TIME VARIATIONS IN GEOMAGNETIC INTENSITY

Jean-Pierre Valet
Institut de Physique du Globe de Paris, Paris, France

Received 15 May 2001; revised 8 October 2001; accepted 8 October 2002; published 19 April 2003.

[1] After many years spent by paleomagnetists studying the directional behavior of the Earth's magnetic field at all possible timescales, detailed measurements of field intensity are now needed to document the variations of the entire vector and to analyze the time evolution of the field components. A significant step has been achieved by combining intensity records derived from archeological materials and from lava flows in order to extract the global field changes over the past 12 kyr. A second significant step was due to the emergence of coherent records of relative paleointensity using the remanent magnetization of sediments to retrace the evolution of the dipole field. A third step was the juxtaposition of these signals with those derived from cosmogenic isotopes. Contemporaneous with the acquisition of records, new techniques have been developed to constrain the geomagnetic origin of the signals. Much activity has also been devoted to improving the quality of determinations of absolute paleointensity from volcanic rocks with new materials, proper selection of samples, and investigations of complex changes in magnetization during laboratory experiments. Altogether these developments brought us from a situation where the field changes were restricted to the past 40 kyr to the emergence of a coherent picture of the changes in the geomagnetic dipole moment for at least the past 1 Myr. On longer timescales the field variability and its average behavior is relatively well documented for the past 400 Myr. Section 3 gives a summary of most methods and techniques that are presently used to track the field intensity changes in the past. In each case, current limits and potential promises are discussed. The section 4 describes the field variations measured so far over various timescales covered by the archeomagnetic and the paleomagnetic records. Preference has always been given to composite records and databases in order to extract and discuss major and global geomagnetic features. Special attention has been devoted to discussing the degree of confidence to be put in the data by considering the integration of multiple data sets involving different techniques and/or materials. *INDEX TERMS:* 1521 Geomagnetism and Paleomagnetism: Paleointensity; 1560 Geomagnetism and Paleomagnetism: Time variations-secular and long term; 1503 Geomagnetism and Paleomagnetism: Archeomagnetism; 1535 Geomagnetism and Paleomagnetism: Reversals (process, timescale, magnetostratigraphy); *KEYWORDS:* paleointensity, paleomagnetism, magnetization, secular variations, archeomagnetism, rock magnetism
Citation: Valet, J.-P., Time variations in geomagnetic intensity, *Rev. Geophys., 41*(1), 1004, doi:10.1029/2001RG000104, 2003.

1. INTRODUCTION

[2] The first magnetic compass was invented 27 centuries ago using a piece of magnetite carved as a spoon. Magnetic directions were measured 1 century later by the Greek philosopher Thales [*Needham*, 1962]. Thus there were 22 centuries between these early measurements of the geomagnetic field and the first direct magnetic observation. At the Earth's surface the present geomagnetic field is roughly equivalent to the field created by a magnetic dipole located at the Earth's center and inclined by 11° on the rotation axis. When averaged over several thousands of years, it becomes equivalent to the field of a centered axial dipole aligned along the rotation axis.

[3] Our knowledge of the geomagnetic field variability was first linked to the direct measurements of the magnetic observatories. They allowed us to map the present field with a resolution of a few tens of kilometers and to describe the most rapid changes occurring over the past 400 years. No information was gained for longer timescales until the emergence of the first archeomagnetic and paleomagnetic measurements in the middle of the last century. Over a long period, paleomagnetists have gathered data from around the world with the aim of covering the geomagnetic spectrum of paleosecular variations in the past. Reversals became a primary subject of interest given their extraordinary character and the need to establish the polarity timescale. When considering hundreds of millions of years, the geomagnetic polarity can change every 0.2 Myr, or it may be stable over periods as long as 80 Myr. However, restraining the geomagnetic history to the short periods (between 1 and 10 kyr at most) covered by field reversals would be very limiting. Reversals must be considered within the entire spectrum of field changes. For similar reasons it is critical to deal with the entire field vector and not just its

8755-1209/03/2001RG000104*$15.00*
doi:10.1029/2001RG000104

directional changes, which might be equivalent to studying climate without considering insolation.

[4] Despite its importance, little attention was initially paid to vector intensity. In fact, the first records were dedicated to absolute paleointensity from archeological materials. The technique was reliable but time-consuming and had a relatively small success rate. Similar experiments using lava flows offered the possibility of exploring the field variations over a much longer period, but they were disclaimed by certain authors [*Thellier*, 1941]. This objection yielded substantial improvements in rock magnetic measurements as well as in dating techniques. The situation changed drastically during the past 10–15 years with significant developments in obtaining relative paleointensity signals from sediments, new perspectives for absolute paleointensity signals from volcanic materials, and the emergence of excellent and precisely dated archeomagnetic records.

[5] The first paleomagnetic studies of paleointensity were devoted to scrutinizing the evolution of the field vector during reversals with the aim of constraining the morphology of the field during the transitional period between the two polarities. It was rapidly established that the field intensity drops considerably, an observation that has been confirmed almost systematically. Because the present field intensity has been decreasing for the past 150 years, it is sometimes speculated that we could well be approaching a reversal. Such an assertion could merit consideration if considered within the long-term variability of the field. This is evidently a good reason for exploring the field changes in the past with as much precision as possible. It has now become possible to constrain field intensity changes over several million years with good precision and to correlate major features recorded from distinct and independent data sets. Besides reversals most, if not all, field structures revealed by various geological records are associated with significant changes in intensity. In fact, all geomagnetic events appear to be intrinsically linked to intensity and could even be seen as resulting from such changes. It is clear that better understanding of the processes that govern the geodynamo will be gained by combining theory and observation. In this respect, recent emergence of numerical simulations of dynamos has provided an alternative way of studying geomagnetic changes. Long data sets are necessary to evaluate theoretical predictions.

[6] The past 10 years have also seen the emergence of global approaches linking various studies of the Earth envelopes. The geomagnetic field is no longer seen as a separate entity; it depends on processes that are coupled to each other. The influence of the mantle on fluid motions in the upper part of the core is a typical example that can only be documented by long-term field changes. Screening of cosmic rays by the field and its influence on production of cosmonuclides requires a combination of geochemistry and magnetism. The suggestion of a relationship between core motions and orbital variations of the Earth rotation axis requires exploration of the spectral content of the field intensity variations. New stratigraphy with unequal resolution for sedimentary sequences based on the recognition of specific features of field intensity changes and various aspects related to rock magnetism are other subjects of interest that are related to paleointensity studies.

[7] In order to extract as much information as possible, homogeneous and high-quality measurements are needed. This first aspect is related to knowledge of the magnetization processes. In the case of volcanic rocks this is linked to rock magnetism and magnetic mineralogy. In the case of sedimentary rocks the situation is more delicate as it depends on conditions prevailing during and after deposition and thus on physical sedimentary parameters as well as on chemistry and mineralogy. The second critical parameter is accuracy in temporal resolution. Uncertainties in dating can generate significant discrepancies and/or introduce large distortions between adjacent records. All these aspects will be extensively discussed in this paper.

[8] Two reviews about paleointensity by *Tauxe* [1993] and *Jacobs* [1998] have been published previously. The first review was dedicated more to methodology, but it also reported interesting results as well as limits and perspectives for future studies. The second review described the construction of the first database, which followed precursory records of relative paleointensity from sediments. First, I will consider the various ways of studying paleointensity and describe some promising technical and methodological developments. Next, I will focus on several new and fascinating geomagnetic features that emerged from the various contributions of the past decade. These results will be presented along with a critical discussion of the overall quality of the records, their resolution in time, and the adequacy of the techniques used. It is obvious that challenges for future studies can be drawn from the present data set as well as from several ongoing and interesting controversies.

2. HISTORICAL FIELD, TIME CONSTANTS, AND RESOLUTION

[9] The main geomagnetic field is subject to changes in its direction and strength on timescales ranging from years to millennia and even millions of years. The first measurements of magnetic field intensity could be due to E. P. E. De Rossel. He was an officer of the 1791–1794 expedition of d'Entrecasteaux, which was searching for the missing La Pérouse expedition [*Merrill et al.*, 1996], and published a narrative description of the trip including some scientific experiments. The six magnetic measurements performed by De Rossel indicated that the field intensity decreases with latitude. However, methods for measuring the total field intensity were not developed before J. C. F. Gauss in 1832. Today an array

of about 150 worldwide geomagnetic observatories continuously record field intensity changes.

[10] Beyond this period a data set has been constructed by assuming that there is a linear relationship between the remanent magnetization carried by archeological materials for the past 10 kyr or by rocks (mostly sediments or volcanics) for the past million (even billion) years and the geomagnetic field that existed when they were formed. The Earth's magnetic field likely never exceeded 100–150 μT, weak enough to make this approximation correct. In order to compare archeomagnetic and paleomagnetic records from different locations and also to analyze long-term field changes it is convenient to refer to an equivalent dipole moment [*Smith*, 1967; see also *Merrill et al*., 1996]. This is basically the same approach as with the virtual geomagnetic pole used for directions, which relies on the assumption that the field vector measured at each site location was generated by a geocentered dipole. If the inclination is available, the virtual dipole moment (VDM) is the most appropriate representation, because it takes into account the tilt of the dipole axis with respect to the rotation axis. Should the field be perfectly dipolar, identical VDM values with identical short-term variations, if any, would be observed everywhere. In fact, it is well known that the dipole moment has decreased at a rate of ~5% per century since the time of the Gauss analysis in 1835. The historical values of the mean VDMs recorded at five observatories during the past 120 years (Figure 1a) vary from 7.5 to 9.2×10^{22} A m^{-2}, which indicates the strong influence of local nondipole components. The enlarged picture in Figure 1b shows different variations for the Australian and French VDM data sets, which are a consequence of time-varying nondipole components.

[11] When there is no inclination record available, archeomagnetists and paleomagnetists use the virtual axial dipole moment (VADM). This calculation assumes that the field was produced by an axial geocentric dipole, which is appropriate if the time period covered by the calculation is long enough (i.e., on the order of at least several millennia, see below). However, for instantaneous or rapid recording acquired during cooling of archeomagnetic material or lava flows, this can generate even larger differences than with VDMs and misinterpretations when comparing the field strength recorded from different areas. The mean VDMs from the five observatories (Figure 1c) do not differ by >20%, but the difference can reach almost 45% between the VADMs. However, this last value is essentially due to the strong Australian field intensity, which reflects the existence of a strong anomaly in this area. The present global dipole moment is 8×10^{22} A m^{-2}.

[12] As field paleointensity studies explore various timescales, from historical to archeomagnetic and paleomagnetic, it is critical to know which features found on the short timescales persist on the longer ones. On the basis of historical and archeomagnetic data, *Hulot and Le Mouël* [1994] and *Hongre et al*. [1998] have shown that terms higher than degree 2 are rapidly averaged out over 150 years. The equatorial dipole is thought to persist over ~500 years, whereas similar analyses over longer time series of several thousand years are needed to assess the stability of these characteristic times. When the field is averaged over 2000 years or more, only the axial dipole remains, to which is added an axial quadrupole term of the order of 5% of the axial dipole [*Merrill and McElhinny*, 1977; *Coupland and Van der Voo*, 1980; *Constable and Parker*, 1988; *Schneider and Kent*, 1990; *Quidelleur et al*., 1996; *Johnson and Constable*, 1997; *Carlut and Courtillot*, 1998]. The existence of higher-order nonzonal terms cannot be totally excluded [*Gubbins and Kelly*, 1993; *Kelly and Gubbins*, 1997; *Johnson and Constable*, 1998] but, if confirmed, would be restricted to rather limited areas, such as the southwestern Pacific [*Constable et al*., 2000; *Elmaleh et al*., 2001].

[13] These observations have important bearings on studies of paleointensity, which essentially rely on two kinds of magnetizations with very different acquisition processes and timings. Thermoremanent magnetization is associated with relatively rapid cooling of magnetic grains that are contained within archeological material or volcanic rocks. Thermoremanence is acquired over a few days for archeological objects or over a few years for volcanic rocks, so that the remanence retains a record of the nondipolar and dipolar contributions of the field at the site location. This information is similar to that provided by direct field measurements from magnetic observatories. Since the process of magnetization acquisition can almost be simulated in the laboratory by heating and cooling (but with a very different rate) the rock samples in the presence of a known field, successful experiments of paleointensity provide us with the absolute value of the total field at the site location. In contrast, the slow acquisition process of magnetization in sediments is long enough to average out the most rapidly varying nondipole components (deposition rates in deep-sea sediments are usually between a few millimeters and a few centimeters per kiloyear, which gives a resolution between 1 and 10 kyr for a typical 2-cm-long sample). In this case the detrital remanent magnetization is predominantly sensitive to the axial dipole field and/or to any possible long-term persistent components, provided they are strong enough (note, however, that the 5% quadrupole contribution reported above lies within noise inherent to most paleointensity records). Owing to complexities inherent in the magnetization processes in sediments, which are dependent on rock magnetic characteristics and typical physical properties (density, porosity, granulometry, mineralogy, etc.), the remanence acquisition cannot be duplicated, so that only relative field intensity can be extracted. Before dealing with geomagnetic characteristics that have been observed from records that belong to these two categories of remanence, it is important to discuss the techniques and methods used to retrieve the paleointensity signal.

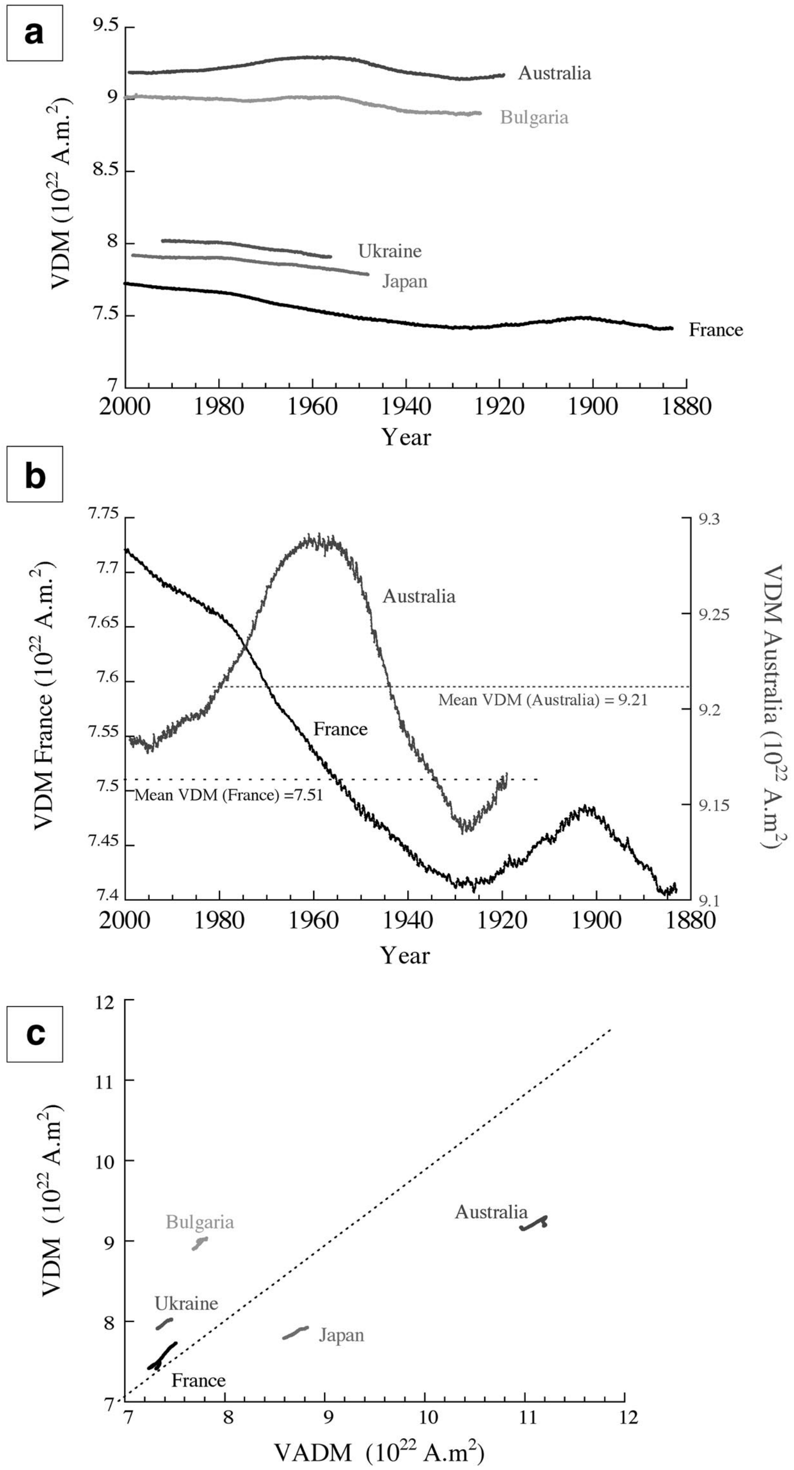

Figure 1. (a) Virtual axial dipole moments (VADM) derived from field measurements at five geomagnetic observatories since 1880. (b) Enlarged view of the virtual dipole moments (VDM) recorded in France and Australia since 1880. The differences between the two records are caused by nondipole components, which are particularly large over Australia. (c) VDM versus VADM values for the five observatories. As expected, there are larger differences between the VADM values, particularly for Australia. These plots indicate the importance of time averaging over at least several hundred years for any meaningfull interpretation of the variations of the Earth's dipole moment.

3. TECHNIQUES AND DEVELOPMENTS

3.1. Absolute Paleointensity

3.1.1. Thellier-Thellier Experiments and Derivatives

[14] Paleointensity techniques involving a series of heatings in a known field can be applied to materials that acquired their magnetization while cooling from high temperatures. The *Thellier* [1941] technique was initially proposed for archeological materials and later extended to volcanic rocks. It relies on a direct comparison between thermal demagnetization of the natural remanent magnetization (NRM) (original magnetization that was acquired during cooling of the material) and acquisition of laboratory partial thermoremanent magnetization (pTRM) in the presence of a known field. The NRM and TRM components are separated by addition of the vectors measured at the same temperature steps. This operation is repeated at incremental temperature steps [*Levi*, 1977] up to the Curie point. The current method of determining the final field value is to calculate the slope of an "Arai" plot [*Nagata*, 1963], which depicts (Figure 2a) the amount of NRM lost as a function of the pTRM gained for each successive temperature interval (in principle, both should keep the same proportion for every temperature interval). In the original Thellier technique, two TRMs with opposite directions are produced in the same field at each temperature step. *Coe* [1967] modified this approach and proposed to first demagnetize the NRM in zero field and then to remagnetize the sample. However, measuring the NRM first does not allow one to detect remagnetization components that could be produced during the subsequent heating in the presence of the field and carried by grains with high blocking temperatures (unless a third heating is performed in zero field to detect alteration). In order to solve this problem it would be preferable to impart the TRM before heating the sample in zero field [*Aitken et al.*, 1988; *Valet et al.*, 1998]. The two techniques are used with the same rate of success, although preference is frequently given to the classical Thellier procedure.

[15] In principle, NRM and TRM of single-domain (SD) grains of magnetite retain the same proportionality in the applied field over the entire temperature spectrum [*Dunlop and Özdemir*, 1993]. In fact, the procedure is possible because of the additivity of the TRMs, which stipulates that the sum of TRMs acquired within successive temperature intervals is equal to the total TRM for the entire interval. However, chemical and mineralogical transformations can affect magnetic grains during heating, and thus the laboratory-induced TRM does not necessarily involve the same grains as the NRM [*McClelland*, 1996]. In order to detect these problems, *Thellier and Thellier* [1959] introduced the pTRM checks that test for duplication of a TRM at one or several previous temperatures. If the new value is different from the previous one, then magnetic grains were affected during the previous heating steps at higher temperatures. Unfortunately, the pTRM checks increase the number of heatings, making the full experiment very time-consuming, but they are unavoidable and necessary to detect alteration. However, in some cases, remagnetization can produce grains with blocking temperatures higher than the heating step. In this case the NRM itself can be affected and partly lost, but the corresponding changes will not necessarily be detected by pTRM checks. Such a situation is illustrated by the Arai plots [*Nagata*, 1963] shown in Figure 2. There is a linear relation between the NRM and the TRM, but the directions of the NRM fail to go through the origin of the demagnetization diagrams when plotted in sample coordinates. This is caused by remagnetization in the field direction of the oven. *Kosterov and Prévot* [1998] made the point that irreversible changes due to alteration can occur even at moderate temperatures. This was observed by measuring hysteresis loops after different heating steps following a previous suggestion by *Haag et al.* [1995]. These changes were not always detected by standard pTRM checks, which again emphasizes the need for multiple and complementary verifications, at least for a limited number of samples per lava flow.

[16] The success rate of paleointensity experiments frequently does not exceed 10–20%. Consequently, it is important to realize that ~50 samples per lava flow should be measured in order to obtain 10 independent field determinations. This reinforces the importance of preliminary investigations aimed at discarding inappropriate specimens. Indeed, the most efficient way to increase the number of determinations by increasing the success rate is to deal with a subset of suitable samples that were selected using non-time-consuming techniques. Note, however, that most experiments must be conducted on adjacent specimens (but from the same core), and there is always a little risk that inhomogeneous magnetic characteristics alter the quality of the selection. It is important to discern techniques for selection that involve directly the NRM and the TRM from other indirect approaches that rely on rock magnetic parameters. Hysteresis measurements and thermomagnetic analyses (measurement of progressive loss of induced magnetization while heating up to Curie temperature and its recovery during subsequent cooling) provide interesting information, but they are not self-sufficient and should be combined or used with other indicators. The "direct" approaches rely mostly on a comparison between characteristics of the NRM and TRM before and after heating at some critical temperature steps. Among other possibilities, one can compare the characteristics of demagnetization by alternating field (af) of the NRM with those of a full TRM (acquired during cooling from above the Curie temperature to room temperature). We will see in section 3.1.2 that other experiments are also efficient.

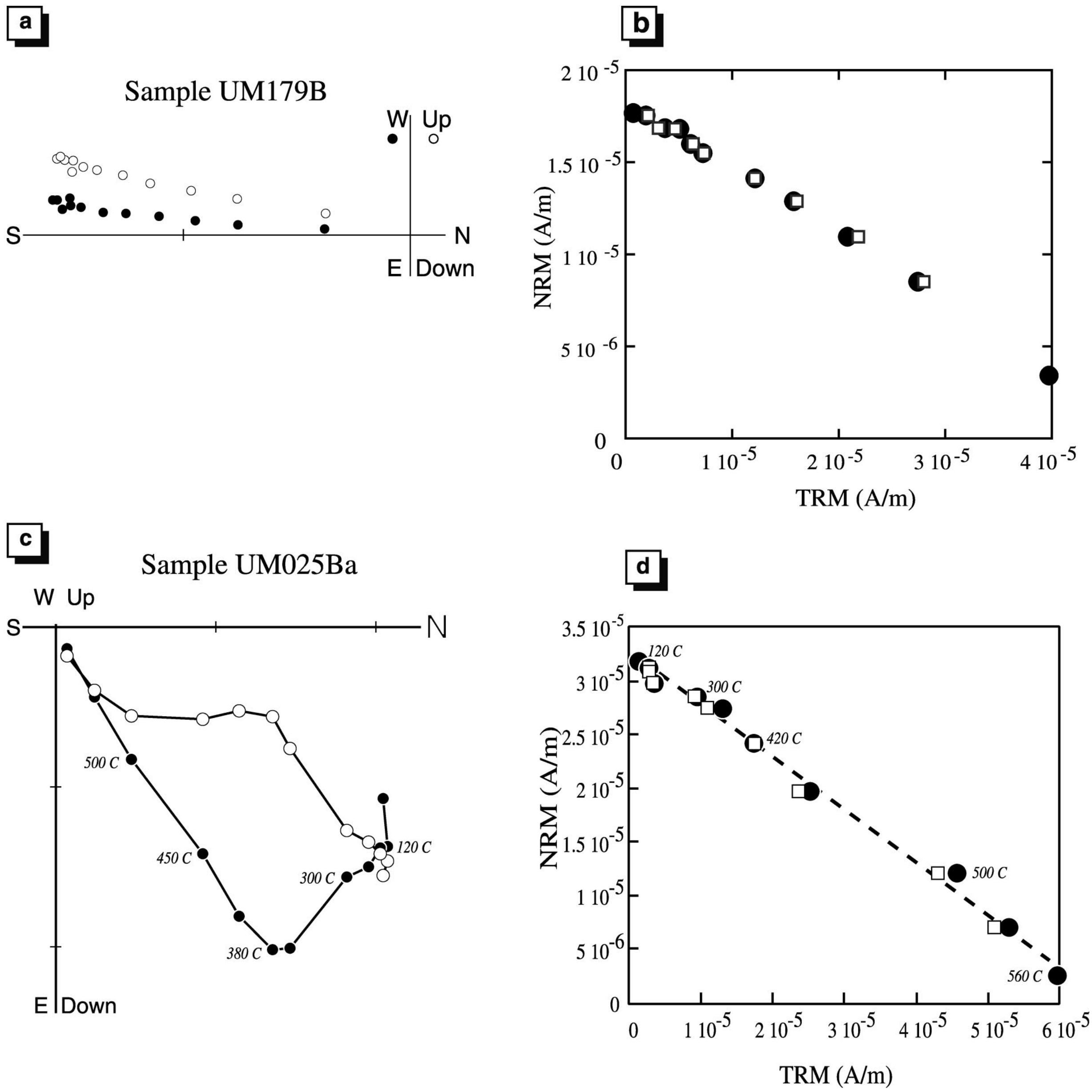

Figure 2. Typical plots used for experiments of absolute paleointensity. (a) Thermal demagnetization diagram of natural remanent magnetization (NRM) with a well-defined high-temperature magnetization component that linearly decreases to the origin. (b) Arai plot [*Nagata*, 1963] for the same sample as in Figure 2a constructed from stepwise demagnetization of the NRM and acquisition of thermoremanent magnetization (TRM) (dots). Because proportionality between the two magnetizations is kept constant throughout the experiment, the plot is characterized by a straight line. The slope of the line indicates the factor by which the oven field must be multiplied to obtain the paleofield value. In addition, pTRM checks (open squares) are performed one step down from the last heating to compare the TRM before and after heating at higher temperature. Positive checks should not differ by more than 5% from the previous TRM value to consider that no significant changes in magnetic mineralogy or granulometry affected the paleointensity experiment. (c) Increasing production of a component in the direction of the oven field during the successive heatings. This is depicted by the demagnetization diagram that does not go through the origin but moves toward the oven field direction. (d) Same sample as in Figure 2c despite significant remagnetization. The Arai plot is quite acceptable with a well-defined straight line and positive pTRM checks. However, the directions show that a component has been acquired in the oven. This emphasizes the importance of considering the entire vector to interpret paleointensity experiments.

3.1.2. Problems Inherent to Multidomain Grains

[17] Another aspect to consider is the behavior of multidomain (MD) and pseudo-single-domain (PSD) grains. Significant progress has been made in this direction. The presence of MD and PSD grains is quite common in lava samples, and they can have important

consequences. A complete review concerning the theoretical and phenomenological developments in this matter is given by *Dunlop and Özdemir* [1997]. Multidomain and PSD grains do not obey the experimental law of additivity of partial TRMs acquired in nonoverlapping temperature intervals by single-domain grains. A direct consequence is the existence of two different slopes in the NRM-TRM plots between low (<350°C) and high (350°–580°C) temperatures. Such behavior is caused by the difference between the blocking temperature (T_b) (temperature required to randomize the orientation of a portion of magnetic grains and thus to create a pTRM during cooling in presence of field) and the unblocking temperature (T_{ub}) (temperature needed to randomize the orientation of the magnetic grains with magnetization acquired at T_b) of MD grains. Heating at temperature T_i produces a wide range of T_{ub}, both lower and higher than T_b, because of repeated domain wall jumps during heating. The anomalous high T_{ub} form a tail extending to the Curie point, which has been widely discussed in the literature [*Bol'shakov and Shcherbakova*, 1979; *Worm et al*., 1988; *Halgedahl*, 1993; *McClelland and Sugiura*, 1987; *McClelland and Shcherbakov*, 1995; *McClelland et al*., 1996]. In a recent paper, *Dunlop and Özdemir* [2000] described how low T_{ub} tails can explain the typical concave shape of the Arai diagrams inherent to multidomain grains. Several theoretical and experimental studies [*Bol'shakov and Shcherbakova*, 1979; *Dunlop and Xu*, 1994; *Xu and Dunlop*, 1994] have shown that demagnetization of the NRM can begin at low temperatures, while a significant part of TRM is acquired at high temperatures. *Dunlop and Özdemir* [2000] suggested that a significant part of the NRM with low T_{ub} can be easily removed, whereas pTRM acquisition is delayed by the presence of high T_b. Thus NRM thermal demagnetization will greatly outweigh pTRM acquisition, and this effect persists until the NRM is about half demagnetized.

[18] The presence of MD grains can be detected in curves of continuous or thermal demagnetization by their characteristic tails [*Bol'shakov and Shcherbakova*, 1979]. Following experiments by *Vinogradov and Markov* [1989], *Shcherbakova and Shcherbakov* [2000] noticed that TRM acquired between room temperature and 300°C after heating to the Curie temperature cannot be removed completely by heating in zero field in the presence of PSD and MD grains, while this is not the case for monodomains. The proportion of magnetization left is a measurement of the tail and depends on the amount of MD grains. This approach can thus be recommended provided that standard hysteresis parameters confirm the diagnostic. It is frequent that a part of the NRM-TRM spectrum be touched by multidomain grains, which in this case are not clearly reflected by concave up Arai diagrams. *McCLelland and Briden* [1996], *Valet et al*. [1996], and *Dunlop and Özdemir* [1997] suggested comparing the NRM measured at T_i before and after heating in the presence of the field. Note that this "NRM check" can only be measured using the Coe version of the Thellier technique, in which the first heating is done in zero field, the second heating is done in the presence of field, and the third one is done in zero field again (which is opposite to the protocol that involves the TRM first to detect alteration but is fully justified with a second NRM measurement). *Riisager and Riisager* [2000] recently demonstrated the efficiency of the test (that they renamed "pTRM-tail check") on Paleocene lavas from the Faeroe Islands as well as on baked sediments. All these results demonstrate that concave diagrams must not be used to extract paleointensity information and that it is important to use appropriate tests to detect MD grains.

[19] In addition to a low success rate caused by frequent irreversible behavior of TRM dependence upon heating, paleointensity experiments are extremely time-consuming. Various approaches have been tentatively proposed, either to speed up the experiments or to restrain or correct undesirable effects produced during heating. The most recent and significant aspects are summarized in section 3.1.3.

3.1.3. Changing the Experimental Protocol

[20] In an attempt to reduce the number of heatings, *Kono and Ueno* [1977] proposed applying a field perpendicular to the NRM direction in order to extract the NRM and the TRM by performing a single heating at each step. In addition to technical difficulties positioning the sample within the oven, Kono and Ueno noticed that this approach requires exceptionally stable components and has greater experimental errors. *Hoffman et al*. [1989] suggested that one method of reducing alteration is to split the sample into a number of subsamples (usually about 10), each heated to one particular temperature, once in zero field and again in a known field. The NRM and the pTRM values are then normalized using the initial intensity of the subsample. This approach relies on the assumption that subspecimens from the same sample are characterized by similar magnetic properties, which is not always true. The natural physical properties of a lava often vary over a few centimeters. *Sherwood* [1991] evaluated the multispecimen approach and found no improvement with respect to the classical Thellier technique and no evidence for reduced alteration.

[21] Several suggestions have been made to increase the number of determinations. *Valet et al*. [1996] proposed a correction technique taking into account the deviations caused by creation or destruction of magnetization. Corrections rely on the hypothesis that the loss (or gain) of TRM induced by mineralogical transformations is indicated by the difference between the pTRM check and the TRM that was initially measured at the same step. In a detailed paper aimed at detecting alteration with high unblocking temperatures, *McClelland and Briden* [1996] reached basically the same conclusions. Corrections can be applied if (1) the NRM was not

affected and (2) the alteration product has unblocking temperatures lower than the temperature of the last heating. These two conditions can be checked by (1) double demagnetization of the NRM (if NRM demagnetization was performed first) and/or (2) scrutinizing the evolution of the NRM in sample coordinates. The suitability of corrections was tested on contemporaneous lava flows. The technique was reasonably successful and gave field determinations within a typical error of 10–20% inherent to paleointensity studies. However, it is frequently questioned because there is no clear indication about the reliability of field determinations. Corrections remain speculative if no other determination can be obtained from the same flow without corrections. In fact, the success rate of corrections mostly depends on the production of components with T_{ub} higher than the most recent heating step at T, which, as noted in section 3.1.2, happens frequently, and may be difficult to distinguish from components with T_{ub} lower than T.

3.1.4. Pretreatment Using Alternating Field Demagnetization

[22] Partial cleaning by af prior to thermal treatment was attempted first by *Coe* [1967] to remove unwanted secondary components of magnetization for the NRM, in particular the contribution of multidomain grains, which have high unblocking temperatures but rather soft resistance to alternating fields. Coe noticed that for three out of four samples the NRM-TRM variations were more concave upward than for af-untreated specimens. However, these last ones were not associated with linear diagrams, and thus the experiment was not fully convincing. In a second experiment, *Coe and Grommé* [1973] considered that the values found were also not as accurate as those without af cleaning. In fact, these results indicated that there was no improvement for curves with a double slope (e.g., multidomain grains). However, the experiments did not incorporate pTRM checks and could thus be dealing with samples that exhibited mineralogical changes. Different conclusions were reached by *McClelland and Briden* [1996], who defended the position that the af pretreatment of their samples did not change the blocking temperature distribution of the characteristic component. Additional insight on this matter was recently provided by *Dunlop and Özdemir* [2000], who reported that 10-mT af precleaning on 20-μm magnetite efficiently removes remanence with anomalously low and high T_{ub}, so that subsequent thermal cleaning is more single-domain-like. Although the changes may be dependent on grain sizes, the approach merits further testing on natural samples.

[23] *Tanaka et al.* [1995a] reported that direct measurements of the total vector at high temperatures can give reliable determinations, while the standard method does not. The sample was heated in a cryogenic magnetometer with a small furnace installed immediately outside one of the access holes, and the measurements were done within the magnetometer. Tanaka et al. noticed that the drop in temperature did not exceed 1°C at 400°C when moving the sample from the furnace into the magnetometer. The technique involves two successive heatings at temperature (with and without field) over successive temperature intervals and does not require cooling the sample to room temperature except at $T_{i\text{-}1}$. The operation is repeated at incremental steps and allows determination of the NRM lost and the TRM gained within each successive temperature interval rather than in a cumulative manner. Tanaka et al. preferred to perform their experiment directly at high temperatures. This interesting technique was hardly used and is certainly worthy of being developed and tested again. The difficulties probably reside in various technical aspects such as the uncommonly large size of the pass-through system used to insert the sample, technical difficulties for construction of the oven as well as the sample holder.

3.1.5. Heating by Microwave Excitation

[24] Instead of using a conventional furnace, *Walton et al*. [1992, 1993] proposed that the spin waves generated by heat in the form of phonons can also be generated within the magnetic grains by direct microwave excitation, which heats the magnetic grains slightly as the spin waves degenerate into phonons. The TRM gained by this method was described as almost identical to one gained by "natural" heating. The main advantage is that the bulk matrix of the sample is not significantly heated, the magnetic grains are slightly heated, and the system that produces the magnetization is heated and cooled very quickly (a few seconds). This approach would offer considerable potential for paleointensity studies. The protocol is similar to the *Coe* [1967] version of Thellier experiments. Therefore there must be a constant relationship between the amount of magnetic grains involved in the NRM demagnetized at any successive microwave power and those involved in the temperature intervals that were related to the initial magnetization of the sample. This is probably the most intriguing question that needs to be answered in order to validate the technique and the underlying theoretical developments. The next question to investigate is whether such relationship remains valid for any kind of sample.

[25] *Shaw et al*. [1996] studied Peruvian ceramics and observed that the microwave results reduced the scatter of the conventional archeomagnetic techniques. In a subsequent study of Chinese ceramics, *Shaw et al*. [1999] obtained better agreement of the microwave results with the global record than the other techniques. To our knowledge, only one study [*Hill and Shaw*, 2000] has been attempted with volcanic samples in two vertical sections from the 1960 Kilauea flow (Hawaii). The results depicted many similarities with the conventional technique, in particular the existence of two slopes for many plots showing the NRM lost against microwave TRM gained. It is therefore not surprising that the paleointensity determinations were rather scattered be-

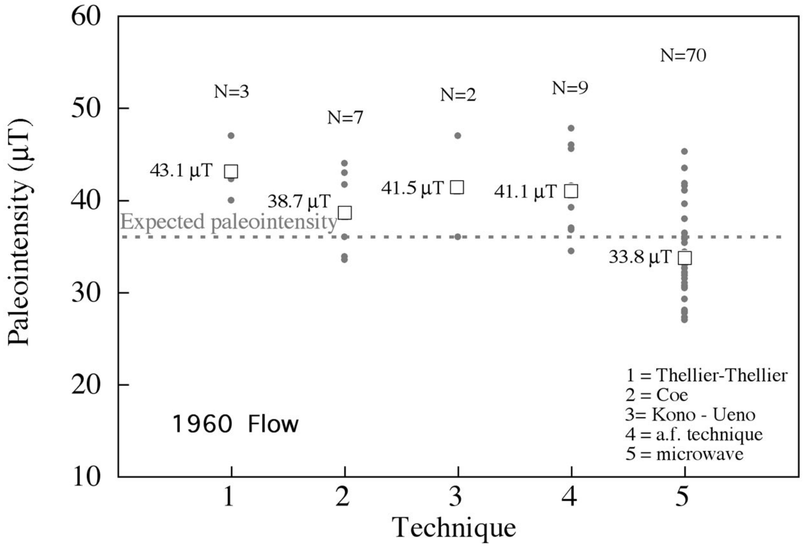

Figure 3. Results of paleointensity determinations from samples of the 1960 lava flow in Hawaii using different techniques. The dotted line gives the expected field value at the site. Data points correspond to a single sample, except for the microwave technique where results have been averaged for several specimens within the same sample. In this last case the results from two sections (see text) have been incorporated. In each case the average paleointensity value is shown by an open square.

tween 25 and 50 μT. In fact, the two sections were characterized by different behavior, and there were changes in the hysteresis parameters at high powers. The samples with two slopes became entirely linear after repeating the microwave experiment. This is indeed similar to the effects of high-temperature oxidation during conventional heating, which is accompanied by reduction of the effective size of multidomain grains to single domains. It is puzzling that the mean intensity of 37.1 +/− 6.4 μT derived from Arai plots with two slopes has been found closer to the expected intensity (36 μT) than the mean value of 31.6 +/− 3.6 μT obtained from plots with an ideal linear behavior. We shall see in section 3.1.8 (and Figure 3) that there is no drastic improvement in the accuracy of the field determination as well as the dispersion around the mean value with respect to standard Thellier techniques. The main advantage of this new approach is rapidity, which, taken alone, is enough to justify the enormous potential interest in the technique. Additional studies on a large collection of recent or contemporaneous flows are needed to definitively assess the promises of microwave experiments.

3.1.6. Improvements Using Alternative Materials

[26] *Pick and Tauxe* [1993a] first proposed that submarine basaltic glasses would be an ideal material for studying absolute paleointensity. The magnetic population of submarine volcanic glasses is consistent with very fine Ti-magnetite grains in the superparamagnetic or single-domain state. There was some concern about the primary origin of the magnetization because of the very fine grain sizes, but consistent determinations were obtained from nicely linear Arai plots by *Pick and Tauxe* [1993a, 1993b], *Meija et al*. [1996], and *Carlut and Kent* [2000], which rules out this possibility. However, there is still a relatively persistent absence of data from contemporaneous samples. These results must be considered along with recent studies by *Zhou et al*. [1999, 2000] using electron microscope analysis, which indicate that submarine glasses do not seem to be sensitive to alteration on a timescale of a few million years. However, there is still a relatively persistent absence of data from contemporaneous samples. *Gee et al*. [2000] obtained excellent agreement between the absolute paleointensities derived from seafloor glass samples collected across the spreading central anomaly of the East Pacific Rise and the well-documented dipole intensity pattern for the past 50 kyr. However, the youngest part of their record is restricted to the 1- to 3-kyr interval. There is no doubt that absolute paleointensities of the youngest axial lava flows will be compared soon with historical measurements of the geomagnetic field and will shed new light on this question.

[27] Unfortunately, various aspects presently prevent using submarine basaltic glasses from several paleomagnetic applications. Indeed, there is a technical problem linked to the difficulty of collecting submarine samples [*Cogné et al*., 1995] oriented with respect to the geographic coordinates. In most cases, samples are not oriented, and the direction of the magnetic field cannot be recovered. It is thus impossible to deal with the full vector (which somehow is a paradox after having measured directions without intensity for a very long period) and also to check for the consistency of the magnetic polarity with the underlying crustal magnetic anomaly. This may change in the future with the development of

new collecting tools. Another limitation is the weak remanent magnetization of the small glassy fragments containing very few crystals. The total moment being of the order of some 10^{-9} A m^{-2} or less, the measurements are frequently noisy, sometimes even impossible with standard techniques. The irregular size of small chips of glass requires special mounting to allow accurate orientation of the samples for the measurements. *Carlut and Kent* [2000] proposed a new mounting technique (based on potassium silicate instead of salt encapsulation) to measure magnetic moments as low as 5×10^{-10} A m^{-2}. Below this value, measurements are seldom possible, and thus the material cannot be used for paleointensity. A last problem is that lavas enriched in magnetic oxides along ridge segments can induce high magnetic anomalies and thus add a significant contribution to the main magnetic field. Glass samples can thus also be delicate for paleointensity. In the absence of detailed chronology and future technical developments to collect oriented samples their use is restricted, but certainly promising, to studying long-term changes in the average field intensity.

[28] *Cottrell and Tarduno* [1999] recently suggested that the use of plagioclase crystals may provide a viable source of paleointensity data. The major argument is that these crystals are less affected by alteration during heating experiments. The plagioclase grains are picked after crushing the samples, but the direction can be obtained from other specimens taken from the same core. The paleointensity results for 10 plagioclase crystals taken from a sample drilled in the 1955 Kilauea lava flow (Hawaii) indicate a mean paleofield value of 33.8 μT, 5% lower than the expected field of ~36 μT measured at the Honolulu magnetic observatory. Since no field measurement has been performed directly above the flow, this could be explained by local anomalies [*Coe and Grommé*, 1973; *Baag et al.*, 1995; *Valet and Soler*, 1999]. It is important to report that the field determinations have been obtained from low-temperature intervals. At higher temperatures, remanence decay was rapid, and measurement errors became significant. Such difficulties for measuring extremely low intensities (sometimes $<5 \times 10^{-12}$ A m^{-2}) may be a limiting factor on the wider use of this technique. In a subsequent study, *Cottrell and Tarduno* [2000] studied 113-Myr-old basalts from the Rajmahal traps using whole rocks and single plagioclase crystals. The magnetic mineralogy was similar, but the paleointensity determinations derived from the whole rock measurements were of lower quality and gave lower values than the plagioclase crystals. According to Cottrell and Tarduno this was caused by a fine-grained magnetic phase formed during the Thellier-Thellier heating experiments. However, to reduce the influence of noise measurement (probably also linked to magnetization of the salt pellets containing the samples, which can reach 1.5×10^{-11} A m^{-2}), the final values were obtained after using a three-point sliding average of NRM and TRM intensities.

3.1.7. Absolute Paleointensity Using Alternating Field Demagnetization

[29] Because paleointensity techniques based on multiple heatings are very time-consuming, it has always been a challenge to improve the speed of the experiments. Techniques involving af demagnetization were the most appealing. Initially proposed by *Van Zijl et al.* [1962], the approach was subsequently revised and referred to as the Shaw method [*Shaw*, 1974]. The direct equivalent of the Thellier approach is to compare the af demagnetization curve of the NRM with the one of a total TRM produced after heating in a laboratory-known field above the Curie temperature [*Van Zijl et al.*, 1962; *Smith*, 1967; *Shaw*, 1974]. If the shapes of the curves are identical, it is assumed that the coercivity spectrum of the NRM has remained unaltered during heating. Indeed, it seems difficult to envisage that magnetization components with identical coercivity spectra would be associated with different mineralogies and grain sizes [*Levi and Banerjee*, 1976]. In this case the value of paleointensity is derived from the NRM/full TRM ratio. The major drawback is that there is a high probability that the total TRM has been affected by thermochemical remagnetization and/or changes in the domain structure. *Doell and Smith* [1969] and *Coe and Grommé* [1973] noted that most samples studied using *Van Zijl et al.*'s [1962] method were characterized by two slopes (low and high coercivity) and that neither provided the right field value. It is indeed frequent that low-temperature segments are characterized by steep slopes because of rapid demagnetization of multidomain grains that are not necessarily involved in the NRM, and thus they provide high field values. In contrast, low-temperature segments can be associated with production of new magnetic phases that increase the TRM values and thus lower the slope.

[30] Further developments and complications happened following attempts by *Shaw* [1974], *Kono* [1978], and *Rolph and Shaw* [1985] to monitor laboratory alteration during TRM acquisition by comparing the ratio of anhysteretic remanent magnetization given before the laboratory heating (ARM_1) and after (ARM_2). *Tsunakawa and Shaw* [1994] noticed that both the uncorrected and corrected versions yielded unreasonable paleointensities. In other words, despite linear behavior over a certain range of coercivities it is not possible to ascertain whether the field determination is correct. For this reason, Tsunakawa and Shaw proposed to detect alteration by heating the samples twice above the Curie temperature in the laboratory. If the difference between the first and the second TRM was larger than the experimental error, the sample was rejected. This selection procedure was applied to 22 specimens and yielded paleointensities within a 10% difference from the expected values. The most significant (and unnoticed) observation was that the selected samples were characterized by linear NRM-TRM diagrams over the entire range of coercivities and not just a limited range of

coercivities. The same conclusions have been reached recently with a larger data set from seven contemporary Hawaiian lava flows [*Valet and Herrero-Bervera*, 2000]. Thirty percent of the NRM-TRM curves obtained after af demagnetization were linear over the entire spectrum of coercivities. The mean field value was <10% higher than the present field, and homogeneous results were obtained within each lava flow. Thus af techniques can certainly be used, provided that the same slope is present over the whole coercivity spectrum without correction.

[31] In fact, the af technique may also be useful to preselect appropriate samples for standard heating techniques by checking that the NRM and the full TRM have identical af demagnetization curves. This procedure is fast and should be quite successful since it relies on the two parameters involved in the thermal experiments.

3.1.8. The 1960 Lava Flow From Hawaii: A Test Laboratory

[32] After considering all proposed improvements, one may wonder to what extent they have been tested and compared with each other. Testing has been conducted in many cases, but comparison is limited to a few samples. Ideally, any technique using paleointensity should be able to provide a large number of field determinations with relatively little intraflow dispersion. Above all, it is necessary to evaluate the accuracy of the field determination. This can evidently be achieved by studying contemporaneous flows, since we know the field that should have been recorded. To our knowledge, the most widely studied flow is the 1960 lava flow from Hawaii. Six different teams [*Abokadair*, 1977; *Tanaka and Kono*, 1991; *Tsunakawa and Shaw*, 1994; *Tanaka et al*., 1995a; *Valet and Herrero-Bervera*, 2000; *Hill and Shaw*, 2000] attempted various techniques, making this flow a real test laboratory. The results obtained for each study have been summarized in Figure 3, which shows the individual sample determinations and the average values. The first observation is that an insufficient number of field determinations were published for most techniques, with the exception of the recent microwave experiments by *Hill and Shaw* [2000]. The second observation is that no technique duplicated the expected intensity of 36 μT but deviated by between 10 and 20%. All values remain consistent within 1σ (the mean value for all sites is 39.6 +/ - 3.6 μT, σ = 3.2). These results suggest that no technique was totally appropriate or, alternatively, that the flow itself may not be very appropriate for paleointensity (a conjugation of the two factors being quite likely). Dispersion could be caused by large inhomogeneities in mineralogy and/or grain sizes or by the presence of complex (re)magnetization processes. Partial remagnetization linked, for example, to production of chemical remanent magnetization (CRM) (components induced by growing of new magnetic grains following chemical transformations) can certainly occur over short timescales after cooling, as evidenced for several volcanic records surrounding reversals [*Hoffman et al*., 1989; *Valet et al*., 1998]. Finally, the existence of field anomalies due to the volcanic underlying terrain [*Baag et al*., 1995; *Valet and Soler*, 1999; *Camps et al*., 1999] can add an external field component, which, in principle, should be averaged out by taking a large number of dispersed samples. In fact, the dispersion of paleointensity values obtained from chips of volcanic glasses from the Juan de Fuca ridge has been attributed to large local crustal magnetic anomalies [*Carlut and Kent*, 2000]. Additional studies are certainly needed from other flows to discriminate complications caused by partial remagnetization or field inhomogeneities from problems inherent to methodology. These studies are, nevertheless, very instructive, as they indicate the precision and thus the degree of confidence that can be expected from determinations of absolute paleointensity. It is certainly reasonable to consider that the present techniques are not able to indicate the field value with a precision better than 20%.

[33] Note, however, that this is not the case for archeomagnetic studies. Indeed, most archeological materials such as bricks or pottery underwent most of the magnetochemical transformations during their original heating and are thus frequently exempt from problems during the laboratory experiments. This explains why archeomagnetic determinations are more accurate and less scattered.

3.1.9. Summary and Suggestions

[34] Because mechanisms yielding mineralogical transformations and/or changes in domain structure remain poorly constrained [*Kosterov and Prévot*, 1998], there can be some subjectivity in the choice of the appropriate segment of the Arai diagram that is supposed to indicate the paleofield intensity. A typical example is the ongoing controversy regarding the existence of two slopes in the Arai diagram. Some authors claim that the low-temperature segment, which is by far the most commonly used, is exempt from any change and can thus be used with confidence, while others propose that this segment incorporates multidomain components that steepen the slope and give values that are too high. Alternatively, the high-temperature segment is frequently affected by irreversible magnetic and mineralogical transformations. In fact, correct and incorrect determinations have been obtained in both cases. It is now clear that no field determination should be done using such diagrams, which are caused by alteration of minerals during heating and/or from the presence of multidomain grains. This prompts one to consider that records of absolute paleointensity should satisfy the following conditions:

1. It is arbitrary to rely on a specific segment of temperature. We can only be confident if a paleointensity estimate has been derived from a single straight line involving low (say, <350°C) and high temperatures (beyond the domain of viscous components). There is grow-

ing evidence that the low-temperature segment (used in most studies) is frequently affected by multidomain grains and is thus inappropriate (or at best delicate) for paleointensity analyses.

2. Measurements of three samples per lava flow are not enough to average out secondary effects. So far, this has been the rule in most studies, but this situation is likely responsible for a large part of the scatter inherent in many studies. At least six (but eight is even better, as is traditionally the case for field sampling procedures) independent determinations per lava flow are necessary to obtain statistically meaningful field value and to average out heterogeneities inherent in the magnetization or in the variability of the local magnetic field. This requires exhaustive sampling of each volcanic unit, given frequently low (around 20%) success rates.

3. Preselection of samples based on their behavior during thermal demagnetization is a reasonable approach. Changes in hysteresis parameters and irreversible thermomagnetic curves provide in most cases very useful indications about production or destruction of mineral phases carrying the initial NRM. Powerful techniques rely on comparisons of partial TRMs acquired at specific temperature intervals after heating specimens to the Curie temperature [*Shcherbakova and Shcherbakov*, 2000] or on comparison of af demagnetization curves of full TRMs. Because such experiments are not time-consuming, they offer the possibility of increasing the success rate and hence reaching a satisfying number of determinations per lava flow.

4. Techniques with af demagnetization may be used only if there is constant proportionality between NRM and TRM over the entire range of coercivities. Parallel experiments using heating methods are needed to confirm the suitability of field determinations by a nonstandard technique. This is valid also (at least for the moment) for the microwave technique.

3.2. Relative Paleointensity

3.2.1. Appropriate Parameter for Normalization

[35] The intensity of natural remanent magnetization (NRM acquired by statistical orientation along the field lines of magnetic moments contained within detrital particles, which represents, in fact, a partial realignment of original NRM vectors) in sediments is primarily a function of both alignment of magnetic grains by the geomagnetic field and concentration of magnetic particles. It is possible to extract relative variations but impossible to extract absolute field intensity. There have been many discussions regarding the choice of the most appropriate magnetic parameter, (1) anhysteretic remanent magnetization (ARM) produced by superimposing a steady or directed field H_0 of small magnitude onto a decreasing alternative field, (2) low field susceptibility (K), (3) saturation isothermal remanent magnetization (SIRM) obtained after application and removal of a steady field that is large enough to align most if not all magnetic moments, to document variations in magnetic concentration and to use as a normalizer of the NRM in order to extract the signal linked to the geomagnetic field. *Tauxe* [1993] provided a comprehensive review on this matter. In their pioneer work, *Johnson et al*. [1975] and *Levi and Banerjee* [1976] proposed to use the remanence whose "demagnetization curve most closely resembles that of the NRM." The anhysteretic remanent magnetization (ARM) was preferred, because this parameter offers the advantage of dealing with single- or pseudo-single-domain grains of magnetite. *Levi and Banerjee* [1976] stated that the NRM and ARM demagnetization curves should be similar to determine the common range of coercivities (and/or blocking temperatures) involved in both magnetizations. Alternatively, *Sugiura* [1979], *King et al*. [1983], and *Tauxe* [1993] considered that IRM and K should preferably be used because acquisition of ARM is not linear as a function of the biasing magnetic field. This is not necessarily true for the range of low steady field values (between 0.05 and 0.1 mT) that are commonly used for paleointensity studies and should not affect the normalization since the biasing field is kept constant over any entire stratigraphic section.

[36] Following this series of innovative papers, no clear consensus emerged in favor of a preferred normalizer. Consequently, authors frequently select one specific parameter (ARM, K, or SIRM) and defend their choice arguing that there is low down-core variability or no correlation with the normalized NRM. The concept of "magnetic uniformity," which was put forward by *King et al*. [1983] as a requirement to obtain suitable paleointensity records, implies the absence of changes in relative grain sizes and concentration. This should be reflected by identical variations displayed by all rock magnetic parameters that can be potentially used as normalizers. Thus, in the absence of any theoretical reason, except for ARM (which involves the same magnetic grains as those carrying the NRM), it would be appropriate to test the coherency of the records obtained with all three normalizers [*Tric et al*., 1992; *Meynadier et al*., 1992]. Climatic influence can control the evolution of magnetic grain sizes and/or magnetic mineralogy and as a consequence the magnetic parameters. It is frequently reported that one given magnetic parameter was more contaminated than others. In this case most authors prefer to use another normalizer. This goes against the rule of magnetic uniformity, at least in a rigorous sense. When K or IRM are preferred, one could even be suspicious, since they also activate a large fraction of magnetic grains that do not carry the NRM. In fact, there is no justification that this alternative parameter will ultimately provide a better normalization. Alternatively, normalization by ARM can be valid but not with other parameters. A typical example is the case of recent sediments from Lake Pepin, Minnesota. *Brachfield and Banerjee* [2000] reported that the NRM/ARM normalized variations match the compilation of archeo-

magnetic measurements in this area [*Lund*, 1996] as well as the record of relative paleointensity from the close-by Lake St. Croix. However, they note that the NRM/SIRM and NRM/K are not coherent with their normalizers but are both strongly coherent with grain-size proxies such as the median destructive field of the NRM. *Brachfield and Banerjee* [2000] conclude that ARM is an appropriate normalizer for sediments dominated by small PSD or single-domain SD grains. As a consequence they notice that we cannot rely on similar profiles of relative paleointensity obtained by the three normalization parameters to consider that they are all adequate normalizers. We suspect that the actual origin of the discrepancies is that grain-size changes induce significant changes in magnetic viscosity, which are extremely delicate to detect in such young sediments. The total NRM vector is the sum of viscous remanent magnetization (VRM) and initial magnetization that reflects the statistical alignment of the grains (viscosity is a gradual change of magnetization with time at constant temperature; as time goes by, there is, indeed, always a fraction of the grains that reach the limit of instability, so that their initial NRM is gradually replaced by a VRM). During periods of high field intensity the NRM increases and the relative part of VRM decreases, which by definition yields a higher median destructive field. In this case the reason why the normalization by ARM was better is that the appropriate range of demagnetization was selected by considering the demagnetizations of the NRM and ARM as well. Keeping a single value (SIRM or K) and normalizing the NRM without considering the amount of grains that are carrying viscous magnetization effectively generates spurious effects. *Brachfield and Banerjee* [2000] proposed a grain-size correction for sediments dominated by larger PSD grains. We would rather not recommend using corrections, unless there is an effective way of testing the final results with other independent data sets. Indeed, the presence of small grains should be seen as a prerequisite for stability of the signal and accurate statistical alignment of the magnetic grains by the field within the sedimentary matrix. It can also be delicate to rely on corrections based on magnetic parameters that do not directly involve the NRM. Note that we are back to the first basic condition, which requires total removal of the viscous components for acquisition of reliable records of relative paleointensity.

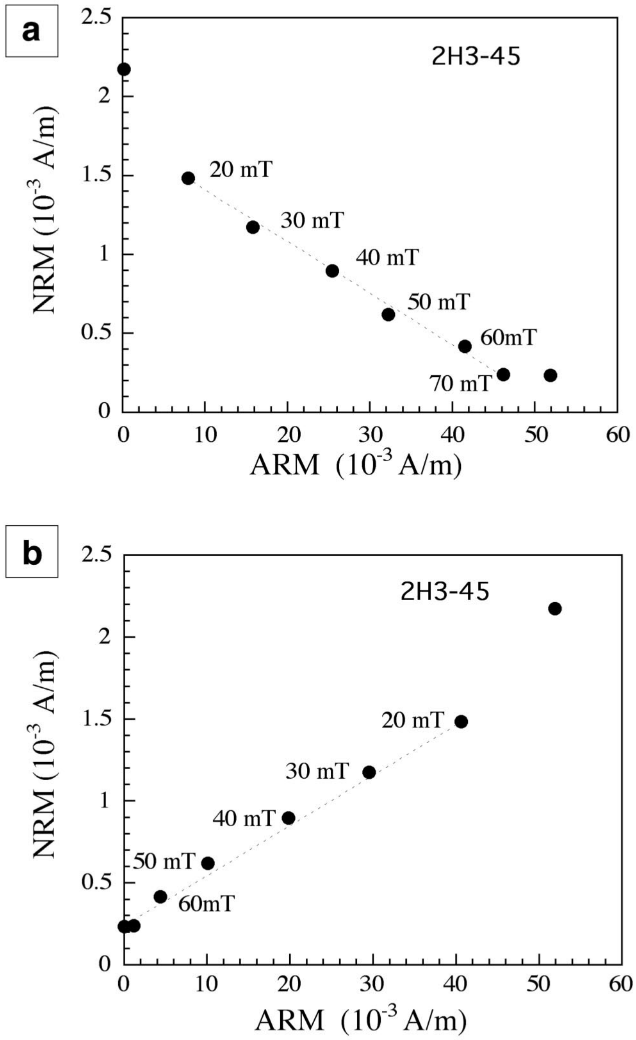

Figure 4. (a) Example of alternating field demagnetization of NRM as a function of anhysteretic remanent magnetization (ARM) acquisition. (b) Same as in Figure 4a but showing NRM values as a function of ARM demagnetization (a full ARM being given after complete demagnetization of the NRM). Both representations show a linear relationship between the two parameters. Viscous components have been completely removed at 20 mT and are only present in the NRM. Either kind of diagram yields identical results [*Valet and Meynadier*, 1998].

[37] *Tauxe et al.* [1995] have suggested that many records could be affected by unremoved viscosity and proposed a "pseudo-Thellier" method for normalization that relies on a comparison of the NRM lost during demagnetization and ARM (or isothermal remanent magnetization (IRM)) acquired within the same range of coercivities. The partial remanence gained in a particular field was plotted versus the remanence left after demagnetizing the total magnetization in the same field, similar to the Arai plot of the Thellier method (Figure 4a). Alternatively, the approach involving demagnetization of the ARM (Figure 4b) is easier to perform and yields similar results [*Valet and Meynadier*, 1998]. *Tauxe et al.* [1995] considered that the difference between the conventional technique involving the choice of a single demagnetization step and the results that they obtained with the pseudo-Thellier approach was caused by a pervasive VRM which biases the traditional technique. Theoretically, the ARM should not be affected by VRM because of the short time span between its production and measurement in the laboratory. This can be deduced also from measurements of recent sediments. *Tauxe et al.* [1995] reported that the NRM versus ARM data measured from freshly stirred remanences showed linear behavior in the NRM-ARM plots, which rules out the

hypothesis of viscous ARM components. Thus appropriate demagnetization of the NRM is the unique important factor. It is essential to perform stepwise demagnetization of the NRM and to construct the signal of paleointensity from the remanent intensity after total removal of the VRM. This is, in essence, the initial requirement proposed by *Levi and Banerjee* [1976]. The same conclusions have also been reached by comparing different techniques of normalization [*Valet and Meynadier*, 1998] with different parameters using alternating fields as well as thermal demagnetization on samples from different locations (Indian and Pacific Oceans). The results were basically unchanged with any technique, provided that the characteristic component of magnetization had been properly isolated and the magnetic grain size had been in the PSD or SD range.

3.2.2. Influence of Climate

[38] The magnetic properties of a sediment, including mineralogy, grain size, and concentration of magnetic minerals can be strongly related to climatic forcing. Such relationships can, in turn, exert an influence on magnetic field estimates [*Lund and Schwartz*, 1999]. *Tauxe and Wu* [1990] were the first to propose a quantitative estimate of the degree to which the normalized intensity records correlate with climatic parameters. They used the coherence function, which can be seen as a generalization of the correlation coefficient in the frequency domain. The statistical significance of certain features in the power spectrum is evaluated by comparing the pattern of the Fourier transforms for different windows. Tauxe and Wu calculated the coherence function between the normalized remanence and the susceptibility of three cores from the western equatorial Pacific. This test between the normalizer and the intensity record has been advocated as a means of assessing relative paleointensity data quality [*Tauxe*, 1993], and it has been used in many subsequent studies. In some cases the relation with other climatic components such as density, carbonate, or $\delta^{18}O$ records, when available, was also investigated.

[39] Most spectral analyses were aimed at detecting coherence between the normalized paleointensity signal and the normalizer, but the climatic content of the initial NRM was barely investigated. It is reasonable to suspect that the remaining climatic content of normalized intensity depends on the spectrum of the NRM (before normalization but after appropriate demagnetization). Indeed, if the NRM shows significant peaks corresponding to periods expected for climatic variations, it is expected that similar peaks would be observed for the normalizer spectrum. In this case it seems reasonable that the normalization will remove most of the climatic peaks and reveal the geomagnetic variations recorded by the NRM. Alternatively, if the NRM does not show climatically related variations while the normalizer does, this could reflect that the magnetization changes caused by field intensity were largely dominant over changes induced by lithological factors. However, a similar situation could be caused by magnetically inhomogeneous sediment or could be generated by remagnetization components. In such cases, there is little chance of any reliable paleointensity estimate. In fact, when the NRM shows no climatic influence, the normalizer should not show climatically related variations. In summary, *Constable et al.* [1998] notice that a good normalizer will be correlated with the initial remanence and should thus be selected accordingly. The extreme situation of a 100% correlation is evidently not expected, as it would indicate that the initial remanence does not contain any record of field intensity.

[40] *Constable et al.* [1998] have noted that the presence of little coherence between the normalizer and the normalized intensity record does not necessarily prove its lack of suitability and that the phase spectrum and the relative power level must also be involved in the diagnostic. Another aspect to consider is that coherence analyses deal with the entire record and do not discard the possibility that some specific parts of the record could be modulated by orbital frequencies. This result somehow questions the use of the coherence functions, which treat the entire record as a whole but do not provide information on the time evolution of the spectral content. In order to achieve this goal we can turn to wavelet analysis. In contrast to the classical Fourier transform, which is computed using a single window of constant width, this multiscale method can distinguish between continuous low-amplitude and nonstationary high-amplitude signals. Another advantage is the use of cross-wavelet spectra, which can be seen as an equivalent of the coherence function. *Guyodo et al.* [2000] analyzed the record from Ocean Drilling Program (ODP) site 983 in order to separate the time intervals over which orbital frequencies are apparent in the record and detected significant and stable covariance between the paleointensity record and the IRM within specific intervals. This relation did not appear using the classical Fourier spectral method.

[41] This point can be illustrated further by constructing a synthetic ARM signal with a climatic component with periods of 100, 41, and 23 kyr superimposed to a time series characterized by a white noise spectrum (Figure 5a). Synthetic geomagnetic field variations are generated by a random process, and the NRM of the sediment is given the following structure:

$$\mathrm{NRM} = \mathrm{ARM}\,(1 + \varepsilon(t)) \times H,$$

where H is field intensity.

[42] The function $\varepsilon\,(t)$ is in the bottom plot of Figure 5a and represents a few percent changes in the response function of the sediment. At any time the NRM is proportional to ARM and to H, but the coefficient of proportionality varies with time depending on ε, which could represent down-core changes in sediment lithology. Significant variations in ε have been imposed between 0 and 350 ka and 700 and 900 ka (Figure 5a). Consequently, there must be some covariance between

the NRM/ARM ratio and the ARM record within these intervals. The power spectra of the ARM and NRM/ARM signals, which were obtained using the Blackman-Tuckey method with a Bartlett window, are plotted in Figure 5b. The three climatic frequencies that were introduced in the ARM signal are effectively present, but they are absent from the NRM/ARM signal. Note also that the coherence values obtained between the ARM and the paleointensity record are below the 95% confidence level (Figure 5b). Therefore these results would be considered as acceptable, and the NRM/ARM variations are a reliable estimate of the changes in field intensity. In contrast, the results in Figure 5c obtained using the wavelet technique show that covariance is indeed present within the expected intervals. We conclude that wavelet analysis was more efficient to determine climatically contaminated intervals.

[43] Once contaminated intervals have been removed, we can wonder if the rest of the paleointensity record can be considered as trustworthy. This is certainly correct for the phase, but we cannot rule out the possibility that the amplitudes could be affected by other persistent climatic influences. Climatically induced lithological changes affect the physical properties of the sediment, which, in turn, generate changes in the response curve of the magnetization. The phase and the amplitude of the signal are not independent; the latter is indirectly affected by climatic components. In other words, we must be cautious when interpreting such records. Comparisons between individual records and stacked curves might help to better constrain and analyze the amplitudes of the signals.

3.2.3. Summary and Suggestions

[44] We are still far away from a detailed understanding of the complexity inherent in the magnetization processes in sediments. However, we shall see in section 3.3 that the basic techniques described above have been sufficient to extract coherent signals of relative paleointensity. In the present state the following guidelines can be proposed to obtain reliable records:

1. Complete removal of all secondary components is a primary condition. Standard af demagnetization has been shown to be effective and, actually, the most reliable technique for demagnetization of sediments (except for land-exposed sections, which are mostly affected by alteration with high-coercivity minerals). The presence of a well-defined characteristic component, which linearly decreases toward the origin, is a basic requirement. There is no need for additional treatments, which can generate additional complexity.

2. Choice of the appropriate parameter for normalization can rely on its coherence with the NRM. Multiple normalizers with identical variations help to strengthen the case of reliability. However, ARM-normalized records should be preferred as they involve PSD or SD magnetic grains.

3. Wavelet analysis is a powerful approach to detect "climatically contaminated" intervals. Coherence analysis is evidently also appropriate and more efficient if it is carried out over successive intervals. Such analyses should include testing of the final record with magnetic concentration and magnetic grain-size proxies.

4. It is important to deal with series that have as accurate time-depth control as possible. Sediments with at least 10% carbonate composition appear most appropriate, since they can be potentially dated by isotope stratigraphy or alternative techniques related to dilution of carbonates. In contrast, magnetization of clay-rich sediments can be affected by misalignment of large aggregates of particles containing small magnetic grains, a concept that has been reinforced by a recent study [*Katari and Bloxham*, 2001]. We see in section 3.3 that this process may not be compatible with current normalization used for paleointensity determinations.

3.3. Paleointensity From Production of Cosmonuclides

[45] Cosmogenic radioisotopes such as ^{14}C ($T_{1/2}$ = 5.73 kyr), ^{36}Cl ($T_{1/2}$ = 300 kyr), and ^{10}Be ($T_{1/2}$ = 1500 kyr) are mostly produced in the stratosphere, by interaction of cosmic ray particles with the Earth's atmosphere, before being incorporated within various geochemical reservoirs. The production rate of these radionuclides is therefore a function of changes in primary cosmic ray flux, solar activity, and shielding by the geomagnetic field. Little is known about variations of the cosmic rays (e.g., the catastrophic explosion of a supernova proposed by *Sonnett et al.* [1987]. In fact, the overall flux appears to have been constant within ~20% on timescales of millions of years [*Vogt et al.*, 1990]. Changes in solar wind modulation are poorly known, although *Raisbeck et al.* [1987] describe brief periods of enhanced cosmogenic production during the past 80 kyr, which they attributed to solar effects. Variations on timescales of <500 years are therefore generally related to solar activity. In most sedimentary records they are considered as negligible since they are not resolvable. Therefore variations in the global dipole moment appear to be the most important factor controlling cosmogenic radionuclide production. In principle, the three isotopes with very different half-lives can provide us with complementary and independent sources of information about changes in geomagnetic activity, according to the first-order relationship Q_m/Q_{m0} = const. $(M/M_0)^{-0.5}$ (where Q stands for the global nuclear disintegration rate for a dipole moment M and M_0 and Q_{m0} represent the current values), established by *Elsasser et al.* [1956] and revisited by *Lal* [1988]. The shielding of the Earth by its magnetic field varies with latitude. Higher production of cosmogenic radionuclides is caused by enhanced penetration of cosmic particles (due to the orientation of the magnetic field lines) at high latitudes (between 60° and 90°), while the lowest production rates are observed at the equator.

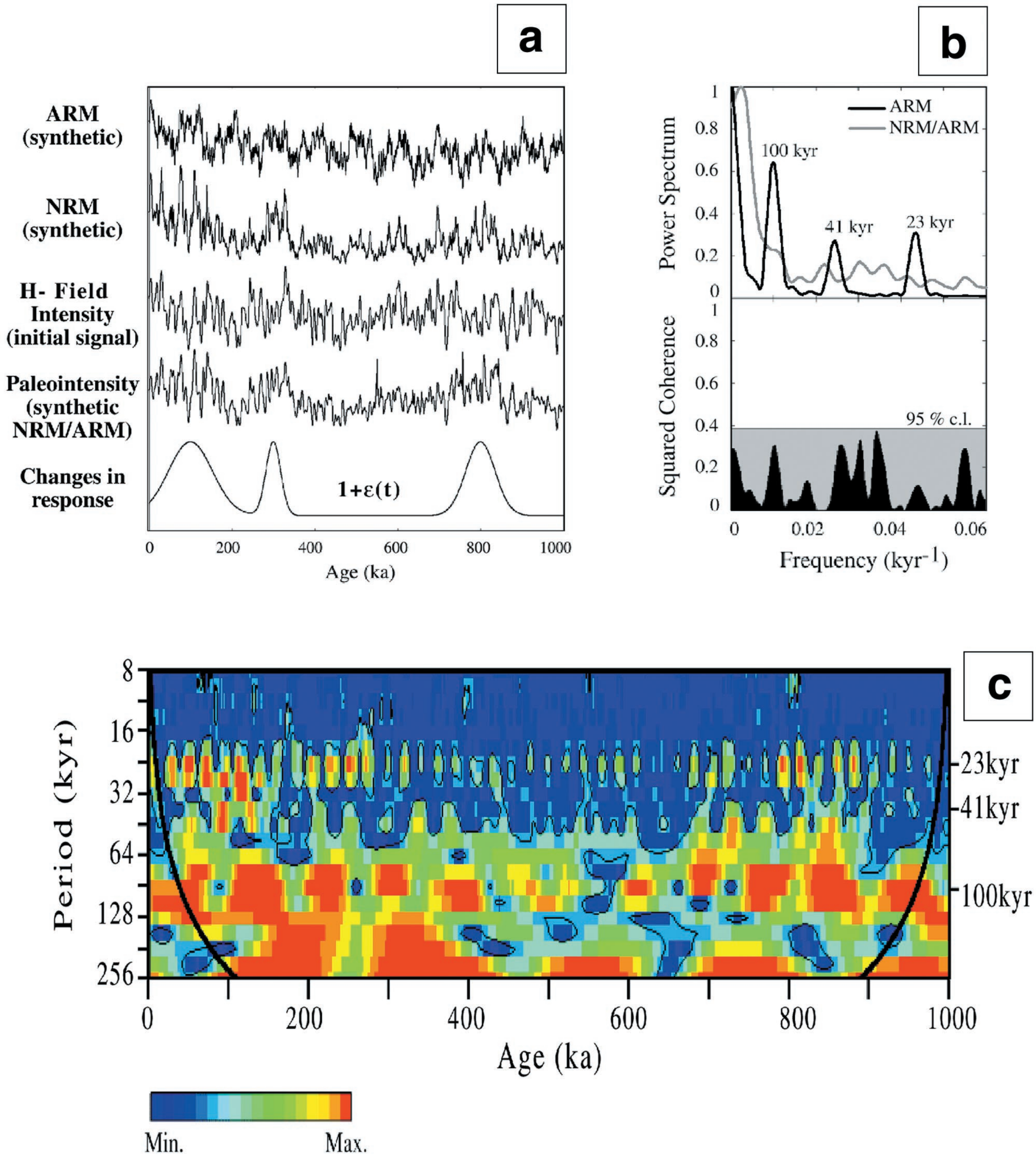

Figure 5. Results of wavelet analysis performed from synthetic signals of relative paleointensity in order to determine possible climatic influence. (a) Anhysteretic remanent magnetization (ARM) variations constructed as a series of 100-, 41-, and 23-kyr periodic functions superimposed to a random signal. A time series representing the field intensity changes was generated by a random process. The NRM depends on concentration (ARM) and intensity changes, but the relationship varies according to a function ε (see text), which could represent down-core changes in lithology. Significant variations in ε have been simulated within the 0- to 350-kyr and 700- to 900-kyr time intervals. Thus covariance between ARM and NRM/ARM is expected within these intervals. (b) Global power spectra of ARM and paleointensity records using a Blackman-Tuckey method with a Bartlett window. The power spectrum of the ARM shows the expected maxima at the periods used in the construction of the signal. However, there is no peak in the paleointensity record despite the existence of a lithologic component. The coherent function between the two signals (ARM and paleointensity) is below the 95% confidence level, whereas the two records are covariant within specific intervals. (c) Cross-wavelet power between the ARM and the paleointensity proxy. Regions of the contour plot in red correspond to local maxima of correlation. Bold black lines on the left and right sides indicate regions where edge effects are significant. As expected, there is covariance between the ARM and the paleointensity record within the time interval 0–350 ka and around 800 ka. These results illustrate the potential of wavelet analysis to detect climatic components.

3.3.1. Beryllium 10 and Chlorine 36

[46] Beryllium 10 has been measured in sediment and also in ice cores. A major advantage of this isotope is that its deposition rate in sediments is not affected by changes in ocean circulation. However, two important factors must be taken into account before ^{10}Be measurements can be interpreted in terms of relative paleointensity. The first factor is a dilution process that can be caused by bottom current or by variations in the biogenic content of the sediments linked to changes in biological activity [*Frank et al.*, 1997]. The second factor is that the 400- to 1000-year-long ^{10}Be residence time in the water column is long enough to be widely influenced by boundary scavenging effects (locations at ocean margins receive a higher ^{10}Be supply than locations in the central ocean gyres). In order to correct for these effects several authors [*Raisbeck et al.*, 1985; *Henken-Mellies et al.*, 1990; *Cini-Castagnoli et al.*, 1995; *Robinson et al.*, 1995] proposed to normalize ^{10}Be by ^{9}Be. In this case the ratio of cosmogenic ^{10}Be to ^{9}Be is a function of the amount of ^{10}Be accumulating as a result of rain and the overall amount of ^{9}Be supplied by terrigenous material entering the oceans and adsorbed on small particles. However, environmental conditions, among which are circulation changes and advection of different water masses with different ^{10}Be/^{9}Be ratios, particularly between glacial and interglacial periods, could influence the signal. However, note that recent results [*Blanckenburg et al.*, 1996] suggest that the ^{10}Be/^{9}Be ratio would not show secular variations due to changes in deep-water flow. In any case, normalization by ^{9}Be requires constant input of detrital material. These problems could explain why several individual records failed to detect a production increase in ^{10}Be records [*Henken-Mellies et al.*, 1990; *Raisbeck et al.*, 1994] at the time of reversals, which are associated with low geomagnetic intensity.

[47] The alternative technique relies on a normalization of the ^{10}Be measurements with respect to initial ^{230}Th introduced in the sedimentary column, which has the advantage of being accumulated at the site, similar to ^{10}Be. The reconstruction of ^{10}Be production rate for the last glacial maximum [*Lao et al.*, 1992] indicated an increase of $\sim$30% in radionuclide production in good agreement with independent estimates derived from U/Th-calibrated ^{14}C ages [*Bard et al.*, 1990]. However, dilution of ^{230}Th in sediment can subsequently be affected by lateral effects. For this reason, *Frank et al.* [1997] considered that averaging the values of several individual remote records is the best approach to normalize the ^{10}Be records.

[48] Production rates of ^{10}Be and ^{36}Cl records have also been derived from ice cores in Antarctica and Greenland. The geographic poles receive a considerable part of their precipitation from lower latitudes, which homogenizes the latitudinally dependent factors so that the flux of cosmogenic isotopes is not dominated by local radioisotope composition. The concentrations must be corrected for variations in accumulation rate and density of ice as constrained by age models. Owing to fast accumulation rates, high-resolution records of geomagnetic changes can potentially be obtained.

[49] In principle, the half-life of ^{10}Be should allow us to explore the past 10 Myr of ^{10}Be production rate and thus to compare two data sets of relative paleointensity for this period. Unfortunately, sedimentary records cannot be reliably corrected [*Frank*, 2000] for sediment redistribution using ^{230}Th beyond 300 ka because of its much shorter lifetime (75 kyr). Because there is a relationship between climatic changes and oceanographic circulation, *Frank* [2000] considers that the alternative approach based on the ^{10}Be/^{9}Be ratio will require similar stacking of individual records from different areas.

3.3.2. Carbon 14

[50] Because of its relatively rapid decay, ^{14}C is applicable to the past 35–40 kyr and thus has less potential geomagnetic interest since this period is relatively well documented by absolute paleointensities from archeological materials and lava flows (see section 4). However, changes in dipole field intensity have been used to constrain the variability imposed on ^{14}C production by the geomagnetic modulation and ultimately to recalibrate the Δ^{14}C timescale. One advantage of ^{14}C is that it is rapidly mixed within the atmospheric CO_2 reservoir and thus loses its latitudinal dependence. However, the variations in ^{14}C measured in archive records are not only influenced by field intensity but also by changes within the global carbon cycle, as CO_2 is exchanged between the atmosphere and the ocean. Processes like changes in thermohaline oceanic circulation, reduction of the biosphere reservoir, or lower atmospheric CO_2 levels during glacials (which may induce an increase of 25–75% in atmospheric Δ^{14}C) must be taken into account before converting the observed Δ^{14}C into field intensity changes. In fact, such effects smooth out the fast changes and introduce a delay and some long-term memory into the system.

4. DATA AND RESULTS

4.1. Archeomagnetic Intensity

[51] The period covering the past few thousand years is critical as it provides a link between the relatively fast changes inherent in the historical field and the long-term variations present in paleomagnetic records. Indeed, this is the appropriate timescale to scrutinize most variations generated by the nondipole field. This requires materials that recorded field changes with a resolution better than a few years. Archeological materials can be found which preserve unambiguous records of field direction and intensity with uncertainties in age, which sometimes do not exceed several tens of years. Relative paleointensity from lacustrine sediments with high deposition rates can also be envisioned. However, the upper stratigraphic levels of sediment cores are often disturbed by drilling,

and their chronology is difficult to establish. It is thus not straightforward to calibrate these records with absolute values given by archeological material or by the historical field. *Brachfield and Banerjee* [2000] attempted this approach by comparing two distinct records from Lake Pepin and Lake St. Croix with the ARCMAG archeomagnetic data set [*Lund*, 1996] that was constructed from absolute paleointensity records on archeological materials and basalts from the western United States. The ARCMAG results were placed in bins that are 50–250 years wide, and they reproduce the same pattern as the sedimentary curves for the past 3000 years. Beyond this limit the sedimentary records are not very coherent, and there is a gap in the archeomagnetic data between 4 and 6 ka. According to the results the field intensity would have been almost twice as high during the time interval 700–1000 kyr B.P., close to the present value in the 1100- to 1300-year time interval, and higher again 2000 years ago.

[52] *Daly and LeGoff* [1996] have compiled an archeomagnetic world data set in intensity and direction. The analysis has been restricted to the past 20 centuries and to sites for which the number of records was large enough to perform a statistical treatment. This yielded results at nine world sites and proper documentation of the secular variation at five sites. For the eastern European sites, new data obtained after 1982 come from Bulgaria-Greece (128 data points) [*Kovacheva and Karnachev*, 1986; *Kovacheva et al*., 1995], Ukraine (259 data points), and the Caucasus (83 data points), many of them already summarized by *Aitken et al*. [1989]. To these has been added the data set from Japan (66 data points). The selected records have errors lower than 10% and uncertainties in age <200 years. In each case the results were reduced to a single regional site (Sofia, Kiev, Gori, and Kyoto, respectively) and were averaged every 25 years within 80-year windows. We must also consider recent determinations obtained from French archeomagnetic sites [*Chauvin et al*., 2000] for time intervals ranging from 65 to 380 A.D. and from 1370 to 1700 A.D. Several studies have also been performed in Finland [*Pesonen et al*., 1995], but the existing data set remains relatively poor compared with other regions.

[53] The VDMs (see definition in section 2) of the past 2000 years for eastern European and Japanese sites have been plotted in Figure 6a in 25-year increments. There are large differences, of up to 20%, between the five records, particularly for the Japanese results. This is not surprising, given the contribution of the nondipole components. In fact, the VADM variations were found to be in slightly closer agreement with each other than the VDMs at the three European sites. This likely reflects the dominance of large variations driven by the axial dipole but more probably reflects that some inclination records would be affected by experimental problems. When compared with eastern Europe, the Japanese data set seems to be similar but more or less phase shifted from the other records. Whether this could reflect the presence of drifting field components requires additional and more detailed results. There is also the possibility of systematic errors in dating.

[54] Shown in Figure 6b is the VDM curve that was obtained after stacking the results of the three European sites, which is compared with a recent and large data set from French sites [*Chauvin et al*., 2000]. Using the same archeomagnetic database as *Daly and Le Goff* [1996], to which they added other sparse records, a few sedimentary records from Argentina and New Zealand and lava data from Hawaii, *Hongre et al*. [1998] computed a time-varying spherical harmonic model of the geomagnetic field between 2000 and 300 B.P. using archeomagnetic and paleomagnetic data from 14 unevenly distributed sites. The curve derived from the model, also shown in Figure 6b, is quite similar to the eastern European stack and confirms that, on average, the dipole field has been decreasing over the past 2 kyr by almost 50%. However, this decrease has not been as regular as previously suggested [*Merrill et al*., 1996] and appears to be punctuated by periods of partial recovery (e.g., between 1600 and 1300 B.P.). According to *Hongre et al*. [1998] secular variation seems to be mostly controlled by the $g_1^{\,1}$ Gauss coefficient. Thus field intensity variations during the past 2 kyr would be dominated by a rather long-term decrease of the axial dipole superimposed on a fluctuating equatorial dipolar component. This pattern is not obvious in the North American data set described above despite the presence of similar variations but not in-phase variations. However, in this case the very limited geographical area covered by these records precludes any direct comparison on a worldwide basis.

[55] It is tempting to compare these eastern European results with the available records of the same period from western Europe. Since data do not incorporate inclination records, the comparison must rely on VADMs. *Chauvin et al*. [2000] attempted this analysis and concluded that many records from western Europe might not be reliable. There are, indeed, differences between the two European data sets shown in Figure 6c, despite some apparent tendency to duplicate the broad trend of the eastern Europe composite curve, but the large gap in western European records between 500 and 1400 A.D. prevents any definitive comparison. Recent results [*Genevey and Gallet*, 2000] obtained from French potteries with dates distributed between the fourth and sixteenth centuries are in satisfactory agreement with those of the French sites studied by *Chauvin et al*. [2000] for the fourteenth and eighteenth centuries. They are also globally similar to the composite record shown for eastern Europe in Figure 6c. The authors observed a shift of 150 years between the variations recorded in France and Ukraine, which they interpret as evidence for eastward drift of geomagnetic sources between western and eastern Europe. Given the temporal resolution of the records and the rate of the variations, it cannot be excluded that the dipole also plays a major role.

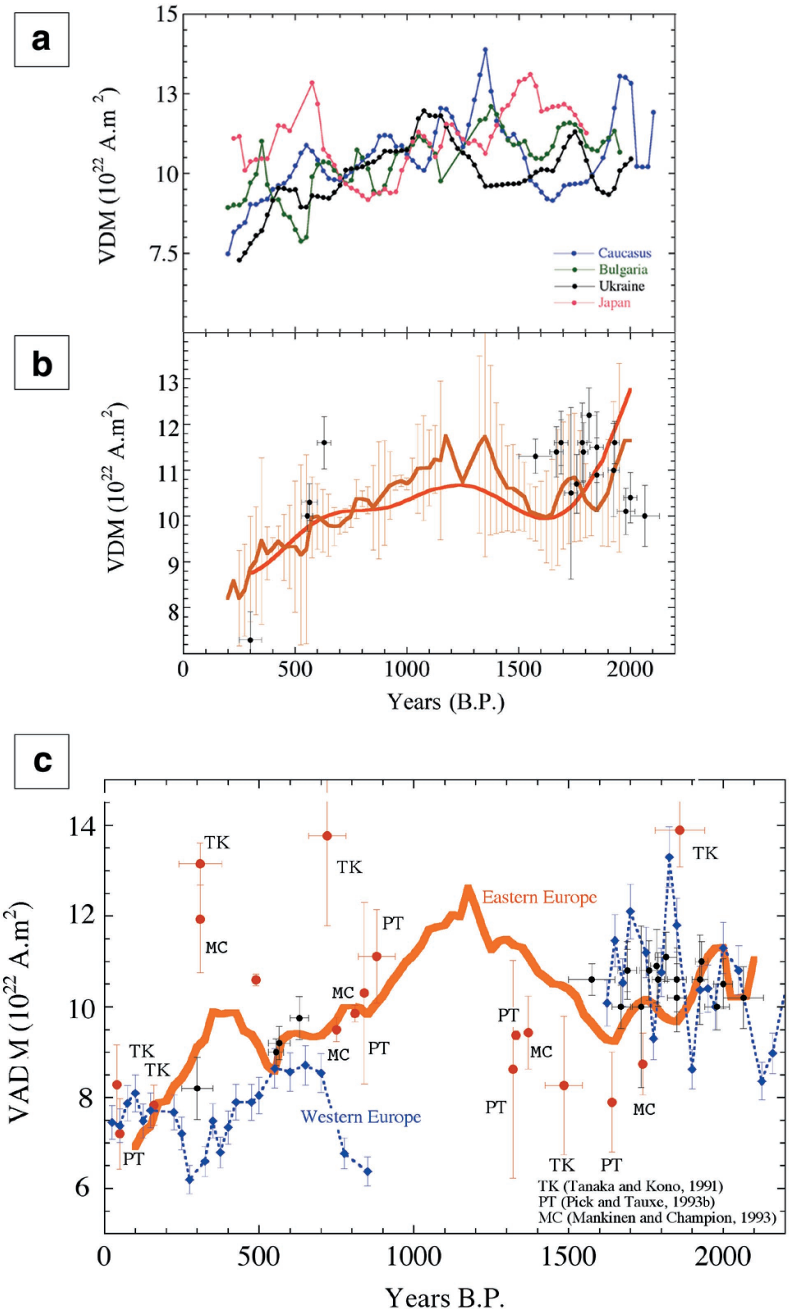

Figure 6. Field intensity changes during the past 2 kyr. (a) VDMs calculated from the *Daly and LeGoff* [1996] database at three European sites and Japan. (b) Composite record of the VDMs for eastern Europe (bold orange line) and predicted VADM variations (red line) derived from the model of *Hongre et al*. [1998] using a wider data set database for the same period. For comparison, recent determinations [*Chauvin et al*., 2000] from French dated sites (dots) are also shown. (c) Comparison between the stacked VADMs of the eastern and western European sites. The western European records are much less detailed and more scattered than the eastern sites, which certainly reflect experimental errors. For comparison, the most recent individual volcanic VADMs (red dots with error bars) obtained for this period [*Pick and Tauxe*, 1993b; *Mankinen and Champion*, 1993; *Tanaka and Kono*, 1995] are also shown.

[56] The next step is to compare the archeomagnetic results with determinations obtained from lava flows. A rapid summary of studies performed during the past decade shows that most volcanic records have been obtained from recent Hawaiian flows [*Tanaka and Kono*, 1991; *Mankinen and Champion*, 1993]. In Figure 6c the corresponding VADMs have been plotted with the archeomagnetic records. VADMs were used in order to

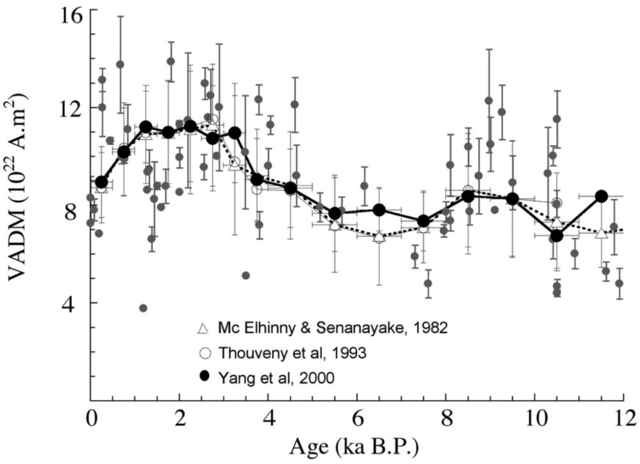

Figure 7. Changes in VADMs during the past 12 kyr B.P. deduced from previous archeomagnetic compilations by *McElhinny and Senanayake* [1982] and *Thouveny et al*. [1993] and the most recent one by *Yang et al*. [2000]. The successive values have been averaged within time intervals of 500 years between 0 and 4 kyr B.P. and 1000 years before. Despite accumulation of results the overall pattern of the curves has not changed significantly in 2 decades. The individual VADMs obtained by Thellier or modified Thellier techniques (grey dots) for the same period have been plotted for comparison.

incorporate data from volcanic glasses [*Pick and Tauxe*, 1993a, 1993b]. Three volcanic values are significantly different, while the others are in relatively good agreement with the archeomagnetic records. Owing to poor temporal resolution and relatively large dispersion it would be very delicate to interpret the field evolution from the volcanic data set. Overall, there seems to be better coherency between the different data sets for the period 1300–1800 years B.P., but this is mostly due to the western European archeomagnetic data set. There is a need for more data covering the past millennium to better document the part of the field spectrum between a few centuries and 1 kyr.

[57] The most recent worldwide compilation of archeointensity for the past 12 kyr has been published by *Yang et al*. [2000]. Despite the presence of almost 3 times more data the results shown in Figure 7 are similar to those depicted by previous compilations [*Mc Elhinny and Senanayake*, 1982; *Thouveny et al*., 1993]. However, two thirds of the data come from European sites, half of them covering the past 2 kyr. The individual values of paleointensity have been averaged within 500- and 1000-year intervals and converted to VADMs. Thus these curves most likely reflect changes in dipole field intensity, despite relatively poor geographical coverage. Between 3 and 1 ka the global field was, on average, about twice as large as between 7 and 5 ka. These results illustrate the potential of archeomagnetic records, which may soon provide us with precise and accurately dated field determinations for the past few thousand years.

4.2. Coherence of Paleointensity Records for the Past 50 kyr

4.2.1. Resolution of the Records

[58] One of the ultimate goals for studies of paleointensity is to produce converging results between cosmonuclides and magnetic measurements in sediments as well as between sedimentary and volcanic records. This is a long and difficult task, because it requires a very large number of data. Before going further into the description of the results, it is thus important to discuss briefly the typical time resolution inherent to sedimentary and volcanic records.

[59] I discuss in section 2 that time constants governing the evolution of the nondipole components (above degree 2) likely do not exceed a few hundred years [*Hulot and Le Mouël*, 1994]. The highest-resolution records from sediments rarely exceed deposition rates of 10–12 cm per thousand years. Deposition above this rate is frequently associated with turbidity currents or a very large amount of clay materials that generate complex magnetization signals and disturbances. Consequently, even with a 2-cm sampling interval, resolution cannot be better than 150 years and may be longer, because many records were obtained from measurements of long sediment cores using pass-through magnetometer systems that generate a convolution of the signal. Smoothing (although most plausibly quite limited) can also be induced by differential alignment of magnetic grains and by bioturbation. In summary, it is certainly not realistic to envisage a time resolution better than 500 years. This implies that the large majority of records should have averaged out most nondipole components and therefore mainly reflect dipole intensity variations. So far, we did not examine the possible role of long-term standing (i.e., with nonzero long-term mean) components. We have seen in section 2 that there is some consensus for the presence of a persisting axial quadrupole term of the order of 5% of the axial dipole, which does not represent more than 3 μT. This value is more likely below the limit of resolution for records of relative paleointensity. Thus, even in the most pessimistic situation of records at the equator with weaker dipole field intensity ($\sim$30 μT) than at higher latitudes, the contribution of a long-term standing component would be very difficult to extract. There is also no reason to consider that large discrepancies between parallel data sets can be caused by such components.

[60] Temporal resolution of the volcanic records is strongly affected by the sporadic nature of volcanic eruptions, where periods of intense activity alternate with long quiescent intervals. The chronology of the flows can be determined by dating techniques, but errors in dating frequently exceed the time span between successive eruptions. In addition, there is no other possibility than linearly interpolating the ages between two dated flows, which (in the presence of irregular activity) can generate serious discrepancies. The simulated paleomagnetic

record resulting from discontinuous and quasi-random reading of field intensity by lava flows has been processed by generating random volcanic sequences with pulses of volcanic activity and quiescent periods [*Guyodo and Valet*, 1999a]. At least four flows per 100 kyr were assigned an age with an error bar. The geomagnetic field variations were simulated by sampling a reference curve that depicts the changes in dipole field intensity with time, to which were superimposed nondipole field changes with the same time constants and variances as the present geomagnetic field. The initial signal could not be reproduced correctly with less than ~10 flows per 100 kyr. A simulation of the global volcanic database for the past 780 kyr was also carried out after generating 20 volcanic sequences of various lengths (between 50 and 500 ka). Each sequence was characterized by four dated flows per 100 kyr after incorporating uncertainties in dating and in the laboratory experiments [see *Guyodo and Valet*, 1999a]. Only paleointensity determinations obtained from the dated flows were incorporated into this simulated database. Many individual paleointensity values derived from this construction in Figure 8a match the dipole field changes, but several show significant discrepancies that exceed the nondipole variability and are caused by experimental uncertainties. One could question whether such data provide significant information about the long-term field changes. A good match with the dipole field variations was obtained after averaging the volcanic data within a 45-kyr-long window (Figure 8b). However, if the distribution of data points is constrained to the resolution of the present global database for paleointensity (Figure 8c), the match remains poor, leaving little hope for extracting significant information in the present state. An interesting illustration of these difficulties is provided by very recent records obtained from drilled cores in the big island of Hawaii. *Teanby et al*. [2002] considered two different age models to correlate the very detailed records of field variations during the past 50 kyrs measured in the drilled cores from scientific observation hole 1 (SOH-1) and from the nearby hole SOH-4 [*Laj et al*., 2002], respectively. The first one is a radiogenic model based on independent Ar/Ar and K/Ar dating, but it provided poor correlation between the two records and also with other Hawaiian data. The second age model relies on direct correlation of common paleomagnetic features. It was fine-tuned by using a stretching algotithm to minimize the misfits. The agreement is much better, but it generated some inconsistencies with the radiometric dates and implied also the existence of extremely high accumulation rates. Despite these discrepancies this model was prefered since it provides a better correlation of the two records. However, the correlations remain inevitably subjective.

4.2.2. Convergence Between Sedimentary and Volcanic Data Sets

[61] A first compilation of volcanic records was attempted by *McElhinny and Senanayake* [1982] for the

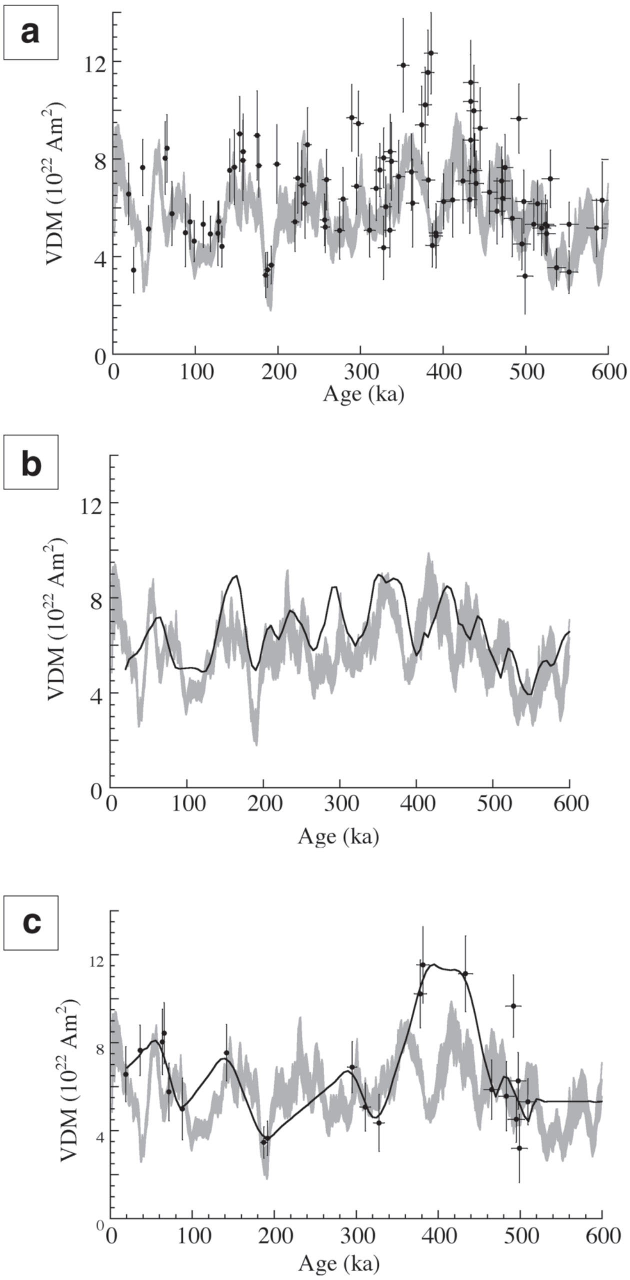

Figure 8. Integration of volcanic and sedimentary data. (a) Simulation of a database for absolute paleointensity that would be constructed from various individual flows with an average of four dated values per 100 kyr [*Guyodo and Valet*, 1999aa. The assumed changes in dipole intensity are shown in grey (and represent what is actually measured in sediments). Nondipole field variations have been generated using the same characteristic time and variance as for the historical field, and they were added to the dipolar field. (b) Smoothing of the simulated database with a moving window of 45 kyr compared with the assumed changes in dipole intensity (grey curve). There is a good match between the two curves, but the correlation is not perfect because of uncertainties in dating. (c) Same smoothing as in Figure 8b with a distribution of data points similar to the one of the present database for paleointensity. The present volcanic database is far from being of sufficient resolution.

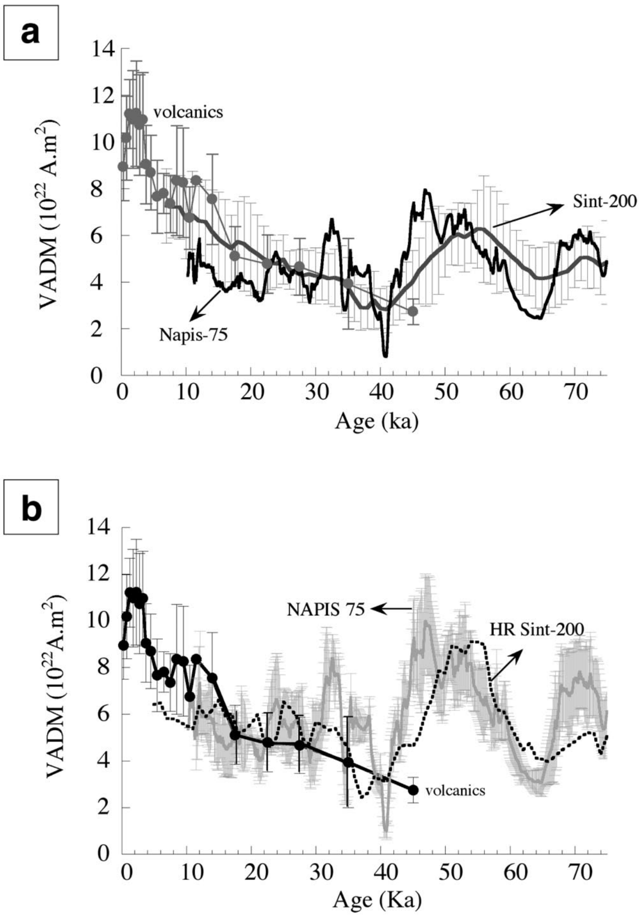

Figure 9. (a) Field variations during the past 75 kyr (North Atlantic paleointensity stack (NAPIS-75)) generated by stacking six independent records from the North Atlantic Ocean [*Laj et al.*, 2000a] plotted with the most recent version of the Sint-200 database (error bars are drawn from the standard deviation around the mean value). These stacks are compared to the composite record of the volcanic database [*Perrin and Shcherbakov*, 1998] obtained after averaging VADMs within 500- and 1000-yr-long time intervals [*Yang et al.*, 2000] for the past 45 kyr. (b) Comparison between NAPIS-75 and a stack of the eight highest-resolution independent independent records that compose the Sint-200 most recent database. Discrepancies between the two records have been considerably reduced; they have mostly been caused by differences in the timescales (see text) used to construct their time-depth correspondence. The past 10–15 kyr remain poorly constrained by the sediment since they correspond to the top of cores that are mostly disturbed by the coring process. Shown again for comparison is the composite volcanic record for the past 45 kyr.

past 50 kyr and then by *Tanaka et al.* [1991], *Merrill et al.* [1996], and *Valet et al.* [1998]. Like for the past 12 kyr, the curve shown in Figure 9a was obtained after averaging the dipole moments within successive time intervals, but their length (between 500 and 5000 years) depended on the resolution of the data set. Like for the past 12 kyr again, despite accumulation of new records, it is striking also that the curve did not change significantly during the last decades. Between 5 and 25 kyr B.P., field variations exhibit some oscillations superimposed on a decreasing trend. Prior to 25 kyr B.P., there are fewer records, but they are sufficient in number to average the virtual moments values within 25-kyr-long time intervals. The situation is more dramatic for the past two intervals, which combine results over 5-kyr-long windows. As a whole, the variations of the VADM depict a rather monotonic decrease during the past 40 kyr, but the length of the intervals prevents observation of shorter-term variations. This decrease cannot be excluded given some dispersion in the results and insufficient records between 25 and 40 ka. For these reasons it is important to compare this data set with the available records from sediments. Note also that the constraints imposed by the overall pattern of the VADMs shown in Figure 9a can be used to calibrate the relative paleointensity of sediments to the absolute values indicated by the volcanics.

[62] Following acquisition of a few sparse records of relative paleointensity, it became crucial to test the coherency of adjacent records from the same geographical area. This was achieved with the emergence of records from the Mediterranean Sea [*Tric et al.*, 1992], the Somali Basin [*Meynadier et al.*, 1992], the Ontong-Java Plateau [*Tauxe and Shackleton*, 1994], the western [*Yamazaki and Ioka*, 1994] and the eastern [*Schneider*, 1993] equatorial Pacific, the Azores area [*Lehman et al.*, 1996], and the Labrador Sea [*Stoner et al.*, 1995, 1998]. Two different and complementary approaches were followed. The first one involves high-resolution records, which can be merged into composite curves but are limited to a specific region. This has been done by *Laj et al.* [2000a] from six independent marine cores [*Channel et al.*, 1997; *Kissel et al.*, 1999] that were incorporated into a North Atlantic paleointensity stack (NAPIS-75) covering the past 75 kyr. The results are shown in Figure 9a. The second approach attempts to provide a global description of the dipole field over a long time period. In this case the records are longer but of lower resolution. The first stack [*Guyodo and Valet*, 1996] has actually been constructed for the past 200 kyr using 18 well-dated marine cores (most of them quoted above) that were sufficiently well distributed geographically to provide a global picture. The resulting curve Sint-200 (signal of intensity for 200 kyr) is also shown in Figure 9a over the past 75 kyr for comparison with NAPIS-75. In both cases the stability of the results has been investigated using bootstrap and jackknife techniques. Overall, the two composite curves display the same global pattern, but, as expected, the NAPIS stack is much more detailed than Sint-200. Note also that there is slightly better agreement between the volcanic composite and the Sint-200 stack, but this must be considered again with the resolution of the two data sets.

[63] Some records from the Sint-200 database reach the resolution of NAPIS, but a few others do not exceed 1cm per thousand years. The stack was thus tuned to the lowest resolution of one data point every 2 kyr. The NAPIS-75 and Sint-200 curves are globally similar, but,

as noticed by *Laj et al*. [2000a], NAPIS-75 does not exhibit the monotonic increase of Sint-200 during the past 40 kyr that is observed in the compilation of the volcanic records (note that the Sint-200 curve reported in Figure 12 of *Laj et al*. [2000a] does not correspond to the actual data set). This problem can be investigated further by selecting the records with the best temporal resolution using an updated database of Sint-200 to construct a high-resolution (HR) Sint profile for the past 75 kyr. The results shown in Figure 9b, which have been obtained by stacking eight independent records, are closer to the NAPIS record. The agreement could be improved further by correcting the two curves for the differences between marine and ice chronologies. Thus, on one side the low-resolution Sint-200 stack is in excellent agreement with the averaged volcanic values, while the NAPIS-75 and HR Sint profiles appear, in turn, consistent with each other. The most direct interpretation is that short-term dipolar variations were smoothed by the low-resolution stacking of Sint-200 and by the averaging process used to construct the volcanic curve. A good exemple is provided by *Carlut and Quidelleur* [2000], who reported absolute paleointensities from dated lava flows at La Guadeloupe Island. They noticed the existence of a 20-kyr-long oscillation present in volcanic data around 80 kyr B.P., which appears clearly in the high-resolution sedimentary records but is heavily smoothed in Sint-200. Prior to 15 ka, the volcanic record is more accurate and is characterized by an overall and almost monotonic decrease. Unfortunately, this period remains very poorly constrained by sedimentary data because it corresponds to the upper part of the records, which are mostly disturbed during coring. Studies using box cores of marine sediments would be helpful.

[64] *Laj et al*. [2000a] discussed the comparison between the NAPIS curve and the nearby record from the Blake Outer Ridge [*Schwartz et al*., 1998]. Both curves show the same large-scale features, but there are also differences, particularly before 50 ka and within the 10- to 20-kyr interval. The comparison has been pursued for other areas: the Mediterranean Sea, the Somali Basin, and the Labrador Sea [*Stoner et al*., 1995, 2000]. Surprisingly, in all cases, there is much better agreement between these independent and remote data sets than with the Blake Outer Ridge. To these records it would be interesting to add the data sets from Lake Baikal [*Peck et al*., 1996] and from Lac du Bouchet [*Thouveny et al*., 1993], which, to our knowledge, are the only long and well-dated records from lacustrine sediments. For some reason, which could be linked to magnetization acquisition in lacustrine sediments and to irregular deposition rates, only the very broad features of the Lac du Bouchet record match the other results. The presence of a large clay content in lake sediments, which favors aggregates of particles through various interactions including Van der Vaals forces, must also be invoked. Viscous drags become very important, and large fields are required to bring their magnetic moments into alignment. Since a significant amount of the small magnetic grains may not have been correctly oriented, it becomes delicate to extract a reliable signal of relative paleointensity using concentration parameters. Indeed, in this case the ratio of the magnetic grains carrying remanence to those involved in the normalizer cannot be considered as constant. Finally, chronology must also be taken into consideration. The NAPIS-75 stack was scaled to the $\delta^{18}O$ Greenland Ice Sheet Project (GISP2) ice core chronology [*Grootes and Stuiver*, 1997], which is different from the SPECMAP [*Imbrie et al*., 1984] or *Martinson et al*. [1988] $\delta^{18}O$ marine chronologies used in the other records. In fact, *Laj et al*. [2000a] obtained a satisfactory correlation between NAPIS and the Mediterranean Sea-Somali Basin composites after adjusting the same features without violating the timescales (at least within the present uncertainties), but they noticed that the same degree of correlation could not be reached for the other records of the stack. In contrast to the Mediterranean Sea and Somali Basin these data sets were not obtained from multiple parallel records. Such deficient correlation emphasizes the importance of stacking adjacent records to eliminate all parasitic effects. Incidentally, this is also a clear indication that most discrepancies between parallel records are random in nature and thus might not only be relevant to climatic components, unless they were generated by complex associations of several factors. The millennial-scale variability present in all results is global in character, which further supports its dipolar origin. Thus, despite its high resolution, nondipole components are most likely not present in NAPIS-75.

4.2.3. Cosmonuclide Signal

[65] The strength of the geomagnetic field is the most important factor controlling cosmogenic production. A precise record of atmospheric $\Delta^{14}C$ variations over the last 30 kyr has been obtained from a comparison of $\Delta^{14}C$ and U/Th ages in corals [*Bard et al*., 1990, 1993, 1998]. The results depict a long-term increase in production rate, which represents a more or less monotonic decrease in geomagnetic intensity from 30 ka until now, and thus they appear to be consistent with the paleointensity stacks. This data set has been further confirmed by a high-resolution $\Delta^{14}C$ record [*Kitagawa and van der Plicht*, 1998] from varved sediments of Lake Suigetsu (Japan). However, this last record suggests that the field intensity would have been relatively high between 45 and 32 ka, in contrast to all other results. This may be caused by a missing part in the varved chronology and/or the fact that this period lies at the limits of ^{14}C measurements (five to eight half-lives). *Voelker et al*. [1998] correlated stable carbon and oxygen isotope records of marine sediments from southwestern Iceland to the GISP2 ice core oxygen isotope record. Despite several possible uncertainties in the various timescales the results confirmed that field intensity was very low between 32 and 42 ka.

[66] Independent measurements of ^{10}Be concentration in several polar ice cores [*Raisbeck et al*., 1992; *Beer et al*., 1992] and in sediments from the Mediterranean [*Cini Castagnoli et al*., 1995] and Caribbean [*Aldahan and Possnert*, 1998] Seas revealed a large peak in ^{10}Be around 35 ka, thus confirming the low relative paleointensities previously reported for this period, which is coeval with the Laschamp event [*Gillot et al*., 1979]. A large increase was also reported during this period in the ^{10}Be/^{9}Be ratio [*Robinson et al*., 1995] measured in an 80-kyr-old sediment core from the northern Atlantic Ocean.

4.3. Stacking Independent Records of Paleointensity

4.3.1. Past 200 kyr: Converging Indications for a Geomagnetic Origin

[67] Presently, the date of 50 ka stands as the oldest limit for which a global pattern of past geomagnetic intensity can reasonably be extracted using volcanic flows. This situation is illustrated in Figure 10a. There are not enough records within successive time intervals (even as large as 10 kyr) to calculate a mean field value that would not be sensitive to local field components. Among recent studies involving lava flows older than 50 ka, one can quote records from the East Pacific Rise [*Pick and Tauxe*, 1993b], the Etna volcano in Sicily [*Tric et* al., 1994], Réunion Island [*Rais*, 1996], the Eiffel volcanics [*Schnepp*, 1996], and the Hawaii Scientific Drilling Project core from Hawaii [*Laj and Kissel*, 1999]. As indicated by the results of the simulation described in section 3.2.1 (see also Figure 8), many additional studies remain necessary to extract global field changes for the past 780 years from volcanic records. We must therefore turn back to sedimentary records.

[68] Despite the overall good agreement of the Sint-200 stack curve with the volcanic records for the past 45 kyr (see Figures 9a and 9b) and the existence of common features with the variations deduced from production rates of cosmonuclides, the geomagnetic origin of the Sint-200 variations remained to be assessed by an independent approach extending over the past 200 kyr covered by the data. For this reason the construction of a global stacked record of ^{10}Be production has been a significant contribution [*Frank et al*., 1997]. The ^{10}Be composite curve was derived from 19 marine cores covering the last 200 kyr and from 18 shorter records covering the Holocene and the last glacial period. All marine cores, as well as their site distributions, were different from the ones used in the Sint-200 stack. The results shown in Figure 10b are in good agreement with the entire Sint-200 record. Independent confirmation was subsequently given by chlorine 36 data obtained from the Greenland Ice Core Project (GRIP) [*Baumgartner et al*., 1998; *Wagner et al*., 2000] after comparison with the stack paleointensity record from the Somali Basin [*Meynadier et al*., 1992], which is identical but with better resolution than Sint-200. Finally, *Laj et al*. [2000a] compared their high-resolution NAPIS-75 data set to the same ^{36}Cl GRIP record (Figure 10c) and observed that both studies match very well within the limits imposed by the timescales (in this last case the timescale of the magnetic record was matched to the GRIP timescale, while the marine timescale was used for ^{36}Cl). In summary, there is very good convergence between independent records that were obtained either from high-resolution (NAPIS and ^{36}Cl) records or from lower-resolution (Sint-200 and ^{10}Be) records.

[69] All these results must be seen as constraints for significance and the future potential of relative paleointensity data from sediments. Indeed, three major observations suggest that the signals have a common geomagnetic origin.

1. The volcanic data for the past 40 kyr are in rather good agreement with records of relative paleointensity, although additional refinement is needed, particularly between 20 and 40 kyr B.P.

2. There is an overall convergence between results obtained with four various and independent techniques (lava flows, sediments, ^{10}Be, and ^{36}Cl).

3. Multiple records from different sites and nearby data sets from the same area show similar variations. Note also that in several cases (e.g., Mediterranean Sea and Somali or South Pacific and Atlantic) the magnetic parameters used as normalizers can have opposite responses to the climatic variations (highs in one case and lows in the other case during glacial intervals). Contaminated records of paleointensity should thus be different between these areas, which is not the case.

[70] A controversial view on this matter was put forward by *Kok* [1999], who considered that both ^{10}Be and Sint-200 stacked records show some climatic modulation at the orbital (41, 23 and 19 kyr) periodicities, although both are relatively short for a correct analysis of the 41-kyr periodicity. This question has been addressed by *Frank* [2000], who scrutinized in detail the independent data sets and parameters for particular periods. There could be some correspondence between the ^{10}Be stack-based paleointensity record and δ^{18}O (e.g., during the climatic substage 5e, i.e., between 130 and 111 ka), but it is clear that the ^{10}Be fluxes are quite similar to the stacked paleointensity record. It is difficult to envisage why a stack of paleomagnetic data, a stack of ^{10}Be measured in sediments, and records of atmospheric Δ^{14}C as well as cosmogenic radionuclide fluxes deduced from ice cores would reflect climatic forcing in the same way and to the same extent. Behind *Kok*'s [1999] suggestion was the hypothesis that all records in question would be contaminated by climatic changes. This leads us to examine the question of a possible influence of Earth's rotation on changes in field intensity.

4.3.2. Records Covering the Brunhes Chron (0–780 ka)

[71] The fast growing database made possible a stacking of the results for the past 800 kyr [*Guyodo and Valet*,

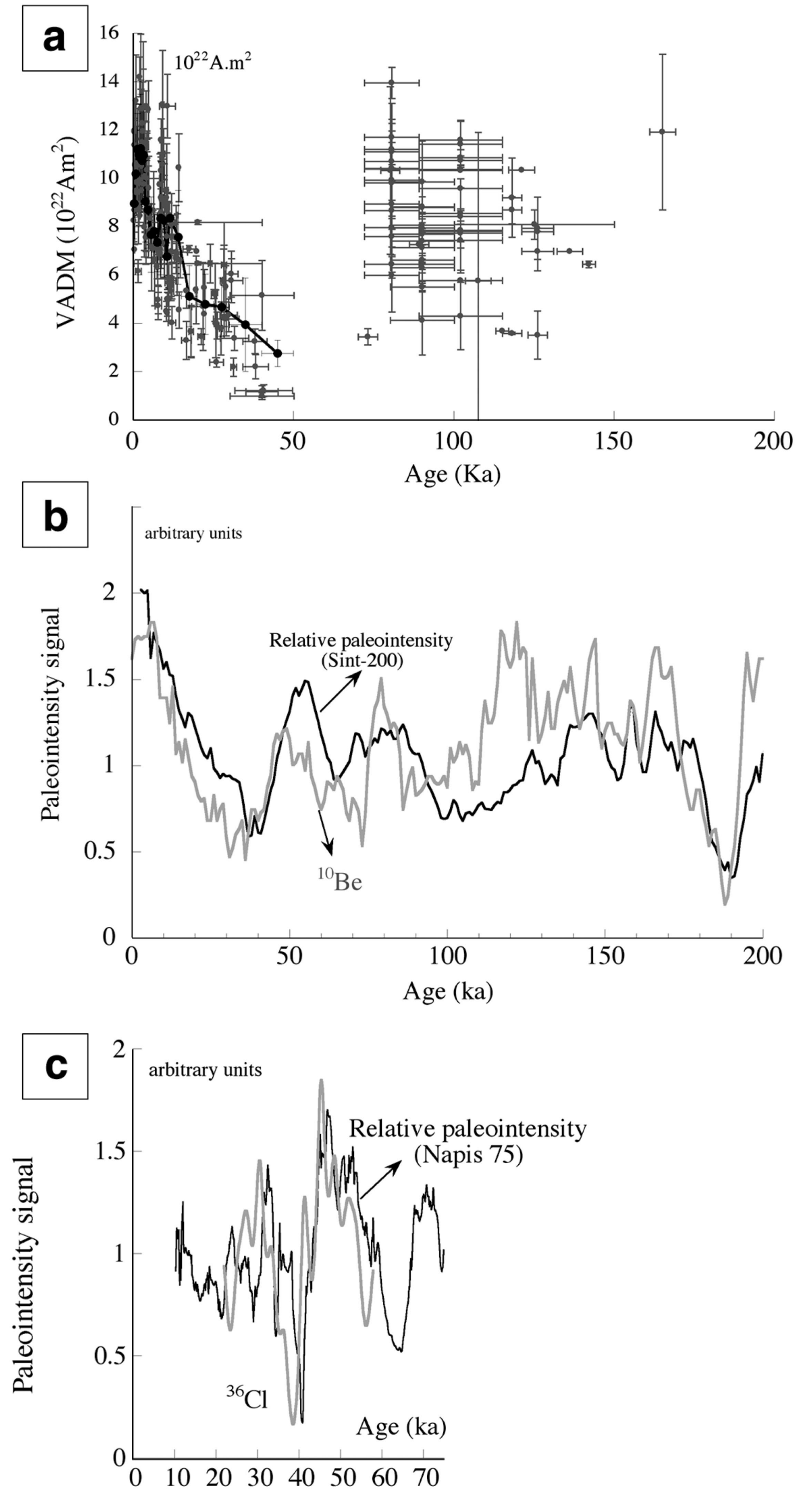

Figure 10. (a) Volcanic VDMs determined from the Thellier measurements extracted from the updated global database of absolute paleointensity. Also shown is the same time-averaged volcanic record as in Figure 9. Note the large scatter partly caused by the absence of reliable dating beyond 50 ka. (b) Composite records derived from measurements of relative paleointensity (dark curve) and ^{10}Be (shaded curve) production rate [*Frank et al.*, 1997]. Dating relies on marine chronology. (c) The ^{36}Cl production rate [*Baumgartner et al.*, 1998] converted in terms of field intensity changes compared with the NAPIS-75 record. The two data sets have been dated using the Greenland Ice Core Project timescale (ice chronology).

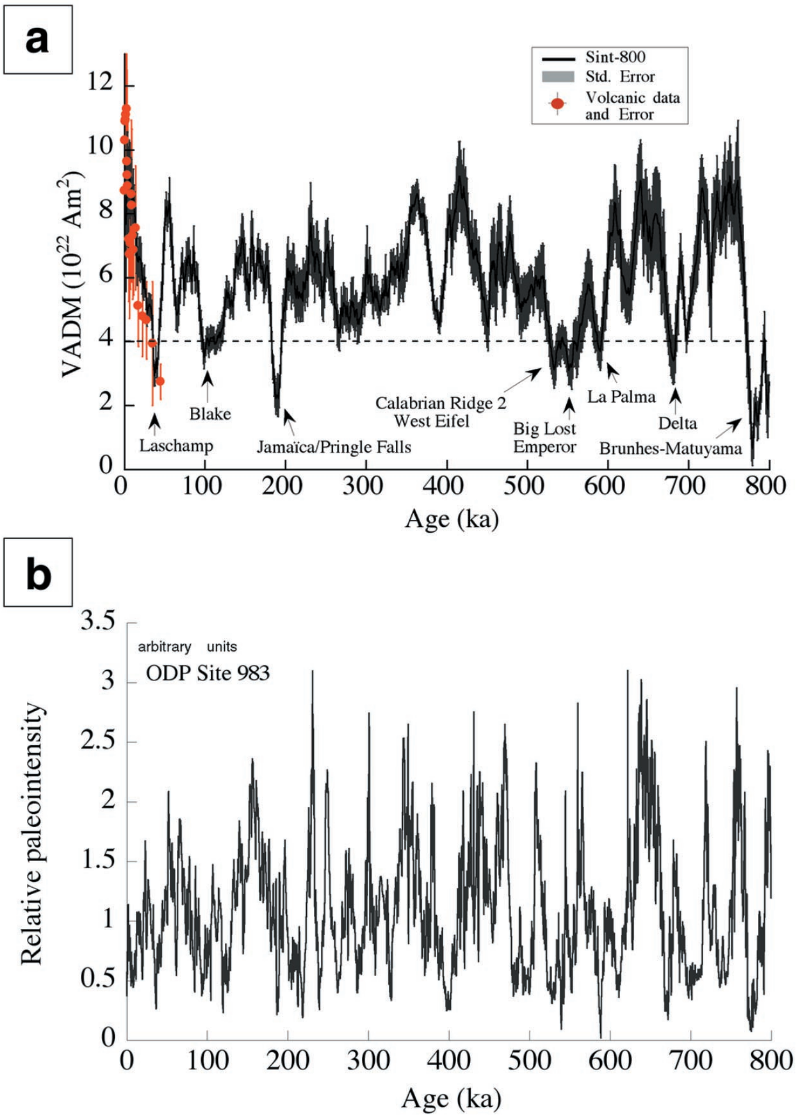

Figure 11. (a) Signal of the axial dipole moment (Sint-800) obtained over the last 800 kyr [*Guyodo and Valet*, 1999b] by stacking 33 records of relative paleointensity (calibrated with time-averaged volcanic VADMs for the period 0–40 kyr B.P.). (b) Record of relative paleointensity from Ocean Drilling Program site 983 over the past 800 kyr [*Channell and Kleiven*, 2000]. Similar features of different amplitudes are common to both records.

1999b]. The resulting curve was constructed from 33 paleointensity records [e.g., *Schneider and Mello*, 1996; *Roberts et al*., 1997; *Yamazaki et al*., 1995; *Channell et al*., 1998; *Guyodo et al*., 1999] and is shown in Figure 11a. The first evident characteristic is the existence of the variable character of the field, which depicts a succession of large 20- to 60-kyr-long oscillations, and changes in amplitude that can exceed a factor of 5. Short periods of very low intensity occur at more or less regular intervals. This observation will be discussed in section 4.3.3.

[72] Since Sint-800 incorporates a very large number of records, it is reasonable to consider that the differences in amplitude between distinct records have been averaged out. A first calibration of Sint-800 can be performed by adjusting the 45-kyr part in the curve with the time-averaged volcanic records shown in Figure 9. This operation gives a mean value of (6.0 +/− 1.5) × 10^{22} A m^{-2} for the past 800 kyr. However, such calibration remains quite limited in time and not exempt from possible uncertainties caused by disturbances that are frequently met in the upper part of the records and thus is to be considered with more caution. In other words, despite overall consistent patterns of the volcanic and sedimentary data sets that indicate their common origin, the amplitudes of the upper part of the sedimentary records remain delicate. The alternative option is to tentatively calibrate the entire record using the few sparse available volcanic records obtained by Thellier technique while keeping the correlation as consistent as possible with the past 45 kyr. In this case the calibration suffers from the large dispersion of the volcanic VADMs and cannot be constrained better than with average

VADMs between 6×10^{22} A m^{-2} and 8×10^{22} A m^{-2}. Attempts for more accurate correlation using sophisticated fitting techniques are not justified given the temporal distribution and the error bars inherent to the volcanic data set.

4.3.3. Intensity Lows

[73] The most direct and common feature present in composite curves of relative paleointensity as well as in all individual well-dated records is that many intensity lows punctuated the evolution of geomagnetic intensity during the past 800 kyr. Another important characteristic is that most large decreases are coeval with acknowledged excursional events as summarized in Figure 11a. This observation strongly reinforces the concept that the presence of "excursional" directions results from the dominance of nondipole components subsequent to the large decrease of the dipole. Several papers [*Champion et al*., 1988; *Nowaczyk et al*., 1994; *Langereis et al*., 1997; *Lund*, 1997; *Lund et al*., 1998] have reported compilations of excursions during the Brunhes (0–780 kyr B.P.) normal chron. According to *Langereis et al*. [1997], six excursions can be reliably considered as worldwide events, whereas five others have not been globally correlated. This suggests that nondipole components did not emerge everywhere in these last five cases, because the dipole intensity remained larger than during the six excursional events. However, this interpretation must be considered with caution, since the definition of intensity lows depends on the resolution of the records, which in the case of Sint-800 is not very high. In fact, all long records obtained so far from sediments of various epochs depict a field evolution dominated by a succession of highs and lows, similar to the past 800 kyr. Thus major characteristics of field intensity variations remained most probably unchanged in the last several million years at least. It is not unlikely that excursions punctuated the entire history of the geomagnetic field and could thus be reflected by the large number of cryptochrons found in the records of marine magnetic anomalies. In any case, paleointensity records depict a much more unstable image of the field than was generally believed.

[74] The volcanic database provides us with a confusing image for this period because of poor temporal coverage and spatial resolution. However, records from detailed volcanic sequences could help constrain field behavior during the very short periods of time covered by excursions and/or reversals. Detailed records of excursions with absolute paleointensity have been obtained during the past few years [*Roperch et al*., 1988; *Chauvin et al*., 1989, 1991; *Camps et al*., 1996; *Carlut et al*., 1999]. All results confirm the existence of a large decrease of the field. The only exception is a 14.3-Myr-old excursion recorded in sequences of basalts from Gran Canaria [*Leonhardt et al*., 2000] with excursional field values almost as large as the present field. The preexcursional and postexcursional directions are characterized by low paleointensity, which eliminates the hypothesis of a rotating dipole. The field values are also much too high to be explained by nondipole components. A different explanation could be that paleointensity values were determined from low-temperature segments and thus two different slopes. In this case the low-temperature segment is characterized by a steep slope, which results from multidomain grains (see section 3.1.2) and evidently generates high values of paleointensity.

[75] *Gubbins* [1999] suggested that during excursions the field could reverse in the liquid outer core but not in the solid inner core, which is characterized by longer timescales (3–5 kyr) than the short dynamics in the fluid core. Although very appealing, this suggestion is extremely difficult to test. Given the time resolution of the records, it is not so obvious that excursions are shorter than reversals. Another feature that is common to all records is the absence of field recovery (during the excursional period) before the field returns to initial polarity. Note that this has been observed also in the sedimentary records, such as Mono Lake [*Liddicoat*, 1992], the Laschamp event [*Laj et al*., 2000b], and the Iceland basin event [*Channell*, 1999]. Maybe the absence of recovery could be a sufficient condition for a rapid return to initial polarity, while intense recovery following a geomagnetic instability could be a necessary condition for generating stable polarity. This suggestion emphasizes the importance of scrutinizing long periods of stable polarity.

4.4. Earth Orbit and Intensity Changes

4.4.1. Geomagnetic Versus Climatic Components

[76] It is now generally accepted that the geomagnetic field is maintained against ohmic dissipation by dynamo action in the electrically conducting liquid outer core. After it had been discarded as a possible driving mechanism by *Bullard* [1949] and *Elsasser* [1950], *Malkus* [1968], *Stacey* [1973], and *Vanyo* [1991] speculated that precession could be powerful enough to stir the core into dynamo action or at least to have a significant impact. The concept was finally rejected as not being energetically adequate by *Rochester et al*. [1975]. It is probably not surprising that a large number of studies have been aimed at detecting the spectral signature of the Earth's orbital parameters in paleointensity records. *Kent and Opdyke* [1977] were the first to find spectral power centered at a period of ~43 kyr in a Brunhes deep-sea core. However, the time resolution and the absence of a detailed chronology were limiting factors. Similarly, the suggestion by *Meynadier et al*. [1992] that the 140-kyr-long records from the Mediterranean Sea and Somali Basin sediments could be modulated by a 23-kyr-long period cannot be validated because of the short time period covered by the record. These studies certainly emphasize the difficulties of extracting any periodic signal with a sufficient confidence and the importance of

analyzing time periods that are long enough with respect to the studied periodicities. The situation is also complicated since we are dealing with climatic variations that can actually contaminate the NRM signal. In this case the presence or absence of a causal link with geomagnetic variations [*Amerigian*, 1977; *Chave and Denham*, 1979; *Kent*, 1982] needs to be investigated. For these reasons, results from a single site must be viewed with caution when related to global geomagnetic changes. Alternatively, composite curves do not provide optimal resolution. This may explain why controversial (and sometimes contradictory) interpretations [*Creer et al*., 1990] have been published on this matter.

[77] Very few records have been obtained with a resolution similar to those from ODP sites 983 and 984 (Figure 11b) and 1021 [*Channell et al*., 1998, *Channell*, 1999; *Guyodo et al*., 1999]. The results from sites 983 and 984 were very appealing not only because of their resolution but also because *Channell et al*. [1998] claimed that they were modulated by 100- and 41-kyr periodicities, arguing that the 41-kyr period was not present in the rock magnetic parameters used as normalizers and thus not attributable to lithological variations. However, this implies that this component was essentially present in the NRM signal, which could thus be driven by a climatic influence as well. Indeed, envisioning a larger climatic influence on lithological variations such as magnetic grain sizes rather than on concentration cannot be excluded. *Guyodo and Valet* [1999b] analyzed different time windows of the Sint-800 stack and found no stable period that could be related to orbital forcing. Stacking individual records, even after proper adjustment of the timescales, could reduce the spectral definition of the results. This is unlikely for time constants as long as 41 kyr, and none of the other records involved in Sint-800 show striking evidence for periodic behavior. Going back to site 983, *Guyodo et al*. [2000] used wavelet analysis and identified intervals over which the paleointensity record showed some coherency with the normalizers. This suggested that the orbital frequencies are the expression of lithological variations and probably not a characteristic of the geodynamo. They also noticed a clear relationship between NRM and grain size, which confirmed the strong influence of climatic forcing on the NRM.

[78] Following a similar approach, *Yokoyama and Yamazaki* [2000] used wavelet transforms for five low- and middle-latitude sedimentary records from the Pacific Ocean and concluded the existence of a 100-kyr period, which they related to the Earth's orbital eccentricity. In this case also, the rock magnetic parameters do not show coherent variations between cores or even negative correlation for two cores, which is evidently in favor of the geomagnetic hypothesis. Yokoyama and Yamazaki indicate that the peaks and valleys of the 128-kyr wavelet component in their composite curve correlate with the eccentricity of the Earth's orbit but with different time lags. They note that, in contrast, there is a constant lag of 18 kyr with the oxygen isotope record of *Bassinot et al*. [1994]. This is surprising because the 100-kyr pulsation of this last data set is supposed to be in phase with eccentricity. Also puzzling is the fact that the intensity dips observed in a previous composite record including three of the present Pacific cores [*Yamazaki et al*., 1995] can be correlated with similar dips (in particular, at 30–50, 110–120, 180–200, 280–300, 380–410, and 520–550 ka) in other records that do not show any periodic change. One could thus wonder whether the apparent 100-kyr period could not be caused by some mismatch between timescales.

[79] An interesting and alternative view has been proposed by *Teanby and Gubbins* [2000], who simulated paleomagnetic samplings of a synthetic 100-kyr-long series. The sequence has the same spectrum as the 12-kyr archeomagnetic record with no period longer than 10 kyr. Teanby and Gubbins observed that sampling of intervals longer than 2 kyr using standard 2-cm-wide cubes causes aliased energy to appear in the power spectrum of paleointensity, in particular long periods that were initially not present. In contrast, they noted that postdepositional reorientations act as a smoothing and should prevent aliasing in low-sedimentation rate cores.

[80] Note that, although not dominant, the presence of eccentricity was also detected at ODP sites 983 and 984 [*Channell*, 1999]. In fact, these results are reminiscent of a study by *Worm* [1997], who reported a relationship between occurrences of geomagnetic events associated with intensity lows and cooling episodes in the eastern equatorial Pacific records from sites 848 and 851 [*Meynadier and Valet*, 1995]. However, the relationship is highly questionable; in any case, no prominent and spectral analyses of these records revealed any dominant climatic period despite the presence of climatic components revealed by δ^{18}O studies in the same cores [*Ravello and Shackleton*, 1995].

[81] A study of three combined records from the Ontong-Java Plateau [*Tauxe and Shackleton*, 1994] spanning the last 400 kyr did not show a dominant period that would fit the values of the orbital parameters but did show a 33-kyr signal, which lies between obliquity and precession. The possibility that this signal could actually be related to the 26-kyr astronomical precession was ruled out by Tauxe and Shackleton. Indeed, this is inconsistent with the existing isotopic ages, unless it is considered that the orbital parameters would primarily dominate the field with respect to the ice volume, which is not possible. *Tauxe and Shackleton* [1994] did not analyze further the origin of this 33-kyr component.

4.4.2. Analyses of Oligocene Records

[82] Analyses of sediments from Deep Sea Drilling Project site 522 [*Tauxe and Hartl*, 1997] showed the presence of a similar 30- to 50-kyr periodicity and led Tauxe and Hartl to consider that the field would pulse at this frequency, not only during the Brunhes but in the

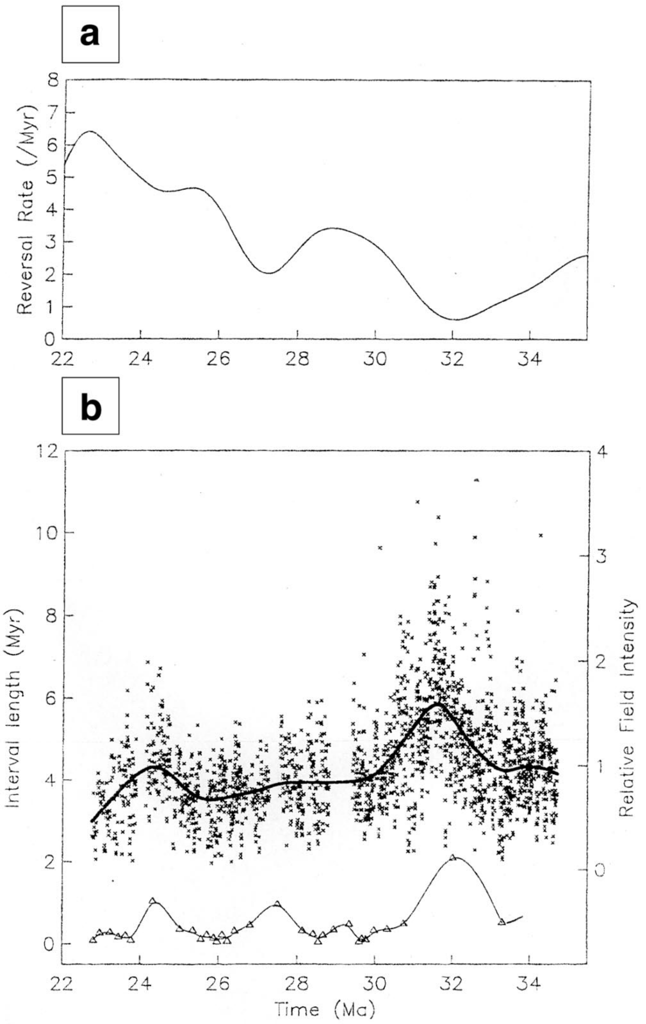

Figure 12. Absolute paleointensity and reversal rate [from *Constable et al.*, 1998]. (a) Estimated reversal rate as a function of time during the Oligocene. (b) Paleointensity derived from a 12-Myr-long Oligocene record [*Tauxe and Hartl*, 1997] after smoothing of individual measurements (crosses) by splines (bold line). The lengths of polarity intervals are shown as triangles (plotted at the midpoint of the interval) in the bottom of Figure 12, while the thin line depicts the corresponding trend in interval length that correlates with the overall trend in paleointensity. Note the presence of higher paleointensity values between 29.4 and 34.7 Ma, which correspond to a lower reversal rate. Constable et al. note that the spectra of intensity variations would indicate a different secular variation regime during this period.

Oligocene as well. This last record (Figure 12) covers 12 Myr of field intensity variations during the Oligocene (between 22.74 and 34.74 Ma) and was sampled at 3- to 4-cm intervals. The sediment accumulation rate varies between 0.5 and 1 cm/kyr, so that the resolution obtained with 2-cm-long standard cubic samples cannot be better than 3–4 kyr and may be as low as 6–8 kyr. Considering that the Nyquist frequency is 16 kyr, the 30-kyr period can be at the limit of detection. To perform their spectral analysis, Tauxe and Hartl used cores that do not span reversals because they considered that the character of the spectrum would be affected by the intensity dips associated with the reversals. In a subsequent analysis of the same sediments, *Constable et al.* [1998] noticed that the short lengths of the records that were investigated by this technique made the peaks that they observed barely significant. Longer sections were analyzed by *Constable et al.* [1998], who concluded that the apparent cyclicities were in fact "of an ephemeral nature."

[83] *Constable et al.* [1998] used the same Oligocene record (Figure 12) to calculate the power spectrum of field intensity and proposed that during the time interval 22.74–28.77 Ma the geomagnetic intensity variations would have been controlled by the reversal process, which took place, on average, 4 times every million years. However, between 34.7 and 29.4 Ma, when the reversal rate is lower, the spectra indicate a different regime with substantially more power overall in the intensity variations and a peak in spectral power at ~8 Myr^{-1}. Among other possibilities this could be caused by many dips related to aborted reversals. *Constable et al.* [1998] wondered why there would be in this case so many consecutive unsuccessful reversals and why the secular variation during this time has so much more power associated with it. One could speculate that the low reversal rate and the existence of higher average field intensity could have some link. In other words, the field would be in too high an energy state to generate complete reversals. Note also that the lower part of the record is associated with higher deposition rates and thus is able to record more rapid variations, which, in turn, induce more power into the signal. In any case, this led us to explore further the origin of intensity dips in records and their possible relationship with field reversals. We conclude this section by stating that at this stage, there does not seem to be conclusive evidence for a periodic modulation of geomagnetic intensity that could be related to the orbital parameters of the Earth.

4.5. Long-Term Trends in Field Intensity

4.5.1. Asymmetrical Variations Across Reversals

4.5.1.1. Sedimentary Records

[84] It is essential to detect whether specific conditions prevail before the field reverses and if there is any influence of reversals on dynamo operation. So far, very few records of relative paleointensity reached periods older than 1 Ma. The major characteristic that emerges from the data covering the past 4 Myr (Figure 13a) is a saw-toothed shape, consisting of a series of long-term decreases of field intensity during intervals of stable polarity, followed by large and rapid recovery immediately after the reversals. This pattern is apparently present in several records from the equatorial Pacific and the Indian Oceans [e.g., *Valet and Meynadier*, 1993; *Meynadier et al.*, 1994, 1995; *Valet et al.*, 1994; *Thibal et*

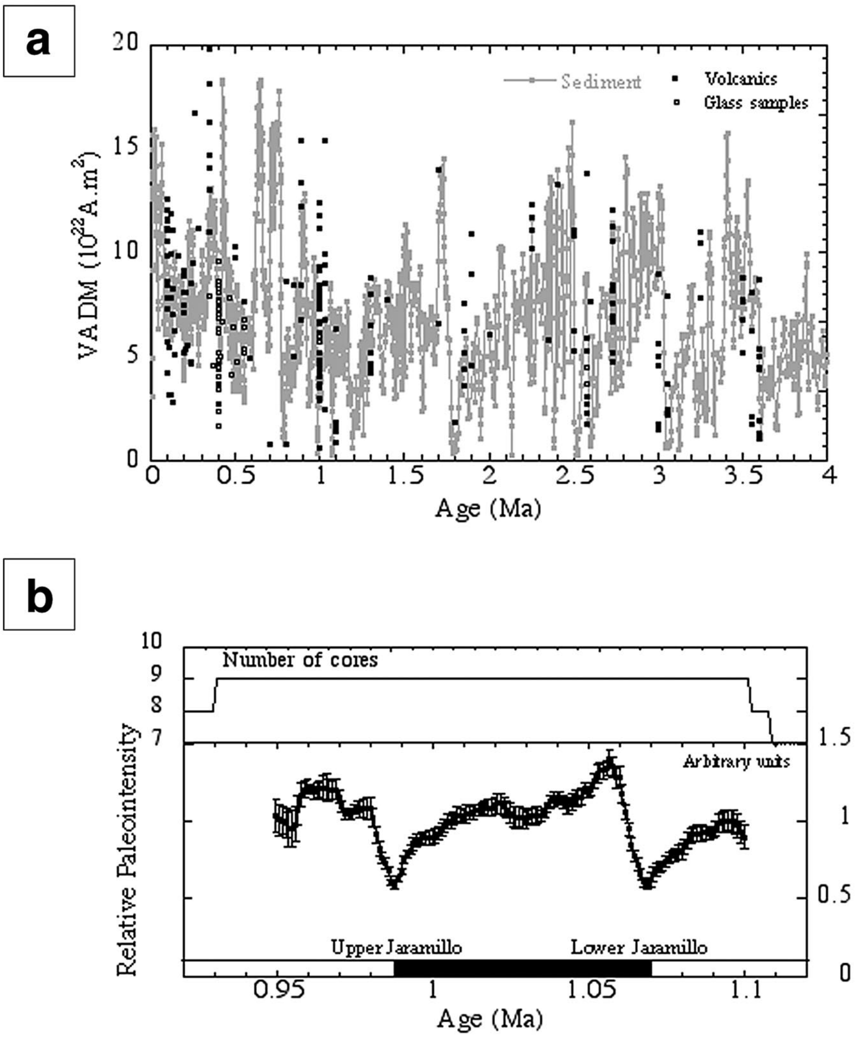

Figure 13. (a) Comparison between the 4-Myr-long record of relative paleointensity from the east equatorial Pacific and Indian Oceans data set [*Meynadier et al*., 1994] and the absolute determinations of VADMs by Thellier techniques extracted from the global database (solid circles). Also, the determinations derived from glass samples are shown by open circles. Polarity is indicated by black/white (normal/reverse) rectangles. (b) Field intensity variations over the 0.95 and 1.1 Ma time interval [*Guyodo et al*., 2001]. The number of cores involved in the stacking procedure is given at the top of Figure 13b. The error bars correspond to the standard error.

al., 1995] but is absent in others [e.g., *Tauxe and Shackleton*, 1994; *Tauxe and Hartl*, 1997]. Before going further in the discussion, it is noteworthy that different authors probably have quite different concepts of the meaning of the term. The sawtooth was originally associated with a long-term decrease of field intensity during the entire polarity interval. However, large-amplitude fluctuations are frequently present over polarity intervals, which in many cases are probably associated with field excursions. In fact, the main issue is certainly to constrain the opposite behavior between the prereversal and postreversal periods, the first one being associated with a long-term decrease and low field intensity, whereas intense and rapid recovery occurs during the second phase. So far, this topic has generated stimulating discussions, new ideas, and rock magnetic experiments, which have contributed to our understanding of paleointensity signals.

[85] An important consequence has been to question the limits of the records. A first typical example was the study of core KK78O30 from the central equatorial Pacific. *Verosub et al*. [1996] noted that the record of the Jaramillo subchron was very similar to the other records from the Pacific with a typical saw-toothed pattern. *Laj et al*. [1996] extended the analysis to the entire core and, in contrast to the previous analysis, reported the absence of asymmetrical changes. However, the mean field intensity derived from core KK78O30 is at least twice as high between 1.2 and 1.8 Ma as during the recent period. This time interval is also characterized by very large changes in amplitude. Such features are unique to this record and thus are likely not associated with geomag-

netic behavior. Another example was the record from site 983 [*Channell and Kleiven*, 2000] in the northern Atlantic. In this case, Channell and Kleiven concluded the absence of monotonic decrease during the Jaramillo, which is the only complete polarity interval recorded in this data set. The record is extremely detailed, but it may not be exempt from climatic contamination [*Guyodo et al.*, 2000] with large periodic variations that increase or introduce highs and lows in the signal, thus making estimates of the amplitudes quite delicate. For this reason the record displays similar features as in other records but with different amplitudes. This is an illustration of the present difficulties in assessing the exact origin of the changes in amplitude. Finally, it is also interesting to consider the record of ^{10}Be production rate that has been obtained by *Raisbeck et al.* [1994] across the last reversal. The results did not show any asymmetrical behavior, but they were inconclusive since the ^{10}Be production curve failed to show any variation during the reversal, while it is well established that a large intensity drop accompanies the polarity change.

[86] If it is assumed that saw-toothed field intensity changes do not exist, there is a need for an alternate explanation of observations. Two nongeomagnetic possibilities have been proposed. For the first possibility it is assumed [*Meynadier and Valet*, 1995; *Mazaud* 1996; *Meynadier and Valet*, 1996] that the magnetization would result from progressive orientation of magnetic grains with depth. However, at least 50% (65% [*Mazaud*, 1996]) of the magnetization must be locked in over a few centimeters below the sediment surface; otherwise, there would be an offset in the position of the reversals, which is not observed in the records [*Meynadier and Valet*, 1996; *deMenocal et al.*, 1990]. In order to preserve the saw-toothed pattern the rest of the magnetization must be locked in over depth scales of several meters (up to 10 m for *Meynadier and Valet* [1996] and 4–6 m for *Mazaud* [1996]). Thus such models rely on specific profiles for magnetization, which seem difficult to link with physical processes in the sediment, at least in the present state of knowledge. The second hypothesis [*Kok and Tauxe*, 1996a] defended the concept of a cumulative viscous component that would not have been removed properly by alternating fields. Notwithstanding the potential consequences for paleomagnetic records in general (most, if not all, sedimentary records have been af demagnetized), this model requires very specific hypotheses about the distribution of relaxation times [*Meynadier et al.*, 1998]. Additional thermal demagnetizations indicated that there was no persistent overprint, whereas discrepant results were reported [*Kok and Tauxe*, 1996b] for the oldest polarity interval. In fact, it turned out [*Meynadier and Valet*, 2000] that the only secondary component found in the samples was a drilling-induced component carried by multidomain grains, and thus the component was much less resistant to alternating fields.

[87] In conclusion, neither of these two models seems to rely on a more robust basis than the geomagnetic hypothesis. These results demonstrate again the importance of stacking records to eliminate factors that prevent one from safely interpreting the origin of changes in amplitude. At present, nine independent records cover the 0.95- to 1.1-Myr period. *Guyodo et al.* [2001] attempted a stack for the time period encompassing the Jaramillo subchron after rescaling the individual records using a common age model. The composite record of relative paleointensity obtained in Figure 13b is consistent with asymmetrical changes across the lower (1.07 Ma) and the upper (0.99 Ma) Jaramillo reversals. According to the depth versus timescale it took the geodynamo between 7 and 12 kyr to restore a maximum field intensity. Thus the first stacks [*Meynadier and Valet*, 2000] across successive reversals seem to indicate that there is an alternation of rapid field restoration and long-term declines with intensity minima coincident with short events or reversals.

[88] *McFadden and Merrill* [1997] refuted that a strong dipole field can inhibit the reversal process, since statistical analyses [*McFadden and Merrill*, 1993; *Merrill and McFadden*, 1994] of reversal chronology showed that inhibition of a future reversal would not exceed 50 kyr. This argument followed the suggestion [*Valet and Meynadier*, 1993] that the duration of polarity intervals would be linked to the amplitude of field recovery following the previous reversal. We have seen above that several factors can affect the amplitudes of the records, which makes this suggestion uncertain as it was drawn from a unique data set. *McFadden and Merrill* [1997] consider that a relationship between field recovery and polarity duration would imply that there has been a massive inhibition in the reversal process just after the Cretaceous superchron (long period without reversals), whereas the statistical analyses of the polarity timescale indicate that the inhibition in the reversal process goes the other way. However, there is no indication that there was a strong inhibition at the end of the Cretaceous. The possibility that the dynamo would have a reversing and a nonreversing state, both associated with different regimes [*Gallet and Hulot*, 1997], cannot be excluded. In fact, a 50-kyr inhibition period does not seem compatible with the existence of excursions (or large directional loops away the initial polarity) that are frequently found closely associated with reversals or with durations (25 kyr) of short events like that of Cobb Mountain [*Clement and Martinson*, 1992; *Gallet et al.*, 1993]. An example could be the volcanic record of the Réunion event [*Carlut et al.*, 1999] from Ethiopia, which is characterized by the absence of intensity recovery following the onset of the event. Actually, low intensity prevailed during the entire interval until the occurrence of the second transition, only 20–30 kyr after the first reversal. If we assume that field stability is related to the amplitude of the recovery, we, indeed, expect rapid occurrence of another reversal after the onset of the Réunion event.

4.5.1.2. Volcanic Records

[89] Because the sedimentary records revealed the existence of very large intensity variations during the past 4 Myr as well as for older periods, it is interesting to attempt a direct comparison between the two data sets. Since the past 4 Myr are by far better documented, the comparison has been restrained to this period using the Pacific and Indian Oceans records [*Meynadier et al.*, 1994] and the Thellier-Thellier records extracted from the global database of absolute paleointensity. The results are shown in Figure 13a. Considering the error bars on the volcanic records and the variability inherent in the nondipolar changes, there is a good first-order agreement between the two data sets. Many more volcanic records are evidently needed, but there is, indeed, a strong hope to improve the convergence. Incidentally, note that the volcanic database supports rather well the asymmetrical pattern displayed by the sediments (Figure 13a). According to this correlation the mean VADMs of $7 +/- 3.6 \times 10^{22}$ A m^{-2}, which is obtained from the volcanic data set, does not differ much from the value of $7.9 +/- 1.8 \times 10^{22}$ A m^{-2} derived from the sedimentary record. This is in better agreement (but very different) from a previous average value of $4.0 +/- 1.9 \times 10^{22}$ A m^{-2}, which was obtained by restraining the calibration of sedimentary record to the past 40 kyr [*Dormy et al.*, 2000].

[90] One of the first observations deduced from a long quest for detailed records of reversals in volcanic and sedimentary sequences was that field intensity dropped significantly during the transitional period. Most sedimentary studies do not allow going much further than this global description. *Lin et al.* [1994] analyzed five reversal records and noticed that large changes in intensity occur while the pole can still be at high latitudes. Only four studies including determinations of absolute paleointensity have been conducted so far from large volcanic sequences without any ambiguity about the succession and the origin of the volcanic units. Strong posttransitional field values have been reported in the Pliocene reversal recorded at Kauai [*Bogue and Paul*, 1993] and in the upper Jaramillo subchron recorded from Tahiti [*Chauvin et al.*, 1990]. The same characteristics emerge also from the 15-Myr-old Steens Mountain reversal [*Prévot et al.*, 1985], from the Matuyama-Brunhes transition recorded from La Palma in the Canary Islands [*Valet et al.*, 1999], and from the transition between chrons 27n and 26r recorded in basalts from west Greenland [*Riisager and Abrahamsen*, 2000]. Thus all five records, which represent so far the unique data encompassing reversals recorded in long sequences of overlying flows, provide support for asymmetrical changes in amplitude between the pretransitional and posttransitional paleointensity values, but the exact pattern of their evolution in time is extremely delicate to establish. This may explain why these results are in contrast with a 3.6-Myr-old excursion record from lava flows in southern Georgia [*Goguitchaichvili et al.*, 2001]. In this case, there is no asymmetry, but the origin of the event is not clear. It could be assigned to a full excursion as well as to some episode during a reversal. In the first case, asymmetry is not necessarily expected for excursions (this, in fact, could be a good way to discern different processes between excursions and reversals), while in the second case the recovery phase would evidently be missing. Goguitchaichvili et al. also argue that the existing data set is not in favor of high posttransitional field intensities, but they rely on incomplete data selection and underestimate high-quality and detailed records. As discussed in section 4.2.1, the major limit is the absence of detailed chronology, which is exacerbated by very irregular eruption rates. In fact, most of these records encompassed only short time periods around reversals. One exception is the La Palma record, which is characterized by five dated flows distributed between the lower and the upper part of the sequence representing a total duration of 0.3 Myr. However, the error bars on dating reach at least 10 kyr (which is expected for 0.8-Myr-old lavas). In such a case, there is no other possibility but to interpolate the ages of flows between two dated points, which inevitably distorts the signal. For this reason, and despite their similar patterns, it has been impossible to unambiguously correlate the La Palma record with the sedimentary curve for the same period.

4.5.2. Intensity Trends From Marine Magnetic Anomalies

[91] The origin of small-amplitude, short-wavelength marine magnetic anomalies, called tiny wiggles, has been debated for over 30 years. Tiny wiggles could result from the complexity of the magnetic properties of the extrusive basalt layer [*Gee and Kent*, 1994], but their correlation over large distances was more in favor of a geomagnetic hypothesis [*Cande and Kent*, 1992]. In this case these short anomalies could either be linked to short-period polarity intervals or to large-scale fluctuations in paleofield intensity. *Gee et al.* [1996] compared synthetic models of anomalies that were constructed from records of relative paleointensity with profiles of sea surface anomaly characterized by various spreading rates. They concluded that the oceanic crust could be a reliable recorder of field intensity changes. *Gee et al.* [2000] confirmed this assumption by showing that near-seafloor magnetic anomalies recorded over the southern East Pacific Rise are very well correlated with the paleointensity variations displayed by the Sint-800 composite curve. They also compared the pattern of the near-bottom magnetic anomaly data with paleointensity measurements of glass samples for the period covering the past 50 kyr and found enough agreement to conclude that large geomagnetic intensity fluctuations are recorded in the oceanic crust. Following a similar approach, *Pouliquen et al.* [2001] stacked six deep-tow profiles and several sea surface profiles of magnetic anomalies. The results were then successfully compared with predicted magnetic profiles derived from records of

relative paleointensity. Many short anomalies could be correlated with known geomagnetic events during the Brunhes and the Matuyama chrons and with several other short events during the Gauss chron. These studies point out the promises of deep-tow measurements to reveal such small-scale features of the field and confirm the overall presence of short-lived periods of low intensity.

[92] In order to constrain further the origin of the tiny wiggles, *Roberts and Lewin-Harris* [2000] presented paleomagnetic results from sediments with two short polarity intervals and another zone of anomalous directions between 9.92 and 10.95 Ma, a period characterized by three tiny wiggles reported from profiles of marine magnetic anomalies. They concluded that some tiny wiggles represent short-period geomagnetic polarity intervals, while others may represent geomagnetic excursions. The distinction is somewhat subjective. Indeed, for sediments as well as marine magnetic anomalies, the resolution of the records certainly plays a major role in the definition of the geomagnetic event. A second reason is that these features are essentially controlled by a large reduction in dipole intensity. If excursions and reversals are primarily periods of weak dipole field, with few statistical changes in the nondipolar terms, the whole spectrum of excursions and other geomagnetic instabilities would simply reflect fluctuations in dipole intensity. The smaller the fluctuation is, the more evasive the excursion is. In that case the changes in the axial dipole are part of the description and definition of such events, with no other special physical causal link.

4.5.3. Field Intensity and Reversal Frequency

[93] Records of relative paleointensity from sediments are much more difficult to obtain for periods older than a few millions of years. This is caused by difficulties in coring long sequences below a few hundred meters, since increased compaction requires more powerful drilling techniques, which then introduce severe physical disturbances in the sediment. Land-exposed sediments can be attractive, but they are frequently affected by alteration and tectonics, which is not compatible with paleointensity experiments. For these reasons, with the exception of one Oligocene record, the data set covering periods older than 5 Ma has been obtained from volcanic rocks.

[94] Among other goals the search for possible relationships between reversal frequency and dipole moment intensity has been challenging. Because of the scarcity of sedimentary records it is not yet possible to perform a real statistical treatment for each polarity interval. In their analysis of an 11-Myr-long Oligocene record from sediments, *Constable et al.* [1998] noted that the average paleointensity per polarity interval weakly correlates with interval length (Figure 14) and also that the variability in paleointensity is proportional to the average. The second relationship was defended on the basis that the variance is too high for measurement errors to be the single cause. The only data set that is long enough for a similar analysis is the 4-Myr-long eastern Pacific record [*Valet and Meynadier*, 1993]. In this case, there is also a relationship between mean paleointensity and polarity duration, which is actually expected from the saw-toothed shape of the record. Following what has been discussed previously, such a relationship is also sensitive to uncertainties related to the amplitudes of the signals, and it would be much more preferable to rely on composite stacks of independent records before drawing any conclusion, despite interesting convergences. We have seen striking evidence from the archeomagnetic and even from high-resolution sedimentary records that large variations in dipole intensity happen on short timescales. Estimates of dispersion are thus also dependent on variations in deposition rate, which affect the amplitudes.

[95] Another matter of interest has been to document the 35-Myr-long normal polarity without reversals called the Cretaceous superchron (118–83 Ma). Superchrons certainly indicate a strong influence of mantle on the core because core dynamics has no natural timescale that long. The idea that independent or different mechanisms would drive the dynamo during this period could be reflected by some typical average field strength and variance. *Pal and Roberts* [1988] relied on the existence of high-amplitude magnetic anomalies in the oceanic crust and on a few discrete determinations of paleointensity to conclude in favor of a stronger field during polarity superchrons. Recently, *Tarduno et al.* [2001] investigated eight time-independent volcanic units from the Rajmahal Traps (113–116 Myr old) through a study of whole rocks and plagioclase crystals. The mean value $12.5 +/- 1.4 \times 10^{22}$ A m^{-2} derived from the plagioclases was 3 times greater than the mean Cenozoic and Early Cretaceous-Late Jurassic estimates, which led Tarduno et al. to suggest a correlation between low reversal frequency and high geomagnetic field. Note that this relationship would go in the same direction as the observations drawn from short polarity intervals. Opposite results have been published recently by *Tanaka and Kono* [2002] from a 90 Ma Cretaceous basalt platform in Inner Mongolia. In fact, there is a small number of determinations obtained by double-heating techniques for this period, and their dispersion makes any interpretation delicate. There is no doubt that, again in this case, significant contributions will emerge from additional paleointensity data to clarify the situation between those who argue that different mechanisms operate in the dynamo at times of long superchrons with respect to the other periods and those who argue that the same very long-term stationary process governs reversal frequency. Thus it is interesting to document the periods preceding and following superchrons as well.

4.5.4. Mesozoic Dipole Low

[96] Considering that most previous studies were derived from unselected data, *Prévot et al.* [1990] carried

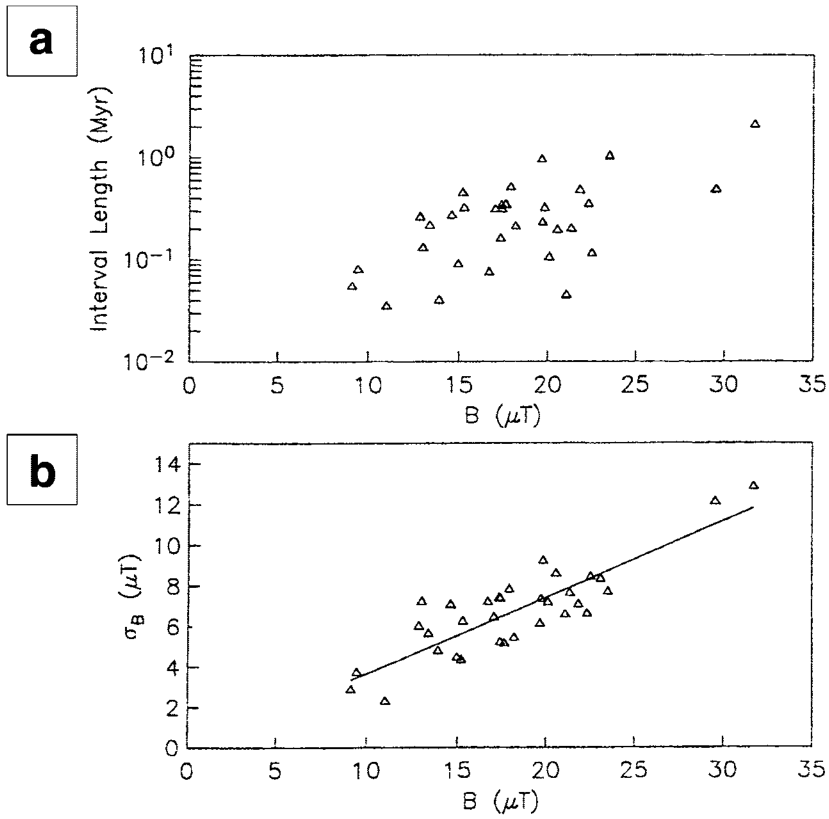

Figure 14. (a) Polarity interval length plotted as a function of average field intensity recorded in Oligocene sediments. (b) Standard deviation in field intensity as a function of average field intensity. Figure 14 is from *Constable et al.* [1998].

out a compilation of absolute paleointensities measured with the Thellier techniques. Figure 15a shows the evolution of the VADMs during the past 250 Myr, which results from a similar selection using the most recent version of the global database. Prévot et al. calculated an average field value for various geological epochs and found a low VDM, close to 3×10^{22} A m^{-2} from the Middle Jurassic to Early Cretaceous (roughly between 180 Ma and 120 Ma), which they defined as a "Mesozoic dipole low." This observation was initially dominated by data from the Caucasus, but additional measurements of a collection of tholeiite basalts from French and Iberian dykes by *Perrin et al.* [1991] reinforced the concept. Using volcanic glasses, *Pick and Tauxe* [1993b] found paleointensities for the beginning and the end of the Cretaceous normal superchron that were lower than today's value by more than a factor of 2. They considered that the "Mesozoic dipole low" should be extended into the Cretaceous superchron and thus cover the time interval between 180 and 80 Ma. However, *Juarez et al.* [1998] subsequently obtained low and high paleointensities during the Cretaceous superchron (Figure 15a). The existence of the dipole low was also examined by *Tanaka et al.* [1995b], who constructed a database from 1123 determinations of absolute paleointensity that were older than 0.03 Myr. Owing to limited site distribution they concluded that the Mesozoic dipole low could not be regarded as definitely established. The same conclusion is heavily suggested by the dispersion inherent in the results shown in Figure 15a, which rely solely on Thellier measurements, while the Tanaka et al. analysis did not make any distinction between various experimental techniques such as the *Van Zijl et al.* [1962] and *Shaw* [1974] af techniques, which can be questioned [*Perrin and Shcherbakov*, 1998; *Juarez et al.*, 1998]. Indeed, many records were based on two-slope Arai diagrams and were not accompanied by parallel Thellier experiments. Following the initial work by *Tanaka et al.* [1995b], *Perrin and Shcherbakov* [1997, 1998] compiled a second global paleointensity database. They used an approach similar to *Tanaka et al.* [1995b] for the past 10 Myr and observed a latitudinal dependence of the VDMs for the past 400 Myr. This is a strong indication in favor of the persistent dominant dipolar nature of the field through time, including the Mesozoic.

[97] *Selkin and Tauxe* [2000] used several criteria to compile a selection of reliable estimates over the past 300 Myr, to which they added 287 determinations (grouped within 53 age classes) derived from modified Thellier experiments from basaltic glasses. Since volcanic glasses do not have inclination records, the mean field values have been interpreted in terms of VADMs after taking into account the paleolatitude of the sites. More than 60% of the samples belong to the young ages

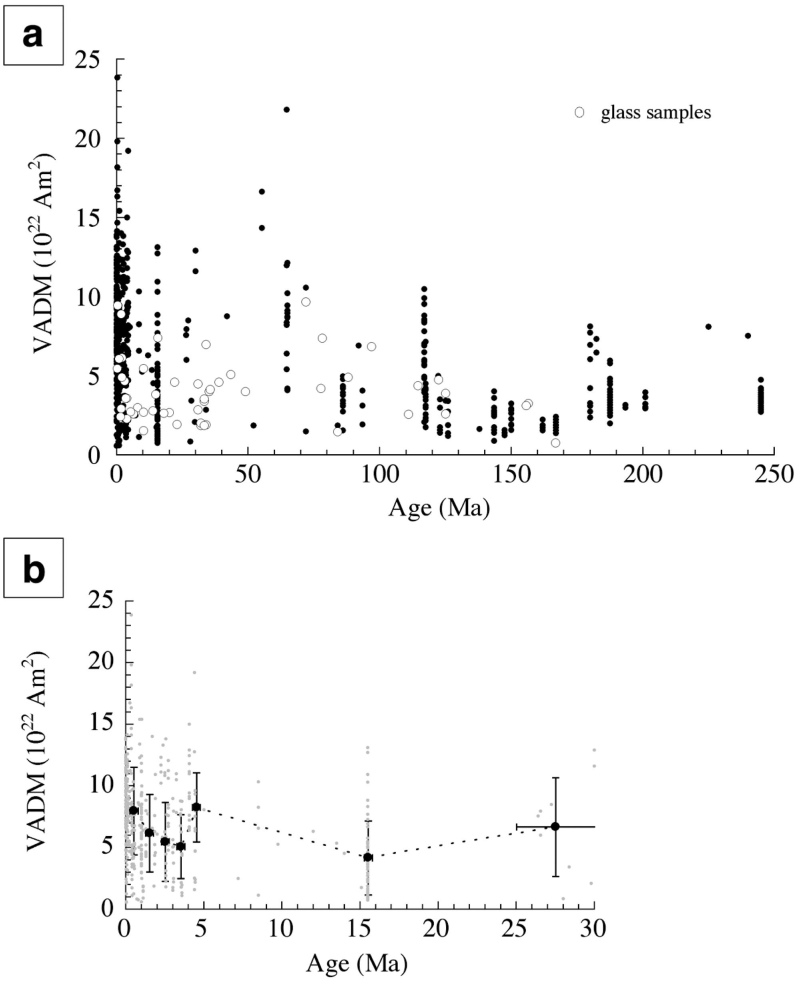

Figure 15. (a) Distribution of the VADMs as a function of time for the past 250 Myr using Thellier experiments from lava flows (solid symbols) and glass samples (open symbols). Given the absence of orientation for glass samples, only VADMs have been plotted. On average, determinations obtained from glass material appear to be lower than those from lava flows. (b) Distribution of the VADMs obtained from Thellier determinations in the global database over the past 30 Myr. Mean VADMs can be calculated within successive 1-Myr-long intervals over the past 5 Myr. Beyond this period the temporal distribution and the number of results is much too poor, most of them with poor dating contraints. Note also the existence of long periods without data.

(younger than 23 Ma), which leaves the older part of the record still poorly documented. The heavily sampled 0–0.3 Ma period has a mean VADM of 8.4 +/− 3.1 × 10^{22} A m^{-2}, which is higher than 4.6 +/− 3.2 × 10^{22} A m^{-2} during the previous 0.3–300 Ma period. *Selkin and Tauxe* [2000] argue that the cumulative distribution functions of the 0–0.3 Ma and 0.3–300 Ma data sets represent significantly different distributions. They also compared distributions of paleointensity during periods of fast and low reversal rate and found no significant difference. Such analyses may be hampered by the temporal distribution of the data. Using the results of Thellier measurements from the updated paleointensity database again, I tentatively averaged out the paleointensity values within successive time intervals. In fact, this operation becomes rapidly meaningless since entire periods are missing, while some others are overrepresented because of poor age constraints. Even within the past 30 Myr, there are long time intervals (e.g., between 5 and 15 Myr) with almost no data points (Figure 15b). Thus it may be delicate to compare a distribution constructed from a sampling density of 520 points per million years (0–0.3 Ma) with one including two points per million years (0.3–300 Ma). It is effectively observed that large VADMs are more common during the recent period. Fifty-four percent of the VADMs are found above 8 × 10^{22} A m^{-2} during the past 0.3 Myr, and only 15% are found during the 0.3–300 Ma interval; thus there is a difference of a factor of 3.5. However, the sampling density of 516 data points per million years within the 0–0.3 Ma interval is 200 times larger than during the 0.3–300 Ma interval (2.2 data points per million years). In other words, any period as long as the past 0.3 Myr would not be described by more than 0.75 data points

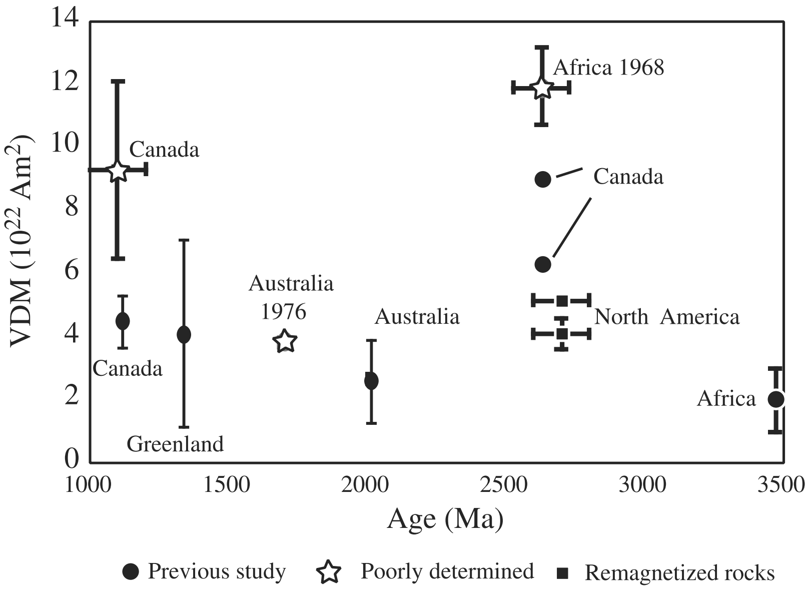

Figure 16. Present database of paleointensity determinations obtained for the Precambrian. The references for the records are indicated with the corresponding data points.

during the 0.3–300 Ma interval. Thus the actual geomagnetic significance of the distributions may be delicate to establish as long as additional heavily sampled time intervals will not be studied.

4.5.5. Paleofield During the Precambrian

[98] In addition to improving our current knowledge about the initial stage of the geodynamo, determining the field intensity during Archean and Early Proterozoic could provide constraints about the formation of the inner core. *Stevenson et al*. [1983] and *Buffett et al*. [1992] suggested that nucleation of the inner core occurred very early in the history of the Earth and might have been accompanied by a sudden increase in geomagnetic intensity. More recently, *Labrosse et al*. [1997] constructed a model for cooling of the Earth's core that shows that the inner core could not start growing earlier than 1.7 Ga (otherwise, it would be larger than its present size). Subsequently, *Labrosse et al*. [2001] incorporated radiogenic elements in the core and proposed that the inner core formation could not have started before 2.5 Ga and that a more realistic age is 1 ± 0.5 Ga. The growth of the inner core may cause important changes in the dynamo process. Following *Gubbins* [1999] suggestion, one could expect more excursions during the early days of field generation because of the stabilizing effect of the inner core and thus maybe some influence on analyses of mean field intensity.

[99] A first compilation of paleointensity determinations from Archean and Proterozoic samples from the Kaapvaal and Superior cratons by *Hale* [1987a] was interpreted as being consistent with a sharp increase in geomagnetic strength between 2.7 and 2.1 Ga. However, this analysis relied on a few data, which did not satisfy most recent quality criteria. The first studies of Precambrian rocks were performed more than 30 years ago, and some results are probably questionable. Remagnetizations and effects of anisotropy [*Selkin et al*., 2000] have probably not always been taken into account. During the past few years the number of paleointensity studies dealing with very old rocks has significantly increased, and there are currently several unpublished studies in progress. The present database shown in Figure 16 does not incorporate so many data, which fully justifies the interest and importance of the upcoming results. A majority of VDMs exhibit low value, but there is a noticeable scatter. A first example is found at 1100 Ma. The Keweenawan lavas from Scandinavia [*Pesonen and Halls*, 1983] and the recent results from the Tudor Gabbro [*Yu and Dunlop*, 2001] in Ontario yield VDM values between 1.36 and 11 $\times$ 10^{22} A m^2. However, the Keweenawan lava flows are characterized by very large error bars and puzzling asymmetrical directions. Such characteristics either represent a very peculiar state of the field or were caused by some artifacts. Many VDMs derived from studies by *Bergh* [1970], *Morimoto et al*. [1997], and *Selkin et al*. [2000] preceding this period indicate low field intensities. This is not the case for the two dikes (9 and 6 $\times$ 10^{22} A m^2) of the Yellowknife greenstone belt from the Slave province, Canada [*Yoshihara and Hamano*, 2000] and the record from South Africa obtained by *McElhinny and Evans* [1968]. The study from Canada meets all selection criteria (except for one negative contact test), and thus there is no reason to suspect the origin of magnetization. Considering the age of their acquisition, the African results remain puzzlingly associated with a very high field intensity, which would certainly justify a new study. Finally, we note that new determinations obtained from 1 to 2.4 Ga Proterozoic dykes [*Macouin et al*., 2003] from the

Superior Province in Canada (not shown in Figure 16) provide additional indications in favor of overall low field intensity during this period. Of particular interest is the study of the Komati formation in South Africa, which provided the oldest paleointensity estimate of the geomagnetic field. Following precursory measurements by *Hale and Dunlop* [1984], *Hale* [1987b] carried out a new study from peridotic and basaltic komatite samples and found a mean field of 5 +/− 2 μT (about 1.5×10^{22} Am^2), which is significantly lower than the average field value for the past 300 Myr. According to *Yoshihara and Hamano* [2000] the values obtained so far would indicate that the dynamo was already as efficient as today at ~2.6 Ga. We would rather conclude that the data appearing in Figure 16 and their time distribution remain too poor to determine any boundary or tendency that would mark some field evolution but that there is also growing evidence in favor of low field intensity during the Precambrian.

5. CONCLUSIONS AND PERSPECTIVES

[100] In less than 20 years our knowledge about field intensity has made a major jump from a composite curve covering the past 12 kyrs to the most recent one covering the past million years, and likely the emergence of a global curve for the past 10 Myr will be next. The most characteristic feature is probably the unstable character of the dipole moment, which depicts a succession of large changes in amplitude with various time constants that can exceed several thousand years. The present decrease of the dipole field by 5% per century does not appear to be atypical with respect to the overall field behavior revealed by the most detailed records. In fact, rapid and large field changes punctuate the recent past (and old) history of the field. This is notably the case for the fast and rapid field recovery following reversals. Apart from the transitional periods, there are strong indications that most intensity lows are associated with excursions or short polarity events. There is no evidence that the nondipole components would diminish similarly but rather that they would become dominant as a consequence of the dipole reduction, thereby suggesting that both sources would not be directly coupled to each other. This reinforces the concept of a continuum between secular variation, excursions, and field reversals. All these features can be seen as manifestations of local nondipole fields in the presence of a strong, weak, or very weak dipole moment. Despite some exciting results, there is no clear demonstration that would favor a relationship between field intensity changes and orbital motions of the Earth's rotation axis. There are also no data indicating a possible relationship between field intensity and external phenomena. Given the overall variance of the field, analyses of the mean field intensity over relatively short time periods are probably not meaningful. In the present state of the data set, there is no robust reason to consider that the mean field intensity changed significantly, say, during the past 5 Myr (the dipole moment oscillated with a mean intensity of between 6 $\times 10^{22}$ A m^{-2} and 8 $\times 10^{22}$ A m^{-2}). Whether or not different values prevailed over short time periods is not excluded, but it requires better sampling to achieve a thorough analysis. The same problem brings up fascinating questions such as the suggestions of a Mesozoic dipole low and a possible link between low (or high) mean field intensity and reversal frequency. The exact pattern of the field intensity changes before and after reversal also remains to be elucidated. There are indications of asymmetrical decay and recovery phases during the pretransitional and posttransitional periods, but their exact duration must be constrained further.

[101] A dominant characteristic emanating from the studies performed during the past 15 years is the permanent quest for improving data quality and hence reliability. In spite of complex processes governing the relationship between field strength and magnetization intensity, it has been quite fascinating to see widely different data sets converging with similar signals. After significant improvements in techniques and measurements, there is now the possibility of using multiple approaches for relative as well as for absolute paleointensity studies.

[102] Relative paleointensity studies should benefit from high-resolution spectral analyses and wavelet analyses of the various components. Future progress will better constrain the factors governing changes in the amplitude of the signals. Duplicate records from nearby sections or cores remains a prerequisite to extracting field variations from other effects (local field contributions, local remagnetization, etc.) and hence to determining the origin of the signals. A more systematic investigation of the magnetic grains involved in the NRM intensity, which can now be achieved more easily with the development of techniques for magnetic separation and transmission as well as reflected light microscopy, is required.

[103] The reliability of determinations of absolute paleointensity has not improved significantly but much care has been taken to include additional checks and criteria before accepting a field determination. The existence of consistent results from different samples within the same flow is not a sufficient condition, and alternatively, some dispersion can be caused by field heterogeneities. The most reliable determinations are evidently those derived from a unique NRM-TRM component involving low as well as high blocking temperatures. This requires the study of a large number of samples and thus a preselection of appropriate samples. Preheating at very high temperatures (with the risk of eliminating a large number of specimens) is an efficient approach. Given the usual success rate of 10–20% it is important to realize that at least 40 samples should be taken in each unit in order to extract eight independent field determinations, which, in any case, is much less time-consuming than

paleointensity experiments applied to 40 specimens. Similarly, detection of multidomain grains using continuous thermal demagnetization should also contribute to the number of determinations, reduce the uncertainties, and hence improve the quality of the records. It remains of note that progress in dating techniques is also expected as age uncertainties are a very limiting factor for studies of absolute paleointensity aimed at describing short-term field variability.

[104] As far as records are concerned, renewal and new development of archeomagnetic studies is essential and very promising. Above all, there is a need for detailed data covering the past millennium in order to explore the stability of features observed in the historical records. Acquisition of high-resolution records covering the past 2 kyr is also a very important target for different areas (western Europe, Asia, and Australia). Field variations occurring with time constants of a few thousand years remain poorly documented. The recent high-resolution profiles from sediments have been shown to be a very promising approach to fill the gap between the timescales inherent to archeomagnetism and the longer time constants inherent to paleomagnetic records. It is clear that sedimentary and many more volcanic determinations are essential to improve the present poor database for at least the past 5 Myr. Both types of records are complementary, and future results should converge with the acquisition of accurately dated volcanic determinations combined with high-resolution sedimentary records. It is hoped that ^{10}Be records will be obtained for the same time interval. Observations of long-term trends in mean field intensity over tens of millions of years can certainly be improved by measurements of new materials such as volcanic glasses or plagioclases, but so far there is no improvement of the success rate, and sometimes there are difficulties ascertaining the fidelity of the results. Additional determinations from terrestrial basalts remain essential as they are for very old time periods. Finally, the early days of the geomagnetic field should be better constrained, thanks to the increasing number of studies on this subject. Such old rocks remain a challenge and could well move us toward different approaches that paleointensity studies may have to face in the future.

[105] **ACKNOWLEDGMENTS.** The author is pleased to thank Vincent Courtillot, Laure Meynadier, Yohan Guyodo, and Julie Carlut for helpful reviews of the manuscript. Yohan Guyodo also helped at several steps of the paper. Phil McFadden and Jim Channell performed comprehensive reviews. Data sets were provided through the courtesy of Juerg Beer, Carlo Laj, Martin Frank, Mioara Mandea, Mélina Macouin, and Jim Stoner. This is IPGP contribution 1796. All data and plots are available on request at valet@ipgp.jussieu.fr.

[106] Louise Kellogg was the Editor responsible for this paper. She thanks Alberto Guadagnini for a cross-disciplinary review and an anonymous technical reviewer.

REFERENCES

Abokadair, A. A., The accuracy of the Thellier's technique for the determination of paleointensities of the Earth's magnetic field, Ph.D. thesis, 346 pp., Univ. of Calif., Santa Cruz, 1977.

Aitken, M. J., A. L. Allsop, G. D. Bussel, and M. B. Winter, Determination of the intensity of the Earth's magnetic field during archeological times: Reliability of the Thellier technique, *Rev. Geophys.*, *26*, 3–12, 1988.

Aitken, M. J., A. L. Adrian, G. D. Bussell, and M. B. Winter, Geomagnetic intensity variation during the last 4000 years, *Phys. Earth Planet. Inter.*, *56*, 49–58, 1989.

Aldahan, A., and G. Possnert, A high resolution ^{10}Be profile from deep-sea sediments covering the last 70 ka: Indication for globally synchronized environmental events, *Quat. Geochronol.*, *17*, 1023–1032, 1998.

Amerigian, C., Measurement of the effect of particle size variation on the detrital remanent magnetization to anhysteretic remanent magnetization ratio in some abyssal sediments, *Earth Planet. Sci. Lett.*, *36*, 434–442, 1977.

Baag, C., C. E. Helsley, S. Z. Xu, and B. R. Lienert, Deflection of paleomagnetic directions due to magnetization of the underlying terrain, *J. Geophys. Res.*, *100*, 10,013–10,027, 1995.

Bard, E., B. Hamelin, R. G. Fairbanks, and A. Zindler, Calibration of the ^{14}C timescale over the past 30,000 years using mass spectrometric U/Th ages from Barbados corals, *Nature*, *345*, 405–410, 1990.

Bard, E., M. Arnold, R. G. Fairbanks, and B. Hamelin, ^{230}Th/^{234}U and ^{14}C ages obtained by mass spectrometry on corals, *Radiocarbon*, *35*, 191–199, 1993.

Bard, E., M. Arnold, B. Hamelin, N. Tisnerat-Laborde, and G. Gabioch, Radiocarbon calibration by means of mass spectrometric ^{230}Th/^{234}U and ^{14}C ages of corals: An updated database including samples from Barbados, Mururoa, and Tahiti, *Radiocarbon*, *40*, 1085–1092, 1998.

Bassinot, F., L. D. Labeyrie, E. Vincent, X. Quidelleur, N. J. Shackleton, and Y. Lancelot, The astronomical theory of climate and the age of the Brunhes-Matuyama reversal, *Earth Planet. Sci. Lett.*, *126*, 91–108, 1994.

Baumgartner, S., J. Beer, J. Masarik, G. Wagner, L. Meynadier, and H.-A. Synal, Geomagnetic modulation of the ^{36}Cl flux in the Grip ice core, Greenland, *Science*, *279*, 1330–1332, 1998.

Beer, J., S. Johnson, G. Bonani, R. C. Finkel, C. C. Langway, H. Oeschger, B. Stauffer, M. Suter, and W. Wölfli, ^{10}Be peaks as time markers in polar ice cores, in *The Last Deglaciation: Absolute and Radiocarbon Chronologies*, edited by E. Bard and W. S. Broecker, pp. 141–153, Springer-Verlag, New York, 1992.

Bergh, H., Paleomagnetism of the Stillwater Complex, Montana, in *Paleogeophysics*, edited by S. K. Runcorn, pp. 143–158, Academic, San Diego, Calif., 1970.

Blanckenburg, F. V., R. K. O'Nions, N. S. Belshaw, A. Gibb, and J. R. Hein, Global distribution of beryllium isotopes in deep ocean water as derived from Fe-Mn crusts, *Earth Planet. Sci. Lett.*, *141*, 213–226, 1996.

Bogue, S. W., and H. A. Paul, Distinctive field behavior following geomagnetic reversals, *Geophys. Res. Lett.*, *20*, 2399–2402, 1993.

Bol'shakov, A. S., and V. V. Shcherbakova, A thermomagnetic criterion for determining the domain structure of ferrimagnetics, *Izv. Acad. Sci. USSR Phys. Solid. Earth, Engl. Transl.*, *15*, 111–117, 1979.

Brachfield, S. A., and S. K. Banerjee, A new high-resolution geomagnetic relative paleointensity record for the North American Holocene: A comparison of sedimentary and absolute intensity data, *J. Geophys. Res.*, *105*, 821–834, 2000.

Buffett, B. A., H. E. Huppert, J. R. Lister, and A. W. Woods, Analytical model for solidification of the Earth's core, *Nature*, *356*, 329–331, 1992.

Bullard, E. C., The magnetic field within the Earth, *Proc. R. Soc. London, Ser. A*, *197*, 433–453, 1949.

Camps, P., G. Ruffet, V. Scherbakov, V. Shcherbakova, M. Prévot, A. Moussine-Pouchkine, Sholpo, L. A. Goguitchaichvili, and B. Asanidzé, Paleomagnetic and geochronological study of a geomagnetic field reversal or excursion recorded in Pliocene volcanic rocks from Georgia (Lesser Caucasus), *Phys. Earth Planet. Inter.*, *96*, 41–59, 1996.

Camps, P., R. S. Coe, and M. Prévot, Transitional geomagnetic impulse hypothesis: Geomagnetic fact or rock-magnetic artifact?, *J. Geophys. Res.*, *104*, 17,747–17,758, 1999.

Cande, S. C., and D. V. Kent, Ultrahigh resolution marine magnetic profiles: A record of continuous paleointensity variations?, *J. Geophys. Res.*, *97*, 15,075–15,083, 1992.

Carlut, J., and V. Courtillot, How complex is the time-averaged geomagnetic field over the past 5 Myr?, *Geophys. J. Int.*, *134*, 527–544, 1998.

Carlut, J., and D. V. Kent, Paleointensity record in zero-age submarine basalt glass: Testing a new dating technique for recent MORBS, *Earth Planet. Sci. Lett.*, *183*, 389–401, 2000.

Carlut, J., and X. Quidelleur, Absolute paleointensities recorded during the Brunhes chron at La Guadeloupe Island, *Phys. Earth Planet. Inter.*, *120*, 255–269, 2000.

Carlut, J., J.-P. Valet, X. Quidelleur, V. Courtillot, T. Kidane, Y. Gallet, and P. Y. Gillot, Paleointensity across the Réunion event in Ethiopia, *Earth Planet. Sci. Lett.*, *170*, 17–34, 1999.

Champion, D. E., M. A. Lanphere, and M. A. Kuntz, Evidence for a new geomagnetic reversal from lava flows in Idaho: Discussion of short polarity reversals in the Brunhes and late Matuyama polarity chrons, *J. Geophys. Res.*, *93*, 11,667–11,680, 1988.

Channell, J. E. T., Geomagnetic paleointensity and directional secular variation at Ocean Drilling Program (ODP) site 984 (Bjorn Drift) since 500 ka: Comparisons with ODP site 983 (Gardar Drift), *J. Geophys. Res.*, *104*, 22,937–22,951, 1999.

Channell, J. E., and H. F. Kleiven, Geomagnetic palaeointensity and astrochronological ages for the Matuyama-Brunhes boundary and the boundaries of the Jaramillo Subchron: Palaeomagnetic and oxygen isotope records from ODP site 983, *Philos. Trans. R. Soc. London, Ser. A*, *358*, 1027–1047, 2000.

Channell, J. E. T., D. A. Hodell, and B. Lehman, Relative geomagnetic paleointensity and ^{18}O at ODP site 983 (Gardar Drift, North Atlantic) since 350 ka, *Earth Planet. Sci. Lett.*, *153*, 103–118, 1997.

Channell, J. E. T., D. A. Hodell, and B. Lehman, Orbital modulation of the Earth's magnetic field intensity, *Nature*, *394*, 464–468, 1998.

Chauvin, A., R. A. Duncan, N. Bonhommmet, and S. Levi, Paleointensity of the Earth's magnetic field and K-Ar dating of the Louchardière volcanic flow (central France): New evidence for the Laschamp excursion, *Geophys. Res. Lett.*, *16*, 1189–1192, 1989.

Chauvin, A., P. Roperch, and R. A. Duncan, Records of geomagnetic reversals from volcanic islands of French Polynesia, *J. Geophys. Res.*, *95*, 2727–2752, 1990.

Chauvin, A., P.-Y. Gillot, and N. Bonhommet, Paleointensity of the Earth's magnetic field recorded by two late Quaternary volcanic sequences at the island of La Réunion (Indian Ocean), *J. Geophys. Res.*, *96*, 1981–2006, 1991.

Chauvin, A., Y. Garcia, P. Lanos, and F. Laubenheimer, Paleointensity of the geomagnetic field recovered on archaeomagnetic sites from France, *Phys. Earth Planet. Inter.*, *120*, 111–136, 2000.

Chave, A. D., and R. C. Denham, Climatic changes, magnetic intensity variations and fluctuations of the eccentricity of the Earth's orbit during the past 2,000,000 years and a mechanism which may be responsible for the relationship—A discussion, *Earth Planet. Sci. Lett.*, *44*, 150–152, 1979.

Cini Castagnoli, G., et al., Evidence for enhanced ^{10}Be deposition in Mediterranean sediments 35 kyr B.P., *Geophys. Res. Lett.*, *22*, 707–710, 1995.

Clement, B. M., and D. G. Martinson, A quantitative comparison of two paleomagnetic records of the Cobb Mountain subchron from North Atlantic deep-sea sediments, *J. Geophys. Res.*, *97*, 1735–1752, 1992.

Coe, R. S., Paleointensities of the Earth's magnetic field determined from Tertiary and Quaternary rocks, *J. Geophys. Res.*, *72*, 3247–3262, 1967.

Coe, R. S., and C. S. Grommé, A comparison of three methods of determining geomagnetic paleointensities, *J. Geomagn. Geoelectr.*, *25*, 415–435, 1973.

Cogné, J. P., J. Francheteau, and V. Courtillot, Large rotation of the Easter microplate as evidenced by oriented paleomagnetic samples from the sea-floor, *Earth Planet. Sci. Lett.*, *136*, 213–222, 1995.

Constable, C. G., and R. L. Parker, Statistics of the geomagnetic secular variation for the past 5 m.y., *J. Geophys. Res.*, *93*, 11,569–11,581, 1988.

Constable, C. G., L. Tauxe, and R. L. Parker, Analysis of 11 Myr of geomagnetic intensity variation, *J. Geophys. Res.*, *103*, 17,735–17,748, 1998.

Constable, C. G., C. Johnson, and S. Lund, Global geomagnetic field models for the past 3000 years: Transient or permanent flux lobes?, *Philos. Trans. R. Soc. London, Ser. A*, *358*, 991–1008, 2000.

Cottrell, R. D., and J. A. Tarduno, Geomagnetic paleointensity derived from single plagioclase crystals, *Earth Planet. Sci. Lett.*, *169*, 1–5, 1999.

Cottrell, R. D., and J. A. Tarduno, In search of high-fidelity geomagnetic paleointensities: A comparison of single plagioclase crystals and whole rock Thellier-Thellier analyses, *J. Geophys. Res.*, *105*, 23,579–23,594, 2000.

Coupland, D. H., and R. Van der Voo, Long-term nondipole components in the geomagnetic field during the last 130 Ma, *J. Geophys. Res.*, *85*, 3529–3548, 1980.

Creer, K. M., N. Thouveny, and I. Blunk, Climatic and geomagnetic influences on the Lac du Bouchet paleomagnetic record through the last 110,000 years, *Phys. Earth Planet. Inter.*, *64*, 314–341, 1990.

Daly, L., and M. LeGoff, An updated and homogeneous world secular variation data base, 1, Smoothing of the archeomagnetic results, *Phys. Earth Planet. Inter.*, *93*, 159–190, 1996.

DeMenocal, P. B., W. F. Ruddiman, and D. V. Kent, Depth of post-depositional remanence acquisition in deep-sea sediments: A case study of the Brunhes-Matuayma reversals and oxygen isotope stage 19.1, *Earth Planet. Sci. Lett.*, *99*, 1–13, 1990.

Doell, R. R., and P. J. Smith, On the use of magnetic cleaning in paleointensity studies, *J. Geomagn. Geoelectr.*, *21*, 579–594, 1969.

Dormy, E., J.-P. Valet, and V. Courtillot, Numerical models of the geodynamo and observational constraints, *Geochem. Geophys. Geosyst.*, *1*, 2000GC000062, 2000.

Dunlop, D. J., and Ö. Özdemir, Thermal demagnetization of VRM and pTRM of single domain magnetite: No evidence for anomalously high unblocking temperatures, *Geophys. Res. Lett.*, *20*, 1939–1942, 1993.

Dunlop, D. J., and Ö. Özdemir, Rock Magnetism, 573 pp., Cambridge Univ. Press, New York, 1997.

Dunlop, D. J., and Ö. Özdemir, Effect of grain size and domain state on thermal demagnetization tails, *Geophys. Res. Lett.*, *27*, 1311–1314, 2000.

Dunlop, D. J., and S. Xu, Theory of partial thermoremanent magnetization in multidomain grains, 1, Repeated identical barriers to wall motion single microcoercivity, *J. Geophys. Res.*, *99*, 9005–9023, 1994.

Elmaleh, A., J. P. Valet, and E. Herrero-Bervera, A map of the Pacific geomagnetic anomaly during the Brunhes chron, *Earth Planet. Sci. Lett.*, *193*, 315–332, 10.1016/S0012-821X(01)00518-0, 2001.

Elsasser, W. M., Causes of motions in the Earth's core, *Eos Trans. AGU*, *31*, 454–462, 1950.

Elsasser, W., E. P. Ney, and J. R. Winckler, Cosmic ray intensity and geomagnetism, *Nature*, *178*, 1226–1227, 1956.

Frank, M., Comparison of cosmogenic radionuclide production and geomagnetic field intensity over the last 200,000 years, *Philos. Trans. R. Soc. London, Ser. A*, *358*, 1089–1107, 2000.

Frank, M., B. Schwarz, S. Baumann, P. W. Kubik, M. Suter, and A. Mangini, A 200 kyr record of cosmogenic radionuclide production rate and geomagnetic field intensity from ^{10}Be in globally stacked deep-sea sediments, *Earth Planet Sci. Lett.*, *149*, 121–129, 1997.

Gallet, Y., and G. Hulot, Stationary and nonstationary behavior within the geomagnetic polarity time scale, *Geophys. Res. Lett.*, *24*, 1875–1878, 1997.

Gallet, Y., J. Gee, L. Tauxe, and J. A. Tarduno, Paleomagnetic analyses of short normal polarity magnetic anomalies in the Matuyama chron, *Proc. Ocean Drill. Program Sci. Results*, *130*, 547–559, 1993.

Gee, J., and D. V. Kent, Variations in layer 2A: Thickness and the origin of the central magnetic anomaly high, *Geophys. Res. Lett.*, *21*, 297–300, 1994.

Gee, J., D. A. Schneider, and D. V. Kent, Marine magnetic anomalies as recorders of geomagnetic intensity variations, *Earth Planet. Sci. Lett.*, *144*, 327–335, 1996.

Gee, J., S. C. Cande, J. A. Hildebrand, J. A. Donnelly, and R. L. Paraker, Geomagnetic intensity variations over the past 780 kyr obtained from near-seafloor anomalies, *Nature*, *408*, 827–832, 2000.

Genevey, A., and Y. Gallet, Intensity of the geomagnetic field in western Europe over the past 2000 years: New data from ancient French potteries, *J. Geophys. Res.*, *107*(B11), 2285, doi:10.1029/2001JB000701, 2002.

Gillot, P.-Y., J. Labeyrie, C. Laj, G. Valladas, G. Guerin, G. Poupeau, and G. Delibrias, Age of the Laschamp paleomagnetic excursion revisited, *Earth Planet. Sci. Lett.*, *42*, 444–450, 1979.

Goguitchaichvili, A., P. Camps, and J. Urrutia-Fucaugauchi, On the features of the geodynamo following reversals or excursions: By absolute geomagnetic paleointensity data, *Phys. Earth Planet. Lett.*, *124*, 81–93, 2001.

Grootes, P. M., and M. Stuiver, Oxygen 18/16 variability in Greenland snow and ice with 10^{-3}- to 10^{5}-year time resolution, *J. Geophys. Res.*, *102*, 26,455–26,470, 1997.

Gubbins, D., The distinction between geomagnetic excursions and reversals, *Geophys. J. Int.*, *137*, F1–F3, 1999.

Gubbins, D., and P. Kelly, Persistent patterns in the geomagnetic field during the last 2.5 Myr, *Nature*, *365*, 829–832, 1993.

Guyodo, Y., and J. P. Valet, Relative variations in geomagnetic intensity from sedimentary records: The past 200 thousand years, *Earth Planet. Sci. Lett.*, *143*, 23–26, 1996.

Guyodo, Y., and J. P. Valet, Integration of volcanic and sedimentary paleointensity data: Constraints imposed by irregular eruption rates, *Geophys. Res. Lett.*, *26*, 3669–3672, 1999a.

Guyodo, Y., and J. P. Valet, Global changes in geomagnetic intensity during the past 800 thousand years, *Nature*, *399*, 249–252, 1999b.

Guyodo, Y., C. Richter, and J.-P. Valet, Paleointensity record from Pleistocene sediments off the California Margin, *J. Geophys. Res.*, *104*, 22,953–22,965, 1999.

Guyodo, Y., P. Gaillot, and J. E. T. Channell, Wavelet analysis of relative geomagnetic paleointensity at ODP site 983, *Earth Planet. Sci. Lett.*, *184*, 109–123, 2000.

Guyodo, Y., G. D. Acton, S. Brachfield, and J. E. T. Channell, A sedimentary paleomagnetic record of the Matuyama chron from the western Antarctic Margin, *Earth Planet. Sci. Lett.*, *191*, 61–74, 2001.

Haag, M., J. R. Dunn, and M. Fuller, A new quality check for absolute palaeointensities of the Earth magnetic field, *Geophys. Res. Lett.*, *22*, 3549–3552, 1995.

Hale, C. J., Palaeomagnetic data suggest link between the Archaean-Proterozoic boundary and inner-core nucleation, *Nature*, *329*, 233–237, 1987a.

Hale, C. J., The intensity of the geomagnetic field at 3.5 Ga: Paleointensity results from the Komati formation, Barberton Mountain Land, South Africa, *Earth Planet. Sci. Lett.*, *86*, 354–364, 1987b.

Hale, C. J., and D. J. Dunlop, Evidence for an early Archean geomagnetic field: A paleomagnetic study of the Komati formation, Barbetone greenstone belt, South Africa, *Geophys. Res. Lett.*, *11*, 97–100, 1984.

Halgedahl, S. L., Experiments to investigate the origin of anomalously elevated unblocking temperatures, *J. Geophys. Res.*, *98*, 22,443–22,460, 1993.

Henken-Mellies, W. U., J. Beer, F. Heller, K. J. Hs, C. Shen, G. Bonani, H. J. Hofman, M. Suter, and W. Wölfli, ^{10}Be and ^{9}Be in South Atlantic DSDP site 519: Relation to geomagnetic reversals and to sediment composition, *Earth Planet. Sci. Lett.*, *98*, 267–276, 1990.

Hill, M., and J. Shaw, Magnetic field intensity study of the 1960 Kilauea lava flow, Hawaii, using the microwave paleointensity technique, *Geophys. J. Int.*, *142*, 487–504, 2000.

Hoffman, K. A., V. L. Constantine, and D. L. Morse, Determination of absolute paleointensity using a multi-specimen procedure, *Nature*, *339*, 295–297, 1989.

Hongre, L., G. Hulot, and A. Khokhlov, An analysis of the geomagnetic field over the past 2000 years, *Phys. Earth Planet. Inter.*, *106*, 311–335, 1998.

Hulot, H. G., and J.-L. Le Mouël, A statistical approach to the Earth's main magnetic field, *Phys. Earth Planet. Inter.*, *82*, 167–183, 1994.

Imbrie, J., J. D. Hays, D. G. Martinson, A. McIntyre, A. C. Mix, J. J. Morley, N. G. Pisias, W. L. Prell, and N. J. Shackleton, The orbital theory of Pleistocene climate: Support from a revised chronology of the marine ^{18}O record, in *Milankovitch and Climate: Understanding the Response to Astronomical Forcing, Part 1*, edited by A. L. Berger et al., pp. 269–305, D. Reidel, Norwell, Mass., 1984.

Jacobs, J. A., Variations in the intensity of the Earth's magnetic field, *Surv. Geophys.*, *19*, 139–187, 1998.

Johnson, C. L., and C. G. Constable, The time-averaged geomagnetic field: Global and regional biases for 0–5 Ma, *Geophys. J. Int.*, *131*, 643–666, 1997.

Johnson, C. L., and C. G. Constable, Persistently anomalous Pacific geomagnetic fields, *Geophys. Res. Lett.*, *25*, 1011–1014, 1998.

Johnson, H. P., H. Kinoshita, and R. T. Merrill, Rock magnetism and paleomagnetism of some north Pacific deep-sea sediment core, *Geol. Soc. Am. Bull.*, *86*, 412–420, 1975.

Juarez, M. C., L. Tauxe, J. S. Gee, and T. Pick, The intensity of the Earth's magnetic field over the past 160 million years, *Nature*, *394*, 878–881, 1998.

Katari, K., and J. Bloxham, Effects of sediments aggregate size on DRM intensity: A new theory, *Earth Planet. Sci. Lett.*, *186*, 113–122, 2001.

Kelly, P., and D. Gubbins, The geomagnetic field over the past 5 million years, *Geophys. J. Int.*, *128*, 1–16, 1997.

Kent, D. V., Apparent correlation of paleomagnetic intensity and climatic records in deep-sea sediments, *Nature*, *299*, 538–539, 1982.

Kent, D. V., and N. D. Opdyke, Paleomagnetic field intensity recorded in a Brunhes epoch deep-sea sediment core, *Nature*, *266*, 156–159, 1977.

King, J. W., S. K. Banerjee, and J. Marvin, A new rock-magnetic approach to selecting sediments for geomagnetic paleointensity studies: Application to paleointensity for the last 4000 years, *J. Geophys. Res.*, *88*, 5911–5921, 1983.

Kissel, C., C. Laj, L. Labeyrie, T. Dodken, A. Voelker, and D. Blamart, Rapid climatic variations during marine isotopic stage 3: Magnetic analysis of sediments from Nordic seas and North Atlantic, *Earth Planet. Sci. Lett.*, *171*, 489–502, 1999.

Kitagawa, H., and J. van der Plicht, Atmospheric radiocarbon calibration to 45,000 yr B.P.: Late glacial fluctuations and cosmogenic isotope production, *Science*, *279*, 1187–1190, 1998.

Kok, Y., Climatic influence in NRM and ^{10}Be-derived geomagnetic paleointensity data, *Earth Planet. Sci. Lett.*, *166*, 105–119, 1999.

Kok, Y. S., and L. Tauxe, Saw-tooth pattern of relative paleointensity records and cumulative viscous remanence, *Earth Planet. Sci. Lett.*, *137*, 95–99, 1996a.

Kok, Y. S., and L. Tauxe, Saw-toothed pattern of sedimentary paleointensity records explained by cumulative viscous remanence, *Earth Planet. Sci. Lett.*, *144*, E9–E14, 1996b.

Kono, M., Reliability of paleointensity methods using alternating field demagnetization and anhysteretic remanence, *Geophys. J. R. Astron. Soc.*, *54*, 241–261, 1978.

Kono, M., and N. Ueno, Paleointensity determination by a modified Thellier method, *Phys. Earth Planet. Inter.*, *13*, 305–314, 1977.

Kosterov, A. A., and M. Prévot, Possible mechanisms causing failure of paleointensity experiments in some basalts, *Geophys. J. Int.*, *134*, 554–572, 1998.

Kovacheva, M., and M. Karnachev, Revised archeointensities data from Bulgaria, *J. Geomagn. Geoelectr.*, *38*, 1297–1310, 1986.

Kovacheva, M., J. Pares, N. Jordonova, and V. Karloukovski, A new contribution to the archeomagnetic study of a Roman pottery kiln from Calahorra (Spain), *Geophys. J. Int.*, *123*, 931–936, 1995.

Labrosse, S., J. P. Poirier, and J. L. LeMouel, On cooling of the Earth core, *Phys. Earth Planet. Inter.*, *99*, 1–17, 1997.

Labrosse, S., J.-P. Poirier, J.-L. Le Mouel, The age of the inner core, *Earth Planet. Sci. Lett.*, *190*, 111–123, 2001.

Laj, C., and C. Kissel, Geomagnetic field intensity at Hawaii for the last 420 kyr from the Hawaii Scientific Drilling Project core, Big Island, Hawaii, *J. Geophys. Res.*, *104*, 15,317–15,338, 1999.

Laj, C., C. Kissel, F. Garnier, and E. Herrero-Bervera, Relative geomagnetic field intensity and reversals for the last 1.8 Ma from a central equatorial Pacific core, *Geophys. Res. Lett.*, *23*, 3393–3396, 1996.

Laj, C., C. Kissel, M. Mazaud, J. E. T. Channell, and J. Beer, North Atlantic paleointensity stack since 75 ka (NAPIS-75) and the duration of the Laschamp event, *Philos. Trans. R. Soc. London, Ser. A*, *358*, 1009–1025, 2000a.

Laj, C., N. Szeremata, C. Kissel, and H. Guillou, Geomagnetic paleointensities at Hawaii between 3.9 and 2.1 Ma: Preliminary results, *Earth Planet. Sci. Lett.*, *179*, 191–204, 2000b.

Laj, C., C. Kissel, V. Scao, J. Beer, D. M. Thomas, H. Guillou, R. Muscheler, and G. Wagner, Geomagnetic intensity and inclination variations of Hawaii for the period 98 kyr from core SOH-4 (Big Island): A new study and a comparison with existing contemporary data, *Phys. Earth Planet. Inter.*, *129*, 205–243, 2002.

Lal, D., Theoretical expected variations in the terrestrial cosmic-ray production rates of isotopes, in *Solar-Terrestrial Relationships and the Earth Environment in the Last Millennia*, edited by X. C. V. Corso, Rep. XCV, pp. 216–233, Soc. Ital. di Fis., Bologna, Italy, 1988.

Langereis, C. G., M. J. Dekkers, G. J. DeLange, M. Paterne, and P. J. M. Van Santvoort, Magnetostratigraphy and astronomical calibration of the last 1.1 Myr from an eastern Mediterranean piston core and dating of short events in the Brunhes. *Geophys. J. Int.*, *129*, 75–94, 1997.

Lao, Y., R. F. Anderson, W. S. Broecker, S. E. Trumbore, H. J. Hofman, and W. Wolfi, Increased production of cosmogenic ^{10}Be during the last glacial maximum, *Nature*, *357*, 576–578, 1992.

Lehman, B., C. Laj, C. Kissel, A. Mazaud, M. Paterne, and L. Labeyrie, Relative changes of the geomagnetic field intensity during the last 280 kyear from piston cores in the Acores area, *Phys. Earth Planet. Inter.*, *93*, 269–284, 1996.

Leonhardt, R., F. Hufenbecher, F. Heider, and H. C. Soffel, High absolute paleointensity during a mid Miocene excursion of the Earth's magnetic field, *Earth Planet. Sci. Lett.*, *184*, 141–154, 2000.

Levi, S., The effect of particle size on paleointensity determination of the geomagnetic field, *Phys. Earth Planet. Inter.*, *13*, 245–259, 1977.

Levi, S., and S. K. Banerjee, On the possibility of obtaining relative paleointensities from lake sediments, *Earth. Planet. Sci. Lett.*, *29*, 219–226, 1976.

Liddicoat, J. C., Mono Lake excursion in Mono basin, California, and at Carlson Sink and Pyramid lake, Nevada, *Geophys. J. Int.*, *108*, 442–452, 1992.

Lin, J. L., K. L. Verosub, and P. A. Roberts, Decay of the virtual dipole moment during polarity transitions and geomagnetic excursions, *Geophys. Res.Lett.*, *21*, 525–528, 1994.

Lund, S., A comparison of Holocene paleomagnetic secular variation records from North America, *J. Geophys. Res.*, *101*, 8007–8024, 1996.

Lund, S., Initial paleomagnetic results from ODP leg 172—Evidence for at least 14 geomagnetic excursions during the Brunhes epoch (abstract), *Eos Trans. AGU*, *78*(46), Fall Meet. Suppl., F181, 1997.

Lund, S., and M. Schwartz, Environmental factors affecting geomagnetic field paleointensity estimates from sediments, in *Quaternary Environmental Magnetism*, edited by B. Maher and R. Thompson, Cambridge Univ. Press, New York, 1999.

Lund, S. P., G. Acton, B. Clement, M. Hastedt, M. Okada, T. Williams, and the ODP Leg 172 Scientific Party, Geomagnetic field excursions occurred often during the last million years, *Eos Trans. AGU*, *79*(14), 178–179, 1998.

Macouin, M., J.-P. Valet, J. Besse, K. Buchan, R. Ernst, M. LeGoff, and U. Schärer, Low paleointensities recorded in 1 to 2.4 Ga old Proterozoic dykes, Superior Province, Canada, *Earth Planet. Sci. Lett.*, in press, 2003.

Malkus, W. V. R., Precession of the Earth as the cause of geomagnetism, *Science*, *160*, 259–264, 1968.

Mankinen, E. A., and D. E. Champion, Broad trends in geomagnetic paleointensity on Hawaii during holocene time, *J. Geophys. Res.*, *98*, 7959–7976, 1993.

Martinson, D. G., N. G. Pisias, J. D. Hays, J. Imbrie, T. C. Moore Jr., and N. J. Shackleton, Age dating and the orbital theory of the ice ages: Development of a high resolution 0 to 300,000 year chronostratigraphy, *Quat. Res.*, *27*, 1–29, 1987.

Mazaud, A., Saw-tooth variation in magnetic intensity profiles and delayed acquisition of magnetization in deep sea cores, *Earth Planet. Sci. Lett.*, *139*, 379–386, 1996.

McClelland, E., Theory of CRM acquired by grain growth, and its implications for TRM acquisition and paleointensity determinations in igneous rocks, *Geophys. J. Int.*, *126*, 271–280, 1996.

McClelland, E., and J. C. Briden, An improved methodology for Thellier-type paleointensity determination in igneous rocks and its usefulness for verifying primary thermoremanence, *J. Geophys. Res.*, *101*, 21,995–22,013, 1996.

McClelland, E., and V. P. Shcherbakov, Metastability of domain state in multidomain magnetite: Consequences for remanence acquisition, *J. Geophys. Res.*, *100*, 3841–3857, 1995.

McClelland, E., and N. Sugiura, A kinematic model of pTRM acquisition in multidomain magnetite, *Phys. Earth Planet. Inter.*, *46*, 9–23, 1987.

McClelland, E., A. R. Muxworthy, and R. M. Thomas, Magnetic properties of the stable fraction of remanence in multidomain (MD) magnetite grains: Single-domain or MD?, *Geophys. Res. Lett.*, *23*, 2831–2834, 1996.

McElhinny, M. W., and M. E. Evans, An investigation of the strength of the geomagnetic field in the early Precambrian, *Phys. Earth Planet. Inter.*, *1*, 485–497, 1968.

McElhinny, M. W., and W. E. Senanayake, Variations in the geomagnetic dipole, 1, The past 50,000 years, *J. Geomagn. Geoelectr.*, *34*, 39–51, 1982.

McFadden, P. L., and R. T. Merrill, Inhibition and geomagnetic field reversals, *J. Geophys. Res.*, *98*, 6189–6199, 1993.

McFadden, P. L., and R. T. Merrill, Sawtooth paleointensity and reversals of the geomagnetic field, *Phys. Earth Planet. Inter.*, *103*, 247–252, 1997.

Mejia, V., N. D. Opdyke, and M. R. Perfit, Paleomagnetic field intensity recorded in submarine basaltic glass from the East Pacific Rise, the last 69 ka, *Geophys. Res. Lett.*, *23*, 475–478, 1996.

Merrill, R. T., and P. L. McFadden, Geomagnetic stability: Reversal events and excursions, *Earth Planet. Sci. Lett.*, *121*, 57–69, 1994.

Merrill, R. T., and M. W. McElhinny, Anomalies in the time-averaged field and their implications for the lower mantle, *Rev. Geophys.*, *15*, 309–323, 1977.

Merrill, R. T., M. W. McElhinny, and P. L. McFadden, *The Magnetic Field of the Earth*, *Int. Geophys. Ser.*, vol. *63*, 531 pp., Academic, San Diego, Calif., 1996.

Meynadier, L., and J.-P. Valet, Relative paleointensity in sediments: What should we believe?, paper presented at XXI General Assembly, Int. Union of Geod. and Geophys., Boulder, Colo., 1995.

Meynadier, L., and J.-P. Valet, Post-depositional realignment of magnetic grains and asymmetrical saw-toothed pattern of magnetization intensity, *Earth Planet. Sci. Lett.*, *140*, 123–132, 1996.

Meynadier, L., and J.-P. Valet, Reply to comment by Y. S. Kok and L. Tauxe on "Saw-toothed variations of relative paleointensity and cumulative viscous remanence: Testing the records and the model," *J. Geophys. Res.*, *105*, 16,613, 2000.

Meynadier, L., J.-P. Valet, R. Weeks, N. J. Shackleton, and V. Lee Hagee, Relative geomagnetic intensity of the field during the last 140 ka, *Earth. Planet. Sci. Lett.*, *114*, 39–57, 1992.

Meynadier, L., J.-P. Valet, F. Bassinot, N. Shackleton, and Y Guyodo, Asymmetrical saw-tooth pattern of the geomagnetic field intensity from equatorial sediments in the Pacific and the Indian Oceans, *Earth Planet. Sci. Lett.*, *126*, 109–127, 1994.

Meynadier, L., J.-P. Valet, and N. J. Shackleton, Relative geomagnetic intensity during the last 4 m.y. from the equatorial Pacific, *Proc. Ocean Drill. Program Sci. Results*, *138*, 779–795, 1995.

Meynadier, L., J.-P. Valet, Y. Guyodo, and C. Richter, Saw-toothed variations of relative paleointensity and cumulative viscous remanence: Testing the records and the model, *J. Geophys. Res.*, *103*, 7095–7105, 1998.

Morimoto, C., Y. Otofuji, M. Miki, H. Tanaka, and T. Itaya, Preliminary palaeomagnetic results of an Archean dolerite dyke of West Greenland: Geomagnetic field intensity at 2.8 Ga, *Geophys. J. Int.*, *128*, 585–593, 1997.

Nachasova, L. E., and K. S. Burakov, The geomagnetic field intensity in Central Asia from 6000 to 3000 B.C., *Izv. Russ. Acad. Sci. Phys. Solid Earth*, Engl. Transl., *36*(5), 358–363, 2000.

Nagata, T., Y. Arai, and K. Momose, Secular variation of the geomagnetic total force during the last 5000 years, *J. Geophys. Res.*, *68*, 5277–5281, 1963.

Needham, J., *Science and Civilisation in China*, vol. 4, *Physics and Physical Technology, Part 1, Physics*, Cambridge Univ. Press, New York, 1962.

Nowaczyk, N. R., T. W. Frederichs, A. Eisenhauer, and G. Gard, Magnetostratigraphic data from late Quaternary sediments from the Yermak Plateau, Arctic Ocean: Evidence for four geomagnetic polarity events within the last 170 ka of the Brunhes chron, *Geophys. J. Int.*, *117*, 453–471, 1994.

Pal, P. C., and P. H. Roberts, Long-term polarity stability and strength of the geomagnetic dipole, *Nature*, *331*, 702–705, 1988.

Peck, J., J. W. King, S. M. Colman, and V. A. Kravchinsky, An 84-kyr paleomagnetic record from the sediments of Lake Baikal, Siberia, *J. Geophys. Res.*, *101*, 11,365–11,386, 1996.

Perrin, M., and V. Shcherbakov, Paleointensity of the Earth's magnetic field for the last 400 Ma: Evidence for a dipole structure during the Mesozoic low, *J. Geomagn. Geoelectr.*, *49*, 601–614, 1997.

Perrin, M., and V. Shcherbakov, Paleointensity database updated, *Eos Trans. AGU*, *79*(16), 198, 1998.

Perrin, M., M. Prévot, and E. A. Mankinen, Low intensity of the geomagnetic field in early Jurassic time, *J. Geophys. Res.*, *96*, 14,197–14,210, 1991.

Pesonen, L. J., and H. C. Halls, Geomagnetic field intensity and reversal asymmetry in late Precambrian Keweenawan rocks, *Geophys. J. R. Astron. Soc.*, *73*, 241–270, 1983.

Pesonen, L. J., M. A. H. Leino, and H. Nevanlinna, Archeomagnetic intensity in Finland during the last 6400 years: Evidence for a latitude-dependent non diple field at AD 500, *J. Geomagn. Geoelectr.*, *47*, 19–40, 1995.

Pick, T., and L. Tauxe, Geomagnetic paleointensities during the Cretaceous normal superchron measured using submarine basaltic glass, *Nature*, *36*, 238–242, 1993a.

Pick, T., and L. Tauxe, Holocene paleointensities: Thellier experiments on submarine basaltic glass from the East Pacific Rise, *J. Geophys. Res.*, *98*, 17,949–17,964, 1993b.

Pouliquen, G., Y. Gallet, J. Dyment, P. Patriat, and C. Tamura, A geomagnetic record over the last 4 million years from deep-tow magnetic profiles across the Central Indian Ridge, *J. Geophys. Res.*, *106*, 10,941–10,960, 2001.

Prévot, M., E. A. Mankinen, R. S. Coe, and C. S. Grommé, The Steens Mountain (Oregon) geomagnetic polarity transition, 2, Field intensity variations and discussion of reversal models, *J. Geophys. Res.*, *90*, 10,417–10,448, 1985.

Prévot, M., M. E. Derder, M. Mc Williams, and J. Thompson, Intensity of the Earth's magnetic field: Evidence for a Mesozoic dipole low, *Earth Planet. Sci. Lett.*, *97*, 129, 1990.

Quidelleur, X., and V. Courtillot, On low degree spherical harmonic models of paleosecular variation, *Phys. Earth Inter.*, *95*, 55–77, 1996.

Rais, A., C. Laj, J. Surmont, P. Y. Gillot, and H. Guillou, Geomagnetic field intensity between 70,000 and 130,000 years B.P. from a volcanic sequence, Indian Ocean, *Earth Planet. Sci. Lett.*, *140*, 173–189, 1996.

Raisbeck, G. M., F. Yiou, D. Bourles, and D. V. Kent, Evidence for an increase in ^{10}Be during a geomagnetic reversal, *Nature*, *315*, 315–317, 1985.

Raisbeck, G. M., F. M. Yiou, D. Bourles, C. Lorius, J. Jouzel, and N. J. Barkov, Evidence for two intervals of enhanced ^{10}Be deposition in Antarctic ice during the last period, *Nature*, *326*, 273–276, 1987.

Raisbeck, G. M., F. Yiou, and S. Z. Zhou, Paleointensity puzzle, *Nature*, *371*, 234–238, 1994.

Ravello, A. C., and N. J. Shackleton, Evidence for surface-water circulation changes at site 851 in the eastern tropical Pacific Ocean, *Proc. Ocean Drill. Program Sci. Results*, *138*, 503–516, 1995.

Riisager, P., and N. Abrahamsen, Palaeointensity of west Greenland Paleocene basalts: Asymmetric intensity around the C27n-C26r transition, *Phys. Earth Planet. Inter.*, *118*, 53–64, 2000.

Riisager, P., and J. Riisager, Detecting multidomain magnetic grains in Thellier paleointensity experiments, *Phys. Earth Planet. Inter.*, *125*, 111–117, 2001.

Roberts, A. P., and J. C. Lewin-Harris, Marine magnetic anomalies: Evidence that 'tiny' wiggles represent short-period geomagnetic polarity intervals, *Earth Planet. Sci. Lett.*, *183*, 375–388, 2000.

Roberts, A. P., B. Lehman, R. J. Weeks, K. L. Verosub, and C. Laj, Relative paleointensity of the geomagnetic field over the last 200,000 years from ODP sites 883 and 884, North Pacific Ocean, *Earth Planet. Sci. Lett.*, *152*, 11–23, 1997.

Robinson, C., G. M. Raisbeck, F. Yiou, B. Lehman, and C. Laj, The relationship between ^{10}Be and geomagnetic field strength records in central North Atlantic sediments during the last 80 ka, *Earth Planet. Sci. Lett.*, *136*, 551–557, 1995.

Rochester, M. G., J. A. Jacobs, D. E. Smylie, and K. F. Chong, Can precession power the geomagnetic dynamo?, *Geophys. J. R. Astron. Soc.*, *43*, 661–678, 1975.

Rolph, T. C., and J. Shaw, A new method of paleofield magnitude correction for thermally altered samples and its application to Lower Carboniferous lavas, *Geophys. J. Int.*, *80*, 773–781, 1985.

Roperch, P., N. Bonhommet, and S. Levi, Paleointensity of the Earth's magnetic field during the Laschamp event and its geomagnetic implications, *Earth Planet. Sci. Lett.*, *88*, 209–219, 1988.

Schneider, D. A., An estimate of late Pleistocene geomagnetic intensiy from Sulu Sea sediments during the last 80 ka, *Earth Planet. Sci. Lett.*, *120*, 301–310, 1993.

Schneider, D. A., and D. V. Kent, The time-averaged paleomagnetic field, *Rev. Geophys.*, *28*, 71–96, 1990.

Schneider, D. A., and G. A. Mello, A high-resolution marine sedimentary record of geomagnetic intensity during the Brunhes chron, *Earth Planet. Sci. Lett.*, *144*, 297–314, 1996.

Schnepp, E., Geomagnetic paleointensities derived from volcanic rocks of the Quaternary East Eiffel volcanic field, Germany, *Phys. Earth Planet. Inter.*, *94*, 23–41, 1996.

Schwartz, M., S. P. Lund, and T. C. Johnson, Geomagnetic field intensity from 71 to 12 ka as recorded in deep-sea sediments of the Blake Outer Ridge, North Atlantic Ocean, *J. Geophys. Res.*, *103*, 30,407–30,416, 1998.

Selkin, P. A., and L. Tauxe, Long-term variations in paleointensity, *Philos. Trans. R. Soc. London, Ser. A, Philos. Trans. Math. Phys. Eng. Sci.*, *358*, 1065–1088, 2000.

Selkin, P. A., J. S. Gee, L. Tauxe, W. P. Meurer, and A. J. Newell, The effect of remanence anisotropy on paleointensity estimates: A case study from the Archean Stillwater Complex, *Earth Planet Sci. Lett.*, *183*, 403–416, 2000.

Shaw, J., A new method of determining the magnitude of the paleomagnetic field: Application to five historic lavas and five archeological samples, *Geophys. J. R. Astron. Soc. 39*, 133–141, 1974.

Shaw, J., D. Walton, S. Yang, T. C. Rolph, and J. A. Share, Microwave archeointensities from Peruvian ceramics, *Geophys. J. Int.*, *124*, 241–244, 1996.

Shaw, J., S. Yang, T. C. Rolph, and F. Y. Sun, A comparison of archeointensity results from Chinese ceramics using microwave and conventional Thellier's and Shaw's methods, *Geophys. J. Int.*, *136*, 714–718, 1999.

Shcherbakova, V. V., and V. P. Shcherbakov, Properties of partial thermoremanent magnetization in pseudosingle domain and multidomain magnetite grains, *J. Geophys. Res.*, *105*, 767–781, 2000.

Sherwood, G. J., Evaluation of a multi-specimen approach to paleointensity determination, *J. Geomagn. Geoelectr.*, *43*, 341–349, 1991.

Smith, P. J., On the suitability of igneous rocks for ancient geomagnetic field intensity determination, *Earth Planet. Sci. Lett.*, *2*, 99–105, 1967.

Sonnett, C. P., C. E. Morfill, and J. R. Jokipii, Interstellar shock waves and ^{10}Be from ice cores, *Nature*, *330*, 458–460, 1987.

Stacey, F. D., The coupling of the core to the precession of the Earth, *Geophys. J. R. Astron. Soc.*, *33*, 47–55, 1973.

Stevenson, D. J., T. Spohn, and G. Schubert, Magnetism and thermal evolution of the terrestrial planets, *Icarus*, *54*, 466–489, 1983.

Stoner, J. S., J. E. T. Channell, and C. Hillaire-Marcel, and J. C. Mareschal, Late Pleistocene relative geomagnetic field paleointensity record from the deep Labrador Sea: Regional and global correlations, *Earth Planet. Sci. Lett.*, *134*, 237–252, 1995.

Stoner, J. S., J. E. T. Channell, and C. Hillaire-Marcel, A 200 ka chronostratigraphy for the Labrador Sea: Indirect correlation of the sediment record to SPECMAP, *Earth. Planet. Sci. Lett.*, *159*, 165–181, 1998.

Stoner, J. S., J. E. T. Channell, C. Hillaire-Marcel, and C. Kissel, Geomagnetic paleointensity and environmental record from Labrador Sea core MD95-2024: Global marine and ice core chronostratigraphy for the last 110 kyr, *Earth Planet. Sci. Lett.*, *183*, 161–177, 2000.

Sugiura, N., ARM, TRM and magnetic interactions: Concentration dependence, *Earth Planet. Sci. Lett.*, *42*, 451–455, 1979.

Tanaka, H., and M. Kono, Preliminary results and reliability of paleointensity studies on historical and ^{14}C dated Hawaiian lava flows, *J. Geomagn. Geoelectr.*, *43*, 375–388, 1991.

Tanaka, H., and M. Kono, Paleointensities from a Cretaceous basalt platform in Inner Mongolia, northeastern China, *Phys. Earth Planet. Inter.*, *133*, 147–157, 2002.

Tanaka, H., J. D. E. Athanassopoulos, J. R. Dunn, and M. Fuller, Paleointensity determinations with measurements at high temperature, *J. Geomagn. Geoelectr.*, *47*, 103–113, 1995a.

Tanaka, H., M. Kono, and H. Uchimura, Some global features of paleointensity in geological time, *Geophys. J. Int.*, *120*, 97–102, 1995b.

Tarduno, J. A., R. D. Cottrell, and A. V. Smirnov, High geomagnetic intensity during the mid-Cretaceous from Thellier analyses of single plagioclase crystals, *Science*, *291*, 1779–1782, 2001.

Tauxe, L., Sedimentary records of relative paleointensity of the geomagnetic field: Theory and practice, *Rev. Geophys.*, *31*, 319–354, 1993.

Tauxe, L., and P. Hartl, 11 million years of Oligocene geomagnetic field behaviour, *Geophys. J. Int.*, *128*, 217–229, 1997.

Tauxe, L., and N. J. Shackleton, Relative paleointensity records from the Ontong-Java Plateau, *Geophys. J. Int.*, *117*, 769–782, 1994.

Tauxe, L., and G. Wu, Normalized remanence in sediments of the western equatorial Pacific: Relative paleointensity of the geomagnetic field?, *J. Geophys. Res.*, *95*, 12,337–12,350, 1990.

Tauxe, L., T. Pick, and Y. S. Kok, Relative paleointensity in sediments: A pseudo-Thellier approach, *Geophys. Res. Lett.*, *22*, 2885–2888, 1995.

Teanby, N., and D. Gubbins, The effects of aliasing and lock-in processes on paleosecular variation records from sediments, *Geophys. J. Int.*, *142*, 563–570, 2000.

Teanby, N., C. Laj, D. Gubbins, and M. Pringle, A detailed palaeointensity record from drill core SOH-1 on Hawaii, *Phys. Earth Planet. Inter.*, *131*, 101–140, 2002.

Thellier, E., Sur la vérification d'une méthode permettant de déterminer l'intensité du champ terrestre dans le Passé, *C. R. Hebd. Seances Acad. Sci.*, *212*, 281–283, 1941.

Thellier, E., and O. Thellier, Sur l'intensité du champ magnétique terrestre dans le passé historique et géologique, *Ann. Geophys.*, *15*, 285–376, 1959.

Thibal, J., J. P. Pozzi, V. Barthès, and G. Dubuisson, Continuous record of geomagnetic field intensity between 4.7 and 2.7 Ma from downhole measurements, *Earth. Planet. Sci. Lett.*, *136*, 541–550, 1995.

Thouveny, N., K. M. Creer, and D. Williamson, Geomagnetic moment variations in the last 70,000 years, impact on production of cosmogenic isotopes, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *7*, 157–172, 1993.

Tric, E., J.-P. Valet, P. Tucholka, M. Paterne, L. Labeyrie, F. Guichard, L. Tauxe, and M. Fortugne, Paleointensity of the geomagnetic field during the last 80,000 years, *J. Geophys. Res.*, *97*, 9337–9351, 1992.

Tric, E., J.-P. Valet, P. Y. Gillot, and I. Le Meur, Absolute paleointensities between 60 and 160 kyrs B.P. from Mount Etna (Sicily), *Phys. Earth Planet. Inter. 85*, 113–119, 1994.

Tsunakawa, H., and J. Shaw, The Shaw method of paleointensity determinations and its application to recent volcanic rocks, *Geophys. J. Int.*, *118*, 781–787, 1994.

Valet, J.-P., and E. Herrero-Bervera, Paleointensity experiments using alternating field demagnetization, *Earth Planet. Sci. Lett.*, *177*, 43–53, 2000.

Valet, J.-P., and L. Meynadier, Geomagnetic field intensity and reversals during the past four million years, *Nature*, *366*, 234–238, 1993.

Valet, J.-P., and L. Meynadier, A comparison of different techniques for relative paleointensity, *Geophys. Res. Lett.*, *25*, 89–92, 1998.

Valet, J. P., and V. Soler, Magnetic anomalies of lava fields in the Canary Islands: Possible consequences for paleomagnetic records, *Phys. Earth Planet. Inter.*, *115*, 109–118, 1999.

Valet, J.-P., L. Meynadier, F. Bassinot, and F. Garnier, Relative paleointensity across the last geomagnetic reversal from sediments of the Atlantic, Indian and Pacific Oceans, *Geophys. Res. Lett.*, *21*, 485–488, 1994.

Valet, J. P., J. Brassart, I. Le Meur, V. Soler, X. Quidelleur, E. Tric, and P. Y. Gillot, Absolute paleointensity and magnetomineralogical changes, *J. Geophys. Res.*, *101*, 25,029–25,044, 1996.

Valet, J.-P., E. Tric, E. Herrero-Bervera, L. Meynadier, and J. P. Lockwood, Absolute paleointensity from Hawaiian lavas younger than 35 ka, *Earth Planet. Sci. Lett.*, *161*, 19–32, 1998.

Valet, J.-P., J. Brassart, X. Quidelleur, V. Soler, P.-Y. Gillot, and L. Hongre, Paleointensity variations across the last geomagnetic reversal at La Palma, Canary Islands, Spain, *J. Geophys. Res.*, *104*, 7577–7598, 1999.

Vanyo, J. P., A geodynamo powered by luni-solar precession, *Geophys. Astrophys. Fluid Dyn.*, *59*, 209–234, 1991.

Van Zijl, J. S. V., K. W. T. Graham, and A. L. Hales, The paleomagnetism of the Stronberg lavas of South Africa, II, The behaviour of the northern magnetic field during a reversal, *Geophys. J. R. Astron. Soc.*, *7*, 169–182, 1962.

Verosub, K. L., E. Herrero-Bervera, and A. P. Roberts, Relative geomagnetic paleointensity across the Jaramillo subchron and the Matuyama/Brunhes boundary, *Geophys. Res. Lett.*, *23*, 467–470, 1996.

Vinogradov, Y. K., and G. P. Markov, On the effect of low temperature heating on the magnetic state of multidomain magnetite, in *Investigations in Rock Magnetism and Paleomagnetism*, pp. 31–39, Inst. of Phys. of the Earth, Moscow, 1989.

Voelker, A., M. Sarnthein, P. M. Grootes, H. Erleunkeuser, C. Laj, A. Mazaud, M. J. Nadeau, and M. Schleicher, Correlation of marine ^{14}C ages from the Nordic Sea with GISP2 isotope record: Implication for ^{14}C calibration beyond 25 ka BP, *Radiocarbon*, *40*, 517–534, 1998.

Vogt, S., G. F. Herzog, and R. C. Reedy, Cosmonuclides in extraterrestrial materials, *Rev. Geophys.*, *28*, 253–275, 1990.

Wagner, G., J. Masarik, J. Beer, S. Baumgartner, D. Imboden, P. W. Kubik, H.-A. Synal, and M. Suter, Reconstruction of the geomagnetic field between 20 and 60 ka B.P. by the cosmogenic radionuclides ^{10}Be and ^{36}Cl from the Grip ice core and a new production model, *Nucl. Instrum. Methods Phys. Res., Sect. B*, *172*, 597–604, 2000.

Walton, D., J. Shaw, J. Share, and J. Hakes, Microwave demagnetisation, *J. Appl. Phys.*, *71*(3), 1549–1551, 1992.

Walton, D., J. Share, T. C. Rolph, and J. Shaw, Microwave magnetisation, *Geophys. Res. Lett.*, *20*, 109–111, 1993.

Worm, H.-U., A link between geomagnetic reversals and events and glaciations, *Earth Planet. Sci. Lett.*, *177*, 55–67, 1997.

Worm, H. U., M. Jackson, P. Kelso, and S. K. Banerjee, Thermal demagnetization of partial thermoremanent magnetization, *J. Geophys. Res.*, *93*, 12,196–12,204, 1988.

Xu, S., and D. J. Dunlop, The theory of partial thermoremanent magnetization in multidomain grains, 2, Effect of microcoercivity distribution and comparison with experiment, *J. Geophys. Res.*, *99*, 9025–9033, 1994.

Yamazaki, T., and N. Ioka, Long-term secular variation of the geomagnetic field during the last 200 kyr recorded in sediment cores from the western equatorial Pacific, *Earth. Planet. Sci. Lett.*, *128*, 527–544, 1994.

Yamazaki, T., N. Ioka, and N. Eguchi, Relative paleointensity of the geomagnetic field during the Brunhes chron, *Earth. Planet. Sci. Lett.*, *128*, 527–544, 1995.

Yang, S., H. Odah, and J. Shaw, Variations in the geomagnetic dipole moment over the last 12,000 years, *Geophys. J. Int.*, *140*, 158–162, 2000.

Yokoyama, Y., and T. Yamazaki, Geomagnetic paleointensity variation with a 100 kyr quasi-period, *Earth Planet. Sci. Lett.*, *181*, 7–14, 2000.

Yoshihara, A., and Y. Hamano, Intensity of the Earth's magnetic field in late Archean obtained from dibase dikes of the Slave Province, Canada, *Phys. Earth Planet. Inter.*, *117*, 295–307, 2000.

Yu, Y., and D. J. Dunlop, Paleointensity determination on the Late Precambrian Tudor Gabbro, Ontario, *J. Geophys. Res.*, *106*, 26,331–26,343, 2001.

Zhou, W., R. Van der Voo, and D. R. Peacor, Preservation of pristine titanomagnetite in older ocean-floor basalts and its significance for paleointensity studies, *Geology*, *27*, 1043–1046, 1999.

Zhou, W., R. Van der Voo, D. Peacor, and Y. Zhang, Variable Ti-content and grain size of titanomagnetite as a function of cooling rate in very young MORB, *Earth Planet. Sci. Lett.*, *179*, 9–20, 2000.

J.-P. Valet, Institut de Physique du Globe de Paris, 4 Place Jussieu, 75252 Paris Cedex 05, France. (valet@ipgp.jussieu.fr)