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Relative paleointensity of the geomagnetic field over the last 200,000 years from ODP Sites 883 and 884, North Pacific Ocean

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Abstract

Ocean Drilling Program Sites 883 and 884 were cored as part of a three-site depth transect down the slopes of Detroit Seamount in the North Pacific Ocean. Continuous mineral magnetic and paleomagnetic measurements were made using u-channel samples for sediments that span the last 200 kyr. Thermomagnetic and high-field hysteresis data indicate that the magnetic mineralogy of the clay-rich sediments at the two sites is dominated by pseudo-single domain magnetite in a narrow range of grain sizes and concentrations, thereby meeting established criteria for relative paleointensity studies. The natural remanent magnetization (NRM) has been normalized using the anhysteretic remanent magnetization (ARM) and the low field magnetic susceptibility (χ). The NRM/ARM and NRM/χ records are nearly identical for both cores. Furthermore, coeval horizons in the two cores can be correlated by matching > 100 magnetic susceptibility features over the 15 m length of both cores. Coherence function analysis indicates that the records are not significantly affected by local environmental conditions. These factors suggest that the large-scale variations in normalized remanence are most likely due to geomagnetic paleointensity fluctuations. Dating constraints are provided by a $\delta^{18}O$ stratigraphy from Site 883. Our North Pacific paleointensity versus age curve is similar to a published record from the western Caroline basin and a recently proposed global paleointensity curve. Offsets (up to 10 kyr) in the timing of paleointensity features between our composite North Pacific record and the global curve may result from imprecisions in the dating of our record. Nevertheless, the correspondence between the paleointensity records suggests that, in suitable sediments, paleointensity of the geomagnetic field can give a globally coherent, dominantly dipolar, signal. © 1997 Elsevier Science B.V.

Keywords: magnetic field; magnetic intensity; North Pacific; Ocean Drilling Program

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1. Introduction

Temporal variations in geomagnetic phenomena can often be well resolved in marine sediments because sedimentation is usually continuous and it is often possible to obtain high-resolution chronologies through $\delta^{18}O$ stratigraphy. Analyses of the ancient intensity of the geomagnetic field are conducted much less routinely than conventional magnetostratigraphic studies because the intensity of the geomagnetic field at the time of remanence acquisition is not a simple function of the remanence intensity of the sediment (e.g., [1–3]). In particular, variations in magnetic mineralogy, concentration, and grain size of the magnetic particles can affect the measured remanence intensity.

Despite the difficulty in assessing the reliability of paleointensity records from sediments, paleointensity studies are important because they can contribute to a more complete understanding of geomagnetic field behavior. Accurate records of geomagnetic field intensity are also necessary outside of geomagnetism: atmospheric production of cosmogenic isotopes is strongly modulated by the intensity of the geomagnetic field, therefore paleointensity determinations are important for calibrating cosmogenic isotopic age determinations of geomorphic surfaces (e.g., [4]), as well as for calibrating the radiocarbon time-scale (e.g., [5,6]).

In paleointensity studies of sediments, the strength of the ancient magnetizing field is usually estimated by normalizing the measured remanence intensity with a magnetic parameter that is dependent on the concentration and type of magnetizable material within the sediment (most commonly, the low-field magnetic susceptibility (χ), the anhysteretic remanent magnetization (ARM) or the saturation isothermal remanent magnetization (SIRM)). Empirical evidence indicates that, with such normalizations, paleointensity studies are most likely to produce reliable results if restricted to sediments in which there are small variations in magnetic properties. A number of criteria for 'magnetic uniformity' have been proposed, to which sediments should comply in order to be considered appropriate for paleointensity studies [2,3]. These criteria are conservative and have been adopted to restrict studies to materials in which the assumption of linearity between magnetizing field and magnetization has empirical support. The criteria are: first, magnetite must be the sole remanence carrier; second, the magnetite grains must range in size between 1 and 15 μm; third, the maximum concentration of magnetite must be no more than ten times the minimum concentration of magnetite [2,3]. Beside these tests for magnetic uniformity, several other criteria should apply, including: (1) a stable, well defined, single component of magnetization should characterize the natural remanent magnetiza-

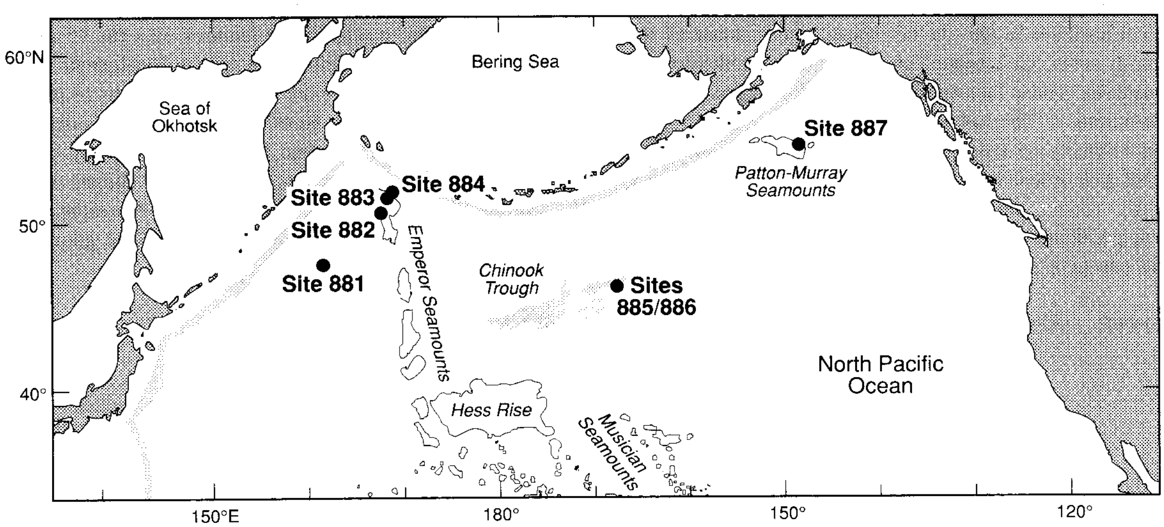

Fig. 1. Location map of the North Pacific Ocean with sites cored during ODP Leg 145, including Sites 883 and 884 (from [29]).

tion (NRM) at the demagnetization level used for paleointensity normalization; (2) the detrital remanent magnetization must be a reliable recorder of the geomagnetic field orientation, with no evidence of inclination error; (3) records obtained from multiple normalization parameters should agree; (4) coherence between paleointensity determinations and the magnetic parameters used for normalization should be minimal; (5) duplicate or multiple records from a given region should agree within the limits of the common time-scales [3].

In many cases, marine sediments that pass the above tests for magnetic stability and uniformity yield reproducible records of relative paleointensity over the last several hundred thousand years [7–18]. When these paleointensity records are plotted on a common time-scale, based on $\delta^{18}O$ dating, remarkable agreement is evident within the resolution of the dating for the respective cores [19]. This suggests that it may be possible to use paleointensity of the geomagnetic field as a global stratigraphic tool.

We have obtained paleointensity records from Holes 883D and 884D, which were recovered from the North Pacific Ocean (Fig. 1) during Leg 145 of the Ocean Drilling Program (ODP), in order to test the hypothesis of whether there is a globally coherent paleointensity signal. The Leg 145 data extend the spatial coverage of sites from which relative paleointensity records are available and they provide a longer data set (to 200 kyr) than many recently published records.

2. Geologic and oceanographic setting

Sites 883 and 884 were cored in close proximity (the sites are separated by 00°26′) along a three-site depth transect (with Site 882) down the slopes of Detroit Seamount (Fig. 1). Of the three sites, Hole 883D was cored at the shallowest water depth (2385 m) and Hole 884D was cored 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 discovery of foraminifer-bearing sediments on the nearby Meiji Seamount [20,21] suggested that Site 883 would be a suitable location for $\delta^{18}O$ studies. Site 884 has lower $CaCO_3$ contents, presumably because of increased dissolution of carbonate at greater water depths. Site 884 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 sediments in Hole 884D are dominated by dark gray clay, while Hole 883D contains greater abundances of diatoms and $CaCO_3$. Sediments at both sites contain intermittent volcanic ash layers. While reliable magnetostratigraphies were obtained from both sites during Leg 145, the clay-rich sediments at Site 884 yielded consistently high quality data (to ca. 13 Ma), while the diatom-rich intervals at Site 883 are not as stably magnetized as the clay-rich intervals and a discernible magnetostratigraphy was determined only to 2.6 Ma [22].

3. Methods

Cores were recovered to a depth of 17.0 meters below sea floor (mbsf) in Hole 883D and to a depth of 14.8 mbsf in Hole 884D. Paleomagnetic measurements were made on u-channel samples with a narrow access pass-through cryogenic magnetometer equipped with high-resolution (about 45 mm spatial resolution) pick-up coils, as described by Weeks et al. [23], at the CFR Paleomagnetism Laboratory, Gif-sur-Yvette, France. The samples were subjected to stepwise alternating field (AF) demagnetization of the natural remanent magnetization (NRM) at applied fields of 10, 20, 25, 30, 40, 50, and 60 mT. An ARM was imparted at a constant AF of 99 mT, with a bias field of 0.05 mT that was produced in a large set of Rubens coils. Stepwise demagnetization of the ARM was carried out at applied fields of 10, 20, 25, 40, and 60 mT. Measurements of χ were made at a spatial resolution of about 3 cm on all of the u-channels in a vertical translation system.

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. Each sample was subjected to high-field magnetic hysteresis analysis up to maximum fields of 1 T. 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.

4. Results

4.1. Dating and correlation of cores

Age control for the last 200 kyr is based on oxygen isotope determinations on the benthic genus *Uvigerina* from Hole 883D [24]. There are many gaps in the $\delta^{18}O$ record (Fig. 2a) because of the low abundance, or complete absence, of foraminifers in some intervals (Fig. 2b). Foraminifers are sufficiently abundant, however, to establish general age control, particularly at the glacial terminations. Accelerator Mass Spectrometer ^{14}C dates are consistent with the $\delta^{18}O$ chronology back to 22 kyr [25]; however, we have relied solely on interpretations of the $\delta^{18}O$ record, as given by Keigwin [24] and Kotilainen and Shackleton [26], to avoid any potential discordance between the two types of age estimate. Stage 1 is poorly represented [24] and the details of Stage 5 have not been resolved due to the absence of foraminifers (Fig. 2). The principal age control points are shown on Fig. 2 and are summarized in Table 1. Dates and nomenclature of the isotopic stages follow Martinson et al. [27]. Dating of the cores was achieved by linear interpolation between the control points listed in Table 1.

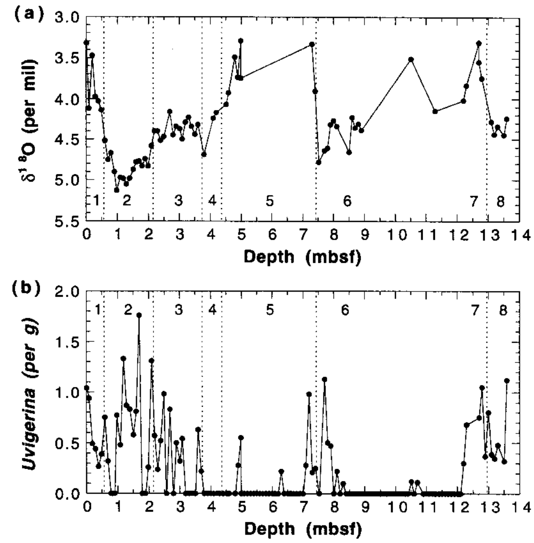

Fig. 2. (a) $\delta^{18}O$ (from *Uvigerina*) results, and (b) number of *Uvigerina* fossils per gram of sediment, from Hole 883D, after Keigwin [24]. Interpreted $\delta^{18}O$ stage boundaries are after Keigwin [24] and Kotilainen and Shackleton [26]. Depths are in meters below sea floor.

Table 1
Chronology of Detroit Seamount sediments from ODP Holes 883D and 884D

Depth in Hole 883D [a] (mbsf)	Depth in Hole 884D [b] (mbsf)	Age [c] (kyr)	$\delta^{18}O$ event [d]	Reference [e]
0.50	0.31	12	2.0	Keigwin [24], K&S [26] [f]
1.40	1.00	20	2.21	Morley et al. [25], K&S [26]
2.10	1.55	24	3.0	K&S [26]
3.65	3.15	59	4.0	K&S [26]
3.79	3.27	64	4.22	This study
4.14	3.58	74	5.0	K&S [26]
7.42	6.48	127	6.0	Keigwin [24], K&S [26]
9.01	8.58	190	7.0	This study (interpolated)
12.90	11.77	244	8.0	Keigwin [24]

[a] Depths (in meters below sea floor (mbsf)) are of $\delta^{18}O$ features reported by Keigwin [24].
[b] Equivalent depths in Hole 884D were determined from susceptibility correlations between cores.
[c,d] Age and name of $\delta^{18}O$ events from Martinson et al. [27].
[e] Reference from which interpretations of $\delta^{18}O$ curve are based.
[f] K&S is Kotilainen and Shackleton.

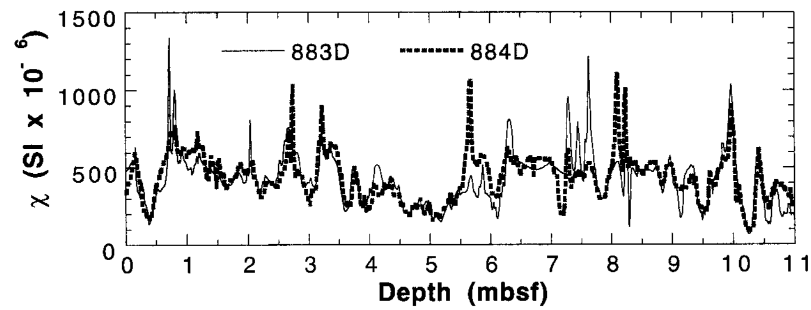

Fig. 3. Low-field magnetic susceptibility (χ) records for Holes 883D and 884D, with the Hole 883D record converted to depths (mbsf) in Hole 884D. The coring gap between cores 883D-1H and 883D-2H is evident at 6.3–6.8 mbsf. The susceptibility records provide the basis for correlation between the two records (R = 0.711).

Detailed correlation between Holes 883D and 884D was achieved by matching more than 100 common features in the respective χ records. Many of the prominent χ peaks are coincident with ash layers. A common depth scale for the two cores was derived by matching χ features using 'Analyserie' software [28]. The data in Fig. 3 are plotted with depths from Hole 883D converted to depths (mbsf) in Hole 884D. Despite minor differences between the records, which would be expected due to differences in depositional environment at the two sites and the presence of coring gaps, such as that in Hole 883D

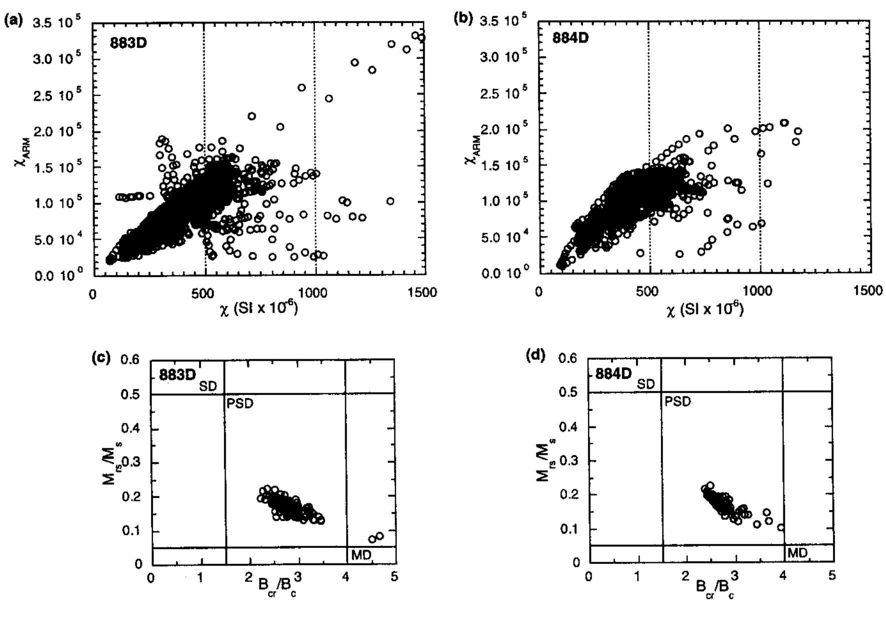

Fig. 4. Magnetic granulometry for Holes 883D and 884D. χ_{ARM} vs χ plots (cf. [30]) for (a) Hole 883D and (b) Hole 884D. Each divergence from the major cluster of points on these plots is due to a stratigraphically restricted departure from mean grain size/concentration, as reported by Roberts et al. [29]. Values of χ_{ARM} have not been normalized to account for the volume of material sensed by the pick-up coils in the magnetometer. (c) and (d) Grain size variations, as indicated by high-field magnetic hysteresis data on bivariate plots of M_{rs}/M_s versus B_{cr}/B_c, for (c) Hole 883D and (d) Hole 884D. Fields are shown for single domain (*SD*), pseudo single domain (*PSD*), and multi-domain (*MD*) grains, after Day et al. [31].

from 6.3 to 6.8 mbsf (Hole 884D depths), the correlation coefficient of 0.711 for the curves indicates good agreement between the records. Overall sedimentation rates at the two sites are comparable (average ca. 5 cm/kyr).

4.2. Mineral magnetic properties of the sediments

A detailed description of the mineral magnetic properties of the Detroit Seamount sediments is given by Roberts et al. [29]. The principal results are briefly summarized here to demonstrate that the sediments meet the established criteria for relative paleointensity determinations.

The χ_{ARM}/χ ratio is useful for estimating relative variations in magnetic grain size [30]. The largest peaks in the χ records, which are usually associated with tephra layers, result in low χ_{ARM}/χ values, which suggests that the ashes are dominated by high concentrations of coarse magnetic particles, relative to the surrounding sediments (Fig. 4a,b). The χ_{ARM}/χ data indicate that the clay-rich sediment in Holes 883D and 884D is of relatively uniform grain size. Distinct trends toward relatively coarse grain sizes are evident; however, Roberts et al. [29] demonstrated that each of these zones of anomalous grain size is stratigraphically restricted and is associated with a tephra layer. Comparison of the range of χ_{ARM} and χ values within the main cluster of points in Fig. 4a,b indicates that the concentration of magnetic grains in the clay-rich sediments varies within a factor of about 9 in Hole 883D and within a factor of about 10 in Hole 884D.

Four hysteresis parameters, M_{rs} (saturation remanence), M_s (saturation magnetization), B_{cr} (coercivity of remanence), and B_c (coercive force), were routinely measured. M_{rs}/M_s and B_{cr}/B_c are highly sensitive to variations in grain size [31]. Detailed measurements of magnetic hysteresis properties indicate little variation in grain size of the bulk clay-rich sediment throughout Holes 883D and 884D (Fig. 4c,d). Clustering of hysteresis data within a restricted area of the pseudo-single domain (PSD) field (cf. [31]) support the χ_{ARM}/χ results, which indicate relatively 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 [29], as would be expected if the magnetic mineral assem-

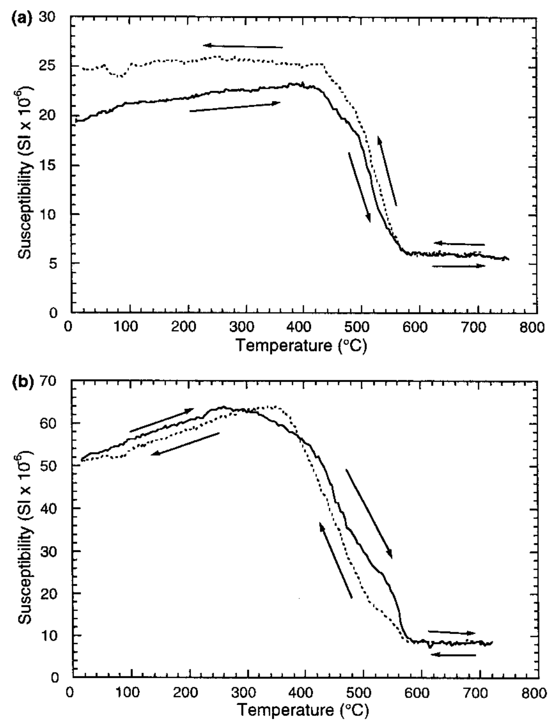

Fig. 5. Temperature-dependence of low-field magnetic susceptibility for representative samples: (a) 145-883D-2H-2, 70 cm, (b) 145-884D-2H-3, 80 cm. Solid line = heating curve; dashed line = cooling curve. The major drop in susceptibility above 550°C is indicative of a magnetic mineral assemblage that is dominated by low-titanium magnetite.

blage is dominated by a single magnetic phase with a restricted range of grain sizes.

Measurements of the temperature dependence of magnetic susceptibility for the Detroit Seamount sediments consistently indicate a major decrease in χ at high temperatures (with the signal disappearing above 550°C), which is indicative of a magnetic assemblage that is dominated by low-titanium magnetite (Fig. 5) (cf. [32]). This is supported by measurements of the S-ratio ($S = -IRM_{-300\ mT}/SIRM_{1.2\ T}$) which vary between 0.88 and 1.00, as would be expected for a ferrimagnetic mineral such as (titano-)magnetite [29]. Together, these results indicate that the magnetic signal in the clay-rich sediments is dominated by magnetite.

The above results indicate that the criteria for magnetic uniformity for relative paleointensity determinations [2,3] are satisfied by the clay-rich sediments from Holes 883D and 884D. Volcanic ash-rich intervals of the core are less appropriate for such studies because of the coarse-grained nature of, and high concentrations of magnetic minerals in, these layers. Data from these intervals that do not meet the criteria for magnetic uniformity have therefore been omitted from our estimates of the relative paleointensity of the geomagnetic field.

4.3. Paleomagnetic stability

Vector demagnetization plots can be constructed at specified positions along the u-channel because

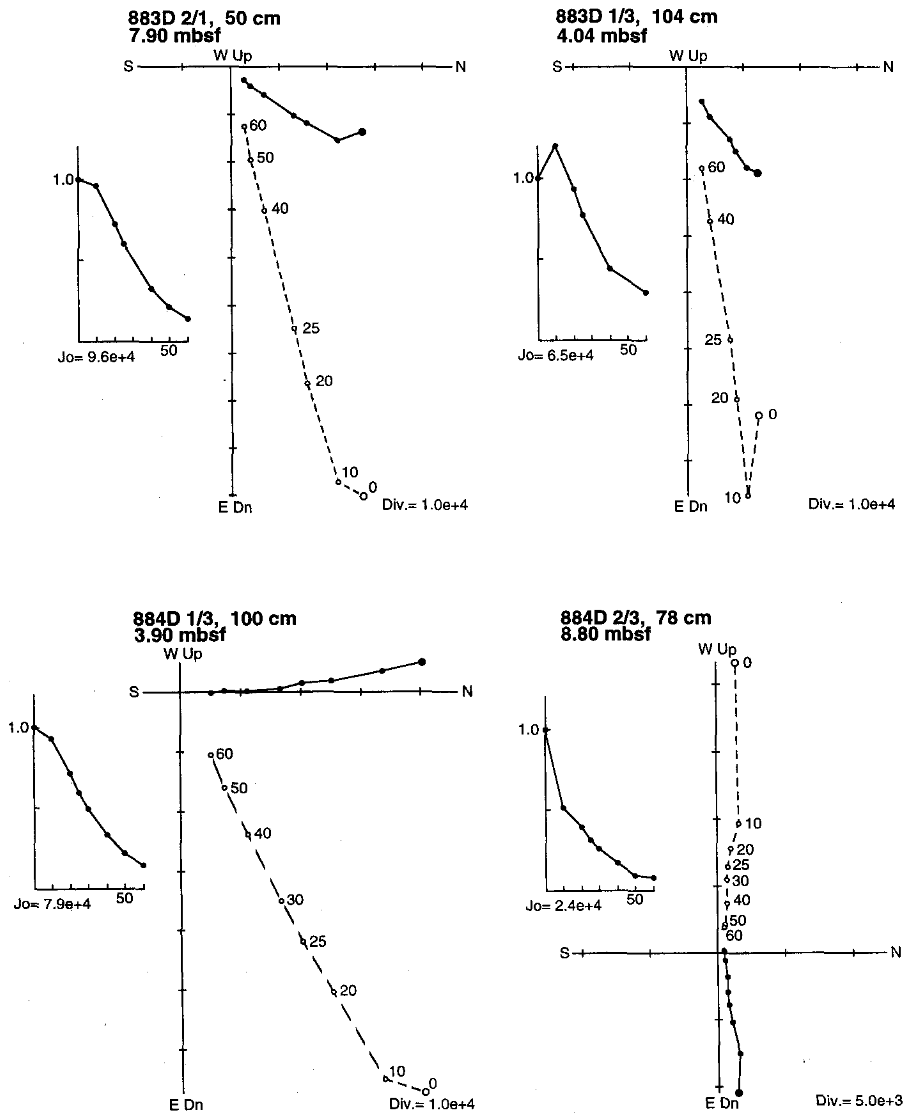

Fig. 6. Representative vector component diagrams for typical samples from Holes 883D and 884D. A single characteristic remanence component is isolated in each diagram by 25 mT. This demagnetization step was therefore used to normalize the relative paleointensity records.

the position of the u-channel is precisely controlled during measurement (cf. [23]). A univectorial characteristic remanence component is reached by peak AFs of 25 mT (Fig. 6). Data at this demagnetization level were therefore used for paleointensity normalization. Furthermore, paleomagnetic declinations and inclinations are consistent with a detrital remanence that faithfully records geomagnetic field directions expected at the site ($D_{exp} = 0°$; $I_{exp} = 68°$), without inclination error (Fig. 7). This result is consistent with the high-quality magnetostratigraphic results that were obtained back to 13 Ma at Site 884 [22].

Low-amplitude secular variation in the geomagnetic field about expected mean field values is evident throughout the Detroit Seamount sediment records (Fig. 7). The stability of the natural remanence provides a further argument for the reliability of the Detroit Seamount sediments for paleointensity determination. An excursion that is evident in each component of the vector waveform occurs at about 190 kyr in Hole 884D. Apparent 'excursions' can be artefacts of the long-core measurement technique in zones where there are large changes in remanence intensity (e.g., [23,33]). We have not conducted measurements of discrete samples to determine whether this excursion is real; however, we interpret it to represent a real excursion because the remanence is stable in this interval (Fig. 6; 884D, 8.80 mbsf), and because there are numerous geomagnetic excursions that have been reported from around the world at this time [14,15,17,34].

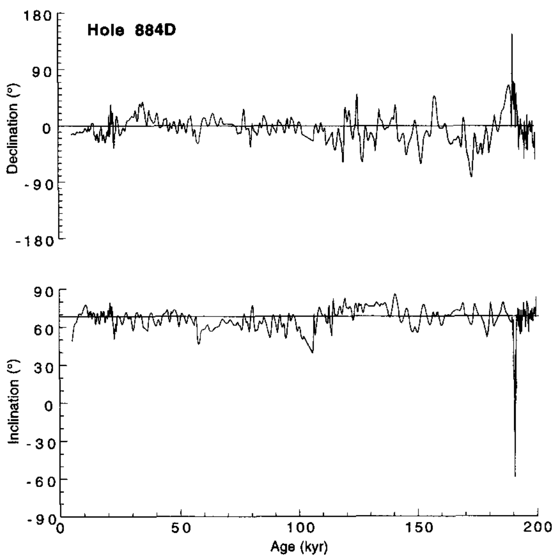

Fig. 7. Declination and inclination versus age (after demagnetization at 25 mT) for Hole 884D. Horizontal lines indicate values about which an axially dipolar geomagnetic field is expected to vary at the site location. Declinations have been corrected for a constantly increasing core rotation with depth and are plotted with the mean corrected declination at 0°.

4.4. Normalized remanence records

Similar results are obtained from NRM/ARM and NRM/χ normalizations at both Holes 883D and 884D (Fig. 8a,b). Furthermore, there is good agreement between the NRM/ARM records from Holes 883D and 884D (Fig. 8c). An additional paleointensity record was obtained from the first of the three sites (Site 882) cored on Detroit Seamount [34]. This record was obtained from discrete samples taken at 15 cm intervals and is therefore of much lower resolution than those reported here. Despite differences in lithology and sedimentary environment, the three records from Detroit Seamount agree well and the normalized remanence features contain similar amplitudes of variation.

4.5. Coherence function analysis

While all of the established criteria for relative paleointensity determinations appear to be satisfied by the records from Holes 883D and 884D, it is still possible that the agreement between the records may be due to environmental control that affects the magnetic signal of these closely spaced sites in the same way. We therefore carried out a coherence function analysis on the record from Hole 884D (cf. [3,8]), to test whether the normalized remanence signal is influenced by environmental (e.g., climatic) factors over specific frequency ranges. If the paleointensity record (NRM/ARM) and related mineral magnetic (e.g., χ) parameters do not display significant coherence, one can have confidence that the paleointensity normalization is not substantially affected by lithological or other environmental factors. Squared coherence between NRM/ARM and χ is plotted versus frequency in Fig. 9. This analysis indicates that there is no significant coherence between the normalized remanence record and the bulk mineral magnetic parameters at any frequency, which suggests that local environmental factors do not have a strong impact on our paleointensity estimates.

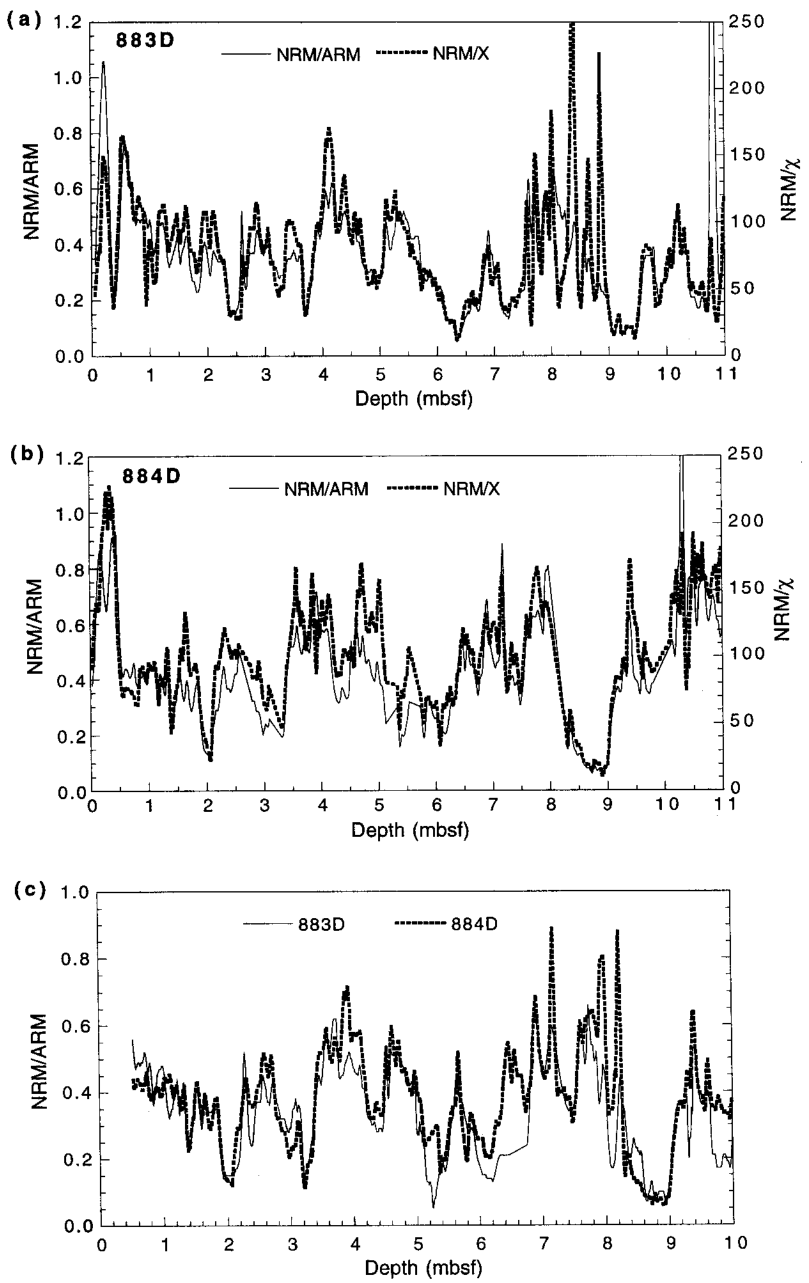

Fig. 8. NRM/ARM and NRM/χ versus depth for (a) Hole 883D and (b) Hole 884D. Comparison of NRM/ARM records versus depth for Holes 883D and 884D is given in (c). Data in (c) are plotted on the depth scale (mbsf) for Hole 884D. Agreement between the curves for both normalizing parameters and between the records from both holes indicates that the sediments are suitable for relative paleointensity investigations.

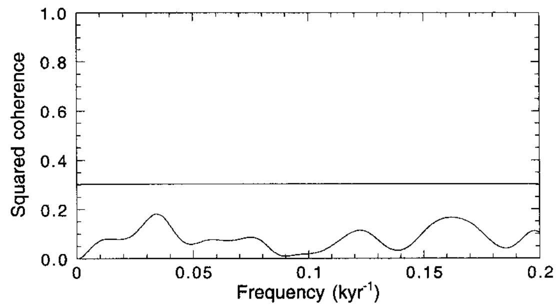

Fig. 9. Squared coherence of the NRM/ARM record with respect to the low-field magnetic susceptibility for Hole 884D. The horizontal line indicates the 95% confidence limit for zero coherence. The entire spectrum lies below this 95% confidence limit, which suggests that the normalized remanence record does not display coherence with local environmental factors (which exert the dominant control on the susceptibility).

4.6. Summary

Compliance with the most commonly applied mineral magnetic criteria, paleomagnetic stability, agreement between results of two paleointensity normalizations (NRM/ARM and NRM/χ), agreement between multiple paleointensity records from Detroit Seamount, and lack of coherence between the normalized remanence and bulk mineral magnetic parameters provide evidence that the records from Holes 883D and 884D meet the strictest criteria for relative paleointensity records (cf. [3]).

5. Discussion and conclusions

All of the above-described tests for assessing the reliability of relative paleointensity records are important; however, the most powerful test is whether there is agreement within the same geographic region, agreement between different depositional environments, and, ultimately, broad scale agreement between records from around the world. A composite paleointensity curve was obtained for Holes 883D and 884D by taking an arithmetic mean of the respective NRM/ARM curves (Fig. 10). No data are shown above 12 kyr because $\delta^{18}O$ stage 1 is poorly represented (i.e., it is much thinner than would be expected [24]).

The composite NRM/ARM curve (Fig. 10) represents an estimate of geomagnetic paleointensity variations at Detroit Seamount and contains several significant features, including: a decline from 200 kyr to a marked intensity low at 190 kyr, several intensity peaks between 180 and 135 kyr, three broad peaks centered at 95, 75, and about 50 kyr, with lows at 105, 90, 60 and 35 kyr. After 35 kyr, the intensity gradually increases toward the present day. The marked intensity low at 190 kyr is coincident with excursionary directions in both declination and inclination (Fig. 7). As stated above, we interpret this as representing a genuine excursion of the geomagnetic field, as has been widely reported from other sites during this time period [14,15,17,34]. The 190 kyr intensity low is apparently geographically

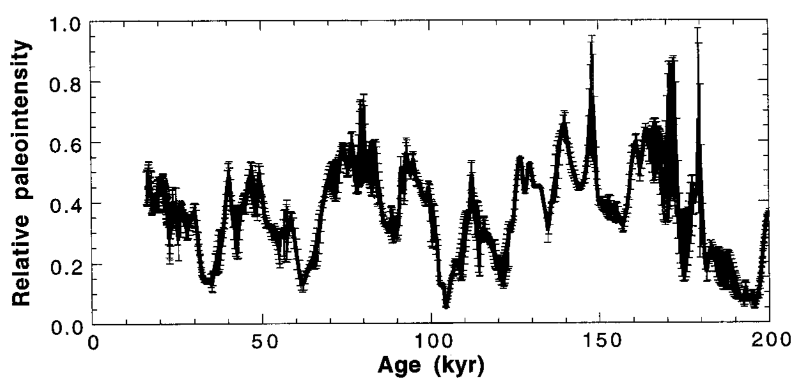

Fig. 10. Relative paleointensity of the geomagnetic field at Detroit Seamount for the last 200 kyr, based on an arithmetic mean of the NRM/ARM records from Holes 883D and 884D. Error bars indicate one standard deviation.

widespread and is evident in all of the published records that meet the reliability requirements for relative paleointensity studies (e.g., [12–15,17,19]). This paleointensity feature is well constrained for most of these records because the boundary between $\delta^{18}O$ stages 7 and 6 lies at 190 kyr [27].

A comparison of our record with many of the published records is shown in Fig. 11: this comparison is restricted to records with $\delta^{18}O$ chronologies (which excludes lacustrine records such as those of Thouveny et al. [35] and Peck et al. [36], and the Labrador Sea record of Stoner et al. [16], which has a ^{14}C-based chronology). There is generally good agreement between our North Pacific composite record and that of Yamazaki and Ioka [14] from the West Caroline basin (western equatorial Pacific). Both records contain the 190 kyr low, and the highs between 180 and 135 kyr, although the latter features are not exactly coincident in time and amplitude. Neither of these records have high-resolution age control in the 130–190 kyr interval and some of the discrepancies in timing could be due to inaccuracies in dating. Regardless, both records contain lows at around 100–105 kyr, 60–70 kyr, and 30–40 kyr, as well as the general increase in intensity toward the present day. There is also a general correspondence between our record and the 'global stack' of Guyodo and Valet [19], which was constructed using some of the other records shown in Fig. 11. Most of the apparent differences between the records shown in Fig. 11a–h are due to differences in interpretation of the relevant $\delta^{18}O$ records by the respective authors. When Guyodo and Valet [19] interpreted the respective $\delta^{18}O$ records in the same way, using the time scale of Martinson et al. [27], many apparent discrepancies between the records decreased. Misalignment between paleointensity features in our record and the global stack may be partially due to the relatively poor age control for our record. However, some of the differences in amplitude of the relative paleointensity features between our North Pacific record and the global stack of Guyodo and Valet [19] may be attributable to attenuation of the global curve that results from stacking records with age control of variable quality. Also, differences in amplitude between the records are to be expected even with a predominantly dipolar field because dipole wobble will produce different intensity variations at sites that are near-sided and far-sided with respect to the geomagnetic pole. Regardless, broad-scale similarities between the timing of the features in the various

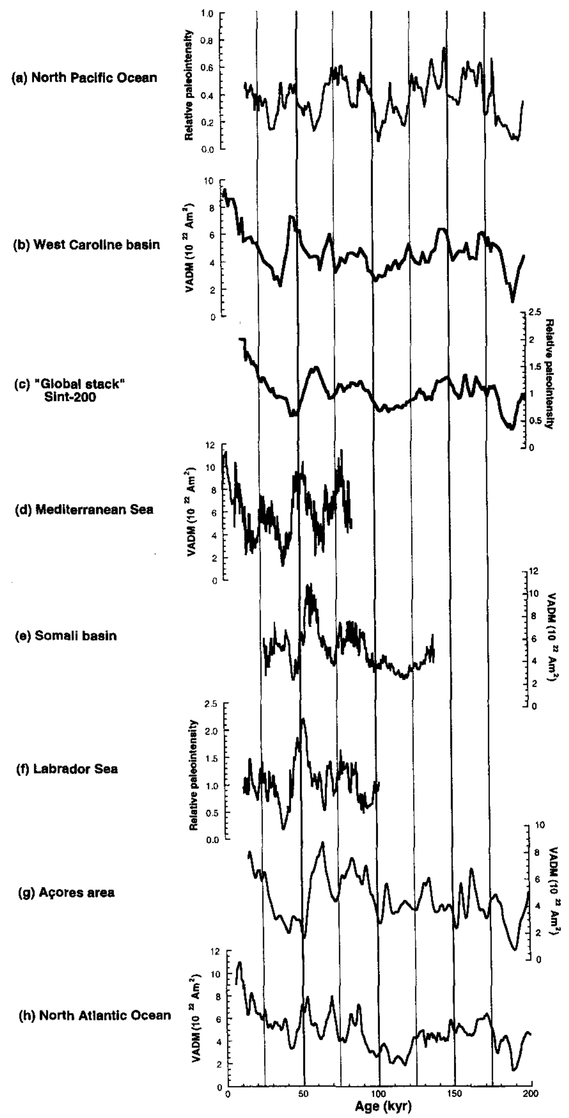

Fig. 11. Comparison of: (a) the Detroit Seamount relative paleointensity record with other records from around the world, including: (b) Yamazaki and Ioka [14] from the western Caroline basin (western equatorial Pacific); (c) the Sint-200 global paleointensity stack of Guyodo and Valet [19]; (d) Tric et al. [10] from the Mediterranean Sea; (e) Meynadier et al. [9] from the Somali Basin; (f) Stoner et al. [16] from the Labrador Sea; (g) Lehman et al. [17] from the central North Atlantic Ocean (Açores area); and (h) Weeks et al. [15] from the North Atlantic Ocean. See text for discussion.

records is encouraging and suggests that a predominantly global signal is recorded by the Detroit Seamount sediments.

Given the evidence that is emerging to support the hypothesis that there is a globally coherent (i.e., predominantly dipolar) paleointensity signal, we suggest that geomagnetic relative paleointensity may provide an alternative temporal framework for the North Pacific Ocean and other areas where harsh environmental conditions or poor $CaCO_3$ preservation prevent determination of reliable $\delta^{18}O$-based chronologies. Such use of paleointensity records has started to gain acceptance. For example, Stoner et al. [16] matched features from the global paleointensity record to those in Labrador Sea cores (Fig. 11) to 'tune' their chronology (because the $\delta^{18}O$ signal departs strongly from the expected signal because of meltwater dilution). Similarly, Peck et al. [36] used paleointensity features to help constrain the age of sediments from Lake Baikal, Siberia, which are difficult to date by other techniques. Roberts et al. [37] attempted to use paleointensity to correlate between marine and lake sediment records, but poor age control made it difficult to make a convincing comparison. Also, records from relatively small lakes are less likely to meet the stringent criteria for paleointensity determinations because of the many sources that can contribute to non-uniformity of the magnetic properties of lake sediments. Nevertheless, paleointensity correlation may enable development of an independent time scale for comparison of continental and marine sediment records, for which there is presently no satisfactory correlation technique. This could produce important benefits in enabling correlation of the paleoclimate records of deep-sea and continental environments and may enable a more satisfactory assessment of the relationship between the driving forces of climate in the respective environments. Development of a robust global paleointensity stratigraphy should therefore be a major goal for future work.

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