, and bulk densities, assuming shipboard values for grain- and water densities [SSP, 2001a,b]. All of the experiments were carried out in a triaxial rig located at Cornell University, under steadily increasing vertical stress with no lateral strain, designed to reproduce uniaxial consolidation and unloading paths assumed in both settings. Both vertical and horizontal stresses, pore pressures, and dimensional and volumetric strains were closely monitored during the tests. Details of the experiments are presented in the references cited
above; the laboratory facilities are described in more detail elsewhere [e.g., Karig, 1996; Ask, 2001].

Summary results of the six reconsolidation experiments conducted on sediments from the Nankai margin are plotted in Figures 14 and 15. A plot of $\Delta \sigma$ against $\sigma_{m}'$ (Figure 14) demonstrates that all six samples exhibit similar elastic reloading paths (Phase II), and undergo yield. Only four of the samples actually reached their peak strengths, $\Delta \sigma_{peak}$. Samples 582-63 and 582-73 (shown in gray) from the USB facies, and Samples 1173-51 and 1174-78 from beneath the décollement horizon. All of the deeper samples from the underthrust section exhibited distinctive precursory yield prior to reaching their peak strengths. Samples 1174-78, 808-74 and 808-85 all exhibit a transient decrease in $\Delta \sigma$ with $\sigma_{m}'$, followed by renewed increase, often at slightly lower slope. Unfortunately, both tests on Site 808 samples were terminated before reaching $\Delta \sigma_{peak}$, and the test results for the former are very noisy and difficult to interpret (Figure 14). In all cases, the tests exhibit stress paths that resemble those of cemented materials that undergo mechanical breakdown in cementation at $\Delta \sigma_{peak}$. Samples 582-63, 1174-78, and 1173-51 also showed a return to a nearly linear path consistent with the normal consolidation path shown in Figure 13b.

The response of sample porosity to increasing vertical stress is also very characteristic of cemented samples (Figure 13). All of the tests show typical elastic reloading phase, with relatively low modified compression index, $C_{cq}$, denoting little change in $\eta$ with increasing $\sigma_{v}'$ (Figure 15). The four samples that reached their $\Delta \sigma_{peak}$ underwent more rapid consolidation, referred to as porosity yield (small white arrows), defined by the intersecting tangents for the two linear portions of the $\eta$- log($\sigma_{v}'$) curves. The corresponding values of $\sigma_{peak}'$ (large white arrows), which correlate with the $\Delta \sigma_{peak}$ values from Figure 14, do not always correspond to the porosity yield stresses. The two samples from Site 808 appear to have never reached porosity yield. In all cases, however, the samples maintained porosities to $\sigma_{v}'$ values in excess of their hydrostatic vertical stresses, $\sigma_{w}'$ (white bars), against which the shipboard porosity data are plotted (Figure 15). Clearly, the sediments are capable of resisting consolidation at all possible in-situ $\sigma_{v}'$, at least at laboratory strain rates.
The enhanced strengths indicated by the laboratory experiments do not necessarily preclude the presence of excess pore pressure in the reference or underthrust sections. The experimental results indicate, however, that porosities alone do not uniquely estimate the in-situ $\sigma'_v$. Enhanced pore fluid pressures could serve to maintain a stable stress state in which intergranular cement can form and develop, thereby reducing the likelihood of further consolidation. Morgan and Ask [2004] consider the possibility that the precursory yield recognized for all of the LSB samples at values of $\Delta\sigma < \Delta\sigma_{\text{peak}}$ denote the in-situ $\sigma'_v$. If true, then all but one of the LSB samples may exist at or close to the predicted $\sigma_{\text{vh}}'$, i.e., subject to minimal values of $P$. Several repeat experiments have been carried on subcores from the same samples presented here, and all show the same general response and precursory yield strengths; these data are the subject of continuing study [e.g., Ask and Morgan, in prep].

The two suites of experiments, using samples from the USB at Site 582 and from the LSB below the protodécollement, yield very distinct results. Despite their much lower porosities, the two Site 582 samples showed initial slopes in $\Delta\sigma - \sigma'_m$ space as high as the most consolidated of the LSB samples, Sample 808-85 (Figure 14). The high elastic moduli for these samples must arise from matrix cementation [Karig, 1993]. The USB samples also lack the precursory yield points detected in the deeper LSB samples. The two suites of samples are also distinguished by their sensitivities to applied load, defined as the ratio of sediment strength and preconsolidation stress predicted from the in-situ consolidation curve [e.g., Mitchell, 1993]:

$$S = \frac{\sigma_{\text{peak}}'}{\sigma'_c}.$$  \hspace{1cm} (2)

The two USB samples from Site 582 exhibit moderate sensitivities of 4.45 and 4.58 respectively (Figure 15). The LSB samples also have measurable, but lower sensitivities of 1.36 for Sample 1173-51, and 2.78 for Sample 1174-78. The limited data for the Site 808 samples suggest that these sediments continue to harden down-dip (Figure 15), implying increased sensitivity in the underthrust sediments with depth, which is likely to influence their deformation behavior within and beneath the prism.
9. Discussion

Sediments within several zones of anomalous porosity beneath and seaward of the Nankai prism exhibit mechanical evidence for some degree of cementation, which allows them to resist consolidation. This property is typical of many natural sediments, which are observed to harden by the combined processes of creep and chemical diagenesis [e.g., Burland, 1990; Mitchell, 1993; Jones, 1994]. The active cementing agents for the Nankai sediments are not known specifically, but appear to be different within the shallower USB sediments (e.g., silica) and the deeper underthrust sediments (e.g., authigenic clay). Changes in stress conditions with the onset of tectonic loading within and beneath the prism, however, can overcome this resistance to deformation, leading to the mix of deformation structures observed within the prism drill cores, and interpreted from seismic reflection data. Below, we explore plausible mechanical histories of the accreting and underthrusting sediments, and their resulting tectonic expressions.

9.1. Deformation within the Prism

Upon accretion, uniaxially deformed sediments are subject to a change in direction of the maximum principal stress, \( \sigma_1 \), from sub-vertical to sub-horizontal. The stress state also changes from axisymmetric to triaxial, defined by three distinct principal stress magnitudes [e.g., Karig and Morgan, 1994]. Consequently, the deformation history of the sediments must now be tracked through three independently varying mechanical quantities: \( \eta \) (or alternatively, void ratio, \( e \)), \( \Delta \sigma \), and \( \sigma_m' \). As summarized by the theory of critical state soil (or sediment) mechanics [e.g., Schofield and Wroth, 1968; Atkinson and Bransby, 1978; Wood, 1990; Jones, 1994], the allowable states in which most sediments exist are represented by a 3D yield surface defined in terms of these three quantities (Figure 16a). This surface is subdivided into a ductile (compactive) yield, or Roscoe, surface, and the brittle (dilative) failure, or Hvorslev, surface. The boundary between the two surfaces defines the critical state line (CSL), to which sediments will evolve with increasing \( \Delta \sigma \) and \( \sigma_m' \). For clarity, example stress paths are typically projected onto one of two planes of the 3D volume: \( \eta-\sigma_m' \) (Figure 16b) or \( \Delta \sigma-\sigma_m' \) (Figure 16c).
Initial uniaxial (vertical) consolidation within the Shikoku Basin leads to increasing compaction with increasing burial. This stress path (1) tracks along the ductile compactive yield surface (Figure 16a and b). The anomalous porosities within the USB (Figure 4c and 13) reflect post-burial cementation and strengthening, denoted by enhanced porosities and $\Delta \sigma$ relative to $\sigma_m'$, i.e., stress path (2), which lies above the ductile yield surfacee. With increased $\Delta \sigma$ and changing diagenetic conditions, the enhanced strength of this unstable material can break down, leading to renewed consolidation as stress path (2) returns to the ductile yield surface (Figure 16). Both stress paths (1) and (2) are consistent with the general decrease in porosities with depth at Nankai (Figure 4), and with penetrative ductile deformation resulting in increasing bedding parallel alignment of clay minerals documented through X-ray goniometry and SEM techniques [Morgan and Karig, 1993; Sunderland and Morgan, 2004].

Within the prism, there is evidence of further distributed ductile deformation, as well as localized deformation bands and microfaults. At first glance, this deformation history implies stress path migration away from the normal consolidation curve, and toward critical state - stress path (3) or brittle failure - stress paths (4) (Figure 16). In fact, the picture is slightly more complicated, due to the change in orientation of $\sigma_1$ from subvertical within the basin, to subhorizontal within the prism. This leads to a reversal in the sign of the differential stress, defined as $\Delta \sigma = \sigma_v' - \sigma_h'$. Consequently, $\Delta \sigma$ is positive within the basin and negative within the prism. Laboratory experiments [e.g., Wong et al., 1997] and theory [e.g., Wood, 1990] suggest that yield and failure conditions within the prism also will be sensitive to this reversal in sign. The yield surface is thought to be elliptical in shape and oriented symmetrically about the initial loading path. Thus, the yield strength for uniaxially consolidated sediments will be lower within the prism than in the basin for the same $\sigma_m'$ (Figure 17a). Given this knowledge, we can consider hypothetical stress paths that will lead to the observed deformation structures within the accreted sediments.

The active mode of deformation within the prism will be highly dependent on the initial cementation state, and the rates of build-up and dissipation of excess pore fluid pressures, $P$. The rotation of clay minerals away from the bedding planes implies tectonically induced ductile...
deformation [Morgan and Karig, 1993; Sunderland and Morgan, 2004], consistent with ductile yield under steadily increasing $\sigma_m'$, or “drained” conditions, i.e., low $P_*$. In this case, stress path (3t) diverges from the $K_0$ consolidation line and approaches the critical state line in the tectonic quadrant ($\Delta\sigma < 0$). The ductile yield surface must grow outward with stress path (3t), leading to penetrative ductile strain associated with porosity reduction (Figure 17a). If tectonic stress is built up more rapidly than $P_*$ can dissipate, however, deformation will occur under decreasing $\sigma_m'$, or undrained, conditions. For example, stress path (4t) retreats from the uniaxially consolidated state, staying within the yield surface and elastic field, bypassing the critical state line to undergo brittle failure (Figure 17a). Associated deformation structures include the low-angle microfaults and fractures found at depth within the proto-thrust zone (e.g., Figure 9c).

Cemented sediments as documented by our mechanical experiments will be subject to even more complicated stress paths during tectonic loading (e.g., Figure 17b). Sediments consolidated to state (1a) and cemented in-situ will sustain high $\Delta\sigma$, under both sub-vertical and sub-horizontal $\sigma_1$. Upon yield, however, these sensitive sediments will be subject to the sudden loss of cohesive strength, leading to rapid increases in $P_*$ and transient low $\sigma_m'$ conditions. If $P_*$ dissipation keeps up with the loading rate, the steady increase in $\sigma_m'$ will lead to penetrative ductile deformation and consolidation, i.e., stress path (3t′) (Figure 17b). If high strain rates and low permeabilities inhibit $P_*$ dissipation, however, the decrease in $\sigma_m'$ will favor localized failure, as shown by stress path (4t′). In contrast to the uncemented analog (Figure 17a), the dissipation of pore pressure following undrained failure of the sensitive cemented sediments will lead to localized compaction, rather than dilation. This offers a plausible origin for compactive deformation bands found within the USB sediments at Sites 1174 and 808 (Figures 9a-b). More consolidated, less sensitive sediments within the underthrust package may follow a similar undrained stress path, but undergo less total compaction due to their lower initial porosities (Figure 17b). This last stress path offers a plausible scenario for underplating of the underthrust sediments, due to the sudden imposition of subhorizontal $\sigma_1$ following seismogenic slip.
9.2. *Structural Progression within the Frontal Prism*

The distribution of deformation structures at the three drill sites along the Muroto transect: Site 1173 within the Shikoku Basin, Site 1174 within the protothrust zone, and Site 808 within the frontal prism, demonstrate a progression in deformation mode and intensity with accretion. Bedding parallel clay mineral fabrics develop early during burial, and undergo progressive rotation perpendicular to the convergence direction within the prism [e.g., *Morgan and Karig*, 1993; *Sunderland and Morgan*, 2004]. At some distance inboard of the deformation front, compactive deformation bands develop, initially at high angles to the subhorizontal $\sigma_1$, but subsequently at lower angles, to be finally overprinted by brittle deformation structures, particularly at depth, and close to throughgoing thrust faults [e.g., *Maltman et al.*, 1993; *Morgan and Karig*, 1995b; *Ujiie et al.*, 2004]. This spatial distribution is indicative of intensification of tectonic loading with distance from the deformation front, but may also denote progressive changes in shear strength and rheology as sediments experience mechanical and diagenetic changes.

One line of evidence in support of evolving material strength and rheology during sediment accretion lies in the occurrence of steeply dipping, reverse sense compactive deformation bands, commonly overprinted by more shallowly dipping deformation bands and microfaults [e.g., *Karig and Lundberg*, 1990; *Maltman et al.*, 1993; *Ujiie et al.*, 2004]. Previous explanations for the steep dips of deformation bands in the prism sediments include post-formation ductile rotation [*Karig and Lundberg*, 1990; *Sunderland and Morgan*, 2004], or alternating dilative and compactive deformation during with sediment dewatering [*Vannucchi and Tobin*, 2000]. Alternatively, compactive deformation bands lie along a continuum of allowable deformation states and geometries, ranging from pure compaction bands oriented normal to $\sigma_1$, to dilative shear bands oriented $\sim 30^\circ$ to $\sigma_1$ [e.g., *Rudnicki and Rice*, 1975; *Issen and Rudnicki*, 2000]. Pure plastic deformation, with neither dilation nor compaction, i.e., critical state, lies between the two end-members at $\sim 45^\circ$ to $\sigma_1$. Laboratory experiments on sand packs and sandstones have reproduced the full range of predicted localized deformation structures and orientations, dependent upon experimental confining pressures and initial porosity [e.g., *Zhang et al.*, 1990;
Wong et al., 1997; Olsson, 1999; Bésuelle, 2001; Fortin et al., 2004]. It is likely, therefore, that the abundance of high-angle compactive deformation bands within the shallower sediments at Sites 1174 and 808 is a result of the initial cemented state of the USB and shallower facies within the Shikoku Basin (Figure 4c). The first deformation bands initiate through ductile compactive yield at high angles (up to 60°) to the subhorizontal \( \sigma_1 \). The overprinting of increasingly shallower dipping deformation bands noted at Site 808 [Karig and Lundberg, 1990] implies a progressive decrease in sediment strength and porosity induced by mechanical or diagenetic breakdown of cement. With increasing consolidation, yield is accommodated by increasing components of shear relative to compaction, culminating in brittle and dilative faulting.

Spatial variations in deformation structures and fabrics correspond to an inferred temporal progression of deformation mode during accretion (Figure 18). Bedding parallel clay mineral fabrics define the reference state within the Shikoku Basin (Figure 18a). Near the deformation front, tectonic loading exceeds the shear strength of the cemented USB sediments, and intergranular cement breaks down heterogeneously along planar instabilities that form deformation bands at high angles to \( \sigma_1 \) (Figure 18b); localized compaction causes strengthening within the bands, arresting their propagation and transferring deformation onto adjacent bands [Rudnicki and Rice, 1975; Aydin and Johnson, 1983]. This produces zones of parallel bands, each with minimal displacement, as is observed in the Nankai prism toe [Karig and Lundberg, 1990; Ujiie et al., 2004]. Deeper more consolidated sediments respond to tectonic loading by the formation of brittle microfaults at low angles to \( \sigma_1 \) (Figures 19b and c). Concurrently, sediments throughout the accreted sediment column are undergoing bulk sub-horizontal consolidation, inducing penetrative clay mineral rotations documented by X-ray goniometry and SEM images [Morgan and Karig, 1993; Sunderland and Morgan, 2004]. In response, later stage deformation bands form at lower angles to \( \sigma_1 \). Eventually, reduced matrix permeability leads to increases in \( P_\perp \), driving deformation into the brittle field to form of microfaults and throughgoing thrust faults observed to cross-cut the ductile and brittle-ductile structures and fabrics (Figure 18c) [Karig and Morgan, 1994; Morgan and Karig, 1995b].
9.3. Implications for Décollement Processes

The histories of the accreted and underthrust structural domains, separated by the décollement, are distinctly different. The entire accreted package has been deformed in some way, as reflected by the abundance of brittle deformation structures, and bulk ductile deformation fabrics (Figure 19). Therefore, the prism sediments are undergoing constant tectonic “reworking.” Their cohesive strengths are lost through mechanical and diagenetic breakdown of intergranular cementation, and through brittle faulting. To date, however, the sediments beneath the décollement at both Sites 1174 and 808 have not encountered significant tectonic shear stresses, and therefore, remain within the upper quadrant shown in Figure 17b and c; presumably, they lie along the normal consolidation line or at slightly higher $\Delta\sigma$ conditions. The underthrust sediments remain coherent, and even strengthen with increasing burial (Figure 19). Due to their high strengths, these sediments can be carried 5 km or more landward without undergoing further consolidation [Morgan and Ask, 2004]. These underthrust sediments can maintain high porosities to even greater depths if ongoing diagenesis allows the mechanical strength to stay ahead of the increased burial loads.

The implied contrast in strength across the décollement is the inverse of that predicted by sediment porosities alone (e.g., Figure 4a-b), but can explain several key observations at the frontal Nankai prism. First, the décollement is mechanically constrained to lie at the boundary between these two domains; it cannot cut downward into the stronger underthrust package, at least beneath the prism toe. Second, the fault must thicken by progressive incorporation of the weaker accreted sediments. This requirement is consistent with a downward intensification of fracture and fragmentation of décollement sediments (Figure 10), and the gradational decrease in lateral anisotropy of the internal clay preferred orientation thought to be inherited from ductile deformation in the prism. These observations suggest that the deepest sediments within the décollement were incorporated into the fault early, undergoing substantial brecciation before significant ductile strain accrued [Morgan and Karig, 1995b].

The intense fracturing and the persistence of very thin shear surfaces of décollement breccia fragments, suggest that décollement slip occurs under high pore pressure conditions [Ujiie et al.,
The décollement fragments also appear to have undergone consolidation under essentially isotropic conditions, as might occur under fluctuating pore fluid pressure conditions if low permeabilities prevent the fluids from permeating into the dense fragments [Morgan and Karig, 1995b]. The confinement of the fragments by high pore pressures may have caused the mechanical breakdown of authigenic clays that cemented the matrix, thereby allowing further consolidation [Morgan and Karig, 1995b; Ujiie et al., 2003].

If the underthrust sediments are anomalously strong beneath the toe of the Nankai prism, then something must change down-dip beneath the décollement to explain the structural and hydrologic changes that we recognize at the up-dip limit of the seismogenic zone (Figure 2). No doubt, continued mineral dehydration, along with low-grade diagenetic and metamorphic reactions, contribute to these changes [e.g., Moore and Saffer, 2001]. Furthermore, intergranular cements may dissolve away under more favorable pressure, temperature, and fluid chemistry conditions at depth (Figure 19). We propose that in addition, sudden stress perturbations, for example, generated in response to seismogenic slip along the down-dip décollement, may help to overcome the cemented strength of the underthrust sediments, leading to rapid loss of shear strength, and rapid increase in pore fluid pressures (Figure 19). The lower effective stress state would favor brittle deformation within the underthrust package, enabling décollement downcutting, ultimately, to the top of the oceanic crust, and underplating of the underthrust package (Figure 19). With time, the underthrust package would consolidate, reducing the porosity contrast across the décollement, and therefore the impedance contrast, as suggested by Bangs et al. [2004]. The lateral changes in physical properties and strength, and associated geochemical and mineralogic changes, could alter the mechanical behavior and frictional properties along the fault [e.g., Zhang et al., 1993; Moore and Saffer, 2001; Brown et al., 2003; Kopf and Brown, 2003], enabling out-of-sequence faults to splay off of the new décollement surface, and possibly favoring unstable, and seismogenic sliding.

This hypothesis is based on characteristics and properties of remote sediments collected from the frontal portion of the Nankai prism along the Muroto transect. Along-strike differences in mineralogy, lithostratigraphy, and thermal and pore fluid conditions, as documented along the
Ashizuri transect [e.g., Moore et al., 1990, 2001; Underwood, this volume], will very certainly influence sediment state and deformation behavior beneath and along the décollement zone. In particular, the presence of sandy turbidites within the deeper USB section elsewhere along the margin, is presumed to decrease the potential for pore pressure build up within the underthrust package [e.g., Moore et al., 1990]; diagenetic processes within these sandy units, and their effects on sediment strength and sliding behavior, are very poorly constrained at this time, but may prove to be as complex or more so, than those we have elaborated here. The model presented here, therefore, still needs to be tested directly, e.g., through laboratory studies on materials from different locations along the Nankai margin, and ultimately through direct drilling and sampling of the seismogenic zone and adjacent fault rocks, an objective that will come to fruition as part of the NanTroSEIZE IODP drilling initiative [Kimura et al., 2003].

10. Conclusions

Clay-rich sediments at the toe of the Nankai accretionary prism, southwest of Japan, undergo diagenetic changes before, during, and after consolidation and deformation within and beneath the prism, leading to spatial and temporal variations in mechanical behavior and deformation modes. Sediments deposited and locally cemented within the Shikoku Basin are partitioned by the basal décollement into accreted and underthrust packages, and subsequently follow distinctive deformation pathways across the margin (e.g., Figure 19). The modes of sediment deformation depend on their initial consolidation and diagenetic state, and locally, their enhanced strengths due to intergranular cementation. Sediments above the décollement encounter increased horizontal tectonic stresses, which induce discrete brittle deformation in the form of microfaults and fractures, penetrative horizontal ductile deformation, and localized brittle-ductile deformation bands. Concurrent with tectonic deformation, sediments undergo progressive diagenetic transformations, causing a breakdown in sediment cementation. Accordingly sediments undergo changes in mechanical behavior, as ductile horizontal shortening associated with clay mineral rotations and porosity reduction overprints previous deformation structures.
Below the active décollement zone along the Muroto transect, slightly enhanced porosities indicate delayed consolidation due to sediment overpressures or cementation. Deformation experiments confirm the high yield strengths of the underthrust sediments, probably reflecting ongoing smectite-illite transformations within the deep sediments, and incipient matrix cementation through growth of authigenic clays. The degree of underconsolidation beneath the décollement increases in the down-dip direction, leading to enhanced sediment sensitivity. The enhanced strengths of the underthrust sediments beneath the toe of the Nankai accretionary prism, contrast with those of the deformed and continually remolded accreted sediments above, helping to guide the décollement horizon and evolution. At Site 1173, mineralogy and sediment fabric differences across the décollement suggest that the fault may ride near the top of a gradual diagenetic transition, possibly in the presence of locally enhanced pore fluid pressures. Beneath the frontal portion of the prism, the stronger underthrust sediments appear to resist décollement downcutting, forcing the décollement zone to thicken by preferential incorporation of accreted strata during shear deformation.

Down-dip, deep beneath the prism, higher pressures, temperatures, and chemical activities enable diagenetic alteration of the strong sediments. In addition, high shear stresses may develop near the up-dip limit of the seismogenic zone, breaking down the strength of the underthrust sediments, leading to loss of cohesion, shear failure, and release of trapped pore fluids. Associated rapid increases in pore pressure and reductions in shear strength along the fault zone will enable décollement down-cutting and the onset of underplating. Subsequent sediment consolidation will reduce contrasts in physical properties across the fault zone. This scenario provides a plausible explanation for the convergence of several structural and hydrological characteristics that have been interpreted close to the up-dip limit of the seismogenic zone beneath the Nankai prism, suggesting that the physical properties changes result in lateral contrast in décollement strength and sliding behavior responsible for the onset of seismogenesis.

The features and interpretations presented here are specific to the Muroto transect, for which the greatest abundance of data are available. Significant along strike variations in thermal structure, lithostratigraphy, and basement geometry will certainly influence the distribution and progression of deformation within the prism and along the decollement. Elsewhere along the
margin, other factors, such as the presence of interbedded turbidites within the underthrust section along the Azshizuri transect, will play important, and potentially, controlling roles in décollement placement and properties. Future drilling investigations will shift to the northeast, along a new transect that lies offshore of the Kii Peninsula (Figure 1), where these mechanical models for the partitioning and evolution of deformation can be put to the test, as sediments enter the accretionary system, and ultimately the deep seismogenic zone.

Acknowledgements

This research used samples and data provided by the Ocean Drilling Program (ODP), collected during ODP Legs 131 and 190, and Deep Sea Drilling Project samples from Leg 87. ODP is sponsored by the U.S. National Science Foundation (NSF) and participating countries under management of Joint Oceanographic Institutions (JOI), Inc. Shore-based experiments were conducted at Cornell University. Dan Karig provided thoughtful advice and throughout the study. George Hade is gratefully thanked for engineering assistance that formed the basis for the well-functioning test equipment. X-ray goniometry analyses were carried out at the Cornell Materials Science Center; Maura Weathers provided technical advice and assistance. We also thank Angelo Benedetto for training and assistance with SEM imaging at Rice University, Kitty Milliken for use of the University of Texas Electron Imaging Facility, and Scott Stookey at CoreLab, Houston. We thank Mike Underwood and Paola Vannucchi for very complete and constructive reviews of an earlier version of this manuscript. Support for the research presented here came from JOI-USSSP grant 190-F001330 to Morgan at Rice University, and past NSF and JOI funding.
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Ask M.V.S., Mechanical tests on claystone from Ocean Drilling Program Hole 1070A (Leg 173); Implications for elevated pore-fluid pressure in sediments within the ocean-continent transition zone, West Iberia, Marine Geology, 177, 395-410, 2001.


Shipboard Scientific Party, Site 1177, *Proc. ODP, Init. Repts.*, 190, Ocean Drilling Program, College Station, TX, 2001c.


Tables

Table 1. Common symbols used in text.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\eta$</td>
<td>Sediment porosity (%), volume of pore fluid/total volume,</td>
</tr>
<tr>
<td>$P_*$</td>
<td>Excess pore fluid pressure (MPa),</td>
</tr>
<tr>
<td>$\sigma_v$</td>
<td>Total vertical stress (MPa),</td>
</tr>
<tr>
<td>$\sigma_v'$</td>
<td>Effective vertical stress (MPa),</td>
</tr>
<tr>
<td>$\sigma_h'$</td>
<td>Effective horizontal stress (MPa)</td>
</tr>
<tr>
<td>$\sigma_{vh}'$</td>
<td>In-situ effective vertical stress for hydrostatic pore pressure (MPa),</td>
</tr>
<tr>
<td>$K_0$</td>
<td>Stress ratio during uniaxial strain, $K_0 = \sigma_h'/\sigma_v'$</td>
</tr>
<tr>
<td>$\Delta\sigma$</td>
<td>Differential stress in tests (MPa), $\Delta\sigma = \sigma_v' - \sigma_h'$</td>
</tr>
<tr>
<td>$\sigma_m'$</td>
<td>Effective mean stress in tests (MPa), $\sigma_m' = (\sigma_v' + 2\sigma_h')/3$</td>
</tr>
<tr>
<td>$\sigma_y$</td>
<td>Effective vertical yield stress in tests (MPa)</td>
</tr>
<tr>
<td>$\Delta\sigma_{peak}$</td>
<td>Peak strength in tests (MPa)</td>
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<tr>
<td>$\sigma_{peak}$</td>
<td>Effective vertical peak stress at $\Delta\sigma_{peak}$ in tests, (MPa)</td>
</tr>
<tr>
<td>$\sigma_c'$</td>
<td>Effective preconsolidation stress of sample (MPa), estimated from $\eta$ and Figure 3</td>
</tr>
<tr>
<td>$\sigma_{y'}$</td>
<td>In-situ effective vertical stress (MPa)</td>
</tr>
<tr>
<td>$C_{c\eta}$</td>
<td>Modified compression index, $C_{c\eta} = -\Delta\eta/\Delta\log(\sigma_v')$</td>
</tr>
<tr>
<td>$V_p$</td>
<td>Compression velocity</td>
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</table>

Table 2. Sample identifications and depths, and principal values and results of reconsolidation tests (from Karig [1993] and Morgan and Ask [2004])

<table>
<thead>
<tr>
<th>Sample (cm)</th>
<th>190-57-64</th>
<th>190-57-64</th>
<th>131-74R-2, 114-130</th>
<th>131-808C-85R-1, 22-38</th>
<th>87-582-63-2, 130-150</th>
<th>87-582-73-1, 68-84</th>
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<tbody>
<tr>
<td>Depth (mbsf)</td>
<td>476</td>
<td>881</td>
<td>1004</td>
<td>1098</td>
<td>644</td>
<td>744</td>
</tr>
<tr>
<td>Depth below décmt (m)</td>
<td>87</td>
<td>41</td>
<td>41</td>
<td>134</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>$\sigma_v'$ (MPa)</td>
<td>3.3</td>
<td>8.6</td>
<td>9.9</td>
<td>10.8</td>
<td>5.6</td>
<td>6.4</td>
</tr>
<tr>
<td>$\sigma_{peak}$ (MPa)</td>
<td>2.95±0.03</td>
<td>4.97±0.03</td>
<td>3.03±0.05</td>
<td>6.40±0.80</td>
<td>6.90±0.20</td>
<td>9.0±0.50</td>
</tr>
<tr>
<td>$\sigma_c'$ (MPa)</td>
<td>3.20</td>
<td>5.20</td>
<td>6.40</td>
<td>7.70</td>
<td>1.8</td>
<td>2.4</td>
</tr>
<tr>
<td>$\sigma_{y'}$ (MPa)</td>
<td>4.0±0.05</td>
<td>13.8±0.2</td>
<td>&gt;13.0</td>
<td>&gt;13.5</td>
<td>8.01±0.20</td>
<td>11.0±0.20</td>
</tr>
<tr>
<td>$P_f^{max}$ (MPa)</td>
<td>0.35±0.03</td>
<td>3.57±0.03</td>
<td>6.87±0.05</td>
<td>4.40±0.80</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>$S$</td>
<td>1.25</td>
<td>2.65</td>
<td>&gt;2.03</td>
<td>&gt;1.75</td>
<td>4.45</td>
<td>4.58</td>
</tr>
</tbody>
</table>

*Sample did not reach state of primary consolidation
**Figure Captions**

**Figure 1.** Location map for of DSDP and ODP drill sites in the Nankai Trough and frontal accretionary prism. Sites 808, 1174, and 1173 lie along the Muroto transect; the Ashizuri transect is located ~100 km to the southwest. A 3D seismic reflection survey parallels the Muroto transect.

**Figure 2.** Interpretive cross-section across the Muroto transect of the Nankai accretionary margin, showing structural domains (modified after *Moore et al.* [2001b]). A step down in the décollement near the frontal out-of-sequence thrust coincides with modeled peak pore fluid pressures (shaded ellipses) from *Saffer and Bekins* [1998]. The aseismic to seismic transition lies just landward of this zone. Boxed region defines frontal prism, enlarged in Figure 3. Stippled region denotes underthrust sediments that become underplated landward of the out-of-sequence thrust.

**Figure 3.** Interpreted cross-section across frontal portion of the Nankai accretionary prism, along the Muroto Transect. Simplified stratigraphy is indicated by numbered units: landward thickening Trench Wedge (1), on-laps the Trench-to-Basin Transition (2). The Upper Shikoku Basin (i.e., USB, 3) and Lower Shikoku Basin (i.e., LSB, 4) facies are underlain by thin volcanoclastic unit (5) and basaltic oceanic crust (6). The décollement zone lies within the Lower Shikoku Basin facies, and is correlated seaward of the deformation front along the age equivalent horizon referred to as the protodécollement. The locations of the three ODP drill sites are shown; reference Site 1173 lies within the Shikoku Basin, Site 1174 passes through the protothrust zone, and Site 808 penetrates the frontal thrust fault. DSDP Site 582 to the southwest is projected into this transect. Locations of samples discussed in this study are indicated by white circles (gray circles for Site 582).

**Figure 4.** Shipboard porosities for the three ODP drill sites, and acoustic velocities for two sites: (A) Site 808, (B) Site 1174, (C) Site 1173 [*SSP*, 1991, 2001a]. Porosity values plotted here are not corrected for smectite interlayer water released during oven drying [*Brown and Ransom*, 1996]; the low smectite abundance in the bulk sediments (~10 wt%) at Nankai results in corrections of much less than 1%. The locations and approximate thicknesses of the
décollement zone and faults are indicated. Heavy dashed line shows calculated effective hydrostatic vertical stress with depth, $\sigma_{vh}'$. Sample positions and labels are denoted to left of each plot.

**Figure 5.** Results of X-ray pole figure goniometry analyses for Site 808 samples [Morgan and Karig, 1995b]. (A) Clay mineral orientation distribution reveals a moderate total anisotropy ($A_3$) between 20-30%, reflecting vertical flattening of sediments during consolidation. Lateral anisotropy ($A_3$) is ~10% above the décollement, denoting rotation of clay minerals away from bedding planes in response to horizontal tectonic shortening. $A_3$ decreases to 0-5% below the décollement, and $A_2$ increases slightly, consistent with uniaxial strain state. (B) Strain ellipsoid shape factor distinguishes the distortion of the sediment in response to tectonic and compaction strains. Nearly all samples lie within the field of flattening, reflecting the persistence of the bedding-planar clay fabric, but increasing constriction with depth implies increasing horizontal shortening. Again, the underthrust sediments exhibit almost pure flattening. Breccia fragments from within the décollement exhibit a gradation from flattening at the base, to modest constriction near the top, suggesting progressive incorporation of increasingly ductilely deformed sediments as the décollement grows.

**Figure 6.** Representative SEM images of LSB sediments from Site 1173. (A) Sample 190-1173A-35X-6, 114-118 cm (330.2 mbsf), taken from above the proto-décollement horizon at the seaward reference site, shows wide size range of clay particles and random microfabric. (B) Sample 190-1173A-51X-2, 42-64 cm (476.3 mbsf), collected below the proto-décollement zone at the reference site, shows more uniform grain size, moderate clay preferred orientation, and moderate sized secondary pores.

**Figure 7.** Representative SEM images of sediments from Site 1174. (A) Sample 1174B-21R-2W, 46-49 cm (335.1 mbsf) derives from the Outer Trench Wedge facies, and reveals domainal microstructure defined by stacks of clay minerals separated by more open microstructure. A wide range of particle sizes are evident, with a distinctive dispersed fine-grained clay phase; up is toward the top. (B) Sample 1174B-58R-3W, 11-15 cm (691.1 mbsf) is from the LSB facies, with well-developed, slightly inclined stacks, and a range of particle sizes; up is toward the top. (C) Sample 1174B-67R-2W, 101-120 cm (775.4 mbsf) from the LSB facies is characterized by
more uniform, dispersed clay minerals of uniform size and shape. Bedding planar preferred orientation is subtle; up is toward the top. (D) Sample 190-1174B-78R-1, 82-104 cm (880.8 mbsf) comes from below the décollement, and exhibits uniform size and shape large grains with a well developed clay preferred orientation and large secondary pores; up is toward top of page.

**Figure 8.** Representative deformation structures from Site 1173. (A) Healed high-angle normal fault showing brecciation in the footwall (Interval 190-1173A-33X-4, 111-122 cm; 308.2 mbsf). (B) Mineralized fluid escape structure, showing zone of sediment brecciation near the base, deflected sediment horizon, and funnel-shaped top (Interval 190-1173A-53X-5, 89-97 cm; 500.6 mbsf). Modified from *SSP* [2001a].

**Figure 9.** Representative deformation structures from Site 1174. (A) Deformation bands from Interval 190-1174B-17R-2, 103-112 cm (297.1 mbsf), show characteristic tabular form, with anastamosing strands of varying widths. Note darker color within the bands, which correlates with decreased porosities, and also high dip angles. (B) Conjugate sets of deformation bands, again with very variable widths from Interval 190-1174B-15R-2, 19-24 cm (277.2 mbsf); one set dips more steeply than the other. (C) Inclined fracture sets from Interval 190-1174B-18R-2, 17-33 cm (306.0 mbsf), possibly contained within a prism backthrust. Modified from *SSP* [2001a].

**Figure 10.** (A) Fracture intensity across the shipboard defined décollement fault zone based on breccia fragment size, from *SSP* [2001a]. Intensity of deformation increases downhole within the recovered intervals. Narrow zones of foliated breccias reflect the most intensely deformed zones; zones of severe shattering are recognized in several locations. (B) – (F) Intervals reflecting types of deformation from within the décollement zone. Base of the décollement is sharply defined, and immediately overlain by a layer of highly foliated breccia. The top of the décollement zone is characterized by high-angle inclined fracture sets that may correspond to the base of incipient thrust zones within the proto-thrust zone.

**Figure 11.** Representative SEM images of breccia fragments from the décollement zone at Site 1174. (A) Sample 1174B-71R-2W, 69-73 cm (813.57 mbsf) shows dense arrangement of large, dispersed, uniform size clay minerals; note lack of well developed clay preferred orientation. Bedding orientation unknown. (B) Sample 1174B-71R-3W, 23-28 cm (814.39 mbsf) shows
similar dense packing of uniform clay minerals. (C) Sample 1174B-72R-3W, 103-105 cm (826.23 mbsf) shows thin zone of well-oriented clays polishing the surface of a breccia fragment; note the dense internal fabric of the breccia fragment. (D) Sample 1174B-71R-2W, 69-73 cm (813.57 mbsf) is close-up of thin oriented clay polish, with deep groove that exposes the less well-developed internal fabric.

**Figure 12.** Shipboard porosities plotted against effective hydrostatic vertical stress, $\sigma_{vh}'$. Only uniaxially consolidated sediments are shown: Site 1173 – open circles; Site 582 – solid diamonds; Site 1174, below décollement – open squares; Site 808 below décollement – crosses. A best fit porosity-stress relationship (solid line, and equation) is derived for porosities from the deeper LSB facies at Site 1173. This relationship predicts the effective vertical preconsolidation stress, $\sigma_c'$, which serves as an estimate for in-situ effective stress, $\sigma_{v*}'$. See text for discussion.

**Figure 13.** Schematic plots of reconsolidation experiments for normally consolidated (1) and cemented (2) sediment. (A) Plot of $\eta$ vs. $\log(\sigma_v')$ transforms the porosity stress relationship into a straight line. Experiment 1 shows initial elastic reconsolidation, characterized by a low initial slope, followed by sediment yield, reduction in $\eta$, and increase in slope, marking a return to normal consolidation; $\sigma_c'$ is estimated at the intersection of tangent lines. Experiment 2 exhibits similar elastic reconsolidation, but overshoots the predicted $\sigma_c'$ due to enhanced matrix strength imparted by intergranular cementation. (B) plot of $\Delta\sigma$ vs. $\sigma_m'$ is more sensitive to changes in sediment stress state. Path 1 is marked by initially low elastic slope followed by yield and return to normal consolidation trend; Path 2 shows more rapid rise in $\Delta\sigma$, precursory yield at $\sigma_y'$, leading to peak strength, $\Delta\sigma_{peak}$ at $\sigma_{peak}'$, and finally strain softening as the sample returns to normal consolidation trend – yield and peak stresses are shown in Figure 13a.

**Figure 14.** Differential stress, $\Delta\sigma$ plotted against effective mean stress, $\sigma_m'$ for all six tests; Site 582 test results are shown in gray [Karig, 1993; Morgan and Ask, 2004]. All tests show similar stress paths with initially steep elastic reloading, followed by yield and reduced rate of loading. Samples 1173-51 and 1174-78, as well as 582-63 and 582-73, both reached peak strength, $\sigma_{peak}'$ conditions, followed by strain softening; three of the tests return to normal consolidation paths. Samples 808-74 and 808-85 appear to have never reached peak strength conditions.
Figure 15. Plot of $\eta$ vs. log($\sigma_v'$) for all four tests; shipboard porosities are also plotted vs. log ($\sigma_{vh}'$) for uniaxially deformed sediments: Site 1173 (open circles), Site 582 (solid diamonds), Site 1174 (open squares), Site 808 (crosses). Solid line shows linear fit to Site 1173 LSB porosities, defining the in-situ consolidation line and predictor for $\sigma_v'$ from Figure 12. Initial yield stress, $\sigma_y'$ from Figure 14 is indicated for each experiment (black arrows); $\sigma_y'$ is observed to approximate $\sigma_c'$ (black line) for all but Sample 808-74. All samples maintained porosities well beyond their predicted $\sigma_c'$, and also beyond $\sigma_{vh}'$ (white bars), indicating substantial matrix strength. Vertical effective stress at peak strength, $\sigma_{peak}'$ is picked for Samples 1173-51, 1174-78, as well as 582-63 and 582-73 (large white arrows). An alternative estimate for peak strength (small white arrows) is obtained from the intersection of tangents to the linear portions of $\eta$ vs. log($\sigma_y'$) data.

Figure 16. (A) A 3D surface defining yield and failure conditions in critical state “soil” mechanics; sediment behavior varies as a function of initial consolidation state and its stress path in $\eta$-$\Delta\sigma$-$\sigma_m'$ space. The light gray surface represents ductile (compactive) yield, the darker gray surface represents brittle (dilative) failure. The junction of the two surfaces defines the critical state line (CSL), which defines pure plastic deformation. Four hypothetical stress paths through this 3D volume are projected onto the $\eta$-$\sigma_m'$ plane (B), and the $\Delta\sigma$-$\sigma_m'$ plane (C): stress path (1) - normal uniaxial consolidation; stress path (2) - cemented sediments; stress path (3) - ductile (compactive) yield to failure along the CSL; stress path (4) - “overconsolidated” elastic sediments, reloaded to brittle failure. See text for discussion.

Figure 17. Projections of the full yield surface onto the $\Delta\sigma$-$\sigma_m'$ plane for both uniaxial ($\Delta\sigma > 0$) and tectonic loading ($\Delta\sigma < 0$) conditions. The yield surface is thought to be elliptical and centered on the consolidation stress path. Prism sediments exhibit lower shear strengths under tectonic, sub-horizontal $\sigma_1$ than under sub-vertical $\sigma_1$. Arrowed paths denote possible stress paths within and beneath the prism. (A) Representative stress paths for uncemented, normally consolidated sediments: stress path (3t) – compactive, ductile yield to critical state; stress path (4t) – brittle, dilative failure. (B) Representative stress paths for cemented, normally consolidated sediments: stress path (2) – progressive breakdown of cementation bonds during
renewed uniaxial consolidation; stress path (3t’) – compactive, ductile yield to critical state; stress path (4t) – brittle failure. See text for discussion.

**Figure 18.** Progression of coupled sediment deformation and diagenesis along the Muroto transect, modified from Sunderland [2003]. (A) Site 1173 is characterized by enhanced sediment porosities throughout the USB facies due to cementation (e.g., opal-A); diagenetic phase change (e.g., to opal-CT) at the base of USB facies allows for renewed vertical consolidation of the deeper sediments, resulting in increased bedding parallel clay mineral alignment. Smectite is found in increasing abundance above the proto-décollement horizon, and drops significantly below, suggesting that the future fault plane coincides with the top of the gradual smectite-illite transition. (B) Deformation front, with facies and diagenetic boundaries interpolated from adjacent drill sites. Both the inferred opal-A - opal-CT and smectite-illite transitions have shifted up-section relative to the décollement horizon, due to increased burial beneath the trench wedge sediments. Tectonic deformation is manifested initially through high-angle compactive deformation bands at low angles to sub-horizontal $\sigma_1$ within cemented USB and shallower facies, and through incipient microfaults within more consolidated LSB sediments. The upward migration of the diagenetic transitions and consolidation of the USB leads to enhanced domainal clay mineral preferred orientations. Displacement along the décollement horizon decouples the deforming sediments above from undisturbed strata below, and allows for continued diagenesis and cementation within the underthrust section. (C) Site 1174, within the protothrust zone, shows more advanced stages of deformation and diagenesis. The opal-A to opal-CT boundary has shifted upward to the base of the trench wedge facies; the top of the smectite-illite transition is now ~140 m above the décollement horizon. Diagenetic and mechanical breakdown of intergranular cement within the USB and trench wedge facies leads to distributed horizontal shortening, rotating clay minerals to steeper orientations. Lower-angle deformation bands overprint earlier generation; brittle proto-thrust faults splay from the décollement zone, overprinting ductile and brittle-ductile deformation structures. Underthrust sediments maintain comparatively high porosities and resist décollement downcutting due to their enhanced strengths from ongoing diagenesis.
Figure 19. Spatial and temporal variations in mode of deformation and tectonic strain that develops during sediment accretion and underthrusting. The progression summarized in Figure 18 produces ductile, brittle-ductile, and brittle deformation above the décollement zone. Sediment properties in the underthrust section are preserved beneath the toe of the prism, due to ongoing sediment diagenesis and enhanced sediment strengths. Continuous tectonic reworking of the overthrust sediments maintains weak prism sediments by comparison, driving deformation upward into the accretionary prism. Stress perturbations near the up-dip limit of the décollement seismogenic zone ultimately exceed the strength of the underthrust sediments, leading to rapid breakdown of cementation, increase in pore fluid pressures, and associated decrease in effective stress, allowing for décollement down-cutting and underplating. Subsequent consolidation and dewatering decreases the physical properties contrast across the décollement zone. Lateral changes in physical properties may result in a change in frictional behavior along the fault, leading to the onset of unstable seismogenic sliding along the dewatered décollement zone. Circled numbers refer to hypothetical stress paths discussed in the text.
Figure 5

Figure 6
Figure 7

Figure 8
Figure 10
Figure 12

1173 in-situ mudstone
\[ \eta = 0.654 - 0.409 \sigma_{vh}' \cdot 100\% \]
Figure 13

Figure 14
Figure 19