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33. MINERAL MAGNETIC PROPERTIES OF MIDDLE AND UPPER PLEISTOCENE SEDIMENTS AT SITES 883, 884, AND 887, NORTH PACIFIC OCEAN[1]

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ABSTRACT

Mineral magnetic studies have been conducted on middle and upper Pleistocene sediments from Holes 883D, 884D, and 887B in the North Pacific Ocean to assess their suitability for determinations of relative paleointensity of the geomagnetic field. Thermomagnetic and high-field hysteresis data suggest that the magnetic mineralogy of the clay-rich sediments at Holes 883D and 884D is dominated by pseudosingle domain magnetite in a narrow grain-size range. The uniform grain size of the magnetic particles in the clay-rich sediment is consistent with sorting and transport by long-distance bottom currents. Coarse-grained magnetic minerals occur in high concentrations in restricted stratigraphic intervals where volcanic ashes occur. Apart from these intervals, the sediments meet the established criteria for relative paleointensity studies. Such studies are now in progress. The sediments at Hole 887B are not suitable for relative paleointensity determinations because of large variations in both grain size and concentration of magnetite. In particular, reductive diagenesis appears to have caused preferential dissolution of fine-grained magnetite in diatom-rich intervals in Hole 887B. Significant variation in the magnetic properties of diatom-, clay-, and tephra-rich parts of all the sequences preclude a straightforward interpretation of the magnetic susceptibility signal as a paleoclimate proxy as is the case in depositional environments that are dominated by climatically modulated $CaCO_3$ fluctuations.

INTRODUCTION

Magnetic susceptibility measurements are widely used in oceanographic studies for correlation of marine sediment cores. The primary causes of magnetic susceptibility variations in marine sediments are changes in the amount and nature of terrigenous materials that reach the drilling site and changes in grain size associated with sorting by currents and eolian processes. These factors are strongly influenced by climate; therefore, mineral magnetic studies often provide important constraints on paleoclimate processes (e.g., Kent, 1982; Mead et al., 1986; Robinson, 1986; Bloemendal et al., 1988, 1992; Doh et al., 1988; Bloemendal and deMenocal, 1989). A wide range of magnetic measurements are used in such studies to obtain information concerning variation in mineralogy, grain size, and concentration of magnetic grains. As well as providing information concerning paleoclimate-induced fluctuations, mineral magnetism offers powerful tools for discriminating between processes that can contribute to the magnetic signal, such as reductive diagenesis (Karlin, 1990a, 1990b; Leslie et al., 1990a, 1990b; Roberts and Pillans, 1993; Roberts and Turner, 1993), deposition of ice-rafted detritus or tephra layers, and determination of the relative contributions from eolian and bottom-current controlled deposition (Bloemendal et al., 1992).

Marine sediments are also important repositories of information concerning temporal variations of the geomagnetic field. Because marine sediments are usually deposited continuously, and because it is often possible to obtain high resolution dating of marine sediments using $\delta^{18}O$ stratigraphy, deep-sea sequences are prime targets for detailed studies of geomagnetic phenomena. Most conventional paleomagnetic studies concentrate on determining the direction of the paleomagnetic vector (declination and inclination). However, it is much more difficult to relate the remanence intensity to the intensity of the ancient magnetic field. In sediments, the intensity of the remanent magnetization in a sample is controlled by a number of factors, including the intensity of the geomagnetic field at the time the magnetization was acquired as well as the concentration and grain size of magnetic carriers. It has been suggested that the effect of concentration can be taken into account by normalizing the measured natural remanent magnetization (NRM) with a parameter that is proportional to the concentration of magnetic grains (Opdyke et al., 1973; Banerjee and Mellema, 1974; Levi and Banerjee, 1976; Tucker, 1981). The resulting ratio is a measure of relative paleointensity rather than absolute field intensity. The three quantities proposed as normalization parameters are the magnetic susceptibility, or χ, the anhysteretic remanent magnetization (ARM), and the saturation isothermal remanent magnetization (SIRM). Several stringent criteria have been proposed to determine the magnetic uniformity of a sediment for relative paleointensity studies (King et al., 1983; Tauxe, 1993). The criteria are (1) the dominant magnetic mineral must be magnetite; (2) the particle size range for the magnetite must be pseudosingle domain (PSD), between 1 and 15 µm; and (3) the maximum magnetite concentration must be less than 10 times greater than the minimum concentration.

Despite the difficulty in obtaining reliable records of relative paleointensity variations from sediments, such studies are important because they should enable the determination of constraints that will lead to a more complete understanding of geomagnetic field behavior. In many cases, sediments that pass various tests for magnetic uniformity appear to yield credible records of relative paleointensity. Some of the features of existing relative paleointensity records appear to be globally coherent. For example, Tric et al. (1992) produced an NRM/ARM record from the Mediterranean Sea that extends back to 80 k.y. This record is in agreement with, and has been calibrated against, paleointensity data from lavas covering the period 0–40 k.y. The record also shows significant agreement with earlier studies from the western equatorial Pacific by Constable and Tauxe (1987) and Tauxe and Wu (1990). Furthermore, Meynadier et al. (1992) have obtained a 140-k.y. record from the Somali Basin that closely reproduces features in the Mediterranean record of Tric et al. (1992).

The coherence of these records from around the world marks a significant step toward the establishment of a credible paleointensity reference curve for the last several 100 k.y. Valet and Meynadier (1992) have even suggested that the paleointensity of the geomagnetic field may provide the basis for a new global stratigraphy. Mineral magnetism has a major role to play in determining and confirming whether the paleointensity of the geomagnetic field may provide

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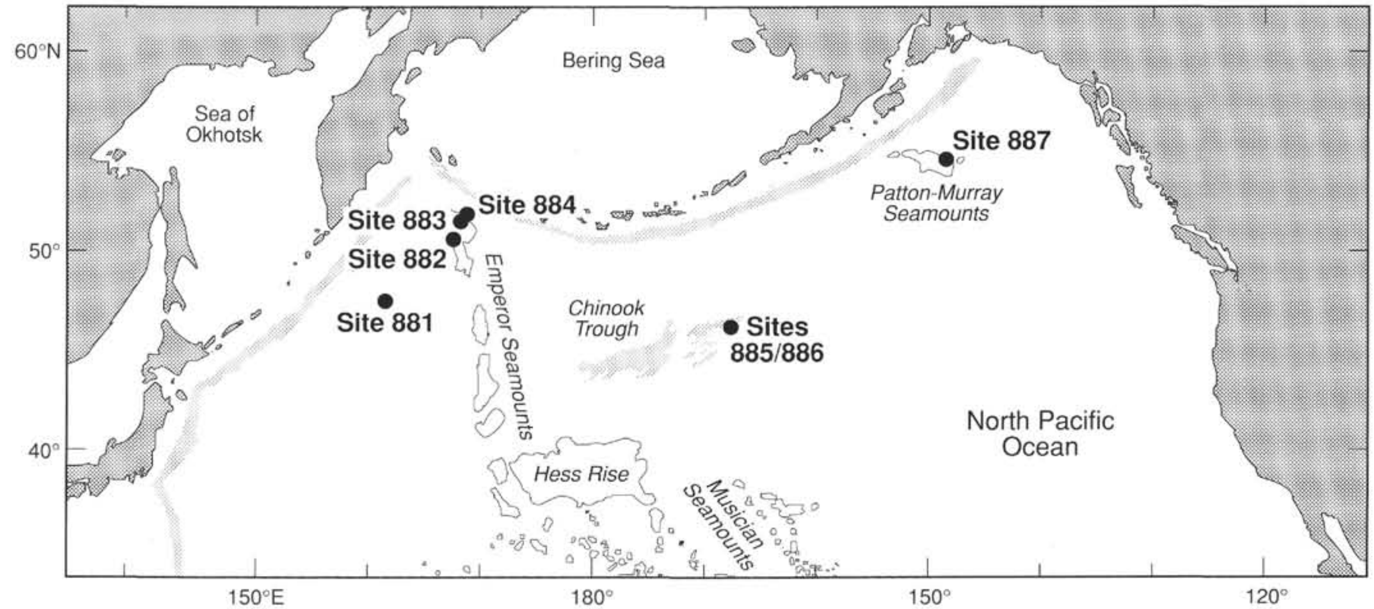

Figure 1. Location map of the North Pacific Ocean showing Leg 145 sites.

the basis for a new global stratigraphy. Despite the apparent agreement between paleointensity records from geographically widely distributed localities, significant differences exist between the records during some time intervals. These differences have been attributed to contributions from nondipole geomagnetic effects by some authors (e.g., Schneider, 1993; Tauxe, 1993); however, it is also possible that factors such as stratigraphically localized variation in grain size and concentration of magnetic minerals could contribute to a poor recording of the paleointensity signal. It is crucial, therefore, that all studies that report sedimentary paleointensity data be subjected to rigorous mineral magnetic characterization. The criteria of King et al. (1983) and Tauxe (1993) should be used to delineate those portions of a sequence that may be capable of yielding a reliable paleointensity record as well as those portions that may be less reliable. In this paper, we report detailed mineral magnetic studies that can be used to assess relative paleointensity data from the middle and upper Pleistocene records from Holes 883D, 884D, and 887B.

OCEANOGRAPHIC SETTING

The locations of Sites 883, 884, and 887 are shown in Figure 1. Sites 883 (51°11.898′N, 167°46.128′E) and 884 (51°27.026′N, 168°20.228′E) were drilled in close proximity along a three-site depth transect down the slopes of the Detroit Seamount. Of the three sites, Hole 883D was drilled at the shallowest water depth (2385 m) and Hole 884D was drilled at the deepest water depth (3826 m). Despite their close proximity, the depositional setting of these sites is distinctly different. In general, the North Pacific Ocean is highly corrosive to calcium carbonate; however, the recent discovery of foraminifer-bearing sediments on the Detroit Seamount (Keigwin, 1987; Keigwin et al., 1992) has suggested that Site 883 would be a suitable location for studies of stable isotope paleoceanography. Site 884 has lower calcium carbonate contents, presumably because of dissolution with depth. Site 884 is unique, however, because it is located on the Meiji Tongue, which is considered to be a drift deposit similar to those in the North Atlantic Ocean, where deep thermohaline currents are responsible for the long-term, long-distance transport of sediment. The Meiji Tongue progrades 2000 km south along the east side of the northern Emperor Seamount Chain (Scholl et al., 1977; Mammerickx, 1985). Mineral compositions indicate that the source of the Meiji deposit is probably the southwestern Bering Sea, with sediment transport occurring by means of a deep passage at the western extreme of the Aleutian Islands (Scholl et al., 1977). The sediments in Holes 883D and 884D are both dominated by dark gray clay with abundant diatoms and intermittent ash layers and dropstones.

Site 887 (54°21.921′N, 148°26.765′W) was drilled on a flat, elevated surface on the eastern part of the Patton-Murray Seamount group in the Gulf of Alaska at a water depth of 3630 m. The location of this site precluded turbidite sedimentation and ensured recovery of pelagic sediments. Determination of the evolution of the North Pacific water mass throughout the latest Quaternary (through stable isotope analysis of foraminifers) was a major goal at this site. Hole 887B was selected for relative paleointensity studies because of the likelihood of obtaining a temporal framework from the stable isotope stratigraphy. The sediments in Hole 887B are dominated by dark gray diatom clays with interbeds of pure diatom ooze and intermittent ash layers and dropstones.

METHODS

Because Holes 883D, 884D, and 887B were the third holes drilled at the respective sites, we were permitted to sample these sediments with U-channels (U-shaped, 2-cm^2 cross-section plastic channels, 1.5 m in length) to make continuous long-core paleomagnetic and mineral magnetic measurements, as described by Weeks et al. (1993). Holes 883D and 887B were drilled to depths of 17.0 and 40.0 mbsf, specifically for high-resolution stable isotope studies. Hole 884D was drilled to a depth of 14.8 mbsf specifically for high-resolution relative paleointensity studies. The entire lengths of Holes 883D and 884D were sampled for this study with U-channels, along with the upper 19.2 m of Hole 887B.

U-channel samples were subjected to pass-through paleomagnetic measurements on a 2-G Enterprises Model 755-R cryogenic magnetometer, equipped with high-resolution pickup coils. The measurements were made at the CFR Paleomagnetism Laboratory, Gif-sur-Yvette, France. Alternating-field (AF) demagnetizations were performed with in-line coils at peak fields of 0, 10, 20, 25, 30, 40, 50, and 60 mT. Because the position of the U-channel on the measurement track is known with high precision, it is possible to create vector component plots at regular intervals along the core to determine the stability of the remanent magnetization. Continuous low-field magnetic susceptibility measurements were made on all U-channels with a vertical translation

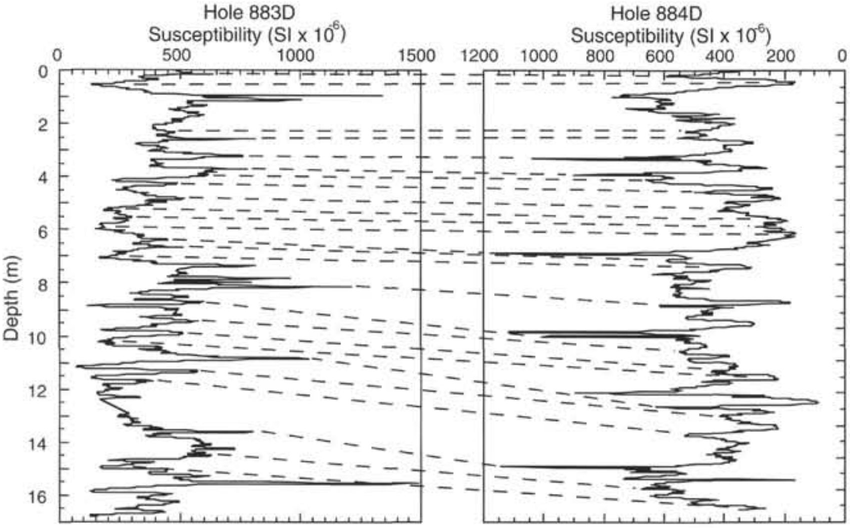

Figure 2. Correlation of magnetic susceptibility records for Holes 883D and 884D from U-channel samples. The Hole 884D record is inverted to facilitate correlation. Note differences in depth scales.

system in which susceptibility measurements are made at 1-cm intervals on a Bartington Instruments Model MS2 susceptibility meter. After AF demagnetization, we produced an anhysteretic remanent magnetization (ARM) in the U-channels using a Schonstedt AF demagnetizing coil within a large set of Rubens coils. The Rubens coils provide a region of constant bias field (0.05 mT) that surrounds the AF coils. The U-channels were moved through the AF coils, which were kept at a constant AF of about 99 mT. Subsequently, the ARM was subjected to an AF demagnetization at peak fields of 0, 10, 20, 25, 40, and 60 mT. All of the U-channel measurements made in this study have been described in detail by Weeks et al. (1993).

Other detailed mineral magnetic studies were made in the Paleomagnetism Laboratory at the University of California–Davis. Small sediment samples (about 0.5 cm^3) were collected at 10-cm stratigraphic intervals throughout the cores studied. Each sample was subjected to high-field magnetic hysteresis analysis up to maximum fields of 1 T using a Princeton Measurements Corporation Alternating Gradient Magnetometer. One bulk sediment sample from each core section (i.e., every 1.5 m) was subjected to temperature dependent susceptibility analysis, to maximum temperatures of 720°C using a Kappabridge KLY-2 with a CS-2 attachment, as described by Hrouda (1994).

RESULTS

Holes 883D and 884D

Major variations in magnetic mineral concentration are evident in logs of magnetic susceptibility for Holes 883D and 884D (Fig. 2). Because of the close proximity of these sites, we use the susceptibility variations as the basis for correlation between the two sites. With detailed peak matching, at least 70 points can be correlated between the two sites. In Figure 2, we show only some of the correlations in order to maintain clarity. This correlation demonstrates that sedimentation rates are similar at the two sites. An offset between correlative depths occurs at 7 mbsf in Hole 883D and at 6 mbsf in Hole 884D. This offset is probably the result of a break between Cores 1H and 2H in Hole 883D.

The largest peaks in the magnetic susceptibility records usually are clearly associated with tephra layers (Figs. 3, 4). The ratio of ARM/susceptibility is particularly useful for estimating relative variations in magnetic grain size (Banerjee et al., 1981). Low ratios indicate a predominance of relatively coarse-grained material, whereas high ratios indicate a predominance of relatively fine-grained material. Because the ARM measurements were made in a pass-through magnetometer system, the volume of material that contributes to the ARM is large and the ARM values are significantly higher than normally reported for such studies. The ARM values reported in this study have not been normalized to account for the volume of material sensed by the pickup coils in the magnetometer. In Figures 3 and 4,

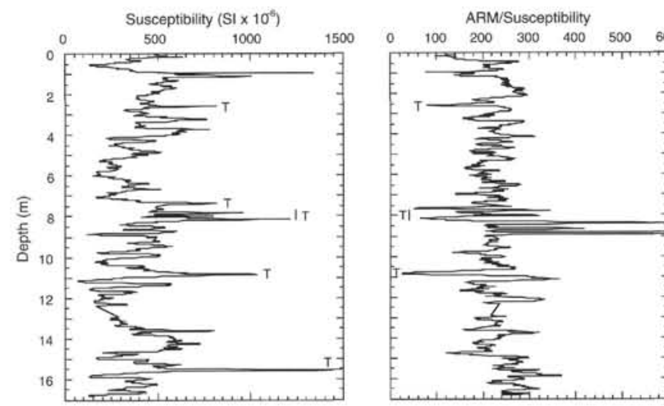

Figure 3. Magnetic susceptibility and ARM/susceptibility records for Hole 883D. Each T indicates a prominent tephra layer.

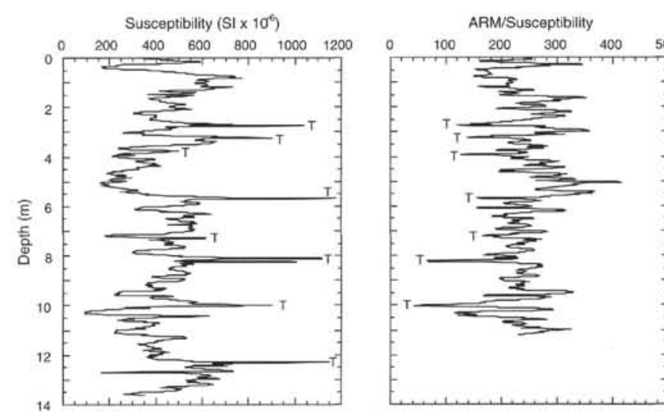

Figure 4. Magnetic susceptibility and ARM/susceptibility records for Hole 884D. Each T indicates a prominent tephra layer.

zones of high susceptibility and coarse grain size (as indicated by low values of ARM/χ) generally correlate, suggesting that the volcanic ashes are dominated by high concentrations of coarse magnetic particles, relative to the overlying and underlying sediments. Alternatively, if higher concentrations of paramagnetic minerals of volcanic origin are present in the tephra layers, they will contribute to the susceptibility and not the ARM and will cause a decrease in the ARM/χ ratio. However, the coarse textures of the tephra layers sup-

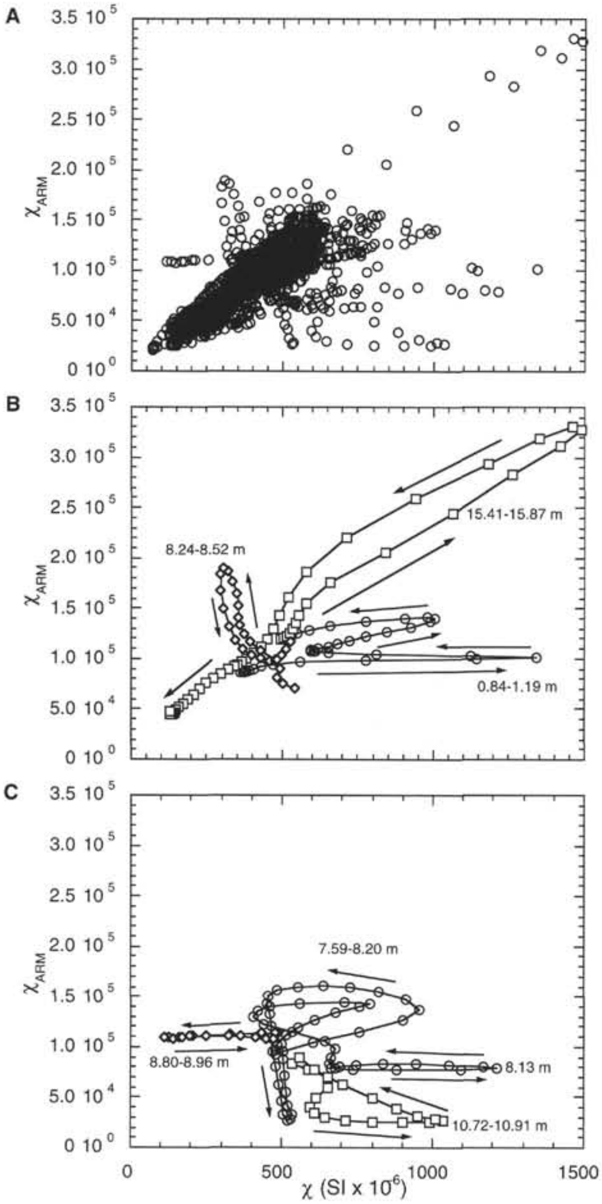

Figure 5. χ_{ARM} vs. χ plots (cf. Banerjee et al., 1981) for Hole 883D. **A.** All data. **B.** Grain-size/concentration excursions for depth intervals 0.84–1.19, 8.24–8.52, and 15.41–15.87 mbsf. **C.** Grain-size/concentration excursions for depth intervals 7.59–8.20, 8.80–8.96, and 10.72–10.91 mbsf. Values of χ_{ARM} have not been normalized to account for the volume of material sensed by the pickup coils in the magnetometer.

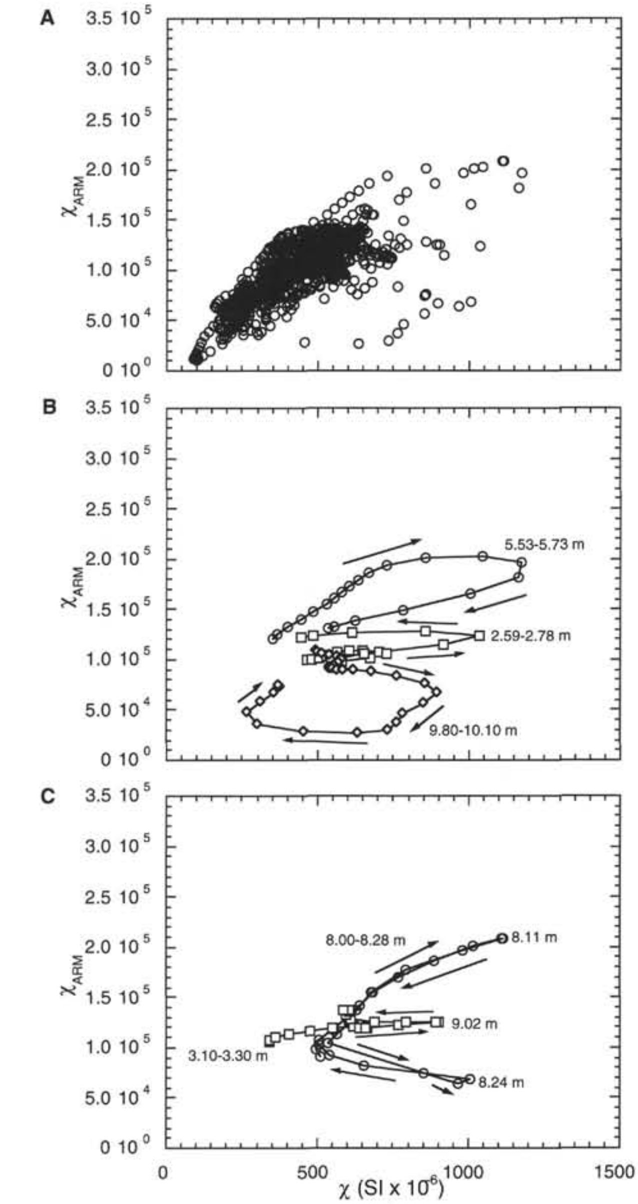

Figure 6. χ_{ARM} vs. χ plots (cf. Banerjee et al., 1981) for Hole 884D. **A.** All data. **B.** Grain-size/concentration excursions for depth intervals 2.59–2.78, 5.53–5.73, and 9.80–10.10 mbsf. **C.** Grain-size/concentration excursions for depth intervals 3.10–3.30 and 8.00–8.28 mbsf. Values of χ_{ARM} have not been normalized to account for the volume of material sensed by the pickup coils in the magnetometer.

port the interpretation that the tephra layers are dominated by coarser magnetic particles than the overlying and underlying sediments.

Plots of ARM vs. χ illustrate the degree of variability of the ARM/χ ratio. In these plots, changes in slope are caused by changes in relative grain size whereas changes along a line of constant slope are caused by variations in concentration of magnetic grains. Such plots indicate that the clay-rich sediment in Holes 883D and 884D is of rather uniform grain size (Figs. 5, 6), with points occurring within a restricted range of slopes. Distinct excursions toward relatively coarse grain sizes are evident, however, and are illustrated in greater detail in Figures 5B–C and 6B–C. Each of these "excursions" is stratigraphically restricted and is associated with a tephra layer delineated in Figures 3 and 4. In Hole 883D, a large counterclockwise-looping excursion, with slightly finer than average grain size and extremely large magnetic mineral concentrations, is associated with a tephra layer at 15.6 mbsf (Fig. 5B). This horizon is anomalous with respect to the other tephra horizons in Hole 883D because of its relatively fine magnetic grain size. A swing toward fine grain sizes is also evident in an interval between 8.80 and 8.96 mbsf (Fig. 5C); however, we have observed no visual differences in the nature of the sediment in this interval relative to those above and below. Furthermore, we see no large variation in grain-size-sensitive hysteresis parameters in this interval (Fig. 7). Comparison of the minimum and maximum values of ARM and χ within the large cluster of points on Figure 5 indicates that the concentration of magnetic grains varies within a factor of about 9.

In Hole 884D, no large excursions are evident toward finer grain sizes (Fig. 6). Several excursions toward coarser grain sizes, which are all associated with volcanic ash layers, are evident (Fig. 6B–C). Comparison of the minimum and maximum values of ARM and χ

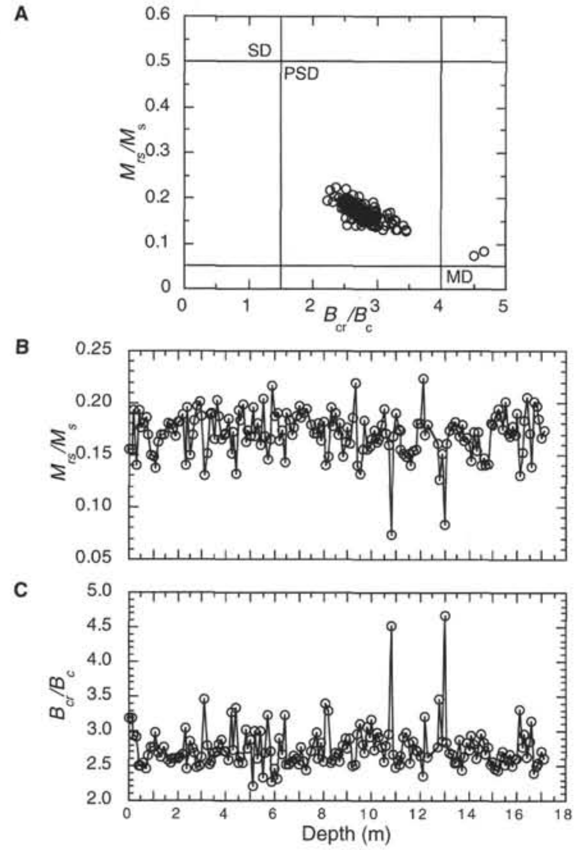

Figure 7. Grain-size variation for Hole 883D as indicated by high-field magnetic hysteresis data. **A.** Bivariate plot of M_{rs}/M_s vs. B_{cr}/B_c. Fields are shown for SD, PSD, and MD grains (after Day et al., 1977). SD = single domain, PSD = pseudosingle domain, and MD = multidomain. **B.** M_{rs}/M_s vs. depth. **C.** B_{cr}/B_c vs. depth.

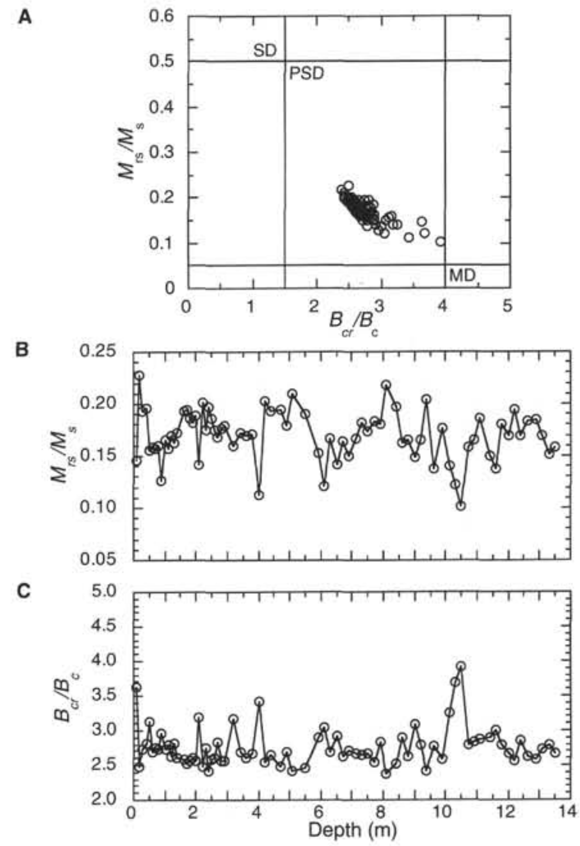

Figure 8. Grain-size variation for Hole 884D as indicated by high-field magnetic hysteresis data. **A.** Bivariate plot of M_{rs}/M_s vs. B_{cr}/B_c. Fields are shown for SD, PSD, and MD grains (after Day et al., 1977). **B.** M_{rs}/M_s vs. depth. **C.** B_{cr}/B_c vs. depth.

within the large cluster of points on Figure 6 indicates that the concentration of magnetic grains varies within a factor of about 10.

Four hysteresis parameters—M_{rs} (saturation remanence), M_s (saturation magnetization), B_{cr} (coercivity of remanence), and B_c (coercive force)—were routinely measured. The ratios M_{rs}/M_s and B_{cr}/B_c are highly sensitive to variations in grain size (Day et al., 1977). Detailed measurements of magnetic hysteresis properties indicate little variation in grain size of the bulk clay-rich sediment throughout Holes 883D and 884D (Figs. 7, 8). Clustering of hysteresis data within a restricted area of the PSD field (cf. Day et al., 1977) confirm the ARM/χ results that indicate minor variability in magnetic grain size. The hysteresis parameters M_{rs}/M_s and B_{cr}/B_c vary inversely with respect to each other with depth, as would be expected if the magnetic mineral assemblage is dominated by a single magnetic phase with a restricted range of grain sizes. In general, tephra layers were avoided when samples were collected for hysteresis analysis to permit characterization of the degree of variability of magnetic grain sizes within the bulk sediment. However, two samples in Hole 883D clearly contain anomalously coarse-grained material at about 10.8 and 13.0 mbsf (Fig. 7). The coarse interval at 10.8 mbsf is associated with a tephra layer (Fig. 3); however, we cannot identify any sedimentary feature responsible for the hysteresis measurement indicative of coarse-grained material at 13.0 mbsf. The range of variability of hysteresis parameters at Holes 883D and 884D is identical (Figs. 7, 8), indicating that magnetic particles at both sites vary within a small, and similar, range of sizes. The range of values shown on Figures 7 and 8 is consistent with a PSD magnetite grain size in the 1–15-µm range (King et al., 1983).

In all the sediments analyzed, measurements of the temperature dependence of magnetic susceptibility reveal a major decrease in susceptibility at high temperatures (above 550°C), which indicates a magnetic mineralogy dominated by magnetite (Fig. 9). This is supported by measurements of the S-ratio obtained on the samples subjected to hysteresis analysis, where $S = -IRM_{-0.3\,T}/SIRM_{1.2\,T}$ (King and Channell, 1991). $SIRM_{1.2\,T}$ is the saturation remanence measured at 1.2 T, whereas $IRM_{-0.3\,T}$ is the isothermal remanence measured by subjecting the sample to a reverse field after the sample has been subjected to the SIRM at 1.2 T. The maximum coercivity for magnetite is 0.3 T; therefore, the S-ratio should be close to 1 for a sample dominated by magnetite. The ratio decreases substantially for samples containing canted antiferromagnets, such as goethite and hematite, which saturate at maximum fields much higher than 0.3 T. Although the number of samples from which the S-ratio was determined is lower than those from which the full set of hysteresis samples was measured, the S-ratios are consistently close to 1, varying from 0.88 to 1.00 (Fig. 10). These measurements indicate that the clay-rich sediments are dominated by a ferrimagnetic mineral which, as indicated by thermomagnetic determinations, is most likely to be magnetite.

The results shown above clearly indicate that the mineral magnetic criteria for magnetic uniformity for relative paleointensity determinations (King et al., 1983; Tauxe, 1993) are satisfied by the clay-rich sediments from Holes 883D and 884D. Volcanic-ash-rich intervals of the core are not appropriate for such studies because of the coarse-grained nature of, and high concentrations of magnetic minerals in, these layers. Therefore, data from these intervals should not be included in estimates of relative paleointensity of the geomagnetic field.

Hole 887B

The magnetic properties of Hole 887B are entirely different from those obtained from Holes 883D and 884D. The susceptibility record

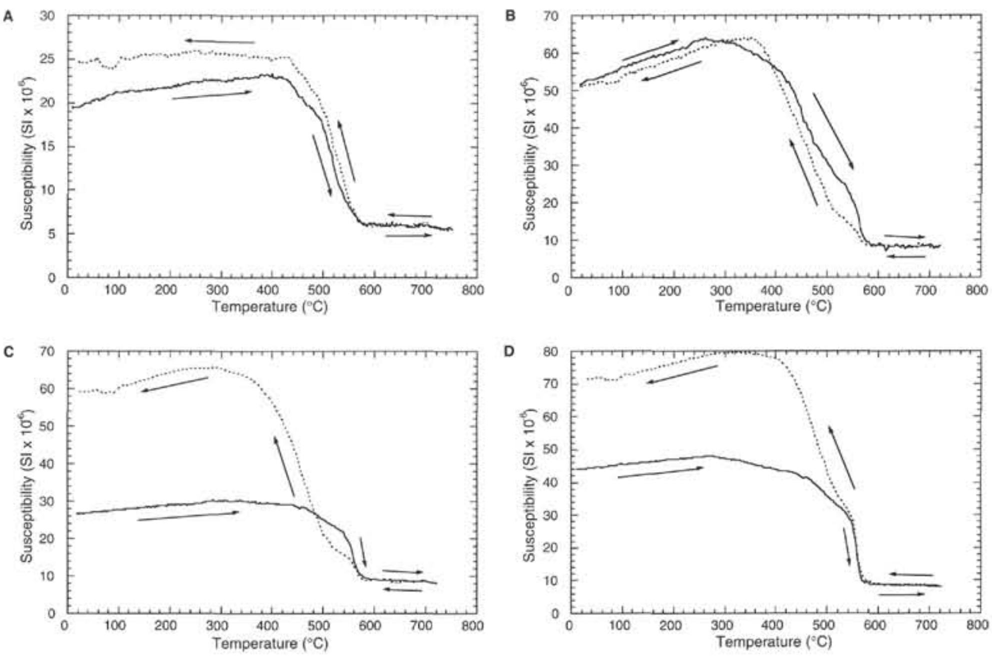

Figure 9. Temperature dependence of magnetic susceptibility for selected samples. **A.** Sample 145-883D-2H-2, 70 cm. **B.** Sample 145-884D-2H-3, 80 cm. **C.** Sample 145-887B-2H-5, 70 cm. **D.** Sample 145-887B-2H-6, 70 cm. Solid line indicates heating curve; dashed line indicates cooling curve. Major drop in susceptibility above 550°C indicates a magnetic mineral assemblage that is dominated by magnetite. Note difference in susceptibility scale for Figure 9A.

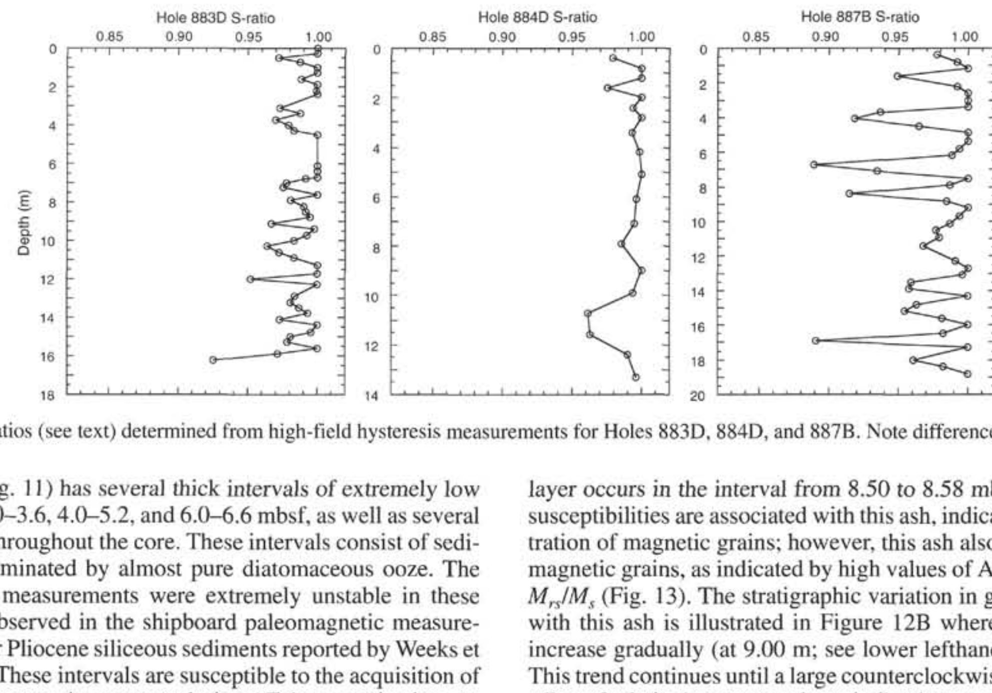

Figure 10. S-ratios (see text) determined from high-field hysteresis measurements for Holes 883D, 884D, and 887B. Note differences in depth scales.

from this hole (Fig. 11) has several thick intervals of extremely low values at about 3.0–3.6, 4.0–5.2, and 6.0–6.6 mbsf, as well as several thinner intervals throughout the core. These intervals consist of sediments that are dominated by almost pure diatomaceous ooze. The NRM and ARM measurements were extremely unstable in these regions, as was observed in the shipboard paleomagnetic measurements of the lower Pliocene siliceous sediments reported by Weeks et al. (this volume). These intervals are susceptible to the acquisition of spurious, random magnetic moments during AF demagnetization. As a result, extensive measurements of the remanence properties were not pursued.

Reliable ARMs were difficult to obtain; therefore, few ARM data are shown in Figure 11. Regardless of the paucity of ARM data, one can see that the diatom-rich intervals are dominated by the coarsest-grained magnetic fraction (i.e., coarser than those in the tephra layers). The sediments from Hole 887B are much more tephra-rich than those from Holes 883D and 884D, probably reflecting the proximity of this site to the Aleutian volcanic arc. The ARM/χ data and hysteresis properties of Hole 887B sediments (Figs. 12, 13) reflect variable grain sizes in different tephra layers. For example, a thick black ash layer occurs in the interval from 8.50 to 8.58 mbsf. Extremely high susceptibilities are associated with this ash, indicating a high concentration of magnetic grains; however, this ash also has extremely fine magnetic grains, as indicated by high values of ARM/χ (Fig. 12) and M_{rs}/M_s (Fig. 13). The stratigraphic variation in grain size associated with this ash is illustrated in Figure 12B where ARM/χ begins to increase gradually (at 9.00 m; see lower lefthand side of Fig. 12B). This trend continues until a large counterclockwise loop occurs along a line of relatively constant slope, indicating a major increase in concentration of magnetic grains. Immediately above the ash, grain sizes decrease until a uniformly coarse value is reached at about 8.25 mbsf. Other ashes, however, such as those at 1.36 and 3.75 mbsf, are much coarser grained (Figs. 11, 13). These fine- and coarse-grained contributions to the grain-size spectrum at Hole 887B represent a much broader range of grain sizes than is observed at Holes 883D and 884D (Fig. 13). Furthermore, two grain-size "envelopes" are apparent on Figure 12A. One of these corresponds to the same size range observed at Sites 883 and 884, whereas the other corresponds to much coarser grain sizes. This suggests a distinctly bimodal population of magnetic grains at Hole 887B. The coarser grains occur in the diatom-rich

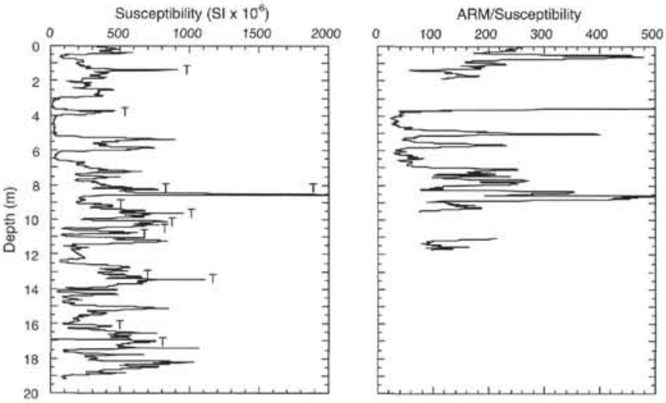

Figure 11. Magnetic susceptibility and ARM/susceptibility records for Hole 887B. Each T indicates a prominent tephra layer.

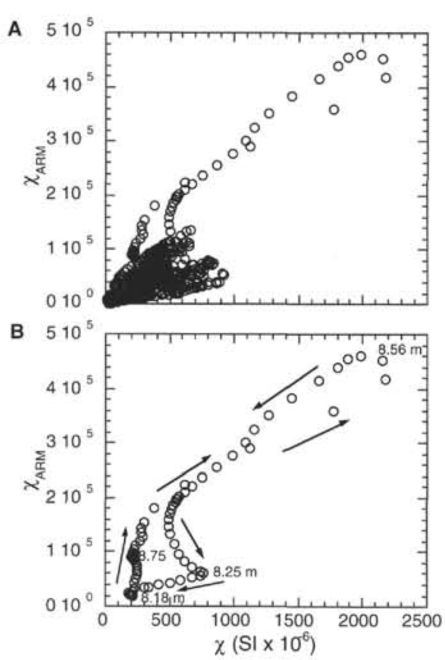

Figure 12. χ_{ARM} vs. χ plots (cf. Banerjee et al., 1981) for Hole 887B. **A.** All data. **B.** Grain-size/concentration excursion for depth interval from 8.18 to 9.00 mbsf. Values of χ_{ARM} have not been normalized to account for the volume of material sensed by the pickup coils in the magnetometer.

intervals. Geochemical analyses (Shipboard Scientific Party, 1993, p. 357) indicate that the diatom-rich zones are enriched in total organic carbon. This suggests that relatively high surface-water productivity and/or rapid burial gave rise to better preservation of organic material in the diatom-rich layers. It is well known that magnetite is highly soluble in organic acids and that reductive diagenesis and magnetite dissolution is common in rapidly deposited marine sediments (e.g., Karlin and Levi, 1983, 1985; Canfield and Berner, 1987; Karlin, 1990a, 1990b; Leslie et al., 1990a, 1990b; Roberts and Pillans, 1993; Roberts and Turner, 1993). It is probable, therefore, that the finest magnetic grains have been preferentially dissolved in the diatom-rich horizons in Hole 887B, giving rise to a magnetic mineral assemblage that is dominated by coarse grains.

Although the magnetic mineralogy appears to be dominated by magnetite (Figs. 9, 10), the large variation in concentration and grain size of magnetic grains, in conjunction with the widespread lack of a stable remanence, suggests that sediments from Hole 887B are inappropriate for studies of relative paleointensity of the geomagnetic field.

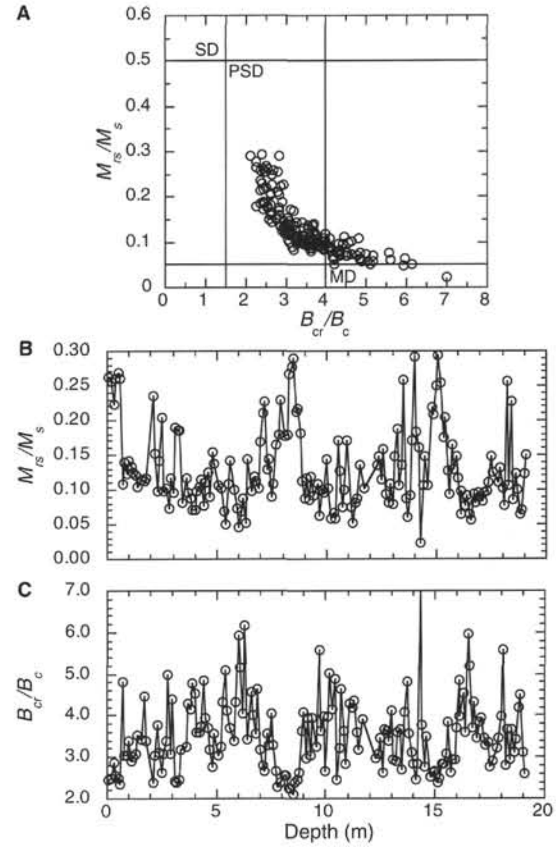

Figure 13. Grain-size variation for Hole 887B as indicated by high-field magnetic hysteresis data. **A.** Bivariate plot of M_{rs}/M_s vs. B_{cr}/B_c. Fields are shown for SD, PSD, and MD grains (after Day et al., 1977). **B.** M_{rs}/M_s vs. depth. **C.** B_{cr}/B_c vs. depth.

DISCUSSION AND CONCLUSIONS

Thermomagnetic and high-field hysteresis data suggest that the magnetic mineralogy at all sites studied is dominated by magnetite. At Holes 883D and 884D, variations in the concentrations and grain sizes of the magnetite appear to be sufficiently low to justify an attempt to obtain relative paleointensity data, except for regions of the sequences where coarse-grained and strongly magnetized volcanic ashes occur. Such studies of relative paleointensity are now in progress and will be reported elsewhere. The sediments at Hole 887B are not suitable for relative paleointensity determinations because of large variations in grain size and in the concentration of magnetite. Hole 887B contains intervals that are dominated by almost pure diatom ooze. The coarse magnetite grain size in these intervals is probably caused by dissolution of the finest grained magnetic particles during reductive diagenesis (e.g., Canfield and Berner, 1987). At Sites 883 and 884, the clay contents of the sediments are high enough that a strong, and apparently reliable, magnetic signal is obtained. The uniform grain size of the magnetic particles in the clay-rich sediment is consistent with sorting and transport by long-distance bottom currents, as has been suggested as the depositional mechanism at these sites (Scholl et al., 1977; Mammerickx, 1985). Significant variation between diatom-, clay-, and tephra-rich parts of the sequence preclude a straightforward interpretation of the magnetic susceptibility signal as a paleoclimate proxy, as is the case in depositional environments that are dominated by climatically modulated $CaCO_3$ fluctuations.

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