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Global Analysis of the Geomagnetic Field: Time Variation of the Dipole Moment and the Geomagnetic Pole in the Holocene

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Global features of the geomagnetic field over the past 10,000 years were studied. Reliable data set of inclination and declination at twelve localities were prepared by combining paleomagnetic data with archaeomagnetic data. We estimated the time variation of the geomagnetic dipole moment by analyzing the virtual geomagnetic pole (VGP) positions calculated from this data set.

The geomagnetic poles were calculated for every 100-year interval by averaging the VGP. The distribution of the geomagnetic poles has an elliptic shape and westward movement was predominant throughout the interval. The obtained time sequence of the movement of the geomagnetic pole can be divided into three intervals: during the period between ca. 7000 and ca. 3700 B.P. (B.P.: before 1950 A.D.), the movement of the geomagnetic pole was inactive, and it was active before and after this period, fluctuating over 10 degrees.

Continuous time variation of the dipole moment was inferred from the angular dispersion of the VGP, by investigating the relationship between the angular dispersion of the VGP and the dipole moment. The result suggests that the dipole moment had sharp peaks of high intensity around 8500 B.P., 4200 B.P. and 1200 B.P.

1. Introduction

Historical measurements of the geomagnetic field, started in the sixteenth century, indicate that the geomagnetic field is slowly changing. Archaeomagnetic studies can extend our knowledge of the geomagnetic field back to about 2000 years B.P. To extend it further back, we need to use sediment data, which are obtained from paleomagnetic studies of unconsolidated sediments. Paleomagnetic studies of lava flows are also used especially in investigating the intensity of the geomagnetic field.

The advantage of using sediments in investigating the geomagnetic secular variation is that sediment data are longer and continuous, compared to archaeomagnetic data. But sediment data have several difficulties in reconstructing the true variations of the geomagnetic field from the measure of the remanent magnetization:

1) The age of sediments does not necessarily coincide with the age of its remanent magnetization due to the time lag during the acquisition process of the remanence (HAMANO, 1980; OKADA and NIITSUMA, 1989). We must also take into account systematic errors in ages in radiocarbon dating due to the hard water effects and/or the reservoir effects (HEDGES, 1983; BARD, 1988).

2) Variations recorded in sediments are smoothed (or filtered) during the acquisition process (HYODO, 1984), and the amplitude of variation is lower than the true variation.

3) In declination measurement, only the relative variation can be measured and its absolute value is unknown in most cases. Offsets of angles in inclination also may occur during coring processes and/or some deformation effects during the compaction of the sediments.

Due to these difficulties, global analyses of the geomagnetic field from sediment data were limited (e.g. CREER and TUCHOLKA, 1982a; THOMPSON, 1984; HAGEE and OLSON, 1989).

In the present study, available sediment data sets for the last 10,000 years were reexamined and corrected by comparing with archaeomagnetic data and lava data. From this data set, we estimated the movement and the intensity of the geomagnetic dipole moment by analyzing the virtual geomagnetic poles (VGP).

2. Geomagnetic Secular Variation Curve

Sediment data all over the world for the last 10,000 years were examined, in which the stability of the remanent magnetization, accuracy of dating, and the rate of sedimentation were checked by the following criteria: 1) stability of the remanence was tested by AF demagnetizing experiments, 2) ages were based on at least five radiocarbon measurements, 3) the average sedimentation rate was higher than 0.4 mm per year. For the radiocarbon age, fluctuation of the radiocarbon level were corrected based on the dendrochronological studies (CLARK, 1975; BECKER and KROMER, 1986; STUIVER *et al.*, 1991).

To test the fidelity and to compensate for the difficulties in sediment data, the sediment data from Japan (OHNO, 1991), Australia (BARTON and BARBETTI, 1982; BARTON, 1983), England (TURNER and THOMPSON, 1982) and Hawaii (PENG and KING, 1992) were compared with the nearby archaeomagnetic and/or lava data, which are considered to be free from the above mentioned difficulties of sediment data. The England data show good consistency with the recently compiled nearby archaeomagnetic data (TARLING, 1989). The Hawaii data are consistent with lava data (HOLCOMB *et al.*, 1986). These four data (Japan, Australia, England and Hawaii) are included in the succeeding analysis without further correction. At the other sediment sites, nearby archaeomagnetic data are not available. Hence, we calculated a model of the geomagnetic field variation from the globally distributed archaeomagnetic data and lava data as reference, in which we applied spherical harmonic analysis, and corrected the sediment data by comparing with the model field. We consider that this is the most reliable method, available up to the present, for the correction of sediment data, although this method cannot properly correct the smoothing effect and the systematic error in radiocarbon dating which are not stationary throughout the data length.

2.1 A model of geomagnetic field for 400–1600 A.D.

Many authors applied spherical harmonic analysis to the historical measurement data for the last 400 years (e.g. BARRACLOUGH, 1974; YUKUTAKE, 1979, 1985, 1987) and obtained the time variation of the Gauss coefficients of up to degree four. We applied spherical harmonic analysis to the archaeo- and paleomagnetic data set between 400 A.D. and 1600 A.D. (that is, between 1550 B.P. and 350 B.P.). The data set was Iceland (BRYNJOLFSSON, 1957), England (TARLING, 1989), France (THELLIER, 1981), Bulgaria (KOVACHEVA, 1980, 1992, personal communication 1992), Ukraina (ZAGNIY, 1981; KOVACHEVA, 1982), China (WEI *et al.*, 1983), Japan (NAGATA *et al.*, 1963; SASAJIMA and MAENAKA, 1966; KONO, 1969; HIROOKA, 1971; SAKAI and HIROOKA, 1986; OHNO, 1988; MAENAKA, 1990), Australia (BARTON, 1983), Hawaii (HOLCOMB *et al.*, 1986), Southwest U.S.A. (DUBOIS, 1989; STERNBERG, 1989) and Meso America (WOLFMAN, 1984; DUBOIS, 1989). In all, 24 elements of the geomagnetic field (11 for inclination, 10 for declination and 3 for intensity) were prepared. For data in Australia, the secular variation curve obtained by combining the paleomagnetic and the archaeomagnetic data was used. We show all the prepared data in Fig. 1, together with the results of spherical harmonic analysis.

KONO (1973; see also 1976) proposed an iterating procedure to calculate relative Gauss coefficients from a data set of inclination and declination. We modified this method and used absolute value at which paleointensity data were available, and calculated the absolute Gauss

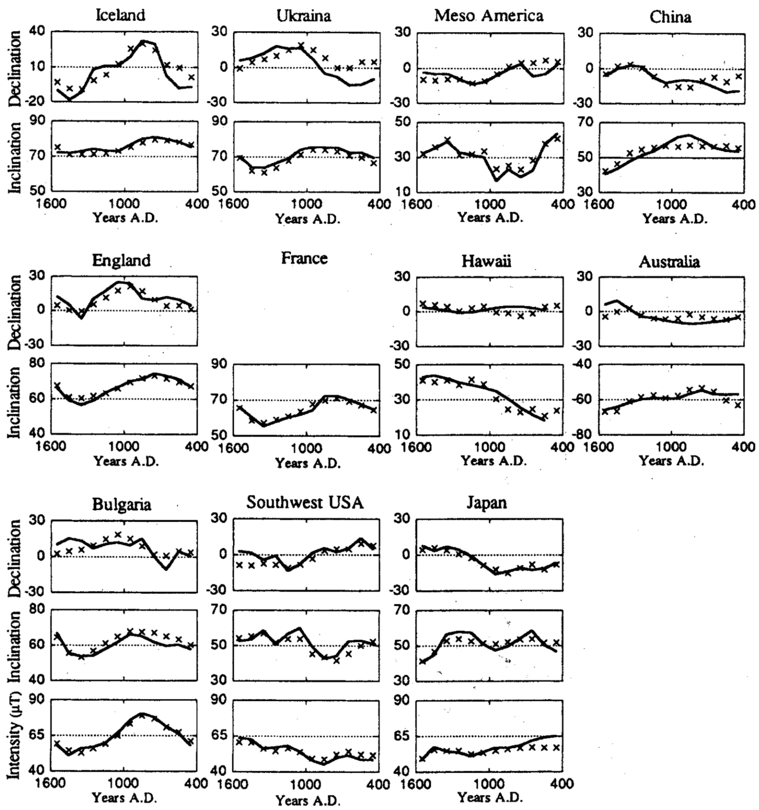

Fig. 1. Archaeomagnetic data set used for the present analysis. Thick lines show the geomagnetic secular variation curves between 400 A.D. and 1600 A.D. employed in the spherical harmonic analysis. Crosses are calculated from the obtained Gauss coefficients in Fig. 2.

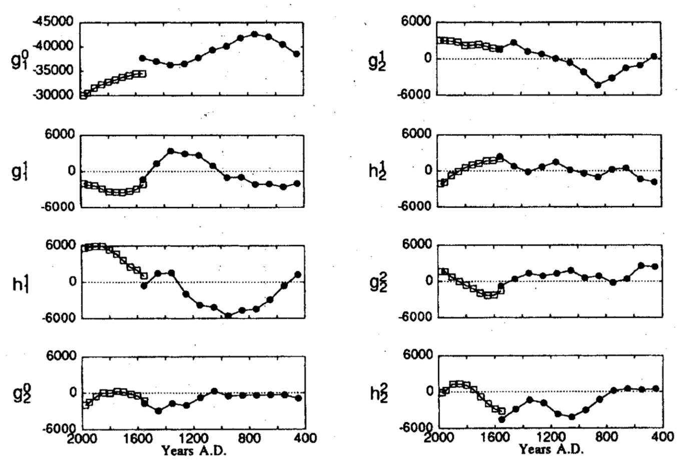

Fig. 2. Gauss coefficients for the last 1600 years. Solid circles show the present results of the spherical harmonic analysis of the archaeo- and paleomagnetic data set in Fig. 1. Open squares show results by YUKUTAKE (1987).

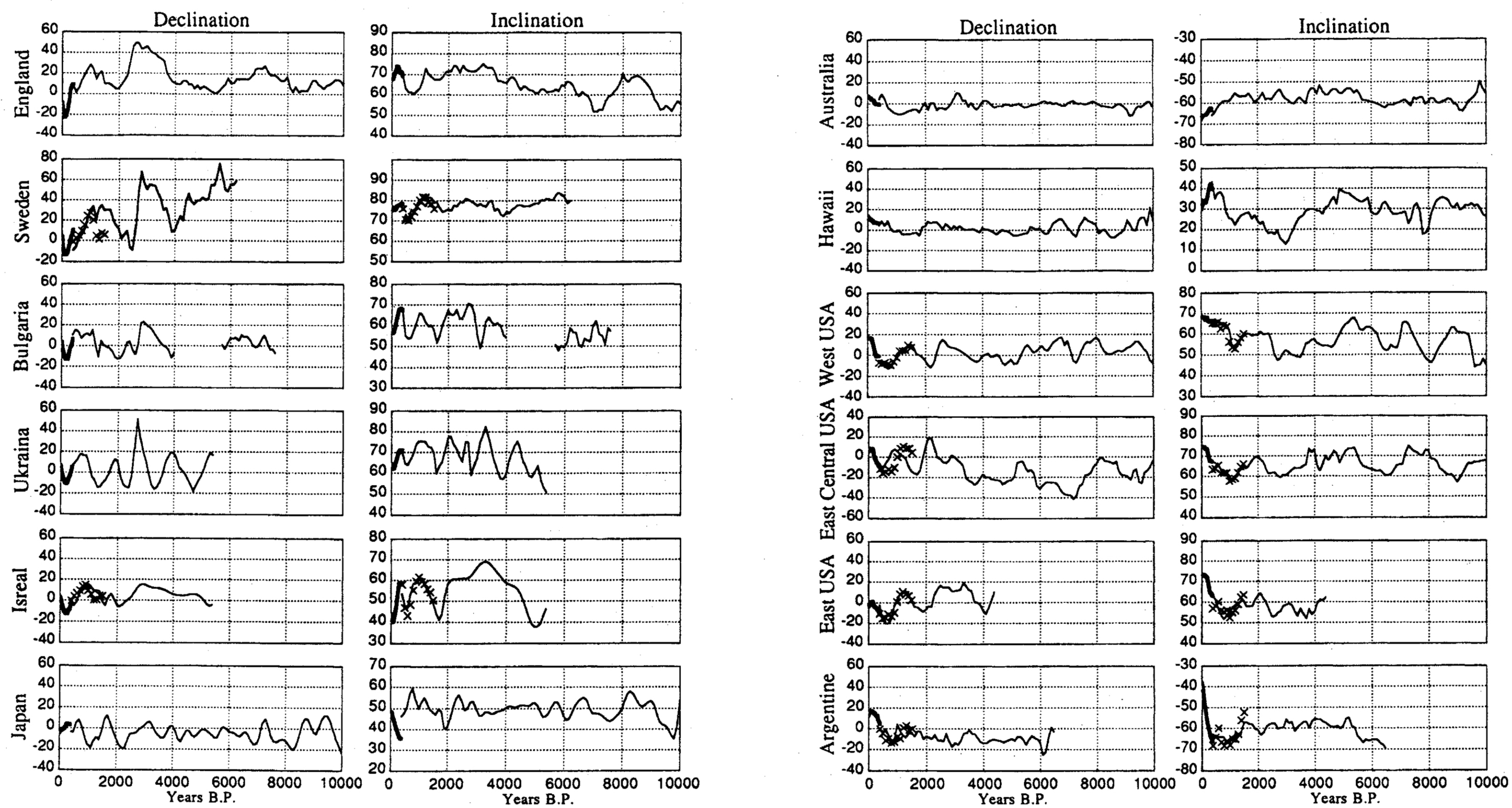

Fig. 3. Geomagnetic secular variation curves employed in the present analyses (solid line). Six sediment data sets were corrected by comparing with the calculated model field (crosses). Thick short lines between 0 and 400 B.P. are calculated from the Gauss coefficients by YUKUTAKE (1987).

coefficients up to degree 2 for every 100-year interval between 400 and 1600 A.D. Figure 2 shows the time variations of the Gauss coefficients. Circles denote the results in the present study, and squares denote the results of spherical harmonic analysis of historical measurement data by YUKUTAKE (1987).

We are mainly interested here in interpolating the directional data, and we cannot say that the paleointensity data used in the present study are sufficient to estimate the global intensity variation precisely. The time variation of the intensity of the geomagnetic field is discussed in Section 4.

In Fig. 1, we calculated the geomagnetic field corresponding to the latitude and the longitude of each data site from thus obtained Gauss coefficients, and compared them with the original values. Although only Gauss coefficients up to degree 2 are considered, the calculated results well reflect the variations of the geomagnetic field observed there.

2.2 Correction of sediment data

Using thus obtained Gauss coefficients, the inclination and the declination at the sediment data sites were calculated. Then, by comparing with these curves, we examined the sediment data and corrected the time lags and the offsets of angles inherent in the inclination and the declination of them. The references for the adopted data set are as follows: Sweden (MÖRNER and SYLWAN, 1989), Isreal (THOMPSON *et al.*, 1985), West U.S.A. (VEROSUB *et al.*, 1986), East Central U.S.A. (CREER and TUCHOLKA, 1982b), East U.S.A. (KING *et al.*, 1983) and Argentine (CREER *et al.*, 1983). In Fig. 3, we show the corrected sediment data. The correction of the time lag was confined within 0 and 400 years, and the correction of the inclination was between -2 and 5 degrees to deepen the inclination. All the relative declinations were corrected to the absolute values.

No amplitude correction was made, due to partly that amplitude corrections were made in conventional methods in the original papers. We did not subtract the trends in declination, since the detrending did not make any significant difference in the succeeding analyses. Besides the above mentioned sediment data, we rejected several data sets from China and Africa because the difference between the data and the model field were too large. In addition, data from northeast Australia were also excluded because this method was not applicable due to the lack of the uppermost part of the sediments.

After all, sediment data from ten locations were used, including the previously mentioned data from Japan, England, Australia and Hawaii. Adding the very long archaeomagnetic results from Bulgaria (KOVACHEVA, 1980, 1992, personal communication 1992) and Ukraina (ZAGNIY, 1981; KOVACHEVA, 1982), a data set of the inclination and the declination at twelve sites were prepared (Fig. 3). Longitudal distribution of the data sites are shown in Fig. 4. Between 4000 B.P.

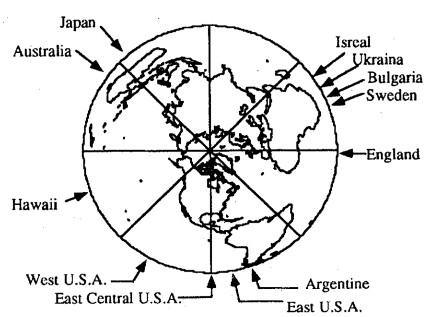

Fig. 4. Distribution of the data sites.

and the present, all the data at twelve sites are available. The data set is biased to the northern hemisphere of middle latitude, and, therefore, the following results mainly reflect the geomagnetic variation in this area. Only six data span back to 10,000 B.P., although the longitudes of them distribute well around the world.

3. Mean VGP Position and Geomagnetic Pole

Averaging of the VGP positions has been employed to determine the location of the geomagnetic pole (CHAMPION, 1980; MERRILL and MCELHINNY, 1983), because the mean of the VGP calculated from several points on the earth's surface is considered to approximate the location of the geomagnetic pole closely, by averaging out the non-dipole field. OHNO and HAMANO (1992) used this method and investigated the polar motion over the last 10,000 years. We revised the data from Bulgaria in the present study, in addition to including the new data from Hawaii, and calculated the geomagnetic pole position.

We calculated the VGP position at each data site for every 100-year interval from the prepared data set, and estimated the locations of the geomagnetic pole by averaging them. In Fig. 5-a, we plot all the positions of the north geomagnetic poles over the past 10,000 years. The

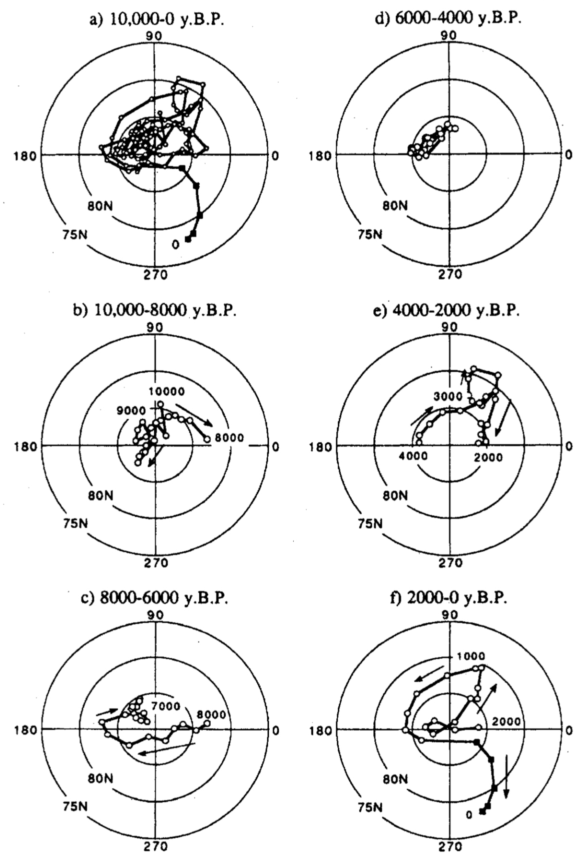

Fig. 5. The estimated location of the north geomagnetic pole (circle). Squares denote the results of the spherical harmonic analysis from the historical observation (YUKUTAKE, 1987). Numbers attached denote the year, and the arrows indicate the directions of the movement from the past to the present.

distribution of the geomagnetic pole over the past 10,000 years has, as a whole, an elliptic shape which is elongated parallel to the meridian of about 45 and 225 degrees. It is also noticeable that the geomagnetic pole moved predominantly in this direction, and the movement perpendicular to this direction was small. In contrast to these characteristics, during the last 400 years the geomagnetic pole moved largely in the direction perpendicular to the predominant direction.

Figures 5-b$\sim$f show the pole positions for every 2000-year interval, respectively. The results of the last 2000 years are similar to the results by MERRILL and MCELHINNY (1983). As shown by the arrows, it is noticeable in the time variation of the longitude that the westward movement was predominant prior to ca. 2000 B.P. and between 400 B.P. and the present. In contrast, from ca. 1800 B.P. to ca. 400 B.P., the geomagnetic pole moved eastward.

The time sequence of the polar motion can be divided into three intervals, the intervals between ca. 10,000 B.P. and ca. 7000 B.P., between ca. 7000 B.P. and ca. 3700 B.P., and between ca. 3700 B.P. and the present. Between ca. 10,000 B.P. and ca. 7000 B.P., the movement of the geomagnetic pole was active, in which it changed its position about 15 degrees. In contrast, during the period between ca. 7000 B.P. and ca. 3700 B.P., the movement of the geomagnetic pole was inactive. It stayed near the geographical pole. After ca. 3700 B.P. the geomagnetic pole moved largely to the circle of 80 degrees of north latitude.

4. Angular Dispersion of VGP and Dipole Moment

The intensity of the geomagnetic field prior to the historical measurements has been studied by measurements on archaeological materials and lava flows. MCELHINNY and SENANAYAKE (1982) estimated the dipole moment over the past 10,000 years by averaging paleointensity data for every 500-year interval between 0 and 4000 B.P. and for every 1000-year interval prior to that period. They used about equal numbers of virtual dipole moment (VDM) data and virtual axial dipole moment (VADM) data. These studies revealed the long period variations in the dipole moment such as the broad maximum from ca. 3000 B.P. to ca. 1000 B.P. and the low dipole moment prior to this period. But variations with short period were smoothed out in their data set. In the present study, we inferred the continuous time variation of the dipole moment from the directional data set, by using the angular dispersion of the VGP.

The VGP positions, calculated from inclination and declination values at several points on the earth's surface, disperse around the geomagnetic pole. A measure of the degree of the angular dispersion is the angular standard deviation (ASD) s defined as

$$s^2 = \frac{1}{n-1}\sum_{i=1}^{n}\delta_i^2 \tag{1}$$

where n denotes the number of measurements, and δ_i is the angular distance between each VGP and their mean position, which we regarded as an approximate geomagnetic pole in the previous section. By its definition, ASD is zero if the geomagnetic field is perfectly dipolar and, hence, all the non-dipole fields are zero. (We neglect here the dispersion due to the errors in measurements for simplification.) Then, as the ratio of the non-dipole field to the dipole moment (N/D ratio) becomes larger, the ASD also becomes larger. We can approximate tangent of the ASD to be proportional to the N/D ratio, provided that the ASD is small. That is,

$$\tan(s) = k\frac{N}{D} \tag{2}$$

where D and N denote the dipole moment and the total non-dipole field, respectively, and k is a constant. We define N to be the square root of the square sum of all the Gauss coefficients except those of degree 1.

We calculated the ASD of VGP from the present data set for every 100-year interval (Fig. 6). In the time variation of ASD, several sharp troughs are noticeable on the trend of stable decrease with time. Especially, the sharp trough around 8500 B.P. is remarkable. Low angular dispersions suggest that the geomagnetic field is dipolar, which depends on either high intensity of the dipole moment or low intensity of the non-dipole field. In Fig. 7, we compared the curve of cotangent of ASD to the VDM curve by MCELHINNY and SENANAYAKE (1982). Both the two curves show similar trends, which suggests that the non-dipole field was mostly constant and, therefore, the variation of ASD mainly reflects the variation of the dipole moment. The similarity is clear in the correlation plot of their trends (Fig. 8), in which we compared the inverse of the VDM and the tangent of the ASD according to Eq. (5) in the following discussion. The tangent of ASD was averaged in the same intervals as those of VDM averaging.

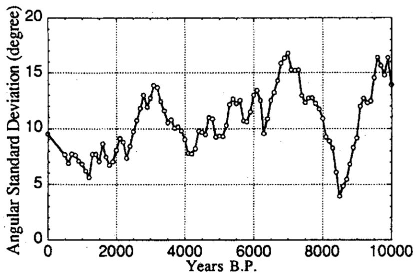

Fig. 6. Time variation of the angular standard deviation (Eq. (1)) calculated for every 100-year interval.

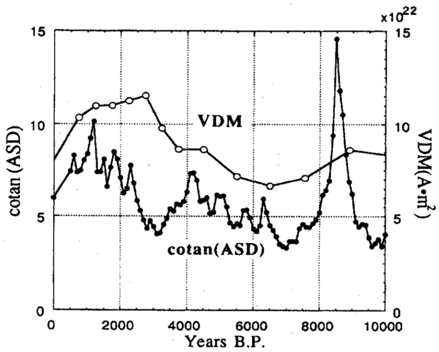

Fig. 7. Comparison of the virtual dipole moment (VDM) and the angular standard deviation (ASD). Solid circles denote the cotangent of ASD. Open circles denote the VDM by MCELHINNY and SENANAYAKE (1982), of which the ages were transformed to coincide with the age of the ASD.

We introduce here a model in which N is combined to D by a linear relationship, because N is much smaller than D. In the model, both D and N have their origins D_0 and N_0, and fluctuate from the origins by ΔD and ΔN, and ΔN is proportional to ΔD. That is,

$$\begin{aligned} D &= D_0 + \Delta D, \\ N &= N_0 + \Delta N, \\ \Delta N &= a\Delta D, \end{aligned} \tag{3}$$

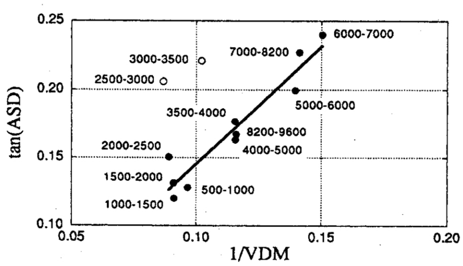

Fig. 8. Correlation plot between $(\text{VDM})^{-1}$ and the tangent of the ASD according to Eq. (5). Numbers attached to the circles denote the period. The line gives the linear fit to the data except for the those between 3500 B.P. and 2500 B.P. (open circles).

where a is a constant. Then,

$$N = aD + b \tag{4}$$

where

$$b = N_0 - aD_0.$$

Inserting Eq. (4) into Eq. (2),

$$\tan(s) = B\frac{1}{D} + A \tag{5}$$

where

$$A = ka,$$

$$B = k(N_0 - aD_0)$$

is derived. We determined the coefficients A and B from the linear regression in the correlation plot (Fig. 8), in which the data between 3500 B.P. and 2500 B.P. were excluded. In this period, the dipole moment was inclined toward the European region (defined as the northern hemisphere from 0–90°E) as shown in Fig. 5-e and, hence, the VADM calculated from the paleointensities in this region was higher than the true dipole moment. Since the distribution of the paleointensity sites used by McElhinny and Senanayake (1982) is biased to the European region, we consider that the dipole moment calculated from their data set may have given spurious high values for these periods.

Inserting thus obtained A and B into Eq. (5), we calculated the continuous variation of the dipole moment from the ASD for every 100-year interval (Fig. 9). The results of the present study coincide with the VDM and, in addition, reveal the short period variations. The coefficient A was calculated to be minus 0.03, and Ohno (1991) estimated the constant k to be 1.0 from the globally distributed observatory data between 1945 and 1980. Therefore, the coefficient a is calculated to be minus 0.03, which means that the total non-dipole field was mostly constant (Fig. 9).

In the time variation of the dipole moment in Fig. 9, the most remarkable feature is the peaks around 8500 B.P., around 4200 B.P., and around 1200 B.P. Several paleointensity studies have reported high intensity values in these periods. High paleointensities were reported between 10,000 B.P. and 8000 B.P. in Japan (Tanaka, 1990), France (Salis *et al.*, 1989), Iceland (Schweitzer and Soffel, 1980) and Turkey (Bucha and Mellaart, 1967). Around 4000 B.P., high paleointensities were reported in Hawaii (Coe *et al.*, 1978) and Japan (Hirooka and Sakai, 1991). In these studies, during the periods when high paleointensity values are reported, low paleointensity values are also reported. These results suggest that the periods of high intensity were very short, and the discrete studies such as those from lava flows could not distinguish the short periods of

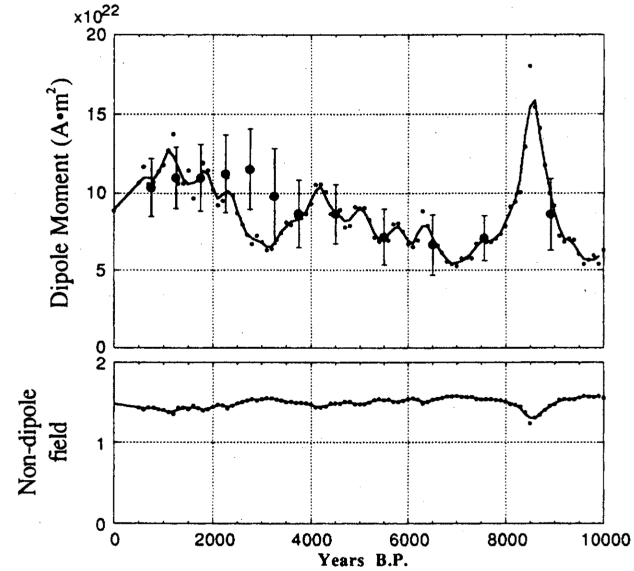

Fig. 9. Time variation of the dipole moment (upper diagram) and the total non-dipole field (lower diagram). Small circles show the calculated results, and the lines indicate their three-point running mean. Solid circles with error bars show the VDM by MCELHINNY and SENANAYAKE (1982).

high intensity from the periods of low intensity before and after the peaks, due to insufficient datings. In the present analysis, we obtained the continuous variation of intensity and made the precise variation clear, because sediment data are continuous and, hence, the order of events is obvious. But the absolute value of the peaks of the high intensity around 8500 B.P. is not so significant, because this is an indirect method and based on several assumptions.

5. Conclusion

Global features of the geomagnetic field over the past 10,000 years were investigated by analyzing the VGP calculated from the data set of twelve localities, in which the deficiency of sediment data were corrected by comparing with archaeomagnetic and lava data.

The distribution of the geomagnetic pole has an elliptic shape and westward movement was predominant throughout the interval. The time sequence of the polar motion can be divided into three intervals. During the period between ca. 7000 and ca. 3700 B.P., the movement of the geomagnetic pole was inactive, and it was active before and after this period, fluctuating over 10 degrees.

We calculated angular dispersion of the VGP for every 100-year interval, and inferred the time variation of the dipole moment by comparing with paleointensity data. The angular dispersion was small for several hundred years around 8500 B.P., 4200 B.P. and 1200 B.P., which suggests sharp peaks of high intensity of the dipole moment considering the high intensity around these periods reported by paleointensity studies. It is interesting to note that the movement of the geomagnetic pole became active after the high dipole moment around 4200 B.P.

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