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A DETAILED RECORD OF THE LOWER JARAMILLO POLARITY TRANSITION FROM A SOUTHERN HEMISPHERE, DEEP-SEA SEDIMENT CORE

Bradford M. Clement and Dennis V. Kent

Lamont-Doherty Geological Observatory and Department of Geological Science, Columbia University

Abstract. A detailed record of the lower Jaramillo (reversed to normal) polarity transition was obtained from a southern hemisphere, deep-sea sediment core (latitude = 35.91°E, longitude = 59.97°E) The record consists of over 850 samples taken across 140 cm of section. The transition itself is recorded across approximately 70 cm and is represented by more than 475 specimens from about 160 levels giving intermediate directions. The transition is identified by a nearly 180° shift from directions in good agreement with a reversed, axial dipole field to those closely aligned with a normal, axial dipole field for the core site latitude. The inclinations shallow gradually early in the reversal and pass through very steep negative values (-80°) late in the transition. The declinations show little appreciable variation until the inclinations have moved through the near vertical, and then slowly approach values in agreement with a normal polarity field. An intensity low accompanies the directional change during which the intensity drops to less than 15% of the maximum values observed in this sample interval. The intensity fluctuation spans a wider interval than the directional change, decreasing prior to any systematic change in the directions and then increasing to pre-transition levels by the same depth at which the directions have stabilized. The VGP path constructed for this reversal is longitudinally constrained to a certain extent, between 140° and 230° for intermediate VGP latitudes and is roughly centered 120° from the site longitude. This path is therefore a far-sided VGP path in Hoffman's [1977] terminology. Assuming a constant sedimentation rate (67m/Ma) through the Jaramillo Subchron, the duration of the transition is estimated to be 11,200 years to 4,500 years (depending on the criteria) for the directional change, whereas the associated intensity variation occurred over 15,000 to 20,000 years. Considered together with records of the most recent reversal (Matuyama/Brunhes) in light of current transitional field models, this record strongly suggests that the lower Jaramillo transitional field was dominated by different harmonics than the Matuyama/Brunhes transitional field.

Introduction

Although the fact that the earth's magnetic field undergoes frequent polarity reversals has been well established for over two decades, the reversal process itself is still very poorly defined. Paleomagnetic records of polarity transitions can provide information about the behavior of the geomagnetic field during a reversal and therefore may be capable of placing constraints on possible reversal mechanisms. In recent years, several polarity transition records have become available, allowing the development of transitional field models (Hoffman, 1977; Hoffman and Fuller, 1978; Fuller et al., 1979; Williams and Fuller, 1981; Hoffman, 1982). These models generally assume that the transitional field was dominated by axisymmetric fields, predicting a dependence mainly upon the site latitude for transitional records. The available records, however, are not from widely distributed sites, and in fact almost all are from northern hemisphere mid-latitudes. This poor distribution, and in particular the lack of southern hemisphere records, has left these models poorly constrained. In addition, there is a poor temporal distribution of these records, with the Matuyama/Brunhes reversal being the only one for which multiple, detailed records exist. The few detailed records of earlier reversals, of both reversed to normal (R-N) and normal to reversed (N-R) sense, do not yet allow definite statements to be made regarding possible differences from one reversal to another.

We resampled the lower Jaramillo transition as it is recorded in deep-sea core RC14-14 in an effort to obtain a detailed southern hemisphere record of a reverse to normal polarity transition. This transition was first reported by Opdyke et al. [1973]. The very high apparent sedimentation rate across critical intervals and the exceptional quality of the magnetic data in this core provided the incentive to study this transition in more detail.

Core Description

Deep-sea core RC14-14 was taken in the Crozet Basin at 35.91° south latitude, 59.97° east longitude. It consists of radiolarian and diatom lutite that, in the interval of interest here, occurs in very fine laminae, suggesting minimal disturbance from bioturbation. The magnetostratigraphy of this core as interpreted by Opdyke et al. [1973], based on the stable natural remanent magnetization (NRM) of the sediment, is defined by four distinct polarity zones. A zone of reversed polarity extends from the top of the core to a depth of 460 cm followed by a normal polarity zone from 460 cm to 940 cm. From 940 cm to 2250 cm the sediment is reversely magnetized with the rest of the core below 2250 cm being normally magnetized. As the top of the core is reversely magnetized it is apparent that the Brunhes polarity zone is missing and therefore the interpretation of the magnetostratigraphy was not straightforward. It was, however, possible to identify the polarity transition at 2250 cm as

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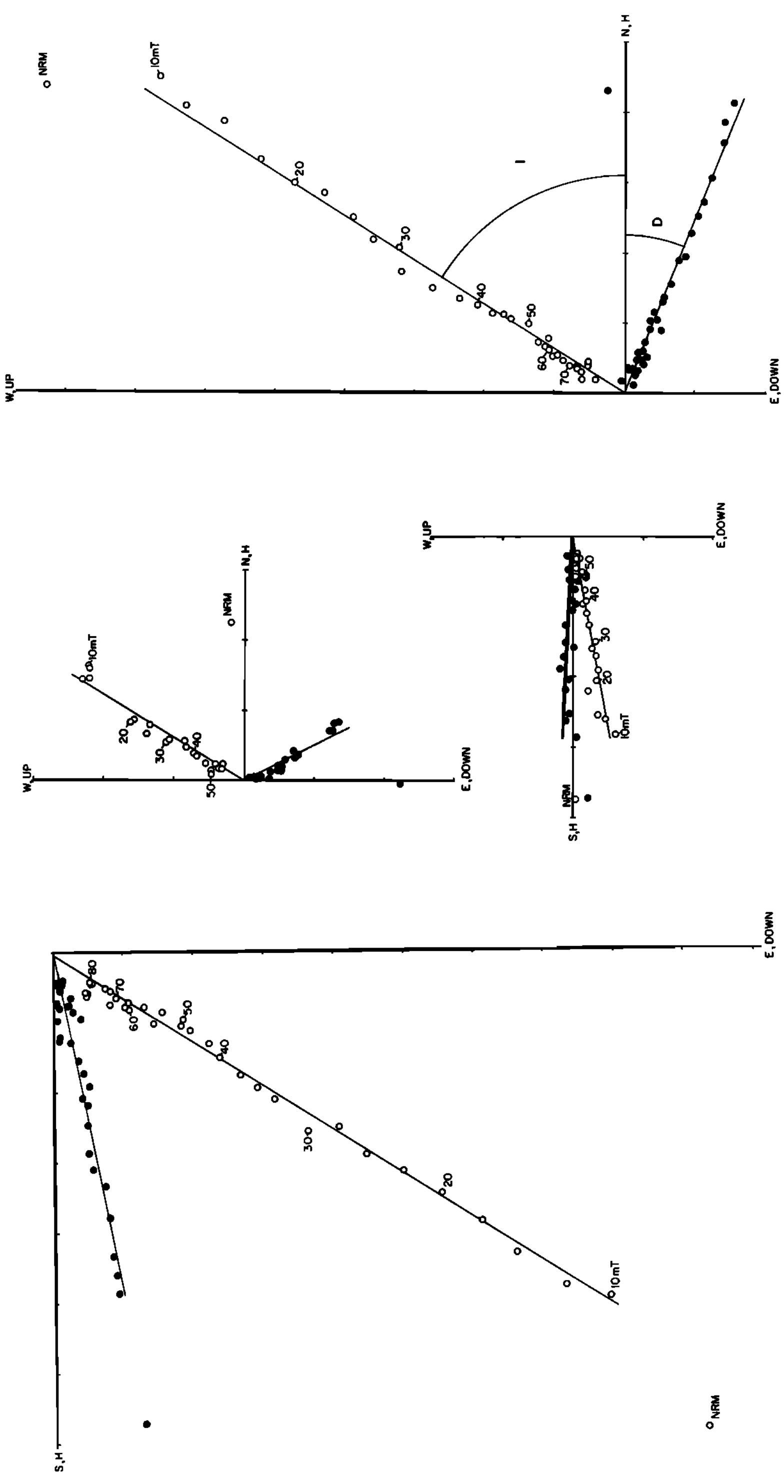

Fig. 1. Results of progressive alternating field demagnetization plotted as inclination-declination vector plots. The numbers next to the symbols are the peak demagnetizing fields in milliteslas. The results from the four specimens are plotted at the same scale with each tick mark spacing representing 1 x 10^{-5} A m^2/kg. Specimen A is from the reversed polarity zone below the reversal, specimens B and C are from within the transition zone, and specimen D is from the normal polarity zone above the reversal.

the upper Olduvai based on the extinction of Clathrocylas bicornus. The normal polarity zone from 2250 cm to 2588 cm (the bottom of the core) was therefore correlated to the Olduvai Subchron and the normal polarity zone from 460 cm to 940 cm to the Jaramillo Subchron.

The thickness of the Jaramillo Subchronozone suggests a high sedimentation rate in this core although the absence of the Brunhes magnetozone and the disproportionately short Matuyama magnetozone suggest that periods of fast sedimentation must have alternated with periods of non-deposition or erosion. The observation of intermediate directions within the transitions by Opdyke et al. [1973], even using standard sampling techniques (2.5-cm samples), suggests high resolution recording of the transitions. In light of these results we resampled the remaining split half of this core by using a much more detailed sampling technique in an attempt to improve the resolution. Data for the lower Jaramillo transition are reported here; similar work on the two N-R reversals (top of the Olduvai and top of the Jaramillo) is in progress.

Experimental Procedures

The lower Jaramillo reversal was placed at approximately 940 cm by Opdyke et al. [1973], leading us to sample the core continuously from a depth of 1010 cm to 870 cm. Successive slices of sediment, 0.3 cm to 0.5 cm thick, were cut from the dried core using a jeweler's saw. Each slice was subdivided to yield three specimens per sampling level. These specimens were weighed and then glued in a known orientation in plastic boxes for measurement. The average specimen measures 8mm x 8mm x 4mm and weighs 0.4 g.

The direction and magnitude of the NRM of each specimen were measured by using a two-axis cryogenic magnetometer with a 6.8 cm access. Progressive alternating field (A.F.) demagnetization studies were carried out at 2.5 mT increments on 20 pilot specimens distributed over the section (1 to 2 specimens from every 10 cm). On the basis of the results of these studies a blanket treatment of 10 mT was applied to the rest of the specimens by using a single axis commercial ac demagnetizer.

Opdyke et al.'s [1973] studies of the anhysteretic remanence (ARM), saturation isothermal remanence and initial susceptibility indicated that there are no dramatic changes of magnetic mineralogy with depth across the polarity transitions. Additional ARM analysis was undertaken here to examine the possibility of changes on the scale of this study. Samples were taken every 5 cm through the sampled section and, on the basis of evidence of approach to saturation values in ARM acquisition experiments, were given an ARM using a 0.05 mT dc field coaxial with a 180 mT alternating field.

Results

The results of progressive A.F. demagnetization of four pilot specimens are shown in Figure 1, plotted as inclination/ declination vector diagrams [Zijderveld, 1967]. These are plotted by using the same scale in order to emphasize the change in directions and NRM intensities observed across the transition. Figure 1a illustrates the behavior of a specimen taken from the reversed polarity zone, about 20 cm below the reversal. Figure 1b and 1c show results from specimens taken from within the transition zone, while Figure 1d illustrates data from a specimen from the normal polarity zone above (about 15 cm) the reversal.

It is clear from these plots that despite the small size of the specimens a stable magnetization can be readily measured by using a cryogenic magnetometer. Treatment at 10 mT removes a small viscous magnetization, isolating a component that decays linearly to the origin with further A.F. treatment up to 80 mT to 90 mT. Such behavior of samples taken from within the transition interval demonstrates that the observed intermediate directions are not the result of complex multicomponent magnetizations. Instead, the univectorial magnetization component isolated after 10 mT is thought likely to represent a record of the direction of the geomagnetic field near the time of deposition.

The transition record obtained after partial demagnetization at 10 mT is shown in Figure 2; the results from the three specimens at each sampling level are plotted to allow judgement of the internal consistency of the data. The transition is identified by a gradual, nearly 180° directional change from reverse to normal polarity across approximately 70 cm of section. A decrease in remanent intensity accompanies the directional change with NRM intensities decreasing to less than 15% of the maximum values observed outside the transition.

The directional data show very tight within and between level grouping both in the reversed polarity interval sampled below (mean D = 180.0°, I = 54.1°, alpha95 = 5.15°, k = 41.2 for n = 20 levels from 979.4 cm to 989.3 cm) and the normal polarity interval above the reversal (mean D = 22.9°, I = 55.2°, alpha95 = 5.6°, k = 67.2 for n = 19 levels from 880.0 cm to 890.8). These mean inclinations for normal and reversed polarity intervals are almost identical irrespective of sign and agree closely with the axial dipole field value of 55° expected for the core site latitude. We regard these agreements as providing strong evidence that these sediments have accurately recorded the geomagnetic field.

The mean declinations for the normal and reversed intervals, however, fall short of being antipodal by 23°. This core was not oriented with respect to azimuth when taken, so the observed declinations were measured relative to the split face of the core. Even relative declination values are therefore dependent upon how well the internal orientation of the core was maintained since its retrieval. Although the declination change with respect to the internal orientation closely approaches that expected for a full polarity reversal, there is the possibility of a core twist or of a core break, perhaps coinciding with the apparent discontinuity in the record observed at 924 cm. This break will be discussed in detail later.

The transition from reverse to normal polarity (Figure 2) occurs across approximately 70 cm of section. The inclination record is characterized by a gradual shallowing trend early in the reversal after which the directions pass through

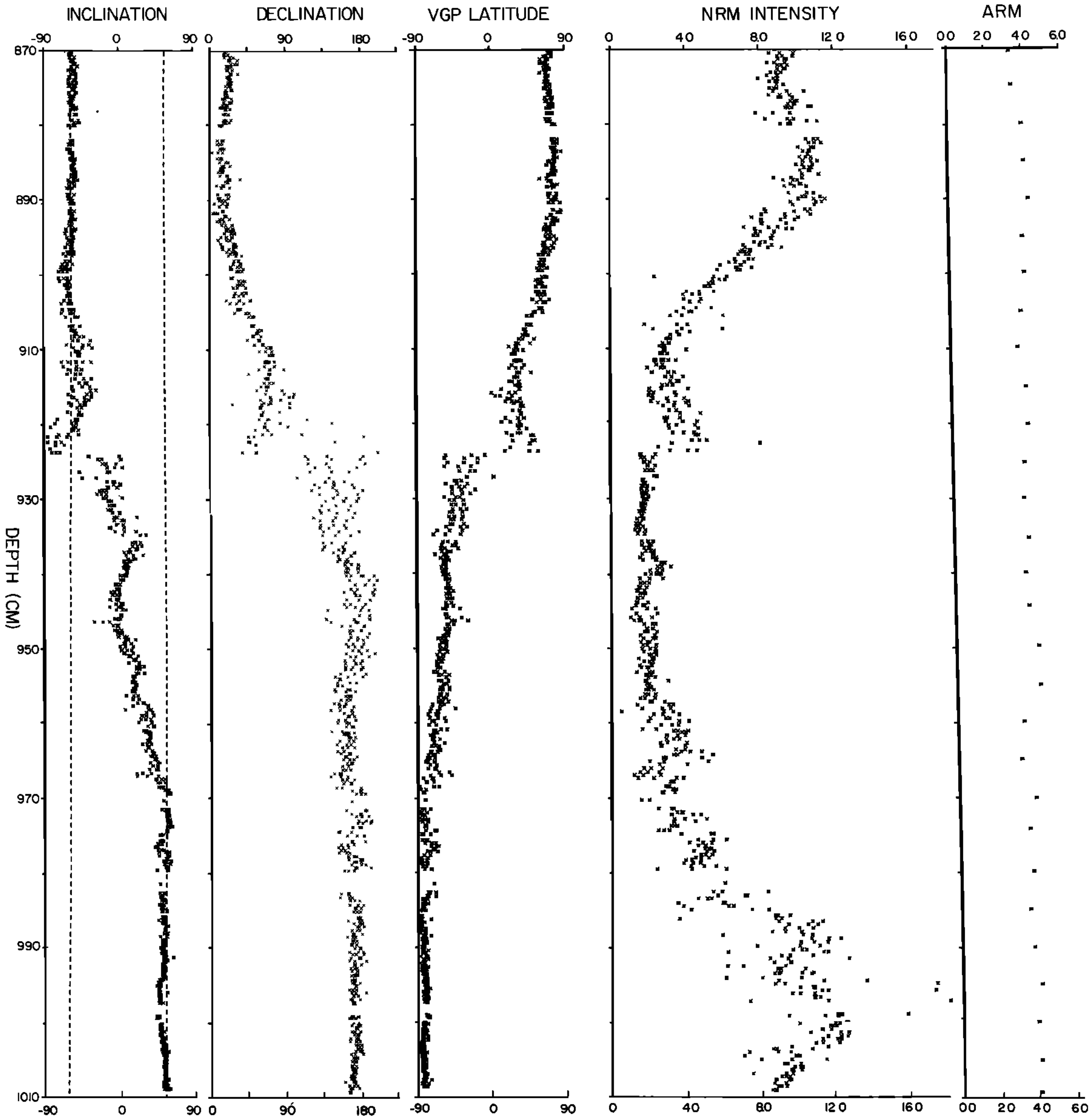

Fig. 2. Paleomagnetic results from RC14-14 after A.F. demagnetization at 10 mT plotted with depth from the core top in centimeters. Inclination (A), declination (B), and VGP latitude (C) are plotted in degrees; ARM intensity (D) plotted in 10^{-4} A m^2/kg; and NRM intensity normalized to specimen weight (E) plotted in units of 10^{-5} A m^2/kg.

very high, negative values before returning to normal polarity dipole inclinations. On the other hand, the declinations show very little change early in the reversal and only begin to change as the inclinations approach near vertical values. The NRM intensity record is characterized by a broad, roughly symmterical trough, which is not well centered about the directional change.

In detail, the inclinations begin to shallow gradually at 970 cm, moving away from values very near to those predicted by a reversed axial dipole field (55.4°) and become nearly horizontal by 948 cm. The directions remain approximately flat from 948 cm to 924 cm with departures of up to 25° at 935 cm and 930 cm. An apparent discontinuity exists in the record at 924 cm across which the inclinations abruptly become very steep in an upward direction (-80°) from 924 cm to 920 cm. The inclinations then shallow to less than dipole values before grouping at -55° by 905 cm.

The declinations remain nearly constant from 1010 cm to 940 cm but begin to rotate toward normal polarity directions above 940 cm. From 935 cm to 924 cm an increasing dispersion appears in the declination data, which is manifest to a lesser extent in the inclination data as well. A portion of the declination dispersion is a result of the very steep inclinations and does not necessarily indicate a poor grouping of the total vector. This, however, cannot account for the total dispersion observed as will be discussed later. From 924 cm to 893 cm the declinations change by more than 35° with no further significant trend apparent in the data above 893 cm.

The intensity low (Figure 2e) associated with the directional change occurs across more than a meter of section. Immediately preceeding and following the low are what appear to be intensity highs. The record, however, is not sufficiently long to determine if these highs are within the range of non-transitional fluctuations. The intensity high preceeding the reversal is observed from 1005 cm to 990 cm as the intensity increases upcore from 9.0×10^{-5} Am^2/kg to a mean of $11.0 \times 10^{-5} Am^2/kg$ before beginning to decrease (at 990 cm). The intensity values decrease by 80% across 20 cm (from 990 cm to 970 cm) before any change in the remanent directions occur and have decreased by 85% at 958 cm where they remain relatively constant from 958 cm to 924 cm at $1.5 \times 10^{-5} Am^2/kg$. A small, abrupt increase at 924 cm marks an apparent discontinuity in the record, above which the intensity drops slightly before beginning to increase in a linear manner. This increase occurs from 910 cm to 892 cm as the values return to a mean of $11.0 \times 10^{-5} Am^2/kg$ before decreasing again to nearly $9 \times 10^{-5} Am^2/kg$ by 875 cm. The intensity increase begins while the remanent directions can still be considered to be transitional and the maximum intensity is reached at approximately the same depth at which the directions stabilize at the normal polarity dipole direction.

Variations in the intensity of magnetization may be reflecting changes in field behavior or changes in the amount or type of magnetic minerals present. Opdyke et al. [1973] used a number of methods, including saturation isothermal remanent magnetization, bulk susceptibility, and ARM measurements, which are used to normalize the intensity record for lithologic variations. Additional work indicates that the ARM method may be the most effective of these as it often appears to be represented by the same magnetic fraction that carries the natural remanent magnetization [Levi and Bannerjee, 1976]. The ARM intensities are plotted in Figure 2d and reveal very little variation, suggesting that there are no marked changes in magnetic mineralogy across the transition (mean = $3.93 \times 10^{-4} Am^2/kg$, standard deviation = $0.39 \times 10^{-4} Am^2/kg$ for n = 29). This implies that the observed NRM intensity record is not strongly controlled by changes in the nature or amount of the magnetic carrier but may instead reflect changes in the geomagnetic field intensity across the polarity transition.

Virtual Geomagnetic Poles

Although this core only has internal azimuthal orientation it is possible to calculate virtual geomagnetic poles (VGPs) for these results by making a deliberate adjustment to the data. The entire declination record across the transition was uniformly reoriented so that the mean reversed declination was set to 180°. Justification for this adjustment is based on the excellent agreement of the mean normal and reversed inclinations with the predicted axial dipole values from which we can infer that the mean declinations for the same intervals should lie along a meridian. The latitude of the VGP calculated for each specimen is shown in Figure 2c. The VGP latitudes move gradually from near -90° below 970 cm to nearly +90° above 893 cm. Again as in the other records (inclination, declination, and intensity) this gradual shift is interrupted by an apparent break at 924 cm.

The VGP path calculated for this reversal is shown in Figure 3. Each VGP was calculated from the unit vector mean [Fisher, 1953] of the three observed directions at each sampling level. The resulting path is somewhat longitudinally constrained, lying between 140°E and 230°E longitude, and is roughly centered over the 180°E meridian. As the path falls more than 90° from the site longitude it is classified as a far-sided VGP path. It is interesting that the points on this path fall in a series of discrete clusters centered at 60°S, 190°E; 35°S, 180°E; 35°N, 155°E and 60°N, 170°E. This may indicate that the geomagnetic field changed in a sporadic manner during this interval or that the recording process was discontinuous on this scale. The latter is thought to be a likely contributing factor as this section of the core is laminated, a feature that in itself suggests nonuniform sedimentation.

The VGP positions for the normal polarity interval appear offset from the north pole because as noted earlier, the normal and reversed polarity declinaitons fall short of being antipodal by 23°. If it is assumed that the discontinutiy observed at 924 cm represents a core break, then it is reasonable to suggest that the declination offset across it may be the result of a differential rotation of the core. An adjusted VGP path was therefore calculated after uniformly rotating the declinations above the discontinuity by 23° (Figure 3). The problems associated with the discontinuity will be discussed in detail later; however, the modified VGP path illustrates the extent of the possible distortion of the record if the core actually did break at this level.

Interpretation of Data

The method of taking three specimens (A, B, and C) for measurement at each sampling level allowed the recognition of a local zone of inhomogenous magnetization in this core. Beginning at approximtely 940 cm and continuing upcore to 924 cm, a divergence, which increases upcore to 924 cm and then decreases to 920 cm, is observed in the declinations of the A, B, and C specimens. In addition, from 924 cm to 920 cm, the A specimen inclinations are up to 20° steeper than the B and C inclinations. While this within level behavior definitely seems to suggest that this section of the core has been deformed, the combined inclination and declination behavior cannot be readily explained by a simple shearing or twisting that might be attributed to the coring process.

The deformed zone is further complicated by an abrupt discontinuity at 924 cm. Since the nearest physical breaks in the dried core were at 913 cm and 926 cm this does not appear to be the result of an improper orientation of the core segments. The discontinuity possibly could have been caused by failure at this level as the core twisted during the coring or extrusion process. Such a break might explain why the observed declinations are not exactly antipodal. The core photographs show a thin, dark lamination at about 924 cm which may be an indication of a sedimentological break in the record and therefore offers an

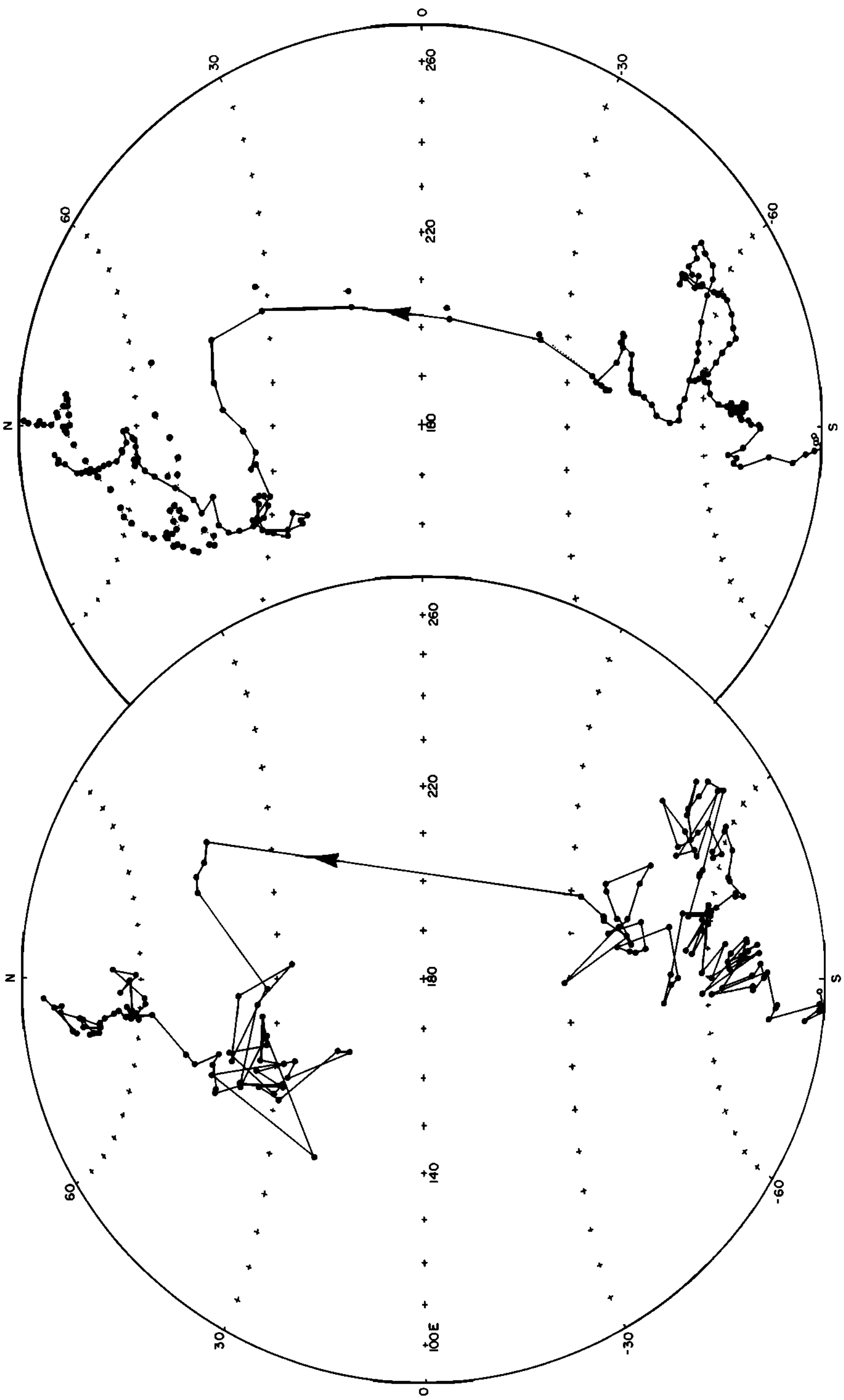

Fig. 3. Virtual geomagnetic pole (VGP) path for the lower Jaramillo transition plotted on an equal area, equatorial projection. The VGP position calculated from the vector mean of the three directions observed at each sampling level is shown in (A), a 5pt running vector mean of the VGPs is plotted in (B). Also shown in (B) is an adjusted portion of the VGP path (connected with dotted lines) as discussed in the text.

alternative explanation for the discontinutiy in the paleomagnetic record. As the discontinuity is observed in the intensity data as well as the directional record, it may be the result of an interruption in deposition and therefore may be revealing the presence of a hiatus in the section. Even if this is a hiatus, some form of inhomogenous deformation must have occurred to produce the observed directional behavior. Perhaps the divergence in decinations was caused by a slight curling of the core about its long axis as it dried. In summary, given the variables involved, it is not possible to determine the exact form of deformation this zone represents. For this reason the directions within this interval (940 cm to 920 cm) may be regarded with some suspicion, although it cannot be ruled out that their general character is representative of field behavior.

The reversal record presented here is of such high resolution that it becomes apparent that the lower boundary of the Jaramillo Subchronozone can vary by tens of centimeters depending on the criteria used to define that boundary. By convention in magnetostratigraphy the boundaries of polarity zones are defined by the mid-points of the transition zone separating the two polarity zones. This method originated when dealing that records which contained very little detail of polarity transitions. In cases where details were available, workers sometimes determined the boundaries on the basis of the directional change, for example, as the point where the inclination passes through zero. When working with records of high resolution the difference between these two methods becomes further complicated by the prediction of current transitional field models that the directional behavior as well as the duration of the directional change will be a function of the site location [Williams and Fuller, 1981; Fuller et al., 1979; Hoffman and Fuller, 1978; Hoffman, 1977]. If these models are valid, then placing the boundary of a polarity zone at the mid-point of the transition or at the point of some given directional characteristic will not be placing the boundary at exactly isochronous levels in records from geographically distributed sites.

The difficulty in pin-pointing transition boundaries must be born in mind when defining and comparing polarity boundaries in high-resolution records from different sites. The choice of methods will be governed by the understanding of transitional field behavior and therefore will be dependent upon the existing reversal models. In the light of the Williams-Fuller model, it appears that using the mid-point method is likely to produce a more accurate estimate of the polarity zone boundary than one based on directional features simply because their model predicts highly asymmetrical directional records that vary more than the transition duration with site latitude. In determining the lower boundary of the Jaramillo Subchronozone we have therefore used the directional mid-point. The mid-point is placed at 930.5 cm if the transition zone boundaries are taken as 968 cm and 893 cm. Opdyke et al. [1973] placed the upper boundary of the Jaramillo at 460 cm using the same method. By this criterion, the Jaramillo Subchronozone is 470.5 cm thick. As discussed by Mankinen et al. [1980] the best estimate for the duration if the Jaramillo Subchron is 70,000 years, which indicates an average sedimentation rate of 67m/Ma for this section of the core.

The duration represented by the thickness of the transition zone can be estimated by assuming constant deposition at the calculated rate. In a high resolution record, the arbitrary decision of where the transition boundaries are placed has an important effect on time estimates. If the transition zone boundaries are placed where the directions tend to exceed the circular standard deviation [Fisher, 1953] of the normal and reversed means (10°, calculated from the mean of the interval 880.0 cm to 890.0 cm; and 13° calculated from the mean of the interval 979.4 cm to 989.3 cm, respectively), then the transition is 75 cm (893.3 cm to 968.0 cm) thick and the resulting duration estimate is 11,200 years. This is a very generous criterion to use to define the transition limits. Previous workers [Fuller et al., 1979] have placed the boundaries at the points where the VGP crosses the $\pm 60^{\circ}$ line of latitude. If we apply this criterion, the transition thickness is 53.5 cm (903.8 cm to 957.3 cm), which yields a duration estimate of 7,900 years. This could be taken a step further by picking the limits as the $\pm 45^{\circ}$ latitude crossing of the VGP path [Hoffman, 1977]. Using this, the transition zone thickness is 30.2 cm, which gives a duration estimate of 4,500 years. These duration estimates would all be too small if the discontinuity at 924 cm is actually a hiatus, although durations estimates obtained agree with those of 4,000 to 10,000 years made for other reversals [Ninkovich et al., 1966; Harrison and Somayjulu, 1966; Hammond et al., 1979; Valet and Laj, 1981; Clement et al., 1982].

An estimate of the duration of the intensity change associated with the reversal can be made in a similar manner. The intensity low extends across a meter of section, from 990 cm to 890 cm, corresponding to a duration of 14,900 years. If the highs immediately before and after the low are also considered to be a part of the transitional behavior, as has been predicted by some reversal models [Levy, 1983], then the change is recorded from 1005 cm to 875 cm with an estimated duration of 19,400 years. In either case the intensity variation associated with the polarity reversal took much longer to occur than the directional change.

Transitional Field Models

The detailed southern hemisphere record of a reversed to normal transition presented in this paper is of interest in evaluating some current transitional field models. Although there now exists evidence that different reversals may have had very different field geometries [Williams and Fuller, 1982; Bogue and Coe, 1982], the lack of other detailed records of the lower Jaramillo reversal from other sites leaves only records of other R-N reversals to compare this one with. Such a comparison may prove useful, either in providing further constraints on these models or in establishing the validity of comparing different polarity reversal transitions.

Hoffman's zonal flooding model [Hoffman, 1977] predicts that a transitional VGP path will be

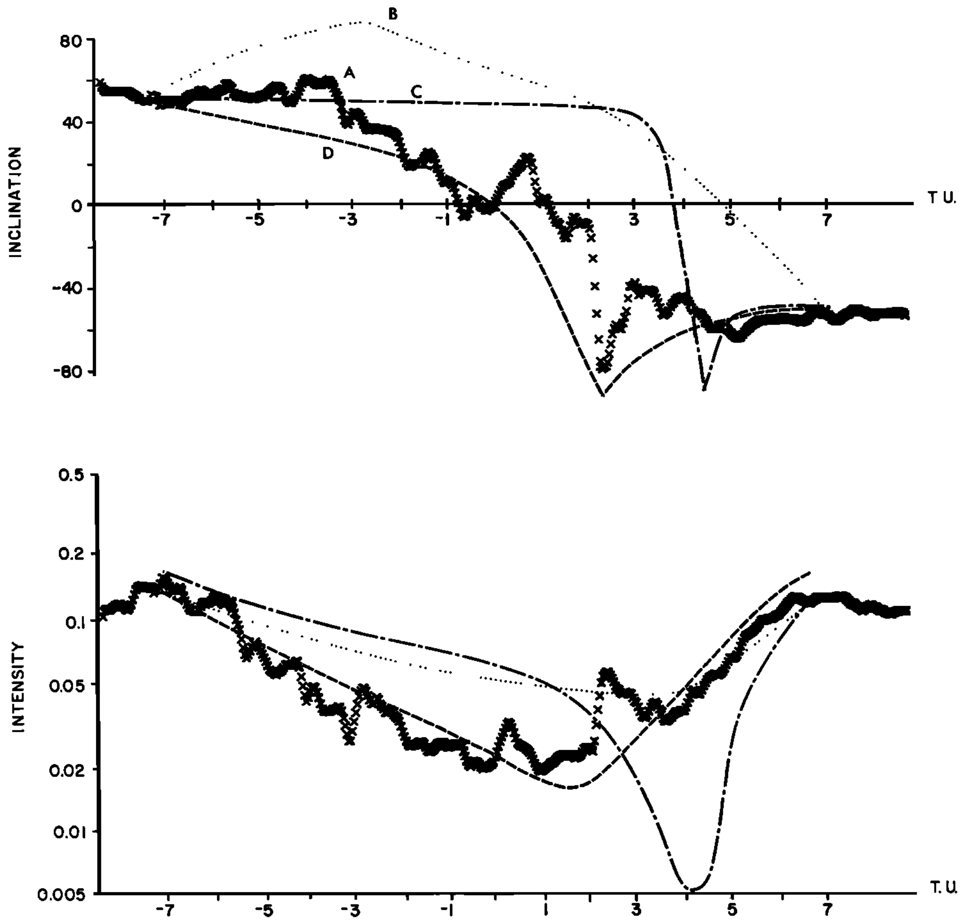

Fig. 4. Five point running mean of the inclination and intensity records from the lower Jaramillo transition (A) plotted together with three synthetic records produced using a Williams-Fuller [1981] type model. Inclinations plotted in degrees and intensities in 1×10^{-4} A m^2/kg versus time units (T. U.). B is the model for the Matuyama-Brunhes field developed by Williams and Fuller (1981) for a site latitude of 35°S. Model C was produced by using their model with the sign of the G_2^0 term changed from positive to negative and with only 30% of the available energy going to G_3^0 and G_4^0. Model D results from adjusting the energy partitioning in model C.

near-sided or far-sided, depending on a number of factors, including whether the site is in the northern or southern hemisphere and whether the transitional field is dominated by a quadrupolar or octupolar field geometry. When Hoffman's [1977] model was developed, the available data (all from the northern hemisphere) exhibited a distinct dependence upon the reversal sense: Normal to reversed transitions showed far-sided VGP paths, while reversed to normal reversals had near-sided paths. Recently reported records of a reversed to normal transition with a far-sided VGP path [Williams and Fuller, 1982] and a normal to reversed transition with a near-sided path [Bogue and Coe, 1982], (both from northern hemishpere sites) make this sense of reversal dependence less apparent.

Nevertheless, the latest reversal (Matuyama/Brunhes, R-N) is still characterized by northern hemisphere transition records yielding near-sided VGP paths. When these records are considered together with the record presented here (of the preceeding R-N reversal), the combined data set consists of near-sided VGP paths from the northern hemisphere and a far-sided path from the southern hemisphere. In terms of Hoffman's [1977] model, this suggests that the transitional fields were characterized by octupolar field geometries. It must be born in mind, however, that only after other detailed records of the lower Jaramillo reversal and southern hemisphere records of the Matuyama/Brunhes reversal become available will it be possible to determine the differences between the lower Jaramillo and the Matuyama/Brunhes transitional fields.

The model developed by Williams and Fuller [1981] assumes that during a reversal, the dipole field decays exponentially to zero and then builds up again in the opposite direction. As the dipole field decays, the energy it loses is redistributed amongst the lower order, nondipole zonal terms of a spherical harmonic expansion of the field. The energy redistribution was based on the results of Verusob 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 nondipole field with terms of degree n = 2,3,4 receiving 72% of that energy. The Williams-Fuller model can predict the inclination and intensity record that would be observed at any site latitude given a particular set of the various parameters involved.

By fitting synthetic inclination and intensity records to the available records of the Matuyama/Brunhes reversal, Williams and Fuller [1981] were able to model that reversal with 20%, 30% and 50% of the dipole energy made available to the G_2^0, G_3^0, and G_4^0 terms respectively. The signs of the nondipole terms were kept constant through the reversal with G_2^0 being held positive, G_3^0 negative and G_4^0 positive. Their model has been reproduced here, and the record that this model predicts for the site latitude of RC14-14 (36°S) is shown in Figure 4 (curve B). When compared with the data presented here we see that their predicted inclination record passes through steep positive values early in the process, while in the data presented here the inclinations pass through very steep, negative values toward the later part of the reversal (after it passes through the horizontal). In addition, the intensity pattern presented here is more symmetric than their synthetic intensity record that reaches a minimum in the latter part of the reversal.

These differences may be an indication that this modeling technique is inappropriate or that the particular model parameters used by Williams and Fuller [1981] for the Matuyama/Brunhes field are not applicable to the lower Jaramillo transitional field. Although the lack of other detailed records of this reversal makes it impossible to uniquely model its harmonic content, the general character of the record presented here can be approximated by making a few adjustments to their model of the most recent reversal.

The feature of this transition record that controlled our modeling is that the inclinations pass through near vertical negative values after passing through the horizonal. By taking the model Williams and Fuller [1981] developed for the Matuyama/Brunhes reversal and changing the sign of G_2^0 from positive to negative and letting it recieve 100% of the energy lost by the dipole term, a synthetic record for this site can be produced in which the inclinations pass through negative, vertical values in the latter part of the reversal. It was found that an inclination record exhibiting steep negative values could be produced as long as no more than 30% of the energy to be distributed to the nondipole terms was made available to the G_3^0 and G_4^0 terms. The site latitude is located very close to the G_2^0 node. This is why the G_2^0 term must be allowed to gain the majority of the energy in order for it to dominate the higher degree terms. For example, if the energy partitioning is set at 20% to a negative G_2^0, 30% to a negative G_3^0, and 50% to a positive G_4^0, as in Williams and Fuller's model for the most recent reversal, but with a cutoff of 30% set for the G_3^0 and G_4^0 terms, the record shown in Figure 4 (curve C) is produced. If adjustments are made to the energy partitioning, a better fitting synthetic record can be obtained. Keeping the 30% cutoff, the resulting model partitions the energy with 70% going to a negative G_2^0, 10% going to a negative G_3^0, and 20% to a positive G_4^0. The inclination and intensity records predicted by this version of the model are included in Figure 4 (curve D). The predicted inclination record passes through -90° in the latter part of the reversal, in excellent agreement with the observed inclinations. In addition, the predicted intensity record appears to fit the observed variation of intensity values quite well.

The results of this modeling support the suggestion [Williams and Fuller, 1982; Bogue and Coe, 1982] that different reversals may have been dominated by very different transitional fields. While there are a number of other variations in the Williams-Fuller model that can approximate this observed record (no attempt was made here to account for all the details of this record), it may be significant that a very good fit is obtained by simply reversing the sign of the G_2^0 and allowing it to dominate the other terms.

Discussion

Analyses of the time averaged paleomagnetic field by Merrill and McElhinny [1977] and Merrill et al. [1979] suggest that the lower order, zonal Gaussian coefficients change sign when the dipole field reverses. They determined that the ratio of G_1^0 to G_2^0 remained positive during both normal and reversed polarity intervals, although during normal intervals the ratio is nearly a factor of 2 greater than it was during reversed polarity intervals.

The results of Williams and Fuller's modeling of the Matuyama/Brunhes reversal and the modeling of the lower Jaramillo data presented here suggest that the G_2^0 term does not change sign during the polarity transition itself. Very detailed records would be required to detect a reversal of G_2^0 during a non-transitional interval, but it may prove useful to examine sets of sequential reversals in order to learn more about the reversal frequency of G_2^0, since zonal harmonic models of transition records appear to be very sensitive to the sign of G_2^0.

Bogue and Coe [1982] presented a set of back to back reversals in the time interval 4.5 to 5.6 M.Y. from Kauai (22°N), both of which exhibit near-sided VGP paths. The inclinations pass through steep downward directions during the latter portion of the R-N transition, but pass through steep downward directions early in the N-R record. Valet and Laj [1981] obtained a set of successive Miocene reversals from Crete (35°N), which yielded a R-N record with a near-sided VGP path and a N-R transition with a far-sided VGP path. Although the VGP path for the reversed to normal reversal is predominantly near-sided, the inclination record passes through very steep negative values early in the reversal. The normal to reverse transition in turn exhibits steeply negative inclinations late in the reversal.

Modeling of these reversal sets reveals that a Williams and Fuller [1981] type model can produce synthetic inclination records that fit the general character of the observed inclination patterns. If the G_2^0 term is allowed to dominate the G_3^0 and G_4^0 terms, as in our modeling of the lower Jaramillo reversal, reasonable fits for the Kauai records can be obtained by keeping G_2^0 positive, G_3^0 negative, and G_4^0 positive for both reversals. In a similar manner the overall character of the Crete records can be modeled by letting G_2^0 be negative (maintaining a negative G_3^0, a positive G_4^0) through both transitions. The most simple interpretation of these results would be that G_2^0 remained the same sign across both reversals in each of these cases. However, the behavior of G_2^0 during the intervening non-transitional periods cannot be resolved with these data.

These modeling exercises seem to indicate that the ratio of G_1^0 to G_2^0 may change sign more frequently than was previously thought. This is supported by geomagnetic observatory records that show that this ratio may have changed sign during historical times [Nagata, 1965]. Better knowledge of the different reversal frequencies of the dipole (G_1^0) and quadrupole (G_2^0) terms may eventually contribute to an explanation for the asymmetry between normal and reversed polarity fields noted by Merrill and McElhinny [1977] when looking at time averaged fields.

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B. M. Clement and D. V. Kent, Lamont-Doherty Geological Observatory, Palisades, NY 10964

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