Magnetization of the oceanic crust: TRM or CRM?

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by

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ABSTRACT

We propose a model in which chemical remanent magnetization (CRM) acquired within the first 20 Ma of crustal evolution may account for 80% of the bulk natural remanent magnetization (NRM) of older basalts. The chemical remanent magnetization of the crust is acquired as the original thermoremanent magnetization (TRM) is lost through low temperature alteration. The CRM intensity and direction are controlled by the post-emplacement polarity history. This model explains several independent observations concerning the magnetization of the oceanic crust. The model accounts for amplitude and skewness discrepancies observed in both the intermediate wavelength satellite field and the short wavelength sea surface magnetic anomaly pattern. It also explains the decay of magnetization away from the spreading axis, and the enhanced magnetization of the Cretaceous Quiet Zones while predicting other systematic variations with age in the bulk magnetization of the oceanic crust. The model also explains discrepancies in the anomaly skewness parameter observed for anomalies of Cretaceous age. Further, our study indicates varying rates of TRM decay in very young crust which depicts the advance of low temperature alteration through the magnetized layer.

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INTRODUCTION

Since the advent of the theory of seafloor spreading, studies of the oceanic magnetic anomaly pattern have concentrated on determining the plate tectonic history of ocean basins, and characterizing the source layer for the anomalies. Results from the Deep Sea Drilling Project have extended our knowledge of the properties of the oceanic crust. These results have shown that CRM should be an important component of the magnetization of the oceanic crust (Smith and Banerjee, 1986; Smith and Banerjee, 1985a and b). Furthermore, the DSDP results support systematic regional variations in the intensity of magnetization of upper crustal basalts. The addition of a satellite magnetic anomaly field to the worldwide sea surface magnetics data base has also enhanced our ability to recognize and characterize systematic variations in the oceanic magnetic anomaly pattern. In this paper we will present evidence that links regional variations in crustal magnetization to low temperature alteration of the crust.

The remarkable ability of the Vine-Matthews-Morley hypothesis to explain the existence of marine magnetic anomalies and associated geophysical signatures has resulted in an almost unquestioning acceptance of its tenets by the earth science community. Twenty plus years of investigation have yielded firm support to the idea that magnetic anomalies on the seafloor are created as the magnetic minerals in the crust cool through their blocking temperature spectra and freeze in the direction of the Earth's magnetic field. This process is supposed to occur continuously and therefore provide a high resolution record of geomagnetic field history. However, a large body of data returned by IPOD, DSDP and ODP drilling suggests that the magnetic structure of the oceanic crust is complex, requiring modifications to the theory in order to account for the evolution of the crust after emplacement.

In this paper, we will consider the effects of crustal alteration on the magnetic signature of the oceanic crust. The principal mechanism for crustal alteration is most likely hydrothermal circulation which is most active in crust younger than 10 Ma (e.g. Staudigel and Hart, 1985; Richardson **et al.**, 1980). It has been shown that oxidation/reduction of magnetic minerals during crustal alteration causes changes in the intensity and direction of the magnetic vector (e.g. Hunt **et al.**, 1986; Bailey and Hale, 1981; Marshall and Cox, 1971). We have investigated the effects of chemical alteration on the original thermoremanent magnetization by means of a model which simulates the acquisition of secondary components of magnetization as the crust ages and alters in composition.

The duration of hydrothermal circulation is on the order of several million years. Therefore, the process of crustal alteration is a long period phenomenon which should affect the intermediate wavelength magnetic field. The Magsat satellite magnetic field has provided an opportunity to study large scale magnetization distributions which reflect pervasive processes that occur during the creation and subsequent aging of the ocean crust. In our previous paper (LaBrecque and Raymond, 1985) we have shown that the principal intermediate wavelength magnetic anomalies seen in the Magsat field over the North Atlantic arise from a magnetization distribution created by the seafloor spreading process. Furthermore, the amplitude of the satellite anomaly field suggests that the magnetization of the crust is more intense in the Cretaceous Quiet Zones, an observation previously alluded to in studies of the sea surface anomaly pattern (Vogt and Einwich, 1979). The large skewness discrepancy observed at the sea surface for anomalies 33 and 34 (Cande and Kristoffersen, 1977) is also observed within the Cretaceous Quiet Zone (KQZ) of the satellite field. These observations of regional variations in crustal magnetization led us to suggest a process of crustal alteration to explain amplitude and skewness variations observed in all the ocean basins.

We now relate three observations concerning the magnetization of the crust to low temperature alteration in young oceanic crust. These are: (1) decay of magnetization with distance from the spreading center , as seen in the NRM intensity and magnetic anomaly amplitudes, which implies a relationship between age and the intensity of magnetization; (2) observations of skewness discrepancies for many polarity intervals, and (3) a more complex variation of magnetization with respect to age in the oceanic crust identified by Bleil and Petersen (1983) from a compilation of NRM intensities of DSDP basalts.

The Model.

The mechanism we propose to account for anomalous skewness and amplitude variations is the acquisition of secondary components of magnetization during the first 20 m.y. as the oceanic crust ages and undergoes alteration. This secondary magnetization is a chemical remanent magnetization. Viscous remanent magnetization is not a likely contributor to anomalous skewness or amplitude variations because its very short relaxation time would result in a unidirectional overprint (Kent and Lowrie, 1977; Dunlop and Hale, 1977). As the crust ages, the primary thermal remanent magnetization decays as the secondary magnetization is acquired. This is accomplished by the alteration of stoichiometric titanomagnetite to cation-deficient spinel phase (titanomaghemite), phase splitting of titanomaghemite to Fe-rich titanomagnetite and titano-ilmenite, or by the growth of secondary magnetic phases. We propose that the secondary magnetization is acquired over a period of several million years **in the direction of the ambient field.** The intensity and direction of the secondary magnetization are determined by the polarity bias of the field during its acquisition. This secondary overprint distorts the original TRM distribution and manifests itself as a component of anomalous skewness.

The acquisition of secondary components of magnetization is modeled as an exponential growth process. Its complement (exponential decay) represents the decay of the original primary thermoremanent magnetization. The secondary components are acquired at the expense of the primary magnetization, therefore the rate of decay of the original remanence is proportional to the rate of growth of the secondary remanence. The process is shown in Figure 1.

The choice of an exponential function to describe the alteration process is supported by field observation sand laboratory analysis. The decay of TRM with age in the oceanic crust has been shown to be an exponential process (e.g. Irving, 1970; Ozima, 1971; Atwater and Mudie, 1973; Johnson and Atwater, 1977; MacDonald, 1977). Chemical reactions which which can be described with an exponential growth/decay function are termed first order and pseudo-first order reactions and they are governed by rate constants expressed in units of 1/time. In particular the alteration of minerals in aqueous solution is controlled by a steady state process in which elementary first-order reactions combine to produce an overall pseudo-first order reaction (Lasaga,1981).

The rate of alteration and consequently the rate of CRM acquisition must be controlled by the concentration of unaltered materials and ions in solution as well as the ambient temperature. Deviations from the model rates will occur due to variations in rate controlling factors such as diffusion rate or temperature. However in the bulk processes which we are considering here it is simplest to assume that the overall reaction will be first order.

Experimental data obtained in laboratory experiments conducted on dredged submarine basalts by Johnson and Merrill (1973) shows a general first order behavior for the acquisition of CRM during oxidiation of the samples (Figure 2). The process was shown to be strongly temperature dependent. These experiments, which were conducted on samples of oceanic crust, are good phenomenological evidence for CRM acquisition as an exponential growth process. Furthermore, the dependence of decay of anhysteretic remanent magnetization (ARM) on the growth of CRM indicates that both phenomena arise from a single process.

Therefore, as a starting point, we will consider the growth of CRM to be exponential and directly related to the exponential decay of the primary TRM component.

For simplicity we have modeled the geomagnetic field history as a scalar variable which varies in a binary fashion (reversed and normal polarities). In this case, the CRM alteration function can be convolved with the geomagnetic reversal time scale to calculate the acquisition of secondary components of magnetization with age. The CRM distribution is superimposed on the bulk magnetization while the original TRM decays exponentially with a characteristic time constant. The sum of these processes results in a new magnetization distribution which describes the effects of alteration:

$$A(T) = M_{trm}\, R(T)\, [1+Pe^{-(T/\lambda)}] + (M_{crm}/\lambda) \sum_{t=0}^{T} R(t)\, e^{-[(T-t)/\lambda]} \Delta t$$

where $A(T)$ is the resultant magnetization distribution as a function of the age of the crust (T).

$R(T)$ is the geomagnetic polarity time scale expressed as a normalized square wave.

M_{trm} is the bulk intensity of surviving thermoremanent magnetization after alteration.

M_{crm} is the bulk intensity of chemical remanent magnetization acquired in a constant ambient field.

P partitions the surviving and transient TRM

The first term in the equation represents the decay of the primary TRM. P/(P+1) is the fractional part of the initial TRM which is lost to the alteration process. TRM decays to an asymptotic value of 1/(P+1) of its original intensity. We chose P=5 for our calculation, resulting in a reduction to 16% of original intensity, a value we determined from examining published results of the the decay in magnetization of dredged axial basalts and inverted near bottom profiles.

The second term characterizes the acquisition of CRM. The geomagnetic field during acquisition is represented by R(t); the final direction and intensity of the secondary magnetization for crust of age T will therefore be a function of the post emplacement polarity history. A random distribution of polarity intervals during the CRM acquisition will result in no significant component of CRM. Conversely, the resultant CRM component acquired during long periods of constant polarity such as the Long Cretaceous Normal will be significant.

The alteration is a function of two parameters, the decay constant (λ) and the ratio of CRM to TRM. The ratio represents the relative contribution of CRM acquired during a constant field to the remaining TRM. If the secondary CRM is acquired during a single long polarity interval then it will achieve a magnetization intensity equal to the fraction M_{crm} of the original intensity of TRM. Similarly, TRM decays to the fraction M_{trm} of the original intensity. If the acquired CRM component is random in direction then the effective bulk magnetization is 100% TRM, but the intensity of the TRM relative to the initial value is M_{trm}. Therefore, the bulk intensity of magnetization employed in the model varies from a minimum of M_{trm} of the original intensity to M_{trm} plus M_{crm} of the original intensity, depending on the coherence of the acquired secondary CRM. The ratio of M_{crm} to M_{trm}

acts as a gain on the amplitude modulation of the resultant magnetization distribution.

The decay constant controls the degree of distortion or skewness of the magnetic anomaly pattern. In Figure 3 a suite of profiles illustrates the phase shift produced by different time constants for anomalies 25 to 34 at Pacific (3a) and Atlantic (3b) spreading rates, and the sequence M0-M15 in the Pacific (3c) and M0-M29 in the Atlantic (3d). A peak in the degree of distortion is reached for a time constant of 1 m.y. This is because most polarity intervals are longer than 1 m.y. and therefore most of the secondary magnetization is acquired during the subsequent oppositely directed polarity interval for crust created near the polarity transition. By slowing down the alteration process and therefore increasing the decay constant, the probability of sampling a random distribution of polarities during acquisition of the secondary magnetization increases and the phase discrepancy becomes localized at the younger boundaries of anomalies formed prior to polarity intervals of long duration. At a time constant of 10 m.y., the secondary components of magnetization are nearly always random and the dominant effect is diminished amplitude. A longer decay constant has the effect of reducing the resultant CRM component in most cases with a resultant reduction in amplitude of the anomalies. Therefore, as the decay constant increases, higher initial intensities must be invoked to produce the observed amplitudes of anomalies on the ridge flanks.

For time constants greater than 5 m.y., the CRM acquired during the Late Cenozoic is very small because rapid reversals prohibit a coherent direction for secondary magnetization; the overall effect is a reduction in the bulk intensity of magnetization during this interval, without any significant phase shift. This is generally what has been observed in young crust. The total magnetization in the model reaches a maximum in the Cretaceous Quiet Zone (KQZ), when the Earth's magnetic field remained in a normal polarity for approximately 26 Ma. Smaller peaks are observed at anomalies 13 and 25 (Late Eocene-Early Oligocene and Mid Paleocene), due to a strong reversed bias in the field during Chron C12R and Chron 24R to C26R. As we can see the growth of CRM as a function of the ambient field can produce intensity modulations of very long period.

The ratio of CRM to TRM (M_{crm} to M_{trm}) also controls, to a lesser degree, the amount of phase shift introduced by the model, by varying the relative amount of undistorted vs. distorted magnetization input to the anomaly calculation. Increasing the ratio also intensifies the amplitude modulation. Figure 4 demonstrates that ratios of CRM to TRM of less than 1:1 produce very little amplitude and phase modulation of the anomaly pattern; at 100% CRM the reversal distribution is severly distorted.

Quantifying the Alteration Model.

We estimated the relative contributions of TRM and CRM as well as the process rate by modeling sea surface and Magsat satellite profiles and matching variation in magnetization with age of geologic samples from dredges and DSDP drill sites. We find a decay constant of 5 m.y. and a ratio of CRM:TRM of 4:1 produces the best fit between the alteration model and observations. The model employs a source layer of 1 km thickness whose depth varies with the square root of age. The initial spreading center magnetization intensity is 8.7 A/m (normalized to equatorial intensity, 12 A/m at 30°). This initial value reflects both the highly magnetized surficial layer and less intensely magnetized intrusive layers. The original remanence (TRM) decays to 16% of the initial value while the CRM will attain 64% of the original TRM intensity if acquired in a constant field, yielding a ratio of CRM:TRM of 4:1. Therefore the maximum intensity possible away from the ridge axis is 80% of the initial TRM value, or 7.0 A/m at the equator. In the event of a randomly oriented CRM, the value drops to 1.4 A/m at the equator, which represents the surviving TRM. The one kilometer thickness choosen for our model source layer can be varied, however the magnetization intensity must change in proportion to the layer thickness to maintain a constant effective magnetization. For example, doubling the thickness of the source layer would require halving the magnetization.

Decay of Magnetization with Crustal Age

It has been recognized for some time that the amplitude of marine magnetic anomalies falls off sharply with distance from the spreading center. This effect has been studied quantitatively by examining the magnetic properties of dredged samples from the rift valley and adjacent flanks (i.e. Irving, 1970; Irving **et al.**, 1970; Prevot **et al.**, 1979; Johnson and Atwater, 1977) as well as near bottom studies (MacDonald, 1977; Klitgord **et al.**, 1975). Irving **et al.** (1970) produced a decay curve based on the intensities of dredged basalts from the Mid-Atlantic Ridge at 45°N which showed a rapid decrease within 1 Ma from a peak value between 50 and 100 A/m to approximately 10 A/m, and a slower decay to a steady state value of 5 A/m on the outer flanks. Klitgord et al., (1975) inverted near bottom magnetic anomaly profiles from six ridge segments in the East Pacific and also found a rapid decrease in magnetization within 1 Ma or less, followed by a gradual decrease over older crust to values less than 10 A/m. Macdonald (1977) determined a half-life of 0.6 Ma for the magnetization decay by inverting near bottom magnetic profiles from the Mid-Atlantic Ridge at 37°N. Johnson and Atwater (1977) report a similar decay curve for drilled and dredged samples from the same area, adding that the magnetization levels off to a value about one-fifth of the peak value.

The steep linear decay in magnetization at the axis can be attributed to low temperature oxidation of the very fine grained highly magnetized exteriors of freshly extruded basalts which are the most vulnerable to contact with seawater and therefore rapidly altered (Watkins et al., 1970; Klitgord, 1976). The coarser grained pillow interiors, together with massive basalt units, are less accessible to the penetration of cold seawater, so it may be conjectured that these materials will take longer to feel the effects of the alteration. Indeed, a more gradual decay is apparent for crust older than 1 Ma. Harrison (1981) examined all the existing data on crustal magnetization from direct sampling and inversions of near bottom profiles and arrived at the distribution shown in Figure 5 (symbols). In Figure 5 we have applied our alteration model with a decay constant of 3 m.y., and a starting intensity of 10 A/m, decaying to 16% of the original intensity without any CRM acquisition (dashed line). Also shown superimposed on the observed magnetization values is an estimate of total magnetization calculated for this decrease in TRM, accompanied by the acquisition of a secondary CRM acquired with a growth rate of 3 m.y. (crosses). We have not included the high and variable magnetization intensities observed at the spreading axis in our consideration of the magnetization decay. The intense high localized at the ridge axis, which reaches amplitudes of 2000 nT in the sea surface magnetic anomaly pattern (FAMOUS area, 37°N) and 100 A/m in NRM intensity of basalt (45°N, Irving, 1970) is not predicted by this decay function, but appears to be controlled by characteristics of the accretion process. The 3 m.y. exponential decay curve predicts intensities that are greater than observed for crust less than 3 Ma old. We believe that this fit indicates that the decay constant characterizing the alteration process is becoming larger as alteration of the more easily weathered surface material goes to completion and the chemical alteration proceeds in deeper and hotter regions of the crust, where presumably the process is quite different than at the surface. In crust older than 3 Ma the decay is characterized by a constant greater than 3 m.y.

A great deal of information on the gradual decay of bulk magnetization of the entire source layer can be gleaned from the sea surface marine magnetic anomaly pattern. We find that the decay of magnetic anomaly amplitudes is best described by an exponential decay function with a decay constant of approximately 5 m.y.

Figure 6 displays the amplitudes of selected anomaly profiles across the Australian-Antarctic and the Pacific-Antarctic spreading systems for crust younger than 20 Ma (anomalies 1 through 6). These profiles were chosen because they display a clear anomaly sequence, with a well defined central anomaly. The profiles have been normalized with respect to the Brunhes or central anomaly amplitude. This analysis is similar to a study by Weissel and Hayes (1972) of anomaly amplitudes on the Australian-Antarctic ridge. Our method of normalizing each profile instead of averaging absolute amplitude gives somewhat different results than those of Weissel and Hayes (1972). These and many other profiles display the gradual decay in anomaly amplitudes with increasing crustal age. Forward modeling of this decrease in anomaly amplitudes reveals an exponential decay in bulk crustal magnetization with a time constant of 5 m.y., as shown by the crosses in Figure 6. The model was calculated for a source layer whose depth increases as the square root of age following the relationship of Parsons and Sclater (1977). The jagged nature of the model curve is a natural consequence of the variable lengths of the polarity intervals.

We find that we can characterize a suite of observations on magnetization decay which manifest differences in crustal alteration rates. The variation in alteration rates are most likely controlled by the depth and temperature of the magnetic source layer. These observations are: 1) the rapid decay of magnetization of fresh axial basalts which occurs very rapidly, with a decay constant of less than 1 m.y., confined spatially to the the spreading center, as identified in previous studies; 2) decay of magnetization of the near surface layer with an average time constant of 3 m.y. which we have identified from a published compilation of magnetization values of samples dredged away from the spreading center and inversions of magnetized topography; and 3) decay of bulk magnetization of the entire source layer with a time constant of 5 m.y. as inferred from the decay of magnetic anomaly amplitudes. These variations in alteration rate are probably depth dependent with the faster rates manifesting the alteration of the surficial oceanic crust.

It is reasonable to assume that there will be some delay before the alteration front pervades the entire magnetic source layer. The gradual decay in bulk magnetization which we model with a 5 m.y. time constant is meant to represent the combined contribution of shallow sources and the deeper layers such as massive basalt, as well as larger grain sizes within the upper (pillow basalt) layer (Marshall and Cox, 1972). Alteration of these materials may be delayed by a lag in the time required for cracks to propagate into the interior, hotter regions of the crust, or as a result of a diffusion process.

Skewness

The shape of marine magnetic anomalies can be quantified by their skewness, or shape asymmetry. The skewness parameter describes the distortion of a symmetric square wave pattern caused by the relative directions of the inclination and declination of the ambient field versus the inclination and declination of the remanent magnetization, as well as the strike of the magnetic lineation. Skewness is mathematically equivalent to a phase shift of the spectral components which comprise the anomaly pattern (Schouten and McCamy,1972; Schouten and Cande, 1976). When an anomaly is reduced to the pole, it is phase shifted to compensate for its theoretical skewness and thus represents the anomaly produced by a particular body in a vertical (polar) magnetic field. In many cases, marine magnetic profiles contain a component of anomalous skewness, which is that asymmetry remaining after the anomalies have been reduced to the pole. Anomalous skewness has been observed in the Neogene and Paleogene anomaly sequence (anomalies 5-20) south of Australia by Weissel and Hayes (1972); in the North and South Pacific Late Cretaceous- Early Tertiary anomaly sequence (27-32) by Cande (1976); for anomalies 33 and 34 in the North Atlantic by Cande and Kristoffersen (1977), the South Atlantic by Cande (1978) and the Agulhas basin by LaBrecque and Hayes (1979), and for anomalies 20-24 in the Philippine Sea (Watts et al., 1977).

Any model which attempts to account for anomalous skewness must satisfy two observations : 1) the amount of skewness discrepancy varies in different ocean basins for a given polarity interval, but is constant when normalized to spreading rate (Cande, 1978), and 2) The degree of anomalous skewness varies with the polarity interval considered; e.g., a large amount (35-45 degrees) of anomalous skewness is observed for anomalies 33 and 34 but the Early Cretaceous M-sequence (M0-M4) displays little or no anomalous skewness (Cande, 1978; Cande and Kent, 1985).

Three hypotheses have been advanced to explain anomalous skewness. These are: 1) rotation of the crust along outward dipping (listric) faults (1^{o} of rotation equals 1^{o} of anomalous skewness) (Cande and Kent, 1976) 2) a two layer model of the magnetic source layer in the oceanic crust in which magnetization is rapidly acquired in the upper layer and slowly acquired along sloping isotherms in the lower layer (Cande and Kent ,1976) 3) a field behavior model in which intensity variations and undetected short reversals near polarity boundaries produce asymmetry in the anomaly pattern (Cande, 1976; Cande, 1978). Cande (1978) favors the field behavior model because it accounts for the variation in skewness between different polarity intervals; however, there is no direct evidence for such field behavior. The two layer model is attractive because it is spreading rate dependent. Evidence from ophiolites (Levi and Banerjee, 1977; Butler et. al, 1976) and transition zone widths (Blakely, 1976 and 1983) suggests the existence of a deeper source layer for marine magnetic anomalies. However, the two layer model cannot explain the absence of anomalous skewness for the younger M-sequence anomalies while the anomaly 33-34 sequence consistently displays over 35 degrees of anomalous skewness. Similarly, tilting cannot fully explain anomalous skewness because it would require that the amount of rotation vary by over 35^{o} within the Cretaceous from M0 to 34 time.

Amplitude

An important observation concerning intensity variations in the ocean crust is that of Bleil and Petersen (1983). Their compilation of the natural remanent magnetization (NRM) intensity of all DSDP basalts normalized to an equatorial intensity displays a strong variation in magnetization intensity with respect to time (see Figure 5(a). The curve shows the rapid decay of NRM with age from approximately 8 A/m at the spreading center to a minimum of 1 A/m at 20 Ma, followed by a steady increase to a peak of 4 A/m in the KQZ which falls off at 120 Ma.

The increased magnetization of Mid Cretaceous basalt samples agrees with observed amplitude variations in the marine anomaly pattern over the KQZ. We have observed high amplitude anomalies within the KQZ in the Magsat field of the North Atlantic, which required magnetization intensities 1.5 that of the Cenozoic sequence to model. Though the sea surface data set lacks coherent lineations within the quiet zone, high amplitude quasi-lineated magnetic anomalies are characteristic of the KQZ in the sea surface field (Hayes and Rabinowitz, 1975; Vogt and Einwich, 1979). In fact, the Cretaceous Quiet Zones are notoriously noisy (Dickson, 1968). This noise has been attributed to highly magnetized crustal relief or intrabasement contrasts by Vogt and Einwich (1979). These observations together constitute strong evidence that a zone of enhanced magnetization extends throughout the KQZ.

Bleil and Petersen (1983) explain the variation in NRM intensity by progressive low-temperature alteration in the crust. Their model considers maghemitization the dominant process for the first 20 Ma, followed by inversion to stable magnetite which continues at a constant rate thereafter and results in an increase in intensity of magnetization. The progressive increase in NRM for the older crust requires that inversion to stable magnetite preserves the original NRM direction. However, as Bleil and Petersen point out, the inversion model cannot account for the decrease in magnetization observed for crust older than 120 Ma.

Comparison to DSDP results

A plot of magnetization intensity vs. age computed from the alteration model is presented in Figure 7 along with the curve of Bleil and Petersen (1983) of NRM intensities of DSDP basalts vs. age. The model curve predicts a minimum in intensity at 20 Ma, and a steady increase in intensity towards a high intensity zone in the Cretaceous Long Normal interval. The intensity remains high for anomalies M0 and M1, then drops off sharply to normal values for anomalies older than M1. High amplitude zones are also predicted at anomalies 13 and 25 (35 and 60 Ma, respectively). There is striking correlation between the intensity modulation observed in sampled oceanic basalts and the alteration model curve. The slower decay of the Bleil and Petersen curve after the Cretaceous Long Normal most likely is caused by uncertainties in the age of the older basalts. The model curve intensity is slightly greater than the observed curve; this is because we are using an equivalent magnetization in a thin source layer to represent the contribution from the entire source layer. The DSDP results, with the exception of hole 504B, sample the upper few hundred meters of the crust, where seawater circulation is most vigorous. As a result, these values are biased towards the most highly oxidized material. Comparison of these two curves demonstrates that the intensity variation with age observed in oceanic basalts can be explained by a model in which alteration tapers off at 10 Ma and ceases after about 20 Ma, which agrees with the results of Staudigel and Hart (1985), from isotopic age dating of alteration products, that the duration of hydrothermal alteration is limited to 10-15 Ma. It also provides a satisfactory explanation for the decrease in intensity after 120 Ma, without invoking a change in the stability of the magnetic phases as was suggested by Bleil and Petersen (1983).

Comparison with the Magsat field

In Figure 8 profiles from the North Atlantic Magsat anomaly field sampled along flow lines are presented along with model profiles calculated at satellite elevation for a standard seafloor spreading model (at 30^{o}N) and for the alteration model. The standard model (.5 km thick layer and 10 A/m intensity) and Magsat profiles were discussed previously by LaBrecque and Raymond (1985). Discrepancies in phase and amplitude between the standard seafloor spreading model and the Magsat profiles occur, which were modeled by adding a phase shift symmetrical about the spreading axis of approximately 35^{o} , as well as an increase in magnetization intensity to 15 A/m, for the crust of the KQZ. However, the alteration model accounts for these variations in amplitude and phase in a systematic fashion, without appealing to undocumented fluctuations in geomagnetic field intensity or systematic crustal rotations. The dominant KQZ anomalies generated by the alteration model agree well in amplitude and phase with the Magsat profiles, with a few exceptions. The phase of the Magsat anomalies on Lines A and B is shifted to the west in relation to the model profiles. The correlation worsens considerably towards the south on the eastern side. However, the low geomagnetic latitude and presence of the Cape Verde platform, an anomalous feature attributed to a thermal anomaly in this region, suggest the poor correlation may be due to the interference of other tectonic processes. Application of the alteration model significantly enhances the correlation between the intermediate wavelength seafloor spreading profiles and the Magsat profiles.

Skewness results

In Figure 9 the alteration model profile at 30°N in the North Atlantic is phase shifted to illustrate the skewness discrepancy produced for anomalies 27-M4. When the model profile is reduced to pole (phase shifted by the amount of its theoretical skewness) it is evident that a significant component of anomalous skewness is present for anomalies M3, 34, 33, 32, and 31. Anomalies M4, M2-M0, 30, and 29-27 do not exhibit a clear departure from a square wave appearance. The magnitude of the skewness discrepancy varies from 35° for M3, 34, and 33 to approximately 20-25° for 32 and 31. Thus, the model is consistent with the fact that anomalous skewness varies in magnitude for different polarity intervals in which it is identified, and also predicts the large skewness discrepancy at anomalies 33 and 34. Since the model predicts little or no anomalous skewness for the late M sequence, except for M3, it provides a satisfactory explanation of the marked difference in skewness at either end of the KQZ. The exception is anomaly M3, which is skewed by the alteration function, and as will be discussed later, may or may not display a component of anomalous skewness in the sea surface field.

The amount of skewness produced by the alteration function was shown to depend on spreading rate (Figure 4). This occurs because the function produces a wavelength dependent phase shift as opposed to a constant phase shift. Therefore, the skewness produced for M3 at North Atlantic spreading rates (<1 cm/yr) is not prominent at Pacific spreading rates (4 cm/yr), where most studies of the M-sequence have been conducted (e.g. Larson and Chase, 1972) In fact, the sequence M0-M15 predicted by the alteration function for Pacific spreading rates and a time constant of 5 m.y. does not display a large enough component of skewness that it could be verified in real data. Furthermore, the very slow spreading rates in the North Atlantic for the Early Cretaceous result in a poor recording of the anomaly pattern, which confuses the identification of anomalous skewness.

Finally, the reduced to pole profiles calculated for Pacific spreading rates shown in Figure 4 reveal a small component of skewness for anomalies 27-32 (except 30) in agreement with Cande (1976).

DISCUSSION

The alteration model has proven encouraging in explaining amplitude and skewness variations in the marine magnetic anomaly pattern as well as variations in NRM intensity of basalts. The model can also shed light on some other problems concerning the magnetization of the crust.

The cornerstone of the model simulating the effects of alteration on the magnetic signature of the crust rests on the assumption that the secondary components are acquired in the direction of the ambient field. This assumption is favored by Smith and Banerjee (1985a and b). Although it is considered invalid by some authors (Petersen et al., 1979; Johnson and Merrill, 1972; Marshall and Cox, 1971; Ozdemir and Dunlop, 1985), This assumption allows us to construct a model which can systematically account for amplitude variations and anomalous skewness in the North Atlantic. The point is that although the inclination of the secondary magnetization may vary widely at a single locality because the components are acquired over a period of time longer than the average polarity epoch due to variations in coercivity and local temperature/pressure conditions, the net magnetization distribution reproduces the typical magnetic anomaly profile.

Previous studies have concluded that remanent directions are not reset by chemical alteration (Hall, 1976; Ozdemir and Dunlop, 1985; Marshall and Cox, 1971). These conclusions have been reached from study of dredged basalts (Marshall and Cox, 1971), experimental studies (Ozdemir and Dunlop, 1985) and cores recovered by DSDP Leg 37 (Hall and Ryall, 1977; Hall, 1976). DSDP sites 332 and 504B which penetrate to 583 and 1076 meters subbasement respectively, have provided the most complete record of magnetic stratigraphy in the crust. However, these sites occupy crust of age 3.5 Ma, therefore we would not expect a large stable component of CRM to be present because several reversals have occurred since crust was formed at these sites. The TRM component should dominate the magnetization vector. Bailey and Hale (1981) report results of an experimental study on Leg 37 basalts (site 332b) indicating that high temperature alteration produces a CRM which is directed along the ambient field for strong fields (50 T) and in an anomalous direction in between the remanent and ambient field direction for weak fields (20 T). It is difficult to interpret these results in terms of real conditions in the oceanic crust. Perhaps basalt sustained at lower temperatures (100°) for longer periods of time would suffer the same amount of alteration as observed in the high temperature, 24 hour experiment, or maybe this type of behavior is confined to the hydrothermally altered basalts residing at depth. In any case if an analogy can be made between this experiment and low temperature alteration (which is likely to dominate in the upper crust where the samples were obtained) then we may conjecture that anomalously shallow or steep inclinations (see for example Harrison and Watkins, 1977; Lowrie, 1979) result from alteration occurring during periods of low field intensity, which are more common during frequent reversal activity. Only a few results exist for very deep crust (i.e. Hole 504B, Iceland Research Drilling Project, Bermuda Deep Drill). These studies indicate that hydrothermally altered basaltic dikes and flows may significantly contribute to the magnetization of the crust (Smith and Banerjee, in press; Hall, 1985; Rice **et al.**, 1980).

It appears that much work remains to be done to assess the magnetic state of the lower crustal layers, and the role of factors such as grain size (S. Halgedahl, personal communication), field strength (Bailey and Hale, 1981) and temperature and oxygen environment in changing the bulk magnetic signature of the oceanic crust during alteration. It also appears that experiments on basalts recovered by the Ocean Drilling Program (ODP) will provide more valuable information than experimental studies for assessing the effects of chemical alteration.

The results of this modeling indicate the importance of CRM as a carrier of the magnetic signal in the ocean crust. If we are correct in assuming that the CRM is acquired as a result of low-temperature oxidation and hydrothermal alteration, then the half-life of the acquisition process which best accounts for observed amplitude and phase variations is an order of magnitude longer than that observed for axial near bottom profiles and direct samples. This result could be interpreted to indicate the time required for the alteration front to proceed throughout the magnetized layer(s). It is consistent with the results of isotopic age dating of alteration products performed by Staudigel and Hart (1985) which shows the duration of alteration to be 10-15 Ma.

The marine magnetic anomaly pattern has been the primary data source for the development of the geomagnetic time scale. The upper limit limit on the resolution of the recording process has been of concern as we seek more detailed geomagnetic field histories. If the major component of magnetization which remains in the oceanic crust is CRM, then why is the field recording so good and what is its maximum resolution?

Distortion of the magnetic anomaly pattern, which is particularly evident in older crust, has been interpreted as resulting from a change in the width of the transition zone as the crust ages. In the model of Blakely (1976, 1983), the smoother polarity boundaries sometimes observed in older crust are interpreted as resulting from the presence of a deeper layer with wide transition zones which gradually dominates the magnetic signal as the upper, highly magnetized layer with sharp transition zones weathers and decays in intensity.

The ability of older crust to preserve the recording of very short polarity intervals bears on the statistical structure of the geomagnetic reversal history. A measure of the resolution of the geomagnetic field recording process is the width of the transition zone which separates the magnetic polarity intervals either within the oceanic crust or within the magnetic anomaly pattern. In the alteration model, the width of a transition zone is a function of the strength of the CRM component, which in turn depends on the post-emplacement polarity history and the rate of alteration. If the variation in width of polarity boundaries is treated as a product of alteration, then all post alteration crust exhibits the same recording resolution. Smith and Banerjee (1985a and b) conclude from rock magnetic studies of DSDP basalts that the original TRM direction is reset by the acquisition of secondary CRM, and this CRM is remarkably stable. Our modeling implies that the signal recorded in all oceanic crust older than 10 Ma exhibits the same recording resolution with the possible exception of regional variations in alteration processes. The possibility that alteration rates vary with crustal depth as discussed earlier implies that the shallower crust exhibit greater fidelity due to its higher acquisition rate for CRM.

The enigmatic anomaly amplitudes associated with the KQZ in the Magsat field of the North Atlantic which provided the impetus for this study are explained by the alteration model without resorting to an increase in the main field intensity during this interval or a change in the magnetic mineralogy of the source layer. It also provides an explanation for the high amplitude noise in the sea surface anomaly profiles of the KQZ as arising from magnetized basement topography, or amplification of intrabasement contrasts, as per Vogt and Einwich (1979).

The anomalous skewness produced by the alteration model is in good agreement with observations made by Cande and Kristoffersen (1977) for anomalies 33 and 34 in the North Atlantic and LaBrecque and Hayes (1979) for anomalies 33 and 34 in the Agulhas Basin. Utilizing a time constant of 5 m.y., our model predicts a skewness discrepancy of 35° for anomalies 33 and 34, which agrees with the average value of 40° reported by Cande and Kristoffersen (1977) and 30° by LaBrecque and Hayes (1979). The results of the modeling show that no significant skewness is predicted for anomalies M0-M2, or for anomalies older than M4, in agreement with the observations of Cande (1978). However, the model predicts a significant skewness discrepancy for anomaly M3, which is contrary to the conclusion of Cande (1978) and Cande and Kent (1985) that the entire M sequence displays no anomalous skewness. Profiles published in Cande (1978) of the sequence M0-M4 in the North and South Atlantic vary greatly in their appearance. Certain profiles (V2413 and V2609) appear to exhibit the same distortion of anomaly M3 as is predicted by the alteration model. Other profiles (V2801, V3014 and especially DSDP 417, Cande and Kent, 1985) indicate that the entire sequence M0-M4 appears deskewed when reduced to the pole. A detailed examination of the shape of anomaly M3 is apparently needed to test the predictive capability of the alteration model.

ACKNOWLEDGEMENTS

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FIGURE CAPTIONS

Figure 1.: Schematic illustration of the conceptual basis for the alteration model. Remanence decays exponentially as a function of crustal age (t) with decay constant (λ). Chemical remanence is acquired as the complement of the NRM decay. In this illustration the CRM is acquired in a constant field, which results in an asymptote of 1. A random field direction during acquisition would result in a very small or zero CRM component. Bias in the polarity of the field will produce a result somewhere between these two extremes. The importance of these two processes to the magnetic evolution of the oceanic crust is explored by examining what decay constant and ratio of CRM to NRM will reproduce observational data.

Figure 2: The time dependent variation of ARM and CRM at various ambient temperatures for natural crystals of titanomagnetite from Johnson and Merrill(1973).

Figure 3.: Reduced-to-pole model profiles are presented to demonstrate the distortion of the anomaly pattern produced by varying the time constant, and the effect of spreading rate. Anomalous skewness is observed in these profiles, which is maximum for a time constant of 1 my, and barely discernible for 10 my. Comparison of the Pacific and Atlantic profiles shows that slower spreading rates produce greater distortion of the pattern. Notice the relatively undistorted shape of anomaly 30, which is the result of the random magnetization distribution of anomalies 29-27. The profiles of anomalies M0-M15 (Pacific) and M0-M29 (Atlantic) exhibit a dramatic spreading rate effect. The high Pacific rates produce a virtually deskewed profile for a time

constant of 5 my and greater, while anomalies M1-M3 appear distorted in the slow spreading Atlantic example, even for a 10 my time constant. Models incorporate a 1 km thick source layer whose depth in meters varies as 2500 + 350 SQRT(T), according to the relationship of Parsons and Sclater (1977). **Atlantic:** Initial intensity = 0.012 A/m; Remanent Inclination : 43; Remanent Declination : -29; Azimuth: 140, corresponding to a flow line crossing the North Atlantic at 30^{o}N. Variable and asymmetric spreading rates where employed to model the spatial magnetization distribution. **Pacific:** Initial intensity = 0.015 A/m; Remanent Inclination: -1.1; Remanent Declination: 9.3; Azimuth: 350, corresponding to the Phoenix lineation pattern. Spreading rates are approximate.

Figure 4. North Atlantic model (same parameters as given in Figure 3) calculated for a time constant of 5 my, with a variable ratio of CRM:TRM. Amplitude is reduced by 50% between 0:1 and 1:1, without any appreciable phase distortion. Amplitude is continually diminished by increasing the ratio, with a dramatic increase in phase distortion occurring between ratios of 2:1 and 4:1. The pure CRM profile appears distorted beyond recognition. Note the change in anomaly amplitude (vertical) scale between profiles 2:1 and 4:1. A ratio of 4:1 best fits the observations of amplitude modulation and anomalous skewness.

Figure 5.: Magnetization values from direct sampling and inversions of near bottom magnetic anomaly profiles as presented in Harrison (1981) are given along with a theoretical model for TRM decay (dashed line) and a model in which TRM decays as CRM is

acquired (fine dots). Magnetization values do not include values over 10 A/m; all values are equatorial. TRM decays exponentially to 10% of its original intensity with a time constant of 3 my. Five point averages for the CRM acquisition model, which will be discussed in this paper, are shown by crosses.

Figure 6.: Anomaly amplitudes normalized to the central anomaly amplitude are plotted for profiles from the Australian-Antarctic and Pacific-Antarctic ridges. These profiles were chosen because they formed at medium spreading north-south ridges at high geomagnetic latitude. The amplitudes of a theoretical model which was calculated for a 5 my decay constant are also given (crosses). An excellent agreement is observed between the observed data and model prediction. Modeling parameters are given in the text.

Figure 7.: Comparison of the intensity vs. age curve of Bliel and Petersen (1983) (top) derived from the NRM intensities of DSDP basalts reduced to equatorial values to the magnetization vs. age curve predicted by the alteration model at the equator (bottom). Stars on the top curve represent interval averages (approx. each 20 Ma). Crosses on the alteration curve indicate 100 point averages. Alteration curve shows intensity decrease at reversal boundaries. Both curves exhibit a minimum at 20 Ma, followed by a steady increase to a peak in the KQZ. The alteration curve shows much finer detail in amplitude modulation . It predicts anomalies M0 and M1 to be of higher intensity than the rest of the M sequence, with intensity decresing until 135 Ma, and high amplitude zones at anomaly 13 (35 Ma) and 21 (60 Ma). 1×10^{-3} G = 1 A/m.

Figure 8.:Comparison of Magsat anomaly profiles sampled along North Atlantic flow lines (solid curves) to alteration model profiles upward continued to satellite elevation generated at corresponding latitudes (dotted curves). Model parameters :

LINE A: Rem. Incl.: 34, Rem. Decl.: -28, Azimuth: 112

LINE B: Rem. Incl.: 38, Rem. Decl.: -28., Azimuth: 115

LINE C: Rem. Incl.: 43, Rem. Decl.: -29, Azimuth: 140

LINE D: Rem. Incl.: 46, Rem. Decl.: -29, Azimuth: 135

Annotations: OCB-Ocean continent boundary (ca. 200 Ma), M25-(ca. 150 Ma), M0-(ca. 108 Ma), 34-(ca. 84 Ma), AXIS-Mid-Atlantic spreading center. Standard seafloor spreading model at 30°N (.5 km thick layer, 10 A/m constant intensity) is shown for comparison to the alteration model (dashed curve).

Figure 9.: Alteration model profile calculated for 30°N is shown at the bottom of the figure and the reduced to pole profile at the top. Three phase shifted profiles are shown to illustrate the magnitude of skewness discrepancy produced by the model for anomalies 27-34 and M0-M4. Anomalies M3, 34 and 33 appear deskewed for an anomalous value of 35°, 32 - 30 for 20-25°, while M4, M2-M0, and 29-27 do not deviate significantly from a square wave appearance in the reduced to pole profile.

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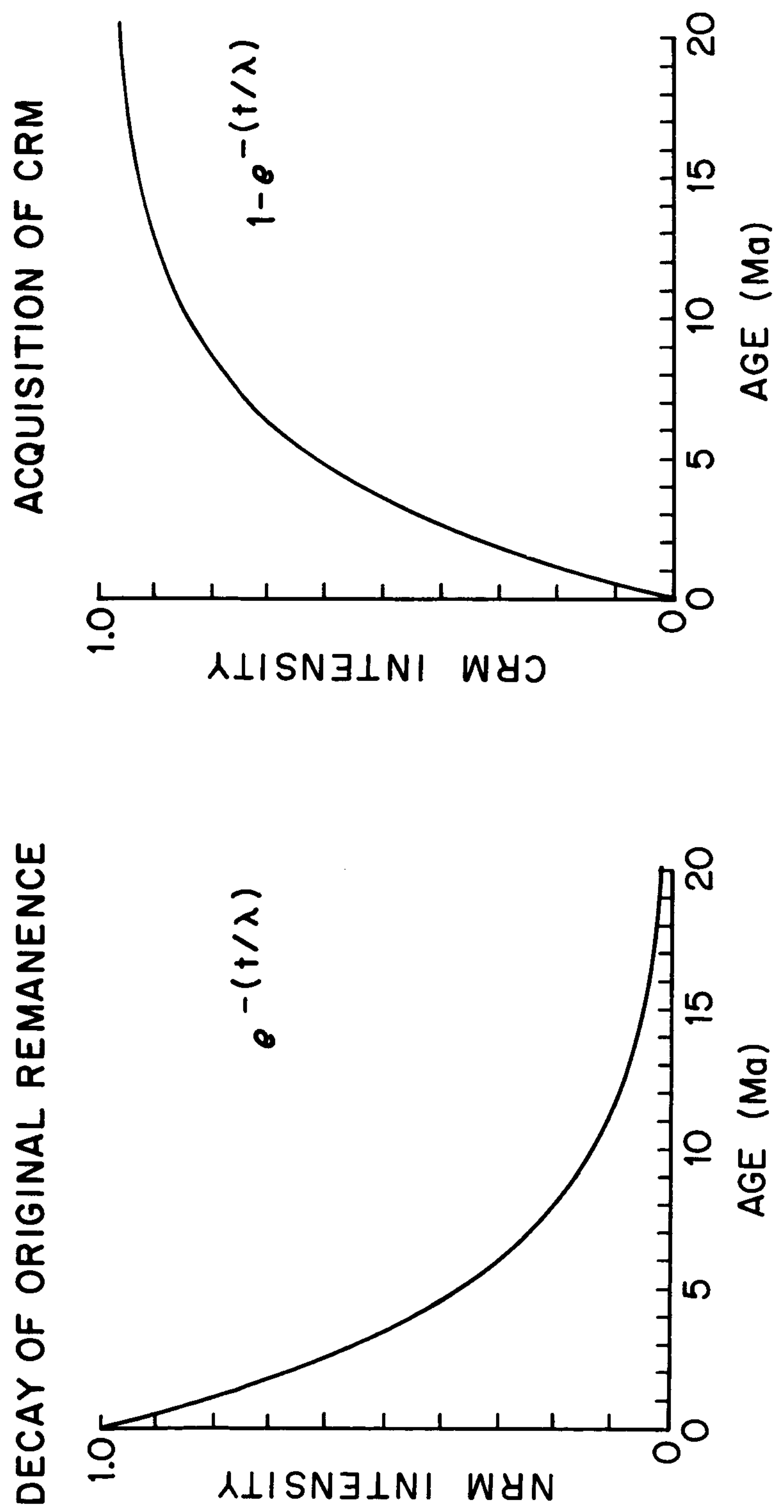
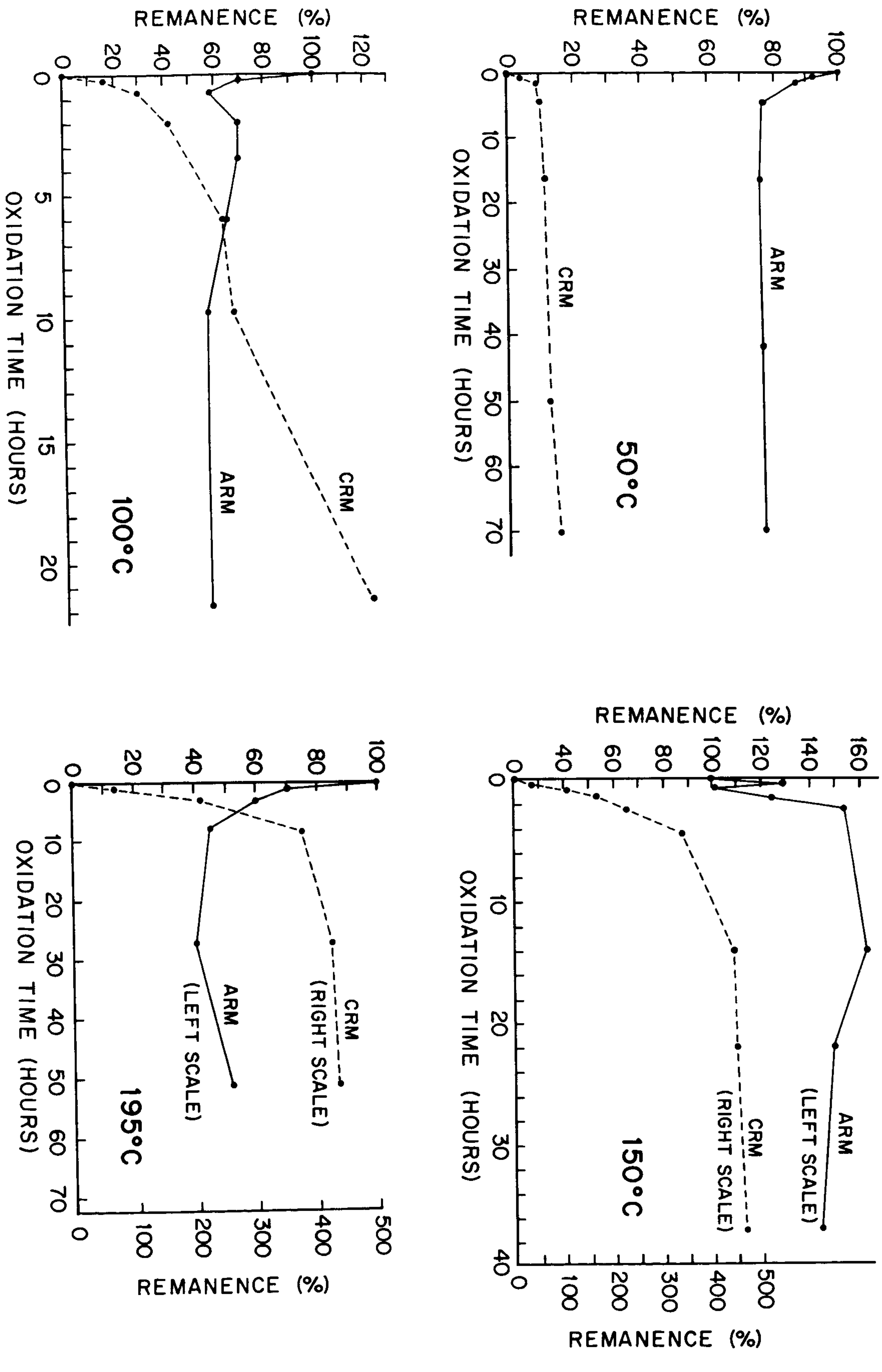

ACQUISITION OF CRM
CRM INTENSITY
1.0
0
5
10
15
20
AGE (Ma)
$1-e^{-(t/\lambda)}$
DECAY OF ORIGINAL REMANENCE
NRM INTENSITY
1.0
0
5
10
15
20
AGE (Ma)
$e^{-(t/\lambda)}$

Figure 1



REMANENCE (%)
OXIDATION TIME (HOURS)
ARM
CRM
50°C
100°C
150°C
195°C
ARM (LEFT SCALE)
CRM (RIGHT SCALE)

Figure 2

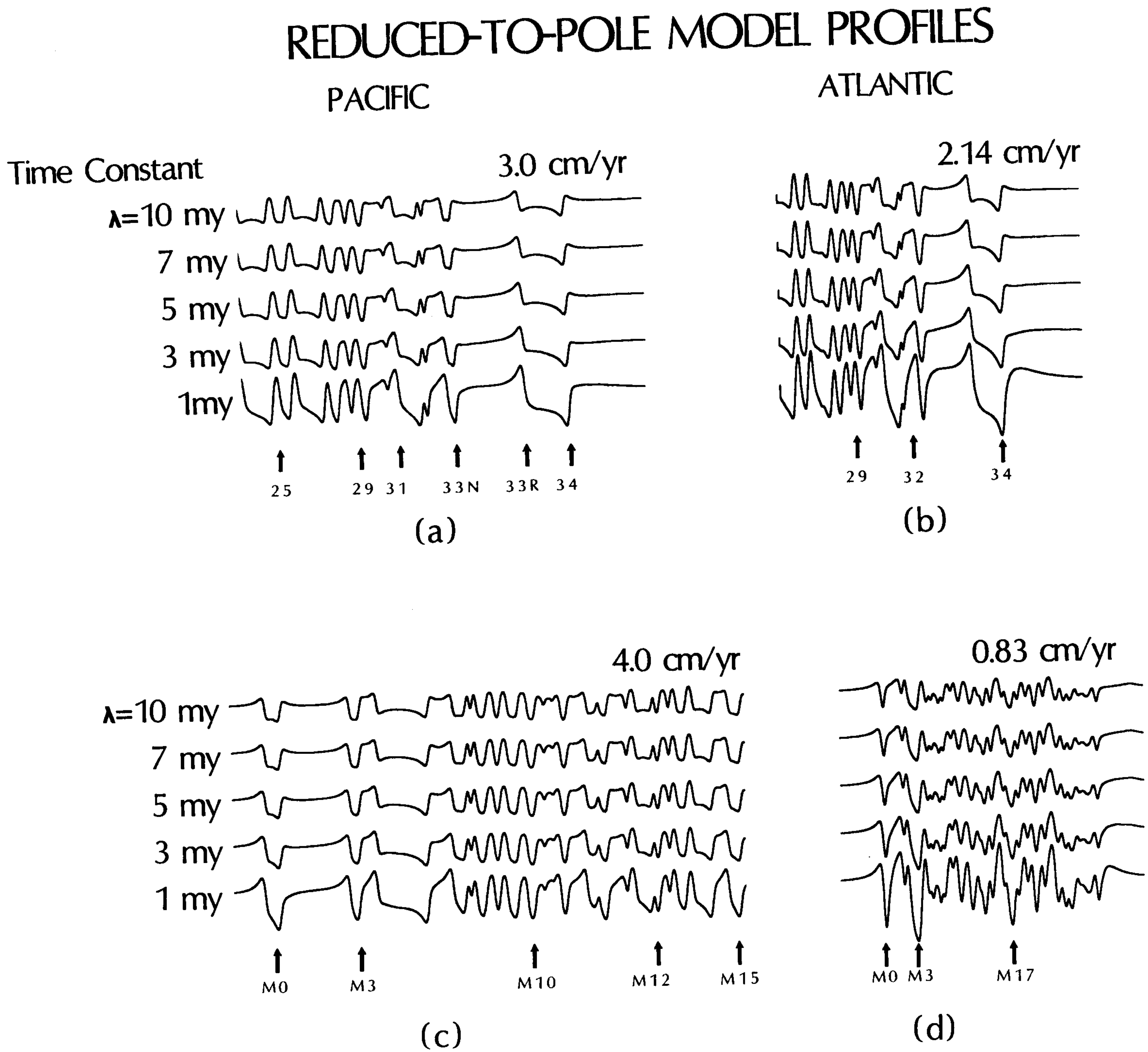

REDUCED-TO-POLE MODEL PROFILES
PACIFIC
ATLANTIC
Time Constant
3.0 cm/yr
λ=10 my
7 my
5 my
3 my
1my
25
29
31
33N
33R
34
(a)
2.14 cm/yr
29
32
34
(b)
4.0 cm/yr
λ=10 my
7 my
5 my
3 my
1 my
M0
M3
M10
M12
M15
(c)
0.83 cm/yr
M0
M3
M17
(d)

Figure 3

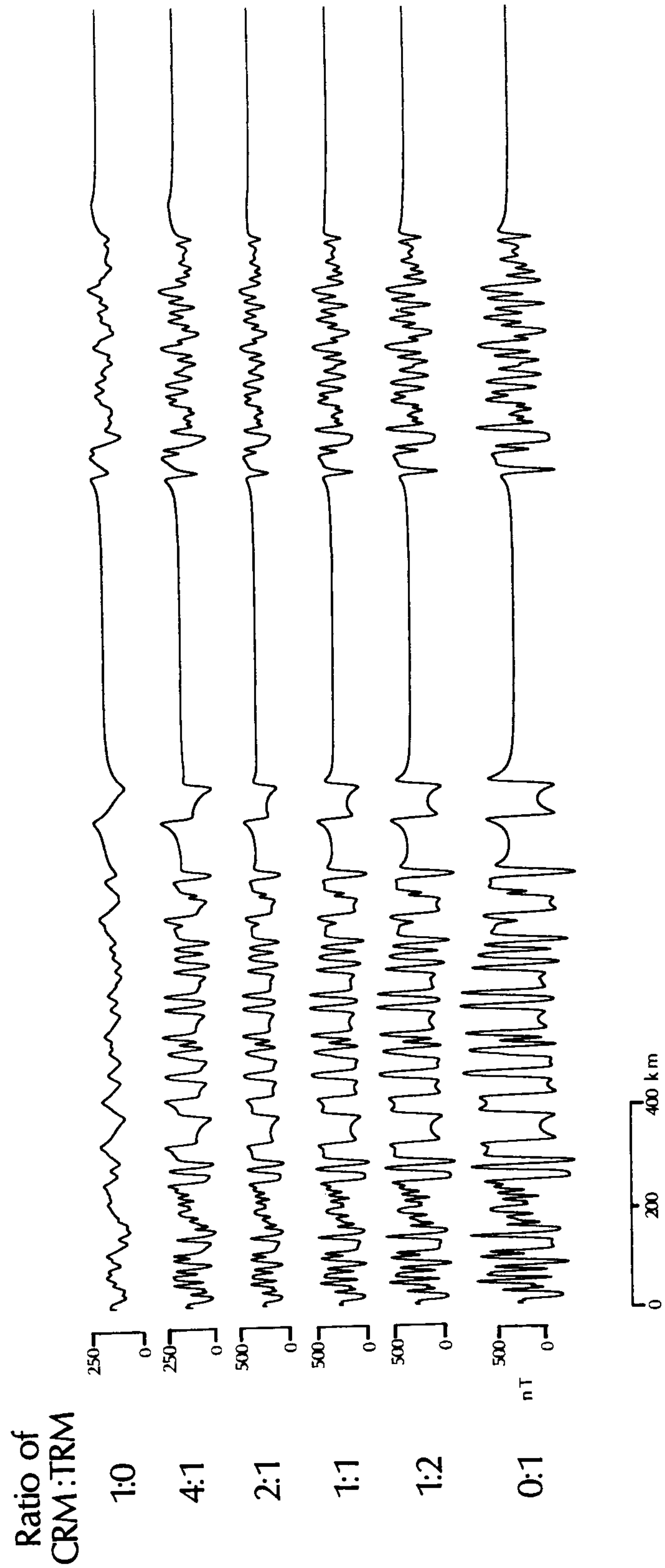


Figure 4

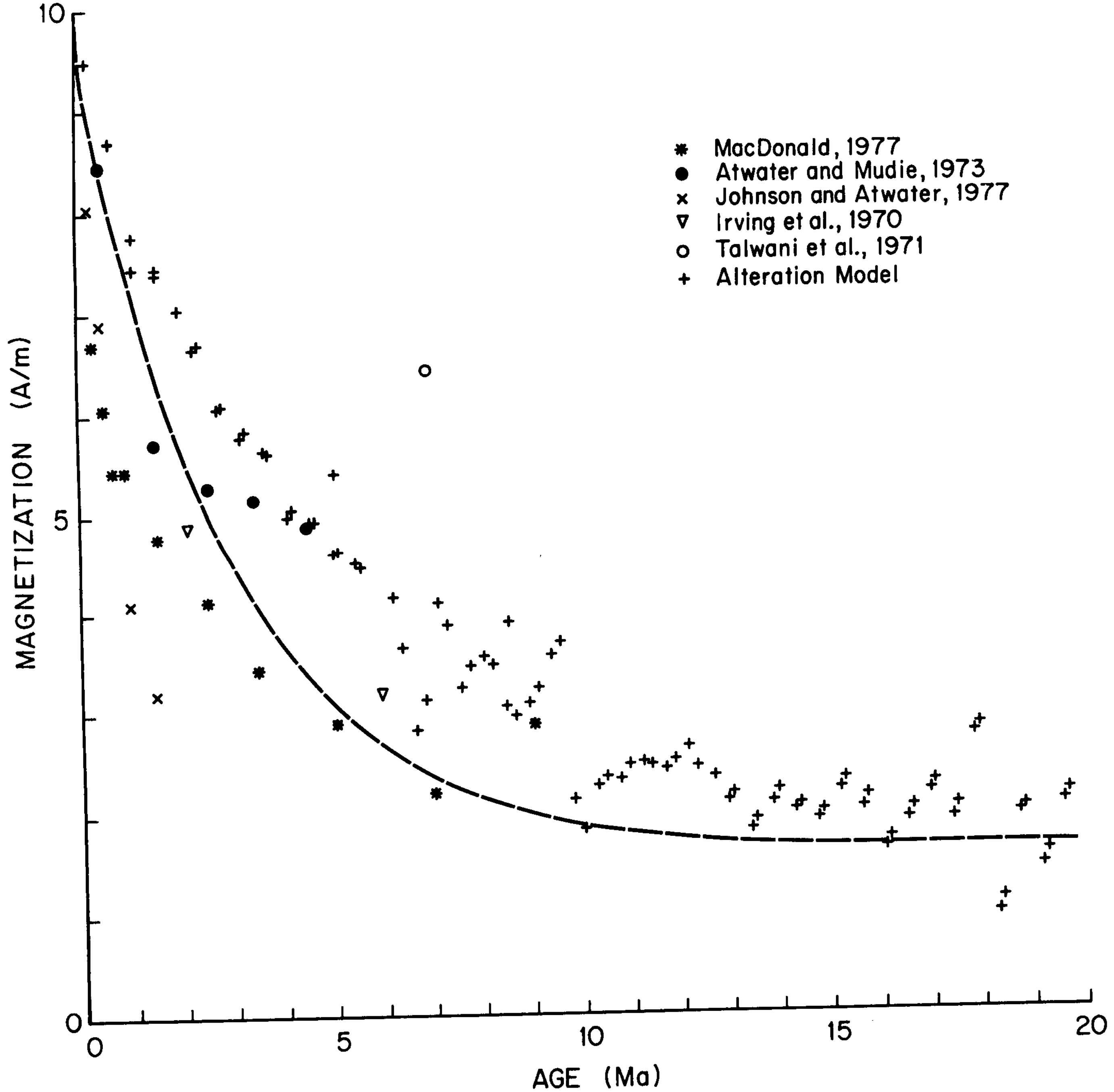

MacDonald, 1977
Atwater and Mudie, 1973
Johnson and Atwater, 1977
Irving et al., 1970
Talwani et al., 1971
Alteration Model
MAGNETIZATION (A/m)
AGE (Ma)
10
5
0
0
5
10
15
20

Figure 5

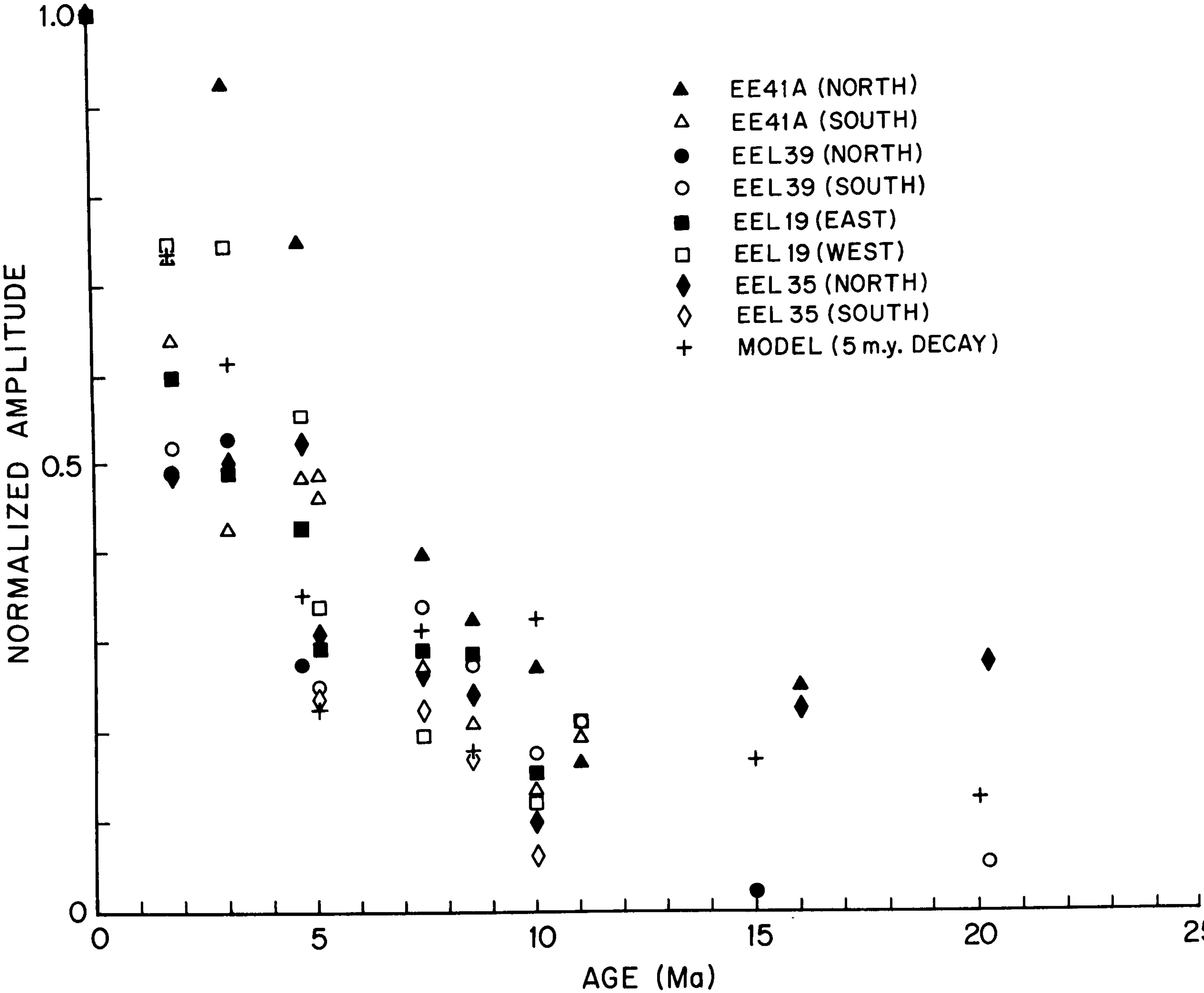
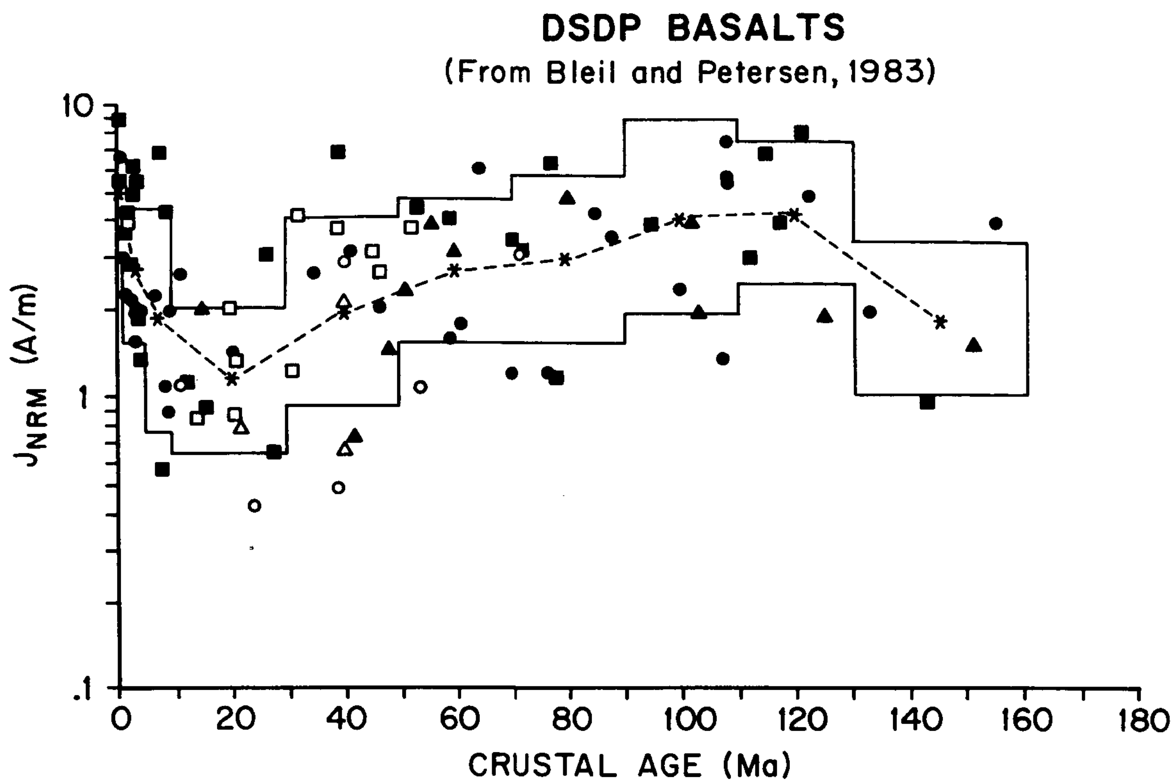

1.0
0.5
0
NORMALIZED AMPLITUDE
0
5
10
15
20
25
AGE (Ma)
EE41A (NORTH)
EE41A (SOUTH)
EEL39 (NORTH)
EEL39 (SOUTH)
EEL19 (EAST)
EEL19 (WEST)
EEL35 (NORTH)
EEL35 (SOUTH)
MODEL (5 m.y. DECAY)

Figure 6


DSDP BASALTS
(From Bleil and Petersen, 1983)
J_{NRM} (A/m)
CRUSTAL AGE (Ma)
10
1
.1
0
20
40
60
80
100
120
140
160
180

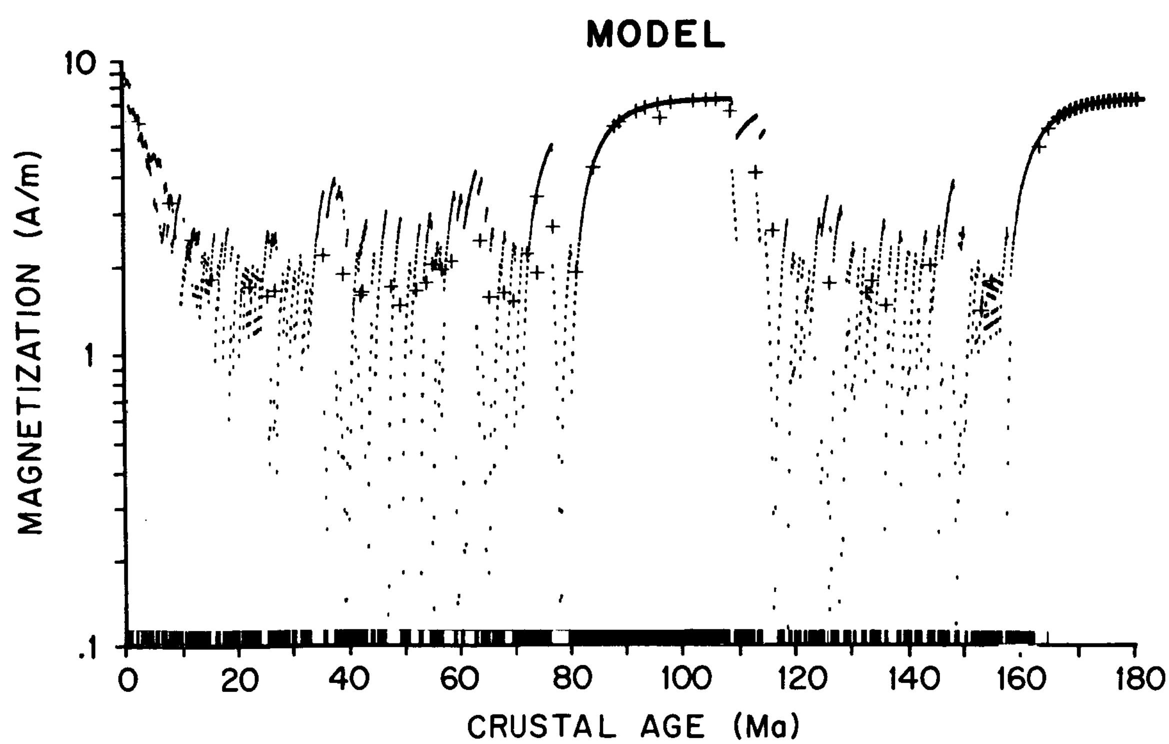

MODEL
MAGNETIZATION (A/m)
CRUSTAL AGE (Ma)
10
1
.1
0
20
40
60
80
100
120
140
160
180

Figure 7

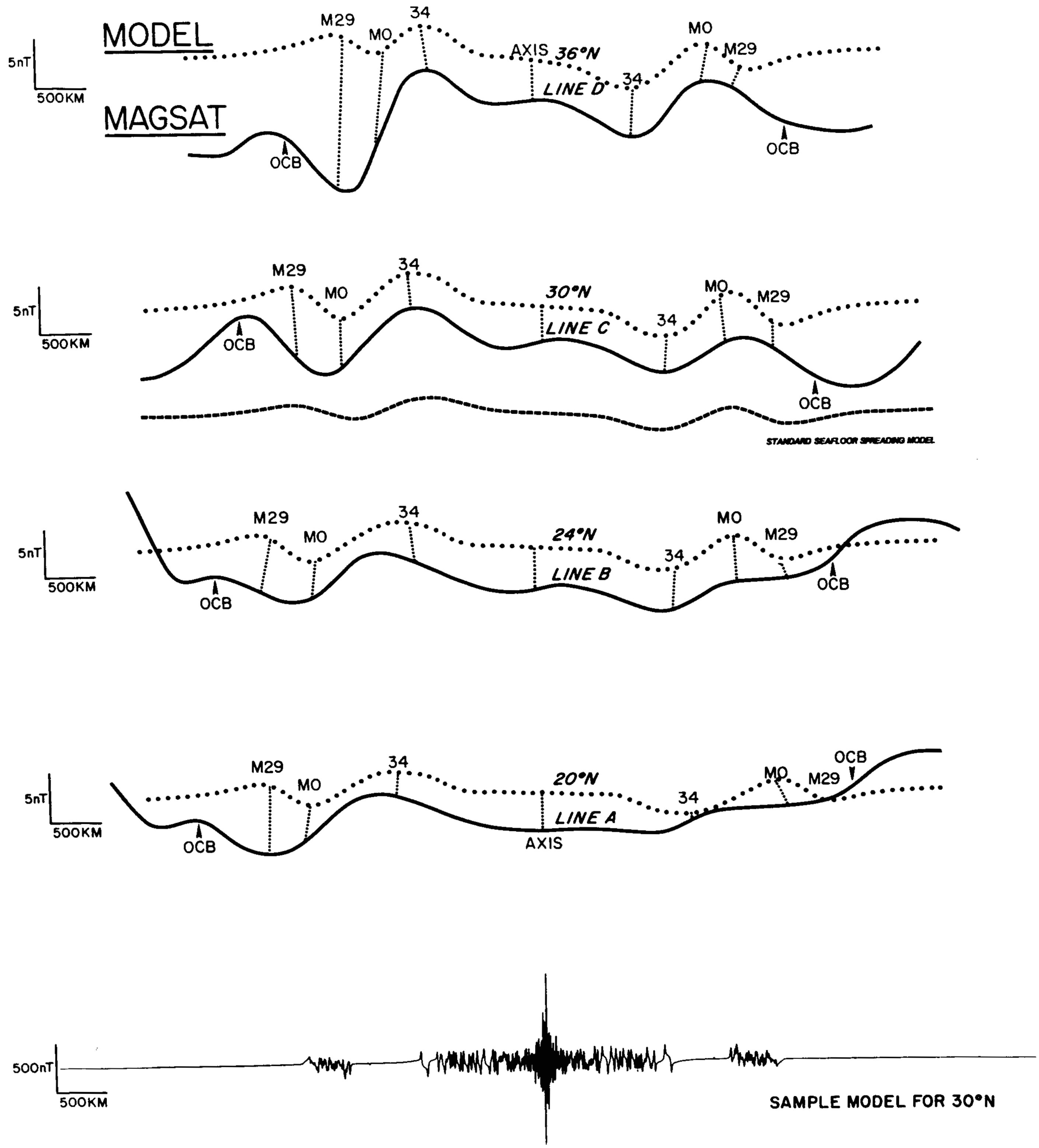
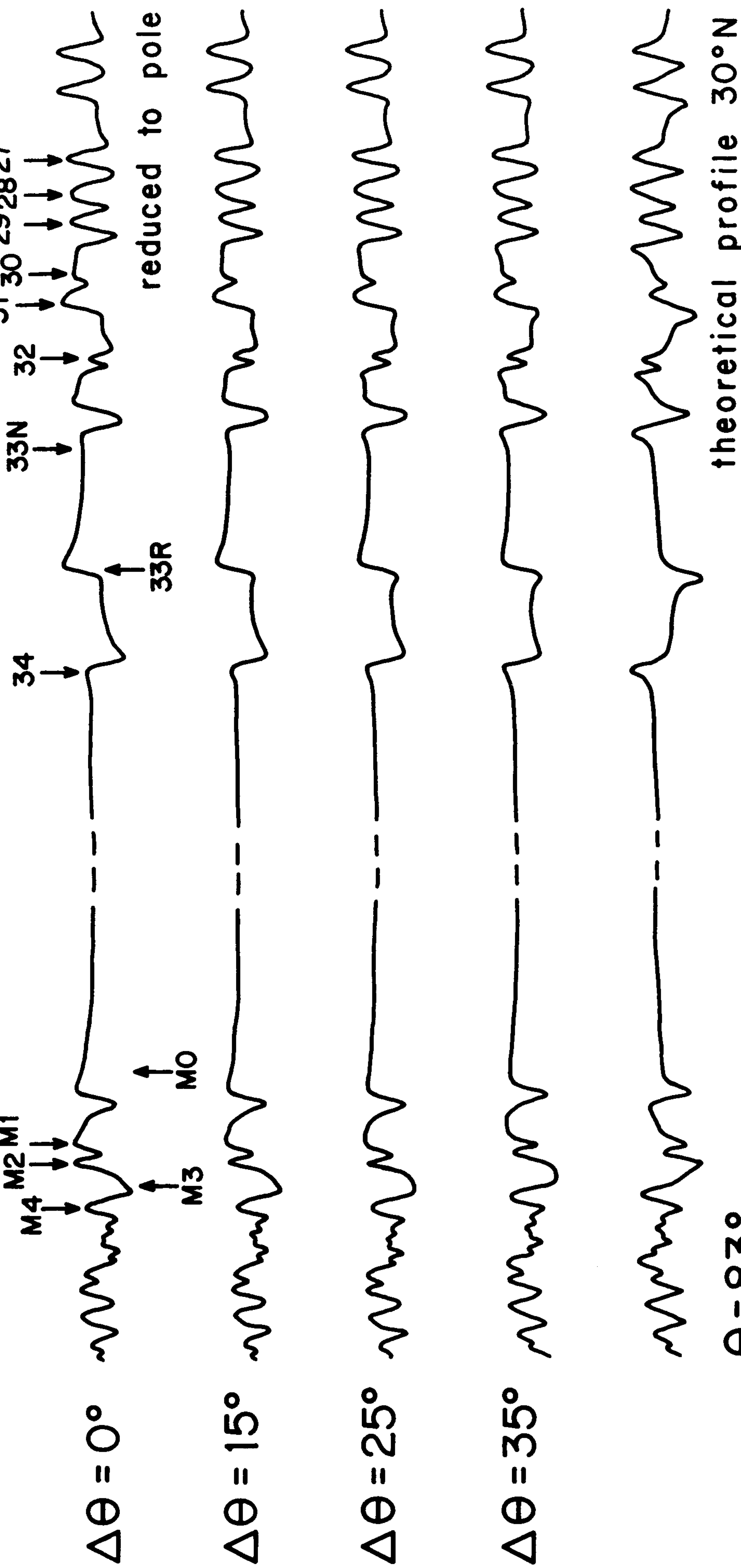

MODEL
MAGSAT
5nT
500KM
M29
MO
34
AXIS 36°N
LINE D
OCB
30°N
LINE C
STANDARD SEAFLOOR SPREADING MODEL
24°N
LINE B
20°N
LINE A
AXIS
500nT
SAMPLE MODEL FOR 30°N

Figure 8



ΔΘ = 0°
ΔΘ = 15°
ΔΘ = 25°
ΔΘ = 35°
Θ = 83°
reduced to pole
theoretical profile 30°N
27
28
29
30
31
32
33N
33R
34
MO
M1
M2
M3
M4

Figure 9