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A relative paleointensity record of the geomagnetic field since 1.6 Ma from the North Pacific

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A paleomagnetic study was conducted on a sediment core KR0310-PC1 taken from the central North Pacific in order to obtain a relative paleointensity record in the Matuyama chron from this region. The core reached to about 1.6 Ma. The age control is based on the correlation of the S ratio ($S_{-0.1\mathrm{T}}$) variations with a global oxygen-isotope stack. Isothermal remanent magnetization (IRM) was used as the normalizer of the relative paleointensity estimation; anhysteretic remanent magnetization (ARM) was not adopted because ARM is sensitive to magnetostatic interaction among magnetic particles, which is evidenced in these sediments by an inverse correlation between the ratio of ARM to saturation IRM (SIRM) and SIRM without significant magnetic grain-size changes. For the last 350 kyrs, the record of core NGC65, which was obtained at practically the same site as KR0310-PC1 and covers the Brunhes chron (Yamazaki, 1999), was incorporated because the upper part of KR0310-PC1 was physically disturbed. In the record of NGC65/KR0310-PC1, the average paleointensity in the late Matuyama chron is not lower than that during the Brunhes chron, which does not support the conclusion of Valet *et al.* (2005) based on their Sint-2000 stack. A spectral analysis on the NGC65/KR0310-PC1 paleointensity record shows a power at the $\sim$100 kyr eccentricity period. The relative paleointensity and magnetic properties of NGC65/KR0310-PC1 were compared with those of MD982185 from the western equatorial Pacific (Yamazaki and Oda, 2002, 2005). The two sites belong to different oceanographic regimes. Coherent variations in the relative paleointensity despite incoherent changes in the magnetic properties suggest that rock-magnetic contamination to the relative paleointensity is small, if any, and the $\sim$100 kyr period in the relative paleointensity records would reflect the geomagnetic field behavior.

Key words: Paleomagnetism, paleointensity, North Pacific, orbital modulation, normalizer, magnetostatic interaction.

1. Introduction

Information on the intensity of the past geomagnetic field (paleointensity) is indispensable for understanding the mechanism of the geodynamo. Marine sediments would be the only media that can preserve continuous records of paleointensity variations. A rapid progress in paleointensity estimation using marine sediment cores in 1990s resulted in the establishment of the global paleointensity stack during the Brunhes chron: the Sint-800 of Guyodo and Valet (1999). Efforts have been made to extend the paleointensity record to older periods, and an increasing number of records in the Matuyama chron have been published (Valet and Meynadier, 1993; Meynadier *et al.*, 1994; Sato *et al.*, 1998; Kok and Tauxe, 1999; Channell *et al.*, 2002; Yamazaki and Oda, 2002, 2005; Carcaillet *et al.*, 2003; Horng *et al.*, 2003). However, agreement among the records is not very good (Yamazaki and Oda, 2005). Furthermore, geographical distribution of the records is still restricted; in the Pacific, for example, the published records are restricted in low latitudes. Hence further accumulation of high quality data is yet needed. In these circumstances, Valet *et al.* (2005) compiled existing (but not all available) relative paleointensity records that reached to the Matuyama chron, and presented a stacked curve during the last two million years called the Sint-2000. It would be necessary to evaluate accuracy and limitations of this stack.

Arguments on possible relationship between the geomagnetic field and climate have started in 1970s (e.g. Wollin *et al.*, 1971; Rampino, 1979), but little convincing evidence was obtained at that time. The rapid progress in relative paleointensity studies revived the orbital modulation issue in the late 1990s. Channell *et al.* (1998) proposed occurrence of the $\sim$40 kyr obliquity frequency from a power spectrum analysis of their relative paleointensity records during the Brunhes chron obtained from ODP Sites 983 and 984 in the North Atlantic. They considered it as geomagnetic field behavior from the observations that no power exists at the $\sim$40 kyr period in bulk magnetic properties and there is no coherence between the relative intensity and the normalizer (IRM), percent carbonate, and a magnetic grain-size proxy at the $\sim$40 kyr period. Yamazaki (1999) instead proposed possible presence of the $\sim$100 kyr eccentricity frequency in his relative paleointensity records from the North Pacific based on the same logic as that of Channell *et al.* (1998).

Thouveny *et al.* (2004) reported the ~100 kyr period in paleointensity records during the last 400 kyr from Portuguese Margin sediments, North Atlantic. Possible occurrence of the ~100 kyr period in paleointensity would not be limited in the Brunhes chron; Yamazaki and Oda (2005) found a significant power at ~100 kyr period in paleointensity records from 0.8 to 3.0 Ma. Recently, Fuller (2006) suggested presence of the ~41 kyr obliquity signal in paleointensity as well as excursions and reversals.

Strong arguments against the orbital modulation of the geomagnetic field come from possible lithological contamination to sedimentary paleointensity records. Guyodo *et al.* (2000) performed a wavelet analysis on the records of paleointensity and magnetic properties from ODP Site 983, which was the same dataset as those from which Channell *et al.* (1998) and Channell and Kleiven (2000) proposed the ~41 kyr periodicity. They found that relative paleointensity has coherency with the normalizer and a magnetic grain-size proxy, the ratio of anhysteretic remanent magnetization (ARM) to magnetic susceptibility (k), and concluded that the orbital frequencies in paleointensity records may be the expression of lithological variations. However, this cannot exclude the possibility of the orbital modulation, because relative paleointensity and magnetic properties of sediments can have coherency if variations of the orbital parameters affect both the geomagnetic field and a depositional environment. To solve this problem, it is important to examine phase relationships in coherency analyses. Patterns of lithological and magnetic property variations induced by paleoclimatic changes may vary place to place; for example, magnetic grain size would increase in a certain period of time in some areas, but in other areas it would decrease in the same period of time. Paleointensity, on the other hand, should be globally synchronous when considering the dipole component. Thus, even when paleointensity and magnetic property variations have coherency, lithological contamination is suggested to be minor if phase angles between the two from various places differ significantly. Using this strategy, Yokoyama and Yamazaki (2000) suggested that the ~100 kyr period would be inherent to the geomagnetic field; five relative paleointensity records from the Pacific during the Brunhes chron showed a good coincidence in this time scale despite significant phase differences in magnetic properties.

In this paper, we present a relative paleointensity record since 1.6 Ma from the central North Pacific. This is the first paleointensity record in the Matuyama chron from this region. We point out some features in our record are significantly different from the Sint-2000 stack. Next, we show possible occurrence of the ~100 kyr eccentricity frequency in our record, and discuss the issue of lithological contamination by comparing with a record from the equatorial Pacific, which belongs to a different oceanographic regime and has different magnetic properties.

2. Sampling and Measurements

A piston core KR0310-PC1 of about 19 m long was taken at 35°14.65′N, 174°59.90′E in the central North Pacific near the Hess Rise (Fig. 1). The water depth of the site, 4951 m, is below the carbonate compensation depth (CCD), and the sediments are comprised of homogeneous siliceous clay with dull yellowish brown to dark brown color. The position of the coring site is almost the same as that of a gravity core NGC65 (Yamazaki, 1999). Because a high-quality relative paleointensity record during the last 650 kyr was recovered from the short gravity core of about 7 m long, we intended to extend further the relative paleointensity record by taking a longer core at the same place.

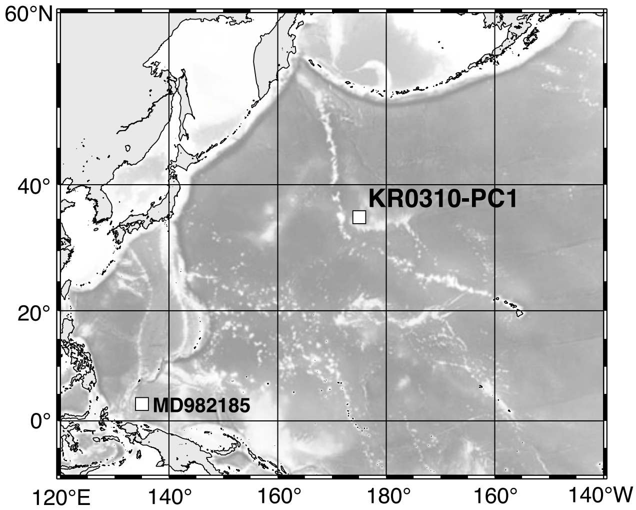

Fig. 1. Location of piston cores.

The upper about 5 m of the sediment core was physically disturbed during coring, which was recognized from the deformation of core liners by suction of a piston. Below this, no evidence of disturbance was observed by visual inspection of half-split core sections onboard. Discrete samples for paleomagnetic and rock-magnetic measurements were taken using plastic cubes of 7 cm^3. We collected 822 samples in total sequentially with no gap between them. The sampling was done onboard R/V Kairei in 2003 immediately after the core recovery, and the samples were tightly sealed and refrigerated.

Magnetic susceptibility (k) of discrete samples was measured using a KappaBridge KLY-3S. Remanent magnetization was measured with a cryogenic magnetometer system (2G Enterprises model 760R) with an in-line alternating-field (AF) demagnetizer, which sits in a magnetically shielded room of the Geological Survey of Japan, AIST. Stepwise AF demagnetization up to 80 mT was performed on natural remanent magnetization (NRM). Anhysteretic remanent magnetization (ARM) was imparted by superimposing a DC biasing field of 0.1 mT on a smoothly decreasing AF with a peak field of 80 mT. Isothermal remanent magnetization (IRM) was given by a pulse magnetizer (2G model 660). IRM of discrete samples was measured using a spinner magnetometer (Natsuhara-Giken SMM-85). First, IRM was imparted at 2.5 T, which is regarded as saturation IRM (SIRM) in this paper. Then, IRMs of 0.1 T and 0.3 T were successively imparted in the direction opposite to the initial IRM. S ratios ($S_{-0.1\mathrm{T}}$ and $S_{-0.3\mathrm{T}}$) were calculated according to the definition of Bloemendal *et al.* (1992). Magnetic hysteresis curves were measured on dried specimens using an alternating-force gradient magnetome-

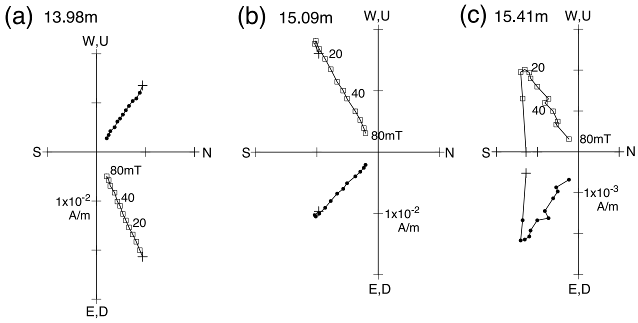

Fig. 2. Examples of stepwise alternating-field demagnetization data. Open (solid) circles are projection of vector end-points on the vertical (horizontal) plane. Cores are not oriented azimuthally.

ter (Princeton Measurements MicroMag 2900); the measurements were done on selected samples of about 20 cm intervals.

3. Results

3.1 Natural remanent magnetization

Stepwise AF demagnetization showed that remanent magnetization of most samples consists of a single component except for the first few demagnetization steps. Magnetic overprint of probably viscous remanent magnetization (VRM) origin could be removed by AF up to 20 mT in general (Fig. 2). Directions of NRM were determined by applying the principal component analysis (PCA) (Kirschvink, 1980). On each sample, we selected a linear portion on the orthogonal plot of the AF demagnetization for PCA, which ranges from 10 or 15 mT to 70 or 80 mT in general, but sometimes the interval is a little narrower. The maximum angular deviations (MADs) are very small in general; 90% of the samples are less than 2° (Fig. 2(a) and (b)). Exceptions are samples with very low remanent magnetization intensities at polarity transitions and possible excursions; an example is shown in Fig. 2(c).

NRM directions (Fig. 3(d)) show that the Brunhes-Matuyama transition occurs at 10.76 m in depth, and the bottom of the core reached between the Jaramillo and Olduvai subchrons. An average inclination calculated from the inclination-only data assuming a Fisher distribution (McFadden and Reid, 1982) is 54.0° (α_{95}=0.6°) during the Brunhes chron except for the uppermost 5 m, and it is −54.5° (α_{95}=1.2°) below the Brunhes/Matuyama transition excluding the Jaramillo subchron. They agree exactly with the inclination expected from the geocentric axial dipole (GAD) field for the site latitude, 54.5°, which indicates that secondary magnetization is not significant. The physical disturbance of the upper 5 m of the core is recognized from incoherent declination swings reflecting severe core-twisting and anomalous inclination changes with steeper values than expected from GAD, and we discarded the data from the upper 5 m. A gradual change of about 160° in declination by core-twisting is observed between 5 and 9 m in depth (Fig. 3(d)), although no disturbance was recognized by visual inspection of the core. The twist within each discrete sample of approximately 2 cm long is less than 1°, and thus we consider it would not have affected NRM intensities. Such gradual core-twisting sometimes happens during coring by a piston corer.

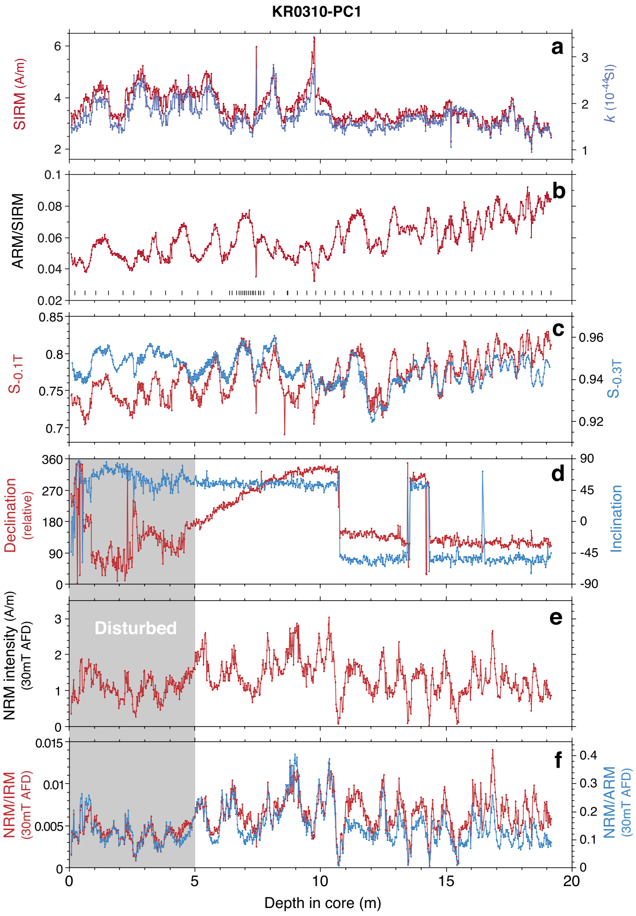

Fig. 3. Remanent magnetization and magnetic properties of core KR0310-PC1. (a) Magnetic concentration represented by saturation isothermal remanent magnetization (SIRM) (red) and magnetic susceptibility (k) (blue). (b) Ratio of anhysteretic remanent magnetization (ARM) to SIRM. Ticks indicate the positions of the specimens for magnetic hysteresis measurements in Fig. 5. (c) S-ratios (red: $S_{-0.1\text{T}}$, blue: $S_{-0.3\text{T}}$). (d) Natural remanent magnetization (NRM) direction (red: relative declination, blue: inclination). The upper 5 m of the core was physically disturbed at coring (shaded in gray). (e) NRM intensity after AF demagnetization at 30 mT. (f) NRM intensity normalized by IRM (red) and ARM (blue). NRM, IRM, and ARM were AF demagnetized at 30 mT.

3.2 Rock magnetism

Magnetic concentration variations represented by k and SIRM are within about three times (Fig. 3(a)), which meets the criteria of reliable relative paleointensity estimation (Tauxe, 1993). The variation is very small in the Matuyama chron, but it becomes larger in the Brunhes chron. There is a decreasing trend with depth.

The ratio of ARM to SIRM shows an increasing trend with depth, which is obvious below about 10 m (Fig. 3(b)). In addition, fluctuations on the order of tens of centimeters to one meter are superimposed on the trend. The ratio of ARM to SIRM or k is often interpreted as a proxy of magnetic grain size (Banerjee *et al.*, 1981; Evans and Heller, 2003). Alternatively, the ratio reflects variations in the strength of magnetostatic interaction among magnetic minerals when magnetic concentration changes, because ARM is very sensitive to magnetostatic interaction

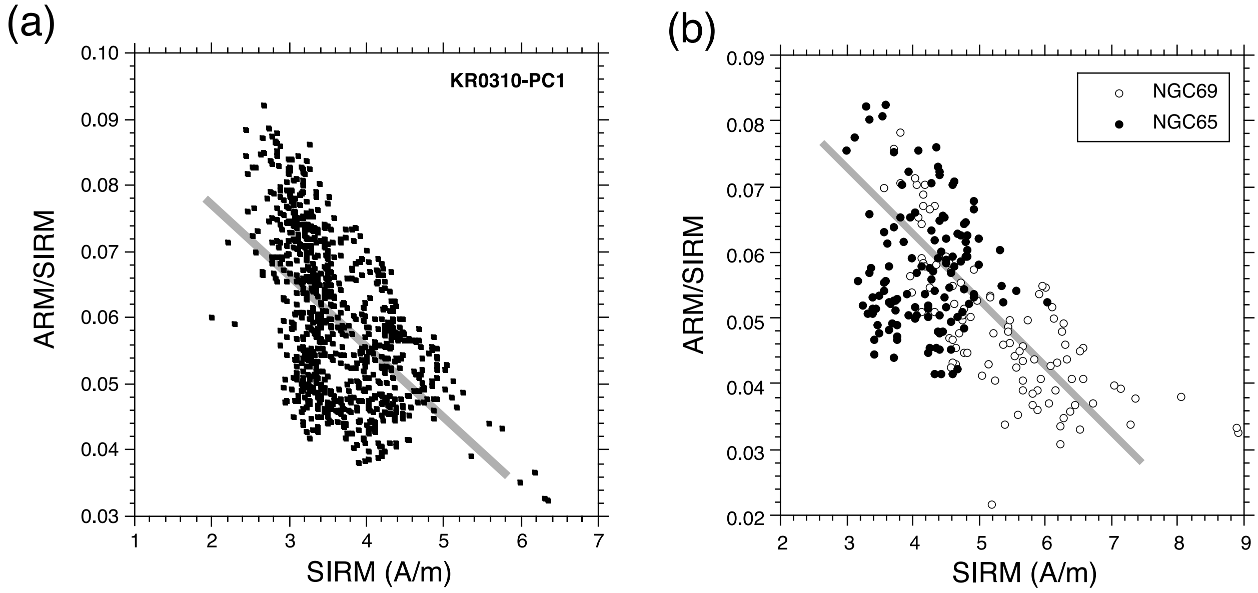

Fig. 4. Relationship between the ratio of ARM to SIRM and SIRM. (a) Core KR0310-PC1. (b) Previously reported cores from the central North Pacific (Yamazaki, 1999): core NGC65 (at the same position as KR0310-PC1 but shorter in length) and NGC69 (at 40°0′N, 175°0′E).

(Sugiura, 1979; Yamazaki and Ioka, 1997a). In this core, the ARM/SIRM ratios are roughly correlated inversely with SIRM (Fig. 4(a)), which is expected from the opposite trend with depth between ARM/SIRM and SIRM (Figs. 3(a) and (b)). This can be interpreted as that an increase of magnetic concentration, which is represented by an increase in SIRM, causes a decrease in the acquisition efficiency of ARM by magnetostatic interaction. The same relationship was observed in the previous cores in the North Pacific (Fig. 4(b)), and interpreted as such (Yamazaki, 1999; Yamamoto *et al.*, 2007). Results of magnetic hysteresis measurements do not show significant variations in magnetic grain size. In a plot of the hysteresis parameters, the ratio of saturation remanence to saturation magnetization (M_{rs}/M_s) versus the ratio of coercivity of remanence to coercivity (B_{cr}/B_c) (Day *et al.*, 1977), most data are well clustered in a pseudo-single-domain (PSD) range (Fig. 5). Hence we conclude that in the present core the ARM/SIRM ratios are dominantly controlled by magnetostatic interaction, in particular for the downward increasing trend. As for fluctuations on the order of tens of centimeters to one meter, the effect of grain size changes seems to be superimposed; variations of ARM/SIRM with no or positive correlation to SIRM, for example around 1 m and from 10.5 to 16 m in depth, can be interpreted as magnetic grain size changes, whereas the variations with inverse correlation to SIRM such as from 8.5 to 10.5 m and from 17 to 19 m would be dominated by magnetostatic interaction.

The core shows high values of $S_{-0.3\mathrm{T}}$, ranging from 0.92 to 0.96 (Fig. 3(c)). This indicates that remanent magnetization is mostly carried by low-coercivity magnetic minerals like magnetite/maghemite. The two S ratios with different back-field intensities, $S_{-0.1\mathrm{T}}$ and $S_{-0.3\mathrm{T}}$, show synchronous peaks and troughs except for between 2 and 5 m in depth. $S_{-0.1\mathrm{T}}$ gradually increases with depth, whereas $S_{-0.3\mathrm{T}}$ does not show such a trend. The variations of the S ratios, in particular $S_{-0.1\mathrm{T}}$, are coeval with that of the ARM/SIRM ratio.

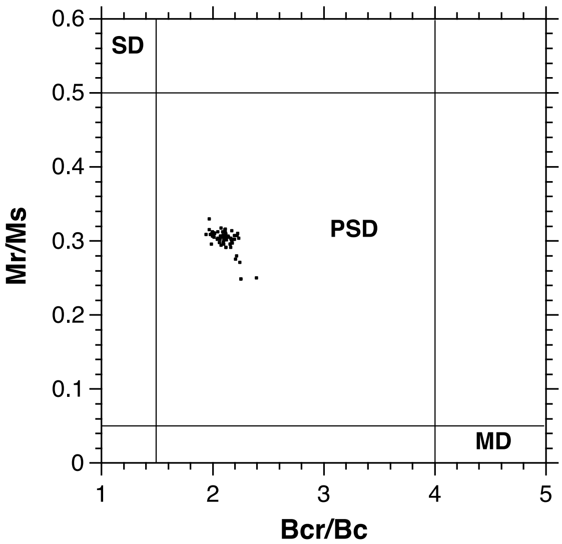

Fig. 5. Plots of hysteresis parameters. Ratio of coercivity of remanence (B_{cr}) to coercivity (B_c) vs. ratio of saturation magnetization (M_s) to saturation remanence (M_r). Depths in the core of the specimens are shown in Fig. 3(b).

Yamazaki and Ioka (1997b) showed that in the pelagic clay province in the middle latitude of the North Pacific, $S_{-0.3\mathrm{T}}$ can be used as a proxy for Asian eolian dust. In their model, the eolian dust, which is transported from arid regions of the Asian continent by westerlies, has a higher concentration of high-coercivity minerals like hematite than other sources of magnetic minerals, and increased inputs of eolian dust during glacial periods caused decreases in $S_{-0.3\mathrm{T}}$. A similar explanation would apply for the variations in $S_{-0.1\mathrm{T}}$; the eolian component would have a higher proportion of maghemite as well as hematite, and a higher proportion of maghemite could cause an increased average coercivity affecting $S_{-0.1\mathrm{T}}$ (Moskowitz, 1980). Iron sulfides are not candidates for the magnetic minerals with coercivity between 0.1 T and 0.3 T

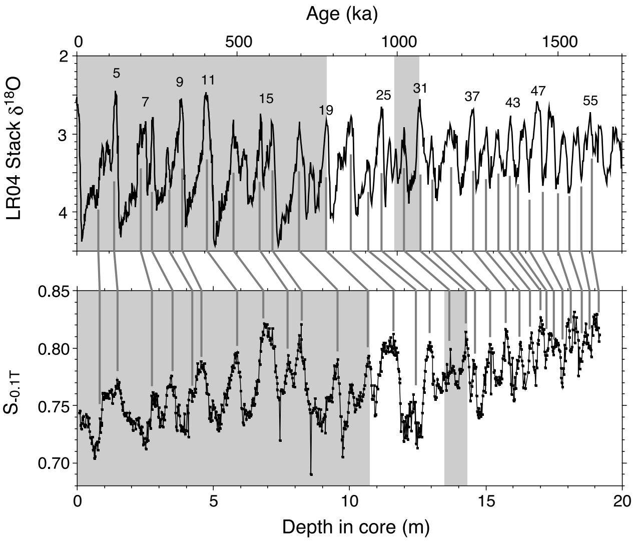

Fig. 6. Age control of KR0310-PC1. Variations of S ratio ($S_{-0.1\mathrm{T}}$) are correlated to the global oxygen isotope (δ^{18}O) stack LR04 (Lisiecki and Raymo, 2005) at the horizons indicated by tie lines. Magnetic polarity is shown by graying normal polarity zones. Ages of the polarity boundaries in the panel of the δ^{18}O curve are after Cande and Kent (1995), and observed depths of polarity boundaries are displayed in the panel of S ratio. The Marine Isotope Stage (MIS) numbers are attached to the δ^{18}O curve.

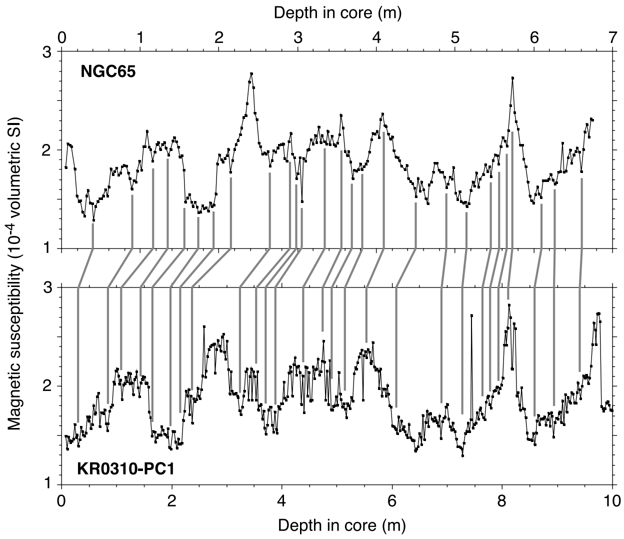

Fig. 7. Intercore correlation using magnetic susceptibility between piston-core KR0310-PC1 and gravity-core NGC65 (Yamazaki, 1999) taken at the same site.

in this core, because the sediments are in an oxidized condition. It is not clear the cause of the coeval variations of the ARM/SIRM ratio and S ratio. Here we present two possible scenarios. In the first model, we infer that the eolian component contains a larger proportion of titanomagnetites with ilmenite lamellae of volcanic or plutonic origin, which has stronger magnetic interaction. Then increased eolian inputs during glacial periods can cause lower ARM/SIRM and S ratios. In the second model, the eolian component is assumed to have a larger average grain size. This also can cause lower ARM/SIRM and S ratios in glacials.

3.3 Age estimation

The core cannot be dated directly by the oxygen isotope (δ^{18}O) stratigraphy because the coring site is below the CCD and little carbonates remain. Instead, we assign ages by correlating S-ratio ($S_{-0.1\mathrm{T}}$) variations to the δ^{18}O stratigraphy (Fig. 6), since it is considered that in this area the variations of the S ratios reflect glacial-interglacial changes as mentioned above. The global δ^{18}O stack LR04 (Lisiecki and Raymo, 2005) was used as a target curve. We first tied magnetic polarity boundaries to Marine Isotope Stages (MISs) using reported correspondence between them: the B/M boundary at MIS 19, the Jaramillo subchron from MISs 28 to 31. Then, we tried to correlate the S ratio variations to the δ^{18}O curve between the polarity boundaries. We assigned interglacial periods (peaks in the figures of the δ^{18}O curves) to highs in the S ratios, that is, higher proportion of low coercivity magnetic minerals like magnetite probably caused by smaller eolian input. We assumed no time lag between the S ratios and the δ^{18}O curve. The variations in S ratio closely resemble the target δ^{18}O curve, and hence assignment of MISs was straightforward. Uncertainty in the age control of these cores is considered to be within one obliquity cycle ($\sim$40 kyr) in general.

3.4 Integration with Core NGC65

As the upper 5 m ($\sim$350 kyrs) of KR0310-PC1 was physically disturbed, we make up for this part using the data of the core NGC65 (Yamazaki, 1999), which was taken from virtually the same site. Using magnetic susceptibility, the two cores can be correlated with each other; the variation patterns are almost identical (Fig. 7). Based on the tie points presented in the figure, depths of NGC65 were converted to those of KR0310-PC1, and then transformed to ages.

From the susceptibility correlation, it is revealed that the uppermost about 20 cm was not taken when the piston corer hit the bottom. A further striking observation is the differences in depths of the equivalent horizons in the two cores; the recovered sediments of KR0310-PC1 are about 50% thicker than those of NGC65 (Figs. 7 and 8). One may image that in piston cores sediments are stretched by suction of a piston, whereas sediments are compressed in gravity cores. However, the values of volumetric susceptibility are almost identical between the two cores (Fig. 7). This indicates that neither stretch nor compaction, which accompany volumetric changes, occurred in these cores. Instead, these phenomena could be explained by "over-sampling" of a piston core and/or "under-sampling" of a gravity core, which was presented by Skinner and McCave (2003) and Széréméta *et al.* (2004). According to their model and observation, a plastic response of sediments to a vertical stress could cause thickening (thinning) of sedimentary layers resulting from incorporation of excessive (deficient) sedimentary material into a corer. An almost constant slope on the plot between the depths of KR0310-PC1 and those of NGC65 (Fig. 8) suggests that "over-sampling" and/or "under-sampling" takes place continuously but not intermittently. This is probably due to the homogeneous lithology of the cores. The age-depth curve of KR0310-PC1 in Fig. 8(b) shows an apparent upward increase of the sedimentation rate with an average of 12 m/m.y., but the increase may have been caused by over-sampling of the piston core.

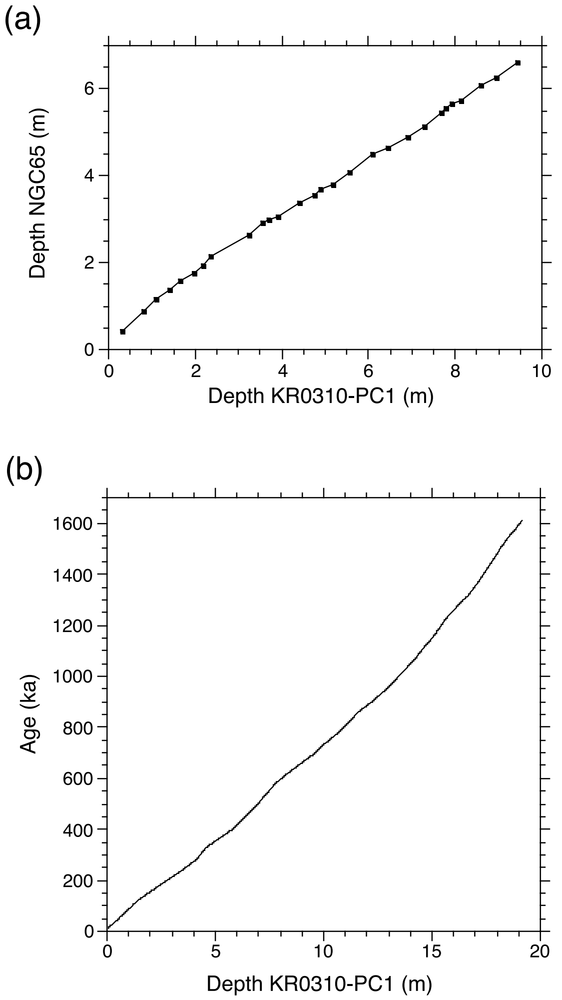

Fig. 8. (a) Comparison of depths in piston-core KR0310-PC1 with corresponding depths in gravity-core NGC65. (b) Depth versus age of KR0310-PC1.

3.5 Relative paleointensity

The present core has magnetic properties suitable for relative paleointensity estimation. The magnetic concentration change is within about three times. The ratio of ARM to SIRM shows some variations, but this mainly reflects changes in the strength of magnetostatic interaction, and magnetic grain-size change is considered to be relatively small. The remanent magnetization is dominantly carried by a low-coercivity fraction, probably magnetite. The fact that the core NGC65 yielded a reliable relative paleointensity record that is consistent with the global stack Sint-800 of Guyodo and Valet (1999) (Yamazaki, 1999; Yamazaki and Oda, 2004) supports appropriateness of the sediments at this site for relative paleointensity estimation. We chose IRM as a normalizer because ARM is sensitive to the magnetostatic interaction, and NRM intensity after AF demagnetization at 30 mT was divided by IRM acquired at 2.5 T and partially demagnetized by AF at 30 mT. Tauxe and Wu (1990) proposed a way to evaluate the normalization process; if the power spectrum of the normalized intensity is coherent with the normalizer, the normalization is inappropriate. On the NGC65/KR0310-PC1 composite record, coherence between the normalized intensity and the normalizer is larger in ARM normalization than in IRM normalization, and the former is significant at a 95% level in many frequency ranges including ∼100 kyrs (Fig. 9). This fact supports the choice of IRM as a normalizer in this core. As variations of SIRM intensity are smaller than those of NRM, the variation pattern of NRM intensity after the normalization does not differ greatly from that before the normalization (Figs. 3(e) and (f)).

The KR0310-PC1/NGC65 composite record of relative paleointensity during the last 1600 kyrs is presented in Fig. 10. This is the first relative paleointensity record of the Matuyama chron in the North Pacific. An intensity low at 1.2 Ma is estimated to be associated with the Cobb Mountain subchron. This subchron is considered to have occurred at MIS 36 (Channell *et al.*, 2002; Horng *et al.*, 2002), and the depth of the intensity low corresponds to MIS 36 based on our age model.

4. Discussion

4.1 Choice of normalizer

In relative paleointensity estimations, proper normalization of the efficiency for depositional remanent magnetization (DRM) acquisition of sediments is essential. Here we discuss more about our preference of IRM to ARM as the normalizer. Because the difference in the variations between ARM and IRM intensities is smaller than those of NRM intensity in the present core, a general pattern of the normalized intensity records such as a succession of peaks and troughs agrees between the normalization by ARM and IRM (Fig. 3(f)). However, the amplitudes of the peaks and troughs do not coincide with each other. In particular, there is a conspicuous difference below the Brunhes-Matuyama transition; the record normalized by ARM shows a lower intensity in average in the Matuyama chron than in the Brunhes chron, but the record normalized by IRM does not indicate such a tendency. Hence the choice of the normalizer is essential for discussing long-term paleointensity intensity changes such as the possible weak average intensity in the late Matuyama chron suggested by Valet *et al.* (2005).

Pioneering works of sedimentary paleointensity proposed to use ARM as the normalizer because coercivity spectra of ARM are closer to those of NRM than IRM in their sediments (Levi and Banerjee, 1976; Kent and Opdyke, 1977; King *et al.*, 1983). Many paleointensity studies since then followed basically these works, adopting ARM as the normalizer (e.g. Valet and Meynadier, 1993; Yamazaki and Ioka, 1994; Roberts *et al.*, 1997; Laj *et al.*, 2000). However, Channell *et al.* (1998, 2004) and Channell and Kleiven (2000) preferred IRM because they found in their sediments that NRM/IRM has smaller dependence on the strength of AF demagnetization fields than NRM/ARM. This observation indicates that coercivity spectra of NRM are not necessarily closer to those of ARM than IRM.

A drawback of ARM as the normalizer is that ARM is sensitive to magnetostatic interaction among grains, as discussed above. Effect of magnetostatic interaction on DRM acquisition has not yet been understood well, but if it is less sensitive than on ARM and similar to IRM, normalization by ARM may overcompensate magnetic concentration changes and would cause significant coherency be-

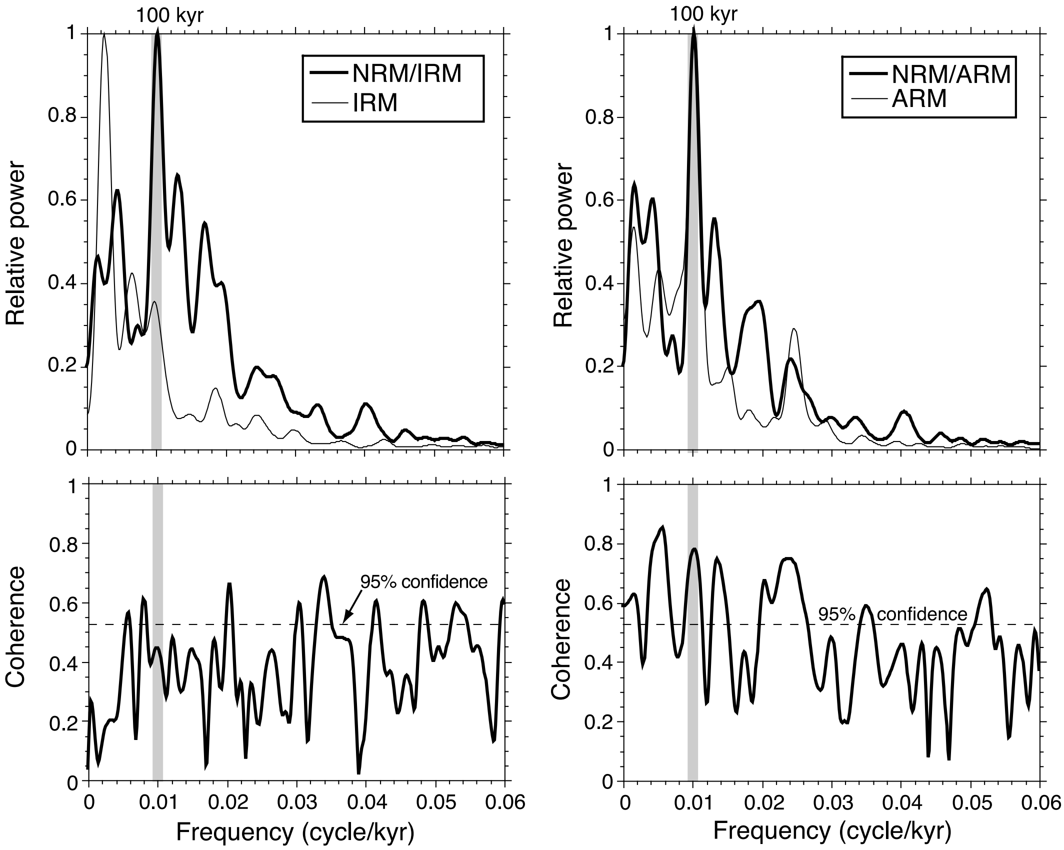

Fig. 9. Power spectra of normalized intensity (left: NRM/IRM, right: NRM/ARM) and normalizer, and squared coherence between the normalized intensity and the normalizer. The AnalySeries software (Paillard *et al.*, 1996) and the Blackman-Tukey method with a Bartlett window were used.

tween the normalized intensity and the normalizer, as is the case of the present core. Considerable number of papers reported that normalization by IRM showed smaller coherence or correlation between the normalized intensity and the normalizer than by ARM (Tauxe and Shackleton, 1994; Lehman *et al.*, 1996; Channell *et al.*, 1998; Williams *et al.*, 1998; Yamazaki, 1999; St-Onge *et al.*, 2003), although some presented the opposite result: ARM normalization resulted in smaller coherency with the normalizer (Guyodo *et al.*, 2001; Gogorza *et al.*, 2004). These examples suggest that being normalized by ARM the overcompensation due to magnetostatic interaction may have occurred in sediments of wide areas. Schwartz *et al.* (1998) selected IRM as the normalizer partly because NRM intensity normalized by ARM seems to correlate with magnetic grain size (ARM/k), and they attributed it to the high sensitivity of ARM to magnetic grain-size variation. Alternatively, magnetostatic interaction may be responsible because ARM/k anti-correlates with k in their sediments.

4.2 Difference from Sint-2000 stack

Valet *et al.* (2005) discussed intensity variations of the dipole field during the last two million years based on their compilation of ten paleointensity records (the Sint-2000 stack). They postulated that the average paleointensity during the late Matuyama chron is smaller than that of the Brunhes chron: the average virtual axial dipole moment (VADM) of 7.5×10^{22} A m^2 during the Brunhes and 5.3×10^{22} A m^2 between 0.78 and 1.18 Ma. In Fig. 11, we compare the Sint-2000 stack with the NGC65/KR0310-PC1 record and that of MD982185 in the western equatorial Pacific (Yamazaki and Oda, 2002, 2005), which is not included in the compilation of Valet *et al.* (2005). In this figure, the average paleointensity during the Brunhes chron is adjusted to unity for all three records. The patterns of relative paleointensity variations agree well with each other within the uncertainty of age. However, only the Sint-2000 stack shows lower paleointensity during the late Matuyama chron. On the two records from the Pacific, the average in this period does not differ significantly from that in the Brunhes chron; In the NGC65/KR0310-PC1 record, the average relative paleointensity for 400 kyrs before the Brunhes/Matuyama transition is 0.94. One possible cause of this discrepancy, the apparently lower relative paleointensity in the late Matuyama chron, may be insufficient removal of an overprint of viscous remanent magnetization acquired during the Brunhes chron. Another possibility would be the

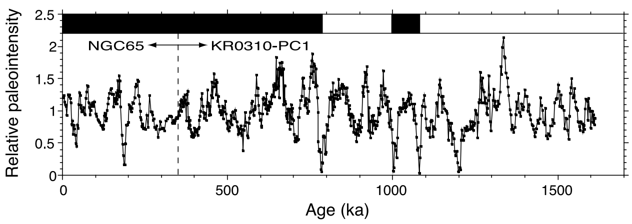

Fig. 10. Relative paleointensity record since 1.6 Ma in the central North Pacific: a composite of cores KR0310-PC1 and NGC65.

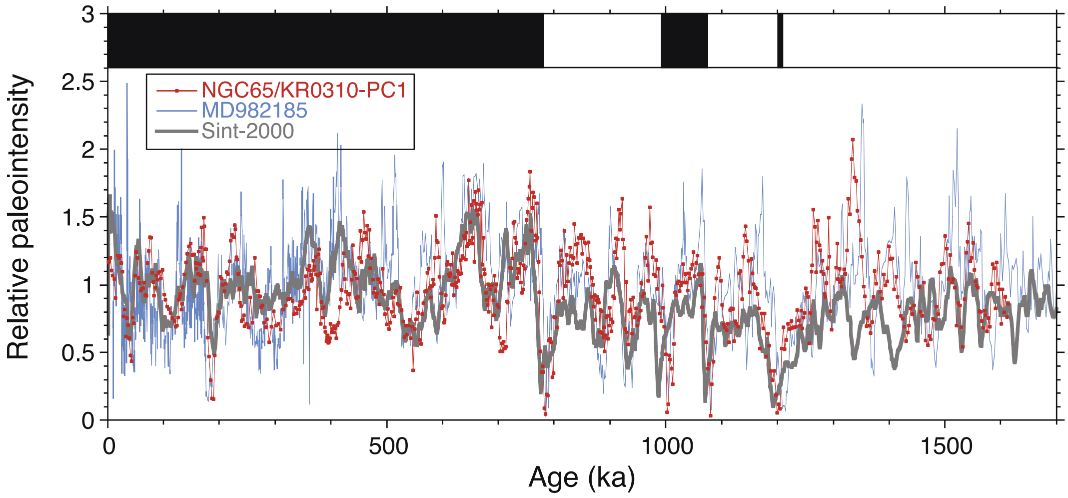

Fig. 11. Comparison of relative paleointensity records: NGC65/KR0310 -PC1 from the central North Pacific (red), MD982185 from the western equatorial Pacific (Yamazaki and Oda, 2002, 2005) (blue), and the Sint-2000 stack of Valet *et al.* (2005) (gray).

choice of a normalizer. As discussed above, a long-term trend in relative paleointensity can be biased depending on the normalizer. In conclusion, we consider the lower paleointensity in the late Matuyama Chron has not yet been established.

Based on the Sint-2000 stack, Valet *et al.* (2005) also suggested an asymmetry of paleointensity near reversals; dipole field intensity begins to decay 60–80 kyr before reversals and rebuilds in the opposite direction in a few thousand years. In Fig. 11, all three records show this tendency at the Brunhes/Matuyama transition. However, at other boundaries such as the upper and lower Jaramillo transitions and the Cobb Mountain subchron, consistency among the records is not good. For example, the asymmetry is only observed in the Sint-2000 at the upper Jaramillo transition. At the lower Jaramillo transition and the Cobb Mountain subchron, the records of Sint-2000 and MD982185 seem to show the asymmetry, but NGC65/KR0310-PC1 does not. The present status of sedimentary paleointensity studies is such that the amplitudes of normalized intensity variations often do not coincide even among cores taken from a narrow area, probably because the amplitudes could be affected by factors that have not been understood well. For example, Tauxe *et al.* (2006) argued that DRM acquisition process could be controlled by flocculation of sediment particles, and DRM intensity may be affected by size distribution of sediment flocs. Furthermore, a change in a sedimentation rate may affect normalized intensity (Yamazaki and Oda, 2005). Besides the problems in the amplitude, a slope of a relative paleointensity decrease/increase in the timescale of a few tens of thousand years may not be constrained well because of uncertainty in age due to possible fluctuation in a sedimentation rate. It would hence be too early to conclude that the asymmetry in intensity variations near polarity reversals is an intrinsic feature of the geomagnetic field behavior.

4.3 Possible orbital modulation

A spectral analysis shows power density around the 100 kyr period in the relative paleointensity record of NGC65/KR0310-PC1 (Fig. 9). The normalizer, IRM, does not have significant power at the frequency, and there is no significant coherency between the normalized intensity and the normalizer. This confirms the previous results from this region during the Brunhes chron (Yamazaki, 1999).

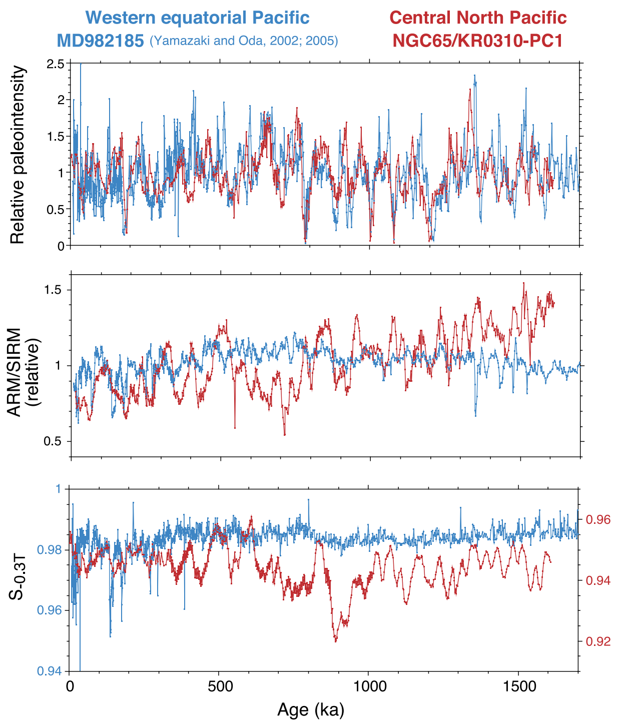

Fig. 12. Comparison of relative paleointensity (top) and magnetic properties (middle: ARM/SIRM ratio, bottom: *S* ratio) between core NGC65/KR0310-PC1 from the central North Pacific (red) and core MD982185 from the western equatorial Pacific. Relative variations are shown for ARM/SIRM; the average of ARM/SIRM is 0.060 for NGC65/KR0310-PC1, and 0.035 for MD982185.

We compare the relative paleointensity and magnetic property records of NGC65/KR0310-PC1 from the central North Pacific with those of core MD982185 from the western equatorial Pacific (Yamazaki and Oda, 2002, 2005). The ~100 kyr period was also found in the relative paleointensity record of MD982185 (Yamazaki and Oda, 2002).

The two relative paleointensity records closely agree with each other (Fig. 12). The period between about 400 and 500 ka is an exception. The disagreement is probably due to the uncertainty of age estimation, which can be ~50 ka at maximum there. The ages of the two records are independently controlled; S-ratio variations were correlated to the LR04 stack of δ^{18}O for KR0310-PC1, and magnetic susceptibility was correlated to the δ^{18}O curve of ODP Site 1143 for MD982185 (Yamazaki and Oda, 2005). Their uncertainty would be larger during the Brunhes chron than the Matuyama chron because ~40 kyr frequency dominates the variations before ~900 ka, whereas ~100 kyr frequency dominates after that.

On the other hand, the changes of magnetic properties are significantly different between the two, which reflects that the two regions are in different oceanographic regimes. First, the amplitude of the variations is considerably larger in the North Pacific than in the equatorial Pacific. In ARM/SIRM, which reflects variations of magnetostatic interaction and/or magnetic grain size, long-term trends of the two records are opposite; NGC65/KR0310-PC1 shows a low at 550 to 650 ka and an increase toward older ages, whereas MD982185 indicates a high and a decreasing trend at the corresponding periods. In the timescales of 10^4 to 10^5 years, the variations are incoherent; some fluctuations are anti-phase (for example, around 520 and 1120 ka), but others are in-phase. Variations in the magnetic mineralogy proxy, $S_{-0.3\mathrm{T}}$, are also incoherent. The coherent variations in the relative paleointensity despite large differences in the magnetic properties suggest that rock-magnetic contamination to the relative paleointensity is small, if any, and the ~100 kyr period in the relative paleointensity records would reflect the geomagnetic field behavior. This problem will be addressed further in an accompanying paper (Yokoyama *et al.*, in this volume) based on more sophisticated analyses using the wavelet transform.

5. Conclusions

(1) A relative paleointensity record during the last 1.6 m.y. was obtained from a sediment core KR0310-PC1 in the central North Pacific, which extends the record of core NGC65 during the Brunhes chron (Yamazaki, 1999) obtained at practically the same site. The age control is based on the correlation of the S ratio ($S_{-0.1\mathrm{T}}$) variations with the global δ^{18}O stack LR04 (Lisiecki and Raymo, 2005).

(2) IRM was used as the normalizer of the relative paleointensity estimation; ARM was not adopted because ARM is sensitive to magnetostatic interaction among magnetic particles, which is evidenced in these sediments by an inverse correlation between the ratio of ARM to SIRM and SIRM without significant magnetic grain-size changes. On ARM normalization, significant coherence is observed between the normalized intensity and the normalizer, but not on IRM normalization. I interpret this result as that the effect of magnetostatic interaction on DRM acquisition would be similar to that on IRM but smaller than ARM, and hence ARM may have overcompensated magnetic concentration changes.

(3) The paleointensity record of NGC65/KR0310-PC1 was compared with that of MD982185 from the western equatorial Pacific (Yamazaki and Oda, 2002, 2005) and the Sint-2000 stack (Valet *et al.*, 2005). Valet *et al.* (2005) suggested based on the Sint-2000 stack that the average paleointensity during the late Matuyama chron is lower than that of the Brunhes chron, but this is not observed in the records of NGC65/KR0310-PC1 and MD982185. The comparison also indicates that the asymmetry of paleointensity near reversals suggested by Valet *et al.* (2005) is observed commonly at the Brunhes-Matuyama transition, but consistency among the records are not good at other reversals.

(4) A spectral analysis on the NGC65/KR0310-PC1 paleointensity record shows a power at the ~100 kyr period. The relative paleointensity and magnetic properties of NGC65/KR0310-PC1 were compared with those of MD982185; the two sites belong to different oceanographic regimes. Coherent variations in the relative paleointensity despite incoherent changes in the magnetic properties suggest that rock-magnetic contamination to the relative paleointensity is small, if any, and the ~100 kyr period in the relative paleointensity records would reflect the geomagnetic field behavior.

(5) Intercore correlation using magnetic susceptibility between the piston-core KR0310-PC1 and the gravity-core NGC65 revealed that the corresponding depth intervals of the former are about 50% longer than those of the latter, although the two were taken at the same site and the values of volumetric magnetic susceptibility are almost identical. This indicates that oversampling of the piston core and/or undersampling of the gravity core occurred, as reported by Skinner and McCave (2003) and Széréméta *et al.* (2004).

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