SOME ASPECTS OF THE GEOMAGNETIC FIELD DURING POLARITY TRANSITIONS

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ABSTRACT

A total of five records (three of them in Fuller's Category "A") of the Olduvai event (1.72 to 1.88 +/- 0.04 my) polarity transitions, from the Northern Hemisphere in the Pacific Ocean, ranging in latitude from 3° to 37° N., were paleomagnetically studied. These reversal records were obtained from deep-sea sediments of azimuthally oriented cores, using a high resolution technique (1 cc "micro-samples" spaced at 5mm intervals).

An overall analysis of these data showed the following features: (1) Intermediate directions of magnetization, (2) A definite directional change of 180°, (3) A pronounced decrease in intensity of magnetization, (4) A correlation between the decrease of magnetization and the directional change, and (5) A gradual decrease in inclination before the actual reversal, coincident with an intensity decrease, in which the inclination first shallows and then, prior to the intensity minimum, changes sign. After the change in sign, high inclinations are present and are associated with the intensity minimum (Olduvai N-R).

The directional changes during these reversals are shorter in duration than the intensity decrease. The observed data give a duration of 2600 to 10000 except for the onset of the Olduvai event (R-N), which is characterized by two oscillations of the geomagnetic field.
Here the intensity records suggest that the complete onset of the Olduvai event may have lasted as much as 50,000 years. Within the transition zones for the low latitude records the inclination values go up to 70°. ARM studies of the five transition records, at 70 mT and 0.5 Gauss coaxial DC fields, reveal a notably flat ARM across the transition zone and throughout the normal and reversed polarities. Moreover, the relative paleointensity study of normalized PDRM100/ARM700, shows characteristic lows during the transition zones, suggesting that such low transitional intensities are not due to mineralogical changes, but are instead due to the actual decrease in surface intensity of the geomagnetic field during polarity transitions.

The calculated VGP paths of the two successive transition records (Olduvai R-N-R) display, for the Olduvai termination (N-R), a near-sided behavior, whereas the R-N transition records are a borderline case of a far-sided behavior. The N-R records support the "flooding approach" model, and the distribution of VGP paths associated with this N-R reversal are roughly simulated by a field geometry consisting of as few as two non-dipole terms, a zonal octupole and a non-axisymmetric quadrupole term at the time of nearly zero dipole contribution.
The R-N VGP transitional paths do not fully support the "flooding approach" model or any other simple model, since the paths are not truly far-sided. This suggests that the field, although non-dipolar, cannot be roughly simulated by a field geometry consisting of only two terms, a zonal quadrupole and a non-axisymmetric octupole at the time of nearly zero dipole contribution. The complexity of these reversal records suggest that all the simple models (the standing field, zonal flooding and frozen flux models) are inadequate to describe the core processes during a magnetic field polarity transition.

Perhaps the data are telling us that no unique model is appropriate and that each reversal is representative of some random combination of several effects. If so, the constraints or core magnetohydrodynamics provided by these single models need to be more closely examined and perhaps discarded and/or replaced.
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INTRODUCTION

A series of previous papers have shown that paleomagnetic results from Tertiary and Quaternary deep-sea sediments generally give reliable records of geomagnetic phenomena, (Opdyke and Foster 1970, Opdyke 1972, Opdyke et al 1973, Hammond et al 1974, Theyer and Hammond 1974, Hammond et al 1979). These records of the behavior of the earth's magnetic field preserved in sediments are generally used to make time-stratigraphic correlations, to extend the geomagnetic time scale, and to provide additional insight into both long- and short-term descriptions of the earth's magnetic field. Until recently it has generally not been feasible to utilize paleomagnetic data from piston-cored deep-sea sediments for studies of short-term description of geomagnetic events such as polarity transitions. Before the advent of very sensitive cryogenic magnetometers such studies were almost entirely dependent on subaerially exposed strata, for the frequently encountered slow sediment accumulation rates and the usually relatively weak magnetization of deep-sea sediments precluded the resolution of detailed field behavior. One of the most important factors inhibiting such studies was our inability to record the in-situ orientation of sediments recovered by deep-sea coring or to ascertain what effects the coring process might have had on the recovered sediments. An instrument designed specifically to record corer behavior and orientation during sediment penetration has been successfully used to obtain this information and is now routinely used aboard Hawaii Institute of Geophysics research vessels (Seyb et al 1977). The
An orienting device utilizes an 8-mm movie camera to record vertical and horizontal orientation, vertical acceleration and time throughout free-fall and penetration.

At the present time, there have been few studies of polarity transitions using deep-sea sediments (Harrison and Somayajulu 1966, Opdyke et al 1973, Freed 1977, Hammond et al 1979, Clement et al 1982, Herrero-Bervera et al 1982a and b, Theyer et al 1982). Of all these studies only four studies have used the in-situ orientation of sediments recovered by deep-sea coring that allows a more reliable record of a short term variation of the earth's magnetic field to be made.

The purpose of this study is to use the paleomagnetism of oriented deep-sea sediment cores to investigate in detail the short-term variation of the geomagnetic field bounding the Olduvai event (R-N and N-R) in order to correlate and/or compare the data with the results obtained for the same reversal in the Indian Ocean by Opdyke et al (1973), and at the same time analyze the behavior of transitional fields at different latitudes.
MAGNETOSTRATIGRAPHY

For this particular study of the polarity transitions bounding the Olduvai event four fully azimuthally oriented piston cores were selected from the extensive HIG collection. Site locations for the four cores K7501, K78019, K78030 and K78113 are shown in figure 1 and are at: 37.3° N, 179.3° W; 8.95° N; 170.5° W; 18.9° N; 160.3° W; 2.7° N, 178.7° W respectively. Water depth at the K7501 site is 5883 m, at the K78019 site is 5270 m, at the K78030 site is 5015 m, and at the K76113 site is 5419 m. The true core length varies from 21.14 m (K78030) to approximately 17.3 m (K78019, K7501). These cores consist of brown pelagic radiolarian and diatom lutytes, with abundant sponge spicules and minor amounts of manganese micronodules, occasionally mottled, but otherwise uniform.

A preliminary paleomagnetic survey of the four cores was carried out taking samples at 10-cm intervals. This survey indicated the presence of five transitions of magnetic polarities which have been identified on the basis of micropaleontological criteria as the Matuyama-Brunhes (K78019 and K78030), the onset and termination of the Jaramillo event (K78030), and the onset (K78113 and K78019) and termination (K7501, K78019 and K78030) of the Olduvai event.

The remanent magnetization of the three cores was measured with a 6-cm three axis ScT cryogenic magnetometer. At least seven pilot samples from each core were subjected to progressive alternating field (AF) demagnetization of the NRM, using fields from 2.5 mT up to 50 mT to selectively eliminate any secondary (viscous) magnetization.
Figure 1. Location map showing the sampling localities in the North Central Pacific Ocean.
components. The resulting AF demagnetization curves and direction of magnetization, as well as the orthogonal intensity Zijderveld plots are shown in Figure 2.

The pilot sample results showed consistent directions of magnetization, to within 5° of the initial value, with demagnetization up to 40 mT. Thus, the results indicated good magnetic stability throughout the cores. Symons and Stupavsky's (1974) method was used to select the optimum demagnetization field intensity necessary to remove the secondary components. These analyses yielded a minimum paleomagnetic stability index of 15 millidegrees per mT at 10 mT for K7501, K76113 and K78030, and at 7.5 mT for K78019. As a result demagnetization fields of 7.5 mT and 10 mT were chosen for blanket treatment of all cores.

The direction and intensity of the remanence after treatment in the 7.5 and 10 mT alternating field were measured for all samples. In this way, the magnetic polarity boundaries were located. The close correspondence observed between the natural remanent magnetization (NRM) and the alternating field (AF) demagnetized values of remanent magnetization indicates that the sediments are magnetically stable.

With the exception of the uppermost 350 cm and lowermost 220 cm of K7501 the magnetostratigraphy is straightforward for all of the cores used in this study and is easily interpreted as shown in Figure 3a. (A relatively high water content, particularly within the upper portion of K7501, resulted in this interval being deformed during handling and sampling procedures and probably accounts for the
Figure 2. Normalized intensity, stability index, equal area stereo plot and orthogonal "Zijderveld" diagrams of progressive a. f. field demagnetization of samples from core K78030. a. Sample located at 1500 cm from top of the core. b. Sample located at 9500 cm from top of the core. c. Sample located at 4000 cm from top of the core. d. Sample located at 7000 cm from top of the core.
4000 cm
Figure 3a. Paleomagnetic data for core K7501 oriented with respect to true north and corrected for corer rotation. Sampling interval is 10 cm. Solid intervals indicate normal polarity in the core, open intervals indicate reversed polarity.
anomalous record for this interval). The sediments in K7501 extend from the Brunhes to the upper Gauss epoch. Considering the average sedimentation rate prevailing during most of the Matuyama epoch (estimated by Hammond et al. 1979, to be about 6 mm/1000 years for this core), and the fact that the core does not contain the Kaena event, the approximate age of the bottom of the core must be about 2.8 m.y.

The magnetic stratigraphy of core K78030 (Figure 3b,c) is easily interpreted as recording the Matuyama-Brunhes boundary, the Jaramillo event and the termination of the Olduvai event. The distances observed between these boundaries gives a sedimentation rate for post Jaramillo times of 2.6 mm/1000. A sedimentation rate is inferred from the distance between the Jaramillo and Olduvai events of 9 mm/1000 years, since these boundaries are separated by .78 my (Berggren et al. 1980). Core K78019 (Figure 3d) recorded the same transitions as K7501 and K78030, with an average sedimentation rate for the entire core of 13 mm/1000 years. In addition to the above mentioned transitions core K78019 records the Reunion event (2.04 m.y. Berggren et al. 1980) yielding an approximate basal age of 2.2 m.y.

Core K76113 (Figure 3e) recorded the same transitions as core K78030 with an average sedimentation rate for the entire Olduvai event of 8 mm/1000 years. Since the lowest polarity change is the onset of the Olduvai event one can extrapolate, using the above mentioned sedimentation rate; the bottom of the core to be of approximately 2.03 m.y. old. (Figure 3 a,b and c).
Figure 3b. Paleomagnetic data for core K78030, oriented with respect to true north and corrected for corer rotation. Sampling interval is 50 cm.
Figure 3c. Routine (at 50cm intervals with conventional 6-cc boxes), and high resolution sampling (1-cc boxes at 2-3mm intervals) in HIG piston core KK78-10-30. The high intensity sampling is at the Matuyama-Brunhes, Jaramillo and Olduvai termination transitions, as well as between the Jaramillo Termination and the Matuyama-Brunhes transition.
Figure 3d. Paleomagnetic data for core K78030, unoriented with respect to true North. Sampling interval 10 cm.
Figure 3e. Paleomagnetic data for core K76113 oriented with respect to true North and corrected for corer rotation. Sampling interval is 10 cm.
Dating these three cores was independently verified by means of their radiolarian and diatom biostratigraphies. As pointed out by Hammond et al. 1979, the occurrence of *Rhizosolenia matuyamai* in core K7501 specifically identifies the Jaramillo event. For the Olduvai event it is important to mention the First Appearance Datum (FAD) of *Pseudonotia doliolus*, the virtually synchronous Last Appearance Datum (LAD) of *Lamprocyclas heteroporos* and the FAD of *Eucyrtidium matuyamai*, all of which occur in the lower part of the Olduvai event (ca. 1.83 m.y.). This is true for core K7501; and for the Equatorial Pacific cores, K78019, K76113 and K78030. The LAD of *Pterocanium prismatum* occurs just above the Olduvai event (ca. 1.7 m.y.). These datum levels or biohorizons have been reported by Berggren et al. 1980 for the same areas (North Pacific and Equatorial Pacific); and for a wide variety of deep-sea cores (e.g. Theyer et al. 1978).
SAMPLING PROCEDURE

After the pilot studies had been completed, the sections of the core containing the Olduvai polarity transitions were sampled as continuously as possible by using plastic cubes of about 1-cc in volume, with sides measuring about 10 mm. The cubes were deployed in rows and columns; in order to achieve both greater vertical (chronological) and horizontal sample density. They were placed as close to each other as physically possible. Two samples were in each row outside the transition zones and three or four were in each row in the transition zones as is shown in Figure 4.

According to Berggren et al (1980); the Olduvai event normal interval age ranges from 1.88 +/- 0.04 m.y. to 1.72 +/- 0.04 m.y. yielding an approximate duration of 160 000 years. Thus, one can use the average sedimentation rates discussed in the previous sections to estimate that each of the 1-cc samples in transition zones represents about 500 years for K7501, about 250 years for K78030 and about 300 years for K78019.
Figure 4 Details of high-resolution sampling in core K78019 using overlapping 1-cc boxes at 2-3 mm intervals.
POSTDEPOSITIONAL REMANENT MAGNETIZATION PROCESSES IN THE MAGNETIZATION OF DEEP-SEA SEDIMENTS

One of the purposes of this study is the analysis of the behavior of the geomagnetic field during the transitions of the Olduvai event recorded in deep-sea sediments. Before the analysis is attempted a fundamental question has to be answered: Are our measurements representative of the behavior of the geomagnetic field when we study deep-sea sediments during a polarity transition?

In order to answer this question; it is necessary to carry out an evaluation of the data available for this study in terms of the nature of the source of the origin of the primary component of magnetization of deep-sea sediments.

The primary component of magnetization of such sediments could be due to a complex phenomenon called detrital remanent magnetization (DRM) involving both depositional and postdepositional effects. The magnetization of sediments is affected by interactions at the sediment/water interface by the dewatering of the sediment after deposition, and by Brownian motion. The totality of these effects, constitutes the detrital remanent magnetism, or DRM of sediments in general (Nagata, 1953). As explained by Verosub (1977), this term covers those processes which influence a detrital magnetic carrier from the moment it starts to settle through the water column until it begins to undergo chemical changes after its incorporation in the sediment. These chemical changes which can include oxidation and reduction of magnetic carriers as well as diagenesis of the matrix, give rise to
chemical remanent magnetization (CRM).

The origin of the NRM of deep-sea sediments has not been resolved completely so far, and studies using such sediments involving cores from nearly all over the world, support the conclusion that CRM is not important or at least is not the most important primary component in the cores investigated (e.g. Keen 1963, Lovlie 1971). Although some isolated and very special cases have been reported in which the NRM is due to a CRM acquired in situ in an indefinite length of time after deposition (Harrison and Peterson, 1965; Haggerty, 1970), this type of magnetization (CRM) does not seem to be the rule. On the other hand an overwhelming number of studies demonstrate that the origin of the NRM of deep-sea sediments may be acquired by postdepositional rotation of magnetic grains to align their individual magnetic moments with the geomagnetic field; e.g. Irving and Major (1964), Lovlie (1971), Kent (1973), Lovlie (1974, 1976).

The terms detrital remanent magnetism and depositional remanent magnetism were used by Nagata (1953) interchangeably in reference to the magnetization of sediments. Irving (1957) recognized that certain processes occur after deposition and introduced the term postdepositional detrital remanent magnetism (PDRM). As could have been expected, a confusion in terminology developed over the classification of postdepositional remanent magnetization. In this study the suggestion given by Verosub (1977) will be followed, namely; the term detrital remanent magnetization, or DRM, refers to the remanent magnetization found in sediments (exclusive of that due to
chemical changes). The term depositional DRM shall be used to refer to the magnetization acquired as the magnetic particles settle through the water column, interact with the substrate at the sediment/water interface, and then come to rest on the substrate. Postdepositional detrital remanent magnetization (PDRM) refers to any magnetization which is acquired after the magnetic carrier comes to rest on the substrate. Postdepositional DRM is acquired on a scale ranging from minutes to years after deposition.

Laboratory experiments and theoretical models have shown that three important effects have to be taken into consideration when one tries to evaluate how accurately sediments record the direction of the applied magnetic field; these are: inclination error, bedding error, and current rotation effects. The first two of these effects were named by King (1955), who studied them in detail, and the current rotation effect has been examined by Granar (1958). These three effects are all aspects of the interaction of the magnetic carrier with the substrate at the sediment/water interface. As such they represent part of the process of depositional detrital remanent magnetization. Graham (1954) and Irving (1957) clearly raised the question of whether or not a sediment could be magnetized (or remagnetized) subsequent to deposition. They were able to demonstrate the possibility of post-depositional DRM, but of course could not prove that it was responsible for the magnetization of sediments. All these laboratory experiments and theories serve to emphasize the fundamental question: How accurately do sediments record the direction of the applied magnetic
field at the time of sedimentation?. As Verosub (1977) states; the explanations of the errors which arise in the acquisition of a depositional DRM apparently involve very basic physical processes. Thus the inclination error should be present in all sediments immediately after deposition in still water. If the inclination error is not found, there are two possible conclusions. The first is that the inclination error has been corrected presumably by a postdepositional remagnetization. The second is that the errors in laboratory experiments are peculiar to the way in which laboratory sediments are deposited and are therefore irrelevant to natural sediments. Two facts could be used to support this point of view. In the first place, the rates of deposition of laboratory sediments are of necessity of the order of at least millimeters per day. Such rates are orders of magnitude greater than those found in natural deep-sea sediments. In the second place, an uncomfortably large percentage of the reported redeposition experiments have involved material from glacial varves. After 40 years of research of this particular issue we cannot yet make the statement that the inclination error has been found in laboratory redeposition experiments on a wide variety of sediments. In fact, Levi and Banerjee (1975) have reported that carefully prepared redeposited sediments from Lake St. Croix, Minnesota, apparently do not exhibit an inclination error".
As it stands right now in the absence of additional support for the second conclusion given by Verosub (1977); the first is usually accepted. The fundamental question of the accuracy of the directions of magnetization of sediments as proposed by Verosub (1977) then takes two forms: 1) Is there evidence for alignment errors in nature? 2) Is there evidence for a postdepositional remagnetization? If postdepositional remagnetization is present it precludes the existence of errors arising from depositional effects. However, the absence of a postdepositional remagnetization does not a priori require the presence of an inclination error, particularly if the laboratory experiments do represent anomalous results.

This study is solely based on deep-sea sediments and thus the analysis will be done taking in account only those problems related to deep-sea sediments. Deep-sea sediments are at the opposite extreme from glacial varves, in terms of rate of deposition, for deep-sea sediments have typical rates of accumulation of a few millimeters to a few centimeters per 1000 years. Keen (1963) carried out the first detailed study of the magnetization of deep-sea sediments using cores from the North Atlantic Ocean. Keen made three important observations, which have been borne out by subsequent work: 1) The inclination is similar to that of the ambient field. 2) Bioturbation destroys any depositional DRM, and 3) Slumped or deformed material can acquire a postdisturbance magnetization.
At the site from which Keen obtained his cores, the ambient field inclination is about the same as that predicted by a geocentric axial dipole model. This inclination is given by the basic paleomagnetic dipolar equations. However, agreement between the observed inclination and that given by the paleomagnetic dipolar equation cannot be taken as evidence for the lack of an inclination error, especially if the data are averaged over a limited temporal and spatial extent. Such an inference requires the assumptions that the time available at a site is sufficiently long to represent a time-averaged geomagnetic field and that the time-averaged geomagnetic field is in fact a geocentric axial dipole.

Work done by Harrison (1966) and later by Opdyke and Henry (1969) showed that the mean core inclinations were in good agreement with the inclinations predicted by a geocentric axial dipole. The excellent agreement led Opdyke and Henry to conclude that averaged over periods of a million years the geomagnetic field was indeed that of a geocentric axial dipole. Since this conclusion assumes that the sediments were accurately recording the earth's field Opdyke and Henry were implicitly concluding that there was no inclination error. The only alternative is that the field is nondipolar in precisely the right way to mask the inclination error. This unlikely situation can be ruled out as a result of a spherical harmonic analysis involving both sediments and volcanic rocks (Georgi, 1974). More recent studies do show that the geomagnetic field may have long-term nondipole features (Wilson, 1970, Watkins 1972, Creer et al 1973; Verosub 1975b, Epp et al
1983). However, these features produce a north-south latitude asymmetry in inclination (Wilson, 1971) rather than the symmetric effect expected for the inclination error. Moreover, the change in inclination averages only a few degrees at most latitudes (Watkins and Richardson 1975), which is unimportant within the context of global analysis of data. Thus, in the case of stably magnetized deep-sea sediments, we can be relatively confident that the direction of the magnetization accurately reflects the orientation of the ambient magnetic field at the time the magnetization is acquired. Studies of such sediments have shown in fact, that small changes in inclination can be used to infer reliably the motion of oceanic plates. Hammond et al. (1974) studied several Central Pacific cores and determined the difference between the geocentric axial dipole inclination at the present core latitude and the actual inclination for samples from the cores. They reported a gradual increase in this difference as a function of depth in the cores which they interpreted as being due to northward movement of the Pacific plate.

As pointed out by Verosub (1977) these observations lead one to conclude that the fundamental nature of the physical processes which give rise to the inclination error, coupled with its absence in deep-sea sediments, implies that the sediments are remagnetized after deposition. Even more compelling evidence for postdepositional DRM comes from the presence of bioturbation of the sea floor. Photographic evidence for the presence on the sea floor of mobile benthic organisms (Laughton 1963) shows that most if not all
deep-sea sediment is susceptible to disturbance by marine organisms. This has been confirmed by photographic studies from deep-sea submersibles (Grassle et al., 1975), faunal studies from box cores and dredge hauls (Hessler and Junars, 1974), and trace fossil studies from DSDP cores (Chamberlain, 1975). Estimates of the depth of bioturbation have been obtained from consideration of the redistribution of ash layers (Ninkovich et al., 1966; Ruddiman and Glover, 1972), the dispersal of microtektites (Glass, 1969), and the incorporation of radioactive contaminants (Noshkin and Bowen, 1973). These techniques lead to the conclusion that fresh surface material can be mixed with subsurface material to a depth ranging from 10 to 60 cm in certain environments (Guinasso and Schink, 1975). Keen (1963) and Harrison (1966) both recognized that bioturbation of this magnitude would destroy any depositional DRM. Since most of the sea floor is bioturbated (Sanders and Hessler, 1969) and since most deep-sea cores display coherent magnetic directions (Opdyke, 1972), this magnetization must represent a postdepositional DRM. Watkins (1968) treated the problem of bioturbation and paleomagnetism in general terms. He used models to show that intermittent faunal disturbance could not only change the location and width of polarity boundaries but could also erase real magnetic reversals or create spurious ones in the paleomagnetic record of a core.
Bioturbation is a process by which a magnetization is destroyed. Thus a logical question is: By what process then does a deep-sea sediment become remagnetized? Keen (1963) noted that small-scale slumps within his deep-sea cores appeared to have been remagnetized. He suggested that the remagnetization occurred because the sediment had been sufficiently fluid to permit reorientation of magnetic particles. These ideas, of course, parallel those of Irving (1957). More recently Kent (1973) demonstrated the feasibility of magnetizing natural sediments by a postdepositional process. Kent stirred slurries of deep-sea material in an applied magnetic field and allowed the material to dry in that field. When the samples were relatively dry, they were found to possess a stable remanence and no inclination error was found. Thus it can be concluded that so far most sediments from the deep sea are remagnetized subsequent to deposition and that this postdepositional process does not lead to errors in the recorded magnetic inclinations.

Another important effect is the water content in deep-sea sediments. Khramov (1968), deposited red clays from suspension; for some samples he increased the intensity of the applied field after sedimentation had been completed. For sediment with a water content of 70 per cent this procedure increased the magnetization, while for a water content of 30 per cent there was no effect. A similar effect was reported by Kent (1973) in his studies of slurries of deep-sea sediments. Relatively dry sediment was found to possess a stable remanence, while unreliable results were obtained from
samples which were "too wet". Thus the water content of a sediment appears to determine whether or not the magnetic carriers have sufficient mobility to rotate within the fluid-filled voids. It is important to mention that the HIC's cores are all collected using liners and are stored wet at 2 to 4°C, in the HIC's core repository after opening. Thus the in situ conditions of the samples tend to be preserved and spurious effects due to drying are minimized.

The importance of pore water to the magnetization of sediments implies that we can distinguish two different magnetic sedimentary environments as pointed out by Verosub (1977); if the initial water content of a sediment is below the critical value, there can be no reorientation of magnetic carriers, and a depositional DRM will be recorded and preserved. On the other hand, if the content is above the critical value, remagnetization will occur subsequent to deposition, and a postdepositional DRM will result, presumably with the absence of an inclination error. The function of bioturbation in the magnetization of sediments now becomes clear. Although in some cases the initial water content of sediment may be above the critical value in the absence of bioturbation, the continual disturbance and mixing of the sediment by benthic fauna can create or maintain the high water content needed for remagnetization. Only when the sediment is dewatered does the magnetization become fixed.
The dewatering of natural sediments occurs not through a process of dehydration as is required for laboratory experiments, but rather through a process of compaction. Yaskawa (1974) has attempted to estimate the lock-in depth of postdepositional DRM based on a simple viscous fluid model in which the viscosity is affected by gravitational compaction. However, it is unlikely that significant advances can be made without careful consideration of the microscopic properties of sediments. Meade (1966) in a review of the factors influencing the compaction of certain clays and sands points out that particle size is the main determinant of water content although clay mineral composition and shape are also important. In the case of postdepositional DRM arising from the rotation of magnetic carriers in water-filled voids, the microscopic properties of sediments which are likely to be important are sediment particle size, shape and mineralogy as well as magnetic particle size and composition. In particular, the ratio of sediment particle size to magnetic carrier will determine the free volume available to the carriers and the ease with which the carriers can rotate.

The influence of the microscopic properties of the sediments on the process of magnetization has already been seen in some studies. Thompson and Kelts (1974) reported that the basal portions of turbidite layers in Lake Zug, Switzerland, have lower magnetic intensities and lower inclinations than the laminated upper portions. One possible explanation for this is that the coarser basal portions preserve a depositional DRM, while the finer material of the upper portions
reflects a postdepositional DRM. This would imply that rapid deposition of coarse material leads to a water content lower than the critical value. A similar dichotomy has been reported for glacial varves from Sweden (Granar, 1958) and Iceland Griffiths et al., (1960). In both cases the silty summer layers have shallower inclinations than the clay-rich winter layers. Clearly more work must be done in the role of grain size in the magnetization of sediments.

Under appropriate circumstances it may be that depositional DRM and postdepositional DRM are competing processes. Postdepositional DRM appears to be potentially a much more complex process than depositional DRM. Magnetic carriers of different size, shape, and mineralogy may combine with sediments of different size and composition to produce a range of competing magnetization processes.

In summary, unconsolidated sediments, particularly those from environments with high rates of deposition, provide remanent magnetic records of very high time resolution. Whereas paleomagnetic data from suites of sedimentary or igneous rocks are usually combined to obtain mean values of geomagnetic parameters, each sample or set of samples from a given horizon in a sediment can be interpreted as a separate magnetic datum (Verosub, 1977). In this way it is possible to obtain from sediments the detailed variation of the geomagnetic field recorded by the processes which lead to the magnetization of a sediment. The problem then is to sort out the true behavior of the earth's magnetic field from the effects of the magnetization process. In order to do this it would be extremely helpful to know the range of inclination,
declination, and intensity actually recorded in the same stratigraphic horizon at different points in a given sedimentary environment. An attempt to do this has been made using red beds of the Chugwater Formation, (Herrero-Bervera and Helsley 1983); and the current study in particular is an effort to study the short-term behavior of the earth's magnetic field specifically during a polarity transition using deep-sea sediments from the Pacific Ocean at different latitudes and longitudes; keeping in mind the above mentioned problems relative to the depositional behavior of sediments.
PREDICTED RELATIONS BETWEEN POLARITY UNITS

As mentioned by Irving (1964), a reversal of magnetization is defined as the actual change from one polarity to another. If the observed reversals of magnetization are, for the most part, records of field reversals, then certain effects should be recognized, and the presence or absence of these may be taken as evidence for or against the hypothesis of field reversals. These predictions can be summarized in the following points:

1) The same polarity should occur in all rocks (for this study, deep-sea sediments) of the same age, irrespective of rock type, in all regions; provided the field has maintained the same polarity for a sufficiently long period of time to establish, by geological methods, the contemporaneity of beds. The time variations of polarity should be the same, period for period, in the standard sections of all regions. Positive and negative polarity zones should be identifiable, and these should occur in stratigraphic succession, so that boundaries between zones of opposite polarity are parallel to the time planes and do not cut across them.

2) At the boundaries between positive and negative levels there should occur, in some instances, beds with transitional directions of magnetization. Because of the fragmentary nature of the record it is not to be expected that transitional levels will occur in all cases. In favorable circumstances, traverses through the same stratigraphic levels at different localities should reveal comparable direction changes.
3) There should be no systematic differences of chemical, mineralogical or physical properties of the magnetic minerals in positive and negatively polarized rocks.

At this point the first prediction given by Irving (1964) has been discussed previously in this study and the correlation of the same reversals has been made at different locations in the Pacific by means of fossils for the entire Olduvai event (R-N and N-R transitions).

For the second prediction and probably the most important of all three (beds with transitional directions of magnetization) the initial remanence (NRM) of all samples was measured on a 6-cm vertical axis ScT cryogenic magnetometer, first, in order to identify the Olduvai event boundaries and second to select the pilot samples to be used for the demagnetization experiments.
PILOT SAMPLE DEMAGNETIZATION EXPERIMENTS

Alternating field demagnetization experiments were performed using a Schonsted GSD-1 demagnetizer; in order to attain the stable component of magnetization of the samples. A number of pilot samples were used and subjected to progressive a.f. demagnetization in order to investigate the magnetic behavior of the samples in the transition zones of all the cores; and the adjacent samples taken from above and below such transition zones. The main purpose of demagnetization is to remove the secondary components of magnetization and to isolate a stable, characteristic magnetization. This characteristic component is identified by a linear trajectory decaying towards the origin of an orthogonal vector endpoint diagrams (Zijderveld 1967). As it has been shown at the beginning of this study the conventional size (6-cc) samples taken at 10 cm intervals displayed a remarkable stability (e.g. core K78030, Figure 2). The high resolution samples (1-cc) were progressively a.f. demagnetized at steps of 2.5 mT and up to fields of 60 mT for most of the samples. In a few special cases; the maximum applied field was less than 60 mT particularly where the samples are within the transition zones, and the magnetization is approximately one order of magnitude lower than that of the samples outside the transition zones. This is the case of sample 90680; (Figure 5) where the signal is comparable with the noise level of the cryogenic magnetometer after 20 mT demagnetization field. This is due to the low intensity of magnetization of the sample, and presumably due to the low intensity of the earth’s magnetic field at that particular
stratigraphic level in the core. Any demagnetization made on these pilot samples at higher alternating fields resulted in an erratic behavior of the magnetic vector, and no reproducibility of results was achieved at fields higher than 20 mT for sample 90680. Although this result could be interpreted as indicating that the sample had an apparent unstable behavior, this is not the case for it was not the magnetization of the sample that was unstable but simply the lack of adequate sensitivity in the magnetometer used to measure the magnetization. Similar results are noted in samples 12644 (maximum field 40 mT) and 13644 (maximum field applied 30 mT).

Only four out of fifteen pilot samples displayed an initial increase in intensity, probably resulting from the removal of a secondary component of magnetization in the direction of the present earth's magnetic field from the more stable reversly magnetized primary component. This secondary component is clearly seen on the orthogonal vector endpoint diagrams of samples such as 12335, 12578 (Figure 8), 20891, 20900 (Figure 7).

After the removal of such secondary components at fields of about 7.5 to 10 mT a univectorial behavior of the characteristic component is revealed by the linear trajectory decaying towards the origin of the Zijderveld diagrams (Zijderveld, 1967). The rest of the normalized diagrams of the pilot samples used for this study show a decrease in intensity of magnetization upon demagnetization instead of an increase of intensity noted in the samples mentioned above.
Samples within the transition zones 10421, (K7501, Figure 6), 12644 (K78019, Figure 8), 13644 (K78019, Figure 9), 20900 (K78030, Figure 7), and 90680 (K76113, Figure 5) all have one common characteristic; in terms of directional changes they remain as intermediate samples relative to their respective transitions. One should also note that samples from core K78019 are less stable than samples from the other intermediate zones shown in Figures 5 to 9.

Another set of pilot samples taken at the lower and upper boundaries of each transition zone show highly stable magnetization. Examples of such behavior are shown by samples 12578 (Upper) and 12701 (Lower) (Figure 6) of core K78019 (Olduvai Termination), 90180 (Upper) and 95770 (Lower) (Figure 5) of core 76113 (Olduvai Onset); 20891 (Upper) and 21079 (Lower) (Figure 7) of core K78030 (Olduvai Termination). These samples are representative examples of the high stability of the specimens at the boundaries of the transition zones. Samples 13367 and 13559 of core K78019 (Olduvai Onset) (Figure 9) are stratigraphically well outside the transition zone.

The purpose of this pilot sample experiment was to test the stability of the samples at different stratigraphic levels in order to compare the magnetic characteristics of the samples within the transition zones versus the magnetic behavior of samples outside the transition zones. The samples outside the transition zones presumably have recorded the behavior of the predominantly dipolar portion of the earth's magnetic field whereas the samples at the boundaries of, or within, the transition zones inferably might have recorded the
nondipolar part of the earth's magnetic field. It can be observed from the magnetic features of the "dipolar" pilot samples, that they are characterized by a high stability upon demagnetization at fields of about 10 to 12.5 mT where the directional changes range from $5^\circ$ declination as well as inclination. Above alternating fields of 10 to 12.5 mT the samples showed a univectorial behavior (Figures 5 to 9) indicating that at that the primary component or characteristic magnetization has been isolated by alternating field demagnetization of 10 mT and that 10 mT probably represented the best field to use for the demagnetization of the rest of the samples.
Figure 5 Normalized intensity, stability index, equal area stereo plot and orthogonal "Zijderveld" diagrams of progressive alternating field demagnetization of transitional samples from core K76113.
Figure 6. Normalized intensity, stability index, equal area stereo plot and orthogonal "Zijderveld" diagrams of progressive alternating field demagnetization of transitional samples from core K78019 and K7501.
Figure 7. Normalized intensity, stability index, equal area stereo plot and orthogonal "Zijderveld" diagrams of progressive alternating field demagnetization of transitional samples from core K78030.
Figure 8. Normalized intensity, stability index, equal area stereoplot and orthogonal "Zijderveld" diagrams of progressive alternating field demagnetization of transitional samples from core K78019.
Figure 9. Normalized intensity, stability index, equal area
steric plot and orthogonal "Zijderveld" diagrams of
progressive alternating field demagnetization of
transitional samples from core K78019.
RESULTS OF THE DEMAGNETIZATION EXPERIMENTS.

The pilot samples showed a median destructive field (MDF) i.e. the alternating field necessary to randomize 50 per cent of the original remanence, ranging from 10 mT to 35 mT (core K7501 MDF=20 mT, K78019 MDF=17 mT SD=4.4, K76113 MDF=16 mT SD=5.2, K78030 MDF=29 mT SD=8). A soft component of magnetization, with a coercivity less than 7.5 to 12.5 mT makes up 10 to 20 per cent of the total NRM for most of these cores. All the specimens for the Olduvai event were demagnetized to 10 mT, which was the field necessary to remove the soft-coercivity, secondary viscous components. Similar results have been reported by other workers who have conducted palomagnetic studies of deep-sea sediments from the Central North Pacific (e.g. Opdyke and Foster 1970, Opdyke et al 1973, and Kent and Lowrie 1974). The cores used for this study do not become magnetically unstable at depth, like the cases reported by Kent and Lowrie 1974. Kent and Lowrie observed a marked magnetic instability in some deep-sea sediment cores, in which the natural remanent magnetization (NRM) of these "red clay" sediments was unstable below several meters depth. They noted that the magnetic instability was related to the presence of a relatively large low-coercivity component of magnetization. The cores selected for the study of polarity transitions in the Olduvai were collected outside the area of magnetic instability reported by Kent and Lowrie 1974, (see Kent and Lowrie 1974, Figure 1).
The termination (upper boundary) of the Olduvai event was recorded in cores K78019, K78030 and K7501 and the onset in cores K76113 and K78019. These two transitions will be evaluated independently since they represent discrete events in the history of the earth's magnetic field. The original name Olduvai was given to the total polarity event and to distinguish each transition the terminology Olduvai onset and Olduvai termination will be used.
OLDUVAI TERMINATION

In order to examine in detail the behavior of the earth's magnetic field during a change of polarity of the field, the declination, inclination and intensity of the Olduvai termination observed in each core K7501, K78030 and K78019 have been plotted in Figures 10, 11 and 12.

All three records have the following features:

1) Intermediate directions of magnetization,
2) A definite directional change of 180°
3) A pronounced decrease in intensity of magnetization
4) A correlation between the decrease in intensity of magnetization and the directional change, and
5) A gradual decrease in inclination before the reversal, coincident with an intensity decrease, in which the inclination first shallows and then, prior to the intensity minimum changes sign.
6) After the change in sign, high inclinations are present and are associated with the intensity minimum.

These characteristics of the polarity transitions are best seen in K7501 and K78019 records, and are not as well recorded in K78030 since it has the lowest sedimentation rate of the three cores.

Since the above mentioned five features are present at all three sites, they presumably correspond to the behavior of the earth's magnetic field during its transition from a normal to a reversed polarity (in this particular case the N-R polarity transition at the termination of the Olduvai event). In order to model the entire record
Figure 10. Stratigraphic plot of inclinations, declinations and intensities of core K7501. Demagnetization level at 10 mT.
CORE NUMBER: K7501
DEDEMAGNETIZATION INTENSITY: 100 OERSTEDS

DEMG
INCLINATION
(DEGREES)

DEMG
DECLINATION
(DEGREES)

DEMG
LOG. INTENSITY
(0.001 A/M)

5,000 yr
Figure 11. Stratigraphic plot of inclinations, declinations and intensities of core K78030. Demagnetization level at 10 mT.
CORE NUMBER: K78030
DESMAGNETIZATION INTENSITY: 100 OERSTEDS

DESMAG INCLINATION (DEGREES)

DESMAG DECLINATION (DEGREES)

DESMAG LOG. INTENSITY (0.01 R/M)

CSL: 18.90
CSLG: -168.30

DEPTH FROM TOP OF CORE IN M:

5,000 yr
Figure 12. Stratigraphic plot of inclinations, declinations and intensities of core K78019. Demagnetization level at 10 mT.
CORE NUMBER: K78019
DEGMAGNETIZATION INTENSITY: 100 OERSTEDS

DEMG
INCLINATION
(DEGREES)

DEMG
DECLINATION
(DEGREES)

DEMG
LOG. INTENSITY
(0.001 A/M)

-90.00 0.00 90.00
-90.00 0.00 90.00
0.00 270.00 -5.40
-5.00 -4.00

14,000 yr. 12,000 yr. 10,000 yr. 15,000 yr. 12,000 yr. 10,000 yr. 17,000 yr. 12,000 yr. 10,000 yr.

DEPTH FROM TOP OF CORE IN M
in terms of dipolar and nondipolar behavior the data set must be separated into a dipolar part and a nondipolar portion. By means of this separation one perhaps can answer one of the main questions mentioned earlier, namely, are we looking at the behavior of the geomagnetic field or are we just seeing an effect of the sedimentation process itself?

Table 1 shows the statistics of the normal as well as the reversed intervals and the combined mean of the two polarities. The mean declination of the two polarities are practically antipodal (180° ± 3°) in each of the three cores. This indicates that a stable magnetization has been isolated. Similar results were obtained by Herrero and Helley, (1983, Appendix A) within sedimentary rocks of the Chugwater Formation. Another important observation is that all the reversed and normal polarities of all the cores show a close similarity to that of the ambient field at the site from which the cores were obtained. This ambient field inclination is about the same (± 5°) as that predicted by the geocentric axial dipole field model. However, as stated earlier agreement between the observed inclination and that given by the dipolar formula cannot be taken as evidence for the lack of an inclination error, especially if the data are averaged over a limited temporal and spatial extent. Such an inference requires the assumptions that the time available at a site is sufficiently long to represent a time-averaged geomagnetic field and that the time-averaged geomagnetic field is in fact a geocentric axial dipole. The minimum time available at the sites is 6500 yr. (one polarity) and the
Table 1

<table>
<thead>
<tr>
<th>Core</th>
<th>Lat. (N)</th>
<th>Long. (W)</th>
<th>D</th>
<th>I</th>
<th>PI</th>
<th>K</th>
<th>$\alpha_{95}$</th>
<th>P. Lat.</th>
<th>dp</th>
<th>dm</th>
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<td></td>
<td></td>
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<td></td>
<td></td>
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<td></td>
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<td>170.3</td>
<td>123</td>
<td>181</td>
<td>-10.5</td>
<td>17.5</td>
<td>18</td>
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<td>179</td>
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<td>17.5</td>
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N, number of samples; D, declination in degrees, east of north; I, inclination in degrees, positive downward; PI, paleoinclination, in degrees, K, Fisher's precision parameter; $\alpha_{95}$ semiangle of cone of 95% confidence for mean directions; and (dp, dm), oval of 95% confidence about the pole position.
The maximum (both polarities) is approximately 45,500 yr. depending upon the respective sedimentation rates of the cores. These estimates are sufficiently long to represent a time-average geomagnetic field and therefore it can be safely inferred that these averages do in fact represent a geocentric axial dipolar field for the respective polarities of the cores.

The significance of the close resemblance between the mean inclinations recorded by the sediments and that of the ambient field as predicted by a geocentric axial dipole field model is that after the removal of the secondary components of magnetization, the primary component or characteristic magnetization is primarily due to a PDRM process. The normal interval of core K78019 (Table 1) where the recorded inclination value is 12° less than the expected value may be an exception to this generalization and may suggest that at least a portion of the primary magnetization may be due to a DRM process and not to a PDRM process like the rest of the intervals analyzed here. It is important to recognize that the estimate of the mean inclination of this particular interval (Normal K78019) was based upon only 21 samples and the time represented by these samples may or may not be adequate to identify the geocentric axial dipolar field.

The important observation made here, namely the lack of inclination error, leads to the conclusion that the sediments were accurately recording the earth's magnetic field during both normal and reversed polarities of the field.
With this important conclusion in mind, and remembering the magnetic stability of the samples, it is unlikely that the limited interval that records the transition zones would behave in a different manner. Thus, it is highly probable that the presumably nondipolar behavior of the earth's magnetic field recorded by the sediments, is in fact real. This is especially true for cores K7501 and K78030. Since the duration of transitional behavior has been independently estimated by Cox et al (1975) to be of the order of 10000 years, and this is the duration of time represented by a range of 8 to 20 cm of sediment (Figures 10, 11, and 12) it is highly probable that the observed transition behavior represents a real record of field behavior. Thus, an evaluation of the possible systematic characteristics of the Olduvai termination transition zone within and between sites is appropriate.
DETAILS OF THE POLARITY TRANSITION.

Well defined intermediate directions of magnetization are present in all three records of the Olduvai termination (see figures 10, 11 and 12). However, it appears that the most detailed and the most regular record of the Olduvai termination transition was recorded by core K7501 (Figure 10). The average sedimentation rate from the top of the Jaramillo event down to the top of the Gauss epoch is about 16 mm/1000 years, in this core. This rate has been reported by Seyb (1977) and Hammond et al (1979) and has been confirmed by the present study.

The transitional field appears to be recorded in approximately 16 cm section of core, beginning at 10480 mm and continuing to 10320 mm (Figure 10). This is based solely on the 180° directional change which appears to have taken approximately 10,000 years. Based only on the intensity record, an estimate of the duration of the Olduvai termination record is about 12,000 years (from 10510 mm to 10320 mm) or 16,000 years (from 10587 mm to 10310 mm) if the estimate is made from the beginning of the intensity period 2 as defined by Herrero-Bervera and Helsley (1983 Appendix A). It is obvious at this point that a different estimate of reversal length might be made from inclination or intensity changes, from one single record. In this case, the observed data give a change in direction which takes place on a shorter time scale than does the intensity change. The intensity from the same core shows a decrease that reaches a minimum between 10356 and 10457 cm and then begins to increase in value (this minimum presumably corresponds to the Period 3, nondipolar dominance interval...
proposed by Herrero-Bervera and Helsley (1983, Appendix A)). The minimum intensity is approximately 25 per cent of the average intensity observed after the dipolar field is completely recovered.

In core K78030 the intermediate directions are observed through 2.4 cm of sediment from 2092.3 to 2089.9 cm (Figure 11). The duration of the directional change based on a sedimentation rate of 9 mm/1000 years is about 2600 years. The accompanying intensity fluctuation occurs between 2095.1 to 2087.7 cm reaching a minimum at 2090.4 cm. The intensity duration took place in about 8200 years and during this 7.4 cm interval the intensity drops to about 15 per cent of the average value after the field was entirely recovered. It is important to point out that the short duration, fewer intermediate directions, and abrupt changes in direction suggest that there may be some time missing in this record.

The transition record obtained from core K78019 from a site at 9° North of the equator is characterized by an upward steepening (Maximum inclination + 30°) in inclination from 1269.6 to 12660.0 cm where it changes sign from positive to negative and undergoes a rapid increase in inclination from 1266.0 to 1260.0 cm. The total direction change occurs across 9.6 cm of sediment which equates to about 6400 years based on a sedimentation rate of 15 mm/1000 years. This change is accompanied by a decrease in intensity between 12719 cm and 12579 cm to about 20 per cent of the maximum intensity measured after the reversed field is established.
The intensity of magnetization remaining after alternating field demagnetization within the zone of intermediate directions is much lower than that in the overlying zone of reversed magnetization and in the underlying zone of normal magnetization for all three cores (Figures 10,11 and 12).

A decrease in the intensity of magnetization can be due to: 1) Less effective magnetic alignment resulting from a decrease in the strength of the magnetizing field, 2) A cancellation effect due to oppositely aligned fractions within the sample; Kobayashi et al (1971), 3) A decrease in the amount of magnetic material present in the core.

Possibility 2 can probably be ruled out in all three cores because the number of adjacent samples in which reduced intensity was observed, this is the same conclusion reached by Opdyke et al (1973) with deep-sea sediments of core RC14-14. In order to test for possibility 3, anhysteretic remanent magnetization (ARM) measurements were made on all samples.

The ARM was induced in each specimen from each horizon using a peak alternating field of 70 mT on a D.C. field of 0.5 Gauss. For all specimens alternating field (AF) demagnetization at 70 mT reduced the magnetization to 10 per cent of its initial value (see Figures 5 to 9). Therefore ARM acquired in a peak alternating field of 70 mT gives an adequate representation of the magnetic grains which carry the bulk of the NRM (Hillhouse and Cox 1976). Anhysteretic magnetization provides a measure of the magnetizability of sediments and may be used to
crudely normalize remanence intensity measurements to yield a number that is more nearly proportional to the ancient variations in field intensity. Figures 13, 14, and 15, show the response the induced ARM, and all three display a remarkably uniform behavior across transition zones and through the normal and reversed intervals. As a result, the ARM experiment indicates that the low intensity of magnetization in the transition zones are probably not due to mineralogical changes. The same results have been reported by other workers in transition zones; e.g. Hillhouse and Cox (1976), Liddicoat (1982), Valet and Laj (1981), Theyer et al (1982).

Another possibility is that magnetization is acquired gradually over some depth range below the sediment water interface; this effect may be due to slow compaction and dehydration of sediments originally deposited as a slurry on the sea bottom. If the sediments become magnetized very slowly over a time interval of 10000 years or more, the transition zone might acquire opposing components of magnetization, resulting in a low net intensity of magnetization. However it is unlikely that the transition intervals that range from about 3 cm to 16 cm with low intensity became magnetized in this way, since core sediment above the transition zones contains a good paleomagnetic record of geomagnetic field behavior (secular variation), which would not be observed unless the acquisition time of remanence were much much less than 10,000 years. (Baag and Helsley, 1974). Assuming that each specimen was magnetized in a total time not much greater than the time required for its deposition and that the
Figure 13. Stratigraphic plot of VGP Latitude and anhysteretic remanent magnetization (ARM) of core K7501. Demagnetization level 10 mT.
Figure 14. Stratigraphic plot of VGP Latitude and anhysteretic remanent magnetization (ARM) of core K78030. Demagnetization level 10 mT.
Figure 15. Stratigraphic plot of VGP Latitude and anhysteretic remanent magnetization (ARM) of core K78019.

Demagnetization level 10 mT.
K78019  
10.0 MILLI-TESLA

VGP  
LATITUDE

A.R.M.  
LOG. INTENSITY  
(0.001 A/M)

70 mT  
0.5 Gauss

5,000 yr.
transitional direction changes occurred over an interval of approximately 2500 years to 10000 years, each specimen represents a 250 to 500 year time average of the ancient field. Thus, one can conclude that the low intensities of the cleaned NRM during transitions in these three cores reflect a decrease in the intensity of the geomagnetic field.
The behavior of the onset of the Olduvai event was studied using two azimuthally oriented cores K78019 (8.95° N and 170.30° W) and K76113 (2.66° N and 173.44° E). The demagnetization field selected on the basis of the pilot study on samples of the two cores was 10 mT. It is important to mention that blanket treatment of the samples at higher fields 20mT and 25mT were performed giving virtually identical results as the results obtained at 10 mT; indicating that a soft secondary viscous component was removed at this low field. The rest of the discussion is based solely at this particular field (10mT).

The two records obtained showed the following features:

1). Intermediate directions of magnetization.
2). A pronounced drop in intensity of magnetization.
3). The drop in intensity of magnetization is a combined effect of the two geomagnetic oscillations prior to the main directional change and the 180° directional change itself where the intermediate directions appear.
4). A definite set of three directional changes of 180° occurred at the onset of the Olduvai event. Two of these sets of directional changes are rapid oscillations of the geomagnetic field prior to the entire reversal of the earth's magnetic field.
Similar findings to those in item 4 have been reported by other workers from different transitions (e.g., Gauss-Matuyama, Liddicoat, (1982) N-R, 2.48 Ma). (Herrero-Bervera, et al 1982, and Theyer et al 1982). Such oscillations have been predicted by Yoshimura (1980), where he presents a reversal mechanism relevant to steady dynamos like that of the earth based on solutions of the nonlinear dynamo equation having an assumed time-delayed feedback process. In particular, his study predicts that during a polarity transition a number of field reversals or oscillations, may take place. Yoshimura (1980) notes that the number of such oscillations, whether odd or even, determines the subsequent polarity. These field reversals or oscillations of the geomagnetic field perhaps can explain the observations shown in figures 16 and 17.

The above features presumably correspond to the behavior of the geomagnetic field during a polarity transition. It is important to point out that this particular polarity transition (Olduvai onset R-N) may have characteristic features quite different from the other transitions reported in the literature so far; namely, two oscillations of the field prior to the reversal that mask the pronounced decrease in intensity seen in other reversals. At the same time a gradual decrease in inclination occurs before the reversal while the intensity drops. Feature number five described earlier as part of the characteristics of the Olduvai termination (N-R) transition is not clearly seen; except for the fact that high inclinations (presumably representing the nondipolar part of the field); after the reversal of the field (see
Figure 16. Stratigraphic plot of inclinations, declinations and intensities of core K76113. Demagnetization level at 10 mT.
Figure 17. Stratigraphic plot of inclinations, declinations and intensities of core K78019. Demagnetization level 10 mT.
CORE NUMBER: K78019

DEMEGNETIZATION INTENSITY: 1000 OERSTEDS

LOG. INTENSITY

DEMEG INCLINATION (DEGREES)

DEMEG DECLINATION (DEGREES)

DEMEG LOG. INTENSITY (A. U.)

7500 yr

DEPTH FROM TOP OF CORE IN CM
Figures 15 and 16) which supports the Williams and Fuller (1981b) hypothesis for low latitude records which is the case (core K78113 is from 2.66° N and core K78019 from 8.95° N).

Shown in Table 2 are the statistical results of the Normal and Reversed intervals as well as a statistical mean of the two polarities. The mean declination of the two polarities are nearly antipodal (180° apart ± 5°) in the two cores indicating that a stable magnetization has been isolated, these results are similar to those obtained during the study of the termination of the Olduvai event discussed previously. As observed during the termination of the Olduvai event the statistical mean corresponding to the reversed and normal polarities of the two cores showed a close similarity to that of the ambient field at the site from which the cores were obtained; the ambient field inclination is about the same (+/- 3°) as that predicted by a geocentric axial dipole field model. This agreement between the observed inclination and that given by the dipolar formula can be taken as evidence for the lack of an inclination error, especially since the data are averaged over a temporal and spatial extent that is satisfactory. In this case (Olduvai onset R-N) the minimum time available at the sites is 16,000 years approximately (one polarity) and the maximum, (both polarities) is approximately 32,000 years depending upon the respective sedimentation rates of the cores, and these estimates are sufficiently long to represent a time-averaged geomagnetic field. The agreement between observed and predicted mean directions also means that the characteristic magnetization is primarily due to a PDRM process. The
<table>
<thead>
<tr>
<th>Lat. Long.</th>
<th>P. Long</th>
<th>K78019</th>
<th>N and R</th>
<th>Reversed</th>
<th>Normal</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lat. Long.</td>
<td></td>
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</tr>
<tr>
<td>Lat.</td>
<td>Long.</td>
<td>N</td>
<td>D</td>
<td>I</td>
<td>PI</td>
</tr>
<tr>
<td>K78019</td>
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<td>170.3</td>
<td>189</td>
<td>1.4</td>
<td>16.2</td>
</tr>
<tr>
<td>N and R</td>
<td>8.95</td>
<td>170.3</td>
<td>151</td>
<td>-15.8</td>
<td>17.5</td>
</tr>
<tr>
<td>Reversed</td>
<td>8.95</td>
<td>170.3</td>
<td>38</td>
<td>9.1</td>
<td>17.4</td>
</tr>
<tr>
<td>Normal</td>
<td>2.7</td>
<td>178.7</td>
<td>125</td>
<td>5.3</td>
<td>9.4</td>
</tr>
<tr>
<td>K76113</td>
<td>2.7</td>
<td>178.7</td>
<td>77</td>
<td>190</td>
<td>-5.1</td>
</tr>
<tr>
<td>N and R</td>
<td>2.7</td>
<td>178.7</td>
<td>48</td>
<td>357</td>
<td>16.2</td>
</tr>
</tbody>
</table>

N, number of samples; D, declination in degrees, east of north; I, inclination in degrees, positive downward; PI, paleoinclination, in degrees, positive downward; K, Fisher's precision parameter; α 95, semiangle of cone of 95% confidence for mean directions; and (dp, dm), oval of 95% confidence about the pole position.
normal interval of core K76113 (Table 2) where the recorded inclination value is 10° higher than that predicted by the geocentric axial dipole field model may be an exception. One explanation could be that there is an inclination error, but the sign of the error is wrong which means that the primary magnetization is probably still due to a PDRM process like the rest of the intervals. The higher inclinations may be evidence that the reversal process is still present, namely high inclinations (up to 90°) during the polarity transition especially at low latitudes as predicted by the zonal harmonic model of reversal transition fields (Williams and Fuller 1981b). This latter explanation seems a plausible one due to the fact that the dipole paleolatitude (2.1°) is in close agreement with the geocentric axial dipole field model as reported by Epp et al. 1983. On the basis of statistics shown in Table 2 it can be concluded that the lack of inclination error indicates that, for the normal and reversed polarities at least the behavior of the earth's magnetic field was recorded accurately.
DETAILS OF THE POLARITY TRANSITIONS AT THE OLDUVAI ONSET.

A detailed stratigraphic plot of the magnetic elements of Core K76113 and K78019 is shown in figures 16 and 17. It can be seen that extremely well defined intermediate directions of magnetization are present in both records. Core K78019 appears to have recorded transitional directions in about 8.6 cm from 13725 mm to 13639 mm. The average sedimentation rate for this particular polarity transition (Olduvai R-N) is about 12 mm/1000 years, so an estimate based solely on the directional change, yields an approximate duration of 7200 years. The two oscillations of the geomagnetic field that took place before the reversal and are located at 13850 mm to 13845 mm, with an approximate duration of 400 years, and at 13793 mm to 13777 mm with an approximate duration of 1400 years. Based on the intensity record and taking into consideration the combination effect produced by the two oscillations of the field on the intensity record, it spans a duration of about 3500 years. It is difficult to correlate this intensity record with the intensity periods proposed by Herrero and Helsley (1983) due to the combined effect mentioned above on the intensity record, especially period one and two; the nondipolar dominance or period 3 reaches its minimum at 13664 mm, representing approximately 15% of the average intensity after the dipolar field is completely recovered.
Core K76113 appears to have recorded intermediate directions of magnetization in about 5cm of sediment (from 9060mm to 9110mm) where a 180° directional change occurs. Based on the sedimentation rate of 0.6cm/1000 years, this transition took about 8300 years. The two oscillations of the field prior to the 180° directional change are located at 9245mm to 9232mm, and 9180mm to 9166mm; representing respectively 2200 years and 2300 years of duration of oscillation of the geomagnetic field. The intensity record shows the same characteristics as the record from core K78019, namely an intensity reduction that is a combined effect of the reversal and the proximity of the two oscillations of the field prior to the 180° reversal of Period 3 (Herrero-Bervera and Helsley 1983). An estimate of the duration of the total decrease of the intensity (end of Period 1) during the reversal yields a maximum of 50000 years. The intensity decrease reaches its maximum between 9068mm and 9110mm; and representing only 13% of the intensity measured after field recovery.

This nearly equatorial record (2.7° N) is characterized by a tendency for high positive inclination intermediate field vectors, which gives a characteristic inclination record. In this particular record the maximum inclination values reached by the intermediate field vectors range from 55° to 61°. The site at 9° N (core K76019) also shows highly positive inclinations, up to 58° before commencing a gradual shallowing (See figure 17).
These important features of the geomagnetic field during a polarity transition also have been observed by Freed (1977) in data from marine sediments from the east equatorial Pacific and by Clement et al (1982). It is important to mention that these other records were obtained from the last reversal, (Matuyama-Brunhes). One should also note that the inclination data from all these records (including the Olduvai R-N and N-R for this study) agree with that predicted for an equatorial site by Williams and Fuller (1981) in that they are characterized by highly positive inclinations (up to 80°).
INTENSITY WITHIN THE TRANSITION ZONE.

The intensity of cleaned magnetization within the zone of intermediate directions is approximately 15 per cent that in the overlying zone of normal magnetization and in the underlying zone of reversed magnetization of the two cores (Figures 16 and 17). As previously discussed for the Olduvai termination, a decrease in the intensity of magnetization can be due to three factors. In these two cores, factor two can be ruled out, and in order to test possibility three, ARM measurements were made on all samples.

The ARM was induced in each specimen from each horizon using peak alternating field (AF) demagnetization at 70mT on a DC field of 0.5 Gauss. Figures 18 and 19 show the response to the induced ARM, and, as the case of the records of the termination of the Olduvai event, they display a pronounced, uniform, flat behavior across the transition zones as well as the normal and reversed intervals. Since ARM provides a measure of the magnetizability of sediments, it can be used to crudely normalize remanence intensity measurements to yield a number that is more nearly proportioned to ancient variations in the field intensity. Figures 18 and 19 show the induced ARM to be uniform for the entire section, indicating that the low intensity of magnetization in the transition zones are probably not due to mineralogical changes. As mentioned earlier similar results have been reported by other workers in transition zones e.g. Hilhouse and Cox (1976), Liddicoat (1981), Herrero-Bervera et al (1982 and 1983), and Theyer et al 1982.
Figure 18. Stratigraphic plot of VGP latitude and anhysteretic remanent magnetization (ARM) of core K76113. Demagnetization level 10 mT.
Figure 19. Stratigraphic plot of VGP latitude and anhysteretic remanent magnetization (ARM) of core K78019. Demagnetization level 10mT.
K78019
18.0 MILLI-TESLA

VGP
LATITUDE

ARM
0.5GAUSS
70 mT
Our understanding of the mechanisms for the acquisition of the natural remanent magnetization (NRM) in sediments is still incomplete, compared to, for example, our knowledge of thermoremanent magnetization (TRM) of basalts. In particular, there is at present no paleointensity technique, analogous to the Thellier method for TRM-bearing rocks (Thellier and Thellier, 1959), for relating the magnetization intensity of a sedimentary rock to the absolute intensity of the geomagnetic field in which the remanence was fixed. Levi and Banerjee (1976), and more recently by King et al (1983), argue that the best one can do is to obtain relative paleointensities by normalizing the observed intensity respect to some independent magnetic property. This will compensate for depth variations in the magnetic mineral content and remanence potential of the sedimentary section. Several attempts have been made to obtain relative geomagnetic paleointensities from sediments. Johnson et al (1975) used isothermal remanence (IRM, \( H = 2000 \) Oe) as the normalizing parameter to plot NRM/IRM versus time, Nakajima and Kawai (1973) used NRM/SIRM (saturation IRM) for a similar purpose. Nesbitt (1966) and Harrison (1966) used initial susceptibility to normalize the NRM intensities. Opdyke et al (1973) compared the NRM fluctuations with variations in IRM, anhysteretic remanent magnetization (ARM), and initial susceptibility (\( \chi \)) to argue that the geomagnetic field intensity decreases during a field reversal. Levi and Banerjee (1976) found that for sediments from Lake Saint Croix the anhysteretic remanent magnetization (ARM) was very similar to the NRM and that ARM was therefore a property that could be used to normalize
the NRM variation. Levi and Banerjee's relative paleointensity data showed the same general features as obtained from the archeomagnetic studies. More recently a similar approach using the ratio of detrital remanent magnetization to anhysteretic remanent magnetization (DRM/ARM), for sediment samples proved that evaluating the DRM/ARM ratio as a measure of relative paleointensity is direct comparison with absolute paleointensity data (King et al 1983). A comparison between LeBoeuf Lake sediments estimates and Thellier-Thellier results from the western United States supports the conclusion that suitable sediments are "uniform" with respect to DRM/ARM ratio if:

1). They contain magnetite (Fe3O4) or a mineral of similar composition (in this study it has been shown that in each transitional record the dominant magnetic mineral is magnetite or perhaps a titanomagnetite with a very small amount of titanium);

2). They were magnetized by a uniform mechanism to produce the original natural detrital remanent magnetization;

3). They are of appropriate uniform magnetic particle size (size in the range of 1-15 um which can record geomagnetic paleointensity fluctuations, King et al (1983)).

Figures 20 and 21 show the normalized NRM 100/ARM 700 behavior of cores K76113 and K78019 respectively and demonstrate once more that a substantial decrease of the NRM 100/ARM 700 ratio is coincident and exactly identical in shape with the cleaned record of the intensity of both records. This indicates that the decrease in intensity is indeed a reflection of the decrease in the intensity of the earth's magnetic
Figure 20. Stratigraphic plot of intensities, anhysteretic remanent magnetization and normalized intensity (10 mT) with respect to ARM (70 mT) of core K76113, corresponding to the Olduvai onset polarity transition.
Figure 21. Stratigraphic plot of intensities, anhysteretic remanent magnetization and normalized intensity (10 mT) with respect to ARM (70 mT) of core K78019, corresponding to the Olduvai onset polarity transition.
field during a polarity transition. Figures 22, 23 and 24 show the same behavior described previously during the termination of the Olduvai event.
Figure 22. Stratigraphic plot of intensities, anhysteretic remanent magnetization and normalized intensity (10 mT) with respect to ARM (70 mT) of core K7501, corresponding to the Olduvai termination polarity transition.
Figure 23. Stratigraphic plot of intensities, anhysteretic remanent magnetization and normalized intensity (10 mT) with respect to ARM (70 mT) of core K78030, corresponding to the Olduvai termination polarity transition.
Figure 24. Stratigraphic plot of intensities, anhysteretic remanent magnetization and normalized intensity (10 mT) with respect to ARM (70 mT) of core K78019, corresponding to the Olduvai termination polarity transition.
LOG

K78-10-19

J  ARM  J/ARM

TRANSITION

OLDUVAI TERMINATION
CHANGES IN VIRTUAL GEOMAGNETIC POLE POSITION

As reported by Herrero-Bervera and Helsley (1983) (see Appendix A) a convenient presentation of the directional changes during a transition can be made by use of the Virtual Geomagnetic Pole (VGP) paths for transitional data sets. This is particularly true if comparisons between different records are to be attempted. Thus, VGP paths were constructed from the data of cores K7501, K78030, K78019 and K76113 and these paths are shown in Figures 25, 26, 27 and 28.

The VGP path from core K7501 certainly falls in the "A" category of Fuller et al. (1979); the reversal is well defined and at least 23 truly intermediate virtual geomagnetic poles have been recorded in the more than 50 successive VGP's. The path appears to be longitudinally constrained between 260° to 300° E.

The older half of the transition is characterized by two loops, one small loop that crosses the equator at 300° E; and another broad loop that crosses the equator at 270° E. For this N-R reversal the VGP's representing the transitional field seem to recover very slowly in comparison with the rapid changes in the VGP's representing the field right after the reversal's onset (See Figure 25). This VGP behavior is similar to the VGP's behavior of the Lower Triassic Chugwater reversal reported by Herrero-Bervera and Helsley (1983) (Appendix A).
Figure 25. Path of virtual geomagnetic poles for the normal to reversed termination of the Olduvai polarity transition of core K7501.
K7501

OLDUVAI TERMINATION

N→R
Figure 26. Path of virtual geomagnetic poles for the normal to reversed termination of the Olduvai polarity transition of core K78030.
K78O30
OLDUVAI TERMINATION
N→R
Figure 27. Path of virtual geomagnetic poles for the normal to reversed termination of the Olduvai polarity transition of core K78019.
K78O19
OLDUVAI TERMINATION
N→R
In core K78030 the virtual geomagnetic poles representing the transitional behavior of the geomagnetic field fall into the "B" category of Fuller et al (1979) classification. The VGP path is longitudinally confined between 200° E and 260° E and is characterized by a big loop that crosses the equator between 200° E and is defined by six intermediate VGP positions (See Figure 26); the majority of these intermediate VGP positions are located in the Southern Hemisphere.

The transition record obtained from the nearly equatorial (9° N) core K78019 (Figure 28); is defined by 24 intermediate VGP positions that fall in the "B" category of Fuller et al (1979) classification and only one of the conditions for such classification is not satisfied (>1 VGP <60° in final Hemisphere) meaning that in the final hemisphere there are no VGP positions between 0° and 60° S. The path is longitudinally confined between 280° E and 320° E. The VGP positions are mainly located in the northern Hemisphere and the small loops do not cross the equator.

VGP paths from core K76113 can be classified in the "A" category of Fuller et al (1979); the reversal is well defined and at least eight truly intermediate virtual geomagnetic poles have been recorded in the more than 26 successive VGP's. The path is definitely longitudinally constrained between 200° E and 260° E. The older half of the transition is characterized by two loops; one big loop that crossed the equator between 220° and 240° E, a small loop located at 80° and 90° N, and between 200° to 260° E that does not cross the equator. For this R-N reversal the VGP's representing the transitional
field seems to linger between $45^\circ$ S and $35^\circ$ N before it changes rapidly to the normal polarity. (See Figure 28).

In core K78019 the virtual geomagnetic poles representing the transitional behavior of the geomagnetic field fall definitely in the "A" category of Fuller et al (1979) classification. The VGP path is longitudinally confined between $240^\circ$ E to $280^\circ$ E and is characterized by several loops that cross the equator between $260^\circ$ E to $280^\circ$ E and is defined by sixteen intermediate VGP positions (See Figure 28).
Figure 28. Path of virtual geomagnetic poles for the reversed to normal onset of the Olduvai event polarity transition of cores K76113 and K78019.
ONSET OF OLDUVAI K76113
R→N

ONSET OF OLDUVAI K78019
R←N
IS THE TRANSITIONAL FIELD DIPOLAR?.

In order to test the hypothesis that the transitional field is dipolar the transitional VGP's recorded in the Central North Pacific were compared with VGP's obtained from the study of the Olduvai Termination reversal in the Indian Ocean (Opdyke et al 1973). If the field was dipolar during the transition, the paths of the virtual poles determined from the Pacific Ocean and the Indian Ocean would be similar.

As described by Opdyke et al (1973). "The movement of the VGP for the upper Olduvai polarity change (Figure 6) is straightforward, going from a position centered on the north pole south through the Atlantic Ocean sector to the south polar regions." The Indian Ocean VGP path and the VGP paths constructed from the Pacific cores used in this study do not overlap, being entirely dissimilar. Thus one can conclude that the geomagnetic field was not dipolar during the upper Olduvai polarity transition. The same conclusion was reached by Hillhouse and Cox (1976) when they compared the pole paths of the same reversal (Matuyama-Brunhes) from two distant localities, Lake Tecopa California and the Boso Peninsula in Japan.

This test has been done with the Olduvai termination VGP paths due to the availability of records published by Opdyke (et al 1973) and Hammond (1979). For the Olduvai Onset (R-N) there are no records available from other oceans or land sections to compare with, except for a record published by Hammond et al 1979. This record was from core K7501 (37.3° N); and shows only one transitional VGP located at
0.4° N and 75° E (See Hammond et al 1979, Figure 3C) in the middle of the Indian Ocean this record due to the lack of sufficient intermediate VGP's does not fit either one of the A or B categories proposed by Fuller et al (1979). Nevertheless, if a comparison is attempted with the records of this study they are entirely dissimilar. K76113 and K78019 cores show paths lying on the American continent, and not even one single VGP close to the Indian Ocean. If this comparison is valid, then the conclusion is the same as the case of the Olduvai termination polarity transition, namely the transitional field was not dipolar. Thus, all recently published transitional data sets (Fuller et al (1979), Clement et al (1982), Herrero-Bervera et al (1982 a,b), Herrero-Bervera and Helsley (1983), and Theyer et al (1982)) indicate that the field is not dipolar during a transition. Even if one rejects Hammond et al's one data point and analyzes just the VGP paths of the Olduvai onset (R-N) transition (cores K76113 and K78019), it is obvious that the paths do not coincide even though they are not far apart (6° in latitude and 8.5° in longitude, see Figure 28). This, independently leads to the conclusion that the field during the Olduvai onset (R-N) reversal was not dipolar.
CHARACTERISTICS OF THE NONDIPOLAR BEHAVIOR OF THE FIELD DURING THE ONSET AND TERMINATION POLARITY TRANSITIONS OF THE OLDUVAI EVENT.

Before a comparison of these data sets is attempted in order either to support or reject and ultimately perhaps to constrain the current models that attempt to explain the transitional behavior of the earth's magnetic field an assessment of the transitional data in terms of comparisons between and within sites has to be made. The transitional data sets are discussed below in chronological order, thus the onset of the Olduvai event is analyzed first.

The behavior of the inclination data depicted in figures 16 and 17 during the polarity transition, shows that the inclination records are characterized by extremely high inclinations that entirely disagree with the expected inclination values at the site latitudes where the cores were collected. For core K76113 the expected inclination is $6^\circ$ and $17.5^\circ$ for core K78019. The observed inclinations reached maximum values of $55^\circ$ to $61^\circ$ for core K76113 and up to $70^\circ$ for core K78019. In these two cases the inclination patterns go from negative inclinations to positives ones, except for the last stretch of the inclination pattern during the transition of core K78019.

A similar observation can be made for the N-R transition records of the Olduvai termination. Cores K7501, K78030 and K78019, shown in figures 10, 11 and 12 respectively, are characterized by inclination patterns that show a clear change from positive values to negative ones reaching high inclination values incompatible with the expected inclination values at the latitudes where the cores were
collected (see Table 1). This is true particularly for cores K78030 and K78019 where inclination values up to 65° and 30° respectively can be observed (see for example figures 11 and 12). In this respect, there seems to be a gradual decrease in maximum inclination values from sites from equatorial latitudes high latitudes sites. For instance core K76113 located at almost equatorial latitudes during the transition recorded a high inclination value of 61° while a near zero value might have been expected and core K7501, located at 37° N, also recorded within the transition zone values up to 60° which are not too much different from the expected inclination values at that particular site latitude. Thus, one can conclude that the high inclinations recorded at the low latitude sites are a distinctive characteristic behavior of the earth's magnetic field during a polarity transition.

The directional changes in all records seems to have recorded a definite 180° directional change. For the case of the two records of the onset of the Olduvai polarity transition the duration in time is very similar (see Tables 3 and 4), despite of their different sedimentation rates, 8000 years for core K76113 and 7500 years for core K78019. These two records do not show a systematic succession of points but rather a tortuous path. This is particularly well defined in the declination record of core K78019 in its beginning part of the polarity transition (see Figure 17). This observation is extremely important due to the fact that the most recently published transition records of lavas, show these distinctive fluctuations in the declination records, see for example Coe et al 1983.
Another very important way to analyze the declination and inclination data in order to compare the behavior of the transitional directional data within and between sites is by converting the declination and inclination into virtual geomagnetic poles. Figures 25, 26, 27 and 28 show that all the VGP paths are sharply constrained in longitude. At least two of the three transitional paths corresponding to the termination of the Olduvai event are found in the hemisphere centered about the meridian antipodal to the site is that from K7501 which is located about 90° to 120° off the site longitude.

The VGP path of core K78019 also is definitely found on the hemisphere centered about the meridian antipodal to the site. It can be said that the location of the path in terms of longitudinal confinement is very similar to the path of core K7501 (figure 25). The path is not as complete as the path of core K7501 but gives very useful and important information nevertheless. The two paths lie on the Americas and both sets of VGP's are in the hemisphere centered antipodal about their site longitude. This result is important because very few results have been reported so far of the same reversal sampled at more than one locality. (Clement et al 1982, Theyer et al 1982, Herrero-Bervera and Helsley 1983). The results reported by Herrero-Bervera and Helsley (1983) give the same consistent answer, namely, three identical VGP paths for a N-R polarity transition of Triassic age, the sites were located from tens of meters apart, up to 110 km apart. The other results recently published are from the Matuyama-Brunhes polarity transition recorded in the Pacific Ocean at
three different localities. These Matuyama-Brunhes records yielded the same answer in that all three sites are near sided, but two of the paths do not exactly coincide as might be expected considering their close geographic proximity (approximately one degree in longitude apart and two degrees apart in latitude, Clement et al 1982). In this study, of the Olduvai Termination two of the VGP paths (K7501 and K78019) are longitudinally coincident, yet the sampling sites are located 10° apart in longitude and approximately 28° apart in latitude.

The results reported here give some indication, that the detailed record during the transition may not be simply related to the geomagnetic field. For example the VGP paths for cores K7501 and K78019 versus K78030 do not exactly coincide as might be expected considering their longitudinal proximity. Similar results were reported for the Matuyama-Brunhes transition by Clement et al (1982). The dissimilarities in VGP paths from nearby cores can be explained either as noise in the record or as a result of incomplete records. (Clement et al 1982). In the case of core K78030 the lack of progressive change in direction within the transition zone may be indicative of an incomplete record (see Figure 11). However this record may have resulted from unresolved multicomponent magnetizations or may be an indication that in this particular section of the core, the sediments have not accurately recorded a geomagnetic field of apparent low intensity.
These observations indicate that for core K78030, and to some extent for core K78019, one must reconsider the possible influence of sedimentological factors such as the effects of burrowing organisms, variation in remanence lock-in depth and small hiatuses which can distort the magnetization record. Since these problems seem to be present in the low sedimentation rate cores, investigation of higher deposition rate cores where the scale of sedimentary disturbances may encompass shorter time intervals, will be necessary to fully assess the reliability of observations of transitional field behavior in sedimentary records.

The VGP path of core K78030 differs significantly when compared to the other two records of the termination of the Olduvai transition. The VGP positions reside almost exclusively in the hemisphere centered upon the site meridian. This path is characterized by a big loop and rapid fluctuations of the presumably transitional behavior of the geomagnetic field. This record is definitely due to the slow sedimentation rate, and not much can be said in terms of transitional behavior of the geomagnetic field.

Despite the uncertainty presented by core K78030, and bearing in mind that most previous records of transitional field have been similarly obtained from sediments (all of the other records are from the Matuyama-Brunhes polarity transition), some general comments regarding the correlation within sites for the termination of the Olduvai event can be made based on the data presented here.
The three transitional VGP paths presented here are each characterized by a portion in which the VGP position moves from high northerly latitudes to high southerly latitudes along a longitudinally confined path, lying more than 90° in longitude away from the site (except for K78030 which is 30° away). Thus the paths tend to be located in the hemisphere centered about the meridian antipodal to the site, but not markedly so. In addition each path contains one or more loops which either precede the longitudinal path (K7501) or follow it (K78030 and K78019). The loops, which are present in all of these paths and are both clockwise and anticlockwise, show a movement of the VGP's from high latitudes to very low latitudes followed by a return to high latitudes.

The demagnetized data show that the VGP paths calculated for the same reversal (Olduvai R-N) at two different sites shown in Figure 28 are very sharply constrained in longitude. The R-N path of core K75113 crosses the equator at about 50° to 80° off the site longitude. This pattern is definitely in the hemisphere centered about the site longitude. This record could be called near sided.

The VGP path of core K78019 crosses the equator between 70° and 90° off the site longitude. This path also is found in the hemisphere centered upon the site meridian; and the path should be called near sided, although this path is a border line case due to the crossing of the site longitude.
These results reported here for the Olduvai onset (R-N) polarity transition; give the same kind of indication as the results of the Olduvai termination (N-R) transition, as well as the results reported by Clement et al (1982) for the Matuyama-Brunhes transition, namely, that the detailed record during this transition may not be simply related to the geomagnetic field. For example the VGP paths for cores K76113 and K78019 do not exactly coincide as might be expected considering their longitudinal proximity (only 8.5° in longitude apart). These dissimilarities, as pointed out by Clement et al (1982) in VGP paths from nearby cores can be explained either as noise in the record or as a result of incomplete records. In the two cases of cores K76113 and K78019 there is a lack of progressive variation in direction within the transition zone indicating that the records are complex. Nevertheless they appear to have recorded accurately the geomagnetic field of apparent low intensity. If one compares the completeness of records from cores K78019 and K76113 with the records of core K78030, then one concludes that the influence of sedimentological factors, such as the effects of burrowing organisms, variation in remanence lock-in depth and small hiatuses which can distort the magnetization record are minimal and should not be considered to be important drawbacks for these particular two records of the onset of the Olduvai polarity transition. Thus, since most of the factors of concern do not seem to be present and based on the excellent site correlation of two oscillations of the geomagnetic field as well as the intermediate directions, one can reliably interpret the two records as a reflection
of the geomagnetic field during a polarity transition.

The two VGP paths presented here are strictly plots of the intermediate VGP's associated with the Olduvai onset (R-N) transition and the VGP's corresponding to the two oscillations of the field are not included in figure 28. The two paths are characterized by a portion in which the VGP position moves from 20 to 80 degrees of southerly latitudes along longitudinally confined paths lying not more than 90° longitude away from the site. The VGP path of core K76113 crosses the equator at approximately 50° away from the sampling site; and not more than 90° from the sampling site; defining a narrow band of about 40° width (see Figure 28). The VGP path of core K78019; is even more longitudinally confined; crossing the equator between 70° and 90° away from the sampling site, defining an even narrower band of about 20° width. These two paths tend to be near-sided, although the path from core K78019 could be called a border line case due to the crossing of the equator at about 90° away from the sampling site. In conclusion the VGP paths constructed for this reversal are longitudinally constrained to a certain extent; and the intermediate VGP's are roughly centered 60° (K76113) and 80° (K78019) away from their respective site longitude. The paths are therefore indubitably near-sided.

It should be pointed out, however, that even though all the transitions can be ascerted to belonging to either a near sided or far sided transition, it can be said even more strikingly that all the transitional paths lie between 45° and 135° from the sampling site. None are truly antipodal or pass near the site itself. Moreover, all
the transitions observed in this study are confined to one quadrant. This remarkable consistency and narrowly confined longitudinally band to which both the onset and termination transitions are confined is the most striking feature of this data set.

One of the most important observations that can be made using the whole data set derived during this study of the entire Olduvai event (R-N-R) is to compare the records of the Olduvai event boundaries sampled from the same core. The record for K78019 (R-N-R) is shown in figures 27 and 28b. The field that controlled these back-to-back or successive transition was definitely nondipolar; the VGP paths do not overlap, even though that the two Subchrons are separated by a time gap of only 160,000 years according to Berggren et al (1980). This is an expected result that could be used as confirmation of the comparison made with the already reported results of the same Olduvai reversals, Opdyke et al (1973) and Hammond et al (1979), that yielded evidence for nondipolar dominance of the geomagnetic field during those two reversals. At the same time this comparison is extremely important because only two other back-to-back reversals from the same locality have been reported, Bogue and Coe (1982) and Valet and Laj (1981). The VGP paths bounding the Olduvai event are characterized by a definite longitudinal confinement of the VGP's but they do not follow an exactly orderly trajectory during their transit from one polarity to another as can be seen in figures 27 and 28b. The back and forth jumps observed in these cores are similar to those observed in the Lower Triassic Chugwater Fm. transition reported by Herrero-Bervera and Helsley.
(1983). Despite this "jumpiness" the Olduvai transitions display a remarkable coincidence of the VGP paths as well as a confined trajectory of the transitional VGP's from a normal to a reversed polarity.

These observations lead to conclude that for the Olduvai event (R-N-R) successive transitions the field that controlled the reversals is definitely nondipolar, but the nondipolar geometries that dominated such transitions were entirely different.

The next step is to compare the transitional paths between sites; and this analysis is at some extent even simpler due to the dissimilarity of all five VGP paths, for instance the two paths that recorded the onset of the Olduvai event are not coincident and again, like in the case of the two successive reversals of core K78019 are entirely dissimilar.

Figure 28 shows clearly the behavior of the transitional VGP's indicating that even between sites of the same reversal, assuming that the transitional directions do represent indeed the behavior of the field during a polarity transition; the nondipolar geometry that controlled the onset of the Olduvai event had to be different at two nearby sites. The same conclusion applies to the transition during the termination of the Olduvai event. The divergent behavior of the two most complete and reliable records, K7501 and K78019, lead to the conclusion that when two successive reversals are compared and when paths paths between sites are compared, either from the same Subchron or from a different one, the transitional fields that dominated the two
Olduvai reversal were of different nature. Moreover the components of the nondipolar field that dominated the transitional field at the latitude of the sampling site of core K7501 were different from the components that dominated the transitional field at the sampling site of core K78019. The same observation applies to the field geometries that dominated the onset of the Olduvai event.

It is valid to say at this point that not one single component of the nondipolar field dominated in time and space the two reversals of the Olduvai event. This is the reason why no one single model or even a combination of model would give us a unique solution to describe the transitional core processes. Perhaps the data are telling us that no unique model is appropriate and that each reversal is representative of some random combination of several effects.

So far, this analysis has been performed utilizing directional data and their respective VGP's. But still another no less important parameter has to be included in the whole analysis of these data sets and this is the intensity of the two reversals of the Olduvai event.

One convenient way to study the intensity is by analyzing the normalized intensity as a function of VGP latitude. This means of analysis has been used to study the Matuyama-Brunhes transition obtained from dry lake sediments in California (Hillhouse and Cox 1976) and provided the only data at the time that was reasonably acceptable for the comparison of observed relative paleointensity and any transitional model available at the time. It is argued by Hillhouse and Cox (1976) that since (1) these sediments contain a good record
(i.e. not obviously smoothed) of paleosecular variation above the transition zone, and (2) acquisition of anhysteretic remanent magnetization (ARM) is generally constant over the entire stratigraphic section, that the observed intensity variation of cleaned natural remanent magnetization (NRM) must reflect geomagnetic behavior.

The relative paleointensity behavior of core K78019 (R-N-R) and core K7501 can be compared using this technique. The comparison of these two relative paleointensity plots of the Olduvai event is given in figure 29. The most relevant feature of this analysis is the lowest relative paleointensity reached by the two reversals. In the case of the onset of the Olduvai only one point located approximately at -15° of the VGP latitude reached a value of about 15 per cent, the rest of the low relative paleointensity points are located in the 20 to 25 per cent vicinity. In the case of the points representing the termination of the Olduvai event the lowest values of the relative paleointensity range from 15 to about 25 per cent.

In conclusion, it can be said that at a site latitude of 9° N the lowest values of the relative paleointensity points range from 15 to 25 per cent of the total intensity, with a median of 20 per cent. At 37° N the lowest values of the normalized intensity are of the order of 22 per cent of the total intensity, and the great majority of points are located in the 30 per cent vicinity.
Figure 29. Normalized remanence intensity versus VGP latitude.

Corresponding to the termination of the Olduvai event core K7501 and core K78019, as well as the onset of the Olduvai event core K78019.
OLDUVAI TERMINATION

K7501
LAT. 37° N

Normalized Intensity

VGP Latitude

OLDUVAI TERMINATION

K78019
LAT. 9° N

Normalized Intensity

VGP Latitude

OLDUVAI ONSET

K78019

Normalized Intensity

VGP Latitude
Table 3

Estimated duration of the transition, length of transition of the termination of the Olduvai event of cores K7501, K78030 and K78019

<table>
<thead>
<tr>
<th>Core</th>
<th>Position</th>
<th>Sedimentation Rate (cm/Ka)</th>
<th>Time Represented by each Sample (yr)</th>
<th>Length of Transition (cm)</th>
<th>Estimated Duration of Transition (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>K7501</td>
<td>179.30 W</td>
<td>1.6</td>
<td>600</td>
<td>16</td>
<td>10000 (Dir)</td>
</tr>
<tr>
<td></td>
<td>37.30 N</td>
<td></td>
<td></td>
<td>19</td>
<td>12000 (Int)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>27</td>
<td>16500 (Int)</td>
</tr>
<tr>
<td>K78030</td>
<td>160.30 W</td>
<td>0.9</td>
<td>1100</td>
<td>2.4</td>
<td>2600 (Dir)</td>
</tr>
<tr>
<td></td>
<td>18.90 N</td>
<td></td>
<td></td>
<td>7.4</td>
<td>8200 (Int)</td>
</tr>
<tr>
<td>K78019</td>
<td>170.30 W</td>
<td>1.5</td>
<td>650</td>
<td>9.6</td>
<td>6400 (Dir)</td>
</tr>
<tr>
<td></td>
<td>8.95 N</td>
<td></td>
<td></td>
<td>14</td>
<td>9300 (Int)</td>
</tr>
</tbody>
</table>
Figure 29, shows that no systematic relative paleointensity low values are found between sites (e.g. K7501 and K78019) and even within sites there is a slight difference of these low intensity values (e.g. K78019 onset and termination).

The divergence of the low values of relative paleointensities between and within sites hints that each reversal has its own particular harmonic content. Again, the intensity record suggest that like the VGP paths the recent models are unable to simulate the complicated behavior of the earth's magnetic field during polarity transitions. The available models can at best explain successfully the reversal process of one single reversal (e.g. Matuyama-Brunhes) but they do not seem to be applicable to all reversal processes.

The next and the last step in this data assessment is to estimate the duration of the polarity transitions based on the sedimentation rates and stratigraphic thicknesses. This estimate should be made for both the duration of directional as well as the intensity changes and must check for possible systematics within and between sites. Table 3 shows an estimate of the duration of the directional changes as well as intensity changes, this is based on the different sedimentation rates of each core and the stratigraphic thicknesses of the transition. It is clear that in all cases the duration of the directional changes as a rule for all cores is less, compared with the duration of the intensity changes; this is a common result for all the transitions reported so far. An other important observation is the latitudinal increase of time of duration (if one
Table 4

Estimated duration of the Onset of the Olduvai event polarity transition of cores K76113 and K78019.

<table>
<thead>
<tr>
<th>Core</th>
<th>Position</th>
<th>Sedimentation Rate (cm/Kyr)</th>
<th>Time represented by each sample (yr)</th>
<th>Length of Transition (cm)</th>
<th>Estimated duration of Transition (yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>K76113</td>
<td>178.7 W</td>
<td>0.6</td>
<td>1600</td>
<td></td>
<td>8300 (DIR)</td>
</tr>
<tr>
<td></td>
<td>2.66 N</td>
<td></td>
<td></td>
<td>5</td>
<td>16600 (INT)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>10</td>
<td>40000 (INT)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>24</td>
<td>50000 (INT)</td>
</tr>
<tr>
<td>K78019</td>
<td>170.30 W</td>
<td>1.2</td>
<td>800</td>
<td>8.6</td>
<td>7200 (DIR)</td>
</tr>
<tr>
<td></td>
<td>8.95 N</td>
<td></td>
<td></td>
<td>15</td>
<td>12500 (INT)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>25</td>
<td>20800 (INT)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>35</td>
<td>29200 (INT)</td>
</tr>
</tbody>
</table>
excludes the incomplete record of core K78030) of the reversal from low latitudes to mid latitudes. Nevertheless, these observations in the northern hemisphere definitely indicate that the intensity changes take place over a much longer time interval than do the directional changes. This result is consistent with the results of other transitions from the same hemisphere (Williams and Fuller 1981b).

Assuming a constant sedimentation rate (0.6 cm/1000 years for K76113 and 1.2 cm/1000 years) through the Olduvai subchron; the duration of the transition is given in table 4.

The estimated duration of the transition is 8300 to 7200 years for the directional change for both cores; whereas the associated intensity variation occurs over a 20,000 to 50,000 years period. The extended period of low intensity may be due to the combined decrease in intensity associated with the two oscillations of the field prior to the real Olduvai onset polarity transition. If one excludes the effect of the intensity caused by the oscillations; the estimates of the duration of the low intensity period associated with the Olduvai onset are, 12,500 yr for core K78019 and 16700 years for core K76113, indicating that the intensity low was longer than the directional change in both cores.

In conclusion, according to the estimates of the duration of the reversals for the Olduvai event (R-N-R), there is a consistent observation; the duration of the directional changes took less time than the duration of the intensity changes. This conclusion applies to the onset and termination of the event.
Irving (1964) stated "There should be no systematic differences of chemical, mineralogical or physical properties of the magnetic minerals in positive and negatively polarized rocks".

In order to test the prediction postulated by Irving (1964) thermomagnetic analyses (temperature dependence of strong field induced magnetization) were conducted on an automatically recording translation balance, where magnetic fields up to 13 Koe could be applied. For the Curie temperature measurement a magnetic field of 1.7 Koe was applied to reduce the influence of paramagnetism. Magnetization was measured in the temperature range between 28° C and 700° C the sample environment being air and argon. From the thermomagnetic curves the Curie temperatures were determined by a graphical method following Grommé et al (1969). The uncertainty in the absolute value of magnetization is about 3 per cent.

The J-T experiments were performed using bulk samples, of 12 specimens of four transitions (Cores K7501 Olduvai N-R, K78019 Olduvai R-N-R and K78113 Olduvai R-N) in order to identify the magnetic minerals. The studied samples were taken from the critical transition zones of the above mentioned polarity transitions, in each case the spacing between samples (bottom to top of the transition zones) exceeds 13cm. All the J-T curves fall definitely into one or another of three types of which the first is illustrated in figure 30 (Core K76113). Here the distinguishing characteristics are:(1) In an argon environment, the J-T curves for heating and cooling are nearly or
Figure 30. Thermomagnetic curves for three samples within the transition zone of core K76113 of the onset of the Olduvai event.
OLDUVAI ONSET
K76113

9000 TOP

INTENSITY ARBITRARY UNITS

9049 IN

9127 BASE

100 200 300 400 500 600 T°C

100 200 300 400 500 600 T°C

100 200 300 400 500 600 T°C
exactly identical. (2) The oxide fraction that is magnetic has a Curie temperature between 500° to 535° C.

Typical type 2 J-T curves are shown in figures 31. These are distinguished by: (1) The cooling curves do not exactly retrace the heating curves, when heated in air. (2) The Curie temperature is usually between 550° C to 595° C, these points are illustrated in figure 31 (core K7501).

Samples from core K78019 gave type 3 J-T curves illustrated in figures 32 and 33. In this type, (1) The curves are completely irreversible when heated in air; (2) The Curie temperature seems to be distributed over the interval from 570° C to 590° C.

In most of these curves the Curie temperatures are clearly defined and the portions of the curves immediately below the Curie temperatures are fairly steep, indicating that the oxide mineral has an essentially uniform composition in each specimen. (Figures 31 to 34). Some nonuniformity of composition may exist in specimens from the temperature range 500° to 535° C (Figure 30), but single Curie temperatures may nevertheless be identified without ambiguity, in most of the cases.

The Curie temperatures indicate that the oxide mineralogy resides primarily (500° to 595° C) in Ti-poor titanomagnetites. For this temperature range (500° to 595° C), and using the general titanomagnetite formula \( xFe_2TiO_4(1-x)Fe_3O_4 \) the value of the composition parameter \( x \) would be between 0.10 to 0.01 (Akimoto et al 1957, Kawai 1959, Nagata and Akimoto 1961). This observation has been confirmed by
Figure 31. Thermomagnetic curves for three samples within the transition zone of core K7501 of the termination of the Olduvai event.
Figure 32. Thermomagnetic curves for three samples within the transition zone of core K78019 of the termination of the Olduvai event.
Figure 33. Thermomagnetic curves for three samples within the transition zone of core K78019 of the onset of the Olduvai event.
Figure 34. Thermomagnetic curves for four samples at different stratigraphic levels from core K78030.
additional studies of parallel samples in the transition zones, such as, magnetic granulometry, x-ray diffraction analysis of final magnetic extracts and AF demagnetization versus IRM acquisition curves indicating that for these particular transitions the only phase on the final magnetic extract is magnetite (most of them SD hexagonal or cuboidal in form) of detrital origin, and these analysis confirmed the idea that there is no trace at all of hematite. Supporting this, among other things is the fact that all the pilot samples as well as the rest of the samples were able to be effectively demagnetized by applying 10 mT as shown in figures 2, and 5 to 9. It is a well known fact that with those small fields it is impossible to demagnetize hematite due to its high coercivity.

The most important conclusion about these Curie analyses is the fact that for each transition zone only one and the same mineral phase exists within it, even though such mineral phases may be slightly different between cores due to the latitudinal and longitudinal distribution of the source areas.
TRANSITIONAL FIELD MODELS

In the past six or seven years several workers have attempted to explain the mechanisms and structure of the Earth's dynamo during some of the past polarity transitions of the geomagnetic field. These new efforts were stimulated by new observational data from paleomagnetic records (VGP records). Most of these published records are from the last reversal of the magnetic field and are geographically located in the Northern Hemisphere, (Fuller et al 1979). More recently, older records from normal to reversed transitions as well as from Southern Hemisphere localities have been reported (Coe et al 1983, Clement and Kent 1983, Herrero-Bervera and Helsley 1983, Liddicoat 1982, Valet and Laj 1981, Williams and Fuller 1982a), along with cases of successive reversals at the same site as well as different sites (Bogue and Coe 1982, Herrero-Bervera et al 1983, Theyer and Herrero-Bervera 1983, Valet et al 1983). All these data sets have been used to either provide further constraints on the models or in establishing the validity of comparing different polarity reversal transitions.

The ideal culmination of transitional field studies would be a deliniation of the harmonic content and its variation in time, throughout a reversal. With the present published data this is clearly unattainable. However certain steps towards this end have been taken. A brief summary of the models recently proposed follows.
Following an earlier suggestion (Smith 1967), Hillhouse and Cox (1976) suggested that transitional fields may arise from components that do not take part in the reversal process. That is, during the decay through zero and the subsequent regeneration of the main field, any stationary non-reversing components would control the intermediate geometry. Provided that such standing field remains for the most part invariant for times far longer than the average recurrence time interval associated with the reversal process, such a hypothesis is associated with a clear prediction: that transitional field behavior experienced at any given site locality is expected to be similar during successive reversals. More specifically, intermediate field geometries associated with sequential reversals should be alike and not dependent on the sense of the transition.

Following the kinematic approach to geomagnetic reversals developed by Parker (1969) and Levy (1971, 1972), Hoffman (1977) showed that the available transition data were compatible with a hypothesized reversal process that starts at low latitudes in the core and subsequently extends, or "floods", north and south (axisymmetrically) until the transition is complete. In contrast to the standing field model, this approach suggests that transitional field geometries arise from the configurational characteristics of the reversal process. Thus, during a transition, the field direction experienced at a given site will rotate either through a vertically downward orientation (producing a near-sided VGP path) or through a vertically upward direction (producing a far-sided path) depending on the location of the
site and the sense of transition. Hence, provided that the configurational aspects of the reversal process remain for the most part invariant for times far longer than the average recurrence interval for transitions, at a given site this model predicts the field vector to follow essentially antipodal paths for consecutive reversals.

More recently the flooding model has been extended to include the distinguishable possibility that reversals originate at high latitudes in the core (Hoffman and Fuller 1978), and also generalized to incorporate non-axisymmetric behavior (Hoffman 1979). However, the prediction of antipodal VGP paths for sequential transitions remains unaltered, provided that the reversal process for both transitions start in the same region of the core. This model is called the "flooding approach model" or the "zonal approach model"

In addition to explaining the low field intensities commonly associated with paleomagnetic records of polarity transitions (see e.g. Smith 1967, Dodson et al 1978, Dagley and Lawley 1974), both the "standing field model" (Hillhouse and Cox 1976) and the "zonal flooding model" (Hoffman 1977, 1978 and 1979) predict, that for a given site, the path of the virtual geomagnetic pole (VGP) will lie along a great circle. Such an occurrence, as pointed out by Hoffman (1979) is consistent to first order with several observations (e.g. Creer and Ispir (1970) and Dodson et al 1978). Hoffman (1979) notes however with regard to the determination of path locations, that the two models are distinguishable. For example, the predicted path associated with the standing field model (Hillhouse and Cox, 1976) is found to lie along
the meridian defined by the longitude coordinate of the VGP determined for the stationary component alone. In contrast, the zonal flooding model (Hoffman, 1977, and Hoffman and Fuller 1978) predicts a path which runs either along the site meridian or along the meridian antipodal to the site depending upon (1) the sense of the transition, (2) the geographic hemisphere in which the site is located, and (3) whether octupole or quadrupole zonal components dominate the transitional field (Hoffman, 1979).

Most available paleomagnetic reversal paths display a site-dependence which suggests that transitional fields are indeed controlled by low-order zonal components (Hoffman 1977 and 1978). In particular, nearly all reverse-to-normal (R-N) VGP paths from northern low- and mid-latitude sites appear on the hemisphere centered about the site meridian. This configurational characteristic of R-N transition fields suggests the predominant zonal term to be either quadrupolar (positive $g_0/2$) or octupolar (positive $g_0/3$), components which are consistent with an axisymmetric reversal process that originates in the southern hemisphere or near the equator of the core, respectively (Hoffman 1977, 1978 and 1979).

Alternatively, the standing field model (Hillhouse and Cox, 1976) can also predict behavior compatible with these records provided that the hypothesized non-reversing part of the non-dipole field possesses the appropriate geometry. Such a situation, however, sharply contrasts with that associated with the flooding approach where a polarity transition involves reversal of the entire geomagnetic field.
Although in general agreement with the zonal flooding model, few of the transitional VGP records plotted in Hoffman (1979, Figure 1) are found in close proximity to the site meridian (i.e., the path locality predicted by the model, Hoffman 1977 and 1978). In addition, even the most well-behaved, detailed paths available before 1979 are seen to deviate from that of a great-circle, often displaying a gradual drift in azimuth (e.g., Hillhouse and Cox 1976). In order to account for this non-axisymmetric behavior, a more general flooding model was proposed by Hoffman (1979) which incorporates both azimuthal as well as latitudinal proliferation of reversed magnetic field throughout the core from a point of origin.

The non-axisymmetric flooding approach to the geomagnetic reversal process is found to successfully simulate transitional field behavior as revealed in available paleomagnetic records. In contrast to the alternative interpretation of the data in terms of standing field components having appropriate assumed geometries, Hoffman's (1979) non-axisymmetric flooding approach, (1) interprets the observed behavior to the time-dependent configurational characteristics of the reversing geodynamo, (2) involves the reversal of the total field, and (3) is critically testable whenever records corresponding to a given transition are available from more than one site locality (Hoffman 1979).
According to Hoffman (1979), the proposed model suggests that a geomagnetic reversal may have its origin at a discrete locality in the core. VGP behavior corresponding to four Cenozoic R-N transitions recorded in the western United States suggests that there may exist a small number of discrete sites in the core at which reversals originate, (Hoffman, 1979). Assuming the existence of such sites, Hoffman (1979) speculated as to whether they are associated with fluctuations in cyclonic activity (Parker (1969), Levy (1971, 1972)), instabilities in angular momentum (Robbins 1976), dramatic changes in convective motion (Busse 1975), or topographic features in the core-mantle interface.

The model proposed by Williams and Fuller (1981b) is called the zonal harmonic model. The model assumes that during a reversal it is unlikely that only a single zonal harmonic will be present during the transition. This problem was simplified by using only terms of $n \leq 4$ and assuming an exponential decay of the dipole term as suggested by the decrease in intensity seen in most of the reversal records published so far. According to the authors the results of this procedure were surprisingly illuminating. The form of the record, in terms of variation of inclination and intensity in time, was found to be highly dependent upon site latitude. The most dramatic effect of this latitudinal variation is seen at equatorial sites where, for axisymmetric fields the inclination must pass through $+90^\circ$ (equivalent to near sided VGP paths or $-90^\circ$ far sided). Furthermore these results suggested that any estimate of the duration of the transition would
vary with site latitude as would the relation between the intensity decrease and inclination change. This could lead to confusion in the definition of the time duration of the reversal.

The above mentioned ideas are based on theoretical grounds, e.g. Parker (1969), Levy (1971, 1972) as well as observational data. The energy redistribution of the non-dipole field was based on the results of Verosub and Cox (1971) which showed that for the last 120 years, 76% of the energy lost by the dipole term has been gained by the non-dipole with terms of degree n=2,3,4 receiving 72% of that energy.

The Williams and Fuller model can predict the inclination and intensity record that would be observed at any site latitude provided sufficient detail is present to constrain the various parameters involved.

The published data (paleomagnetic records) have been used to test these variations of the geomagnetic reversal models. In the review by Fuller et al (1979) 15 records were found to satisfy requirements to be considered category A. However, with regard to a total understanding of the geomagnetic reversal process, this data set was found to be deficient in two important ways. First, the geographical distribution of sites at which the recordings were obtained is restricted. More specifically, all category A records are associated with sites located in the Northern Hemisphere. Moreover, apart from three Icelandic records and two records from a site in the east-equatorial Pacific, all were obtained from mid-latitudes. Second, reversal records associated with the transition sense from normal to reverse (N-R) are poorly represented.
Up until today the need for additional records has been clear, especially N-R records obtained from the Northern Hemisphere and records of either transition sense from the Southern Hemisphere (at the present only one record from the Southern Hemisphere has been published, the Lower Jaramillo polarity transition by Clement and Kent 1983). Nevertheless, observed paleomagnetic behavior associated with the relative abundance of R-N recordings from low and mid-northern latitudes listed by Fuller et al (1979) suggest that transitional fields are controlled by axisymmetric non-dipole term(s), Hoffman (1982). The data, when viewed as paths of the virtual geomagnetic pole (VGP), have been found to display a dramatic dependence on the locality of the recording site consistent with this hypothesis (Hoffman 1977; Hoffman and Fuller 1978; and Fuller et al 1979). Specifically, transitional paths of the VGP associated with R-N transitions are found to reside almost exclusively on the hemisphere centered about the site longitude, the so-called "near-side". Similarly, VGP paths corresponding to N-R transition records from low and mid-latitude sites tend to be found on the "far side". Notwithstanding the relatively small number of transitions represented by available data, such an overall dependance of the geometry of reversing fields on both the locality of the site of observation and the sense of the transition, if confirmed, would indicate that (1) intermediate fields associated with polarity transitions are significantly axisymmetric (Hoffman, 1982), (2) configurational characteristics of the reversal process in the core are not dependent on the transition sense, and (3) the entire field
takes part in the reversal (e.g. stationary, non-reversing field components, if present, do not significantly affect the geometry of transitional fields). Moreover, it has been pointed out (Hoffman and Fuller 1978) that given the recognition of site-dependent characteristics associated with records obtained from the Southern Hemisphere, the geometry of the controlling axisymmetric term (whether quadrupole or octupole), as well as particular configurational systematics of the reversal process (e.g. the region(s) in the core where reversals originate) would, according to Hoffman (1982), be identifiable. Implicitly, the question of whether the dominating zonal geometry is similar for all reversals would be answerable as well.

As pointed out by Hoffman (1982), the question now is: in what way, if any, do the recently acquired transition records reported since the 1979 review alter the claim of a general site-dependence? Interestingly, these newly available data include (1) three sets of sequential transitions— one pair from marine sediments of Miocene age, and two different R-N reversals recorded at the same site from Crete (Valet and Laj 1981, Valet et al 1983) and the other pair from basalts of Pliocene age from Kauai (Bogue and Coe 1982), (2) a low latitude R-N recording from an intrusion of Miocene age from the Philippines (Williams and Fuller 1981a), (3) a record corresponding to the Gauss-Matuyama (N-R) transition obtained from sediments in California, and (4) a N-R data set of several sites of the same reversal in the Lower Triassic Chugwater Formation (Herrero-Bervera and Helsley 1983, Appendix A). In addition we now have the transitions reported here for
the Olduvai onset and termination at several localities.

Of particular interest with regard to the possible site dependence of R-N transition data is the finding of an unambiguous far-sided VGP path associated with the newly acquired record from the Philippines (Williams and Fuller 1981a). Such an observation is exceptional to this particular subset of transition data from low and mid-northern latitude sites.

There is also a possibility that the N-R transition recorded at Kauai (Bogue and Coe 1982) produced a near-sided VGP path at that site. However, the data lack low-latitude VGP's and cannot be classified as category A according to Hoffman (1982).

Another interesting set of results is the Chugwater Formation polarity transition reported by Herrero-Bervera and Helsley (1983, Appendix A). For a N-R transition according to the "flooding approach model" prediction, the VGP paths would have to be far-sided where the data set shows that the VGP's are neither far-nor near-sided paths but simply at the boundary of the two regions.

In general, the overall effect of including newly acquired paleomagnetic transition records with those previously available is to make somewhat less convincing any claim of a general site dependence. Indeed, as Williams and Fuller (1981b) point out, there is no guarantee that each reversal has the same harmonic content. This idea is strongly supported by the results published by Clement and Kent (1983) indicating that their transition record strongly suggest that the Lower Jaramillo transitional field was dominated by different harmonics than
the Matuyama/Brunhes transitional field. Consideration of the enlarged data set now suggests that such an overall characteristic, if it exists at all, may be time-dependent or may be determinable only in a statistical manner.

On the other hand, this conclusion does not alter the claim of an axisymmetric site dependence of VGP paths corresponding to particular transition fields. This appears to be the case for the Matuyama-Brunhes (R-N) transition. The axisymmetric control of this transition field is clearly seen in Figure 1 of Hoffman (1982) where representative data from each of five site localities are plotted with respect to a common site longitude.

Such a figure is of particular interest due to the observation that the distribution of VGP's about the site longitude shows far less dispersion during the initial stage of the transition (Southern Hemisphere VGP's) than those associated with the final stage (Northern Hemisphere VGP's). This feature suggests that the field is more strongly controlled by zonal components during the onset of the reversal process. Such an observation has been pointed out previously (Fuller et al 1979). More recently Hide (1981) came to the same conclusion on purely theoretical grounds, suggesting that axisymmetric terms dominate transition fields primarily during the decay phase of the main field.
The important point to be made here is that, in general, the standing field model and the flooding models predict field behavior associated with sequential reversals that theoretically is easily distinguishable. Fortunately, there are sets of transition data suitable for such a test (see for instance discussion by Bogue and Coe 1982, Valet and Laj 1981, Valet et al 1983).

In this regard, Valet and Laj (1981) present transition data associated with sequential reversals from Miocene sediments in Crete for which the VGP's are plotted with respect to site longitude (Figure 35a). As can be seen, the VGP behavior associated with the prior (R-N) transition is what might be termed borderline near-far to the west of the recording site. In contrast, the later transition from N-R (this path although detailed, is not rigorously classifiable as category A) is clearly far-sided with VGP's residing almost exclusively to the east of the site. Moreover, the final equatorial crossings by the VGP for these sequential records are seen to be nearly antipodal. Valet and Laj (1981) conclude from this observation that the generalized flooding model of the geomagnetic reversal process is supported. More recently Valet et al (1983) reported on two different R-N geomagnetic reversals with identical VGP paths recorded at the same site, reaching the conclusion that (excluding the possibility of an accidental coincidence of identical VGP paths) the basic mechanisms in the geodynamo leading to a reversal might persist unchanged over a long period of time. Moreover, the authors conclude, that the two VGP paths are largely constrained in longitude along a mean great circle located
approximately 80° W of the site, so that the transitional field cannot be correctly described in terms of pure axial symmetry.

Recently Bogue and Coe (1982) presented an analysis of consecutive reversal data from the early Pliocene basalts from Kauai for which the VGP's are shown (Figure 35b). Bogue and Coe suggest that both reversal paths are similar and near-sided. However, the paucity of intermediate VGP's, especially with regard to the N-R transition, leaves this interpretation in some doubt. Bogue and Coe suggest that these VGP data are more easily explained by the standing field model, but note that the flooding approach can also simulate such directional behavior provided that the R-N transition is assumed to start in the Southern Hemisphere of the core while the N-R transition is assumed to start in the Northern Hemisphere. Furthermore, they argue that the variation in paleointensity with VGP latitude is more consistent with the flooding model (Bogue and Coe 1981).

In addition to the above studies, the Gauss-Matuyama (N-R) polarity transition has recently been studied in sediments from Searles Valley, Calif. (Liddicoat, 1982). Following Liddicoat, these VGP data are reproduced in Figure 35c together with those associated with the Matuyama-Brunhes (R-N) transition recorded at nearby Lake Tecopa (Hillhouse and Cox 1976). Although not successive reversals (they differ in age by some 1.7 m.y.), the similarity of the transition fields is striking. Bogue and Coe (1982) point out that these data, together with intermediate paleomagnetic behavior revealed in Matuyama epoch volcanics from nearby Clear Lake (Mankinen et al 1980),
support more convincingly the long-term standing field hypothesis. On the other hand, the most recently published record of the Lower Jaramillo polarity transition from a Southern Hemisphere, deep-sea core (Clement and Kent 1983) can be considered together with records of the most recent reversal (Matuyama-Brunhes). Clement and Kent (1983) strongly suggests that the Lower Jaramillo transitional field was dominated by different harmonics than the Matuyama-Brunhes transitional field.

Thus available records of reversal pairs do not yet furnish sufficient evidence to eliminate either approach from consideration. The acquisition of additional records of sequential transitions would be most helpful, thus, one of the aims of this study is to increase the number of reliable records of polarity transitions of the geomagnetic field and particularly of those Subchrons that have not yet been studied such as the case of the Olduvai event (N-R-N) and compare the results in order to: (1) find possible similar characteristics of the short-term behavior of the field within and between sites of the same reversal in this case the back-to-back reversals of the Olduvai event. (2) a test of the "standing field model", "flooding approach model", and "zonal harmonic model". (3) the possible discrimination among these models as well as the setting up of the highlights for a different model, if necessary.
Figure 35. Virtual geomagnetic pole paths, plotted with respect to site longitude, corresponding to reversal pairs: (a) sequential Miocene transitions from Crete (Valet and Laj, 1981); (b) sequential Pliocene transitions from Kauai (Bogue and Coe, 1982); (c) Gauss-Matuyama (Liddicoat, 1982) and Matuyama-Brunhes (Hillhouse and Cox 1976). Taken from Hoffman (1982).
At this point several questions about geomagnetic reversal, can be asked:

1) Does the morphology of transitional fields associated with available records of Cenozoic reversals display consistent, recognizable systematics, as has been claimed (e.g. Hoffman 1977)?

2) Do transitional fields arise primarily from time-dependent configurational characteristics associated with the reversal process in the core, or does a major contribution arise from a standing (i.e. non-reversing) portion of the field?

3) What dominant field geometries are associated with particular polarity transitions?

In order to shed light on these aspects of the geomagnetic reversal process a brief summary of the behavior of the most studied polarity transition, the Matuyama-Brunhes (R-N), will be presented before the discussion of the Olduvai event reversals.
FIELD GEOMETRIES DURING SELECTED REVERSALS

Testable Models

According to Hoffman (1982), a refined description of the transitional field corresponding to a particular reversal will require multiple recordings obtained from distant sites. At present there exist two testable quantitative models that may be applied to such data sets. First the generalized flooding approach (Hoffman 1979) involves a phenomenological simulation of the reversal process in the core. As mentioned before the model utilizes a Busse-type geometry (Busse 1975) for the magnetic source region and simulates a reversal through the movement of magnetic poles about the surface of this region from a point of initiation. Both north-south as well as east-west flooding of reversed flux is involved. Furthermore, Hoffman (1981b) has shown that the variation of the dominant transitional field components can be calculated directly. Hence, when successful, this model provides a quantitative description of the variation of the most important axisymmetric as well as non-axisymmetric non-dipole terms present in the model solution. From this description specific predictions can be made about the behavior of the transitional vector field (e.g. VGP paths, plots of intensity against VGP latitude) for a given site.

Williams and Fuller (1981b) have taken a different approach to the problem, providing a model that quantifies the zonal harmonic content throughout a reversal. Rather than simulating the path of the virtual geomagnetic pole, Williams and Fuller analyse the inclination records associated with a given reversal. They argue that since
axisymmetric terms appear to dominate transitional fields, the observed time-dependent features in inclination at a given site can be regarded as a signature of the zonal geometries present. Such a claim is clearly demonstrated by the authors through the production of synthetic inclination records based upon a redistribution of energy assuming an exponential decay of the dipole field into $g_0$, $g_0^2$, and $g_0^3$ terms.
The Matuyama-Brunhes (R-N) transition.

There currently exist no less than ten paleomagnetic records corresponding to the Matuyama-Brunhes transition associated with five distinct site localities in the Northern Hemisphere. However, owing to the total lack of data from the Southern Hemisphere, some ambiguity must exist with regard to distinguishing the axisymmetric harmonic content during the transition. For this reason, the parameters associated with the general flooding model for this reversal were only loosely fitted (Hoffman 1979, 1981b). In particular, by placing the starting point for this transition along the equator of the core, a zonal octupolar \( (g_0^3) \) geometry was assumed with no quadrupole \( (g_0^2) \) contribution. Transition paths predicted by the modelled solution are reproduced in Figure 36, together with representative VGP data obtained from each site locality. To a first order, the simulation appears to be successful. Moreover, the predicted intensity behavior associated with this modelled solution reflects well that observed in the single, credible intensity record currently available for this transition (see Hoffman, 1979 for further details).

It should be pointed out that a simulation that fits as well as that shown in figure 37 may be generated by placing the initiation point for this reversal at low latitudes within the Southern Hemisphere of the core. Therefore a significant contribution to the axisymmetric geometry by \( g_0^2 \) cannot be ruled out, but can only be quantified with the attainment of records from sites at southern latitudes. (Hoffman 1982.)
Finally, regardless of the latitudinal placement of the starting point for the reversal, the dominant non-axisymmetric geometry associated with the model is that of a \( g_2^1, h_2^1 \) quadrupole. This result is consistent with the comparison by Liddicoat (1982) of VGP paths associated with two records of the Gauss-Matuyama (N-R) transition, one from California and the other from Turkmenia. The sites have nearly the same latitude while separated in longitude by approximately 180°. Interestingly, the two paths are seen to traverse the equator nearly 180° apart and, when plotted with respect to a common site longitude, appear quite similar. Such a result is most easily explained by the presence of a non-axial quadrupole term in the transition field.

Williams and Fuller (1981b) point out that the form of the inclination records associated with equatorial sites, in particular, places strong constraints on the possible distribution of non-dipole zonal harmonics present in the transition field. Finding a good fit for the Matuyama-Brunhes data to involve a redistribution of dipole energy to \( g_2^0, g_3^0 \) and \( g_4^0 \) components in the ratio 2:3:5, Williams and Fuller produced synthetic inclination logs that simulate very well the time-dependent behavior experienced at most site localities. More specifically, this solution is associated with a dominant quadrupole \( g_2^0 \) whose effect is tempered in the Northern Hemisphere by a \( g_3^0 \) contribution of opposite sign and to a lesser extent the \( g_4^0 \) component. An interesting prediction for transitional field behavior at mid-southern latitudes is a minimal variation in intensity.
Figure 36. Virtual geomagnetic pole paths corresponding to the Matuyama–Brunhes (R-N) transition: (a) observed, (b) predicted by the generalized flooding model. From Hoffman (1981b).
throughout the reversal (Williams and Fuller 1981b). Figure 35b shows this simulation for the case of the low-latitude records from the east-equatorial Pacific as well as for a mid-latitude record from eastern Europe. Application of the same model to the high-resolution R-N Tertiary transition record from Mt. Rainier (Dodson et al. 1978) renders an excellent fit in both inclination as well as relative paleointensity (Williams and Fuller 1981b). A test to this model has been attempted using core oriented data in time and space of the Matuyama-Brunhes, Jaramillo and Olduvai events with an excellent fit in both inclination and paleointensity (Theyer et al. 1984).
Figure 37. Virtual geomagnetic pole paths, plotted with respect to site longitude, corresponding to the Olduvai termination transition, core K7501 and Olduvai onset, core K76113.
OLDUVAI TERMINATION

K7501
SITE

OLDUVAI ONSET

K76113
SITE

R→N

N→R
DISSCUSION

In a strict sense the Olduvai event comprises two reversals, a reversed to normal (R-N), located at 1.88 m.y. and a normal to reversed (N-R) at 1.72 m.y. in the geomagnetic time scale of Berggren et al (1980). Based on the paleomagnetic characteristics such as univectorial behavior of the samples within the transition zones, uniform and single phase magnetic mineralogy in the critical transition zones, and a definite lack of inclination error in the normal and reversed intervals just above and below the transition zones, the records that are entirely reliable for a transitional geomagnetic interpretation are from core K76113 (R-N) and core K7501 (N-R). These records at the same time show VGP data definitely as categorized according to Fuller et al (1979).

The records from core K78019, even though they cannot be disregarded for the interpretation of the behavior of the geomagnetic field during polarity transitions, seem to fail at least one of the rigorous rock-magnetic tests imposed on the transitional paleomagnetic records. For instance K78019 N-R records show a possible inclination error in the normal interval; and the VGP path falls in the B category of Fuller et al. 1979; in addition, the record from core K78030 is an incomplete record of the transitional behavior of the field of that particular Subchron; even though it satisfies the rigorous rock-magnetic tests imposed on the above mentioned records. The only record left is from core K78019 R-N transition, which shows features of the geomagnetic field susceptible to correlation with the other record.
of the same Subchron (Olduvai R-N, record K7113). This record satisfies the rock-magnetic tests imposed on the other records, (and is classified in the A category of Fuller et al. 1979), except that the pilot sample in the transition zone displays a not too convincing univectorial behavior.

With the evaluation of the data, one can reliably attempt to answer the questions posed previously, find the possible systematics and test the current models of transitional fields using the transition records of the Olduvai event (R-N-R) reversals.

One of the most obvious characteristics of the records is the conspicuous difference in terms of the behavior of the geomagnetic field before the transition as shown by the directional data (Figures 10 to 12, 16 and 17). The records of the onset of the Olduvai show two characteristic oscillations of the field right before the transitional directional change; these are unique characteristics of this particular reversal (Olduvai R-N) and there is no other similar report published so far. The only other transition with solely one oscillation of the field prior to the reversal is the Gauss-Matuyama transition reported by Liddicoat (1982). At this point it is important to mention that Yoshimura (1980) noted that the number of such oscillations, whether odd or even determines the subsequent polarity.

When the directional data are converted to VGP's and plotted with respect to a common site longitude, the VGP paths expected according to Hoffman 1977, Hoffman and Fuller 1978 and Fuller et al 1979; would be a near-sided path for a R-N reversal and a far-sided
path for a Northern Hemisphere reversals. As shown in figure 37 the R-N path (Olduvai Onset) shows the site dependence characteristic of the R-N reversals, the most typical example of this behavior of the field is the Matuyama-Brunhes polarity transition, which in this case yields a near-sided path. In the case of the termination of the Olduvai event (N-R) the corresponding VGP path renders a far-sided path, although the majority of VGP's are found to reside on the antipodal hemisphere about the site meridian on the far side, at least 25 per cent of the VGP's are located approximately 90° from the site longitude. When this VGP path is compared with other VGP paths of back to back reversals, for instance with the transition reported by Valet and Laj (1981) one can conclude that the two data sets look different, (see figures 35a and 37). The two figures show a pair of sequential reversals, where the Valet and Laj reversal paths display the final equatorial crossing by the VGP's for these sequential records a nearly antipodal behavior. This probably suggests that particularly the Olduvai N-R transition is not exclusively controlled by an axisymmetric transitional field dominated by low order zonal spherical harmonics. It is possible that these transitions are controlled by non-axisymmetric fields, and this idea is probably supported by the VGP paths of the same Subchron. The Olduvai onset (R-N), cores K78019 and K78030, figures 26 and 27, and Olduvai termination (N-R) core K78019 figure 28, provide paths of N-R transitions that are not exactly coincident even if they are converted for site latitudes differences and they are not far-sided as predicted by the flooding approach model.
For the case of the Olduvai R-N transition analyzed from the behavior of the VGP path of core K78019, it could be considered as a borderline near sided case. This is also supported by data obtained from the same core (K78019) of the two sequential reversals of the Jaramillo event (Herrero-Bervera et al 1983b).

In conclusion; the data presented here suggest an apparent \( g_3^0 \) (octupolar) dominance for the R-N Olduvai transition, due to the near-sideness of the VGP paths; even though that the paths do not overlap suggesting the dominance of non-axisymmetric terms. In particular the quadrupole component controls the non-axisymmetric characteristics of the transition field and so is responsible for the observed deviation in a given VGP path from the site longitude during such transition. The N-R Olduvai transition, even though the two paths are far-sided, shows characteristics of a borderline case (particularly core K78019 and at least 25 per cent of the VGP’s of core K76113) and the paths, like the paths of the Olduvai R-N, do not overlap suggesting in this case (Olduvai N-R) that the field during this polarity transition was dominated by not only \( g_2^0 \) (quadrupolar) fields but possibly by a dominant non-axisymmetric geometry associated with the flooding approach model such as of a \( (g_{\frac{1}{2}}, h_{\frac{1}{2}}) \) quadrupole field.

The relative paleointensity (PDRM100/ARM700) studies performed on all the samples of all the sites showed that: (1) these sediments contain a good record (i.e., not obviously smoothed) of paleosecular variation above and below the transition zones, and (2) acquisition of anhysteretic remanent magnetization is invariably constant over the
entire stratigraphic sections. Thus, the observed intensity variation of cleaned natural remanent magnetization (NRM) must reflect geomagnetic behavior (Hillhouse and Cox 1976, Hoffman 1979). With this important conclusion in mind, not only directional data converted to VGP’s can be handled in order to analyze the geomagnetic behavior of transitional fields, but reliable intensity data show that the decrease in intensity is indeed that of the decrease of the intensity of the earth’s magnetic field. One way to do it is by normalizing the intensity and plotting it versus VGP latitude. This aspect of the transitional field is illustrated in figure 29, where in all cases there is a dramatic decrease in intensity particularly at high VGP latitudes.

The relative paleointensity values reach a minimum of about 15 to 25 per cent of the full polarity field strength; with few individual samples giving minimum field strength of 10 per cent (i.e. in core K78030 the intensity drops to 15 per cent and for K76113 is only 13 per cent). When the modelled Matuyama-Brunhes field was analyzed in terms of field strength derived from intensity studies so as to reveal the predicted world wide distribution in minimum field strength experienced during the such transition, the results showed that for equatorial latitudes (0° up to 10°) the minimum value is less than 5 per cent and for the rest of the Pacific Ocean in the Northern Hemisphere the minimum would be between 5 to 10 per cent. Although highly speculative according to Hoffman (1981) the findings illustrate the effect of both non-dipole constituents of the modelled Matuyama-Brunhes transitional
field. If purely axisymmetric, the minimum intensity would be solely a function of site latitude. However, since the transition field (at least for the Matuyama-Brunhes and Olduvai event transitions) contains a strong non-axisymmetric term as well, some longitudinal dependence is also expected and indeed is seen that for the area of the Kamchatka peninsula and the sea of Okhostsk the values are less than 5 per cent; and for the area of central Atlantic Ocean at latitudes between 30 to 60° for both hemispheres, the values are greater than 15 per cent of the full polarity strength, (Hoffman 1981).

It must be pointed out again according to Hoffman (1981) that the modelled solution (for the Matuyama-Brunhes transition) assumes that the reversal starts at the equator of the core, producing a zonal transition field which is predominantly octupolar. This octupole term is responsible for the very weak predicted intensities on the Earth's surface near the equator when compared to some mid-latitude sites. In contrast, a modelled transition field containing a dominant zonal quadrupole term, through an assumed reversal process which starts deep in the Southern Hemisphere of the core, would have produced transitional field intensities at the equator that are relatively strong compared to those experienced at mid-latitudes. Because of the present indeterminacy of the $g_0^2 / g_0^3$ ratio, discussion of the nature of the intensity variation during this field reversal (Matuyama-Brunhes) must be restricted to mid-northern latitudes. At these sites Hoffman (1981) found that the minimum transitional field strength averages about 10 per cent of the full polarity field
(consistent with some observations, see Fuller et al 1979) but to vary by more than a factor of three depending upon site longitude. More specifically, for a particular mid-latitude, the minimum transitional field strength is seen to have the largest value at the longitude corresponding to that in the core where the reversal process is assumed to begin and to have the smallest value 180° in longitude away from that point.

For the case of the Olduvai event transitions (R-N-R) again the average field strength is greater than 10 per cent (for the Matuyama-Brunhes in the North central Pacific, the field is less than 10 percent) and showing the highest value of approximately 20 to 25 percent at a latitude of 37° corresponding to the Olduvai termination (core K7501). This indicates that for the Olduvai event transitions the field dominating was non-dipolar, indeed, but a combination of a quadrupolar, octupolar and hexadecapolar geometries, with their respective non-axisymmetric terms included in each case of the axisymmetric geometries.

Analysis of the modelled solution to the reversing geodynamo for the Matuyama-Brunhes polarity change (from Hoffman 1979 and 1981) renders a low-order quantitative description of the transition field. More specifically, available paleomagnetic records associated with this reversal are roughly simulated by a field geometry consisting of as few as two non-dipole terms (a zonal and a non-axisymmetric quadrupole) at the time of zero dipole contribution.
Since the reversal process is assumed to start at the equator of the core, the results from the analysis performed by Hoffman (1981) must be considered to be non-unique.

Thus, at this point the first question mentioned above, namely, does the morphology of transitional fields associated with available records of Cenozoic reversals display consistent, recognizable systematics, as has been claimed (e.g. by Hoffman 1977) can be at least partially answered. If one compares the modelled Matuyama-Brunhes solution with the Olduvai data, particularly with the Olduvai R-N transition, one can say that the Olduvai records do not display consistent, and recognizable systematics as has been claimed by Hoffman 1977; this is true for other Cenozoic transitions such as, the Lower Jaramillo transition studied by Clement and Kent 1984; and the Kauai transition studied by Bogue and Coe 1982, and Bogue 1983, as well as the Steens Mountain reversal studied by Coe et al 1983.

Obviously the complexity and variety of paleomagnetic reversal records are difficult to reconcile with simple standing field, zonal flooding and frozen frozen-flux models. The paleomagnetic data force one to conclude that transitional core processes, if similar to those implied by the simple models, must vary considerably between reversals. It may thus be misleading to base interpretations on the systematics of the data from more than a single reversal.
CONCLUSIONS

Analysis of the data collected with the primary objective of the study of the short-term behavior of the Earth's magnetic field indicates that the use of a high resolution sampling technique, particularly for deep-sea sediments, can be a powerful technique that is able to reveal the time-averaged behavior of transitional fields.

The records obtained with this technique proved to be highly reliable and demonstrated that the samples within the transition zones display a univectorial behavior. The specimens studied have a uniform mineralogy single-phase mineralogy within each one of the transition zones and a lack of inclination errors in the intervals outside the critical transition zones. This establishes that a post-depositional remanent magnetization mechanism was operating and that this mechanism should yield reliable records suitable for transitional field studies.

Of the five transitional paleomagnetic records studied three display a high degree of reliability (K7501 (N-R), K78019 (R-N), and K76113 (R-N)). These records are category "A" according to Fuller et al 1979 and the other two fell in category "B". As a rule all the N-R records are the near-sided, whereas the R-N records are either a border line case (K78019) or a far-sided case. The N-R records are consistent with the Hoffman (1977) and Hoffman and Fuller (1978) flooding approach model, and the distribution of VGP paths associated with this N-R reversal are roughly simulated by a field geometry consisting of as few as two non-dipole terms, a zonal octupole and a non-axisymmetric quadrupole at the time of zero dipole contribution. For the case of
the N-R VGP paths they do to not fully support the flooding approach model; since the paths are not truly far sided. This suggests that the field even though non-dipolar cannot be roughly simulated by a field geometry consisting of less than two terms, one a zonal quadrupole and one a nonaxisymmetric octupole term at the time of zero dipolar contribution. This transition is rather unique and does not show similarities when compared with other reversals published so far for it is characterized by two oscillations of the field immediately before the 180° directional change of the field. On the other hand relative paleointensity studies of all the samples when plotted as normalized intensity versus VGP Latitude, show that the minimum field strength does not support either the simple two non-dipolar dominance of the transitional field, or a zonal octupole and a non-axisymmetric quadrupole or a zonal quadrupole and non-axisymmetric octupole, but rather a combination of quadrupole, octupole, and hexadecapolar geometries including their respective non-axisymmetric components of the nondipolar field. Thus, the complexity and variety of paleomagnetic reversal records are difficult to reconcile with the currently available simple models, (standing field, zonal flooding and frozen flux), forcing one to conclude that transitional core processes, if similar to those implied by simple models, must vary considerably between reversals. Because of this difficulty and more generally, the nonuniqueness of any reversal model, it is important to consider what additional factors can be learned with data sets reversals, specially those that can be observed and correlated accurately over large distances as is the case with marine deep-sea sediments records.
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APPENDIX.
Paleomagnetism of a Polarity Transition in the Lower (?) Triassic Chugwater Formation, Wyoming

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A total of 156 oriented specimens, taken at stratigraphic intervals of 3 to 20 cm, have been collected from the Red Peak Member of the Chugwater Formation at three localities approximately 25 km southeast of Dubois, Wyoming. The stratigraphic interval containing the polarity transition in the remanent magnetic field is about 1.0 m. All specimens were thermally demagnetized at 500°, 550°, 620°, and 660°C and show directions of magnetization of normal, reversed, and transitional character as well as a remarkable decrease of the remanent intensity in the transition zone. The section sampled is divided into five intensity periods: two periods recording the dipolar field, one recording a dipolar field of decreasing intensity, one recording a dipolar field of increasing intensity, and a period of low intensity recording the field present during the 180° directional swing. A virtual geomagnetic pole (VGP) path obtained for a N → R polarity transition makes a fairly smooth transit, with the path confined to a sector of longitude between 85°W and 110°E. The reversal's onset is first indicated by a slow movement of the VGP across northern latitudes. The VGP then lingers to two positions 60°N and 40°N (35°E to 70°E) and 20°N and 40°N (335°E to 5°E), suggesting brief periods of field stability (or nondipolar field dominance). The onset of the transition appears to be more rapid than the recovery. The VGP path observed at a northern hemisphere site is located about 90° from the site. The pole position derived from stable normal and reversed portions of this study (115.5°E, 46.4°N, $d_m = 4.1, dp = 2.2, dm = 4.2$) lies approximately in the center of the main group of poles reported for other Triassic units.

INTRODUCTION

One of the most interesting features of the earth's magnetic field is that worldwide simultaneous polarity changes have occurred. These polarity changes provide the most compelling evidence that the observations of reversed magnetization in rocks is related to reversals of the earth's magnetic field. Other evidence comes from measurements showing that the polarity changed continuously from one polarity to the other in a rock sequence, a polarity change or polarity transition [McElhinny, 1973]. Directions of magnetization observed in rocks during such changes are called intermediate or sometimes transitional and have been reported by several workers. They are not common, however, indicating that the duration of a transition is short [Goldstein et al., 1969].

Most early estimates of the duration of the transition tended to give times of the order of $10^3$ or $10^4$ years. However, much longer transition times have been suggested [Momose, 1963]. Cox and Dalrymple [1967] estimated statistically that the time required for a reversal is 5000 years or less. Oceanic cores have yielded independent estimates of 4000 years [Harrison and Somayajulu, 1966] and 4600 years [Opdyke et al., 1973], and oriented cores reported more recently by Hammond et al. [1979] suggested directional changes of approximately 6000 years for the Jaramillo and Olduvai events. The Boso Peninsula records give 4700 years [Niiyama, 1971]. The common element linking these phenomena is that they occur very quickly, so quickly in fact that it is difficult to find rocks which have recorded them completely enough and accurately enough to provide useful magnetic records [Cox et al., 1975].

As pointed out by Cox et al. [1975], a difficulty encountered in much of this research is that as magnetic recorders, many rocks have a poor high-frequency response. Volcanic rocks, for example, are erupted too sporadically to provide a record with a needed degree of continuity. Sedimentary rocks also contain many stratigraphic gaps corresponding to times of nondeposition or erosion. Moreover, an initially complete and accurate paleomagnetic record in sediments may be obliterated or (what is worse) altered by chemical diagenesis and bioturbation occurring after initial deposition [Løvlie, 1976; Kent and Lowrie, 1974]. The magnetic recorders on the earth have undergone erasure and overprinting and in general have generated records with numerous gaps. Furthermore, the paleomagnetic record from any single sediment core or geologic section is generally far from being complete or perfect. As a result, our picture of the fine structure of the ancient magnetic field is still hazy.

The present work was undertaken in spite of the above enumerated difficulties to determine whether or not the paleomagnetism of red sediments could provide additional insight into the nature of the geomagnetic field during periods of transition between normal and reversed polarity.

GEOLoGY

The Chugwater Formation (Triassic) is well exposed in central Wyoming where it is a thick, conspicuous sequence of red beds exposed around the flanks of most uplifts. Its regional stratigraphy and history are discussed by Love [1939], McKee et al. [1959], Picard [1967], and more recently by High and Picard [1969]. The nomenclature of the Chugwater Formation proposed by Love [1939] and followed by McKee et al. [1959] is used in this paper along with the modifications proposed by High and Picard [1969]. From oldest to youngest, Love's units of the Chugwater Formation are: Red Peak Member, Alcova Limestone Member, Crow Mountain Sandstone Member, and Popo Agie Member. Picard [1964] subdivides the Red Peak into five informal rock units; from oldest to youngest these are: silty claystone...
facies, lower platy facies, alternating facies, upper platy facies, and variegated facies.

The section sampled for these studies is in the alternating facies of the Red Peak Member located in west central Wyoming (Figure 1) southeast of the town of Dubois. Bedding within the Chugwater Formation dips at 9° to 15° in the region of study, and lithologic correlation between adjacent outcrops can be made easily. In the area of study, the Chugwater Formation overlies the Phosphoria Formation (Permian) and in the same area is overlain by the Nugget Sandstone (Jurassic?). Thus the age of the Chugwater Formation of this region is established as Triassic [High and Picard, 1969], although the formation itself contains few fossils. Despite the lack of abundant or diagnostic fossils, detailed stratigraphic relations have been established in large areas by means of physical correlations. On the basis of regional correlation with other Permian and Triassic strata, High and Picard [1969] conclude that the lowermost member of the Chugwater Formation may be of Early Triassic age. The uppermost units of the Chugwater Formation are con-

Fig. 1. Location map showing the sampling locality (solid circles) within the Chugwater Formation. Pp, Permian Phosphoria Formation; Tru, Triassic Chugwater Formation; Ju, Jurassic rocks, undivided (includes Nugget sandstone). Geology is from Love et al. [1955].
sidered by the same authors to be as young as Late Triassic. Thus one can only say with fair certainty that the rocks studied are of Early Triassic age. No samples were taken from the portion of the section considered to be possibly of Middle or Upper Triassic corresponding to the Crow Mountain Member and the Popo Agie Member, respectively. All samples were collected below the Alcova Limestone, which separates the Red Peak Member and the Crow Mountain Member [Love, 1957; High and Picard, 1967, 1969; Pipirinos, 1968].

Sampling Procedure

Oriented samples were collected approximately 20 km southeast of Dubois, Wyoming (Figure 1). All samples were collected by means of a hand-held gasoline-powered drill [Helsley, 1967b] and were oriented by means of a magnetic compass.

The initial search for the reversal was done at a locality known as 'Red Grade,' 109°26.6'W, 43°24.8'N, on the south side of 'Little Red Creek' at elevations between 1975 and 1987 m where samples at widely spaced stratigraphic levels were drilled in order to identify the intervals of normal, reversed, and intermediate polarity. Succeeding samples at closer intervals focused on one reversal that occurred between two locally identifiable stratigraphic marker horizons (in this paper referred to as the upper clay seam and lower clay seam) that allowed stratigraphic correlations between sampling localities. Once the possible transition zone was stratigraphically identified, a total of 113 samples (approximately 156 specimens) were drilled at three localities: profile 1; profile 2, located 90 m west of profile 1; and profile 3, located 25 m west of profile 2 in the Red Peak Member of the Chugwater Formation. This sampling covered a stratigraphic interval of approximately 3.5 m of fine-grained red sediments and was drilled at 3- to 20-cm spacing between samples. The samples were drilled with inclinations ranging from 8° to 82° with respect to horizontal and were up to 25 cm in length. This procedure allows a more or less continuous sampling and in some cases provides an independent observation of the same stratigraphic level of the reversal. Since the Red Peak Member is about 200 to 300 m thick in west central Wyoming [Picard, 1964], the detailed results presented below cover only about one fifteenth of the total Chugwater section.

Samples collected during the initial search for reversals (of the order of 58 samples) were cut to 2.5-cm (1 inch) cylinders and measured in the field with a portable air turbine spinner magnetometer [Helsley, 1967a]. After the stratigraphic identification of the possible transition zone, subsequent samples taken in the continuous stratigraphic sequence were cut to 2.5-cm (1 inch) cylinders upon return to the laboratory and were initially measured on a PAR SM-1 magnetometer. Measurements made during the thermal demagnetization experiments, including remeasurement of the initial magnetization, were on a 6-cm vertical axis SCT cryogenic magnetometer.

Discussion of NRM Results

The natural remanent magnetization (NRM) of all samples is summarized in Figures 2 and 3. Several features are immediately apparent. The reversed samples have a declination near 150° and a slightly negative (upward) inclination. The normal sequence has a declination of about 340° and a positive inclination of about 23°. The initial grouping of the reversed and normal groups is remarkable.

Another conspicuous feature is what apparently looks like a polarity transition zone recorded from about 1.0 m to approximately 4.0 m for the three profiles (Figures 6 through 9). At this point this feature could represent either a true polarity transition or a superposition of normal and reverse components, or overprints of the present geomagnetic field, or a diagenetic feature such as a multiple antiparallel component acquired through chemical alteration over long time intervals. However, upon closer examination, none of these samples are near the field observed today and only one is near the axial dipolar field (Figure 2a).

Directions of both reversed and normal polarity are well grouped (Figures 2a–2c and 3a), and both groups have dispersions characteristic of many red bed sequences [e.g., Helsley, 1969]. The intermediate directions corresponding to the apparent polarity transition definitely do not show obvious evidence of streaking toward the present earth's magnetic field. The normal and reversed (NRM) groups give results that are very similar to other results from the Chugwater Formation reported by Grubbs and Van der Voo [1976] and Van der Voo and Grubbs [1977] and from other Lower Triassic units from the Colorado Plateau [Irving, 1964; Helsley, 1969; Helsley and Steiner, 1974; Elston and Purucker, 1979; Purucker et al., 1980]. Thus both groups can be considered to be essentially stable and to represent a true direction for the Triassic period.

Demagnetization Experiments

Alternating field demagnetization of seven pilot samples was found to be ineffective in removing systematic spurious components even at field values of up to 280 mT. Similar results were found in red sediments of the Moenkopi Formation [Helsley, 1969]. This is probably due to a 'hard' viscous remanent magnetization (VRM) in hematite [Dunlop and Stirling, 1977]. Thus magnetic stability of the Chugwater samples was studied primarily by thermal methods.

Thermal demagnetization was accomplished in a conventional noninductively wound electrical furnace, the samples being cooled in a region where the residual field was less than 10 nT. Temperature was monitored in the furnace by means of a Chromel-Alumel thermocouple that was referred to an ice bath and read on a Beckman digital multimeter and a Love temperature controller.

Five pilot specimens were selected for a progressive thermal demagnetization experiment. These experiments produced changes in the direction of magnetization as well as effects on the observed intensity, until temperatures in excess of 550°C were reached. Figures 4, 5, and 6 show the response of the specimens to progressive heating experiments.

The normal specimens show a minor change in declination (7° to 14°) and inclination (6° up to 20°) (Figure 4a), and their intensities decrease by 30 to 40% at temperatures of less than 550° (see Figures 6 to 9). Transitional specimens, for example, specimen 126, show larger changes in declination (20°) and inclination (13°). The reversed sample shown in Figure 4 did not display systematic changes of more than 3° in declination and about 6° in inclination and there is little change in intensity at temperatures of less than 620°C (Figures 4b and 6). These results are comparable to the results from the Lower Triassic Moenkopi Formation sam-
samples obtained by Helsley [1969] (normal samples); they show an entirely different magnetic behavior from pilot samples reported by Lienert and Helsley [1980], which show regular decreases in their intensities up to 650°C. Moreover, Lienert and Helsley [1980] observed three samples that define a change in magnetization characteristics close to 550°C, which may indicate the presence of magnetite (Curie point 578°C). Such evidence for magnetite has not been observed in Chugwater samples. The intensity of these specimens (normal and reversed) falls off rapidly at temperatures of greater than 660°C, indicating a Curie temperature close to that of hematite (675°C) (see Figures 4b and 5).

The demagnetization curves presented in Figure 4b show an apparently uniform behavior; all of them are characterized by thermally discrete components [Irving and Opdyke, 1965], especially specimens 93, 126, and 127, in which the intensity remains unchanged up to temperatures near the Curie point. These characteristics agree with the results reported by Collinson [1974, Figure 2] and Elston and Purucker [1979] in red beds of the Moenkopi Formation, where the progressive thermal and chemical demagnetization analyses indicated that the most stable magnetization resides in specularite. Similar behavior is seen in samples of the Moenkopi Formation at Bears Ears Utah [Lienert and Helsley, 1980] in which a group of pilot samples shows a slight decrease in intensity up to 200°C and then remains fairly constant up to temperatures of 650°C, where the intensities fall off rapidly, indicating a Curie temperature close to that of hematite.

The directional changes observed in four of the samples are shown in stereographic projection in Figure 4a and the Zijderveld plots of two of these are shown in Figure 5. In these two cases the samples show a stable normal direction at temperatures exceeding 550°C; this is supported by the behavior of the point representing temperatures above 550°C that trend toward the origin of the Zijderveld diagram [Zijderveld, 1967]. This suggests that the NRM consists of two components: a very small secondary component (viscous?) and a dominant component that could be compared favorably with the expected Triassic direction for North America [Helsley and Steiner, 1974; Van der Voo and Grubbs, 1977]. The predominant component is the only component seen above 620°C as shown in Figure 5. Transitional samples 125 and 133 are essentially univectorial, as are the normal and reversed samples 93 and 141. This univectorial behavior strongly suggests that the transition samples consist of a single component of magnetization.

On the basis of the results of the demagnetization experiments with individual samples, the rest of the specimens were demagnetized at temperatures of 550°C (profiles 1 and 2), and profile 3 results at 500°C, 620°C, and 660°C are shown in Figures 2, 3, and 6–9.

Many samples could not be thermally demagnetized to 500°C, for a number of the specimens had broken during drilling in the field, and it was necessary to glue them together with epoxy, which breaks down at relatively low temperatures (see difference in samples populations in Table 2).
Fig. 3. Equal area plots of the NRM, 500°C, 620°C, and 660°C directions of profile 3. Crosses indicate the axial dipolar field; solid triangles represent the present earth's field direction.

Fig. 4. Response of selected samples to progressive thermal demagnetization: (a) equal area plot of changes in direction; (b) changes in intensity. Arrows indicate the direction of change with increasing temperature. Cross indicates the axial dipolar field direction; solid triangle represents the present earth's field direction.
Figure 3 shows data for 500°C and 620°C, and visual examination of the data suggests that the best step of demagnetization is 620°C, although Table 2 indicates that the 500°C step yields slightly the best statistical parameters. This suggests that demagnetization in the 500°C to 600°C range is both appropriate and adequate for this sample suite. Table 2 summarizes the mean directions and statistical parameters calculated for the normal and reversed intervals from the NRM and thermal demagnetization data at 500°C, 550°C, 620°C, and 660°C. The normal interval shows a movement away from the present earth’s field toward the mean Moenkopi normal direction reported by Helsley (1969, Table 7) and Lienert and Helsley (1980, Table 1). The reversed interval is tightly grouped and shows a discrepancy of 18° (profile 3, 620°C) from the mean declination and zero degrees in mean inclination with respect to the reversed (500°C)
Fig. 6. Stratigraphic plot of inclinations, declinations, and intensities of the NRM and 550°C of profile 1. The hachured areas represent a marker horizon.
Fig. 7. Stratigraphic plot of inclinations, declinations, and intensities of NRM and 550°C of profile 2. The hachured areas represent marker horizons.
Fig. 8. Stratigraphic plot of inclinations, declinations, and intensities of NRM and 550°C. The hachured areas represent a marker horizon.
Fig. 9. Stratigraphic plot of inclinations, declinations, and intensities of 620°C and 660°C. The hachured areas represent a marker horizon.
selected mean reported by Helsley [1969, Table 7] for the Moenkopi Formation. The reason for this difference is unknown, but it should be noted that the Moenkopi samples used in Helsley's 1969 study are not demagnetized above 500°C and appear to have considerable scatter remaining after demagnetization. The only significant effect of the additional steps of demagnetization (620° and 660°C) was a small reduction in scatter in the normal interval.

At this point the most striking feature of the stereographic projections and the stratigraphic plots of the three profiles is the presence of intermediate directions of magnetization between the normal and reversed intervals that could be easily correlated over tens of meters by means of marker horizons (hatched lines in Figures 6 to 9). The intermediate directions represent a polarity transition due to (1) a definite directional change of 180° and (2) a conspicuous decrease in intensity of magnetization (in this particular case coincident with the directional change). Figures 6 to 9 show stratigraphic plots of the declination, inclination, and intensity data both before and after thermal demagnetization.

**Behavior of the Geomagnetic Field During a Polarity Transition**

**Intensity of Magnetization of the Samples**

Picard [1964] analyzed remanent intensities of undemagnetized samples from the Chugwater Formation in terms of lithologic types and in terms of normal and reversed intervals. He found that the distributions within normal and reversed groups were identical and that the only differences present were between lithologic types. His mean intensity for all samples was $6.5 \times 10^{-3}$ A/m, and his range was from 0.6 to 70.4 $\times 10^{-3}$ A/m. Table 1 presents the mean intensity of all samples and the range of NRM intensities is from 4.04 to 246 $\times 10^{-3}$ A/m. Figures 6 to 9 show stratigraphic plots of the entire sections. Careful examination of these data, their running averages, and the mean intensities in Table 1 suggests that the intensity of the magnetic field can be subdivided into five intensity periods. These intensity periods, identified by either apparent changes in intensity or direction, are somewhat subjective but nevertheless form a basis for model discussions. From the base of the sections up, these periods are as follows: period 1—typical normal field intensity from 0 to approximately 1.20 m; period 2—generally a time of decreasing intensity from 1.20 m to approximate-ly 2.0 m; period 3—a period of very low field intensity from 2.0 m to 3.0 m; period 4—a period of increasing field intensity from 3.0 m to approximately 3.5 m; and period 5—a period of typical reversed field intensity from 3.5 m to approximately 4.0 m (top of profile 3). (See Figures 6 through 9.)

There is a slight difference in the mean intensity of magnetization between the normal and reversed groups, but there is a remarkable difference between both the NRM and thermally demagnetized mean intensities of the reversed and normal groups and the mean intensity in the transition zone (see Table 1). Two features are immediately apparent. First, there is a pronounced decrease in the intensity of magnetization upon demagnetization of the samples in the transition zone with respect to the normal and reversed intervals (Figures 6 to 9). Second, the values of mean intensities for the three profiles after heating to 550°C and 620°C are lower with respect to the mean intensities of their respective NRM data (see Table 1). As mentioned above, the transition zone is divided into five periods depending upon the intensity and direction of magnetization of the samples. The true dipolar field is represented by periods 1 and 5. The lower intensity in periods 2 and 4 in the transition zone probably reflects a decrease in the intensity of the earth's magnetic field without any significant change in direction and thus supports the idea of a dipolar field reducing to a minimum value close to zero followed by an increase in the opposite sense, following the transition. Period 3 coincides with the onset of the directional change, and its maximum intensity decrease is coincident with the steepest inclination (Figures 6 to 9). This is consistent with a model in which the dipolar field has decreased to near zero and the field then becomes dominated by the nondipolar portion. Low intensities of magnetization have been generally noted in polarity transitions (Lawley, 1970; Dagley and Lawley, 1974; Baag and Helsley, 1974; Hillhouse and Cox, 1976), and thus it is worth noting once again that within period 3 the intensity of magnetization drops to a value that is about 25% of the average intensity observed after the dipolar field has recovered; this is in close agreement with the 20% of the nondipole component reported by Cox et al. [1975], and thus it might be correct to infer that period 3 is a period of nondipole dominance. On the other hand, if instead of taking the mean value of the nondipole dominant period one takes the mean of the minimum values of magnetization (at least three samples per profile) in the

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**TABLE 1. Mean Intensity of Magnetization ($1 \times 10^{-3}$ A/m)**

<table>
<thead>
<tr>
<th>Profile</th>
<th>Normal Period 1</th>
<th>Transition: Period 3</th>
<th>Reversed Period 4</th>
</tr>
</thead>
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<td></td>
<td>Dipole Decreasing</td>
<td>Nondipole Dominance</td>
<td>Increasing Period 5</td>
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<tr>
<td>Profile 1</td>
<td></td>
<td></td>
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<td>0.52</td>
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</tr>
<tr>
<td>Profile 3</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>NRM</td>
<td></td>
<td>3.07</td>
<td>2.85</td>
</tr>
<tr>
<td>620°C</td>
<td></td>
<td>2.48</td>
<td>1.94</td>
</tr>
<tr>
<td>620NRM</td>
<td></td>
<td>0.81</td>
<td>0.68</td>
</tr>
</tbody>
</table>
transition zone and compares this value with the mean value of the reversed field after demagnetization to 550°C and 660°C, it represents about 9% of the dipole field in Early Triassic time. This result agrees with the result given by Irving [1964], who reports that the rms strength of the nondipole field at the surface amounts of about 5% of today’s dipole field. Although this does not prove that period 3 is a nondipole dominant period, it is strongly suggestive.

Duration of the Reversal and Timing of Phenomena Within the Reversal

The data set indicates that a slow change in intensity, without change in direction, begins long before the marked intensity decrease associated with the change in direction. The observed intensity decrease accompanying the intermediate directions in these profiles is shown in Figures 6 to 9. The major decrease in intensity of magnetization appears to be coincident with the directional change and recovers shortly after the directional change is complete. The present results appear to indicate that the major portion of the intensity change took the same amount of time as the directional change. The less pronounced intensity lows of periods 2 and 4, however, precede and follow the directional change and suggest that the total time involved in the transition may be three times more than that associated with the period of rapid directional change.

Based on profiles 1 through 3, the average thickness of the transition zone is 1.0 m. From this one can estimate the duration of the directional change and intensity excursion. Using the same type of argument as that presented by Picard [1964], namely, thickness divided by an estimate of the number of years represented by the formation, the estimated average depositional rate for the formation is about 41,000 y/m of sedimentary rock. However, it is unlikely that any one part of the formation is deposited at this extremely slow rate. If one assumes that most of the time is represented by hiatuses, i.e., periods of nondeposition or erosion, then the well-bedded upper portion of the formation should represent more time while the poorly bedded lower part of the formation (Red Peak Member sampled for this study) would represent a considerably higher than normal sedimentary rate. Thus the duration of the transition phenomenon would have to have taken place in substantially less than 36,900 years. How much less, unfortunately, cannot be determined. Without any better information on rates of sedimentation of parts of the Chugwater Formation, no conclusion can be made about the length of time represented by the 1.0 m of sediments in which the intermediate directions are recorded.

From speculation based on average rates of shelf sedimentation (Picard [1966] considers the ‘alternating facies’ to be an alternating of paralic and shallow marine environments), a sedimentation rate of about 15 to 700 mm/1000 years [Fischer, 1969] can be assumed. Under this assumption the polarity transition might have taken as long as 27,000 years or as short as 600 years. Thus the range of estimates is compatible with the estimates made by others but in no way contributes any additional understanding of the time required for transitions.

Comparing the transition pattern (declination and intensity) of this study with the transition recorded in the Moenkopi Formation reported by Baag and Helsley [1974], the observed data from the Chugwater Formation fit the model of a rapid acquisition pattern that is due to a rapid acquisition component only. This rapid acquisition pattern can be exhibited by a detrital remanent magnetization (DRM) and also by any rapid chemical magnetization, either with or without a certain time lag.

The results obtained in this study indicate that the duration of the intensity change (period 3) is the same as the duration of the directional change and that the polarity transition takes place in approximately 1 m of sediment. These results are in agreement with the results of transition studies in the red bed sequences reported by Khramov [1960], Picard [1964], Petrova [1965], Helsley [1969], and Baag and Helsley [1974], as well as other studies of sedimentary rocks [McElhinny, 1970; Pulliaah and Verma, 1970]. Moreover, the intensity patterns for the transition interval recorded by these red beds are also compatible with those recorded by deep-sea sediments, lava flows, and intrusive rocks, such as the results of Ninkovich et al. [1966], Harrison and Somayajulu [1966], Dunn et al. [1971], York et al. [1971], Steinhauser and Vincenz [1973], Opdyke et al. [1973], Hillhouse and Cox [1976], Dodson et al. [1978], and Hammond et al. [1979]. The results support the hypothesis that the time duration of field intensity decrease is the same as that of direction changes and consequently indicate that the dipole field ‘collapsed’ rather than ‘flipped over’ during the reversal, assuming that the low intensities of the demagnetized data during the transition reflect a decrease in the intensity of the geomagnetic field [Hillhouse and Cox, 1976].

Virtual Geomagnetic Poles

As an alternative presentation of the data, the results are plotted as a succession of virtual geomagnetic poles (VGP) defined as the position of the equivalent geomagnetic pole calculated from a spot reading of the paleomagnetic field. They represent only an instant in time, just as the present geomagnetic poles are instantaneous [McElhinny, 1973]. It should be noted that this is not intended to imply that the field was dipolar during the reversal. As shown by Dodson et al. [1978], VGP’s afford a convenient presentation, particularly if comparisons between different records are to be attempted.

The reversal is best defined in profile 3 and is defined by 83 VGP’s (48 are intermediate). The record certainly falls in the ‘A’ category of Fuller et al. [1979]. The VGP’s make a fairly smooth transit from normal to reversed polarity, with the path confined to a sector of latitude between 85°W and 110°E. Following the reversal’s onset, the virtual pole moved slowly across northern latitudes, then lingered briefly between 60°N and 40°N and between 35°E and 70°E, then again between 20°N and 40°N and from 335°E to 5°E, suggesting the presence of brief periods of stability of the field (or nondipolar field dominance) or fluctuations in sedimentation rate. Before completing its transit to reversed polarity, the VGP’s representing the transitional field seemed to recover very slowly in comparison with the behavior of the VGP’s representing the field right after the reversal’s onset (see Figure 10).

The systematic and reproducible transition VGP path determined in this Chugwater reversal may be the result of a sluggish paleomagnetic recorder having smoothed a more rapidly changing magnetic field, each horizon representing an integration of the field over a period of several years or several centuries. The sedimentation rate may not have been rapid enough and the remanence acquisition time not short.
enough to permit full resolution of the transition pole path. A hint of this ‘swinging’ pattern is present if a sample by sample analysis is made. A similar result has been reported by Hillhouse and Cox [1976] on Quaternary sediments at Lake Tecopa, California. These relatively smooth pole paths can be contrasted with other well-documented transitions, for example, the sequence of 26 lava flows spanning an Upper Miocene polarity transition reported by Watkins [1969], where he found large swings in magnetic direction between successive flows, or the results reported by Dunn et al. [1971] within a Miocene granodiorite intrusion which also showed rapid swings. The above cited results on lavas suggest that lavas are instantaneous recorders of the geomagnetic field and these rapid swings are probably real and should probably be expected in any set of data. Thus one is left with the conclusion that the lack of fluctuation may indicate that the field has been imperfectly recorded and that the time of magnetization was not ‘instantaneous’ but simply rapid or penecontemporaneous, as suggested by Baag and Helsley [1974].

Interpretation of Transitional Fields in Terms of Axisymmetric Fields

Assuming that the behavior of a transitional field is nondipolar, one could compare the behavior of this polarity transition with the polarity transitions summarized by Fuller et al. [1979] in order to see if this transition fits the transitional field configuration models proposed by Hoffman and Fuller [1978]. Analysis of polarity transition records, through the determined transitional path of the VGP’s, has
TABLE 2. Statistical Analyses of the Normal and Reversed Intervals

<table>
<thead>
<tr>
<th>Profile</th>
<th>Latitude, °N</th>
<th>Longitude, °W</th>
<th>Demag.</th>
<th>N</th>
<th>D</th>
<th>I</th>
<th>K</th>
<th>ø₉₅</th>
<th>Longitude, °E</th>
<th>Pole Latitude</th>
<th>dp</th>
<th>dm</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 R</td>
<td>43-40</td>
<td>109-40</td>
<td>NRM</td>
<td>9</td>
<td>140</td>
<td>23</td>
<td>34.4</td>
<td>7.5</td>
<td>301.0</td>
<td>33.8S</td>
<td>5.0</td>
<td>10.1</td>
</tr>
<tr>
<td>1 N</td>
<td>43-40</td>
<td>109-40</td>
<td>NRM</td>
<td>12</td>
<td>342</td>
<td>-11</td>
<td>204.5</td>
<td>4.2</td>
<td>289.4</td>
<td>45.6S</td>
<td>2.2</td>
<td>4.3</td>
</tr>
<tr>
<td>Mean</td>
<td>43-40</td>
<td>109-40</td>
<td>NRM</td>
<td>21</td>
<td>332</td>
<td>13</td>
<td>15.1</td>
<td>8.5</td>
<td>112.4</td>
<td>45.9</td>
<td>4.4</td>
<td>8.6</td>
</tr>
<tr>
<td>1 R</td>
<td>43-40</td>
<td>109-40</td>
<td>NRM</td>
<td>550</td>
<td>7</td>
<td>154</td>
<td>-11</td>
<td>204.5</td>
<td>4.2</td>
<td>289.4</td>
<td>45.6S</td>
<td>2.2</td>
</tr>
<tr>
<td>Mean</td>
<td>43-40</td>
<td>109-40</td>
<td>NRM</td>
<td>550</td>
<td>12</td>
<td>337</td>
<td>23</td>
<td>34.4</td>
<td>7.5</td>
<td>103.2</td>
<td>45.8</td>
<td>4.2</td>
</tr>
<tr>
<td>2 R</td>
<td>43-40</td>
<td>109-40</td>
<td>10</td>
<td>6</td>
<td>151</td>
<td>-6</td>
<td>101.4</td>
<td>6.7</td>
<td>291.3</td>
<td>41.8S</td>
<td>3.4</td>
<td>6.7</td>
</tr>
<tr>
<td>Mean</td>
<td>43-40</td>
<td>109-40</td>
<td>10</td>
<td>6</td>
<td>344</td>
<td>22</td>
<td>41.4</td>
<td>7.6</td>
<td>99.1</td>
<td>54.9</td>
<td>4.2</td>
<td>8.0</td>
</tr>
<tr>
<td>2 R</td>
<td>43-40</td>
<td>109-40</td>
<td>NRM</td>
<td>16</td>
<td>339</td>
<td>16</td>
<td>30.0</td>
<td>6.8</td>
<td>104.6</td>
<td>50.0</td>
<td>3.6</td>
<td>7.0</td>
</tr>
<tr>
<td>Mean</td>
<td>43-40</td>
<td>109-40</td>
<td>NRM</td>
<td>16</td>
<td>339</td>
<td>16</td>
<td>30.0</td>
<td>6.8</td>
<td>104.6</td>
<td>50.0</td>
<td>3.6</td>
<td>7.0</td>
</tr>
</tbody>
</table>

N, number of samples; D, declination in degrees, east of north; I, inclination in degrees, positive downward; K, Fisher's precision parameter; ø₉₅, semiangle of cone of 95% confidence for mean directions; and (dp, dm), oval of 95% confidence about the pole position.

shown certain characteristics to be common to many transitions. It is of particular interest that most detailed transitional VGP paths are largely constrained in longitude during the excursion from one polarity to the other. Two such paths from widely separated sites which recorded the Matuyama-Brunhes reverse-to-normal (R→N) transition were found to be widely dissimilar, thus providing clear evidence that the ambient field during this reversal was predominantly nondipolar [Hillhouse and Cox, 1976]. Further clarification of the geometry of transitional fields has been made recently with the recognition of a dependence of the transitional VGP path behavior on site location [Hoffman, 1977]. Moreover, this site control has been shown to be consistent with a transitional field which is predominantly zonal, a field geometry suggestive of a hydromagnetic reversal process which starts at certain latitudes in the core [Hillhouse and Cox, 1976]. Further clarification of the geometry of transitional fields has been made recently with the recognition of a dependence of the transitional VGP path behavior on site location [Hoffman, 1977]. Moreover, this site control has been shown to be consistent with a transitional field which is predominantly zonal, a field geometry suggestive of a hydromagnetic reversal process which starts at certain latitudes in the core [Hillhouse and Cox, 1976].

Both octupole [Hoffman, 1977] and quadrupole [Dodson et al., 1978] dominated transitional field geometries have been hypothesized by Hoffman and Fuller [1978]. Furthermore, they have shown that these field configurations and, hence, primary characteristics of the reversing dynamo process from which each may arise are paleomagnetically distinguishable. They also indicated more specifically which cases are possible, although the lack of available data precludes any firm conclusion.

![Fig. 11. Summary of Triassic poles for North America, including the mean pole for a section sampled in the Red Peak Member in the Chugwater Formation.](image-url)
TABLE 3. Summary of Lower Triassic Poles

<table>
<thead>
<tr>
<th>Formation</th>
<th>Latitude, °N</th>
<th>Longitude, °E</th>
<th>D</th>
<th>I</th>
<th>K</th>
<th>α95</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chugwater (Lander)</td>
<td>47</td>
<td>100</td>
<td>344</td>
<td>27</td>
<td>55</td>
<td>4</td>
<td>Collinson and Runcorn [1960]</td>
</tr>
<tr>
<td>Chugwater</td>
<td>48</td>
<td>112</td>
<td>344</td>
<td>17</td>
<td>51</td>
<td>5</td>
<td>Irving [1964]</td>
</tr>
<tr>
<td>Chugwater</td>
<td>48.5</td>
<td>112.2</td>
<td>154</td>
<td>-15.5</td>
<td>86</td>
<td>5.9</td>
<td>Van der Voo and Grubbs [1977]</td>
</tr>
<tr>
<td>Moenkopi</td>
<td>57</td>
<td>107</td>
<td>338</td>
<td>19</td>
<td>21</td>
<td>13</td>
<td>Collinson and Runcorn [1960]</td>
</tr>
<tr>
<td>Moenkopi</td>
<td>57</td>
<td>98</td>
<td>346</td>
<td>17</td>
<td>85</td>
<td>5</td>
<td>Helsley [1969]</td>
</tr>
<tr>
<td>Chugwater</td>
<td>45.4</td>
<td>115.3</td>
<td>330</td>
<td>27</td>
<td>47</td>
<td>4.1</td>
<td>this study</td>
</tr>
</tbody>
</table>

At this point it is worth noting that several major features of this transition are similar to those summarized by Fuller et al. [1979]. These features are as follows:

1. The reversal is accompanied by a decrease in intensity of field.
2. The reversal in direction is accomplished by a VGP path, which is confined in longitude; however, the mean path from pole to pole is accompanied by a certain degree of looping (possibly a period of stability or nondipolar field dominance) about certain points, so that the path is traversed somewhat irregularly.
3. The VGP path observed at a northern hemisphere site is about 90° from the sampling site; that is, the transitional poles straddle the near-side far-side boundary if one plots up the VGP’s relative to the sampling site.
4. The decrease in intensity seems to be coincident with the time taken for the change in direction, and the duration of the reversal is comparable in length to that reported by other studies (e.g., 10,000 years proposed by Cox et al. [1975]).

Fig. 12. Correlation between the N → R VGP average paths of profiles 1, 2, and 3 and a model intensity of magnetization of the Chugwater Formation.
Comparison With Triassic Paleomagnetic Poles for North America

A summary of the Triassic paleomagnetic pole positions for North America is given in Figure 11 and Table 3. The pole determined in this study lies approximately in the center of the main group of poles reported for other Triassic units (see Table 3 for references). The results reported by Collinson and Runcorn [1960] are based on samples not treated thermally and thus should not be given the same reliability as those for the more recent studies in which demagnetization studies have been made.

Summary and Conclusions

Stable directions of magnetization of the samples taken for this polarity transition study show once again that red sediments are good recorders of geomagnetic phenomena. The clear intermediate directions of magnetization after demagnetizing the specimens to 550°C and 660°C represent a true polarity transition and not two opposed magnetizations of compatible intensity superimposed on each other. This conclusion is supported by the correlation of intermediate directions of magnetization of the same reversal at three different sampling localities that gave the same paleomagnetic answer (identical magnetic behavior upon demagnetization, and VGP paths) displaying a remarkable decrease in intensity coincident with a 180° directional change during the polarity transition. The intensity of magnetization of the section can be subdivided into five intensity periods, two dipolar periods, one dipolar decreasing period, a nondipolar dominant period, and a dipolar increasing period (Figure 12). The nondipolar period took place in about 1.0 m of red sediments, and the mean intensity of magnetization drops to a value that is about 25% of the mean intensity observed after the dipolar field has recovered; or if one compares the values of the mean of the least magnetized samples with the mean value of the reversed field after demagnetization to 550°C and 620°C, it represents about 9% of the dipole field in the Early Triassic. These conclusions lead us to think that the geomagnetic field during this Chugwater polarity transition was nondipolar and support the hypothesis that the time duration of field intensity decrease was longer than that of direction changes and consequently indicate that the dipole field ‘collapsed’ rather than flipped over during the reversal.

The VGP path, confined to a 20° wide sector of longitude between approximately 90° from the site (Figure 10), shows that it is located about 90° from the site, straddling the boundary between the near side and the far side, and thus does not assist in constraining the hypothesis of Hoffman and Fuller [1978].

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