Magnetic fabric analyses of ocean floor sediments: characterizing depositional processes in the Nankai Trough and Shikoku Basin

Beth Novak
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Magnetic Fabric Analyses of Ocean Floor Sediments: Characterizing Depositional Processes in the Nankai Trough and Shikoku Basin

By

Beth Novak

Accepted in Partial Completion

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Master of Science

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MASTER’S THESIS

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Beth Novak
October 12, 2012
Magnetic Fabric Analyses of Ocean Floor Sediments: Characterizing Depositional Processes in the Nankai Trough and Shikoku Basin

A Thesis
Presented to
The Faculty of
Western Washington University

In Partial Fulfillment
Of the Requirements for the Degree
Masters of Science

By
Beth Novak
October, 2012
ABSTRACT

Magnetic fabric analysis was conducted on accretionary prism sediments and on sediments taken from the Shikoku Basin during Integrated Ocean Drilling Program (IODP) Expedition (Exp) 333 in order to characterize the sedimentary processes and depositional mechanisms active along the Nankai Trough. Sedimentary ages, in the form of a magnetostratigraphy, were also determined for these sediments during the course of this study. IODP Exp 333 focused coring efforts off the coast of the Kii Peninsula in Japan on three sites, one site within the Nankai Trough accretionary prism (Site C0018) and two sites seaward of the Nankai Trough in the Shikoku Basin (Sites C0011 and C0012). Exp 333 was designed to characterize the sedimentary sequence on the down-going plate and the uppermost igneous basement prior to their arrival at the Nankai Trough subduction front (Sites C0011 and C0012) and to constrain the submarine landslide history along the forearc slope as it relates to the megasplay fault (Site C0018). Sedimentary ages obtained from the magnetostratigraphy of the 314 m section at Site C0018 range from 0 to approximately 1.7 Ma (Pleistocene to Holocene). Sediments from Site C0011, a 380 m section, range in age from 0 to 7.12 Ma (Miocene to Holocene) and the sediments from Site C0012, a 180 m section, range in age from 0 to 8.25 Ma (Miocene to Holocene).

Anisotropy of magnetic susceptibility (AMS) measurements have been used to identify sediment disturbance and compaction disequilibria, to characterize the depositional mechanisms for each site, and to identify sediment transport directions. Oblate AMS fabrics are dominant at all three locations. At Site C0018 significant sediment disturbance due to submarine landslide events was recognized. Current deposition is an important mechanism at Sites C0018 and C0011. Depositional mechanisms at Site C0012 are dominated by slope
gravity deposition, which is gravity-controlled deposition of sediment in the water column onto an inclined surface, and the interaction of slope gravity with grain collision. The sediment transport direction at Site C0018 is predominantly to the SE. The paleocurrent at Site C0011 has had two distinct stages: SW from ~7.12 to ~5.24 Ma, and predominantly NW-SE from ~5.24 Ma to present, altering the accumulation of sediment at Site C0011. Site C0012 sediment transport directions are generally to the SE from 7.6 to 7.5 Ma and ESE from 7.5 Ma to 7.14 Ma. There is no distinct sediment transport direction recorded in the sediments from 7.14 to present at Site C0012. Compaction disequilibria were noted at Sites C0011 and C0012 and have been attributed to the presence of silica cement in the sediments.

Analysis of the paleomagnetic record with depth has been used to determine the effects of compaction and slumping on the paleomagnetic signal. In order to accurately define a relationship between inclination and compaction in these sediments, it is necessary to understand the depositional history and consolidation history at each site. Due to sediment disturbance caused by submarine landslide events at Site C0018, inclination values have been significantly altered and therefore are unusable in paleomagnetic reconstructions. Significant inclination shallowing is not recorded at Sites C0011 and C0012, which may be due to reduced compaction caused by the possible presence of silica cement in the sediments from both sites. Inclination error is not universally found in sediments and therefore sediments can record the geomagnetic field without significant bias.
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INTRODUCTION

Statement of Purpose

The goals of this study are to determine the ages and transport processes of sediments cored at three sites during Integrated Ocean Drilling Program (IODP) Expedition (Exp) 333: Sites C0018, C0011, and C0012 (Fig. 1). Magnetostratigraphy is used for age determination, and magnetic anisotropy is used to characterize sediment transport processes. The secondary goal of this study is to characterize the processes under which the geomagnetic field is recorded in clastic marine sediments. It is likely that sediment accumulation has occurred at a faster rate at Site C0018 than at Sites C0011 and C0012 due to Site C0018’s proximity to the Japanese coastline. A comparison of the sediments from the three sites will help reveal the changes in sediment deposition and behavior before and after entering an active subduction zone.

Geologic Setting

IODP Exp 333 cored ocean floor sediments and basalts in the Shikoku basin, a back-arc basin, off the coast of the Kii Peninsula in Japan (Fig. 1; Henry et al., 2012). Drilling and sampling were done aboard the deep sea-drilling vessel D/V Chikyu. The Shikoku Basin is a broad, fan-shaped depression in the northeastern part of the Philippine Sea plate. The basin is bounded to the north by the Nankai Trough and to the south by the Parece Vela Basin (Okino et al., 1993). Early studies of the magnetic anomaly lineations and topography of the Shikoku Basin suggest active spreading in the basin from 30 Ma to 15 Ma (Okino et al., 1993). Spreading of the Shikoku Basin ceased around 15 Ma. Following this period of tectonic activity, the Shikoku basin became inactive and is currently characterized by low
seismic activity (Okino et al., 1993). The basin is currently experiencing deposition of sediments and subsidence (Okino et al., 1993).

Coring during Exp 333 took place at three separate locations. Site C0018 is located in the Nankai Trough accretionary prism, while Sites C0011 and C0012 are located seaward of the Nankai Trench in the Shikoku Basin on the Kashinosaki Knoll, a prominent seafloor high (Fig. 2). Site C0011 is located on the northwestern flank of Kashinosaki Knoll next to a blind thrust, while Site C0012 is located near the summit of the knoll (Fig. 3; Henry et al., 2012).

Expedition Background

IODP Exp 333 Sites C0011 and C0012 are part of the second stage of the Nankai Trough Seismogenic Zone Experiment (NanTroSEIZE) program aimed at understanding subduction inputs and heat flow. Site C0018 was selected for the Nankai Trough Submarine Landslide History (NanTroSLIDE) program in order to obtain data on the up-dip terminus of the megasplay fault. The main objectives of the NanTroSEIZE program are 1) characterizing the mechanisms and properties controlling the up-dip aseismic-seismic transition along the megathrust and faults at the plate interface, 2) understanding the mechanisms associated with earthquake and tsunami generation, 3) characterizing strain accumulation and release, and 4) understanding the mechanical strength and hydrogeologic behavior at the plate boundary and within the megasplay fault (Henry et al., 2012). Exp 333 was initiated to address these questions by characterizing the composition and material properties of the ocean floor sediments and uppermost basalts of the ocean crust prior to deformation and subduction along the Nankai Trough (Henry et al., 2012). The specific goals of Exp 333 were to: (1) drill and core previously unsampled intervals of sediment and basalt at IODP Sites C0011
and C0012 in the Shikoku Basin, and (2) drill and core at Site C0018 near the up-dip terminus of the megasplay fault (Henry et al., 2012).

Research Goals

The main goal of my research is to characterize the sediment transport and depositional mechanisms at Sites C0018, C0011, and C0012 using the magnetic fabric record and to provide ages for the sediments at each site. These same analyses may be useful in characterizing the triggering events for mass transport deposits (MTDs) through the identification of surfaces of weakness within the sediments, which include sediments with higher water content, periods of faster sediment accumulation, and possible sediment overloading. Depositional processes play a large role in sediment behavior. Characterizing these processes at Sites C0011 and C0012 may also clarify the reason that the Site C0012 section appears to be condensed relative to the Site C0011 section (Saito et al., 2010; Henry et al., 2012). The depositional differences between these sites may lead to distinct behavioral differences when these sediments reach the deformation front.

The final aspect of my work is to investigate the effects of burial and compaction on magnetic inclination in ocean floor sediments. The (more or less) continuous accumulation of sediments can provide high temporal resolution records of the geomagnetic field, which cannot be obtained in terrestrial environments (Arason and Levi, 1990a). The sediments from Sites C0011 and C0012 are relatively undisturbed and are ideal for studying how magnetic inclination is recorded, and possibly altered, over long periods of time. The current understanding of the ancient geomagnetic field comes from studies of magnetism recorded in rocks and sediments. But the possibility of inclination error, especially in deep-sea sediments, may indicate the current record of the geomagnetic field is inaccurate (e.g.,
Applications of paleomagnetism to tectonic problems operate on the assumption that the time averaged geomagnetic field is approximately a geocentric axial dipole (GAD) and is rotationally symmetric around the North-South axis of the Earth (e.g. Tauxe and Kent, 2004; Tauxe et al., 2008). The GAD was developed based on inclination values from samples taken from well-distributed locations around the world. Sediment samples are a major component in these data sets. Because of the importance of sediments for the GAD hypothesis, our current geomagnetic field models may be influenced by significant inclination error (Arason and Levi, 1990b; Li et al., 2004; Tauxe and Kent, 2004; Tauxe, 2005). Variations in inclination due to burial and compaction were identified through changes in magnetic inclination values with regard to depth during the course of this study.

**DRILLING SUMMARY**

IODP standard procedure dictates a specific naming convention for each drill site and sample. A specific drilling location is referred to as a Site followed by an identification number (such as Site C0018). Each hole drilled at a site is assigned a letter, starting at the beginning of the alphabet (i.e. Hole C0018A). At each site and hole, 9.5 m lengths of cores are collected, numbered, labeled with a reference to the type of coring system used, and split into 1.5 m sections. The types of coring systems and the abbreviations are outlined in Table 1. Individual samples collected from a core are labeled as follows: Site, Hole, Core, Section, and the depth interval the sample was collected from in each section. For example, C0018A-1H-2-36-38 cm would indicate a sample taken from Site C0018, Hole A, Core 1 (collected with a hydraulic piston coring system as indicated by the ‘H’), Section 2, at a depth of 36-38
cm from the top of section 2. In an effort to condense sample names, reported sample depths are the center of each sample.

Drilling at Site C0018 reached a depth of 314.15 meters below sea floor (mbsf) with a recovery of 271.44 m of sediment (86.4% of the drilled interval). A total of 36 cores were collected at Site C0018 (Table 2). The lithologic units consist mainly of silty clay with varying amounts of volcaniclastic sand and ash (Fig. 4; Henry et al., 2012).

Within the cored sediments at Site C0018, six intervals with MTDs were recognized. The largest of these MTDs (MTD six) occurred below 127.55 mbsf and extended downward for 61.03 m (Fig. 4; Henry et al., 2012). The scientific party of Exp 333 relied heavily on visual evidence to identify the presence of an MTD (Henry et al., 2012). These visual cues included the absence of bioturbation, the presence of mixed sediments of differing grain sizes, and the recognition of convoluted strata within the MTDs. These same features were found to be indicative of remobilization within the MTDs.

At Site C0011 two holes (C0011C and C0011D) were drilled with approximately 380 m of sediment recovery (Fig. 5). Hole C0011C included three cores that reached a maximum depth of 22.5 mbsf with 22.99 m (102.2%) of sediment recovery (Table 2). Sediment recovery exceeds 100% for cores when the length exceeds the drilling depth, perhaps due to the in-flow of drilling fluids during the coring process. At Hole C0011D, 52 cores were collected for a total of 359 m of sediment cored with 351.49 m (97.9%) recovered. The lithologies cored at Holes C0011C and C0011D include, in order of abundance, silty clay, clay interbedded with volcanic ash, and minor quantities of volcaniclastic sand (Henry et al., 2012). The sediments range in induration from soft sediment to mudstone.
Site C0012 included five holes (C0012C, C0012D, C0012E, C0012F, and C0012G). The recovered lithologies are shown in Figure 6. Fifteen cores were collected at Hole C0012C with 124.5 m cored and 128.62 m (103.3%) recovered. Roughly 63.8 m were cored at Hole C0012D in 13 cores and 66.38 m (104%) were recovered for sampling (Table 2). Lithologies in Holes C0012C and C0012D include, in order of abundance, greenish gray clay and silty clay, silt interbedded with volcanic ash, and minor quantities of volcanioclastic sand with the frequency of ash layers changing throughout the extent of the cored sediments (Henry et al., 2012).

The remaining three holes (C0012E, C0012F, and C0012G) were planned to recover material from the transitional zone between the Shikoku Basin sediments and the basaltic ocean crust. Sediments from 178 to 500 mbsf were cored during Exp 322 and were not sampled during Exp 333 (Saito et al., 2010; Henry et al., 2012). Coring at C0012E began at roughly 500 mbsf. Lithologies in these cores include red calcareous claystones overlying pillow basalts (Henry et al., 2012). At Hole C0012E, three cores were drilled and 28.5 m of sediment were collected with recovery of 13.85 m (48.6%). The two cores collected at Hole C0012F cored 5.5 m of sediment with recovery of 2.65 m (48.2%). Finally, Hole C0012G included fifteen cores for a total of 115.5 m cored. Roughly 25.92 m (22.4%) of basalt were recovered (Table 2). Due to the large gap in the recovered sediments between Holes C0012D and C0012E, the samples from Holes C0012E-C0012G will not be covered in detail here.
PREVIOUS WORK

Previous IODP expeditions have sampled and studied sediments from the Nankai Trough accretionary prism and the Shikoku Basin. The IODP Exp 316 science party concluded that active deformation is occurring in the accretionary wedge. With the addition of 2-D seismic reflection surveys, the structure and thickness of the sedimentary units throughout the basin have been classified (Park et al., 2002). The seismic reflection data were key in identifying the optimal location for sampling submarine landslide deposits for Site C0018.

In an effort to characterize the sedimentary inputs to the subduction zone along the Nankai Trough, the NanTroSEIZE drilling program has cored twelve sites during previous expeditions (IODP Exp 314-316, 319, 322, 332). Sites C0011 and C0012 were previously cored during IODP Exp 322. The sediments were collected during Exp 322 with a rotary coring system, which resulted in poor recovery of the softer, typically shallower, sediments (Saito et al., 2010). The sediments at these sites were characterized as hemipelagic muds with red calcareous clay at the transition to the igneous basement (Saito et al., 2010; Henry et al., 2012). The sedimentary sequences at Sites C0011 and C0012 are comparable, but the Exp 322 science party noted that the sedimentary section at Site C0012 appeared to be condensed relative to the sedimentary section at Site C0011 (Saito et al., 2010). Sediments collected directly above the sediment-basalt interface at Site C0012 and those found below 340 mbsf at Site C0011 are dated as Miocene in age (Saito et al., 2010). Due to the poor recovery at both Sites C0011 and C0012, a detailed characterization of the sediments was not completed until Exp 333.
Multiple generations of seismic reflection studies have been conducted over many years along the Nankai Trough region (e.g. Park et al., 2002). These surveys have allowed for more specific site selection for drill sites and characterization of sediment thickness and deposit types before drilling. Site C0018 was selected based on these studies due to the identification of MTDs. These MTDs are characterized by the absence of a clear stratigraphy within the deposits. Site C0018 represents the downslope depositional center for multiple MTDs and appeared to include a 150 m thick MTD. A characterization of the fabric within these deposits will be useful in clarifying the deposition and remobilization behavior of submarine landslides.

METHODS

Sampling

The samples collected for this study during Exp 333 consist of 556 oriented sediment cubes. A summary of samples and the depths of recovery can be found in Tables 1A, 2A, and 3A. Discrete cubic samples (~7 cm$^3$) or minicores (~11 cm$^3$) were taken one per section (each section is roughly 150 cm in length) from the working halves of each core (Fig. 7; Henry et al., 2012). The sample spacing varied, depending on the properties of the core material (e.g., to avoid flow-in of drilling fluid, coring disturbances, sandy sediment, etc.) and the distribution of interbedded lithologies. The paleomagnetic sampling was concentrated on undisturbed sediments, typically hemipelagic mud and mudstone. The orientation framework of the discrete samples is shown in Figure 7 (Henry et al., 2012).
Paleomagnetism and Magnetostratigraphy

Standard practice in paleomagnetism is to measure Natural Remanent Magnetization (NRM) and to subject the sample to a series of demagnetization steps of increasing strength in an effort to isolate the original magnetization direction recorded within the sediments. Each demagnetization step removes a portion of the remanent magnetization vector. The remanent magnetization vector represents the orientation and intensity of the magnetism in a sample after demagnetization and is represented by a point on the orthogonal vector projection. Each point is the end of a vector that extends from the origin to the point; a set of points defines a demagnetization path for a specimen (Kirschvink, 1980). The direction of the remanent magnetization vector changes with each progressive step until a stable direction has been achieved. When a stable component is isolated, the vector will progress toward the origin, where remanent magnetization is zero, in a straight line. The stable vector that progresses toward the origin is typically referred to as the Characteristic Remanent Magnetization (ChRM). The ChRM is representative of the original magnetization direction recorded in the sediments after deposition.

A D-2000 Alternating Field Demagnetizer (AFD), which has a maximum demagnetizing field of 200 mT, was used to perform all demagnetization steps. The demagnetizer applies a reversing, anti-parallel magnetic field along the x, y, and z-axes of a sample and magnetic remanence is gradually removed with increasing peak fields (Stephenson, 1993). Each measurement is taken after the sample has been demagnetized along all three axes. Demagnetization was done in increasing field steps, and measurements were made between each step on a 2-G cryogenic magnetometer.
In some cases, specimens can gain magnetization during the demagnetization process due to mineral alignment or the magnetic mineralogy present in the sample (Stephenson, 1993; Dankers and Zijderveld, 1981). This special case has been termed ‘gyroremanent magnetization’ (GRM) and multiple methods exist for avoiding the effect of GRM (Stephenson, 1993). GRM is often recognized at fields higher than 60 mT with a systematic increase in the remanent magnetization and a curved trajectory of the remanent magnetization vector away from the origin in the orthogonal vector plot (Fig. 8A). Samples that gain a GRM are typically characterized by a decrease in the remanent magnetization in the early demagnetization steps, but between 60 and 80 mT the samples acquire a remanent magnetization direction along the last axis that is demagnetized and the remanent magnetization of these same samples will begin to increase (Fig. 8B; Dankers and Zijderveld, 1981, Stephenson, 1993). In order to avoid the effects of GRM, specialized demagnetization techniques, often referred to as Anti-Gyroremanent Magnetization (AGRM) techniques, must be used.

The AGRM method employed in this study involved systematically demagnetizing and measuring each sample along each axis prior to the demagnetization of the next axis. This was done for each of the three axes (x, y, and z) at each demagnetization level (typically starting between 60 and 80 mT). This process differs from the typical demagnetization technique because each sample is measured three times for each demagnetization level, rather than just once. The changes in the remanent magnetization vector after demagnetization along each axis are used to calculate a resultant vector. The resultant vector calculated from each component provides a result that is not affected by a GRM and can be
used to determine paleomagnetic directions (Dankers and Zijderveld, 1981; Stephenson, 1993).

The remanent vectors for all demagnetized samples are plotted on orthogonal vector projections (Fig. 9-11). Each diagram contains two projections of the vectors. Data are shown on the horizontal plane with solid symbols, while data on the vertical plane are denoted by open symbols. Declination and inclination are determined for each sample from the least-squares fit lines or planes, which are fit to the orthogonal vector projections.

The Mean Angular Deviation (MAD) value is used in order to evaluate the colinearity or coplanarity of a group of points based on the geometry of the data set (Kirschvink, 1980). Low MAD values indicate a stronger colinearity or coplanarity than large MAD values. Kirschvink (1980) used a MAD value of 15° as the standard for evaluation of the colinearity or coplanarity of the best-fit line or plane. Samples with MAD values below 15° have low vector direction variance, and the directions determined from the best-fit line or plane are estimates of the remanent magnetization vector. MAD values of 15° were used as the standard for evaluation throughout this study.

The inclinations of the ChRM directions, defined on the orthogonal vector projections (i.e. Figures 9-11), are used to define the polarity of each sample and polarity zones for each site. The polarity zones are used to define a magnetostratigraphy and are then correlated with the Geomagnetic Polarity Time Scale (GPTS). Due to the repeated reversal of the magnetic pole over time, inclination directions can be broken into two distinct polarity groups: normal and reverse. When the samples from one site are analyzed as a cohesive group, the polarity zones can be used to create a detailed magnetostratigraphy and date the sediments.
The magnetostratigraphy for each site was completed following the GPTS from Gradstein et al. (2004). Specific chron and subchron ages were assigned to the sediments. Boundaries between normal and reverse chronos were not placed at specific sample locations, but instead at the average depth between the samples. Tephrochronology and biostratigraphy ages have also been used to correlate the magnetostratigraphy and GPTS.

Tephrochronology and biostratigraphy ages provide constraints on the magnetostratigraphy ages and aide in defining more accurate age correlations between the magnetostratigraphy and GPTS. The tephrochronology ages were assigned based on the identification of specific ash layers within the sediments. The preliminary biostratigraphic ages for Sites C0011 and C0012 were determined from samples taken roughly every 10 m from the core catcher of each core during Exp 333 (Saito et al., 2010; Henry et al., 2012). The Site C0018 preliminary biostratigraphy included all core catcher samples as well as additional samples taken between the core catchers (Henry et al., 2012).

One of the primary assumptions in paleomagnetic studies is that the magnetic field averages to the GAD over a significant period of time. In order to assume GAD, the sediments collected for this study must meet the requirements for paleosecular variation (PSV). Sediment accumulation rates calculated during this study were used to determine if the 9.5 m cores spanned enough time to meet the PSV requirement. The averaging time for secular variation is not well agreed upon, but $10^4$-$10^5$ years are commonly noted as sufficient time periods (Tauxe, 2010). Individual cores from Site C0011 represent no less than 130 kyr, and cores from Site C0012 represent no less than 150 kyr based on the sediment accumulation rates. Accumulation rates for C0018 were taken from the Exp 333 biostratigraphy estimates and ranged from 6 cm/kyr (0-25 mbsf) to 23 cm/kyr (25-314 mbsf;
Henry et al., 2012). The cores from Site C0018 therefore represent no less than 40 kyr and meet the PSV requirement. Thus the GAD hypothesis is assumed for all cores and for the analysis of inclination with depth.

**Magnetic Susceptibility**

One of the primary methods that can be used to examine mineral alignment and to infer sediment transport characteristics and variations in compaction or deformation of sediments is the anisotropy of magnetic susceptibility (AMS). All rocks have a magnetic susceptibility regardless of their magnetic mineral composition (Tarling and Hrouda, 1993). The magnetic susceptibility (K) defines the relationship between the induced magnetization (M) and the small applied field (H) and can be written as:

\[ M_i = K_{ij} H_j \]

(Tauxe, 1998).

The letters \( i \) and \( j \) represent components of the susceptibility tensors. The intensity of the induced magnetization can vary by direction and is referred to as the AMS. AMS is geometrically represented by three principal axes: maximum (\( K_1 \)), intermediate (\( K_2 \)), and minimum (\( K_3 \)) axes of magnetic susceptibility (Fig. 12; Tarling and Hrouda, 1993). These three principal axes are also referred to as eigenvectors and are used to describe the orientation of the susceptibility ellipsoid. The susceptibility tensor (\( K_{ij} \)) is also used to calculate eigenvalues. The eigenvalues are used to describe the shape of the susceptibility ellipsoid (Tauxe, 2010). The shape of the ellipsoid is defined by the susceptibility values for each axis relative to the other axes.

The magnitude of anisotropy of susceptibility depends on two factors: the anisotropy of the particles themselves, and the degree of alignment of the particles (Tarling and Hrouda, 1993). Although each particle in a sample has distinct K axes, AMS measurements record
the average K axes for the entire sample. Samples with particles that are highly anisotropic and well aligned have a greater magnitude of anisotropy between the length of the $K_1$ and $K_3$ axes than samples with highly anisotropic particles that are randomly oriented. Grains with shape anisotropy typically fall with the $K_3$ axis perpendicular to the depositional plane, especially in the case of clay-rich sediments (Tauxe, 1998). The $K_1$ and $K_2$ axes typically fall sub-horizontal to the depositional plane. In a quiet environment, these axes will be randomly oriented in the horizontal plane resulting in an oblate anisotropy (Fig. 13; Tauxe, 1998). The susceptibility axes will be the most isotropic—reflecting the more random alignment of these particles—in the shallowest sediments. The degree of anisotropy is expected to increase with depth due to increased alignment of the particles caused by increased burial and compaction (Tarling and Hrouda, 1993; Tauxe, 1998; Kawamura and Ogawa, 2004; Schwehr et al., 2006).

AMS has become a well-established technique for studying mineral orientation fabrics in geologic materials (Tarling and Hrouda, 1993; Tauxe, 1998). AMS measurements are sensitive to the compaction state of clay-rich sediments (i.e. Housen et al., 1996; Kawamura and Ogawa, 2004; Schwehr et al., 2006). The most common uses of AMS in sediments include identifying consolidation trends in sediments, characterizing paleostresses, identifying disturbances within sediments, and determining paleocurrent direction.

Down-hole variations in the degree of magnetic anisotropy can be used to determine if there are variations in the mineral alignment produced by burial and compaction, and possibly by initial (but visually cryptic) layer-parallel shortening (Housen and Kanamatsu, 2003; Ujiie et al., 2003). Post-depositional compaction reduces porosity and may rotate detrital grains toward horizontal, producing oblate fabrics with vertical minimum axes and
horizontal maximum axes in the sediments. These fabrics can be detected through AMS and scanning electron microscopy (Arason and Levi, 1990a; Tarling and Hrouda, 1993; Kawamura and Ogawa, 2004; Li et al., 2004).

Magnetic fabrics are useful for examining compaction in clay-rich sediments (Arason and Levi, 1990a). Compaction due to loading can lead to changes in the state of consolidation and the clay microstructure (Kawamura and Ogawa, 2004). In some cases, magnetic particles (such as small grains of magnetite that may also carry, in part, the remanence of the sediments) can attach to clay particles via electrostatic or Van der Waals forces early in the depositional process. These magnetic particles eventually become incorporated into the clay fabric as the clay particles attach to one another (Arason and Levi, 1990a; Sun and Kodama, 1992). As compaction increases, porosity decreases, eliminating void space in which magnetic particles could be randomly oriented and clay flakes rotate from edge-face contacts to face-face contacts (Fig. 14B; Kawamura and Ogawa, 2004). The particles become oriented perpendicular to the direction of compaction leading to further alignment of grains and the production of AMS fabrics in sediments.

As compaction occurs, anisotropy continues to become more pronounced and the $K_3$ axis rotates towards vertical (Tauxe, 1998). The process of compaction and consolidation begins within a few centimeters below the seafloor and typically results in a shaly structure, wherein all clay flakes have rotated to have face-face contacts, in clay-rich sediments below 150 m (Kawamura and Ogawa, 2004). The process of compaction is ongoing and should progress with depth. By examining the degree of compaction within clay-rich sediments, it should be possible to identify areas where sharp increases or decreases in the level of
compaction occur. These unexpected changes in the compaction level are collectively referred to as compaction disequilibria.

Sediment consolidation is typically accompanied by changes in porosity and permeability (Hamilton, 1959; Kawamura and Ogawa, 2004). Laboratory-run consolidation tests, examining the relationship between the void ratio and the total pressure applied to the sediments, have been used to create curves depicting the expected settlement behavior for varying sediment types (Hamilton, 1959). These curves will be referred to as settling curves for the remainder of this study. As the sediments are consolidated, the void ratio and porosity decrease. As the state of consolidation increases, porosity and permeability are gradually reduced. This pattern of gradually reduced porosity with increased compaction is commonly referred to as ‘normal’ consolidation (Kawamura and Ogawa, 2004). Normal consolidation can also occur in naturally deposited sediments. Nobes et al. (1986) documented consistent porosity decreases with depth in ocean sediments collected globally during Deep Sea Drilling Program Legs 1-86. Based on these findings, porosity is expected to decrease with depth at the IODP Exp 333 sites.

AMS can be used to detect unconformities in the form of apparent overconsolidation of sediments. Apparent overconsolidation is a term used to define a zone of sediments which are more compact than predicted based on experimental settling curves (Hamilton, 1959; Schwehr et al., 2006). Apparent overconsolidation is expected to be accompanied by an abrupt increase in the degree of anisotropy. The increase in the degree of anisotropy can be recognized by an increase in separation of the maximum and minimum eigenvalues (Fig. 14A), \( \tau_1 \) and \( \tau_3 \) respectively, or by an increase in the AMS P value \( (K_1/K_3) \), both of which are measures of anisotropy (Schwehr et al., 2006). Unconformities, especially those associated
with the removal of sediment, should be associated with an area of increased anisotropy due to the contrast in the amount of sediment compaction above and below the unconformity. Zones of apparent overconsolidation are thought to be the result of erosional unconformities, landslide events, or earthquake-induced dewatering (Lee et al., 2004; Schwehr et al., 2006). Detection of unconformities may therefore aid in the recognition of erosional events and earthquake occurrences at a specific location.

On the other hand, underconsolidated zones are created when overlying layers of low permeability prevent or retard the dewatering of the sediments. A decrease in the degree of anisotropy recognized by a decrease in $P (K_1/K_3)$ or a decrease in the separation of $\tau_1$ and $\tau_3$ reflects underconsolidation (Fig. 14B). Underconsolidated zones commonly represent zones with rapidly deposited sediments, because newly deposited clay-rich sediments can confine and trap water in the underlying strata (Fig. 14B; Schwehr et al., 2006; Kawamura and Ogawa, 2004). Zones of underconsolidation also have been found in accretionary prism fault zones where high-pressure fluids play a role in decreasing mechanical coupling (e.g., Moore et al., 1995; Housen et al., 1996; Maltman et al., 1997).

Determining how sediments were originally deposited is an important part of understanding the processes active in a basin. As previously mentioned, in quiet water deposition, the $K_3$ axis of an anisotropic grain is deposited with sub-vertical orientation (Fig. 15A). But when a current is present at the time of deposition, grains are commonly aligned and the magnetic foliation plane is often imbricated relative to the current direction (Fig. 15B; Taira, 1989; Tauxe, 1998). The azimuth of the offset of the $K_3$ direction from vertical indicates the imbrication direction and can be plotted on lower hemisphere equal area (EA) diagrams for relatively simple analysis (Fig. 15B; Taira, 1989). The relationship between
current and imbrication azimuth has been used in determining both paleocurrent direction and depositional mechanisms (Taira, 1989; Pares et al., 2007).

AMS measurements can also be used to distinguish depositional mechanisms such as gravity, current, grain collision, and viscous suspension in q-β diagrams (Taira, 1989). In these diagrams, magnetic lineation/foliation ratio (q) is plotted against the imbrication of the magnetic foliation plane relative to a horizontal bedding surface (β) (Fig. 16; Taira, 1989). The plot is divided into sections representing one of the above-mentioned depositional mechanisms, which were investigated by Taira (1989) using a combination of laboratory and natural sediment transport examples. The depositional mechanisms are then assigned based on position and scatter of the data within the plot.

All low field susceptibility measurements used for this study were conducted at the Pacific Northwest Paleomagnetism Laboratory. Low field susceptibility AMS measurements were done on a Kappabridge KLY-3 Spinning Specimen Magnetic Susceptibility Anisotropy Meter. This instrument subjects a specimen to an applied magnetic field and measures the resulting induced magnetization in multiple orientations. These measurements are used to characterize the magnetic fabric in the sediments. Susceptibility data generated from the AMS measurements include, eigenvectors (K₁, K₂, K₃), eigenvalues (τ₁, τ₂, τ₃), lineation (L=K₁/K₂) values, foliation (F=K₂/K₃) values, and the degree of anisotropy (P=K₁/K₃). Tables 1A-3A list the AMS parameters (K₁, K₂, K₃, F, L, and P) for each sample from each site.
AMS Rotation

Cores collected during IODP expeditions are typically recovered in 9.5 m lengths, which are not oriented relative to one another. Care is taken to maintain a common orientation for all sections within an individual core. Therefore, samples taken from one core should have a common orientation. In order to characterize the orientation direction of the AMS ellipsoid accurately on the EA plots, reorientation is necessary.

Reorientation is accomplished using the geomagnetic field recorded in the sediments. The mean declination and inclination of the samples obtained from analyses of step-wise demagnetization experiments serve as an estimate of the GAD field at the time of deposition (e.g., Owens, 1993; Housen et al., 1996; Pares et al., 2007). Reorientation methods varied depending on the type of coring system used.

For piston-cored and punch-cored sediments, the rotation value was determined by the average declination of all samples taken from one core. In order to obtain an accurate average declination for each core, all samples from a given core were rotated so that they were oriented into the same hemisphere. This reorientation process involved rotating some samples 180° and reversing the polarity of the inclination to ensure all samples from a core were all normal or all reverse. Reorienting the samples reduced the error caused by polarity flips within a core. In this manner, all piston and punch cored sediments were reoriented to geographic north.

In the case of shoe coring, in which sediments are subjected to rotation during the coring process (i.e., Table 1 for coring system details), cores and sections are less well preserved; and the 9.5 m cores are typically broken in ‘biscuits’ (Fig. 17C). The individual samples taken from the ‘biscuits’ do not individually represent a long enough period of time
to average PSV. The orientation of these samples would have larger uncertainty than the samples taken from cohesive cores. For the cores with ‘biscuits’, samples were rotated individually to geographic North based on their declination. After the completion of individual sample rotation, the samples were regrouped by core for analysis. The assumption of orientation continuity within a core can then be applied to all cores for all eigenvector analyses.

The magnetic directions recorded in individual samples must also be evaluated prior to being used for average declination calculations. Directions near polarity boundaries may reflect the transitional state of the geomagnetic field; therefore samples found to be within roughly one meter of a polarity change were excluded. Samples taken from the core catcher have a higher likelihood of being misoriented prior to sampling. Core catcher samples, which fell as outliers, were rotated to the average declination of the samples from the core. By evaluating and excluding samples with spurious directions, the calculated average declinations should be accurate and the reoriented eigenvectors should accurately represent the AMS ellipsoid from each sample.

**Bedding Correction**

When sediments are tilted after deposition, it is necessary to correct the magnetic directions back to horizontal. At Sites C0011 and C0018, bedding was sub-horizontal and the sediments did not require bedding corrections. Bedding corrections were only necessary at Site C0012. The magnetic inclination directions for each sample were oriented to horizontal using the strike and dip measurements taken during Exp 333. A sample was only reoriented if a strike and dip measurement was available for the same section of core from
which the sample was collected. The bedding corrections were also applied to the AMS data to ensure all values were reoriented to horizontal.

**Rock Magnetism**

High field measurements, magnetic hysteresis, isothermal remanent magnetization (IRM), and direct current demagnetization (DCD), were measured with a MicroMag 3900 Vibrating Sample Magnetometer (VSM). These measurements are used to characterize the magnetic mineralogies and magnetic domain states found in the individual samples. A representative group of samples from each site was selected to undergo these high field susceptibility measurements (Tables 4 A-6A). Sediment was collected from the plastic cubes and packed into gel caps for magnetic hysteresis measurements after alternating field demagnetization was complete. Magnetic hysteresis was measured under an applied field of 6000 kOe that varied at a 200 Oe increment with an averaging time of three seconds. These settings were kept constant for all samples.

IRM is the remanent magnetization acquired at one temperature, typically room temperature, over a short period of time in an external applied field (McElhinny and McFadden, 2000). The acquisition of an IRM is used to determine the magnetic properties (magnetic particle size and magnetic mineralogy) of samples. IRM was measured on a logarithmic scale from an initial field of 10 Oe to a final field of 9000 Oe with 30 measurement points and an averaging time of four seconds.

**Porosity and Pore Water Volume**

Combining an AMS study with water content data may give additional information regarding the compaction history of a sedimentary sequence. The amount of water within the sediments is related to the amount of compaction the sediments have undergone, which
affects the interpretation of fabric shape and inclination with depth. As compaction increases and pore space is reduced, the pore water volume is expected to decrease (Hamilton, 1959; Kawamura and Ogawa, 2004). Compaction should expel water from the sediments and reduce the water content. When the water content does not decrease with depth as expected, compaction may not increase with depth and further investigation is necessary.

Porosity can often be used as a proxy for the degree of compaction within sediments. Increased compaction reduces the porosity of the sediments. During Exp 333, the physical properties team collected porosity data on each core. These measurements will serve as the measure of compaction for the sediments from each site (Fig. 18D, 19D, 20D; Arason and Levi, 1990a; Kawamura and Ogawa, 2004). Inorganic chemists from Exp 333 collected pore water values by squeezing sections of core and collecting the expelled water (Henry et al., 2012). These measurements are discussed in more detail in the discussion and interpretation section.

RESULTS

Paleomagnetism and Magnetostratigraphy

NRM, the remanent magnetization acquired under natural conditions, was measured for each discrete sample. Following the NRM measurements, samples were step-wise demagnetized. The demagnetization steps of 2.5, 5, 7.5, 10, 15, 20, 25, 30, 35, 40, 50, 60, and sometimes 90 mT, provided a well resolved data set for determining a paleomagnetic vector for most samples from Sites C0011 and C0012. Samples from Site C0018 underwent the same early demagnetization steps as mentioned above, but higher steps were necessary (up to roughly 140 mT). Many samples from Site C0018 also required the use of the AGRM
technique, beginning between 60 and 80 mT, to avoid the acquisition of a GRM and to resolve a ChRM.

After the alternating field demagnetization of all of the sediments was complete, best-fit lines were selected for each orthogonal projection. When a ChRM was not resolvable from the orthogonal projection, the data for the specimen were thrown out. In many cases, a steeply inclined vector is resolved prior to the ChRM. This sub-vertical vector, the first removed component, is likely a magnetic overprint imparted to the sediments by the weak magnetic field generated by the rotating drill-string and is referred to here as the drill-string overprint. This overprint is typically removed between 15 and 25 mT and does not appear to affect the ChRM vector at any of the sites.

The orthogonal vector diagrams for Site C0018 are broken into three types (Fig. 9). Type one is characterized by two components, with the second component progressing toward the origin and MAD values less than 15° (Fig. 9A). The second type includes all samples that underwent AGRM with a MAD value of less than 15°. Type two is also typically characterized by two components, the second trending toward the origin (Fig. 9B). Type three includes all samples that did not progress toward the origin, and therefore were not useful for defining a ChRM direction, or had a MAD value above 15° (Fig. 9C). First and second removed component directions for each sample from Site C0018 are listed in Tables 7A and 8A, respectively.

The classification of the type two samples required the recognition of GRM acquisition during the demagnetization process. Type two samples were distinguished by an increase in magnetic moment and a progression away from the origin above demagnetization steps of 60 mT (Fig. 9B). These same samples had scattered declinations and inclinations
with increasingly strong demagnetization steps, indicating that the samples were acquiring a GRM. These samples were therefore chosen to undergo AGRM in the hopes of defining a ChRM.

Orthogonal vector projections for samples from Site C0011 typically have two components (Fig. 10). The first component is generally very steep (average of 86.8° for Hole C0011D) and removed between 10 and 25 mT. The ChRM progresses toward the origin and is less steep (average of 72.3°) than the first removed component. MAD values were less than seven degrees for 89% of the samples at Site C0011 and both components are well defined. Tables 9A and 10A contain the first and second removed component directions for each sample from Site C0011.

Site C0012 was similar in demagnetization characteristics to Site C0011. Most samples consisted of a two-component orthogonal projection, the first component being steep (C0012C average: 88.8°; C0012D average: 85.4°), almost vertical, and the second component becoming shallower (C0012C average: 34.5°; C0012D average: 47.3°) and progressing to the origin with progressive demagnetization (Fig. 11). The first component is removed between 15 and 40 mT. MAD values for the second removed component were less than seven degrees for 82% of the samples. The remaining samples had only one component or a ChRM was not determined. The first removed components for the Site C0012 samples are listed in Table 11A and the second removed components are listed in Table 12A.

The data obtained from the orthogonal vector diagrams were used to create detailed magnetostratigraphies for each site. Portions of Site C0018 could not be used to assign ages to the sediments due to disturbance (MTDs), which likely altered the original magnetization direction recorded in the sediments (Fig. 21; Rosenbaum et al., 2000). Magnetostratigraphies
for Site C0011 and Site C0012 are illustrated in Figures 22 and 23, respectively. Inclination values were evaluated at each site in effort to identify and classify the changes in inclination with depth.

**Sediment Accumulation Rates**

To further illustrate the differences in sedimentation at Sites C0011 and C0012, a comparison of accumulation rates was completed. After assigning ages to samples at specific chron and subchron boundaries, diagrams of age versus depth were created (Fig. 24). Strata of ash with known ages have been included and provide further age constraints. These well dated ash layers include the Azuki (0.85 Ma), the Pink (1.05 Ma), the Habutaki (2.85 Ma), and the Ohta (4.0 Ma) ashes (Hayashida et al., 1996; Satoguchi et al., 2005; Nagahashi and Satoguchi, 2007; Henry et al., 2012). Biostratigraphy ages are also included in Figure 24. The changes in accumulation rates are well documented at both Sites C0011 and C0012.

**Inclination versus Depth**

Inclination values for each site were plotted against depth to determine if compaction causes significant inclination shallowing (Fig. 18C, 19C, 20C). Expected inclination at all three sites is ~52° (or -52°) assuming that no latitudinal migration of the basin has occurred during the time of deposition (Henry et al., 2012). Bedding at Site C0012 varied in strike and in degree of dip. Bedding corrections were applied to samples from Site C0012 to account for these variances.

The trends in inclination with depth can be separated into two sections at Site C0018 (Fig. 18C). In the first 127 m of sediment, inclination is near the expected 52°, with a mean inclination of 52.2° and a standard deviation of 11.9°. Below 127 mbsf magnetic inclination
varies from less than 10° to nearly 90°. The mean inclination for the sediments in this section is 42.6° with a standard deviation of 24.5°. Neither region is characterized by steepening or shallowing inclination.

At Site C0011, there are three major divisions in the trend in inclination with depth (Fig. 19C). The inclination from 0 to 25 mbsf decreases from an average inclination of 51° (calculated from sediments between 0 and 25 m) to an average of 44° (average calculated from sediments between 25 and 50 m). Inclination begins to steepen gradually from 25 mbsf to 100 mbsf, reaching a peak with an average inclination of 66.3° between 100 mbsf to 125 mbsf. Following this peak the inclination values are relatively constant, but a transition occurs around 200 mbsf (Fig. 19C). Average inclination in the bottom-most sediments (358 m to 378 m) is roughly 51.5°.

Five distinct regions are identified in the Site C0012 inclination versus depth data after corrections for bedding dips have been applied (Fig. 20C). The first region is from zero to 18 mbsf. Inclination values in this section of the core are scattered from 12° to 80° with an average inclination of 46.5° and a standard deviation of 22°. The second region (19-50 mbsf) is characterized by a fairly constant inclination angle, with a mean inclination of 26° and a standard deviation of approximately 15°. Inclination steepens in the third section which extends from 50 to 82 mbsf. Average inclination in this region is 49° with a standard deviation of 13°. The fourth region, 85 to 109 mbsf, begins with a shallowing of the magnetic inclination (average inclination of 18°; standard deviation of 15°). At approximately 112 mbsf, inclination begins to steadily increase again, reaching an average inclination of approximately 59° and a standard deviation of 12° between 134 and 137 mbsf. The remaining cores (137-178 mbsf) did not require bedding correction and have also been
included in this analysis. Finally, the inclination values between 137 mbsf and 178 mbsf average to approximately 53° with a standard deviation of 15°.

**Anisotropy of Magnetic Susceptibility (AMS)**

**Eigenvalues and Eigenvectors**

Eigenvalues ($\tau_{1,2,3}$) and eigenvectors ($K_{1,2,3}$) are used to represent the magnitude and orientation, respectively, of the AMS ellipsoid. The eigenvalues are normalized and sum to unity with $\tau_1$ representing the largest eigenvalue and $\tau_3$ representing the smallest (Tables 13A-15A). The general increase of anisotropy expected with depth is measured by the difference between $\tau_1$ and $\tau_3$ (Schwehr et al., 2006). This difference can be recognized by the gap between $\tau_1$ and $\tau_3$ in Figures 7B, 8B, and 9B. The larger the gap, the more anisotropic the fabric shape. The directions of the eigenvectors can be used to examine fabric shape and identify sediment disturbance. $K_1$ represents the maximum eigenvector and $K_3$ represents the minimum eigenvector.

The eigenvalues at Site C0018 can be split into two sections (Fig. 18B). In section one $\tau_1$ and $\tau_3$ do not indicate an increase in anisotropy. The anisotropy of the sediments is relatively constant to approximately 225 mbsf. The difference between $\tau_1$ and $\tau_3$ varies greatly between samples, but the overall trend is consistent. In section two, anisotropy is reduced, but appears to increase near the bottom of the hole.

At Site C0011, visual inspection of the eigenvalue data revealed four separate sections (Fig. 19B). An increase in the difference between $\tau_1$ and $\tau_3$ is noted in section one from 0 to 35 mbsf. In section two, 35 to 100 mbsf, anisotropy decreases. The third section is characterized by a relatively constant anisotropy. Section three extends from 100 to 250...
mbsf. Below 250 mbsf the minimum and maximum eigenvalues indicate increasing anisotropy.

The eigenvalue data from Site C0012 has been separated into two sections based on trends in \( \tau_1 \) and \( \tau_3 \) (Fig. 20B). In one the difference in \( \tau_1 \) and \( \tau_3 \) is relatively constant. These relatively constant eigenvalues are recognized from 0 to 80 mbsf. Below 80 mbsf the difference between \( \tau_1 \) and \( \tau_3 \) begins to increase. The increase in anisotropy below 80 mbsf continues until the base of the sedimentary section at 180 mbsf.

**Degree of Anisotropy**

In order to identify regions of sediment disturbance; the degree of anisotropy (P), \( K_1/K_3 \), was plotted versus depth (Fig. 18A, 19A, 20A). P is expected to increase with depth gradually as compaction occurs in the clay-rich sediment (Kawamura and Ogawa, 2004). A one to one ratio of \( K_1 \) to \( K_3 \) (P=1) indicates an isotropic fabric. As \( K_1 \) and \( K_3 \) become more distinct, the P values increase. A larger P value indicates a more anisotropic fabric within the sediments.

The degree of anisotropy varies throughout Site C0018 (Fig. 18A). Three distinct regions have been identified. Section one extends from 0 to approximately 230 mbsf. The degree of anisotropy in section one is relatively constant. Section one transitions into section two around 230 mbsf where P begins to decrease. The decrease in P continues to a depth of 250 mbsf. The third section is marked by an increase in P. The increase in P continues from 250 mbsf to the bottom of the hole at Site C0018. The boundaries that mark the beginning of section two and section three (Fig. 18A) may be related to the increase in sand and the frequency of silt turbidites (Fig. 4).
Trends in P are more easily identified at Site C0011 (Fig. 19A). The core can be split into four separate sections. Section one, 0 to approximately 35 mbsf, is characterized by increasing P values. From 35 to 100 mbsf there is a decrease in P. This decreasing section is section two. The third section extends from 100 to 250 mbsf and is marked by relatively constant P. At 250 mbsf P begins to increase. The gradual increase in P continues to the bottom of Site C0011. The lithologic unit change and the presence of sand, sandstone, and conglomerate found from approximately 345 to 380 mbsf may explain the variance in the degree of anisotropy between samples in these sediments (Fig. 5).

At Site C0012, the degree of anisotropy with depth can be separated into two distinct regions (Fig. 20A). Section one is defined by relatively constant P from 0 to 80 mbsf. P begins to increase gradually around 80 mbsf and continues to increase to the bottom of Hole C0012D. This section of increasing P is section two. Samples between 135 and 180 mbsf have P values ranging from 1.01 to 1.1, a much larger variance between samples than in the shallower sediments. This may be due to the presence of abundant ash layers in the sediments from 125 to 150 mbsf and the transition from lithologic unit one to lithologic unit two at 150 mbsf (Fig. 6).

When evaluating anisotropy, bioturbation must be taken into consideration due to the likelihood of disturbance in the sediments. Significant bioturbation was noted during Exp 333 at Sites C0011 and C0012 (Henry et al., 2012). At Site C0011, subunit 1B, which corresponds to a depth interval from 251 to 348 mbsf, is heavily bioturbated. Unit 1 from Site C0012 varies from moderately bioturbated at the surface to heavily bioturbated at the bottom of the unit (0-151 mbsf). Although significant peaks can be identified in the P values
for Site C0011 around 275 mbsf and for Site C0012 at roughly 90 mbsf and 138 mbsf, the bioturbation has not caused changes in anisotropy values that are readily noted.

**Fabric Shape Classification**

Flinn diagrams of the degree of lineation ($K_1/K_2$) and foliation ($K_2/K_3$), generated from the AMS measurements, can also be used to analyze fabrics in sedimentary environments (Fig. 25-27; Tarling and Hrouda, 1993). Sedimentary fabrics are expected to become more oblate with depth due to compaction. The effects of compaction reach a critical point if sediments are consolidated with a large degree of grain-to-grain contact. At this point, foliation within the sediments cannot increase any further because the grain contacts are all face-face and the sedimentary fabric is classified as oblate (Tarling and Hrouda, 1993; Kawamura and Ogawa, 2004). Flinn diagrams were used to classify the dominant fabric shape within a core, but there are other data sets that can also help distinguish fabric shape.

Oblate ellipsoids consist of flat spheroid shaped grains (pancake shaped) with AMS axes $K_1 \approx K_2 > K_3$. In clay-rich sediments deposited on a horizontal surface, $K_3$ is expected to be sub-vertical (Tarling and Hrouda, 1993; Tauxe, 1998). $K_1$ and $K_2$ are girdled on the EA plot while $K_3$ is well defined. The 95% confidence bounds for the $\tau_1$ and $\tau_2$ distributions will overlap for oblate AMS ellipsoids, while the $\tau_3$ distribution will be distinct (Fig. 28A; Tauxe, 1998). Finally, oblate ellipsoids are characterized by a strong foliation ($K_2/K_3$) relative to the lineation ($K_1/K_2$).

Prolate AMS ellipsoids consist of elongate (football shaped) grains with $K_1 > K_2 \approx K_3$. The 95% confidence bounds for the $\tau_1$ distribution is distinct and well defined while the distributions of $\tau_2$ and $\tau_3$ overlap (Fig. 28B). The $K_2$ and $K_3$ eigenvectors are typically
distributed in a girdle for a prolate ellipsoid, while the $K_1$ eigenvector is well defined. Prolate AMS ellipsoids are characterized by a strong lineation relative to the foliation.

The third ellipsoid shape is ‘spherical’. The ratio of foliation to lineation for spherical ellipsoids is near equal. The eigenvectors are generally poorly defined and plot in a cloud on EA plots. There is no statistical difference in the distributions of the eigenvalues when the ellipsoid fabric is spherical (Fig. 28C). Ellipsoids may be misinterpreted as spherical if the sediments have been disturbed. There is no clear method for distinguishing between disturbed sediments and a spherical distribution.

The combination of Flinn diagrams, EA diagrams, and the distribution of eigenvectors can be useful in determining the shape of the magnetic fabric in sediments (Tauxe, 1998; Tarling and Hrouda, 1993). Figures 29, 30, and 31 include EA diagrams with eigenvectors, Flinn diagrams, and histograms of eigenvalues generated from a parametric bootstrap with the 95% confidence interval for each $\tau$ also shown. The parametric bootstrap is best suited for a homogeneous sample set (Tauxe, 1998). Based on the assumption that one core represents a section of sediment with a common orientation, the data were grouped by core when plotted on the Flinn and EA diagrams and the eigenvector distribution histograms.

**Depositional Mechanisms**

Depositional mechanisms for each site were determined from $q$-$\beta$ diagrams (Fig. 16; Taira, 1989). Figures 32A, B, and C illustrate how depositional mechanisms varied by site. Average $q$, the degree of lineation to foliation, and average $\beta$, the imbrication angle of the grains from horizontal, for each core were used. The $q$ and $\beta$ values for each sample from each site are listed in Tables 16A-18A. Cores with $K_3$ inclination below 55° have very steep
imbrication angles (greater than 35°) using the Taira (1989) method (Fig. 32). The Taira (1989) study focused on modern fluvial, eolian dune, and beach sand sediments. It is possible that the imbrication angles over 35° are due to a sedimentary process that was not replicated in the Taira (1989) study. Further investigation is necessary in order to understand the cause of the steep imbrication angles.

The q-ß diagram for Site C0018 is shown in Figure 32A. The dominant mechanisms of deposition at Site C0018 are current and grain collision. Slope gravity deposition, gravity controlled deposition of sediment from the water column onto a slope, was dominant in cores 27, 28, and 29. Ten cores fall outside the defined regions and have high standard deviation (greater than 20°) of $K_3$. These samples are from cores within MTD six and are disturbed. Average $q$ and $ß$ are not applicable for these cores.

The dominant depositional mechanism at Site C0011 is current (Fig. 32B). Much of the data that can be classified as slope gravity deposition can also be classified as current deposition due to the overlap of the defining fields. A subset of the cores (11 of 52) lie in the grain collision mechanism region, but grain collision does not seem to be the dominant depositional mechanism at Site C0011.

At Site C0012, a large portion of cores fall outside of the regions defined by Taira (1989; Fig. 32C). Nine cores out of a total of twenty-one cores have imbrication angles above 35°. Bioturbation is documented in five of the cores with steep imbrication angles. Of the remaining 14 cores, seven fall within the slope gravity and grain collision interaction zone. Grain collision interaction with slope gravity is the dominant depositional mechanism at Site C0012. The remaining seven cores are distributed between slope gravity, current, grain collision, and viscous suspension.
**Sediment Transport Directions**

Sediment transport directions can be defined for each site using AMS K_1 and K_3 directions (Ellwood and Ledbetter, 1977; Ellwood and Ledbetter, 1979; Liu et al., 2001; Pares et al., 2007). In the presence of a current, grains can become imbricated and the K_3 axis can be offset in the direction of current flow (Fig. 15B). The grain shape influences how the grains react in the presence of a current. Paleocurrent direction is determined using the primary AMS fabric in the sediments (Fig. 33). Primary fabrics in this study were classified as having oblate fabrics and average K_3 imbrication angles for each core within 25° of vertical (Hamilton and Rees, 1970; Hrouda, 1982). The AMS data have been rotated to geographic north in order to obtain paleocurrent flow directions in geographic coordinates. Rose diagrams of rotationally corrected K_3 shown in Figures 34-36 depict the azimuthal distribution of K_3 axes. The data in each rose diagram are grouped by lithologic unit. Sediment transport is assumed to be in the direction of the dominant K_3 azimuth.

Seventeen of the 34 cores collected at Site C0018 meet the primary fabric criteria discussed above (Cores 4-9H, 24T, 27T-36X). The sediment transport direction in Unit Ia (Cores 4-9H; 25-80 mbsf; 0.3-0.45 Ma; Fig. 4) is generally to the SE (Fig. 34A). Unit Ib is separated into two sections: piston and punch cores (Core 24H and Cores 27-30H) and shoe cores (Core 31-36X). This separation was done to reduce the effect of error introduced during the rotation of individual shoe-cored samples. The sediment transport direction is not well defined in the Unit Ib piston and punch cores (Fig. 34B), but sediment transport appears to be to the SE in the Unit Ib shoe cores (250-310 mbsf; ~1.3-1.67 Ma; Fig. 34C). The dominant transport of sediment to the SE at Site C0018 indicates downslope movement of the accretionary sediments towards the Nankai Trough.
The azimuthal directions of $K_3$ axes can also be used to characterize the sediment transport behavior within MTD six at Site C0018. The cores taken from MTD six (Cores 15-23) do not meet the primary fabric criteria, indicating that the sediments were likely disturbed during the deposition of the MTD. The original sedimentary fabric is not retained, but the disturbed sediments within the MTD record the direction of sediment transport. The sediment transport direction within MTD six is predominantly downslope to the SE (Fig. 34D).

Twenty-nine of the 53 cores collected at Site C0011 meet the primary fabric criteria (Cores C0011C-3H, C0011D-4H, 5H, 11H-15H, 17H, 21H, 25T-28X, 32X-45X, 49X, and 52X). The sediment transport directions in the remaining cores can be broken into four categories based on the lithologic unit boundaries (Fig. 5). The paleocurrent direction is not well defined below 347 mbsf (>7.12 Ma), which is Unit II (Cores C0011D-49X and 52X; Fig. 35A). The sediment transport direction in Subunit Ib from 250-350 mbsf (Cores C0011D 35X-45X; 7.12-5.24 Ma) is to the SW roughly parallel with the Nankai Trough (Fig. 35B). Due to the transition from piston and punch cores to shoe cores in Subunit Ia, the subunit was broken into two sections. The dominant current direction in the shoe cores of Subunit Ia (Cores C0011D-26-28X and 32-34X; 5.24-4.3 Ma) is to the SE (Fig. 35C). The sediment transport direction in the piston and punch cores of Subunit Ia (Cores C0011C-3H, C0011D-4-5H, 11-15H, 17H, 21H, and 25H) represents the sediment transport direction from 4.3 Ma to present (0 to approximately 200 mbsf). The direction recorded in these sediments is NW-SE (Fig. 35D). The SE and NW sediment transport directions recorded in Subunit Ia are generally perpendicular to the Nankai Trough and parallel the direction of convergence between the Philippine Sea Plate and Japan (Fig. 1).
At Site C0012, sediment transport direction analysis was completed on cores which met the primary fabric criteria. In some cases, this limited the analysis to cores that had undergone bedding correction, rotation correction, or a combination of both corrections. The most recent sediments at Site C0012 are part of Subunit Ia (0-71 mbsf; 0-4.4 Ma; Fig. 6). These sediments are characterized by random axes orientations and do not meet the primary fabric data. No sediment transport direction was determined for Subunit Ia. Subunit Ib, Cores C0012C-9H through C0012D-7H (13 cores total), met the primary fabric criteria. There is no distinct paleocurrent direction in Cores C0012C-9H to 15H and C0012D-2H (71-124 mbsf; 4.4-7.14 Ma; Fig. 36A). In Subunit Ic (124-150 mbsf; 7.14-7.5 Ma), the sediment transport direction is ESE (Fig. 36B). The sediment transport direction recorded in Unit II, Cores C0012D-6H and 7H (150-160 mbsf; 7.5-7.6 Ma), is SE (Fig. 36C). The ESE and SE sediment transport directions in Subunit Ic and Unit II indicate sediment movement toward the Kashinosaki Knoll and away from the Nankai Trough.

**Hysteresis and Day Plots**

In order to evaluate grain size and composition of the magnetic particles within the sediments, magnetic hysteresis, IRM, and DCD were measured on select samples from each site. The ratio of saturation remanent magnetization (Mrs) to saturation magnetization (Ms) versus the ratio of coercivity of remanence (Hcr) to the coercivity (Hc) is often plotted following Day (1977) in an effort to characterize the domain state and grain size of the magnetic grains (Tables 4 A-6A). Identifying the types of magnetic particles within the sediments is necessary to ensure that the demagnetization techniques are appropriate; certain magnetic minerals behave differently under particular demagnetization techniques. By
noting the magnetic minerals in a sample it is possible to explain the differences in demagnetization behavior of the samples.

The Day plot for Site C0018 (Fig. 37A) indicates that the sediments have a magnetic mineralogy dominated by a superparamagnetic (SP) component, which may be due to sulfides in the sediment. Greigite, Fe₃S₄, is an authigenic ferrimagnetic mineral often found in muds and mudstones. Greigite forms as a precursor to pyrite in anoxic sulfide-reducing sedimentary environments, but it can also grow during digenesis if dissolved iron and sulfide are available (Roberts, 1995; Roberts et al., 2011). Greigite falls along a single domain-SP mixing line on the Day plot; this is also where the sediments for Site C0018 fall (Fig. 37A; Roberts et al., 2011). Roberts et al. (2011) argue that magnetic hysteresis loops for greigite samples will become progressively more “potbellied” with an increase in SP grains. Potbelly hysteresis loops are characterized by a widening of the loop around the origin. The uncorrected magnetic hysteresis loop for sample C0018-34X-9-22cm, shown in Figure 38, is one example of the potbellied hysteresis loops from Site C0018. Three of the ten samples measured displayed potbellied behavior. During AFD, greigite often acquires a GRM in fields above 40-60 mT. The acquisition of a GRM is accompanied by an increase in remanence (Snowball, 1997; Roberts et al., 2011). A review of the type two orthogonal vector diagrams and the corresponding relative intensity plots (J-plots) from Site C0018 reveal increased remanence above 60 mT (Fig. 9B). J-plots are a graphical representation of the relative magnetic intensity of a sample. In an ideal situation, the plot will have a characteristic reverse J shape with the relative intensity of the sample decreasing with each progressive demagnetization step. The relative intensity does not decrease as expected in many of the Site C0018 samples. The demagnetization behavior of the sediments from Site
C0018 is likely due to the presence of greigite, which appears to be the dominant magnetic mineralogy at Site C0018.

The dominant magnetic mineralogy at Sites C0011 and C0012 is single domain and multi-domain magnetite. Characteristic hysteresis loops for Sites C0011 and C0012 are shown in Figures 39 and 40 respectively. All of the measured samples are subparallel to the single domain-multi-domain mixing curve, as shown in Figure 37B and C, likely caused by a large coercivity contrast in the samples (Roberts et al., 1995). The larger the coercivity contrast in a sample, the higher the Hcr/Hc ratio will be and a high ratio leads to a systematic offset of the data to the right on a Day plot (Roberts et al., 1995). The mixtures of coercivities present in Sites C0011 and C0012 may explain why all of the samples fall to the right of the single domain-multi-domain mixing line. Samples from Site C0011 and C0012 contain similar mixtures of single domain and multi-domain magnetite, but one sample from each site was found to exhibit predominantly multi-domain behavior (Fig. 37D).

The samples that exhibit the strong multi-domain behavior are C0012D-13H-3-40 cm and C0011D-51X-4-71 cm. The C0012D-13H-3-40 cm sample was specifically taken to test a portion of red-stained sediment, which upon analysis is interpreted to be multi-domain magnetite that has oxidized to either hematite or maghemite. Sample C0011D-51X-4-71 cm was taken from a sandy lithology which may lead to the presence of larger magnetite grains in the sediment. Magnetite grains larger than 10 micrometers in diameter are typically classified as multi-domain (Butler, 1992). The magnetite in C0011D-51X-4-71 cm is interpreted to be multi-domain.
DISCUSSION AND INTERPRETATION

Magnetostratigraphy

Analyses of the polarity zones from both the onshore and shipboard data at Site C0018 revealed two polarity chrons. The boundaries between the normal and reverse zones do not correlate well between the shipboard and shorebased data. Samples were collected from the same cores; the same sedimentary horizons were not sampled. Short reversals may have been sampled for shipboard analysis, but due to the sampling interval may not have been sampled for the shore-based study or vice versa. The normal polarity chron from 0 to approximately 125 mbsf coincides well with the Brunhes Chron from Gradstein et al. (2004). The reverse polarity zone that begins around 125 mbsf and extends to the bottom of the hole is interpreted as the early Matuyama Chron. Based on biostratigraphy ages, the oldest sediments at Site C0018 are about 1.67 Ma (Gradstein et al., 2004; Henry et al., 2012). The GPTS, shipboard magnetostratigraphy, shorebased magnetostratigraphy, and the biostratigraphy were difficult to reconcile below 190 mbsf due to sediment disturbance and gaps in sediment recovery (Fig. 21).

Magnetostratigraphy was not completed within MTD six, because the polarity zones are poorly defined due to sediment disturbance, but tephrochronology of the Azuki and Pink ashes (0.85 Ma and 1.05 Ma respectively) provides an age bracket for the MTD (Hayashida et al., 1996; Henry et al., 2012). The Azuki ash (0.85 Ma) was identified approximately three meters above the upper boundary of MTD six; the base of MTD six lies directly above the Pink ash (1.05 Ma). MTD six therefore occurred between 0.85 Ma and 1.05 Ma. The correlation of magnetostratigraphy and tephrochronology was essential for dating the sediments at Site C0018.
At Site C0011, ten polarity chron and subchrons were identified in the shorebased data as well as in the shipboard data (Fig. 22). Four ash layers, the Azuki (0.85 Ma), the Pink (1.05 Ma), the Habutaki (2.85 Ma), and the Ohta (4.0 Ma), were used to help constrain the sediment ages and correlate the ship and shorebased magnetostratigraphies with the GPTS from Gradstein et al. (2004; Hayashida et al., 1996; Satoguchi et al., 2005; Nagahashi and Satoguchi, 2007; Henry et al., 2012). The Brunhes Chron is identified in the first 15 mbsf and the Matuyama Chron is well defined from 15-75 mbsf. Within the Matuyama Chron, the Jaramillo (0.99-1.06 Ma at ~26 mbsf) and Olduvai (1.77-1.95 Ma between 55 and 65 mbsf) subchrons are recognized. The Gauss Chron, a normal chron extending from 2.58 to 3.60 Ma, is identified between 75 and 126 mbsf, while the Gilbert Chron (3.60-6.03 Ma) is identified between 126 and 282 mbsf. The bottom 100 m of sediment have been correlated with the C3An.1n, C3An.1r, C3An.2n, and C3Ar Chrons from Gradstein et al. (2004) based on the shipboard magnetostratigraphy. Shore based samples did not record any reversals below 335 mbsf. The sedimentary sequence at Site C0011 appears to be continuous to 380 mbsf (~7.12 Ma). It is difficult to assign ages beyond 350 mbsf due to the lack of a lower constraining boundary.

At Site C0012, the sediments range in age from 0 to approximately 8.25 Ma with ten polarity zones in both the shipboard and shorebased magnetostratigraphies (Fig. 23). Some of the shorter reversals (less than 1 Ma) do not correlate between the ship and shorebased magnetostratigraphies, perhaps due to sampling recovery, especially with regards to shorebased samples taken approximately once per meter and a half.

The defined polarity zones at Site C0012 are similar to those of Site C0011, but there are a few significant differences. The Brunhes Chron (0-0.78 Ma) is recognized from 0 to 5
mbsf. This is a condensed section relative to Site C0011. The transition to the Matuyama Chron occurs around 5 mbsf and the Azuki (0.85 Ma) and Pink (1.05 Ma) ashes were identified within a meter of one another, just below the chron boundary (Hayashida et al., 1996; Henry et al., 2012). The Matuyama Chron is also significantly shorter than expected (only three meters of sediment for a 1.8 Ma time span). A significant section of the Matuyama Chron, and possibly part of the Gauss Chron, is missing between five and ten mbsf, indicating the presence of an unconformity, an unconformity not seen at Site C0011 (Fig. 23). Below the unconformity, the sedimentary record appears to be complete, with the identification of the Gauss Chron from just below the unconformity to approximately 14 mbsf and the Gilbert Chron from 14 mbsf to 101 mbsf. The C3An.1n, C3An.1r, C3An.2n, C3Ar, and C4r.1n Chrons are identified between 125 and 178 mbsf and are correlated with the Gradstein et al. (2004) GPTS. An age of approximately 8.25 Ma has been assigned to the sediments at the bottom of Hole C0012D.

**Sediment Accumulation**

By combining the ages of sediments with the depths at which the sediments were collected, rates of accumulation can be calculated for both Sites C0011 and C0012. As previously noted, the rates reported have not been corrected for compaction. Compaction is minimal in much of the sediment at Sites C0011 and C0012, likely due to the presence of silica cement. Decreased density and resistivity values and high silica content in the water collected from Site C0011 between 80 and 240 mbsf and Site C0012 between 12 and 70 mbsf led the scientific party of Exp 333 to believe that silica cement, possibly related to opal diagenesis, was present in the sediments (Spinelli et al., 2007; Henry et al., 2012). Compaction correction is unnecessary for much of the sedimentary section sampled at both
sites due to the minimal compaction present in the sediments. The accumulation rate should reflect the differences in deposition, possibly due to differences in site location or depositional mechanism, between Sites C0011 and C0012.

The rate of accumulation appears to have shifted significantly three times at Site C0011 (Fig. 24A). The oldest (7.14 to 5.24 Ma) sediments record an average accumulation rate of approximately 4.3 cm/kyr, but the accumulation rate may have been as low as 3.0 cm/kyr and as high as 6.6 cm/kyr during this period. It is difficult to characterize the rate of accumulation in these older sediments due to uncertainties in the correlation to the GPTS. The rate of sediment accumulation increased to roughly 7.5 cm/kyr around 5.24 Ma. This period of increased accumulation continued until 3.03 Ma, at which time sediment accumulation decreased to a rate of only 1.7 cm/kyr. From 1.95 Ma to 0.78 Ma, the rate of accumulation increased to approximately 4.2 cm/kyr. The present (0.78 Ma to 0 Ma) accumulation rate is approximately 2.2 cm/kyr.

At Site C0012, there are five distinct time periods defined by the rate of deposition (Fig. 24B). From roughly 8.23 Ma to 7.14 Ma the average rate of accumulation was 4.9 cm/kyr. A shift in the sediment accumulation rate occurred around 7.14 Ma, with the rate slowing to 2.0 cm/kyr. This reduced rate of sediment accumulation continues to approximately 5 Ma. Accumulation between 5 Ma and 4.63 Ma slowed even further to 0.30 cm/kyr. Following this period of very slow sediment accumulation the rates of accumulation increased greatly. In the period between 4.63 Ma and 3.6 Ma the sediments record a rate of accumulation of 6.2 cm/kyr. This period of rapid accumulation is immediately followed by a hiatus of deposition for roughly 2.5 Ma. The 2.5 Ma hiatus and the period from 5-4.63 Ma (accumulation rate of 0.30 cm/kyr) may represent a lack of deposition or the removal of
sediment after deposition. The most recent sediments (1.05 Ma to present) record a rate of approximately 0.70 cm/kyr.

At present, sediment accumulation at Site C0012 occurs more slowly than sediment accumulation at Site C0011. Even though accumulation rates vary for both sites, this trend holds consistent over the past 8.23 Ma. Site C0011 is lower in the basin on the Kashinosaki Knoll, allowing for sediments to accumulate quickly with very little disturbance, whereas the distal position of Site C0012 on a bathymetric high lends itself to slower sediment accumulation and, in certain cases, erosional processes.

The two depositional hiatuses at Site C0012 are prime examples of the erosional processes that may have been active at the site. Both hiatuses are likely unconformities created by the removal of once overlying sediment. The hiatus between 5 and 4.63 Ma (81.2-82.2 mbsf) is short and not well documented by the available data. For the purposes of this study, further interpretation will be focused on the hiatus from 3.6 to 1.1 Ma due to the availability of supporting data.

The likelihood of an unconformity between five and ten mbsf is strongly supported by much of the data collected during this study. The hiatus of sedimentation for roughly 2.5 Ma, as noted in the magnetostratigraphy and by the missing Habutaki ash (2.85 Ma), was the first evidence for an unconformity at Site C0012 (Fig. 23). There is also a strong tendency toward oblate AMS ellipsoids at Site C0012, which may be related to the compaction of the sediments before the removal of overlying sediments (Fig. 27). In the Exp 333 shipboard physical properties studies, the porosity values at Site C0012 (below 240 mbsf) were found to generally be lower than the porosity values at Site C0011 for the corresponding interval, suggesting that the sediments at Site C0012 have undergone more compaction (Fig. 19D and
The anisotropy data suggest a similar trend. A significant increase in P does not occur until 125 mbsf at Site C0012. A similar increase does not occur at Site C0011 until 250 mbsf (Fig. 19A and 20A). The shift of anisotropy at shallower depths, the decreased porosity, along with the sedimentation hiatus, and the AMS ellipsoid shape strongly support the existence of an unconformity at Site C0012, most likely caused by the removal of sediment due to submarine landslides or erosion due to ocean currents.

**Sediment Transport**

Sediment transport directions appear to be linked to the changes in sediment accumulation rates at Sites C0011 and C0012. Sediment accumulation rates at Site C0011 below 250 mbsf are approximately 4.3 cm/kyr. This rate correlates with the timing of a paleocurrent direction to the SW (250-350 mbsf), along the flank of the Kashinosaki Knoll, possibly carrying sediment away from Site C0011 (Fig. 3). The paleocurrent direction shifts to a SE direction around 250 mbsf and stays relatively consistent from 5.24 to 4.3 Ma. During this period (~5.25 to 3.12 Ma), sediment accumulation increases to a rate of 7.5 cm/kyr. A SE current would be flowing normal to the Kashinosaki Knoll and possibly deposit an increased sediment load at Site C0011. The accumulation rate from ~90 to 0 mbsf at Site C0011 ranges from 3.4 cm/kyr to 1.7 cm/kyr. Sediment transport directions during this period are to the NW-SE. The NW transport of sediment away from the Kashinosaki Knoll may reduce sediment accumulation at Site C0011. At Site C0012, the rate of accumulation stays consistent at approximately 2.0 cm/kyr from 7.14 to 4.4 Ma. During this same time period no distinct paleocurrent direction is recorded. With no directional paleocurrent carrying sediment toward Site C0012 or away from the site the sediments appear to settle at a consistent rate. The accumulation of sediment is increased (~4.9 cm/kyr)
from 7.14 Ma to 8 Ma when the paleocurrent direction is predominantly to the SE, toward the knoll. The ocean currents appear to have a significant effect on the accumulation of sediment at Sites C0011 and C0012.

The paleocurrent directions recorded at Sites C0018, C0011, and C0012 in the Shikoku Basin may be controlled by the Kuroshio Current and the Antarctic Intermediate Water (AAIW) Current. The Kuroshio Current is a wind-driven western boundary surface current that flows N-NE through the Shikoku Basin and has been active for the past 15 Ma (Sawada and Handa, 1998; Qiu, 2001). The current has varying paths, one straight and the other a meander, causing the current to affect different portions of the basin at different times (Sawada and Handa, 1988; Qiu, 2001). When the Kuroshio Current flows close to the coast, the path of the current is straight. The Kuroshio Current is periodically deflected away from the coast and these periods are referred to as large meander periods. During the straight path periods the Kuroshio Current may have less impact on the sediments at Sites C0011 and C0012, but during large meander periods the current may play a role at Sites C0011 and C0012. The sediments, which record the generally N paleocurrent direction, may represent periods when the Kuroshio Current had a greater impact on sediment deposition. The AAIW is a strong S-SE flowing bottom current that may be the cause of the southern paleocurrent directions noted prior to 4 Ma at both Sites C0011 and C0012 (Underwood and Steurer, 2003). Although these two currents are likely related to deposition in the Shikoku Basin, based on the current data set a direct link cannot be conclusively drawn.
Depositional Mechanisms

The differences in depositional mechanisms identified previously are most likely due to the environments in which the sites are located. Deposition at Site C0018, located in the accretionary prism, is dominated by current and grain collision (Fig. 32A). Grain collision results from the layer-by-layer collision of grains, typically on an inclined surface, or the bombardment of grains resting on a surface by transported grains. The Site C0018 sediments were likely deposited by SE downslope movement of the sediments on the actively deforming accretionary wedge. Deposition at Site C0011 is located on the flank of the Kashinosaki Knoll in the Shikoku basin (Fig. 32B). Site C0011 it is not actively deforming and the depositional environment is quieter then the depositional environment at Site C0018. Deep ocean currents also appear to be important at Site C0011 as the rate of sediment accumulation is tied to current direction. And finally, the dominant depositional mechanism at Site C0012 is a combination of slope-gravity and collision interaction (Fig. 32C). This may reflect the site’s location near the summit of the Kashinosaki Knoll, where the slope is steep and grains may roll due to gravity. The slope at Site C0012 may have been unstable, as indicated by the two unconformities identified in the sedimentary section. No one process is dominant at these deep ocean sites. Deep ocean deposition is variable and is controlled by plate movement, ocean currents, and slope stability.

Steep imbrication angles (above 35°) recognized in many of the cores are accompanied by highly variable K₃ inclinations. Bioturbation does not appear to be the cause of these steep imbrication angles at Site C0018 or Site C0011 but may play a role at Site C0012. The bioturbation in the upper unit at Site C0012 may be the cause of the fabric
and grain reorientation identified in Site C0012 Cores 1H-8H AMS data, but this is difficult to ascertain for certain.

Because average q and β were used to generate the q-β plots, inhomogeneous sediments and cores cannot be used in the analyses of depositional mechanisms. Cores with a K₃ standard deviation greater than 20° are represented with open symbols in the q-β plots (Fig. 32) and may represent inhomogeneous sediments, while samples with a standard deviation lower than 20° may represent homogeneous sediments. At Sites C0018 and C0011, all cores with high imbrication angles also have a K₃ standard deviation greater than 20°. The high imbrication angles in the sediments from Sites C0018 and C0011 are likely due to sediment inhomogeneity. The samples from Site C0018 with high imbrication angles were all taken from within MTD six and contain a mixture of sediment types and grain sizes. Due to the chaotic nature of these deposits, defining a specific ‘MTD’ region on the q-β diagram is difficult. The MTD intervals are not well-suited for this analysis, because of the effects of slumping and sediment disturbance on the fabric orientation. The steep imbrication angles in the Site C0011 sediments may be due to the presence of ash and volcaniclastic sand within the sediments (Fig. 5). Depositional mechanism analyses using q and β values must be limited to homogenous sediments.

At Site C0012, four samples have high imbrication angles, but standard deviations below 20°. Site C0012D Core 13H was not corrected for bedding and therefore the imbrication angles may not be accurate. In the sediments collected from Site C0012C Cores 3H, 5H, and 6H, the standard deviations of K₃ are high (15°, 13°, and 19° respectively) but do not meet the exclusion criteria. The EA plots for these cores also indicated sediment disturbance. These samples may have also been disturbed after deposition due to sampling
or coring processes. Imbrication angles over 35° may be useful as indicators for disturbed clay-rich sediments.

The type of depositional mechanism dominant at each site was expected to be related to the sediment transport direction. Upon comparison of the q-β diagrams and the sediment transport direction results, it was determined that the depositional mechanism and sediment transport direction were independent of one another. The depositional mechanisms at a site vary more frequently than the sediment transport direction. Although the sediment transport direction plays a role in determining the amount of sediment deposited at a site, it does not appear to control the type of deposition occurring.

**Consolidation Trends**

The amount of compaction and consolidation within the sediments influences the way inclination shallowing is predicted to behave. Porosity measurements, as well as anisotropy and water content measurements, are used in characterizing compaction. A comparison between these data sets provides a platform for understanding inclination trends at Sites C0018, C0011, and C0012.

The dominant ellipsoid shape at all of the sites presented here is oblate (Fig. 25-27). Although some individual samples display a tendency toward a prolate ellipsoid, no core has an overall prolate ellipsoid. The τ₃ 95% confidence bounds are distinct from τ₁ and τ₂ for many of the cores and the degree ratio of F to L increases with depth, most notably at Site C0011 and Site C0012 (Fig. 26 and 27). Samples from Hole C0012E and C0012G, taken from 500-525 mbsf (325 m below the samples from C0012D), are characterized by strongly oblate ellipsoids (Fig. 27). The increasingly oblate fabrics in the cores from below 500 mbsf may be related to increased compaction and consolidation. It is not clear that compaction
caused the oblate ellipsoids in the shallower sediments. The degree of foliation does not appear to vary greatly by depth. The oblate fabric found in most of the sediments may simply be a result of platy clay-grain deposition.

However, not all of the AMS ellipsoids are oblate. At Site C0018, cores one through four and 13 through 23 do not fit the criteria for an oblate ellipsoid (Fig. 29B). Eigenvectors are not well clustered and the 95% confidence bounds overlap for these specific cores. The 95% confidence bounds cannot be referenced for these samples because the ratio of foliation to lineation is not homogeneous throughout the core. Various cores from Site C0011 and C0012 also display this behavior (Fig. 30B and 31B). Although these AMS ellipsoids could be classified as spherical, it is more likely that the AMS ellipsoids are a mixture of prolate and oblate ellipsoids and that they are not well represented by the core averages.

The inhomogeneity within Cores 13H-23H from Site C0018 suggests the sediments within these cores have been disturbed. The AMS ellipsoids for these cores were classified as “spherical” (Fig. 29B), but the sediments are inhomogeneous with regards to ellipsoid shape and an average AMS ellipsoid is not representative for these cores. Average imbrication and lineation values for sediments within the MTD are often high and can be seen on a q-ß diagram (Fig. 32A). Inclination angles within this MTD are scattered (Fig. 18C). The disturbance of sediments as seen from the AMS data and inclination begins two cores earlier, core 13 rather than 15, than originally documented.

The sedimentary fabric disturbance in Site C0018 Cores 13H and 14H is caused by drilling disturbance and not sedimentary deposition. At first it appeared that MTD six was thicker than originally documented, but the upper boundary of MTD six is well documented and constrained by overlying ash layers (Henry et al., 2012). The likely cause of the
disturbance is the drilling process itself. These cores were collected with a hydraulic piston coring system. As sediments become consolidated they react in a brittle fashion to the coring, and as seen in Figure 17B the sediments break into large pieces separated by voids. The fabrics recorded in these sediments may be disturbed and therefore the data obtained from these cores may be unreliable.

Normal consolidation of the sediments at Sites C0011 and C0012 was expected due to the relatively quiet environment in which the sites are located. This is not the case. Magnetic anisotropy should increase with depth during normal consolidation, but the anisotropy of the sediments at Site C0012 does not begin to increase with depth until approximately 80 mbsf. The increase in magnetic anisotropy becomes more noticeable at approximately 125 mbsf. This depth also marks where water volume per section is reduced to 1 mL/cm. Prior to 125 mbsf, the water volume per section is relatively constant. Water volume per section at Site C0011 declines to 1 mL/cm at roughly 250 mbsf according to Henry et al. (2012), the same depth at which anisotropy begins to steadily increase. Because of the inverse relationship between water content and anisotropy, measurements of anisotropy such as P values and eigenvalues may not be particularly useful until water content of the sediments has been reduced to a threshold value. There appears to be little consolidation in the uppermost 250 m and 80 m of sediment at Sites C0011 and C0012 respectively.

The trends in anisotropy with depth, as determined by the difference in $\tau_1$ and $\tau_3$, can be used to identify compaction disequilibria (Schwehr et al., 2006). Site C0018 and C0011 exhibit evidence for possible underconsolidation (reduced anisotropy). These zones are less compacted than expected, which may be due to lithology changes or sediments with low
permeability trapping water within the underlying sediments and retarding compaction. At Site C0018, the zone of decreased anisotropy corresponds with an increase in sandy lithologies. The samples in section and two from Site C0011 were collected from an interval with abundant volcanioclastic sand and ash. Because these intervals are not predominantly clay, the shape behavior of the grains does not react as predicted by Schwehr et al. (2006).

Apparent overconsolidation was not identified at any of the sites. Expectations were that anisotropy would increase sharply below the 2.5 Ma unconformity at Site C0012, but no increase in the difference between $\tau_1$ and $\tau_3$ is noted. The sampling interval may be too large and may not provide the detailed record of the subtle eigenvalue changes necessary for identifying an unconformity. It is also possible that the unconformity is not associated with any significant difference in sediment consolidation.

The sediments from Site C0018 record a period of normal consolidation (Fig. 18D; Henry et al., 2012). Porosity at Site C0018 generally decreases down hole, with increases occurring at the base of MTD two, three, five, and six (Henry et al., 2012). A possible increased compaction gradient relative to the average porosity-depth trend was noted within the MTDs during Exp 333 (Henry et al., 2012). Pore water volumes above 130 mbsf are relatively constant (1.95 to 2.15 mL/cm). Below 130 mbsf, the amount of pore water decreases rapidly, most likely due to compaction of the sediments and the increase of sand deposits (Fig. 4; Henry et al., 2012). Even though normal consolidation is recorded at Site C0018, no inclination shallowing was identified.

Magnetic anisotropy and paleomagnetic inclination do not seem to be directly correlated with porosity and compaction at Site C0018 (Fig. 18). The P values are relatively constant from 0-200 mbsf, likely due the disturbance of the sediments as indicated by the
high degree of scatter in the paleomagnetic inclination values beginning around 130 mbsf (Fig. 18C). Magnetic anisotropy within MTD six may not record the original compaction state of the sediments due to disturbance caused by mass wasting. The P values and eigenvalues indicated no areas of compaction disequilibria. The magnetic fabric analyses did not lead to the characterization of surfaces of weakness or MTDs within the sediments at Site C0018.

Inclination does not appear to be strongly linked to porosity at Site C0011 (Fig. 19C and D). Porosity decreases, reflecting normal consolidation, from 0-50 mbsf, but inclination shifts from a decreasing trend to an increasing trend at 25 mbsf. As discussed earlier, inclination decreases from 200 to 250 mbsf and is relatively constant from 250-378 mbsf. Trends in porosity are more complicated in this region with porosity holding constant from 80-240 mbsf and suddenly decreasing, followed by a normal consolidation trend (Henry et al., 2012).

Due to the scatter in the inclination values, it is difficult to draw concrete relationships between inclination and porosity at Site C0012. Inclination values appear to be loosely related to porosity at Site C0012. Shallowing occurs in the inclination values around 15 mbsf; near the same depth, a sudden increase in porosity is noted. Around 137 mbsf, inclination values steepen and then hold constant around the expected inclination of 52°. Porosity around 137 mbsf begins to decrease, following the normal consolidation trend. Porosity and inclination are influenced by similar factors at Site C0012, but what these factors are is not completely understood.

Anomalously high porosity and low anisotropy at Sites C0011 and C0012 may be due to the presence of silica cement (Henry et al., 2012; Yujin et al., In Review). Porosity and
anisotropy remain relatively constant from the ocean floor to 250 mbsf and 80 mbsf at Sites C0011 and C0012 respectively (Fig. 19A, 19D, 20A, and 20D). The constant porosity and anisotropy in the shallow sediments is attributed to impeded compaction due to cementation. Similar cementation has been found in the Shikoku Basin (Spinelli et al., 2007). This cement strengthens the sediments and inhibits mechanical compaction. The cement may prevent an increase in anisotropy and in turn prevent inclination shallowing. There is a reduction in porosity and an increase in anisotropy below 250 mbsf at Site C0011 and 80 mbsf at Site C0012. These changes in the sediment physical properties may be a result of compaction in the sediments following the breakdown of the cement.

Multiple sources of the possible silica cement have been suggested. The source of the cement may be biogenic opal-A as suggested by Spinelli et al. (2007) or the alteration of volcanic glass in the hemipelagic mud as suggested by White et al. (2011). Spinelli et al. (2007) posit that when biogenic opal-A transitions into opal-CT silica cement can be produced which prevents compaction. Upon the diagenesis of the opal-CT to quartz the cement can no longer prevent compaction and the sediments can consolidate (Spinelli et al., 2007). These processes are thermally dependent and the sediments from Sites C0011 and C0012 may be too shallow to meet the thermal requirements for opal diagenesis (Yujin et al., In Review). Biogenic opal-A cannot be ruled out as the source of the cement because the initial conditions are variable. Volcanic glass is another possible source of the silica cement. Volcanic glass can alter to amorphous silica due to dehydration and heating. The amorphous silica can coat grain boundaries and form a resistant framework that would impede compaction (White et al., 2011). White et al. (2011) notes that the processes of opal-A transformation to opal-CT and the alteration of volcanic glass may coincide, making both of
these explanations for the possible cement present in the sediments at Sites C0011 and C0012 viable.

The inclination steepening noted in the sediments at Sites C0011 and C0012 is not an isolated phenomenon. Steep inclinations have been noted in previous studies, such as Ocean Drilling Program Leg 190 (Moore et al., 2001). Steep inclinations angles are often attributed to deformation of the sediments after deposition (Tan and Kodama, 2002). In the case of Sites C0011 and C0012, block rotation is unlikely given the stability of the Shikoku Basin and Philippine Sea Plate over the past 15 Ma. Deformation may be the cause of the steep inclination angles recorded in the sediments, but the presence of sediment deformation is not clear at either site. Non-dipole fields may also contribute to discrepancies between the observed and predicted magnetic inclination values (Schneider and Kent, 1986; Kent and Smethurst, 1998; Si and Van der Voo, 2001). Global or regional quadropole or octupole fields can cause steeper, or shallower, then expected magnetic inclination values. Corrections may be applied to reduce the influence of a non-dipole field on the magnetic inclination record, but were not conducted during this study (Schneider and Kent, 1986; Si and Van der Voo, 2001). The cause of the steepened inclinations at Sites C0011 and C0012 cannot be fully explained at this time.
CONCLUSIONS

Magnetic fabric analyses proved to be useful in characterizing the sedimentary fabric, the sediment transport direction, and the depositional mechanisms in the sediments collected during Exp 333. Specific findings include the following.

- The fabric shapes at Sites C0018, C0011, and C0012 are all dominantly oblate and begin to become more oblate with increasing depth. The oblate sedimentary fabrics are probably due to late-stage compaction as well as the high clay content in the sediments.

- Sediment transport, including MTD movement, on the accretionary prism at Site C0018 is downslope to the SE. At Site C0011, sediment transport is predominantly to the SE from 5.24 Ma to present and prior to 7.12 Ma. The period from 7.12-5.24 Ma is marked by sediment transport to the SW. Finally, the sediments at Site C0012 record no distinct sediment transport direction from 7.14 Ma to present. A current direction to the SE is recorded at Site C0012 from 7.14-7.6 Ma.

- Current deposition is dominant at Sites C0018 and C0011 but is not significant at Site C0012, where the interaction between slope-gravity and collision deposition dominate. The current deposition at Site C0018 reflects the downslope movement of sediments on the accretionary wedge.

The characterization of these sediments will aid in the understanding of what is entering the subduction zone at the Nankai Trough and will aid in future NanTroSEIZE research.

Water content and the probable presence of cement greatly impact the compaction of the sediments collected from Sites C0011 and C0012. These ‘real world’ factors complicate
the AMS interpretations. Sediments are not deposited in closed systems, and therefore many factors can and do affect the preservation of the magnetic fabric. The application of the Schwehr et al. (2006) method for identifying compaction disequilibria through AMS may not be universally applicable due to the large number of variables affecting compaction in naturally deposited sediments, especially when detailed sampling (i.e. closely spaced samples) is not available.

The inclinations recorded by the undisturbed sediments from Sites C0018, C0011, and C0012 vary greatly. There is no trend toward inclination shallowing at any of the sites, but steepened inclinations were noted. Silica cement may have locked the inclination values in for sediments from Sites C0011 and C0012, and the steep inclinations may be a manifestation of this behavior. The overly steep inclinations may indicate long-term non-dipole field behavior and in turn suggest that the GAD hypothesis may be incorrect. Corrections for non-dipole fields, such as those completed by Schneider and Kent (1986) and Si and Van der Voo (2001), may reduce the possible influence of a non-dipole field and result in a more accurate paleolatitude or paleogeographic reconstruction. But the progression of shallow inclinations in the upper sections of Sites C0011 and C0012 to steeper inclinations in the lower, more deeply buried and consolidated sediments is not fully understood. Regardless, this study does demonstrate that marine sediments can record the geomagnetic field without significant inclination bias. Therefore the paleomagnetic record from sediments and sedimentary rocks does not universally suffer from inclination error, for example as argued by Butler et al. (2001), so paleolatitudes and paleogeographic reconstructions that use sediments and sedimentary rocks can be used with greater confidence.
REFERENCES


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<table>
<thead>
<tr>
<th>Coring Systems</th>
<th>Core Letter Abbreviation</th>
<th>System Explanation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hydraulic Piston Coring System (HPCS)</td>
<td>H</td>
<td>HPCS inner core barrel is run to the bottom of the coring wireline. Pump pressure is applied to the drill pipe, severing the share pins and strokes the inner core barrel 9.5 m into the sediment. The inner core barrel is retrieved by wireline.</td>
</tr>
<tr>
<td>Extended Punch Coring System (EPCS)</td>
<td>T</td>
<td>EPCS is used to provide near insitu core in formations too firm for HPCS instead of ESCS, which creates more disturbance in sediments. The coring system punches ahead of the bit in a similar fashion to HPCS, but only 2-5 inches of advancement are made at a time. The inner core barrel does not rotate.</td>
</tr>
<tr>
<td>Extended Shoe Coring System (ESCS)</td>
<td>X</td>
<td>Used for recovery of 9.5 m long core samples from soft to moderately hard formations. ESCS is typically used when sediments are too hard for EPCS, but too soft for RCB recovery. The ESCS cutting shoe extends ahead of the main bit and retracts into the main bit with added weight. ESCS relies on rotation of the drill string to advance the hole.</td>
</tr>
<tr>
<td>Rotary Core Barrel (RCB)</td>
<td>R</td>
<td>The RCB bottom hole assembly (the bit and outer core barrel) are rotated with the drill string while bearings allow the inner core barrel to remain stationary. The RCB bit trims 2.312 in. core. The 9.5 m core sections are retrieved by wireline.</td>
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Table 2- Expedition 333 Drilling Summary (Henry et al., 2012)

<table>
<thead>
<tr>
<th>Hole</th>
<th>Water depth (mbsl)</th>
<th>Cores (N)</th>
<th>Cored (m)</th>
<th>Recovered (m)</th>
<th>Recovery (%)</th>
<th>Drilled (m)</th>
<th>Penetration (m)</th>
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<td>C0018A</td>
<td>3084.35</td>
<td>36</td>
<td>314.15</td>
<td>271.44</td>
<td>86.4</td>
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<td>C0011C</td>
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<td>3</td>
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<td>102.2</td>
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<td>15</td>
<td>124.5</td>
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<td>63.8</td>
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<td>C0012E</td>
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<td>48.2</td>
<td>525.5</td>
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<td>15</td>
<td>115.5</td>
<td>25.92</td>
<td>22.4</td>
<td>630.5</td>
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<tr>
<td><strong>Expedition 333 totals:</strong></td>
<td><strong>139</strong></td>
<td><strong>1033.45</strong></td>
<td><strong>883.34</strong></td>
<td><strong>76.625</strong></td>
<td><strong>2705.65</strong></td>
<td><strong>2705.65</strong></td>
<td></td>
</tr>
</tbody>
</table>

Notes: N = number.
Figure 1- Bathymetric map of the Shikoku Basin off the coast of the Kii Peninsula in Japan adapted from Henry et al. (2012). The colors represent depth in meters below sea level (mbsl). Brown= 1000 mbsl, Yellow= 2000 mbsl, Green= 3000 mbsl, and Blue= 4000 mbsl. Sites from Expedition 333 are marked by red circles. The yellow arrow depicts the estimated convergence between the Philippine Sea plate and Japan (Seno et al., 1993; Heki, 2007; Henry et al., 2012).
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Figure 3- Detailed bathymetric map of Kashinosaki Knoll and the Nankai Trough adapted from Henry et al. (2012). The locations of Sites C0018, C0011, and C0012 are shown. The colors represent depth below sea level. Green= 3000 mbsl and Blue= 4000 mbsl.
Figure 4- Core recovery log and lithologic column from Site C0018 adapted from Henry et al. (2012). MTD = Mass transport deposit. The numbers one through six on the right side of the lithologic column are the identifying numbers for each MTD (i.e. MTD 1, MTD 2, etc.). Distinct ash deposits (the Azuki and Pink ash layers) are highlighted.
Figure 5- Core recovery log and lithologic column from Site C0011 adapted from Henry et al. (2012). MTD= Mass transport deposit. Distinct ash deposits (the Azuki, Pink, Habutaki, and Ohta ashes) and their ages are indicated.
Figure 6- Core recovery log and lithologic column from Site C0012 adapted from Henry et al. (2012). Distinct ash deposits (the Azuki, Pink, and Ohta ashes) and their ages are indicated.
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Figure 9-Site C0018 orthogonal vector diagrams and relative intensity diagrams. Data are shown on the horizontal plane with solid squares, while data on the vertical plane are denoted by open squares. A) Type 1 with a steep first removed component and a shallower second removed component progressing toward the origin. Relative intensity decreases with progressive demagnetization. B) Type 2 with two components. The second removed component progresses away from the origin and relative intensity progressively increases starting around 60 mT. AGRM techniques were applied to samples with these characteristics. AGRM techniques were applied to this specific sample at the 90 and 100 mT steps where a decrease in the relative intensity can be seen. The second removed component begins to trend toward the origin after the AGRM methods are applied. C) Type 3 with no well-defined ChRM.
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Figure 11- Site C0012 select orthogonal vector diagrams and relative intensity plots. Data are shown on the horizontal plane with solid squares, while data on the vertical plane are denoted by open squares. Examples of A) A sample with normal polarity and B) a sample with reverse polarity both with two components. The second removed component progresses toward the origin and relative intensity decreases with progressive demagnetization.
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Figure 15- Illustration of grain imbrication. Equal area projections are upper hemisphere adapted from Taira (1989). Circles=K1, the maximum eigenvector; X=K3, the minimum eigenvector
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Figure 17- Example images of cores collected during Exp 333.  A) An undisturbed section of hemipelagic mud collected from Hole C0018A Core 12H Section 8 with a hydraulic piston coring system. B) A disturbed core section with cracks and voids (no rotation) collected from Hole C0018A Core 13H Section 4 using a hydraulic piston coring system. C) A highly disturbed core section from Hole C0011D Core 36X Section 1.  The core was collected using an Extended Shoe Coring System (ESCS) which rotates and causes the core to break into ‘biscuits’ or individual sections of sediment separated by drilling fluid. Each biscuit has been outlined in white. The material between and around each biscuit is drilling fluid.
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Figure 20- Site C0012 data versus depth. A) Degree of Anisotropy (P=K1/K3). Blue lines separate the sections with differing trends. The black trendline is a moving average using five data points. B) Eigenvalues for the AMS ellipsoid. τ1 is shown in blue and τ3 is shown in red. The black lines highlight the areas with eigenvalue trends that vary from the expected trend as discussed in the text. C) Inclination. The red vertical lines indicate the expected inclination of 52 or -52 degrees. Changes in the inclination angle occur multiple times with significant changes occurring at 20, 50, 85 and 112 mbsf. Solid diamonds indicate data have been corrected for bedding dip and open diamonds represent data that did not undergo bedding corrections. D) Porosity data from Exp 333 (Henry et al., 2012).
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Figure 22- Site C0011 magnetostratigraphy. The on-shore magnetostratigraphy from this study and the Exp 333 shipboard magnetostratigraphy are provided for comparison with the GPTS from Gradstein et al. (2004). Tephrochronology and biostratigraphy ages provide further age constraints (Henry et al., 2012).
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Figure 24- Accumulation Rates. A) Site C0011 sediment accumulation rates and B) Site C0012 sediment accumulation rates with sedimentation hiatuses also marked. Red squares represent tephrochronology ages, blue diamonds represent magnetostratigraphy ages, and green triangles represent biostratigraphy ages. Accumulation rates for each section are included on each plot.
Figure 25- Site C0018 Flinn Plots. The x-axis is foliation and the y-axis is lineation. The black diagonal line represents a one to one ratio of lineation to foliation. Data points along this line are characterized by sedimentary fabrics with spherical ellipsoids. Oblate fabrics fall below the diagonal line and prolate fabrics fall above the diagonal line. Each color represents a separate core (c1=core 1, c2=core 2, etc.) Cores 13 through 23 do not follow the expected trend of increasingly oblate fabrics with depth. Cores 24 through 36 have increasingly oblate fabrics.
Figure 26- Site C0011 Flinn Plots. The x-axis is foliation and the y-axis is lineation. The black diagonal line represents a one to one ratio of lineation to foliation. Data points along this line are characterized by sedimentary fabrics with spherical ellipsoids. Oblate fabrics fall below the diagonal line and prolate fabrics fall above the diagonal line. Each color represents a separate core (c1=Hole C, Core 1; c2=Hole C Core 2; d1=Hole D, Core 1, etc.) The fabrics are predominantly oblate.
Figure 27- Site C0012 Flinn Plots. The x-axis is foliation and the y-axis is lineation. The black diagonal line represents a one to one ratio of lineation to foliation. Data points along this line are characterized by sedimentary fabrics with spherical ellipsoids. Oblate fabrics fall below the diagonal line and prolate fabrics fall above the diagonal line. Each color represents an individual core (c1= Hole C, Core 1; c2=Hole C core 2; d1=Hole D, Core 1, etc.) The fabric becomes more oblate with depth. The bottom Flinn plot in the right column is a representation of the fabric shape in sediments taken from Site C0012 Holes E and G (depths of ~350 mbsf and deeper). Fabrics in samples from Holes C0012 E and G have distinctly oblate fabrics.
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Figure 32- q-B diagrams for A) Site C0018, B) Site C0011, C) Site C0012 after Taira (1989). Each point represents the average for an individual core, blue diamond= samples that did not require any corrections, red open square=K3 inclination with standard deviation greater than 20 degrees, green triangle=samples corrected for bedding dip. Each data point represents a distinct core.
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Figure 34- Site C0018 sediment transport directions determined rose diagrams representing the azimuthal orientations of the K3 axes. A) Unit Ia with a paleocurrent direction to the SE, B) Unit Ib piston cores with no distinct paleocurrent direction, C) Unit Ib rotary cores with a paleocurrent direction to the SE, and D) MTD 6 with a paleocurrent direction to the SE. Sector angles of the rose diagrams = 10 degrees. Each rose diagram is labeled with number of data points, as well as the mean azimuthal orientation and the 95% confidence interval for the mean azimuthal orientation. The tick interval indicates the value of each tick mark and is generally a fraction of one (e.g., tick interval 0.2 data indicates that each tick is equal to 20% of one data point and 5 ticks equal one data point.)
Figure 35- Site C0011 sediment transport directions determined from rose diagrams representing the azimuthal orientations of the K3 axes. A) Unit II with no distinct paleocurrent direction, B) Subunit Ib with a paleocurrent direction to the SW, C) Subunit Ia rotary cores with a paleocurrent direction to the SE, and D) Subunit Ia piston cores with a paleocurrent direction of NW-SE. Sector angles of the rose diagrams = 10 degrees. Each rose diagram is labeled with number of data, as well as the mean azimuthal orientation and the 95% confidence interval for the mean azimuthal orientation. The tick interval indicates the value of each tick mark and is generally a fraction of one (e.g. tick interval 0.2 data indicates that each tick is equal to 20% of one data point and 5 ticks equal one data point.)
Figure 36- Site C0012 sediment transport directions determined from rose diagrams representing the azimuthal orientations of the K3 axes. A) Subunit Ib with no distinct paleocurrent direction, B) Subunit Ic with an ESE paleocurrent direction, and C) Unit II with a SE paleocurrent direction. Subunit Ia is not shown because the data did not meet the primary fabric criteria. Sector angles of the rose diagrams=10 degrees. Each rose diagram is labeled with the number of samples, as well as the mean azimuthal orientation and the 95% confidence interval for the mean azimuthal orientation. The tick interval indicates the value of each tick mark and is generally a fraction of one (e.g. tick interval 0.2 data indicates that each tick is equal to 20% of one data point and 5 ticks equal one data point.)
Figure 37 - Day Plots of high field magnetic parameters measured for representative samples from A) Site C0018, B) Site C0011, C) Site C0012, and D) combined plot of Sites C0018, C0011, and C0012. Along the y-axis is the ratio of Mrs/Ms (saturation magnetization/saturation remanent magnetization) versus Hcr/Hc (remanent coercivity/coercivity) along the x-axis (figure after Dunlop, 2002).
Figure 38-Site C0018 magnetic hysteresis loops for select representative samples. A) and D) C0018A Core 5H Section 8 58cm, B) and E) C0018A Core14H Section 6 53.5cm, and C) and F) C0018A Core 34X Section 9 22cm. Plots A)-C) are uncorrected loops. Plots D)-F) are corrected loops for the same samples with the paramagnetic signature removed, allowing the magnetic hysteresis plot to provide values for magnetic parameters such as coercivity and remanence. All three samples show a mix of ferrimagnetic carrier with a paramagnetic component. Plot C is characterized as potbellied.
Figure 39-Site C0011 magnetic hysteresis loops with the paramagnetic signature removed. A) C0011C Core 1H Section 3 34cm, B) C0011D Core 51X Section 4 71cm, C) C0011D Core 44X Section 3 13cm. Saturation magnetization varies between samples.
Figure 40-Site C0012 representative magnetic hysteresis loops with the paramagnetic signature removed A) C0012C Core 1H Section 1 60cm, B) C0012G Core 1R Section 4 25cm, C) C0012D Core 13H Section 3 40cm, D) C0012D Core 9H Section 1 120cm.