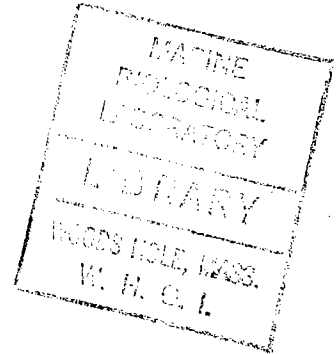


THE ORIGIN AND TECTONIC HISTORY  
OF THE  
SOUTHWEST PHILIPPINE SEA

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1976

by

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TABLE OF CONTENTS

Subject	Page No.
1. List of Figures .....	3
2. List of Tables .....	6
3. Biographical Sketch .....	8
4. Acknowledgements .....	10
5. Abstract .....	11
6. Section I - Introduction .....	12
7. Section II - Magnetic Anomalies in the West Philippine Basin .....	28
a. Abstract .....	29
b. Introduction .....	30
c. Analysis of the Magnetics .....	34
d. Bathymetric Data .....	57
e. Conclusions .....	67
f. References .....	70
g. Acknowledgements .....	75
8. Section III - Paleomagnetism of DSDP Sediments and Basalt, Phase Shifting of Magnetic Anomalies, and Rotations of the West Philippine Basin .....	76
a. Abstract .....	77
b. Introduction .....	79
c. Paleomagnetism of Site 292 Basalts .....	84
d. Paleomagnetism of DSDP Sediments .....	109
e. Paleopoles and Relative Rotations .....	144
f. Conclusions .....	156
g. References .....	159
h. Acknowledgements .....	165
9. Appendix A - Heat Flow, Depth and Crustal Thickness of the Marginal Basins of the South Philippine Sea ..	166
10. Appendix B - Plots of Magnetic Anomalies along Track in the Philippine Sea .....	177

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## LIST OF FIGURES

	Page No.
SECTION II	
1. Ship tracks in the Philippine Sea .....	33
2. Magnetic anomalies in the Philippine Sea.....	37
3. Profiles of magnetic anomalies in the West Philippine Basin .....	40
4. Phase shifted magnetic anomalies and their iden- tifications in the West Philippine Basin .....	47
5. Age vs. distance from the Central Basin Fault .....	55
6. Profiles of bathymetry across the Central Basin Fault.....	57
7. Locations of anomalies in the West Philippine Basin .....	61
8. Depth vs. age for the West Philippine Basin compared to the North Pacific .....	66
SECTION III	
1. Bathymetric map and DSDP site locations in the West Philippine Basin .....	82
2. Demagnetization curves for selected site 292 basalts .....	90
3. Stereonet projections of magnetization vectors for selected site 292 basalts .....	92
4. PSI vs. peak alternating field for selected site 292 basalts .....	94
5. Stability parameter (d) vs. $J_{nrm}$ for site 292 basalts .....	97
6. VRM acquisition vs. time for selected site 292 basalts .....	102

7.	(a) S vs. d and (b) S vs. $J_{nrm}$ for selected site 292 basalts .....	106
8.	Lithological and paleontological summary for site 292 sediment showing positions and ages of detailed sampling locations .....	114
9.	Demagnetization curves and stereonet plots of magnetization vectors for selected site 292 sediments .....	116
10.	PSI vs. peak alternating field for selected site 292 sediments .....	118
11.	Inclination vs. depth down core for site 292 sediments .....	127
12.	Histograms of stability index $\Delta$ for site 292, 290 and 294 sediments .....	130
13.	Age vs. paleolatitude for site 292 sediments and basalt .....	132
14.	PSI vs. peak alternating field for site 290 and 294 sediments .....	141
15.	Age vs. paleolatitude for site 290 and 294 sediments...	143
16.	Ship tracks and magnetic anomaly identifications in the southwest Philippine Basin .....	146
17.	Phase shifted and projected magnetic anomalies with identifications in the West Philippine Basin .....	149
18.	Stereonet projections of paleomagnetic pole positions for the West Philippine Basin .....	153
APPENDIX A		
1.	Topographic features of the Philippine Sea with locations of Antipode and Tasaday ship tracks .....	168
2.	Heat flow in the Philippine Sea .....	170

3. Profiles of gravity, magnetics, bathymetry  
and sediments along Antipode and Tasaday ship  
tracks .....171

4. Histograms of heat flow values in the  
Philippine Sea .....172

5. Heat flow vs. age for the southwest Philippine  
and Parece Vela basins compared with the North  
Pacific .....174

6. Depth vs. age for regions in the Philippine Sea  
compared with the northeastern Pacific .....174

APPENDIX B

1-8 Magnetic anomalies along ship track in the  
Philippine Sea .....183-90

LIST OF TABLES

	Page No.
SECTION II	
1. Paleontological and magnetic anomaly ages for DSDP sites 290, 291, 19 and 32 .....	52
2. Distance and ages of magnetic anomalies across the Central Basin Fault .....	53
3. Smoothed bathymetric depths across the Central Basin Fault .....	64
SECTION II	
1. Summary of remanent magnetic properties for DSDP site 292 basalts .....	87,88
2. Summary of viscous magnetic properties for DSDP site 292 basalts .....	103
3. Remanent magnetic properties for DSDP site 292 sediments .....	120-25
4. Remanent magnetic properties for DSDP sites 290 and 294 sediments .....	136-39
5. Magnitude and location of major east-west sea-floor spreading magnetic anomalies .....	151
APPENDIX A	
1. Heat flow stations in the western Pacific .....	169
2. Heat flow statistics .....	170
3. (a) Depth and age from JOIDES Deep-Sea Drilling sites .....	173
(b) Depth and age from magnetic anomalies and topographic profiles in the Philippine Basin .....	173
4. (a) Comparison of velocity and thickness of the crustal layers in the West Philippine and Parece Vela basins and the North Pacific .....	175

- 4. (b) Expected increase in depth resulting from  
thinner crustal layers ..... 175

## BIOGRAPHICAL SKETCH

I was born in Washington, D.C. on July 11, 1948. My youth and adolescence was spent in Falls Church, Va. except for 1964-65 when I lived in Baghdad, Iraq and attended Al-Hikma University. I attended Oberlin College, Oberlin, Ohio during 1966-70 where I received a B.A. with honors in physics. From 1970 to 1972 I taught physics and 9th grade science at Swarthmore High School, Swarthmore, Pa. and attended Temple University, Philadelphia, Pa. where I finished an M.Ed. in 1972. In 1972 I entered the Dept. of Earth and Planetary Sciences at MIT, Cambridge, Ma. and transferred into their Joint Program in Oceanography with the Woods Hole Oceanographic Institution, Woods Hole, Ma. in 1973. In Oct. 1976 I shall become a Senior Assistant in Research in the Dept. of Geodesy and Geophysics, Cambridge University, Cambridge, England.

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- (1) Sclater, J.G., D. Karig, L.A. Lawver, and K. Loudon, Heat flow, depth, and crustal thickness of the marginal basins of the South Philippine Sea, J. Geophys. Res., 81, 309-318, 1976.



(2) Louden, K.E., Magnetic anomalies in the West Philippine Basin, in The Geophysics of the Pacific Ocean Basin and Its Margin, Geophys. Monogr. Ser., vol. 19, AGU, Washington, D.C., in press, 1976.

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(4) Louden, K.E., Paleomagnetism of DSDP sediments and basalt, phase shifting of magnetic anomalies, and rotations of the West Philippine Basin, submitted to J. Geophys. Res., 1976.

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ABSTRACT

This thesis is a collection and analysis of sea-floor magnetic anomalies, bathymetry, and the paleomagnetism of DSDP sediments and basalt in the West Philippine Basin, in an attempt to resolve questions about its origin as a marginal basin. Our results suggest that this basin was formed in an Eocene pulse of rapid spreading ( $v_{1/2} = 41-44$  mm/yr) in a direction (N 21°E) significantly different from later pulses which opened the more eastern basins of the Philippine Sea. The Central Basin Fault appears to be intimately associated with this spreading by nature of its structure and trend, and it may be a remnant of a former ridge system. Our preliminary calculation of paleopole positions also suggests that there was a large amount (60°) of clockwise rotation between this basin and the magnetic pole. This is consistent with rotations of the Pacific plate with respect to the magnetic pole and current directions of Philippine-Pacific relative rotations. Basement depths of 6 km in the West Philippine Basin imply that its crustal and/or lithospheric structure is different from Pacific structure of the same age.

Thesis Supervisor: John G. Sclater  
Title: Associate Professor of Geophysics

SECTION I

INTRODUCTION

Much of the West Pacific margin consists of oceanic basins separating its deep trenches and their associated volcanic island arc chains from the adjacent continents of Australia and Eurasia. The origins and history of these marginal basins have long remained a controversial enigma to both past and present global tectonic theories. Pre-plate tectonic hypotheses considered these regions to be former continental areas that had subsided (e.g. Marshall, 1910; Kuenen, 1935; Glaessner, 1950). Mechanisms for this subsidence included subcrustal erosion (Lawson, 1932), crustal warping due to compression (Umbgrove, 1947) and oceanization (Belousov and Ruditch, 1961), among many others. It wasn't until seismic refraction techniques became available in the deep oceans during the late 1950's and early 1960's that it was discovered that the crust beneath these basins was similar to that of normal oceans (Gaskell et al., 1959; Shor, 1964; Murauchi et al., 1968). These results have disproven many of the earlier theories, and while some (e.g. Belousov, 1967) still remain possible, they are currently not widely accepted in the West.

Theories contrasting to those with continental origins include the continental drift theory of Wegener (1929) which treated these basins as extensional gaps left

behind the trailing edge of a moving continent, and the expanding earth theory of Carey (1958) which considered them as extensional 'rhombochasms' between large scale lithospheric units. There were also various compressional theories which explained the arcs within and adjacent to the marginal basins as manifestations of convection currents (Vening Meinesz, 1951; Holmes, 1965; Wright, 1966; Menard, 1964). These were predated by the earlier tectogene folding concepts of Hess (1948).

The advent of global plate tectonic concepts (Wilson, 1965; McKenzie and Parker, 1967; Morgan, 1968; LePichon, 1968), in which trenches were considered to be sites of lithospheric subduction, added more justification to these compressional theories. Problems remained, however, since it was found that high heat flow and shallow water depths, which normally are associated with active intrusion and crustal extension (Sclater and Francheteau, 1970), were reported well to the west of the island arc chains (Yasui and Watanabe, 1965; Vacquier et al., 1966; Yasui et al., 1968a, 1968b, 1970; Sclater and Menard, 1967). This high heat flow has presented major problems to the plate model (McKenzie and Sclater, 1968; Oxburgh and Turcotte, 1970).

It was not until Karig (1970) reported new evidence suggesting that crustal extension did in fact exist behind the Tonga-Kermadec island arc, that the earlier extensional theories (Deitz, 1954; Carey, 1958; Murauchi and Den, 1966; Tanner, 1968) were revived. This pioneering study was later supported by additional strong evidence of extension behind the Mariana island arc (Karig, 1971a), and led to several attempts at placing this 'inter-arc spreading' hypothesis within a plate tectonic framework (Karig, 1971b; Pacham and Falvey, 1971; Sclater, 1972; Moberly, 1972). The primary result of this work was that there appeared to be significant differences between values of heat flow and depth vs. age in the older and now no longer active marginal basins and typical results for the normal oceans. This seemed to suggest that either the crust (Karig, 1971b) or the entire lithosphere (Sclater, 1972) was thinner than normal.

Such attempts at synthesizing and comparing geophysical data in marginal basins to pre-existing global plate tectonic models have been severely hampered by several problems. Evidence for extension within active marginal basin troughs is limited and depends primarily on identification of certain topographic morphologies and sediment distribution patterns. Geochemical analysis on a few basalts

dredged from the Mariana and Lau basins also shows strong similarities to rocks from major ocean ridge systems. The amount of detailed geophysical data is limited to a few well surveyed regions in these same active troughs, but the recognition of adequate magnetic anomaly patterns that can be unambiguously identified to world-wide magnetic reversals is confined to other areas such as the Scotia (Barker, 1972) and Tasman (Hayes and Ringis, 1973) seas, which do not conform very well to the type inter-arc spreading model of Karig (1971b). This has meant that unique evolutionary histories of spreading within even the most clearly defined and well studied basins is not yet possible. Controversies exist between possibilities of these basins being either old, trapped oceanic crust or younger crust formed by inter-arc spreading. It is not yet clear whether inter-arc spreading is any different from ridge spreading in normal oceans, whether the older marginal basins have been formed in the same manner as now occurs in the active troughs landward of their island arcs, and whether the many separate marginal basins have a common origin.

Until there exist more unambiguous data, particularly basement ages from both magnetic anomaly identifications and DSDP cores, current geophysical and geological disputes



over the origins of specific marginal basins cannot be resolved. It is the purpose of this thesis to add new constraints to the history of one such marginal basin, the West Philippine Basin, in hopes that this data can help to resolve present controversies over its history (see the introduction to section II) and contribute toward the over-all goal of understanding the basic processes at work in all such marginal basins. Toward this end we have done the following:

- (1) Collected and plotted most of the existing magnetic data in the Philippine Sea (Appendix B),
- (2) Mapped and identified a set of magnetic lineations across the Central Basin Fault in the West Philippine Basin (Section II), and
- (3) Performed a detailed paleomagnetic study on basalt and sediment samples from DSDP sites 292, 290 and 294 (Section III).

Analysis of these additional data leads to the following results:

- (1) The West Philippine Basin was formed south of the equator by rapid spreading ( $v_{1/2} = 41-44$  mm/yr) during the mid-to-late Eocene (35-50 mybp).
- (2) The Central Basin Fault seems to be intimately associated

with this spreading by nature of the structure and trend of its topography. In addition, its deep central rift is also characteristic of the fossil ridges near the Galapagos spreading center (Anderson et al., 1976). Thus the typical association in marginal basins of magnetic anomalies with bathymetric features similar to mid-ocean ridges is continued by the results of this study.

(3) The direction of this former spreading must have been distinctly different from that which exists today along the Mariana and Izu-Bonin arcs. Thus a more complicated tectonic development than originally suggested by Karig (1971a) is inevitable.

(4) Phase shifting of the magnetic anomalies and analysis of the paleomagnetic data for DSDP samples indicates that a significant rotation of the West Philippine Basin has occurred since it was formed. This result agrees with the hypothesis of Uyeda and Ben-Avraham (1972) that this basin might be a remnant of the former Kula-Pacific ridge system of Larson and Chase (1972).

(5) The well-defined age determination of the West Philippine Basin leaves little doubt that there is a significant discrepancy between its depth of over 6 km and normal depths of 5 km for oceanic crust of a similar age. The reasons for

such an offset are as yet unknown, although as earlier suggested by Karig (1971b) and Sclater (1972), the crust and lithosphere might be significantly different than normal (Appendix A; Sclater et al., 1976; Seekins and Teng, 1976).

(6) There is no need in our analysis to require either non-rigid plate interactions between the Philippine and Pacific plates nor significant rotation due to inter-arc spreading. Paleomagnetic pole positions for the West Philippine Basin are consistent with present day plate motions as determined by earthquakes on the margins of the Philippine Sea (Katsumata and Sykes, 1969; Fitch, 1972; Karig, 1975). Furthermore, the paleopole positions for Guam (Larson et al., 1975) which lies on this boundary is also consistent with such a polar wandering curve. This implies that rotations along the island arc chain need not be of separate origin to the rotation between the major plates.

What becomes clear from this study is the necessity for further measurements in the Philippine Sea of the basic physical properties of its crust, upon which all other analyses rely. These include the seismic structure of the deep water basins and the magnetic properties beneath sea-floor spreading anomalies. When these measurements are

made we can begin to form additional concrete constraints on the geophysical processes which have formed this one example of a marginal basin. The uniqueness or generality of this one example among many will remain uncertain for the immediate future.

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SECTION II

MAGNETIC ANOMALIES IN THE WEST

PHILIPPINE BASIN

## ABSTRACT

We present a comprehensive collection of magnetic anomalies in the West Philippine Basin together with several bathymetric profiles and nannofossil age dates for JOIDES sites 290 and 291. From this data we locate symmetric magnetic anomalies that have strikes roughly parallel to the Central Basin Fault. They can be identified as numbers 17-21 if the basin originated south of the equator and opened at a rate of 41-44 mm/yr. This adds justification but not total proof for the theory that the Central Basin Fault may be an extinct spreading center which slowed and then ceased spreading 40 m.y. ago. It is still possible that island arc spreading created these anomalies, but if so, the origin and evolution of such an arc cannot be simply connected to the later development of the Parece Vela and Shikoku Basins to the east. Other results show that the elevation of this region is 1 km lower than would be observed for similar aged crust in the North Pacific. If the JOIDES dates and our anomaly interpretations are correct, then the lack of a correspondingly large negative gravity anomaly suggests major differences between this extinct ridge and those known to be currently spreading.

## INTRODUCTION

One of the most significant difficulties in the original formulation of plate tectonics was its failure to account for tensional features on the concave side of island arcs. Karig's later analysis of these features (Karig, 1970, 1971a) which he synthesized into his theory of inter-arc spreading (Karig, 1971b, 1974) has therefore been a major contribution to our understanding of these marginal basins. Unfortunately, there is as yet little geophysical proof that older, and now no longer active, sections of marginal basins were formed by the same processes that we observe in active troughs. This is important since only a small fraction of these basins are presently active.

The study of magnetic anomalies which could presumably yield new insights into the origins of inactive marginal basins have so far had mixed success. Results fall into two general categories:

- (1) Characteristic anomalies are not apparent, but there are low amplitude lineations parallel to the trench.
- (2) Characteristic anomalies that can be correlated to world-wide events are observed together with a well-defined bathymetric ridge.

The Japan Sea typifies the first case (Isezaki and Uyeda, 1973) and is similar to what is observed in the actively spreading basins of the Tonga (Lawver, personal communication) and Mariana (Karig, 1971a) island arcs. The apparent lack of characteristic magnetic signatures is explained by the diffuse

nature of inter-arc as opposed to mid-ocean ridge spreading, and is one of the differences that distinguish these two spreading modes (Karig, 1970; Matsuda and Uyeda, 1971). The second type is observed in the Tasman (Hayes and Ringis, 1973) and Scotia (Barker, 1972) seas and is typical of mid-ocean ridges.

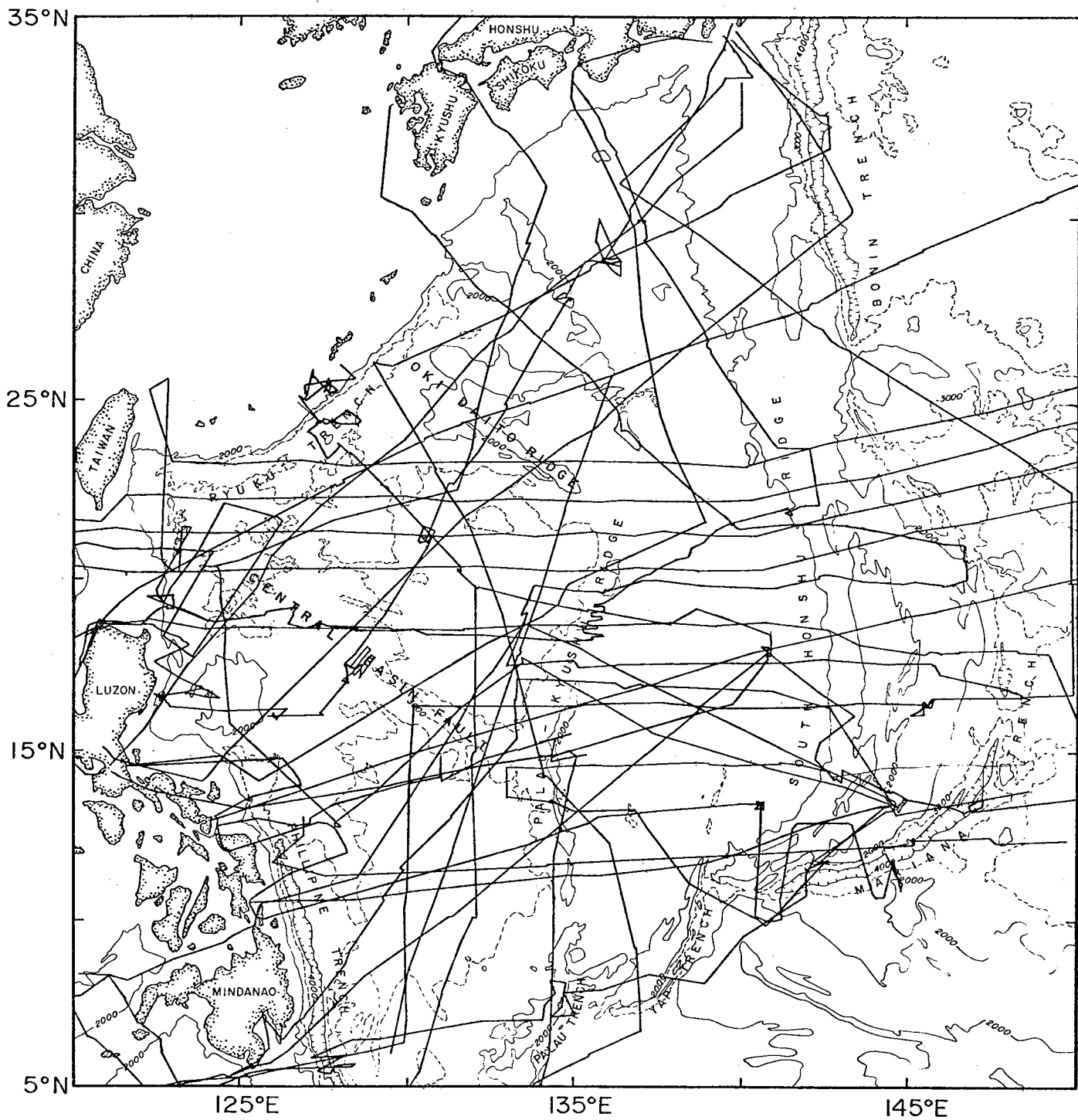
The possibility of ridge spreading in marginal basins raises several questions. If ridge spreading exists, is it related to inter-arc spreading or are there two separate processes that together determine the history of these basins? Can such ridge spreading be related to mid-ocean ridge systems of the past or does it have a separate origin?

A good place to seek the answer to these questions lies in the Western Philippine Sea. This region (Figure 1) is separated from the eastern sections of the Philippine Sea by the Oki-Daito and Palau-Kyushu Ridges. Its main bathymetric feature is the Central Basin Fault (Hess, 1948) which runs NW-SE and splits the area into nearly equal halves. This trend is quite different in direction to that observed in the ridges to the east, which run north-south and parallel to the Mariana and Bonin trenches. What makes the West Philippine Basin of interest is that it is the only region where ideas favoring alternative modes of spreading have been in direct dispute. Karig proposed that this region was formed in an Eocene pulse of inter-arc spreading - an older manifestation of what is presently occurring in the Mariana Trough (Karig,

FIGURE 1

Ship tracks superimposed on a 2000 m, 3000 m, and 4000 m contour map of the Philippine Sea (Chase and Menard, 1969). Major bathymetric features are identified.





1971a). He treated the Central Basin Fault as a strike-slip fault with several hundred kilometers of left-lateral offset (Karig et al., 1973). Counter to this was evidence presented by Ben-Avraham et al. (1972) of NW-SE trending magnetic anomalies symmetric about the Central Basin Fault. Uyeda and Ben-Avraham (1972) therefore proposed that the Central Basin Fault was an extinct spreading center which is related to the Kula-Pacific ridge system of Larson and Chase (1972).

This paper is a study of this region as a particularly interesting example of an inactive marginal basin. By such a study we hope to clarify the processes that have shaped its history. Specifically, we synthesize magnetic, bathymetric and Deep Sea Drilling data in an attempt to:

- (1) locate any magnetic lineations and confirm or reject those published by Ben-Avraham et al. (1972),
- (2) correlate specific anomalies with world-wide reversals,
- (3) determine the nature of the Central Basin Fault, and
- (4) relate our findings to possible theories of this basin's origin.

#### ANALYSIS OF THE MAGNETICS

##### Lineations

Our first step was to collect all the known magnetic data in the Philippine Sea. Those lines which are utilized for this work are shown in Figure 1. While processing this data we were able to run our own line during leg 7 of the

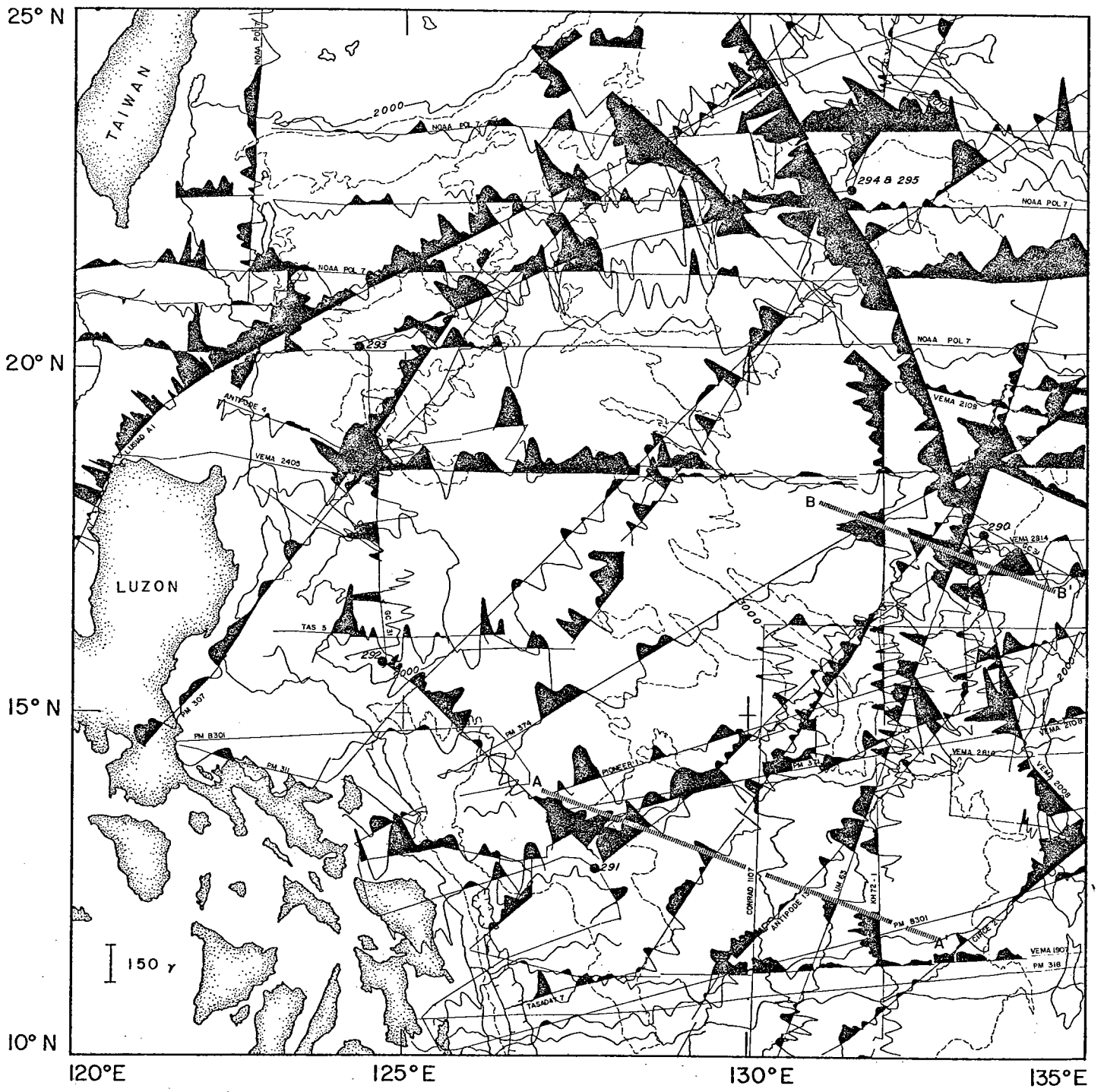
Tasaday Expedition of Scripps Institute of Oceanography (Nov.-Dec., 1973). We ran this line at right angles to the Central Basin Fault and close to the lines which Ben-Avraham et al., (1972) had earlier published. We hoped that if lineations existed, they would at least be traceable in this limited area of denser coverage.

Figure 2 shows the magnetic anomalies in the Western Basin plotted at right angles to each ship's track. The scale is smaller than normal because of the low amplitudes of the anomalies. There are two difficulties that such low amplitude anomalies pose. One is the considerable effect that diurnal and semi-diurnal variations in the Earth's present field has upon the amplitude. This might be on the order of 10-30 gammas for each component, so that the total effect could be as large as 50 gammas (Chapman and Bartels, 1940). The other factor that affects the zero level of the anomalies but which leaves the shape unchanged is the secular variation, which in this region is as great as tens of gammas per year (Vestine, 1967). Both of these effects can be seen in the large offsets in certain profiles in Figure 2, but can be neglected in this study since our anomalies have a wavelength of 50 n.m. or less and we are only matching shapes and not contouring equal values.

We have concentrated our study on the anomalies located between lines A-A' and B-B' in Figure 2. This is where our

FIGURE 2

Magnetic anomalies in the West Basin of the Philippine Sea plotted at right angles to track together with an identification of each ship track, 2000 m and 3000 m contours, and the location of JOIDES sites 290 through 295 (JOIDES, 1973).



coverage is greatest and where we can use JOIDES data taken in the West Philippine Basin during leg 31 of the Glomar Challenger (JOIDES, 1973) most effectively. The long wavelength anomalies close to lines A-A' and B-B' are particularly noticeable and distinctive. It is possible that there are more lineations to the north of this area; however, if so, they are certainly more confused than those in the south and we feel that the orientation and density of our tracks is not sufficient to unravel the confusion into an indisputable pattern. In Figure 3 we project the anomalies between lines A-A' and B-B' along a direction of  $N21^{\circ}E$ . This direction is determined primarily by the strike of the large wavelength anomaly near line A-A'. Notice also that in Figure 2 the anomalies for tracks in a direction perpendicular to the projection show low amplitudes and long wavelengths, as expected when running parallel to a mid-ocean ridge. But caution must still be taken at accepting this result since offsets between tracks can produce a spurious trend. At the present we have no evidence from either magnetics or topography that such offsets occur in this restricted area, and have therefore positioned our projected tracks only by the crossings between them in Figure 2.

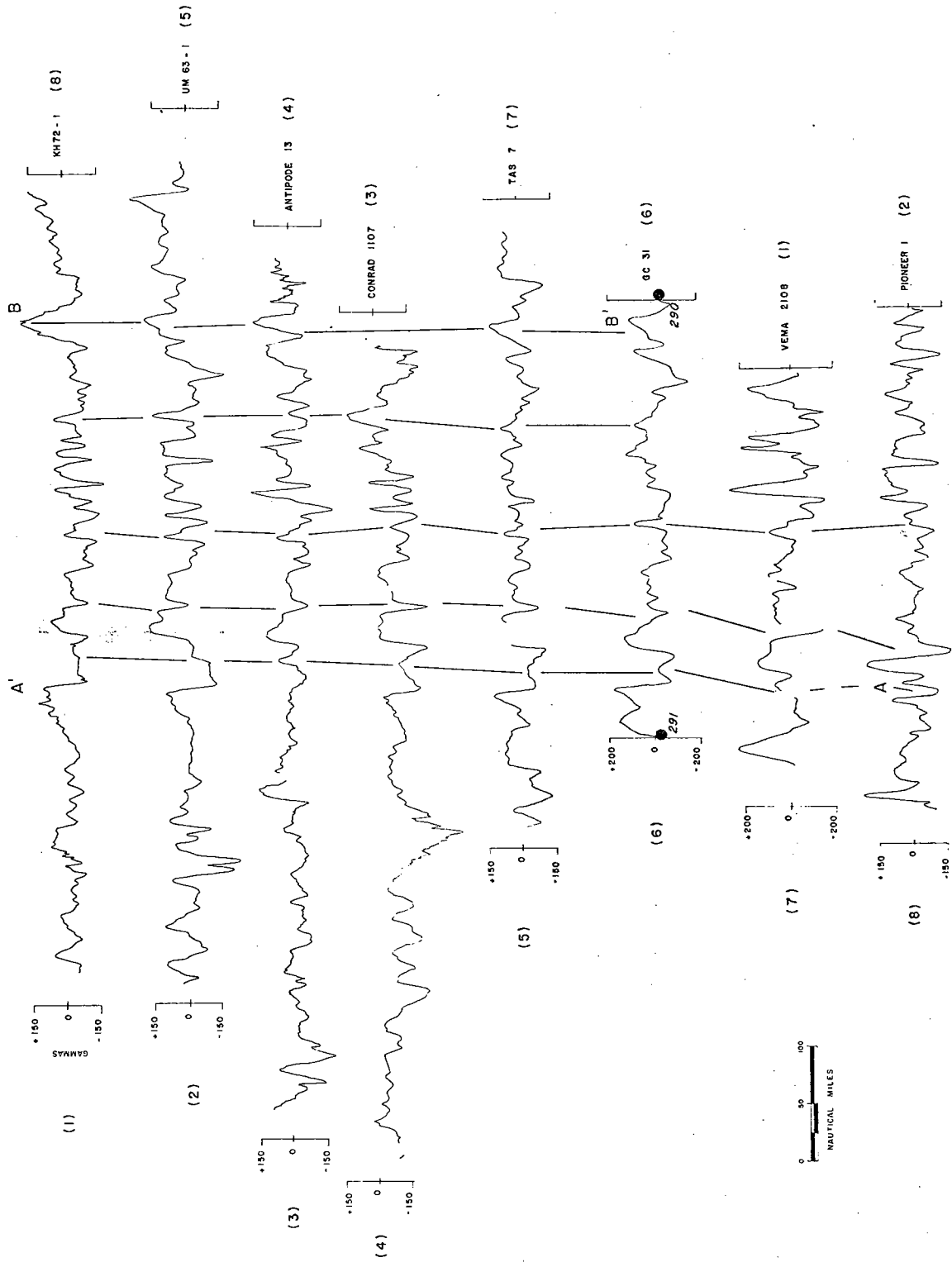
Even with consideration for these precautionary words, we feel that these lineations are real, and that our measured trend of  $N69^{\circ}W$  (at right angles to the direction of projection)

FIGURE 3

Stacked magnetic anomalies which occur between lines A-A' and B-B' of figure 2 projected along a direction N 21°E. The relative position of each profile is indicated by the location of points A, A', B, B' as well as by relative numbers to the right and left of each profile. These numbers increase in an eastward direction. JOIDES sites 290 and 291 are located on the GC 31 profile.

NORTH

SOUTH





is valid. This confirms that those anomalies of Ben-Avraham et al. (1972) form traceable lineations (but in a direction 20° counterclockwise to their original prediction of N50°W). The extent of these lineations is rather limited, but one may still conclude from this evidence that at least the southern part of the West Philippine Sea opened in a nearly north-south direction. This is contradictory to the early inter-arc spreading hypotheses, which would predict a west to east direction of opening - the same as occurred in the Parece Vela Basin to the east (Karig, 1971a). Another objection is the lack of a uniform progression in the dates for JOIDES sites 290 through 295 that were drilled during leg 31 of the Deep Sea Drilling Project (Karig et al., 1973). We agree with the more recent statement of Karig et al. (1973) that this basin must have had a more complex history than was originally thought. If inter-arc spreading occurred at all, it must have been in a direction significantly different than is observed to the east.

The absence of recorded seismic events within the West Philippine Basin (Barazangi and Dorman, 1969; Fitch, 1972; Shimamura et al., 1975) probably indicates that there is no current generation of crust. It is most likely older than the Parece Vela Basin to the east (Karig, 1971a) as indicated by its thicker sediment cover (Ingle et al., in press). We conclude that the West Basin opened before the start of inter-arc spreading

in the Parece Vela Basin by one of the following two processes:

- (1) symmetric spreading by the Central Basin Fault, or
- (2) spreading from a ridge that no longer exists, such as in a piece of trapped oceanic crust.

In the first case we would see symmetry about the Central Basin Fault in our magnetic anomalies. This is the contention of Ben-Avraham et al. (1972) which we will refer to as the symmetric model. In the second case there would be no symmetry and our anomalies would be matched by a model of spreading from only one side of a ridge. This we will refer to as the non-symmetric model. Notice that each model could have originated either in a mid-ocean ridge system or as spreading behind an island arc.

#### Phase Shifting

Relative poles of rotation for the Pacific-Philippine plate system have been determined by fault plane solutions (Fitch, 1972), but they are poorly constrained and still a matter of debate (Karig, personal communication). Otherwise very little is known about the plate motions of the Philippine Sea. Such knowledge is important if we wish to compare our magnetic anomalies to those representative of other oceans or to a theoretical model based on a standard time scale. Changes in paleolatitude and paleostrike produce changes in the

inclination of the remanent magnetization which can cause large changes in the amplitude and shape of the anomalies, making valid correlations difficult or impossible. This is especially true when close to the equator as we are in this case.

A very useful approach to the problem of matching anomalies when little is known about their original orientation is a method developed by Schouten and others (Cande and Schouten, 1971; Cande et al., 1973; Cande and Schouten, in prep.). In this method the anomaly is formulated in terms of an angle  $\theta$  called the skewness parameter:

$$\theta = I'_r + I' - 180^\circ \quad (\text{Cande and Schouten, in prep.}) \quad (1)$$

where  $I'_r$  = effective inclination of the remanent field  
 $I'$  = effective inclination of the ambient field

The influence of both the magnetic inclination and the strike of the anomaly structure are combined in the angle  $\theta$  since the effective inclinations are defined by,

$$\tan I' = \tan I_0 / \sin \alpha \quad (\text{Gay, 1963}) \quad (2)$$

where  $I_0$  = inclination of the magnetic field

$\alpha$  = strike of the lineations.

The Fourier transform of an anomaly formulated in this manner can be represented as,

$$M(k) = C \cdot J(k) \cdot E(k) \cdot e^{j\theta}. \quad (\text{Schouten, 1971}) \quad (3)$$

$$\text{where } C = (\sin I \sin I_r) / (\sin I' \sin I'_r) \quad (4)$$

is a real constant.  $J(k)$  is the Fourier transform of the function that represents the distribution of magnetization in the anomaly producing layer.

$$E(k) = 2\pi (e^{-ak} - e^{-bk}) \quad (5)$$

where  $a$  and  $b$  are the depths to the top and bottom of the magnetized layer. If two sets of anomalies have the same origin but are now separated from each other,  $E(k)$  and  $J(k)$  would be identical. Changes in amplitude and shape from one set to the other would be determined solely by the term  $C \cdot e^{j\theta}$ . Multiplication of this term by  $e^{-j\theta}$  would make  $M(k)$  real and independent of  $\theta$ . That would leave us with deskewed (symmetric) anomalies which have different amplitudes but identical shapes.

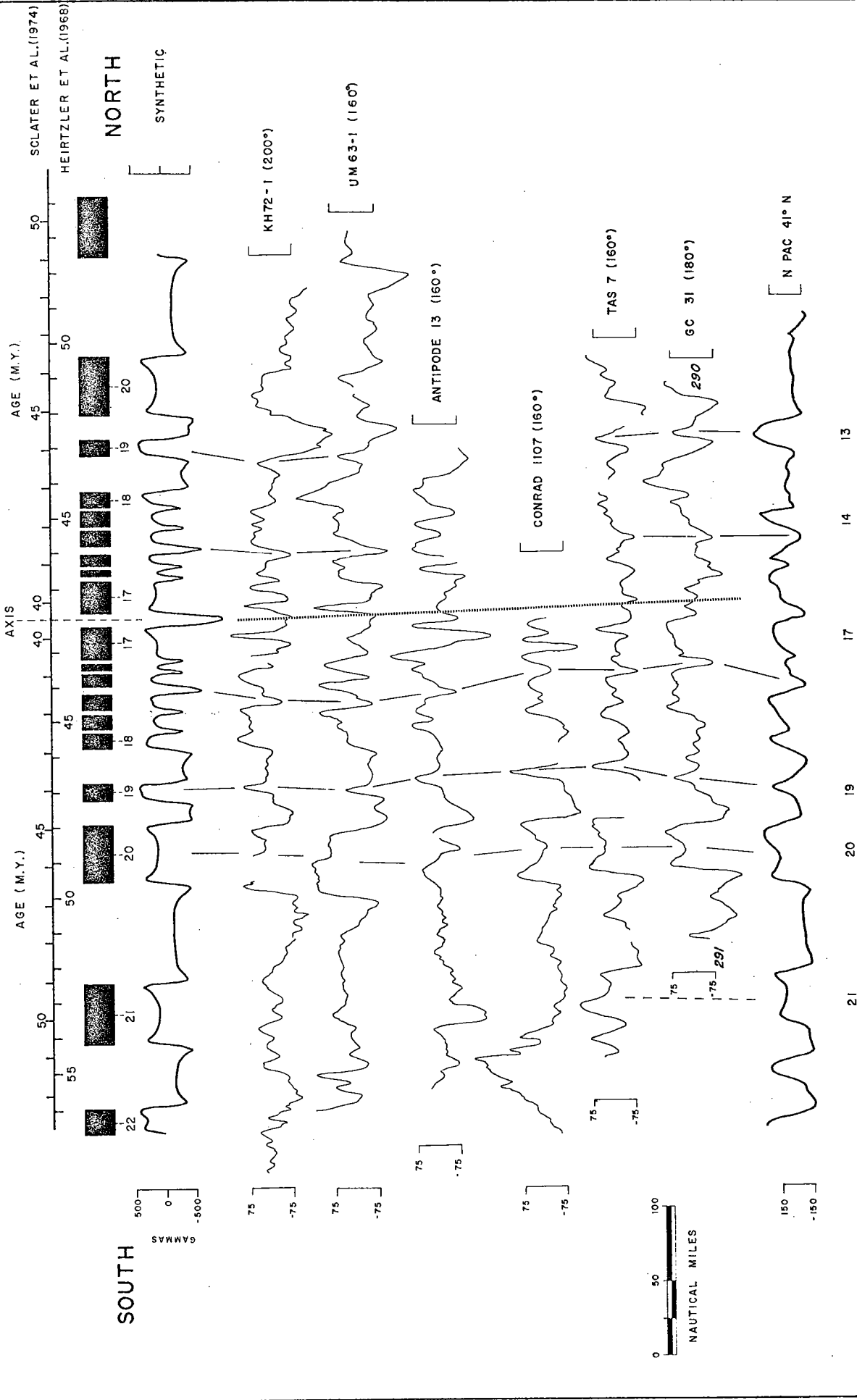
What is done is to input a profile of magnetic anomalies and then change the value of  $\theta$  until no skewness is observed in the output profile. It is therefore the actual shape of the anomalies that determines a value for  $\theta$ . Using eq. (1) and eq. (2) one can then calculate possible sets of values for the paleolatitude and strike of the anomalies. In the most general case one can determine two possible values for  $\theta$ .

These would be  $180^\circ$  apart and represent two deskewed solutions, one of which is the inverse of the other and occurs when inverted positive anomalies become negative anomalies. Recognition of the possibility that anomalies can be inverted was the key to the Larson and Pitman (1972) correlation of the Keithley lineations to three other sets of anomalies in the Pacific. This established a worldwide correlation of Mesozoic anomalies and led to the first attempt at a Mesozoic time scale. In their case the Pacific anomalies were inverted with respect to the Keithley lineations, and was the main reason why they had been invalidly correlated in an earlier attempt by Vogt et al. (1971). We believe that a similar situation led to earlier unsuccessful attempts by Ben-Avraham and the author to correlate the Philippine anomalies of Ben-Avraham et al. (1972).

Results of our application of Schouten's method to the anomalies of Figure 3 are shown in Figure 4. These anomalies represent the deskewed inverse of the original anomalies. The other, non-inverted solution has been eliminated because it would lead to values of  $I'_r$  greater than  $90^\circ$  and therefore is not a physically real solution. This leaves us with only one possible  $\theta$  value, resulting from the inversion of our original anomalies and ranging from  $-160^\circ$  to  $-180^\circ$ . In our case  $I_o = 28^\circ$  and  $\alpha = 110^\circ$  which means that  $I' = 30^\circ$  from eq. (2).

FIGURE 4

Phase shifted magnetic profiles compared with a symmetric spreading model at top and a non-symmetric model (spreading from only one side of a ridge) at the bottom. The degree of phase shifting is indicated in parentheses next to the identification of each profile. The dotted line locates the central valley of the Central Basin Fault. Anomaly numbers are based on the system of Heirtzler et al. (1968)



Thus  $I'_r$  can range from  $-10^\circ$  to  $-30^\circ$  according to eq. (1). If  $\alpha$  has remained constant, then these anomalies were formed between  $-4.5^\circ$  and  $-15^\circ$  latitude; if  $\alpha$  has changed, the value of the paleolatitude could range from  $0^\circ$  to  $-16^\circ$ . These results disagree with earlier solutions given by Ben-Avraham et al. (1972) which predict formation north of the equator. It is unfortunate that we cannot determine if  $\alpha$  has changed since these anomalies were formed. Changes in  $\alpha$  near the equator affect the amplitude of the anomalies (the value of  $C$  in eq. (3)) rather than their skewness (Schouten, 1971), and too little is known regarding the predicted amplitudes of magnetic anomalies for us to use that as a criterion (Cande and Schouten, in prep.).

#### Identification of Anomalies

The nannofossil ages reported from leg 31 of the Glomar Challenger (JOIDES, 1973) were all in the Eocene or younger, except for hole 294/295 which was reported to be Paleocene. This last date has been recently revised and now seems to be much the same as the others (Karig, personal communication). At first these dates were thought to rule out the possibility of symmetric spreading from the Central Basin Fault (JOIDES, 1973; Karig et al., 1973). We do not subscribe to such a conclusion. These dates do rule out a speculation by Ben-Avraham et al. (1972) that his Philippine anomalies might be part of the



Mesozoic. But beyond telling us in what age province to look for matching our anomalies to standard reversals, they tell us little. Spreading rates must have been large, but the direction of spreading is not clearly indicated. We must therefore assume that the anomalies get younger as they approach the Central Basin Fault from the south, so as not to rule out the symmetric model from the outset. We cannot say a priori whether this is solid ground upon which to build further analysis. If we assume this nonetheless and then look at sea-floor spreading anomalies of 50 m.y.b.p., we find that there exists from anomaly numbers 22 to 17 a progression from longer wavelengths to shorter. This fits the general trend of long to short to long wavelengths observed in the anomalies of Figure 3 as one crosses the Central Basin Fault. By using the standard North Pacific profile at 41°N (Pitman et al., 1968), shifting it to remove its slight skewness and changing its horizontal scale to form a best fit, we produce the model at the bottom of Figure 4. This is a non-symmetric model starting at anomaly number 21 and ending at number 13, where only the direction of spreading has been assumed. To contrast it with the symmetric model, we have constructed the top model of Figure 4 from a standard magnetics program (McKenzie and Sclater, 1971). It adheres to the same fit in the region south of the Central Basin Fault as does the non-symmetric model. Notice however that they are not

identical reproductions. The synthetic anomalies include some additional short reversals not found in the Pacific data. This is to be expected since the reversal history for that age is not known with great precision.

Except for these few short reversals, both models are the same south of the Central Basin Fault. If we now look at the north side where the models should differ, it appears that the symmetric model forms a better match to our lineations. The primary basis for this is the presence of some positive events that do not appear in the non-symmetric model. It should be noted however, that both models show the same general trend from long to short to long wavelengths already noted in the real data. We therefore cannot be absolutely certain which model to discard on the basis of anomaly matching alone.

#### Testing the Models

The location of JOIDES sites 290 and 291 occur at either end of the anomalies presented in Figure 4. This allows us to use the dates of these sites for confirmation of our identifications to the south of the Central Basin Fault, and as a test for the age that each model would predict to the north. In order to do this we must: (1) analyze the fossil record for sites 290 and 291 to determine their range of possible

ages, and (2) date our anomalies by the use of a standard time scale.

Table I presents the JOIDES data for sites 290, 291, 19 and 32. The latter two are sites of similar age to the former but taken from other ocean basins. They were used by Sclater et al. (1974) to revise the Heirtzler et al. (1968) time scale, which was found to predict progressively older ages than the fossil record. We will refer to their revision as the modified Heirtzler scale. Notice that this revision still does not account for all of the discrepancies between fossil and magnetic ages. Site 19 has a magnetic anomaly age based on the modified Heirtzler scale of 1-6 m.y. older than its nannofossil age as determined according to the Berggren (1972) Cenozoic time scale. A similar discrepancy appears in the ages for site 291, where we predict a magnetic anomaly age 3-10 m.y. older than the nannofossils allow. Since site 19 has well defined magnetics but still shows a similar variance, we believe that this justifies our magnetic identifications to the south of the Central Basin Fault.

Let us now analyze what our two spreading models predict for site 290. The magnetic vs. fossil ages should exhibit a similar pattern as do those of site 32. This means that they should agree to within 3 m.y. At first glance this would preclude the symmetric model, but actually this evidence is not very convincing. In Figure 5, which is a graph of

SITE	LAT	LONG	CONTACT CORE	DIAGNOSTIC CALCAREOUS ZONE	NANNO- FOSSIL AGE <sup>1</sup>	MAGNETIC ANOMALY	HEIRTZLER et al.(1968) AGE	SCLATER et al.(1974) AGE
19	28°32.1'S	23°40.6'W	12	Chiphragmolithus quadratus	46-48	Just before 21	53-56	49-52
291	12°48.4'N	127°49.9'E	5	Discoaster barbadiensis Cyclococcolithus formosus	38-43	End of 21	50-52	46-48
32	37°07.6'N	127°33.4'W	14	Coccolithus bisectus	34-37.5	13	37-39	35-37
290	17°44.9'N	133°28.1'E	9	Reticulofenestra hillae R. umbilica	36-38	Just before 19 Just after 13	47-49 36-38	44-46 34-36

<sup>1</sup> From Berggren (1972) and Bukry (1974)

TABLE I

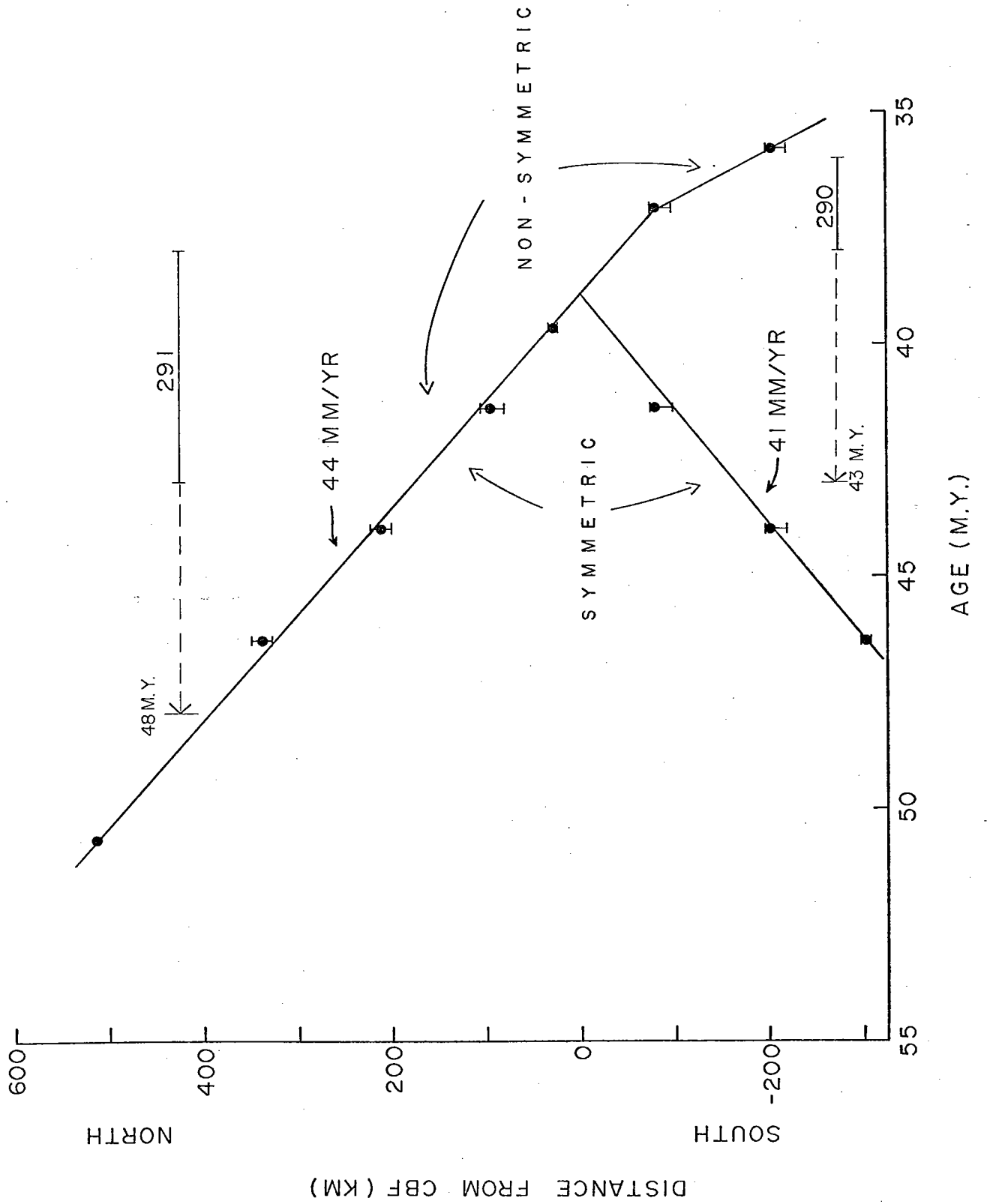
Anomaly	DISTANCE FROM CBF (NM)						Average		AGE (MY)	
	KH 72-1	UM 63-1	ANT 13	C 1107	TAS 7	GC 31	N.M.	K.M.	Heirtzler	Sclater
Beginning of 21					277		277	513	54.2	50.7
Beginning of 20	183	189	177	180	180	183	183±6	339±11	49.6	46.4
19	114	117	114	109	109	120	114±6	211±11	47	44
Between 17 & 18	54	57	54	26*	46	43	51 <sup>+6</sup> <sub>-8</sub>	95 <sup>+11</sup> <sub>-15</sub>	44.1	41.4
End of 17	14	14	31*		17	14	15 <sup>+2</sup> <sub>-1</sub>	28 <sup>+4</sup> <sub>-2</sub>	42.3	39.7
Between 15 & 14 or Between 17 & 18	-43	-46	-34		-46	-43	-43 <sup>+3</sup> <sub>-9</sub>	-80 <sup>+5</sup> <sub>-17</sub>	39.5	37.1
19 or 13	-100	-109			-112	-112	-109 <sup>+3</sup> <sub>-9</sub>	-202 <sup>+5</sup> <sub>-17</sub>	47	44
Beginning of 20	-166	-160					-163±3	-302±5	49.6	46.4
JOIDES Site 290					-149		-149	-276		
JOIDES Site 291					229		229	424		

\* Not included in average

TABLE II

FIGURE 5

Graph of distance from the Central Basin Fault vs. age, taken from values measured in figure 4 and listed in Table II. Symmetric vs. non-symmetric models for ages given by the revised Heirtzler time scale (Sclater et al., 1974) are compared to nannofossil ages for JOIDES sites 290 and 291. Arrows indicate ages adjusted to better represent the basement contact.



measurements taken from Figure 4 and tabulated in Table II, neither the symmetric nor the nonsymmetric models uniquely fit the JOIDES data. The primary reason for this is the difficulty in dating the JOIDES holes due to a lack of fossils close to basement for site 291 and failure to hit basement for site 290. If sedimentation rates are extrapolated to the basement of site 291, one obtains an age 5 m.y. older than the lower most nannofossils indicate (Ingle et al., in press). If one does the same for site 290, an age in excess of 43 m.y.b.p. results.

These paleontological ages, once corrected to better represent true basement (dotted arrows in Figure 5), have now become much more definitive. Site 291 becomes an even better fit to our magnetic identifications. Site 290 now adds further evidence in favor of magnetic symmetry across the Central Basin Fault, since it is unreasonable to suggest that magnetic basement could be younger than the age of the earliest sediments.



#### BATHYMETRIC DATA

A summary diagram (Figure 6) shows our magnetic lineations and their possible interpretation. The lineations of Figure 4 are mapped onto the anomalies of Figure 2 for several specific reversals. We also include the bathymetric trace of the Central Basin Fault indicated in Figures 4 and 7. Notice that in this limited area it runs at a  $20^\circ$  angle to its overall trend, and is nearly parallel to our magnetic lineations. It seems highly unlikely that this would occur unless both magnetics and bathymetry are genetically related. This could only mean that this section of the Central Basin Fault was a former spreading center. Further confirmation is seen in Figure 7, which includes profiles of bathymetry projected along the same direction as the magnetics of Figures 3 and 4. We observe that:

(1) In all but the Tasaday 7 profile the Central Basin Fault is a sharp trough bounded by two peaks, which bottoms at a depth equal to that of the outer flanks. The Tasaday 7 profile has this same sharp trough but also has a broad trough next to it, which somewhat distorts the pattern.

(2) If a smooth average is drawn through the profiles, the area immediately adjacent to the Central Basin Fault is elevated relative to the flanks. This can be observed even though the flanks of several profiles are distorted by the outer rise of the Philippine Trench (Watts and Talwani, 1974)

FIGURE 6

Map of magnetic lineations (heavy lines) and their possible interpretation according to the symmetric spreading model . Less certain identifications are dotted. Heavy dashed line indicates the central valley of the Central Basin Fault which occurs between peaks on either side. Total extent from peak to peak is indicated by shaded region.

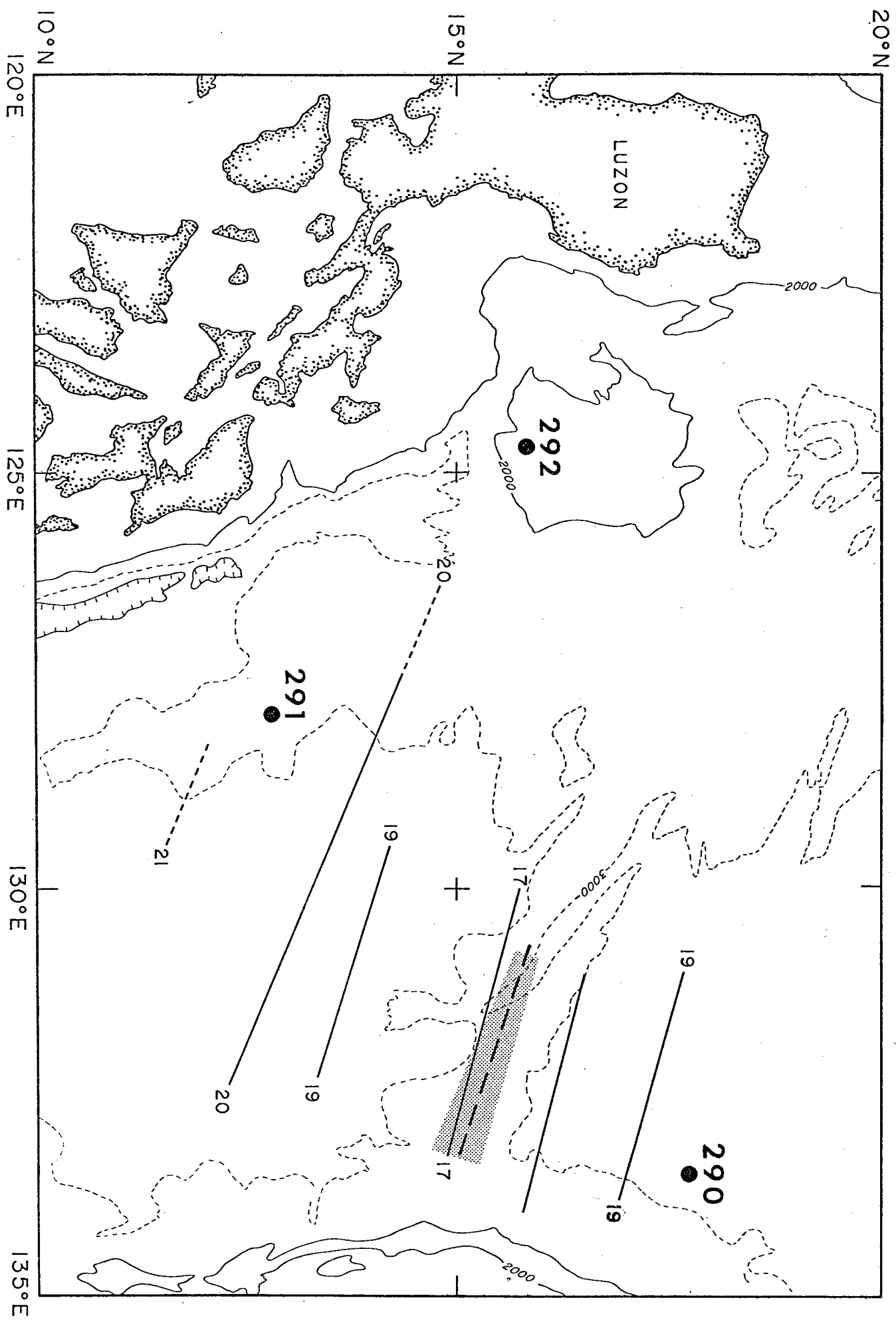


FIGURE 7

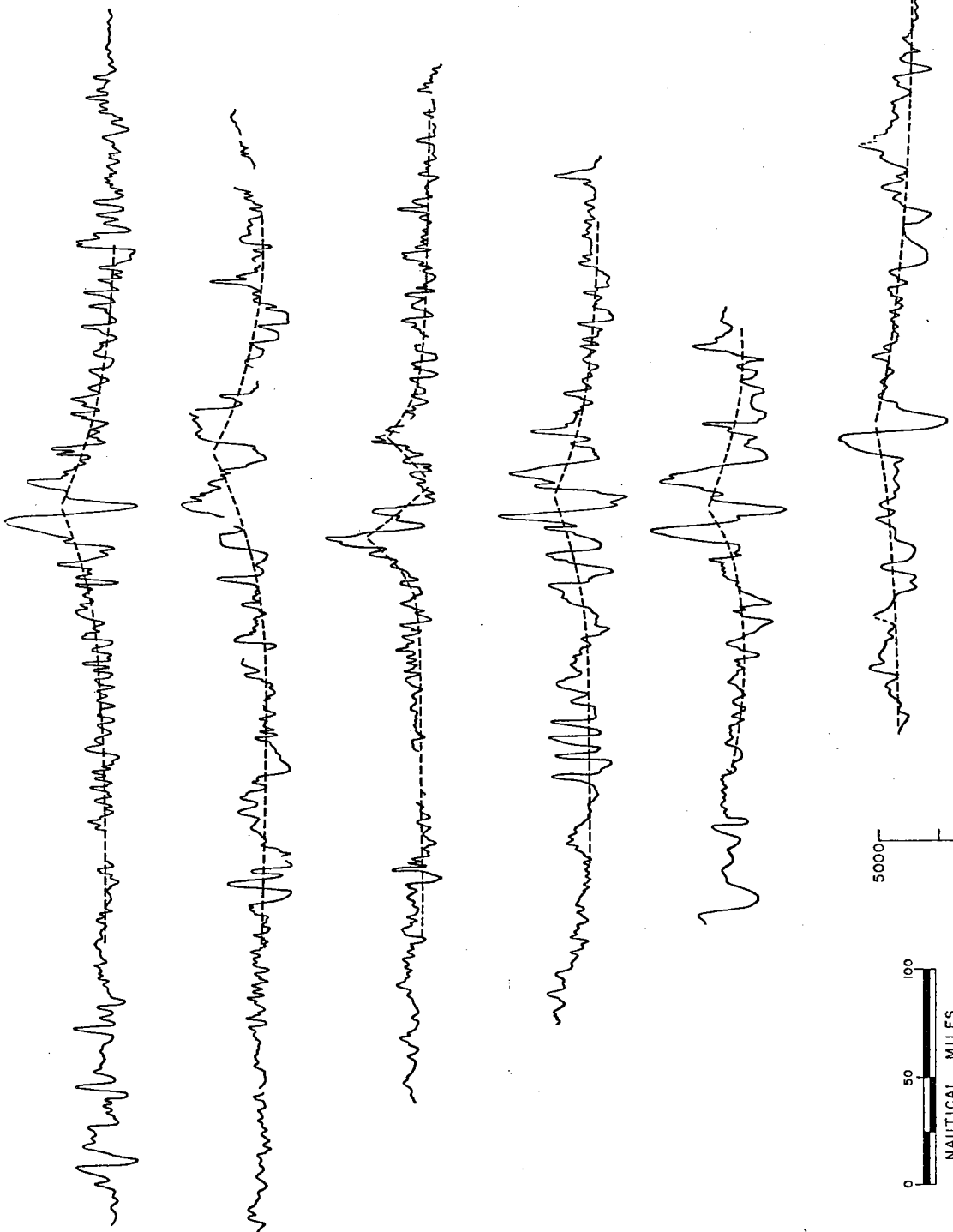
Bathymetric profiles across the Central Basin  
Fault projected along a direction of N 21°E.  
Dashed line represents visual smoothing of  
overall topographic trends.

SOUTH

NORTH

5000  
7000  
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7000



0 50 100  
NAUTICAL MILES

DEPTH (UNCORRECTED METERS)

(southern sections of profiles ANT 13, TAS 7, GC 31 and PIONEER 1) and the Palau-Kyushu Ridge (northern sections of profiles ANT 13 and PIONEER 1). However, the rate of drop in elevation differs for each profile. In the GC 31 profile it is somewhat obscured by large amplitude, short wavelength fluctuations. The TAS 5 line which is located to the north of the magnetic lineations (Figure 2) shows a slight drop that is not well determined. The Central Basin Fault has an anti-symmetric shape in this profile which differs from those to the south.

A detailed survey during leg 5 of cruise Tasaday indicates that at that point the Central Basin Fault is a N-S en eschalon feature which appears highly fractured (Karig, personal communication). This could have occurred during a later tectonic activity postdating its spreading history and thereby changing its original character. If such activity was strongest to the north, it might explain why the bathymetry and magnetic patterns become more confused there. The remaining question is whether our region of study to the south represents a small remainder of the original configuration of the entire Central Basin Fault or whether the West Philippine Basin was always a collection of distinct tectonic provinces. Unfortunately the answer is at present beyond the limits of our data to decide.

The profiles south of the TAS 5 track suggest two

further complications. First, the deep trough at the axis of the Central Basin Fault is morphologically similar to slow spreading ridges ( $v_{1/2} < 20$  mm/yr) which contradicts our magnetic anomaly evidence (Fig. 5) of a faster spreading rate ( $v_{1/2} = 41-44$  mm/yr). The faster spreading rate is also supported by the sediment pattern which shows a thin (<200 m) and uniform layer. Perhaps we can explain this conflict by suggesting that the spreading decreased just before it stopped, thereby producing a limited extent of greatly elevated and rifted topography. This would also help explain why the magnetic pattern changes so drastically from one profile to the next in the vicinity of the Central Basin Fault. Second, if we plot depth vs. age, we see from table III and Figure 8 that all the depths, both for JOIDES ages and for ages based on our magnetic anomalies, are universally greater than for the North Pacific (Sclater et al., 1971). Thus according to these plots, the age of this basin should be greater than 80 m.y.b.p.

We do not believe that the depth of this basin gives a true indication of age for the following reasons:

(1) Nowhere in JOIDES cores 290-295 are there sediments older than the Paleocene (Ingle et al., in press).

(2) The correspondence of the younger sediment ages for JOIDES cores 290 and 291 to a plausible magnetic anomaly identification seems too coincidental to be an artifact. In our view this rules out the older "depth" age even if possible magnetic anomaly identifications could be found in the Mesozoic.

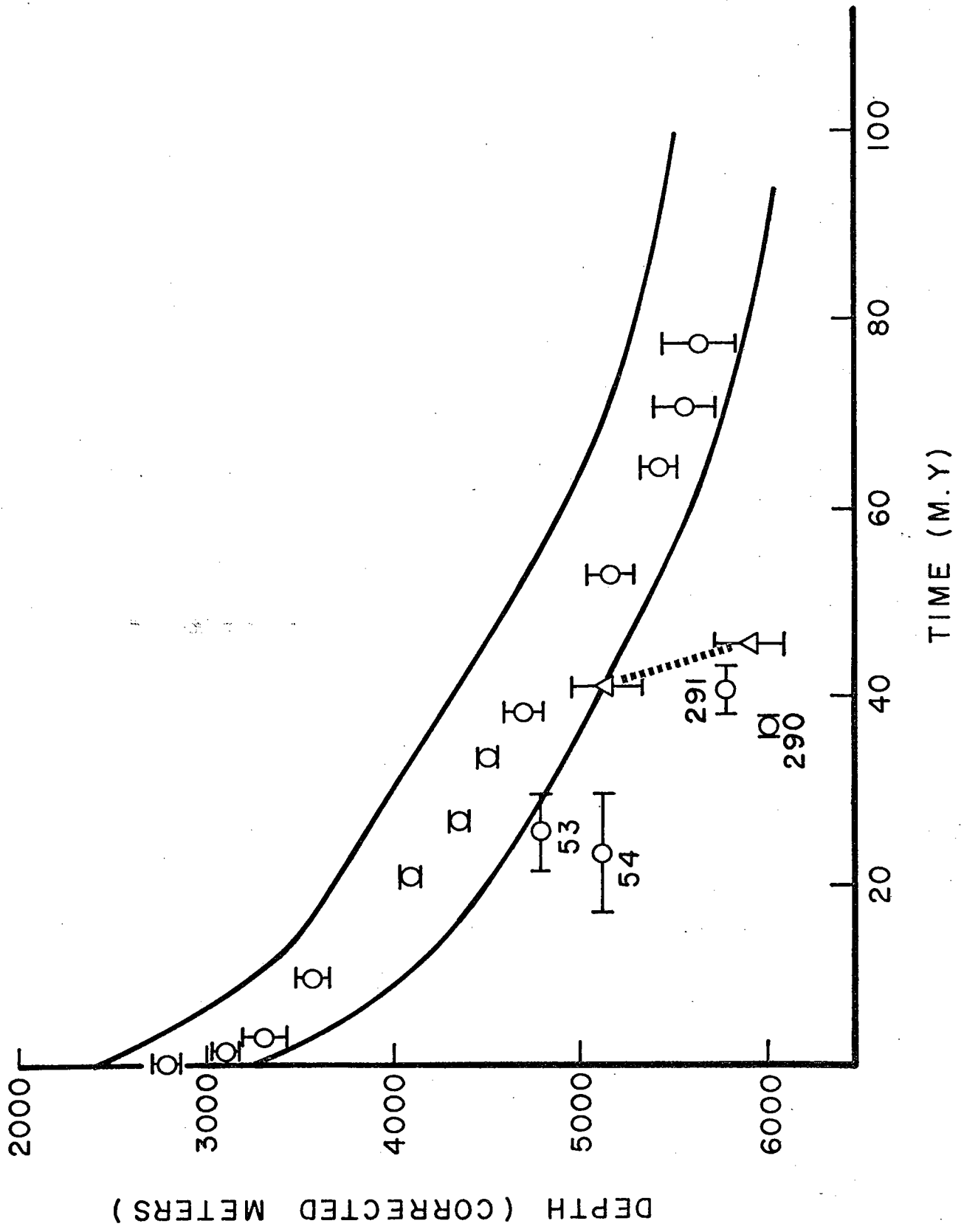
CRUISE ID	SMOOTHED BATHYMETRIC DEPTHS (uncorrected km)		
	South Flank	Ridge Crest	North Flank
KH 72-1	5.8	5.1	6.0
ANT 13	5.7	4.8	5.7
TAS 7	5.6	4.7	5.7
GC 31	5.6	5.0	5.8
PIONEER 1	—	4.9	—
TAS 5	5.3	4.9	5.6
AVERAGE	5.6±0.3	4.9±0.2	5.8±0.2

TABLE III



FIGURE 8

Depth of smoothed topography vs. age. Data for the North Pacific (circles between heavy lines) is from Sclater et al. (1971) and is compared to values for JOIDES sites 53, 54, 290 and 291 (JOIDES, 1969, 1973). We also plot values treating the Central Basin Fault as a symmetric spreading ridge (triangles) taken from Table III and figure 5. Dashed line connects ridge crest to flank.



(3) There are recent seismic studies which indicate that the crust in the West Philippine Basin is not similar to North Pacific crust (Shimamura et al., 1975; Henry et al., in press). In particular, the thin "oceanic" crust suggested by the refraction studies of Henry et al. (in press) can help to explain the occurrence of a 1 km negative anomaly without a correspondingly large negative gravity anomaly (Sclater et al., in prep.).

#### CONCLUSIONS

All of the data presented in this paper can be accounted for by treating the Central Basin Fault as a spreading center that slowed and stopped 40 m.y. ago. The ridge-like bathymetry and symmetry in the magnetic lineations are the strongest support for such a conclusion. The ages of JOIDES sites 290 and 291 also can be explained with this model, but because of the incomplete agreement between fossil ages and magnetic time scales for this period, coupled with the limited extent of our lineations and the fast spreading rates needed to match the anomalies, they cannot remove all uncertainty in our identification.

It is very tempting to go beyond the limit of such a conclusion and propose a comprehensive tectonic model that connects the formation of this basin with a proto-Pacific reconstruction. Such an attempt has already been made by Uyeda and Ben-Avraham (1972) and Uyeda and Miyashiro (1974). They propose that the Philippine Ridge (Central Basin Fault)

was connected by means of a long transform fault to the Kula-Pacific ridge system. When the Pacific plate changed its direction of motion from north to north-west 42-44 m.y. ago (Claque and Jarrard, 1973), the transform fault became a subduction zone, leading to the start of inter-arc spreading and the formation of the Parece Vela Basin. There are several specific problems with such a model:

(1) The Palau-Kyushu Ridge is asserted to be the current expression of the transform fault between the Kula-Pacific and the Philippine Ridges. If so, the fault must have extended for several thousand kilometers. This is much greater than for most presently active transform faults. The only purely oceanic transform fault of such a large extent that has been magnetically documented is the Ninety-east Ridge (Sclater and Fisher, 1974), which is no longer active. There are also other interpretations of the origin of this feature (Bowin, 1973). Other long transform faults are either on continental crust (San Andreas Fault) or are not yet supported by unequivocal data (the connecting link between the South Atlantic Ridge and the West Indian Ocean Ridge). It is possible that the mechanics of ocean floor spreading make the possibility of such a long transform fault lasting over an extended period of time unlikely (Lachenbruch, 1973).

(2) The magnetic anomalies identified in this paper appear to have originated near to or just south of the equator. This

invalidates the location of the ridge in its present latitude when spreading ceased, which is central to the Uyeda and Ben-Avraham (1972) model. Their model also predicts very different subsequent motions between the Philippine and Pacific plates. It is interesting to note, however, that our results (formation of anomalies from  $0^{\circ}$  to  $-16^{\circ}$ ) can be explained by using the same pole of "absolute" rotation for the Philippine plate as has been determined for the Pacific plate by the hot-spot theory (Clague and Jarrard, 1973). A pole at  $72^{\circ}\text{N}$ ,  $83^{\circ}\text{W}$  and a rate of  $13.10^{-7}$  deg/yr yield a paleolatitude of  $-10^{\circ}$  for 50 m.y.b.p.

(3) The elevation of the West Basin is depressed by 1 km from standard oceanic depths of corresponding ages taken from the North Pacific (Sclater et al., 1971). Such a negative residual depth anomaly should produce a negative gravity anomaly of  $30 \pm 10$  mgals if it follows a similar relationship as the North Atlantic (Sclater et al., in press). This is not observed. It indicates that the crustal structure of the West Basin is different from that of the North Atlantic and possibly from other ocean basins. It follows that it could not have originated in an identical process.

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SECTION III

PALEOMAGNETISM OF DSDP SEDIMENTS AND BASALT, PHASE  
SHIFTING OF MAGNETIC ANOMALIES, AND ROTATIONS OF THE  
WEST PHILIPPINE BASIN

ABSTRACT

A paleomagnetic study of sediments from DSDP sites 290, 292 and 294 suggests that the West Philippine Basin originated between 5°S and 10°S. Results for site 292 are particularly consistent; a group of 20 closely spaced samples at the sediment-basalt interface has a mean paleolatitude and 95% confidence interval of  $\pm 4.9^\circ \pm 2.0^\circ$ . This is much more definitive than a group of 36 basalt samples from the same site which shows a possible range of  $\pm 11.4^\circ \pm 21.0^\circ$ . This large interval is due to an incomplete sampling of secular variation which is not a problem for groups of sediment samples. The basalt is less stable than the sediment and may be contaminated by viscous components, although a study of VRM acquisition during 1000 hrs. in both a 1.0 and 0.5 oe field suggests that laboratory measurements may overestimate the actual viscous growth which occurs in geologic time.

Crustal origin of the West Philippine Basin in southern latitudes is also consistent with the phase shifting of marine magnetic anomalies which must be inverted to be matched with a set of world-wide Eocene reversals. Two well-defined paleomagnetic pole positions are determined if the low amplitude of these East-West anomalies, when compared to similarly oriented anomalies in the Pacific and Indian oceans, is due to a rotation of this plate. One of these is consistent with paleopoles calculated from the "absolute" motion of the Pacific and directions of relative motion between the Philippine and Pacific plates. This suggests a clockwise rotation of 60°

between the Philippine plate and the magnetic pole and is similar to results from the study of Miocene rocks on Guam. This may imply that both Guam and the West Philippine Basin have undergone the same past plate motion and that the rotation of Guam is not due to a simple bending of the Mariana arc.

## INTRODUCTION

There are essentially four methods by which we can determine the rotations of the plates relative to an external frame of reference. These are: (1) the paleomagnetic study of basalts and sediments, (2) the phase or shape analysis of linear, sea-floor spreading and seamount magnetic anomalies, (3) the bathymetric mapping of hot-spot traces, and (4) the location of deep water regions of high sedimentation, which are assumed to be produced within  $\pm 5^\circ$  of the paleoequator. If it is then assumed that the earth's paleomagnetic field has time-averaged over  $10^4$  to  $10^5$  years to an axially geocentric dipole concordant with the spin axis [Opdyke, 1972], and that hot spots have remained fixed with respect to this same axis, then each of these measurements are directly comparable. Such 'absolute' motions have been determined in this manner for the South Atlantic [Phillips and Forsyth, 1972; Lowrie et al., 1973b], North Pacific [Francheteau et al., 1970; Claque and Jarrard 1973; Harrison et al., 1975] and Indian oceans [McKenzie and Sclater, 1971; Blow and Hamilton, 1975], and results based on the different techniques are generally consistent [Peirce, in press]. These studies have been particularly successful in recording the degree of northward motion since the Cretaceous, which is an especially important characteristic of the Pacific and Indian plate motions.

It is the attempt of this paper to use some of these same techniques for placing constraints on the early history

and subsequent motion of the Philippine plate. A comparison of the 'absolute' motions of the Pacific and Philippine plates can hopefully lead to some understanding of how these two plates have moved relative to each other. This is important since the nature of relative motions produced by inner-arc spreading is not well understood. Determinations of instantaneous motion are based almost entirely on fault plane solutions [Katsumata and Sykes, 1969; Fitch, 1972], and the results are only poorly constrained as evidenced by the many different locations of possible Philippine-Pacific poles [Karig, 1975]. Almost nothing certain is known about motions over a finite length of time.

The primary reason for this is the absence of easily identifiable magnetic anomalies in both active (young) and inactive (old) basins of the Philippine Sea. Recently, however, Loudon (1976) has identified some anomalies in the West Philippine Basin across the Central Basin Fault (Figure 1 and 16). This region is one of the older sections of the Philippine Sea, but controversy still exists over its exact age and origin. One theory forms the West Philippine Basin by inner-arc spreading during an Eocene pulse similar to, but in a different direction from, that which later opened the more eastern Parece Vela and Shikoku basins [Karig, 1975]. Another explains it as a remnant of the Mesozoic Kula-Pacific ridge system which was trapped when the 'absolute' motion of the Pacific changed direction 40 m.y. ago [Uyeda



FIGURE 1

Bathymetric map of the Philippine Sea from Chase et al. (1969) with uncorrected thousand fathom contours and selected DSDP site locations. The dashed line encloses area of Figure 16.

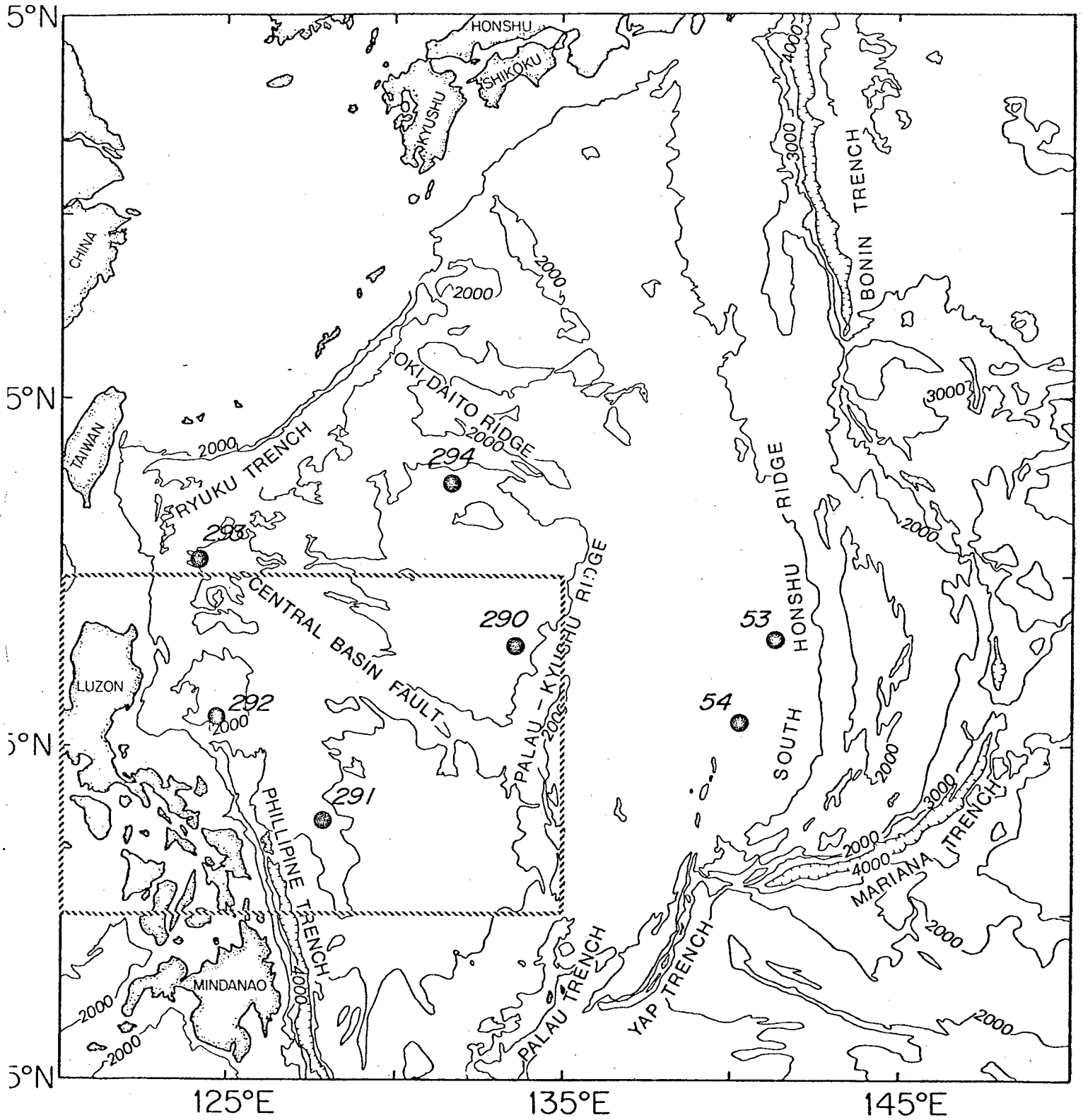


FIG 1

and Ben-Avraham, 1972]. This later theory is inconsistent with ages based on JOIDES cores, magnetic anomaly identifications and heat flow averages, which all indicate an Eocene age [Louden, 1976]. Only the 6 km depth of the basin which gives a minimum age of 80 m.y. from the depth vs. age curves of the Pacific [Sclater et al., 1971], is consistent with this interpretation; but it is not certain whether depth vs. age curves for major ocean basins can be applied to marginal basins [Sclater et al., 1976].

Initial analysis of the skewness of the magnetic anomalies indicated that they must have been formed 0-16° south of the equator. The total amount of northward motion is very similar to that of the Pacific plate over the same time period (40-50 m.y.). In this paper we will add to these results a study of paleolatitudes derived from sediments taken at three DSDP sites (290, 292 and 294) and basalt from one site (292). We also have reanalyzed the magnetic anomalies using a slightly larger data set. These two analyses (phase shifting and paleomagnetism) can be combined to determine geomagnetic pole positions well constrained in latitude but not in longitude. In addition, the low amplitudes of our magnetic anomalies suggest that they must have undergone significant declinational change since they were formed, if their origins are similar to those of anomalies in other ocean basins. One of the two possible pole positions that are consistent with these magnetic anomaly phase and amplitude criterion plus the paleomagnetic inclinations which we

measure in DSDP sediments and basalt, would require significant rotations ( 80-100°) about a relative Pacific-Philippine pole similar to those determined from present day motions. Paleomagnetic results for basalts from Guam [Larson et al., 1975) can also be explained by the same "polar wandering" curve.

## PALEOMAGNETISM OF SITE 292 BASALTS

### Introduction

Previous studies of DSDP basalts often show well grouped inclinations which are consistent with plate reconstructions [Lowrie, 1974; Blow and Hamilton, 1975; Peirce, in press]. These results have not, however, been nearly as useful as paleomagnetic data from land basalt upon which the reconstructions are based, because they lack declination control and often do not sample separate flow units. This means that they may not record pole positions for a time interval long enough to average out non-dipole and dipole fluctuations away from the spin axis. Angular standard deviations of such variations can be as large as 20° [Baag and Helsley, 1974]. Another complication is that many DSDP basalts show a large amount of viscous remanent magnetization (VRM), which in several cases has made their inclinations unreliable [Lowrie et al., 1973a; Peirce et al., 1975].

The only land based paleomagnetic measurements relevant to the Philippine Sea are those of Larson et al. [1975]

for Miocene volcanic rocks on Guam. Their results suggest that Guam has not changed latitude since the Miocene, but has rotated clockwise by 50°. This may record the buckling of the island arc as the Parece Vela Basin opened and does not necessarily apply to the earlier history of the West Philippine Basin. Guam is situated on the boundary between the Pacific and Philippine plates and its motion may represent distortions in that boundary rather than the actual plate motions themselves. It is therefore necessary to look at the paleomagnetic record of DSDP basalts in the West Philippine Basin, keeping in mind the probable limitations.

Site 292 on the Benham Rise is the only large volume of basalt recovered during DSDP leg 31: 76 m of basalt was cored of which 72 m were recovered. This is a uniform intersertal to subophitic tholeiite with plagioclase and augite phenocrysts. Some sections show a slightly greenish color due to alteration effects. Although a few individual units can be distinguished, there are no major discontinuities and the entire 72 m probably represents a single flow unit [Ingle et al., 1975]. The following sections describe the results of our paleomagnetic study of 36 samples of this basalt. Measurements of basalt magnetization were made on a PAR spinner magnetometer operating at 15 Hz.

#### Alternating Field Demagnetization

Eleven of our 36 samples with a wide range of NRM intensities (12.0 to  $1.8 \times 10^{-3}$  gauss) and inclinations

(+13° to -77°) were subjected to alternating field (AF) demagnetization of up to 200 oe. peak field. In most cases measurements were repeated once again at each field value. We used a two-axis tumbler and a 60 Hz alternating field which decreased from peak field to zero in 2.5 min. Results are shown in figures 2 and 3. By 200 oe inconsistent directions or intensities under double demagnetization made results at higher fields suspect. Median destructive fields (figure 2) ranged from 42 to 139 oe; 8 of 11 values fell between 75 and 95 oe. This loss of intensity with increasing demagnetization is much less variable from sample to sample than were the changes in direction of magnetization (fig. 3). Some directions varied little from NRM to fields of 200 oe while others had large variations. Symons and Stupavsky (1974) have suggested a stability index, PSI, for such cases in which the variation in magnetization direction is of greatest interest. The PSI is a measure of the angular change (in millidegrees) per oersted increase in demagnetizing field; the lower this value is, the more stable is the remanence. Figure 4 shows PSI vs. peak demagnetizing fields for our 11 pilot samples. Low values are reached around 50-75 oe and most fall within the moderate to low stability range of Symons and Stupavsky [1974].

One disadvantage of the PSI calculation is that it does not account for inconsistencies between multiple measurements at the same demagnetizing field. For instance, samples 42-3-123 and 42-5-121 show low PSI readings at 200 oe (figure 4), but the stereo-net plots (figure 3) show large angular

TABLE I continued:

ID	$J_0$	$J_{75}$ ( $10^{-3}$ Gauss)	$J_{100}$	$D_0$	$D_{75}$ (Deg)	$D_{100}$	$I_0$	$I_{75}$ (Deg)	$I_{100}$	$\lambda_{75}$ (Deg)	$d$	$k$ ( $10^{-3}$ Gauss/oc)	$Q_n$
45-1-76	<u>3.19</u>	1.28	.725	<u>53.1</u>	44.4	43.5	<u>-13.1</u>	-8.9	-10.9	(-4.5)	14.1	1.34	2.38
45-2-28	<u>5.67</u>	2.57	1.61	<u>65.0</u>	52.8	54.3	<u>-20.0</u>	-16.9	-22.5	-8.6	3.9	1.29	4.40
45-2-78*	<u>.900</u>	<u>.478</u>	<u>.306</u>	<u>172.4</u>	<u>170.5</u>	<u>167.8</u>	<u>-33.4</u>	<u>-22.8</u>	<u>-22.1</u>	<u>-11.9</u>	2.6		
45-2-132	<u>8.90</u>	6.53	4.83	<u>164.2</u>	174.6	175.3	<u>-24.5</u>	-24.9	-24.0	-13.1	1.8	1.26	7.06
45-3-7	9.69	7.50	6.32	38.8	33.3	32.7	-14.6	-18.1	-18.2	-9.3	0.57	1.01	9.59
45-3-126	13.9	11.7	10.1	44.6	44.0	42.9	-31.8	-33.6	-34.2	-18.4	1.1	1.33	10.5
46-1-25	<u>2.70</u>	1.95	1.35	<u>154.0</u>	163.3	164.3	<u>-18.5</u>	-17.9	-22.3	-9.1	2.6	.83	3.25
46-1-136	<u>1.04</u>	.714	.323	<u>46.9</u>	37.0	41.1	<u>-18.6</u>	-16.3	-11.3	(-8.3)	13.1	.82	1.27
46-2-116	2.00	1.42	1.11	13.4	357.5	354.6	-21.5	-10.8	-13.2	-5.4	4.4	1.32	1.52
47-1-47	6.06	1.46	.880	93.9	90.4	82.8	-22.2	-32.5	-36.4	(-17.7)	5.4	1.20	5.05
47-2-31	3.75	<u>.734</u>	<u>.447</u>	97.4	<u>135.9</u>	<u>107.1</u>	-1.9	-4.3	-43.7	(-2.2)	46.9	1.16	3.23
47-2-130	<u>2.81</u>	.943	.489	<u>44.3</u>	23.3	22.0	<u>-15.7</u>	-27.2	-38.4	(-14.4)	7.1	<u>1.27</u>	<u>2.21</u>
	6.32									mean -11.4	(-10.3)	1.28	5.26
										$\sigma$			
										$\sigma/\bar{N}$			
										3.8	(6.4)		
										0.8	(1.1)		

The ID gives the core no., section no. and position in cm. from the top of the section to the top of the sample. Each sample is approximately 10 cm<sup>3</sup>. Subscripts refer to fields of peak AC demagnetization.  $\lambda_{75}$  is the paleo-latitude after 75 oe demagnetization.  $d$  is a measure of the reliability as defined in the text. Values in parentheses are only included in the bracketed mean and standard deviation. Underlined values represent one measurement; all other values are averages of two separate measurements.

\* volume is not measured, but is assumed to be 10 cm<sup>3</sup>.

TABLE I: Summary of Remanent Magnetic Properties for DSDP Site 292 Basalts.

ID	$J_0$		$J_{75}$ ( $10^{-3}$ Gauss)		$J_{100}$		$D_0$	$D_{75}$ (Deg)	$D_{100}$	$I_0$	$I_{75}$ (Deg)	$I_{100}$	$\lambda_{75}$ (Deg)	d	k ( $10^{-3}$ Gauss/oc)	$Q_n$
	$J_0$	$J_{75}$	$J_{75}$	$J_{100}$	$D_0$	$D_{75}$										
40-1-103	11.8	6.60	4.29	194.2	197.8	198.6	12.7	13.8	14.0	7.0	1.4	1.17	10.1			
41-3-32	3.55	1.84	.959	348.2	333.3	322.5	-21.5	-25.2	-20.1	(-13.2)	9.1	1.52	2.34			
41-3-127	4.99	3.27	2.06	217.9	221.1	219.3	-22.1	-21.5	-14.7	-11.1	2.9	1.13	4.42			
41-4-51	5.81	2.34	2.09	129.7	130.1	128.0	-13.4	-14.7	-14.6	-7.5	2.0	1.27	4.57			
41-5-8	17.2	13.4	10.3	156.2	157.7	157.6	-17.1	-17.2	-17.7	-8.8	0.87	.85	20.2			
41-5-117	14.1	12.0	10.2	321.7	320.1	320.5	-17.2	-18.6	-18.0	-9.6	0.57	1.66	8.49			
42-2-12	2.44	1.43	.905	119.9	115.1	110.8	-3.4	-7.8	-6.5	-3.9	4.5	1.43	1.71			
42-2-144	2.15	1.22	.602	288.9	275.9	271.9	-26.1	-26.2	-23.4	(-13.8)	14.3	1.34	1.60			
42-3-123	3.97	2.40	1.79	340.6	321.0	332.0	15.5	24.0	25.0	(12.6)	10.1	1.64	2.42			
42-4-62	9.03	6.42	4.18	298.1	295.4	294.3	-20.8	-20.6	-22.3	-10.6	1.5	.89	10.1			
42-5-22	10.7	9.38	8.66	294.7	292.2	290.6	-29.2	-28.7	-29.0	-15.3	1.1	1.18	9.07			
42-5-121	12.0	10.4	8.72	223.9	225.9	226.4	-29.7	-29.3	-30.3	-15.7	1.2	1.06	11.3			
43-2-41	7.10	4.61	3.15	120.2	124.9	127.0	-10.5	-12.0	-8.5	-6.0	2.9	1.21	5.87			
43-2-120	5.69	2.02	1.47	31.8	349.4	350.2	-28.9	-27.2	-27.7	-14.4	4.6	1.52	3.74			
43-3-90	5.43	2.33	1.15	206.5	214.0	219.6	-25.0	-33.1	-34.9	-18.0	3.9	1.26	4.31			
43-4-50	4.39	2.18	1.42	76.2	72.6	71.6	-19.9	-22.5	-18.1	-11.7	4.5	1.36	3.23			
43-4-102	6.05	2.09	1.30	82.3	72.6	71.1	-16.2	-18.8	-16.1	(-9.6)	9.7	1.63	3.71			
44-2-19	4.02	2.51	1.65	146.2	154.7	153.4	-26.9	-21.3	-21.4	-11.0	3.8	1.19	3.38			
44-2-138	3.71	3.03	1.93	197.0	217.0	215.6	-17.2	-27.3	-24.6	-14.5	4.4	1.44	2.58			
44-3-128	6.34	4.39	2.97	244.5	246.0	247.6	-30.9	-27.0	-26.5	-14.3	2.3	1.06	5.98			
44-4-23	7.55	4.44	2.71	162.1	169.6	166.8	-32.6	-28.9	-27.4	-15.4	1.6	1.49	5.07			
44-4-141	5.25	2.12	1.30	151.2	162.8	162.7	-25.5	-29.0	-33.0	-15.5	4.3	1.87	2.81			
44-5-121	1.80	1.23	.769	321.5	258.8	258.7	-76.8	-32.6	-33.3	(-17.7)	13.2	1.49	1.21			



FIGURE 2

Demagnetization curves for selected site 292  
basalts. All repeated values are plotted  
except in a few cases where they are identical.

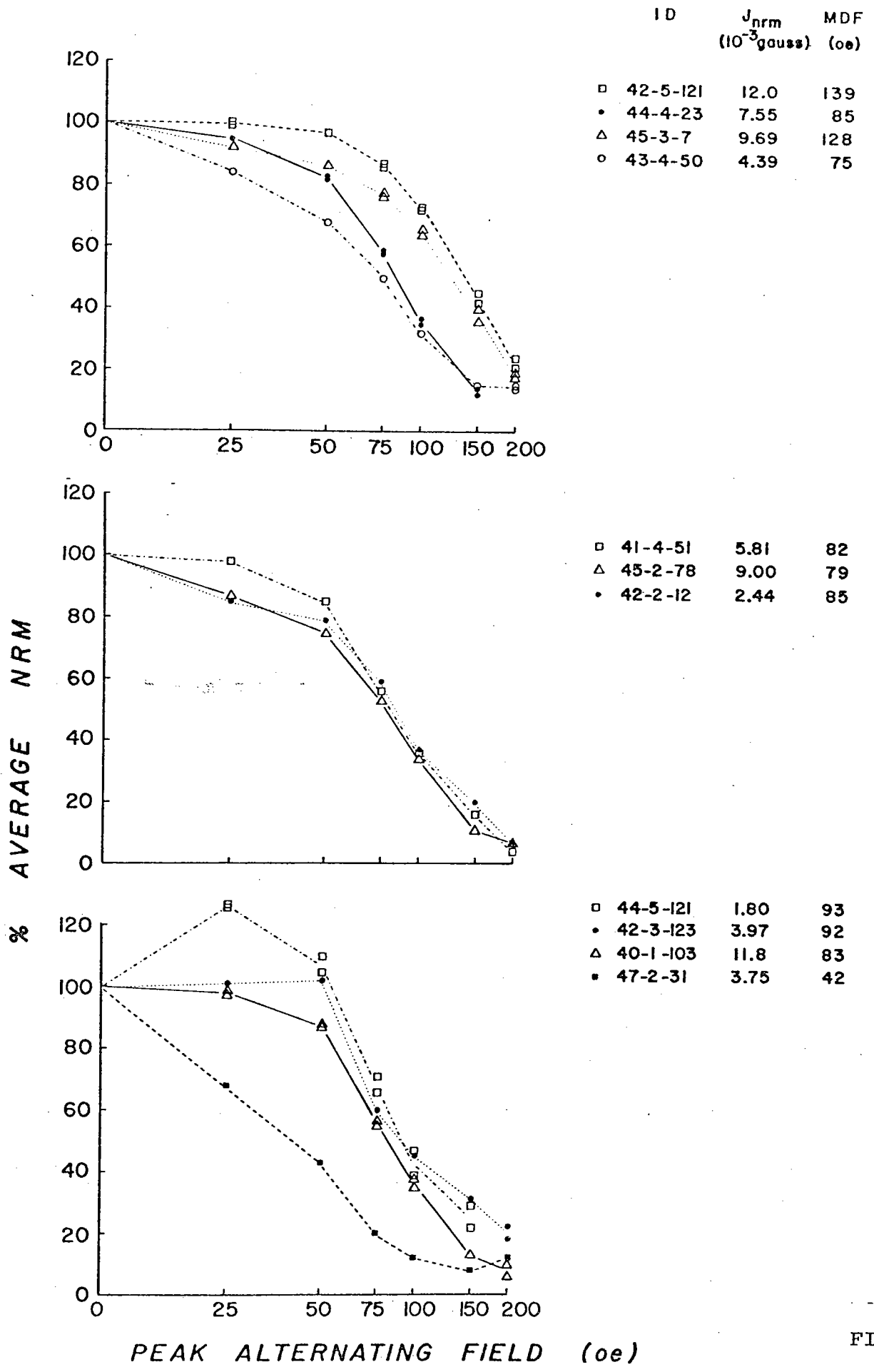


FIG 2

FIGURE 3

Stereonet plots of relative magnetization directions for the same samples as in Figure 2. All repeated measurements are shown and joined by dotted lines where greatly different. Parentheses enclose values of positive inclination except for samples 42-3-123 and 40-1-103 which have all positive inclinations (indicated by the + next to their ID number). All other inclinations are negative. Solid lines and arrows outline the path that magnetization directions follow after demagnetization at progressively higher fields.

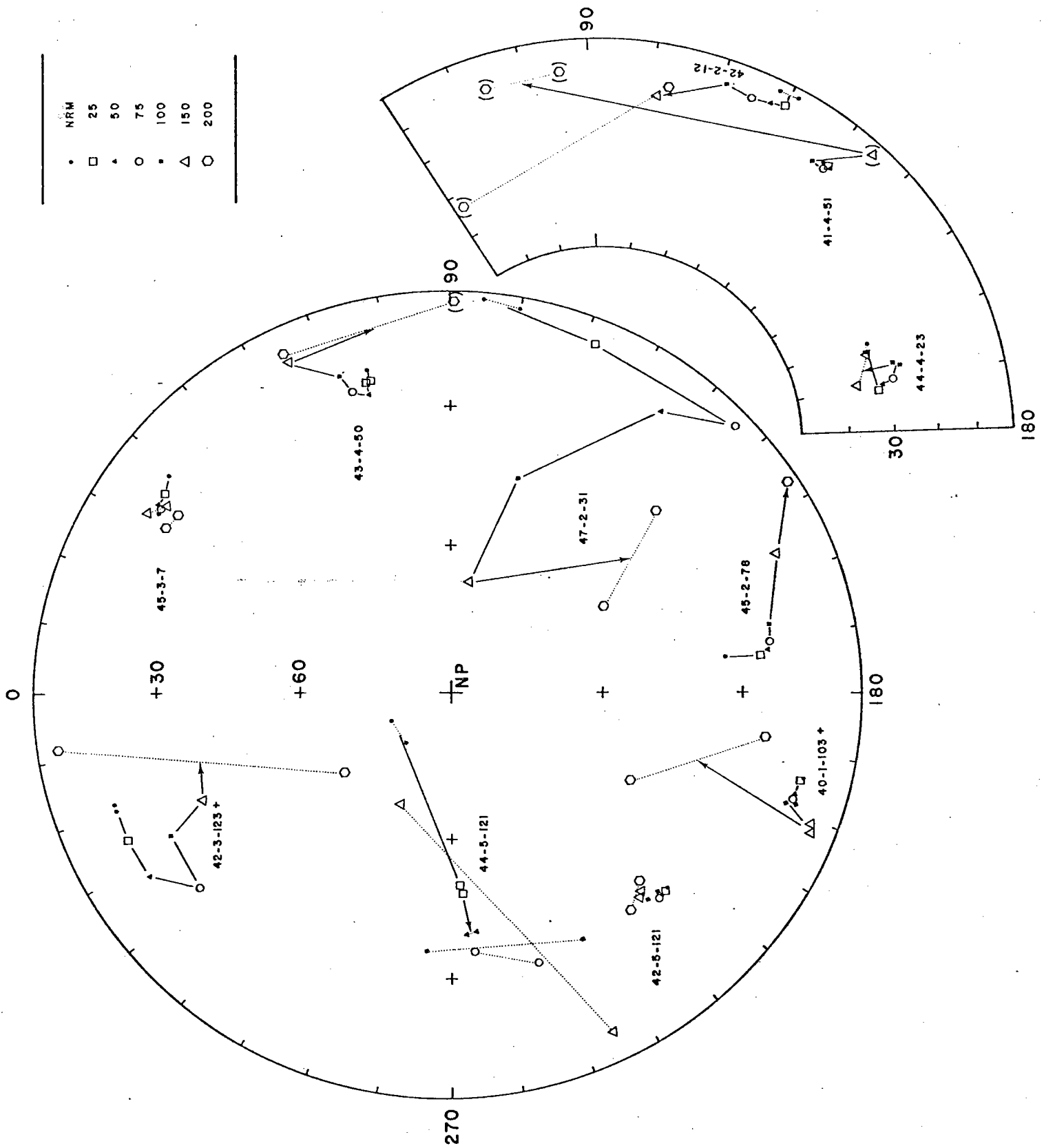


FIG 3

## FIGURE 4

PSI values as a function of demagnetizing field for the same site 292 basalts as in figures 2 and 3. The PSI index is a measure of stability as defined by Symons and Stupavsky (1974). Lowest values fall between 75 and 100 oe demagnetizing field and lie in the range of low to moderate stability.

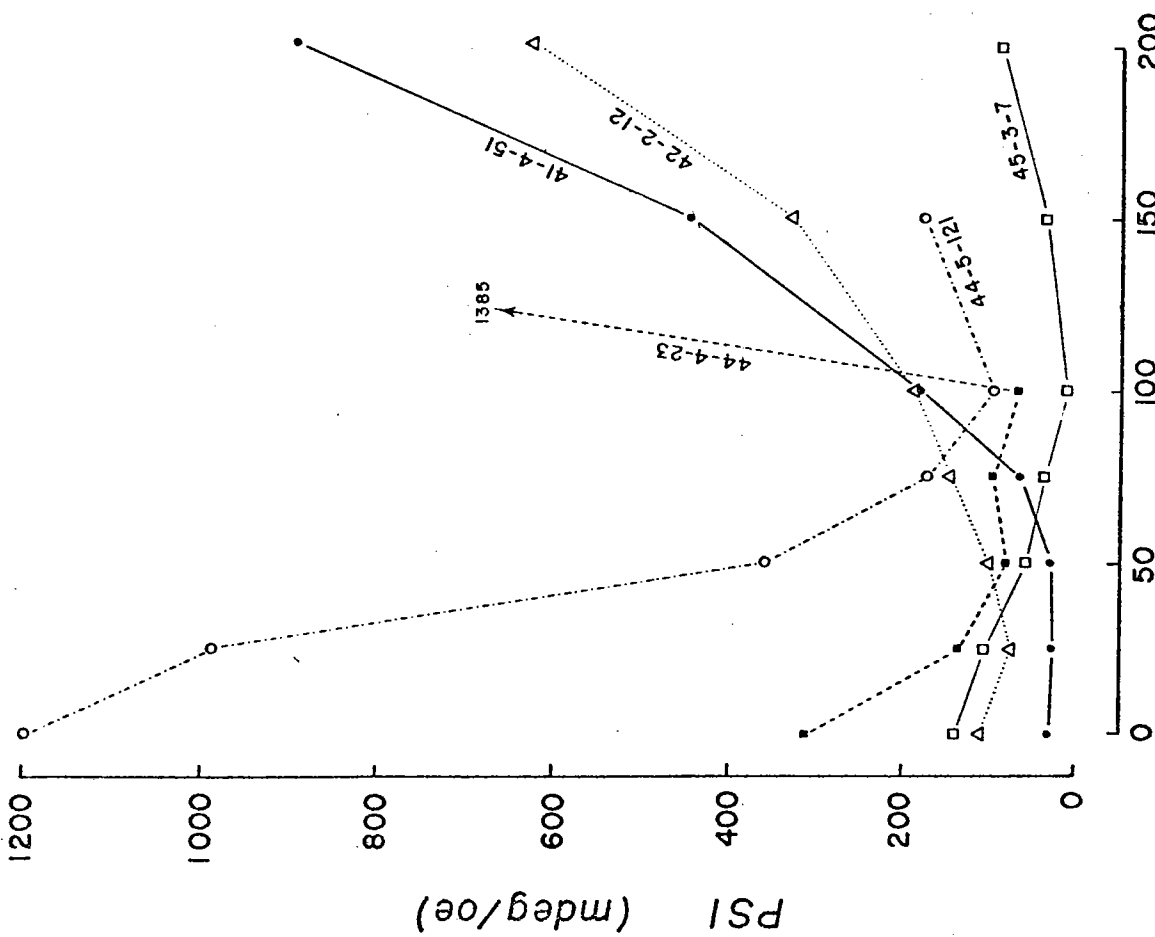
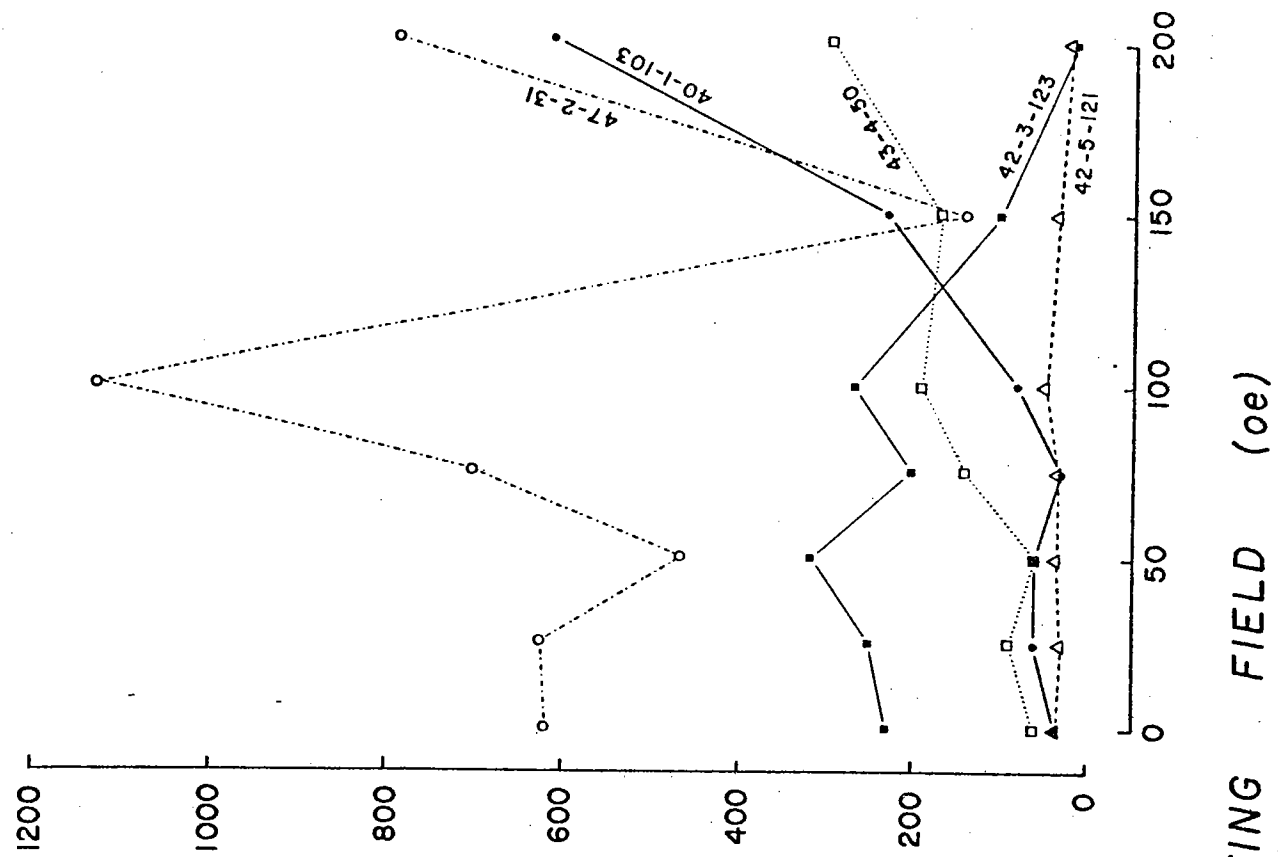


FIG 4

differences between the two separate determinations of the magnetization vector. In order to include this additional information about the consistency of magnetization direction, we calculate a stability index,  $d$ , given by the following equation:

$$d = (\alpha_{75-100} + \alpha_{75} + \alpha_{100})/3 \quad (1)$$

where,  $\alpha_{75-100}$  is the angular change between average magnetization directions at 75 and 100 oe, and  $\alpha_{75}$  and  $\alpha_{100}$  are the angular changes between repeated magnetization measurements at 75 and 100 oe respectively. Low values of  $d$  represent small angular changes in magnetization direction between 75 and 100 oe (as do the PSI values for these fields) in addition to small changes in repeated measurements at each field. We use an average of the three separate angular changes because repeated measurements were not always taken.

A plot of  $d$  vs. NRM intensity ( $J_{\text{nrm}}$ ) is given in figure 5. It appears that there is a general increase in stability (i.e. lower values of  $d$ ) for samples with higher values of  $J_{\text{nrm}}$ . Acquisition of viscous magnetization might be one explanation for this. All but two of our samples are reversely magnetized, with inclinations close to, but of opposite sign from that of the present field. This means that viscous components acquired since the beginning of the Bruhnes epoch would reduce the amplitude of the originally reversed magnetization. We know that it is difficult to remove VRM by AF demagnetization

FIGURE 5

Stability index  $d$  (defined in the text) vs.  
NRM intensity  $J_{\text{nrm}}$  for all site 292 basalts.  
Low stability (high values of  $d$ ) at low  $J_{\text{nrm}}$   
suggests that viscous components may be a partial  
cause of their instability.



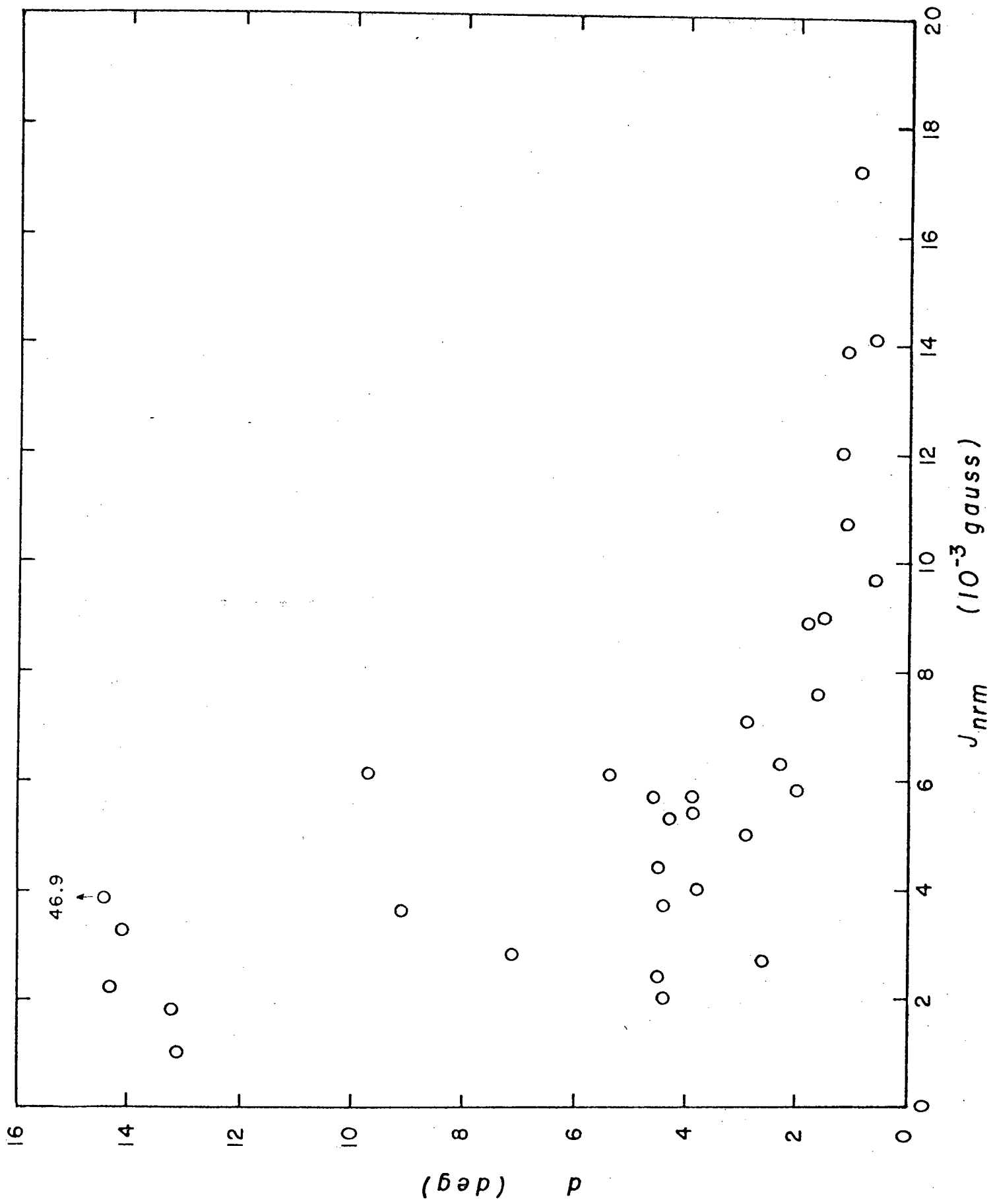


FIG 5

[Biquand and Prevot, 1971], so it is quite possible for it to contaminate even our demagnetized results. If, however, we can show that there is a relationship between the degree of VRM acquisition and our stability parameter  $d$ , then we might be able to confidently reduce the effects of VRM in our paleolatitude determination by using only those samples with small  $d$ .

#### Acquisition of Viscous Remanent Magnetization

The predominant feature of viscous remanent magnetization (VRM) is its acquisition in an applied field  $H$  as a function of the logarithm of time:

$$J(t) = J_0 + S (\log t - \log t_0) \quad (2)$$

where  $J$  is the magnetic intensity,  $t$  is the time spent in the field  $H$ , and  $S$  is the magnetic viscosity coefficient. Such behavior is predicted by all theories of thermally activated magnetic viscosity, primarily due to Neel [1950] and Stacey [1963], and has been observed in both single and multi-domain terrestrial materials. Lowrie [1973] has shown that it is also an important factor in the instability of some DSDP basalts. On the basis of laboratory VRM growth during 1000 hours, he extrapolated values of  $J(t)$  to 0.69 m.y. as an estimate of the viscous growth during the Bruhnes normal epoch. If such a value is a large portion of the total NRM, this indicates that an initial remanence may have been destroyed by the later accumulation of viscous components.

There are several additional features of VRM acquisition which complicate the significance of such an extrapolation. First, the  $\log t$  relationship is critically dependent on the nature of the grain distribution  $f(v, H_{CO})$  for non-interacting SD grains, where  $v$  is the volume and  $H_{CO}$  the microscopic coercive force (Dunlop, 1973). Consequently, if a sample contains discontinuities in  $f(v, H_{CO})$ , the slope  $S$  may change with time. Grain sizes and coercivities affected during 1000 hours are quite different from those affected over  $10^6$  years, and extended extrapolations in time based on constant  $S$  could become inaccurate. Second, the temperature at which acquisition occurs has an important influence on  $S$  through its effect on the relaxation time by which an initial magnetic moment decays. This relaxation time  $\tau$  depends exponentially on the thermal energy, and if the temperature changes from  $T_1$  to  $T_2$ ,  $\tau$  will shift from  $\tau_1$  to  $\tau_2$  according to the relationship

$$T_1 \ln(C\tau_1) = T_2 \ln(C\tau_2) \quad (3)$$

where  $C=10^{10} \text{ sec}^{-1}$  [Neel, 1950]. A change from oceanic bottom water ( $T=275^\circ\text{K}$ ) to laboratory ( $T=293^\circ\text{K}$ ) conditions decreases  $C\tau$  to  $(C\tau)^{0.94}$ , which will have the effect of increasing  $S$ . Third, our measured values of  $S$  may not correspond to the real geological acquisition rate. Our laboratory procedure, in which we first demagnetize the sample in an alternating field, may activate VRM susceptibility by rearranging domain walls into less durable positions, from which subsequent VRM is more

easily acquired. All of these effects are not well understood and we should proceed with caution when using laboratory VRM studies in analysis of NRM measurements.

We have conducted some traditional VRM experiments to determine the extent to which 6 samples acquired magnetization during a period of up to 1000 hours after initial demagnetization in a field of 500 oe. Two experiments were conducted on the same samples, the first using a field of 1.0 oe to produce the VRM and the other using a 0.5 oe field. Results are shown in figure 6. For each sample the shape of the best fitting curve (using the least squares fit of a 3rd order polynomial) differs between the 1.0 oe and 0.5 oe cases. In most instances there are significant changes in slope during the 1000 hours, with the 1.0 oe results showing increasing slopes similar to those reported by Creer et al. [1970], and the 0.5 oe showing a three stage pattern previously observed by Lowrie [1974]. Such changes in slope have been attributed to the dependence of the viscosity coefficient  $S$  on grain size distribution. The region of large slope is thought to be due to growth of VRM by single domain grains; the smaller slopes due to larger grains, possibly in the region of pseudo-single domains. But the fact that the shapes of our curves depend on the field strength acting on the samples suggests that the true reason for this behavior does not lie entirely with the grain size distribution.

After 30-50 hours the 1.0 oe slopes have become approximately twice the 0.5 oe slopes, as determined by a least squares straight line fit (see figure 6 and Table II). This confirms

FIGURE 6

Acquisition of viscous remanent magnetization as a function of the logarithm of time for selected site 292 basalts. Solid circles represent acquisition in a 0.5 oe field and open circles in a 1.0 oe field. Curves are the best least squares fit by a 3rd order polynomial to all of the points; dashed lines are the best linear fit to values for large t. Numbers indicate slopes of these straight line segments.

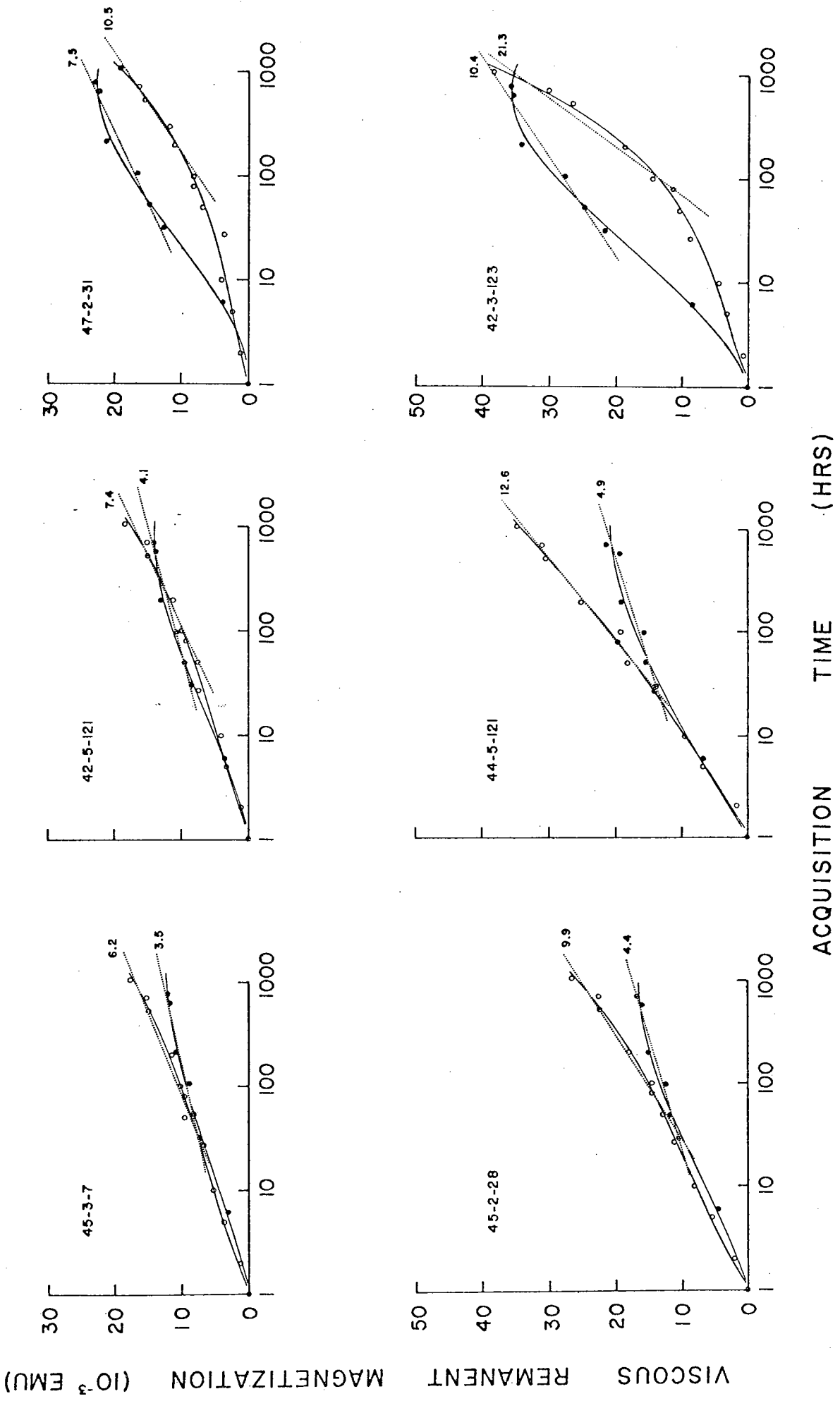


FIG 6

TABLE II: Summary of Viscous Magnetic Properties for DSDP Site 292 Basalts

ID	$H_{1000}^{(1)}$	$H_{700}^{(2)}$	$H_{0.69}^{(2)}$	$H_{nrm}$	$S^{(1)}$	$S^{(2)}$	$M_{700}^{(2)}$	$M_{0.69}^{(2)}$	$M_{nrm}$	$d$ (deg)
	(oe)						$(10^{-4} \text{ Gauss-cm}^3)$			
42-5-121	19	28	96	139	7.39	4.10	14.0	42.4	118.	1.2
45-3-7	37	28	96	128	6.22	3.54	12.2	36.7	93.	0.6
42-3-123	18	27	93	92	21.3	10.4	35.7	108.	37.	10.1
44-5-121	16	19	65	93	12.6	4.93	21.3	55.5	17.5	13.2
45-2-28	21	26	89	66	9.91	4.43	16.9	47.6	53.4	3.9
47-2-31	45	24	83	42	10.5	7.46	22.8	74.5	35.2	46.9

$H_{1000}^{(1)}$ , median destructive field (MDF) of viscous magnetization acquired in 1000 hours in a 1.0 oe field;  $H_{700}^{(2)}$ , MDF of viscous magnetization acquired in 700 hours in a 0.5 oe field;  $H_{0.69}^{(2)}$ , calculated MDF for a linear extrapolation of  $H_{700}^{(2)}$  to 0.69 my;  $H_{nrm}$ , measured MDF.

$S^{(1)}$  and  $S^{(2)}$ , viscosity coefficient in 1.0 and 0.5 oe fields respectively;  $M_{700}^{(2)}$ , magnetization acquired after 700 hours in a 0.5 oe field;  $M_{0.69}^{(2)}$ , calculated magnetization after 0.69 my using a linear extrapolation of  $M_{700}^{(2)}$  with a slope of  $S^{(2)}$ ;  $M_{nrm}$ , measured NRM magnetization.

$d$ , stability index as defined in the text.

previous results [Creer, 1957], which show that the magnitude of  $S$  depends upon  $H$ , in the Rayleigh region of small field strengths (<20-50 oe). To test whether the magnitude of these slopes is related to the instability of our samples under AF demagnetization, we plot  $S$  vs.  $\log d$  and  $S$  vs.  $J_{\text{nrm}}$  in figure 7. We find that in both 1.0 oe and 0.5 oe fields, the larger values of  $S$  are associated with more unstable rocks (larger  $d$ ) and with lower NRM intensities. The one exception is sample 42-3-123, but this specimen has a normal inclination and would not represent removal of a reversed magnetization by a positive field as would the others.

The extent to which an original remanent intensity might be destroyed by such rates of VRM growth are calculated in Table II. Using extrapolations based on magnetizations at 1000 hours in the 1.0 oe field ( $M_{1000}$ ) and 700 hours in the 0.5 oe field ( $M_{700}$ ), together with our linear determinations of  $S$ , we can calculate magnetizations acquired over the duration of the Bruhnes epoch ( $M_{0.69}$  m.y.). In Table II we only calculate  $M_{0.69}$  for a 0.5 oe field which should be closer to the actual Bruhnes earth's field acting on the ocean crust [McElhinney, 1973]. In two cases (47-2-31 and 44-5-121)  $M_{0.69}$  was larger than  $M_{\text{nrm}}$  although  $I_{\text{nrm}}$  was still reversed. This may indicate that our extrapolations are too large and that true viscous effects are smaller than indicated by our laboratory experiments. This is not surprising since the lack of a simple  $\log t$  relationship during the 1000 hours of our experiment, makes a linear extrapolation out to



## FIGURE 7

Viscosity coefficient  $S$  of selected site 292 basalts as a function of (a) stability index  $d$ , and (b) NRM intensity  $J_{\text{NRM}}$  for viscous acquisition in 1.0 and 0.5 oe fields. The direct relationship in (a) between  $S$  and  $d$  suggests that samples with low  $d$  are the least affected by viscous components. The inverse relationship in (b) suggests, as does figure 5, that acquisition of VRM causes decreased NRM intensities. This would be expected for samples such as these, whose inclinations are opposite in sign to that of the present field. The only dissimilar points in both (a) and (b) are for sample 42-3-123 which is the only one with positive inclination.

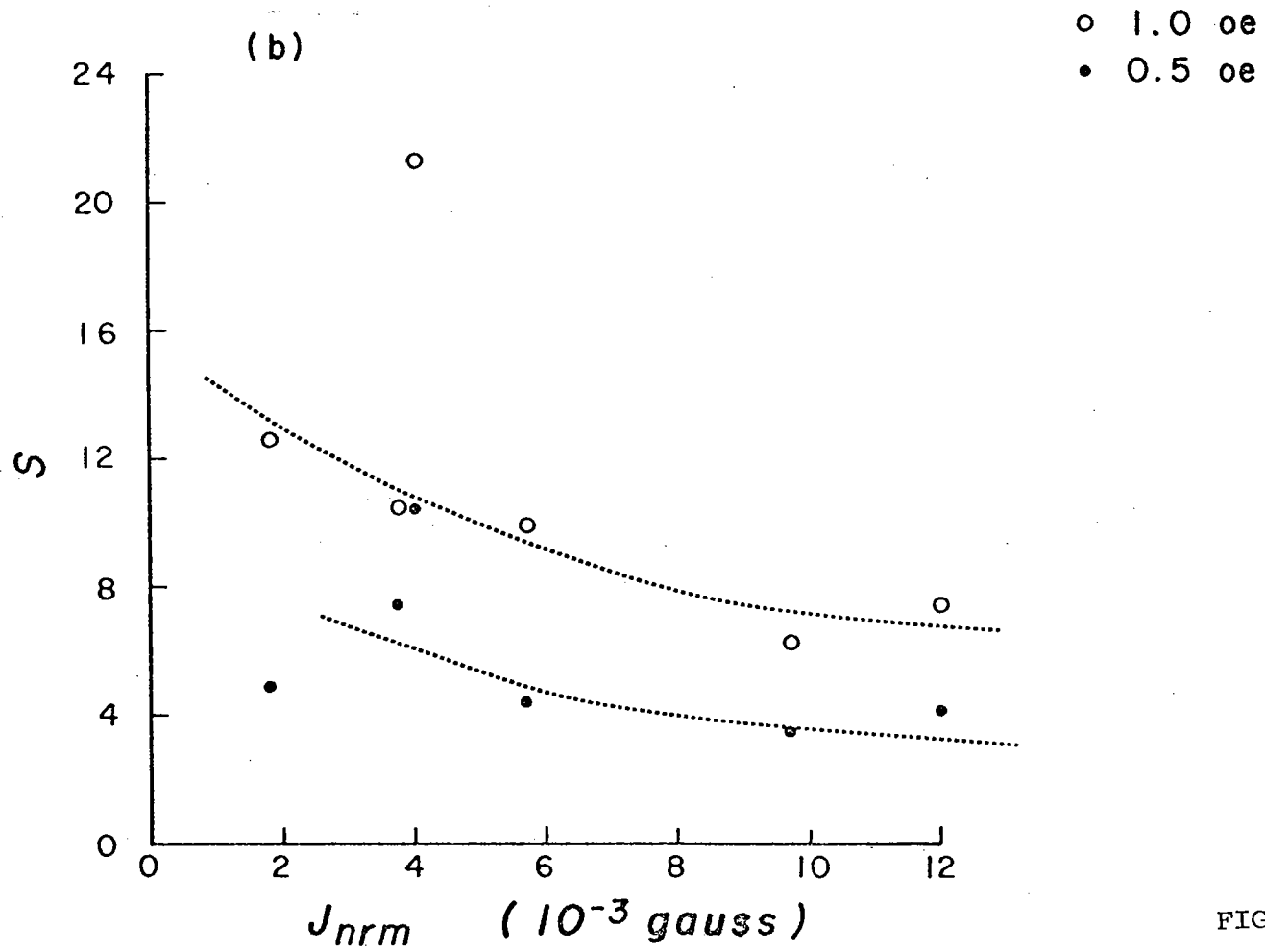
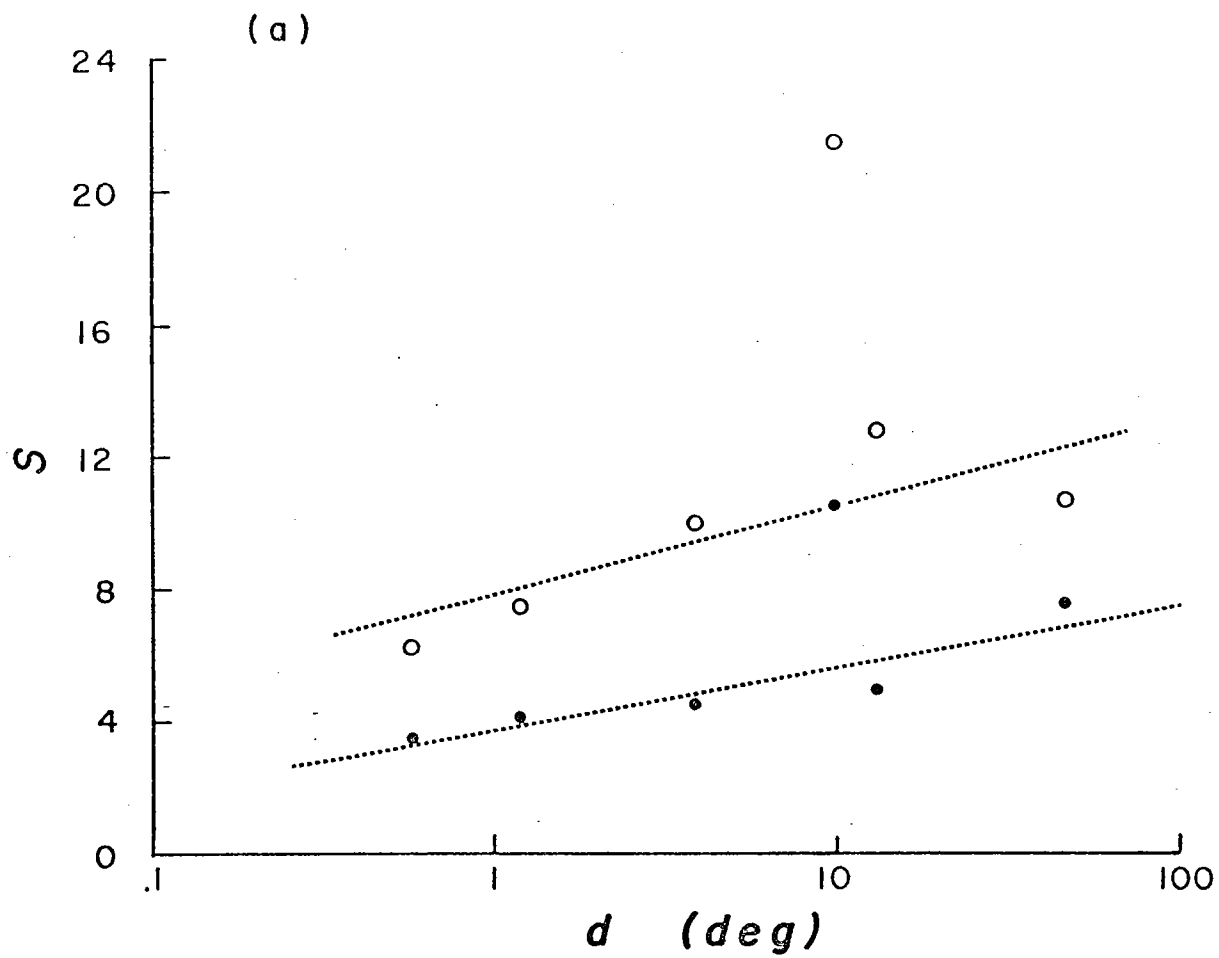


FIG 7

0.69 m.y. highly questionable. We still are not certain of the extent that viscous effects have hindered accurate determinations of remanent inclinations, but we do have some justification for expecting that samples with low values of  $d$  may be the least effected. The large values of  $S$  as measured in the laboratory must, however, add a certain skepticism as to the true meaning of our basalt inclinations.

### Paleolatitudes

In Table I we calculate a paleolatitude for each paleoinclination after AF demagnetization in 75 oe. We only consider values for  $d < 5$  and also have disregarded our one positive inclination with small  $d$  (sample 40-1-103). This serves to reduce the scatter in the result, but does not greatly shift the average paleolatitude calculated using all the data (Table I). Our result is that the mean paleolatitude is  $\pm 11.4^\circ$ . We use the arithmetic mean even though our numbers should actually be treated by the statistics on a sphere when only inclinations are known. Theoretical analysis of borehole data indicates that for high latitudes (particularly above  $50^\circ$ ) the true spherical mean should be as much as  $20^\circ$  higher than that given by the arithmetic average of the paleolatitudes, but that for low latitudes simple averaging of the paleolatitudes gives the same result as when they are treated on a sphere [Briden and Ward 1966; Cox, in preparation].

The southern latitude is consistent with the author's previous analysis of magnetic anomalies in the West Philippine Basin [Louden, 1976], but it is not unique because we do not

know whether the field was reversed or normal during the formation of these rocks. They do not lie on an identifiable magnetic anomaly. Also, we must add onto the experimental variance (given by the square of the standard deviation of our numerical population), a variance determined from probable models for secular variation. Our samples only record a moment in time, and the standard deviation of the mean paleolatitude cannot contain information concerning the secular variation. We use the relationship,

$$\text{S.E.} = \sqrt{(\sigma_{sv}^2 + \sigma_{exp}^2)/N_i} = 10.5^\circ \quad (4)$$

where, S.E. = standard error in our mean paleolatitude,

$$\begin{aligned} \sigma_{sv} &= 9.8^\circ \text{ [Cox, 1970]}, \\ \sigma_{exp} &= 3.8^\circ \text{ (from Table I), and} \\ N_i &= \text{number of flow units} = 1. \end{aligned}$$

The possible range of paleolatitudes,

$$\lambda_r = \underline{+11.4^\circ} \underline{+21.0^\circ},$$

represents the interval of 95% confidence, and extends from +33.4° to -33.4°. This large range tells us that basalts which do not adequately sample secular variation must have paleolatitudes greater than 30° for either polarity to be unique.

We must look to the sediment record for more significant data. In that case, where groups of samples span time intervals greater than  $10^4$  years,  $\sigma_{\text{exp}}$  does contain information concerning secular variation, and it is not necessary to add the additional term  $\sigma_{\text{sv}}$  to it.  $N_i$  would be the total number of values contained in the mean (i.e. the number of samples in each group) rather than the number of flow units. These changes should significantly reduce the standard error.

## PALEOMAGNETISM OF DSDP SEDIMENTS

### Introduction

The paleomagnetic study of deep-sea sediment cores has led to significant advances both in our knowledge of the magnetic reversal stratigraphy over the last 20 m.y., and in understanding the nature of secular variation and excursions in the earth's magnetic field of much shorter duration [Opdyke, 1972; Harrison, 1974; Creer et al., 1972]. The major advantage of sediments is that, unlike the basalts, they can offer a semi-continuous record of the magnetic field direction and intensity. The detail of such a record depends directly upon the rate of sedimentation. Low rates ( $0.1 \text{ cm}/10^3 \text{ yr}$ ) will average out most of the secular variation in a standard sample 2.54 cm thick, whereas samples with high sedimentation rates ( $10 \text{ cm}/10^3 \text{ yr}$ ) can record the details of variations over several hundred years.

Successful studies of the sediment paleomagnetic record are possible even though there is relatively little known

about the mechanisms by which the magnetic pattern is recorded. The slow deposition of sediment grains on the ocean floor (DRM), post-depositional mixing such as caused by bioturbation (PDRM), and chemical alteration (CRM) may all play a role. Measurements on many cores of differing lithologies show that some offer coherent patterns in agreement with results from land basalts and sea-floor spreading anomalies, while others are scattered with unrecognizable patterns. We do not understand why this happens. Our confidence in the reliability of any particular sediment record is based more on macroscopic properties, such as the statistics of the scatter in inclination and declination directions, than it is on any fundamental paleomagnetic properties of the minerals which carry the signal. This poses a significant problem for the analysis of DSDP cores. Because of probable rotations during the drilling process, declination values are always highly scattered, and reversal chronologies can only be determined by changes in sign of the inclination which may be unreliable for low latitudes. Results often have not been easy to interpret and attempts at utilizing these cores to extend the sequence of reversals back through the Tertiary have made only slow progress.

Recently there have been attempts at using the JOIDES data for mapping changes in paleolatitude back in time [Sclater and Cox, 1970; Hammond et al., 1975; Peirce, in press]. Results can become particularly useful when averages

are taken of many closely spaced samples, with the standard deviation of the mean forming one measure of our confidence in the result [Green and Brecher, 1974]. Collections of high density sampling can be taken in several locations down core to give averages of paleolatitude as a function of age. With such a relationship we can hope to distinguish between origins north and south of the equator. A pattern of steep to shallow to steep inclinations should define an equatorial crossing even though the polarity of the inclination averages alone are not significant. Such a pattern has been observed in equatorial Pacific piston cores by Hammond et al. [1974]. We only must assume that the plate motion is continuous and that abrupt reversals in its direction do not occur. The two fundamental limitations in our basalt analyses (i.e. the large uncertainties produced by secular variation and the non-uniqueness of the polarity) can be overcome by detailed analysis of sediment cores.

#### Site 292

Drill cores from site 292 contain the most complete sediment section to basement of any of the JOIDES holes in the West Philippine Basin. The lithology of this sediment is a uniform calcareous ooze/ chalk rich in nannofossils, indicating that the bottom (now at 2943 m) has remained above the carbonate compensation depth (CCD) throughout its history [Ingle et al., 1975]. Sedimentation rates as determined by these fossils are fairly uniform at approximately  $1.0 \text{ cm}/10^3 \text{ yr}$ .

Levels of bulk magnetization ( $0.3$  to  $4.0 \cdot 10^{-4}$  gauss-cm<sup>3</sup>) are relatively high due to the large ash content, which probably owes its origins to nearby island arc volcanoes. All of these factors plus the undisturbed nature of many of the cores, makes this a most promising site for paleomagnetic study.

We selected six relatively undisturbed areas at different locations down core for detailed sampling. These positions together with their associated paleontological ages are shown in Figure 8. At each position we took a suite of 10-20 samples of approximately 10 cm<sup>3</sup> each. If the sediment was sufficiently indurated, as it was in cores 31-33, 34-38 and 39, we could sample by diamond drilling, just as for the basalts; if not, we took an oriented sample by pressing a non-magnetic plastic tube (approximately 1 in. by 1 in.) into the core by means of a small piston apparatus. The ends of the tube were sealed by epoxy to avoid later disturbances. A total of 94 samples were taken by these techniques and later studied with the use of C.E. Helsley's cryogenic magnetometer and 400 Hz demagnetizer at the University of Texas at Dallas.

We selected 5 of our samples for detailed AF demagnetization experiments, results of which can be seen in figures 9 and 10. The PSI stability index indicates moderate to high stability at a demagnetization field of 88 oe. Median destructive fields (MDF) range from 135 to 310 oe. On the



FIGURE 8

Lithological and paleontological stratigraphy for DSDP site 292 sediment from Ingle et al. (1975). Shaded areas locate positions of recovered sediment. Numbers identify the location and age of cores which were densely sampled. Paleomagnetic measurements were made on 10-20, 10 cm<sup>3</sup> samples as closely spaced as possible in each of these 6 locations, for a total of 93 individual samples.

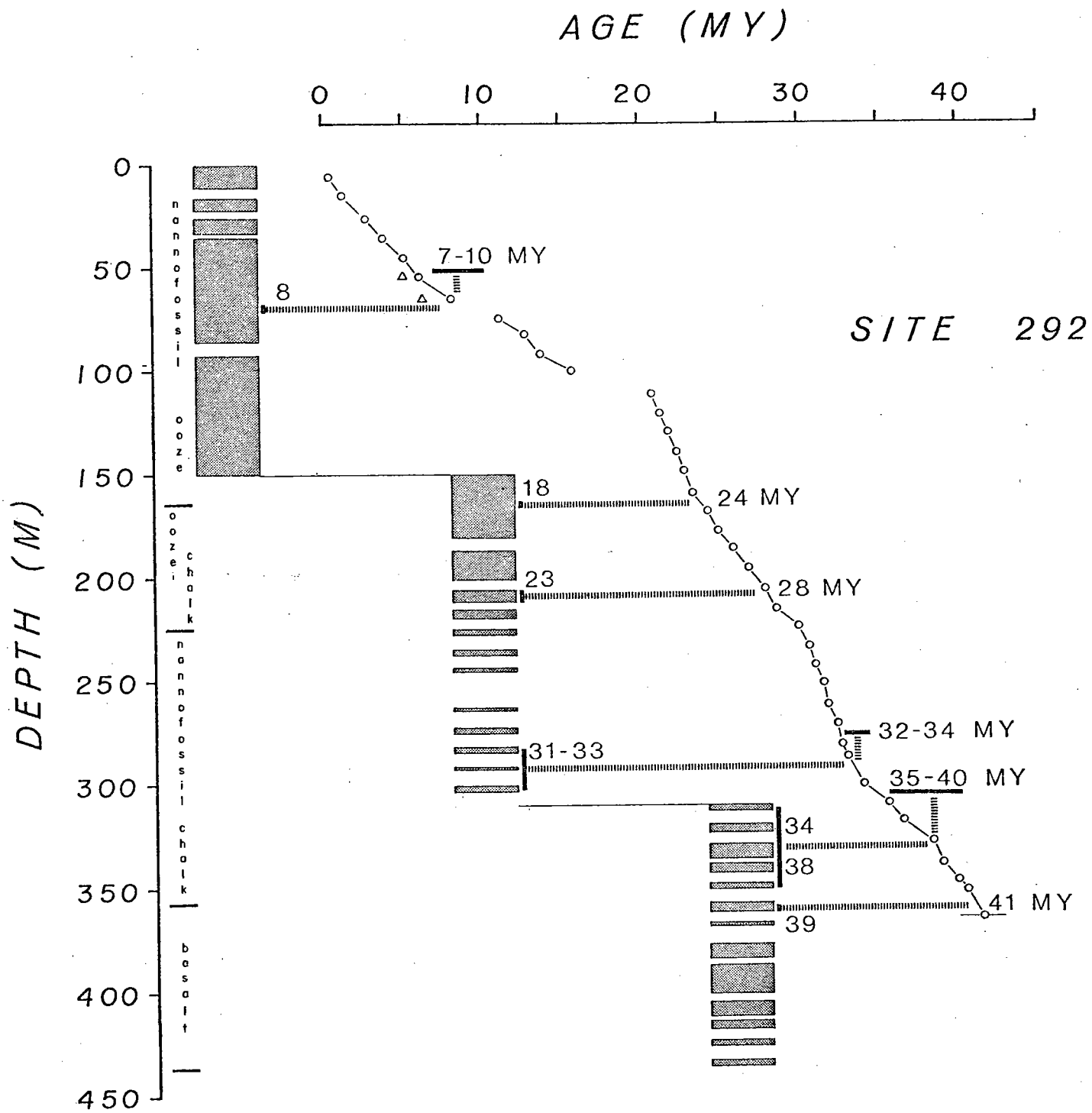


FIG 8

FIGURE 9

Demagnetization curves and stereonet plots for selected site 292 sediment intensities and directions. Higher MDF values and well grouped directions indicate higher stability than for the site 292 basalts of Figures 2-4.

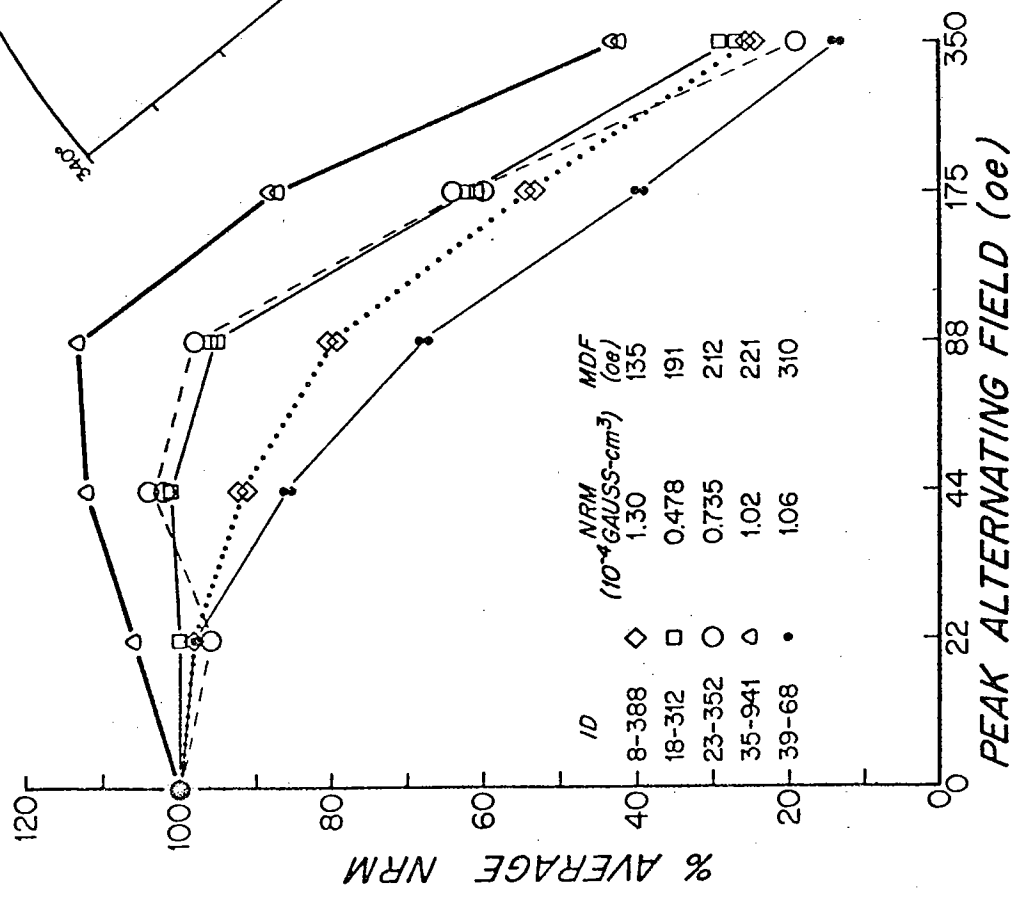
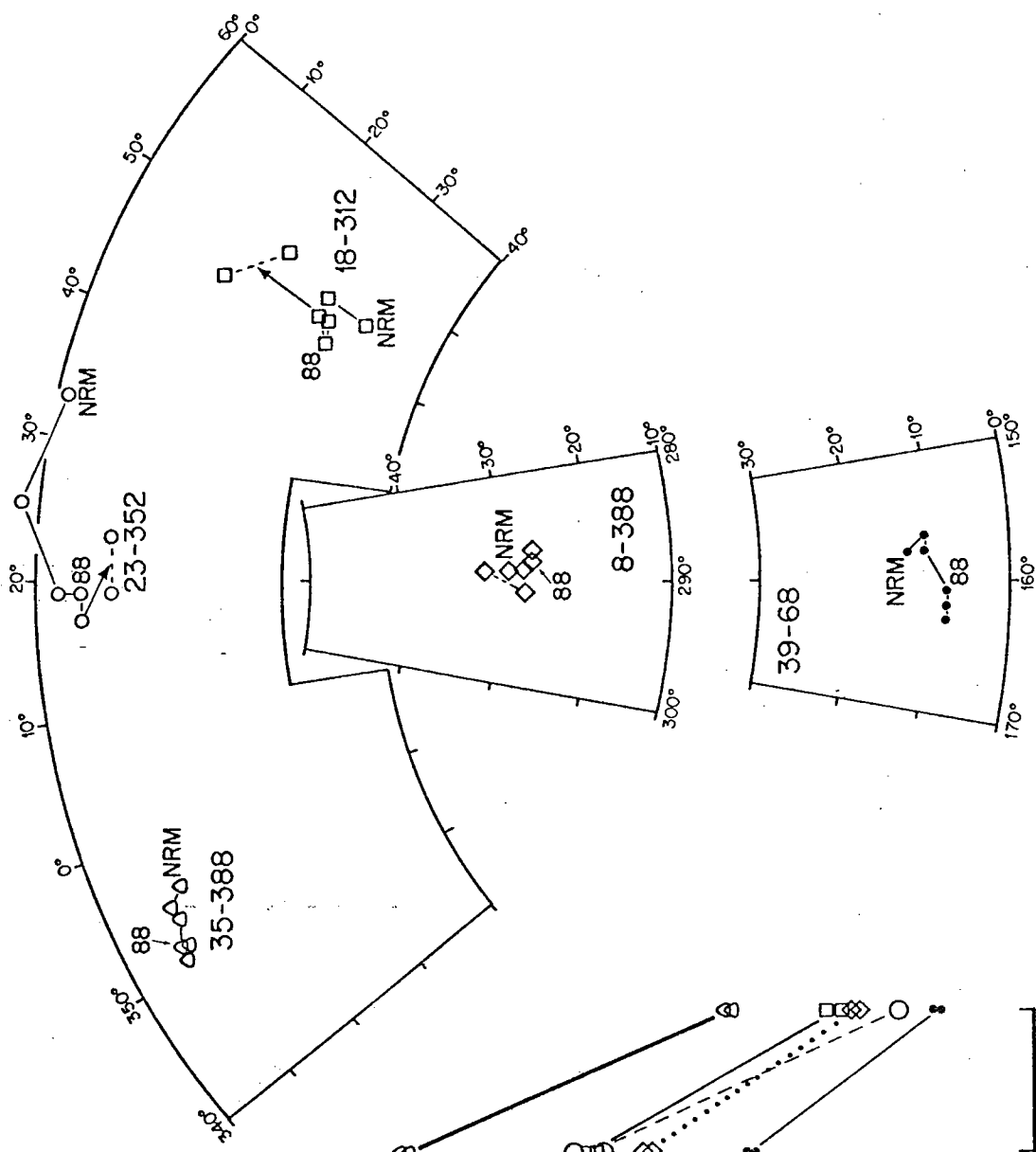


FIG 9

FIGURE 10

PSI stability index as a function of demagnetizing field for the same selected site 292 sediments as in Figure 9. Low values at 88-175 oe indicate moderate to high stability.

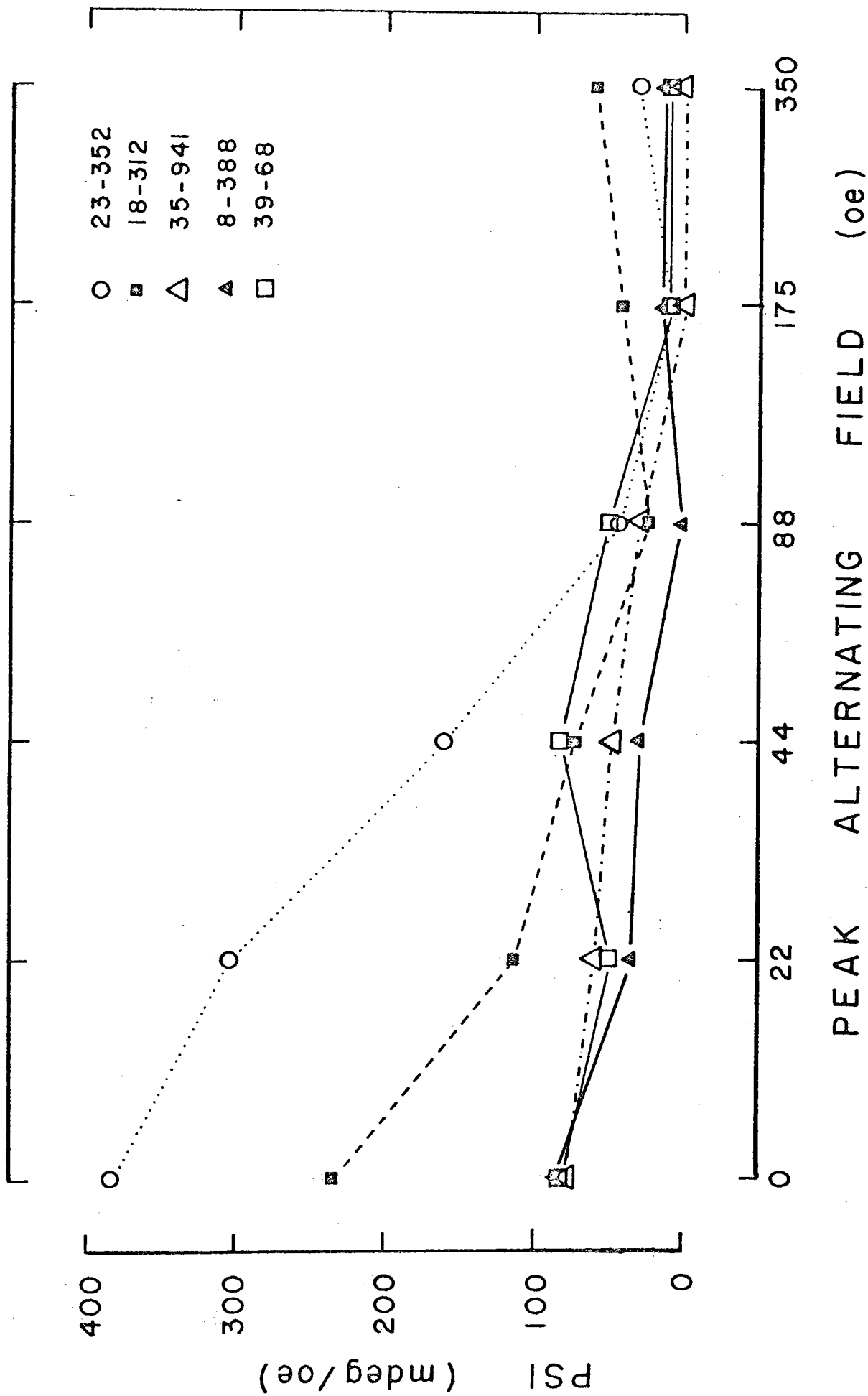


FIG 10

basis of these results we demagnetized all samples at 88 oe (see Table III for a listing of all our paleomagnetic measurements). Figure 11 is a graph that shows all the demagnetized inclinations in 5 of our 6 groups of samples. Values for cores 34-38 were omitted because they span a depth range much greater than do the others. Histograms of inclinations in each group show a marked decrease from core 8 to 33, followed by an increase from 33 to 39. This is precisely the pattern that we hoped might be associated with an equatorial crossing.

A more statistical analysis can be accomplished by determining means and their deviations for each group of samples. So as not to bias our averages we first want to discard any results which are either very different from the others or represent more unstable behavior. Unfortunately, we cannot calculate as detailed a stability index as we could for the basalts, since we only know two magnetization vectors (NRM and 88 oe). We do notice that many samples with extreme inclinations also have magnetization directions which change greatly under demagnetization from NRM to 88 oe, whereas most samples show little change. We quantify this by calculating the angular change between magnetizations at NRM and 88 oe. This number,  $\Delta$ , is not a strictly valid criterion of stability although in general higher values of  $\Delta$  correspond to higher values of PSI. We utilize it only as a means of limiting our results so that they all represent similar magnetic behavior. A histogram of  $\Delta$  values (Fig.

TABLE III: Remanent Magnetic Properties for DSDP Site 292 Sediments

292-8  
N=13

Depth below sea floor: 66.67 m Age: 7-10 mybp

DEPTH (CM)	M ( $10^{-40}$ GAUSS-CM <sup>3</sup> )	M <sub>88</sub>	I (DEG)	I <sub>88</sub> (DEG)	D <sub>0</sub> (DEG)	D <sub>88</sub> (DEG)	Δ (DEG)	λ <sub>88</sub> (DEG)
0	0.518	0.486	22.3	21.1	25.4	23.5	2	10.9
53	1.06	0.821	21.0	18.8	269.7	280.4	10	9.7
152	0.652	0.607	-9.7	-8.6	352.8	354.0	2	-4.3
170	1.04	0.735	19.5	19.0	186.1	185.8	1	9.8
187	0.629	0.531	28.7	30.4	90.1	83.7	6	16.4
226	1.20	0.978	30.1	31.7	119.5	117.0	3	17.2
251	1.43	1.18	17.0	17.1	290.7	297.7	7	8.7
273	0.996	0.768	1.8	0.7	227.8	232.2	5	0.4
291	0.332	0.244	-26.9	(-36.5)	209.3	254.3	39	(-20.3)
328	0.469	0.348	11.4	9.8	245.7	255.5	10	4.9
388	1.30	1.04	27.0	26.4	109.0	106.7	2	13.9
424	1.37	1.13	17.9	20.4	78.2	77.6	3	10.5
470	1.29	1.10	9.5	9.8	335.9	338.5	3	4.9
496	1.05	0.820	21.7	23.0	116.8	115.9	2	<u>12.0</u>
							mean	8.8 (6.8)
							σ	5.9 (9.4)
							σ/√N	1.6 (2.5)



TABLE III continued:

292-18  
N=14

Depth below sea floor: 163.35 m Age: 24 mybp

DEPTH (CM)	$M_{40}$ ( $10^{-4}$ )	$M_{88}$ GAUSS-CM <sup>3</sup> )	$I_{40}$ (DEG)	$I_{88}$ (DEG)	$D_{40}$ (DEG)	$D_{88}$ (DEG)	$\Delta$ (DEG)	$\lambda_{88}$ (DEG)
0	1.00	0.876	14.8	11.7	128.0	133.1	6	5.9
30	1.55	1.14	16.2	13.9	157.4	161.8	5	7.1
58	0.842	0.789	9.8	4.2	213.1	213.2	6	2.1
109	0.564	0.427	-9.6	-8.6	199.1	191.1	8	-4.3
138	0.265	0.225	4.8	-2.3	161.4	135.9	26	-1.2
176	0.102	0.104	0.5	(-2.0)	327.4	212.9	115	(-1.0)
191	0.244	0.139	48.6	(67.8)	210.6	166.1	29	(50.8)
209	0.078	0.124	20.1	(-7.5)	217.4	240.3	36	(-3.8)
222	0.192	0.119	-31.7	(-53.4)	169.2	110.9	47	(-34.0)
241	0.313	0.261	-21.3	-18.8	175.9	164.0	12	-9.7
256	1.30	1.14	-15.5	-13.4	172.7	172.9	2	-6.8
288	0.657	0.715	-11.9	-16.6	213.2	206.9	8	-8.5
312	0.480	0.461	-30.4	-27.7	225.6	222.6	4	-14.7
342	0.612	0.644	-18.8	-26.9	263.9	245.1	19	<u>-14.2</u>
							mean	-7.4 (-4.0)
							$\sigma$	4.3 (17.0)
							$\sigma/\sqrt{N}$	1.4 (4.0)

TABLE III continued:

292-23

N=14

Depth below sea floor: 206.03 m Age: 28 mybp

DEPTH (CM)	$M_0$ ( $10^{-4}$ GAUSS-CM <sup>3</sup> )	$M_{88}$	$I_0$ (DEG)	$I_{88}$ (DEG)	$D_0$ (DEG)	$D_{88}$ (DEG)	$\Delta$ (DEG)	$\lambda_{88}$ (DEG)
0	0.832	0.513	4.7	-8.8	340.4	324.4	21	-4.4
6	0.627	0.701	-16.4	-19.9	297.6	313.9	16	-10.3
18	0.638	0.500	3.7	-13.7	307.3	277.7	34	-7.0
31	0.893	0.884	-5.1	0.0	79.2	69.3	11	0.0
95	1.70	1.51	0.1	-1.5	186.3	184.3	3	-0.8
115	0.621	0.636	-18.7	-21.7	55.9	77.7	21	-11.3
125	1.30	1.16	-8.7	-13.1	147.3	138.7	10	-6.6
219	0.552	0.654	-8.2	4.4	72.8	60.5	18	2.2
261	1.07	1.24	-4.7	-6.6	349.5	347.0	3	-3.3
282	0.762	0.735	-19.3	(-27.4)	116.5	95.3	21	-14.5
352	0.733	0.722	0.8	-4.2	212.0	198.6	14	-2.1
426	0.749	0.727	3.1	2.2	290.9	295.3	5	1.1
471	1.21	1.04	-25.8	(-28.5)	172.9	173.3	3	-15.2
539	0.422	0.379	22.1	4.2	61.3	81.1	26	<u>2.1</u>
								mean -5.0
								$\sigma$ 5.7
								$\sigma/\sqrt{N}$ 1.5

TABLE III continued:

292-31 to 33

N=19

Depth below sea floor: 283.69 m . Age: 32-34 mybp

DEPTH (CM)	$M_0$ ( $10^{-4}$ )	$M_{883}$ GAUSS- $CM^3$ )	$I_0$ (DEG)	$I_{88}$ (DEG)	$D_0$ (DEG)	$D_{88}$ (DEG)	$\Delta$ (DEG)	$\lambda_{88}$ (DEG)
0	0.683	0.738	7.1	(-21.6)	322.4	321.6	29	(-11.2)
70	0.850	0.819	-8.0	-7.7	263.7	265.1	1	-3.9
154	0.334	0.347	-30.3	(-34.8)	33.4	35.5	5	(-19.2)
241	1.27		11.1		52.7			5.6*
283	1.63	1.57	-5.4	-5.3	4.7	3.4	1	-2.7
288	1.57	1.50	-9.2	-9.0	14.2	12.7	2	-4.5
292	0.875	0.854	-2.2	-4.8	166.1	164.8	3	-2.4
297	0.905	0.877	-6.8	-6.7	160.0	160.4	0	-3.4
302	0.783	0.770	-4.9	-5.5	159.8	158.8	1	-2.8
311	0.714	0.701	-2.3	-5.4	170.8	169.2	4	-2.7
317	0.836	0.832	-12.8	-12.4	261.5	260.8	1	-6.3
323	0.695	0.686	-4.9	-3.4	269.9	269.3	2	-1.7
328	0.559	0.537	-0.8	-1.0	37.8	36.3	2	-0.5
334	0.455	0.445	-8.4	-8.3	40.3	41.3	1	-4.2
343	1.14	1.17	-1.8	-4.3	129.9	131.2	3	-2.2
394	0.197	1.00	23.8	(3.6)	46.7	81.0	39	(1.8)
398	0.509	0.401	4.6	7.4	43.6	47.3	5	3.7
404	3.02	2.88	-5.5	-4.2	280.0	276.0	4	-2.1
408	3.98	3.79	-12.9	-10.6	280.5	280.4	2	-5.4
419	1.35	1.25	-0.6	-0.9	271.7	270.1	2	-0.5
423	0.835	0.802	4.7	2.9	57.1	61.0	4	1.5
428	0.932	0.905	4.0	5.6	51.9	52.3	2	2.8

mean-1.7 (-)

 $\sigma$  3.0 ( $\sigma/\sqrt{N}$  0.7 (

TABLE III continued:

292-34 to 38

N=9

Depth below sea floor: 312.17 m Age: 35-40 mybp

DEPTH (CM)	$M_0$ ( $10^{-4}$ GAUSS-CM <sup>3</sup> )	$M_{88}$	$I_0$ (DEG)	$I_{88}$ (DEG)	$D_0$ (DEG)	$D_{88}$ (DEG)	$\Delta$ (DEG)	$\lambda_{88}$ (DEG)
0	1.89	1.69	-2.6	-3.0	286.5	284.9	2	-1.5
106	1.11	1.04	2.5	2.8	97.4	96.3	1	1.4
941	1.02	1.15	-9.7	-6.4	355.6	351.8	5	-3.2
1044	0.358	0.461	-0.6	0.5	14.7	8.9	6	0.3
1929	1.18		8.2		111.7			4.1*
1964	2.39	2.41	0.4	-0.2	293.3	296.8	4	-0.1
2865	1.17	0.969	5.4	2.5	207.4	207.3	3	1.3
2918	0.956		9.1		61.1			4.6*
3794	0.442		0.0		274.3			0.0*
3897	0.0881		(29.2)		261.0			
							mean	0.8
							$\sigma$	2.3
							$\sigma/\sqrt{N}$	0.8

TABLE III continued:

292-39  
N=20

Depth below sea floor: 359.57 m Age: 41 mybp

DEPTH (CM)	$M_0$ ( $10^{-4}$ )	$M_{883}$ GAUSS-CM <sup>3</sup> )	$I_0$ (DEG)	$I_{88}$ (DEG)	$D_0$ (DEG)	$D_{88}$ (DEG)	$\Delta$ (DEG)	$\lambda_{88}$ (DEG)
0	0.767		-7.6		99.6			-3.8*
11	0.806		0.5		256.9			0.3*
24	0.554		19.0		52.9			9.8*
35	0.825	0.543	17.7	13.2	59.2	65.1	7	6.7
45	0.688	0.413	22.9	20.7	61.6	68.1	6	10.7
68	1.05	0.719	11.9	8.3	158.0	160.4	4	4.2
83	0.675	0.427	17.7	12.5	162.7	168.3	8	6.3
101	1.25	0.801	14.6	9.2	98.2	101.8	6	4.6
117	1.15	0.778	24.5	19.9	94.3	93.6	5	10.3
130	1.26	0.843	11.0	8.4	106.1	100.6	6	4.2
147	0.645	0.361	0.0	-8.2	18.6	19.2	8	-4.1
154	0.478	0.346	1.8	-7.1	30.6	33.1	9	-3.6
169	1.24	0.842	8.2	6.6	290.8	286.0	5	3.3
185	1.13	0.799	12.2	9.6	17.2	21.0	5	4.8
205	1.59	1.14	11.4	14.0	288.6	292.6	5	7.1
211	1.46	1.11	11.1	13.0	292.9	292.9	2	6.6
228	1.05	0.803	15.6	14.6	0.0	3.4	3	7.4
236	0.653	0.481	22.0	20.5	355.9	8.3	12	10.6
258	0.665	0.447	15.4	15.3	327.3	343.2	15	7.8
266	0.732	0.417	8.3	9.0	339.1	346.5	7	4.5
							mean	4.9
							$\sigma$	4.5
							$\sigma/\sqrt{N}$	1.0

The depth is the distance from the top of the core. Each sample is approximately 10-20 cm<sup>3</sup>. Subscripts refer to fields of peak AC demagnetization.  $\Delta$  is a measure of the reliability as defined in the text.  $\lambda_{88}$  is the paleolatitude after 88 oe demagnetization. Values in parentheses are only included in bracketed values for the mean and standard deviation.

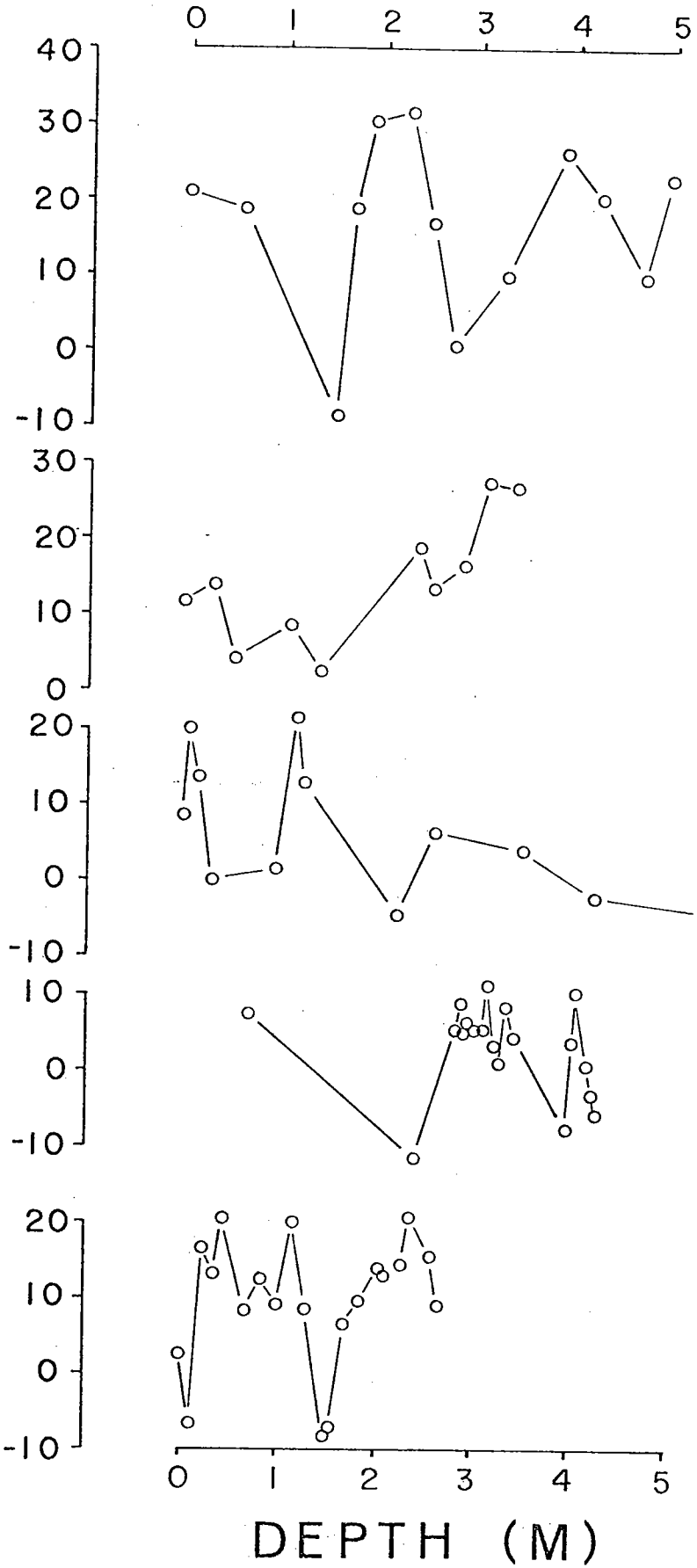
\* NRM value

## FIGURE 11

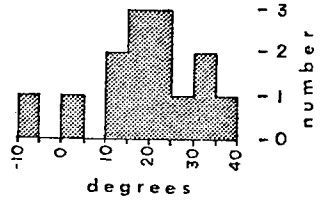
Paleoinclinations after demagnetization in 88  
oe for site 292 sediments. Large numbers identify  
the sampling location (see Figure 8) for each  
group of closely spaced samples. The depth is  
relative to the position of the top sample in  
each group. Histograms show the decrease in  
inclination from cores 8 to 33 and increase from  
33 to 39 which we attribute to a crossing of the  
paleomagnetic equator. Inclinations from cores  
34-38 are not included due to the large depth  
range which they span.

# SITE 292

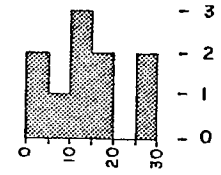
INCLINATION (DEG)



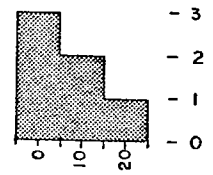
8



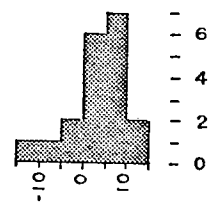
18



23



31-33



39

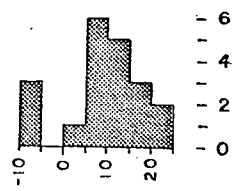


FIG 11

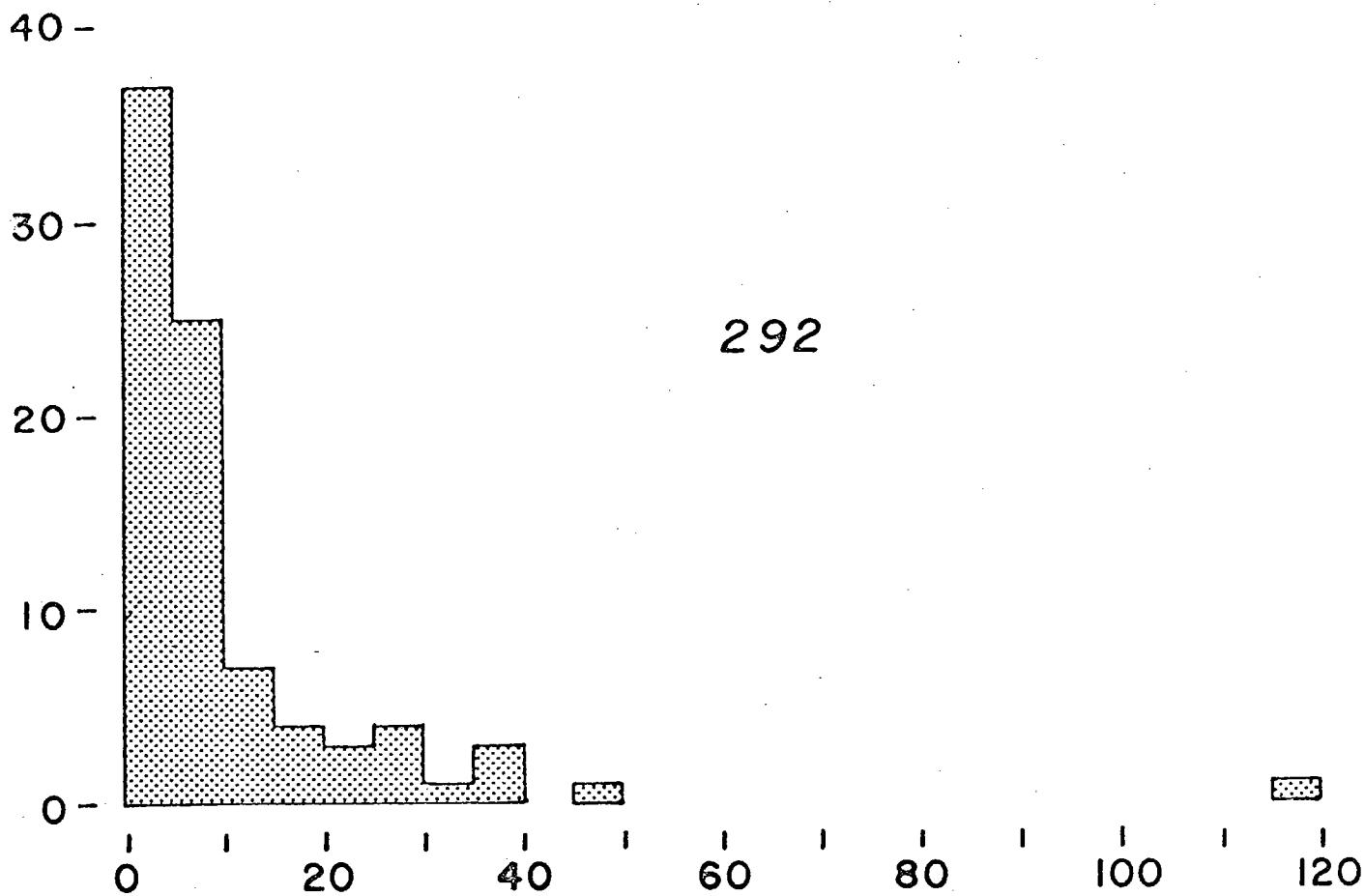
l2) shows that most are small ( $<20^\circ$ ), although from table III we can see that samples from different cores show different behavior. Some, like those in core 8, have very low values except for the few that also have extreme values of inclination; others, like those from core 23, have all rather high values. In general we have discarded data from samples with  $\Delta > 25$ . In only one case (sample 31-154) have we discarded an extreme inclination that also had low  $\Delta$ . Mean inclinations of these selected samples are not greatly different from those which include all our samples (Table III), but standard deviations are in some cases greatly reduced. In all but one case we use the actual sign of the inclinations, and assume that the values in each group are not affected by reversals of the earth's field. By so doing we produce distribution patterns (Figure 11) more similar to Gaussian than if we took absolute values and assumed that all changes in sign were caused by reversals. The only exception to this is in core 18, where a sequence of three positive inclinations shifts to negative values. This we attribute to a reversal.

Mean paleolatitudes and their 95% confidence intervals are plotted in figure 13 as a function of age. Here we can clearly see the northward motion and equatorial crossing of site 292. The two primary difficulties with our basalt results are not present in this analysis of sediments. First, the ability of the sediments to average out secular variations in paleopole positions allows us in the best



FIGURE 12

Histograms of  $\Delta$  (defined in the text) for sediments of site 292, 290 and 294. Higher values for sites 290 and 294 may indicate greater instability in remanent magnetization than for site 292. This is also evidenced by the larger standard errors in mean paleolatitude determinations for sites 290 and 294 (Figure 15) than for site 292 (Figure 13).



NUMBER

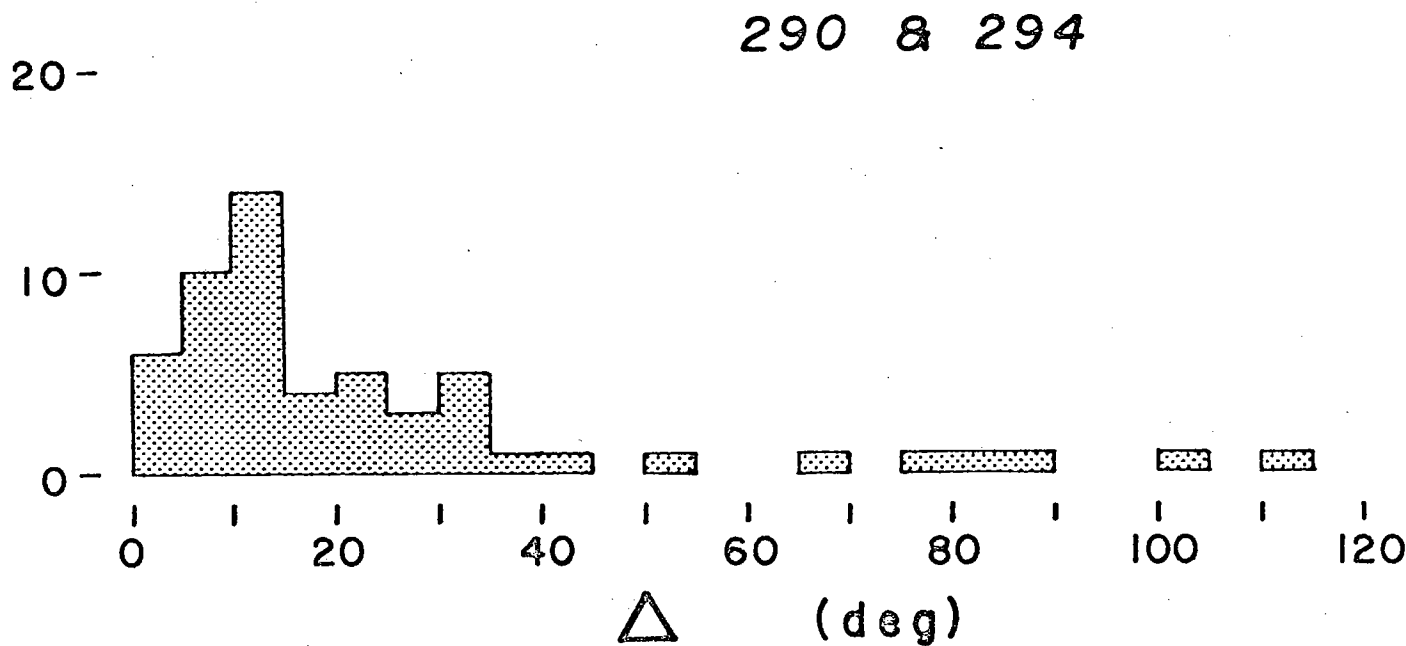


FIG 12

## FIGURE 13

Age vs. mean paleolatitude for site 292 sediment (circles) and basalt (triangles). Error bars represent  $\pm 2$  standard errors and include deviations due to theoretical secular variation. Each mean paleolatitude could be plotted on either side of the equator as indicated by the two locations for the basalt and the oldest sediment (core 39), but a southern origin is suggested if plate motion is to be uniform. Line A is a visual best fit constrained to go through the present latitude of site 292 (solid circle), while line B fits only the 5 oldest sediment values. The large range in basalt paleolatitudes is due to its insufficient sampling of secular variation, and shows that in the absence of multiple basalt flows the sediment data can be much more conclusive.

# SITE 292

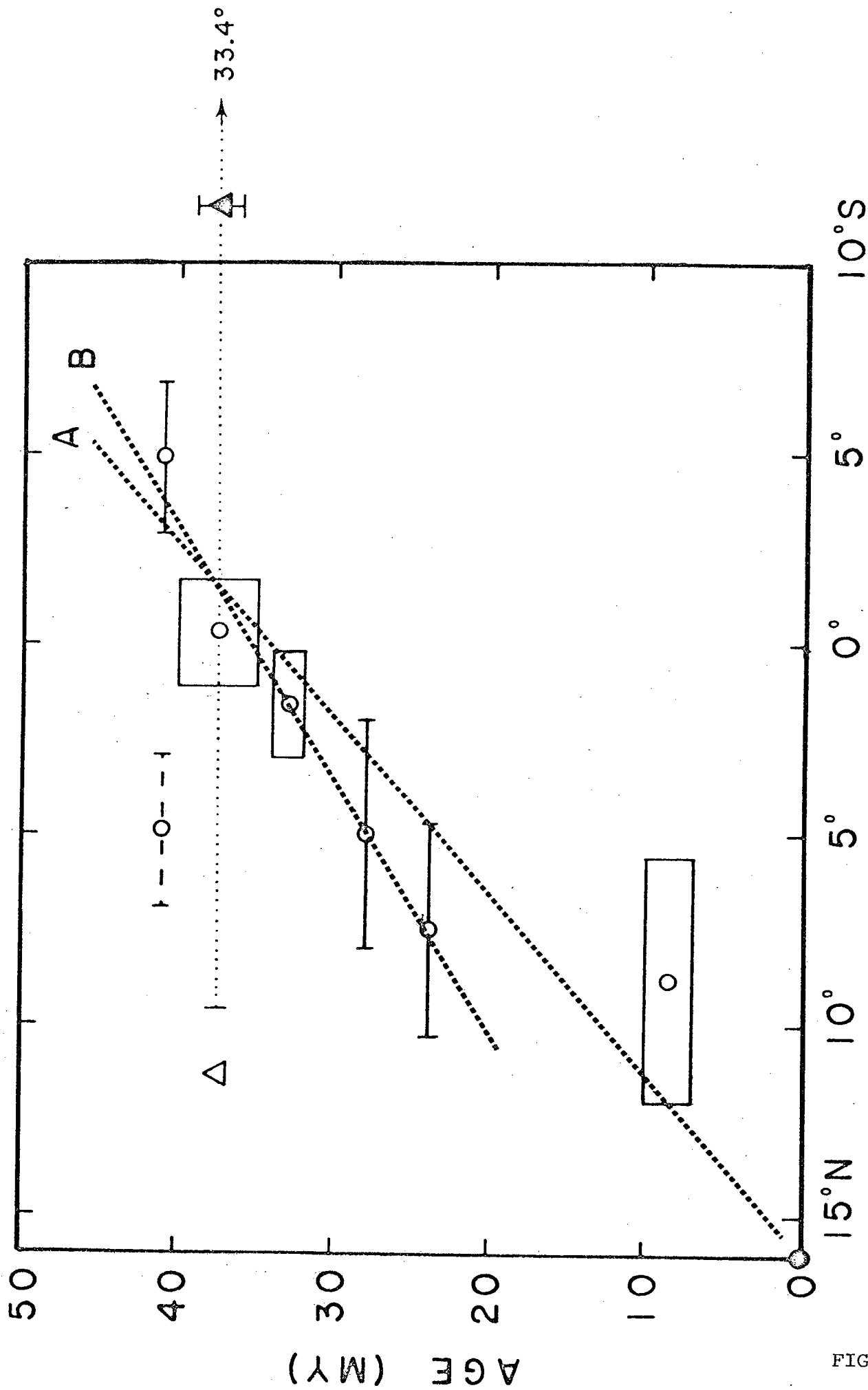


FIG 13

cases a 95% confidence interval of less than 5°. Second, we deal with the problem of non-uniqueness in paleolatitude sign by considering the trend as a function of age. Only if rather discontinuous plate motions occurred could the northern latitude location for core 39 be possible.

Of course, when we calculate paleolatitudes from our paleoinclinations, we use the equation for a geocentric dipole,

$$\tan I = 2 \tan \phi \quad (5)$$

where  $I$  is the inclination and  $\phi$  is the latitude. If the dipole is not geocentric, but has been shifted northwards in the past, as is suggested by several authors [Wilson, 1970; Wilson and McElhinny, 1974], then that shift would produce an apparent northward motion for a fixed site when analyzed with the geocentric equation. We should really use the equation for an offset dipole,

$$\tan I = \frac{2 \cos \theta + 2 r^2 \cos \theta - r \cos^2 \theta - 3 r}{\sin \theta (1 + r \cos \theta - 2 r^2)} \quad (6)$$

where  $\theta$  is the colatitude of our site and  $r$  is the northward offset in the dipole expressed as a fraction of the earth's radius. However, to produce as large enough an effect for  $\theta$  fixed at 75° as to fit Figure 13 by increasing  $r$ , we would need a maximum offset larger than 1000 km. This is several

times as large as values which have recently been suggested [Wilson and McElhinney, 1974]. Also, our purpose is to compare our paleomagnetic results with those from magnetic anomalies. Northward shifts in the dipole would affect both results and would not interfere with such a comparison.

#### Sites 290 and 294

The one major assumption we do make in order to directly compare our site 292 paleomagnetic results with those from the phase shifting of magnetic anomalies is that they are both part of the same rigid plate. They do lie near each other, but this still does not make this assumption entirely obvious. Site 292 is situated on an anomalous and unexplained feature, the Benham Rise, while the magnetic anomalies are all located in the deep water regions of the basin. We must therefore show that our paleomagnetic analysis of site 292 is not unique to it, but is also valid for other sites in the West Philippine Basin.

Sites 290 and 294 [Figure 1] offer the only other relatively undisturbed sedimentary sections in the West Philippine Basin, but these do not consist of as promising a material as does site 292. Only a few cores were recovered at each site, sedimentation rates are low ( $0.2-0.3 \text{ cm}/10^3\text{yr}$ ), and the sediment contains few micro-fossils. The lithologies are variable and quite different from site 292. Site 290 consists of a zeolite rich clay/ash in core 2 and a nannofossil rich volcanic ash in core 7; site 294 contains a brown clay

in core 2 and a clay with palagonite ash in core 4. Most of these differences are due primarily to the greater water depths (6063 m at site 290 and 5784 m at site 294) which places the sedimentary regime well below the CCD [Ingle et al., 1975].

We collected a total of 58 samples from two cores in each of sites 290 and 294. A pilot study of 8 of these shows [Figure 14] that they are less stable than are the site 292 samples, and that 175 oe may have been a better choice for a standard field of demagnetization than the 88 oe which was our value for all non-pilot samples. Values for  $\Delta$  (Table IV) are higher than for site 292, and show a different distribution in the histogram of figure 12. It is therefore not surprising that we find the standard deviations of mean paleolatitudes from these sites to be much larger than those for site 292.

Figure 15 shows mean paleolatitudes for sites 290 and 294 as a function of time. The paleolatitudes themselves are not very definitive owing to their large intervals of 95% confidence, but we can still use them as a test for values predicted by the motion of site 292. These predictions are based on the two lines A and B, which are determined in figure 13. Line A is a visual best fit that is constrained to go through the present latitude of site 292, while line B fits only the oldest 5 paleolatitudes. These same lines, plotted in figure 15 to go through the present latitudes of

TABLE IV: Remanent Magnetic Properties for DSDP Sites 290 and 294 Sediments

290-2  
N=14

Depth below sea floor: 75.05 m Age: 31-33 mybp

DEPTH (CM)	$M_0$ ( $10^{-4}$ GAUSS-CM <sup>3</sup> )	$M_{88}$	$I_0$ (DEG)	$I_{88}$ (DEG)	$D_0$ (DEG)	$D_{88}$ (DEG)	$\Delta$ (DEG)	$\lambda_{88}$ (DEG)
0	1.70	1.72	-6.8	-7.1	61.5	92.9	31	-3.6
21	0.865	0.643	16.6	-0.2	139.8	106.1	37	-0.1
41	1.31	0.930	-23.4	-25.8	240.6	256.0	14	-13.6
61	1.30	0.832	-5.7	-10.1	253.0	258.0	7	-5.1
81	1.85	1.32	15.2	13.1	240.9	252.5	11	6.6
101	1.10	0.664	1.0	-4.4	237.5	260.4	24	-2.2
121	2.38	1.91	-30.5	-31.9	239.2	252.2	11	-17.3
141	1.23	0.852	-16.9	-23.5	236.3	248.8	13	-12.3
151	0.288	0.370	39.8	28.6	155.2	107.5	41	15.3
174	0.332	0.314	-11.1	(-53.8)	177.5	109.3	68	(-34.3)
191	0.791	0.907	-6.8	-15.2	133.9	101.7	33	-7.7
211	0.688	0.780	-37.8	-34.6	143.2	102.0	33	-19.0
229	1.44	1.26	7.1	8.9	273.3	288.7	15	4.5
251	1.97	1.54	40.0	34.8	256.7	273.6	14	19.2
271	0.745	0.370	-39.9	-40.7	152.6	114.4	29	-23.3
							mean	-4.2 (-5.4)
							$\sigma$	12.1 (14.1)
							$\sigma/\sqrt{N}$	3.2 (3.8)



TABLE IV continued:

290-7  
N=15

Depth below sea floor: 205.07 m Age: 45-46 mybp

DEPTH (CM)	$M_0$ ( $10^{-40}$ )	$M_{883}$ GAUSS-CM <sup>3</sup> )	I (DEG)	$I_{88}$ (DEG)	D (DEG)	$D_{88}$ (DEG)	$\Delta$ (DEG)	$\lambda_{88}$ (DEG)
0	0.359	0.523	40.3	37.6	274.4	206.8	51	21.1
15	0.636	0.613	-15.0	-8.9	268.6	248.6	21	-4.5
25	1.25	0.852	3.3	4.6	22.4	41.1	19	2.3
50	3.78	1.31	7.6	8.2	161.3	151.6	10	4.1
98	0.909	1.08	79.3	61.1	260.8	192.3	27	42.2
115	2.23	2.10	21.7	32.4	314.5	317.0	11	17.6
129	2.40	1.92	23.5	34.4	342.1	346.2	12	18.9
156	2.15	1.66	9.1	11.3	6.1	6.4	2	5.7
285	2.71	2.61	60.4	69.2	344.9	343.6	9	52.8
333	3.20	2.87	44.4	51.7	284.8	274.5	10	32.3
353	2.70	2.49	65.3	71.3	292.2	276.1	8	55.9
371	2.92	2.65	61.5	68.0	321.7	302.8	10	51.1
385	2.54	2.57	64.4	63.6	261.3	239.2	10	45.2
408	2.45	2.50	63.9	58.9	119.9	139.8	11	39.7
423	1.80	1.94	20.1	29.2	77.4	102.2	24	15.6
							mean	26.7
							$\sigma$	19.5
							$\sigma/\sqrt{N}$	5.0

TABLE IV continued:

294-2  
N=11

Depth below sea floor: 47.39 m Age: 24 mybpt

DEPTH (CM)	$M_0$ ( $10^{-4}$ GAUSS-CM <sup>3</sup> )	$M_{88}$ (GAUSS-CM <sup>3</sup> )	$I_0$ (DEG)	$I_{88}$ (DEG)	$D_0$ (DEG)	$D_{88}$ (DEG)	$\Delta$ (DEG)	$\lambda_{88}$ (DEG)
0	1.34	1.01	60.4	(72.5)	149.2	107.9	20	(57.8)
7	1.59	1.20	45.7	43.0	74.7	73.4	3	25.0
17	0.575	0.202	49.2	46.1	171.6	336.5	84	27.5
26	0.491	0.169	-14.3	(-77.4)	151.4	339.3	88	(-65.9)
35	0.360	0.177	12.3	4.8	213.9	224.7	22	2.4
40	0.717	0.616	-32.6	-39.0	147.7	105.8	34	-22.0
46	0.544	0.643	-21.3	-26.0	108.5	80.1	26	-13.7
57	1.03	0.736	-1.6	-18.2	11.5	2.1	19	-9.3
67	1.11	0.811	14.2	10.8	112.6	95.2	17	5.5
72	0.688	0.450	39.4	53.4	116.5	51.4	45	34.0
85	0.209	0.0762	0.0	15.1	261.5	13.0	111	7.7
93	0.556	0.378	19.6	25.1	168.2	81.7	79	13.2
99	0.443	0.190	10.3	29.0	194.4	301.7	100	<u>15.5</u>
							mean	7.8 (6.0)
							$\sigma$	16.9 (29.1)
							$\sigma/\sqrt{N}$	5.1 (8.1)

TABLE IV continued:

294-4  
N=15Depth below sea floor: 99.05 m Age: 50 mybp<sup>+</sup>

DEPTH (CM)	$M_0$ ( $10^{-4}$ GAUSS-CM <sup>3</sup> )	$M_{88}$ (CM <sup>3</sup> )	$I_0$ (DEG)	$I_{88}$ (DEG)	$D_0$ (DEG)	$D_{88}$ (DEG)	$\Delta$ (DEG)	$\lambda_{88}$ (DEG)
0	1.58	1.21	2.1	-5.2	301.7	313.5	14	-2.6
5	1.63	1.06	-0.8	1.5	238.5	238.8	2	0.8
11	1.21	0.754	23.3	20.6	169.0	165.4	4	10.6
16	1.77	1.15	58.7	59.7	86.5	77.3	5	40.6
21	1.78	1.41	27.8	21.5	341.9	347.3	8	11.1
26	2.86	1.70	33.6	34.5	186.9	187.7	1	19.0
37	3.12	2.30	9.1	6.6	298.8	305.9	8	3.3
52	4.82	3.29	27.6	30.9	73.6	65.1	8	16.7
58	2.05	1.72	22.2	19.4	349.3	350.6	3	10.0
68	1.28	0.897	-2.6	-3.3	222.6	330.6	8	-1.7
78	2.39	1.66	56.2	51.8	70.4	58.7	8	32.4
88	0.733	0.475	2.0	-5.7	192.8	202.4	12	-2.9
98	0.328	0.200	47.3	48.5	128.7	80.0	32	29.5
108	0.366	0.0815	21.5	24.2	190.3	185.0	6	12.7
118	0.442	0.149	58.5	32.8	276.2	336.8	47	17.9
							mean	13.2
							$\sigma$	12.8
							$\sigma/\sqrt{N}$	3.3

The depth is the distance from the top of the core. Each sample is approximately 10-20 cm<sup>3</sup>. Subscripts refer to fields of peak AC demagnetization.  $\Delta$  is a measure of the reliability as defined in the text.  $\lambda_{88}$  is the paleolatitude after 88 oe demagnetization. Values in parentheses are only included in the bracketed means and standard deviations.

+ assuming a sedimentation rate of 2.0 m/my.

FIGURE 14

PSI vs. field of demagnetization for selected site 290 and 294 sediments. Lowest values at 175 oe indicate only moderate to low stability.

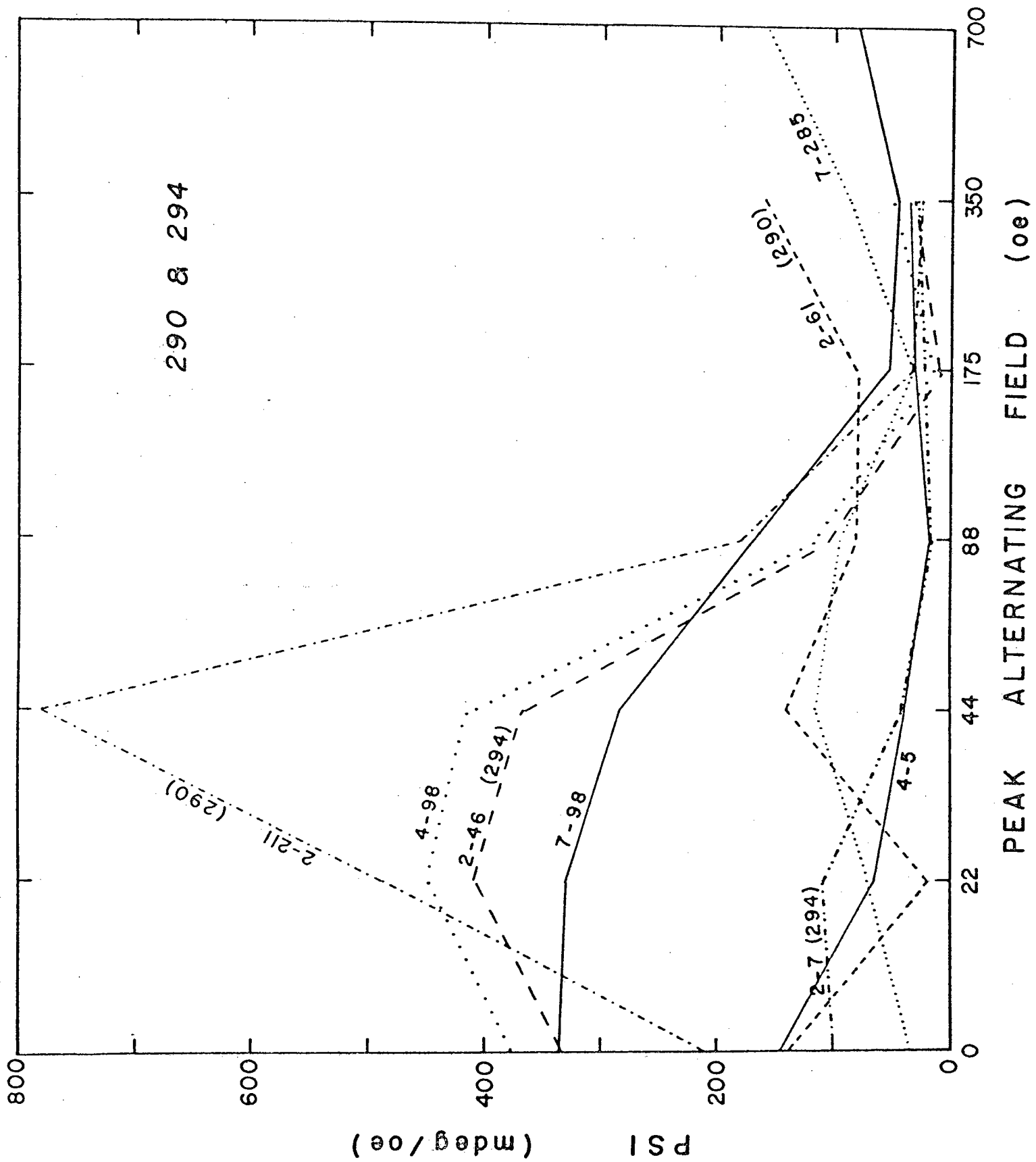
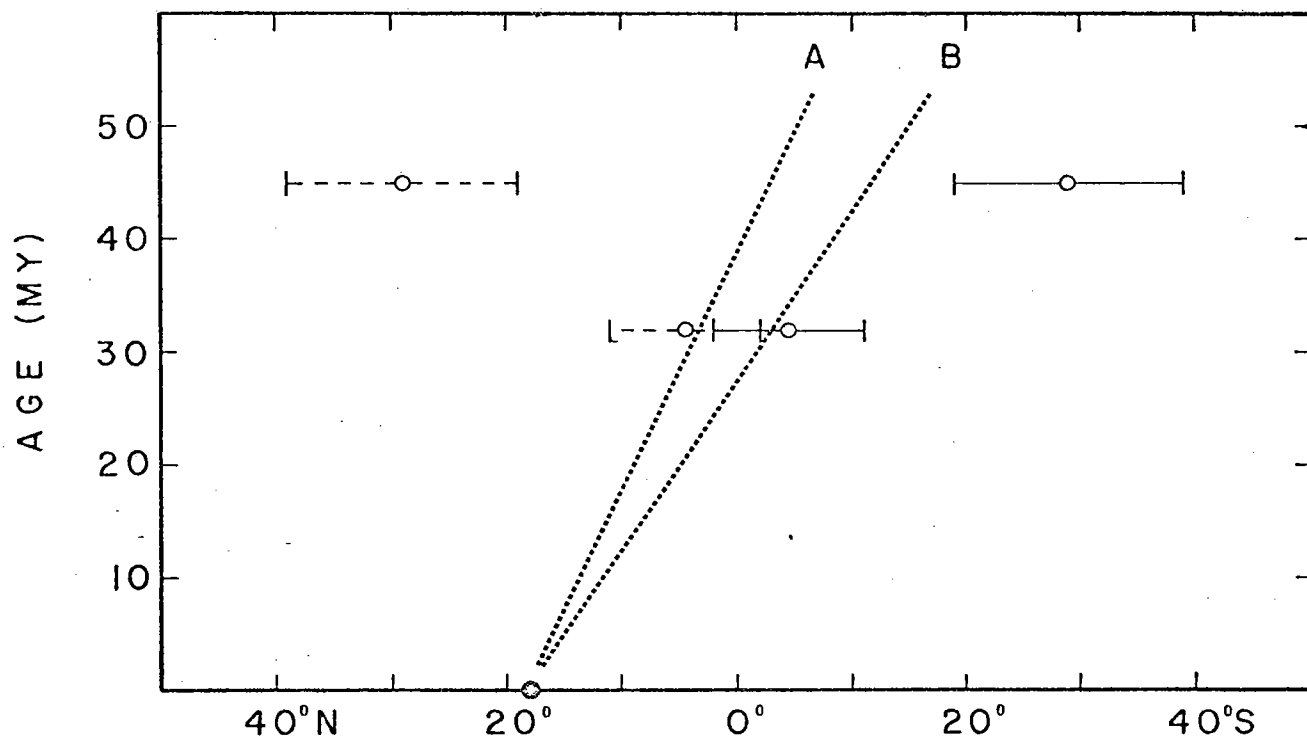


FIG 14

FIGURE 15

Paleolatitudes vs. age for sites 290 and 294. Means and  $\pm 2$  standard errors are plotted for two groups of samples at each site. Values on either side of the equator are equally possible and both are plotted for each determination. Origins south of the equator are consistent with lines A and B defined in Figure 13 for the motion of site 292. This agreement suggests that paleolatitude determinations for site 292 are not governed by the possibly anomalous tectonic setting of the Benham Rise.

SITE 290



SITE 294

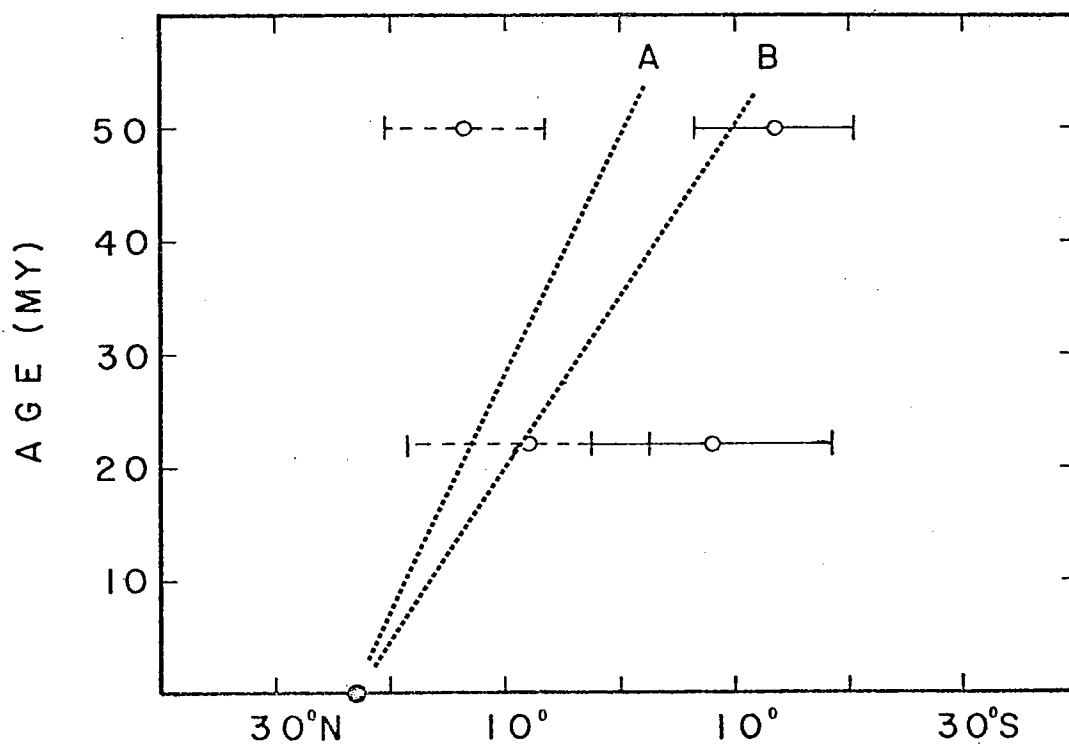


FIG 15

sites 290 and 294, also fit the older paleolatitude measurements. The one exception is the oldest paleolatitude determination for site 290, which is located further south than its prediction.

#### PALEOPOLES AND RELATIVE ROTATIONS

Paleopoles can be determined from the shape of sea-floor spreading magnetic anomalies [Larson and Chase, 1972; Schouten and Cande, 1976; Louden, 1976]. The technique is to describe the anomaly in the wave number domain,

$$M(k) = C.J(k).E(k).e^{-i\theta}, \quad (7)$$

where  $M(k)$  is the Fourier transform of the magnetic anomaly  $m(x)$ ,  $C$  is a constant,  $J(k)$  is the transform of the magnetization distribution  $j(x)$  and  $E(k)$  is the magnetic earth filter [Schouten and McCamy, 1972]. The effect of phase filter  $e^{-i\theta}$  can be removed by multiplying  $M(k)$  by  $e^{i\theta}$  for the appropriate value of  $\theta$ , and a "deskewed" (i.e. squarewave looking) pattern should result. This  $\theta$  gives the set of all possible remanent inclinations and declinations, which map as a great circle on a stereonet plot of paleo-north poles (Schouten and Cande, 1976).

We apply this technique on a short sequence of anomalies in the West Philippine Basin first described by Louden (1976). Figure 16 gives the location and identification of these anomalies. These identifications are consistent with



FIGURE 16

Map of DSDP sites, ship tracks and magnetic anomalies in the West Philippine Basin (see Figure 1 for location). Alphabetic labels identify ship track segments used in Figure 17. Numbers identify anomalies as in Heirtzler et al. (1968)

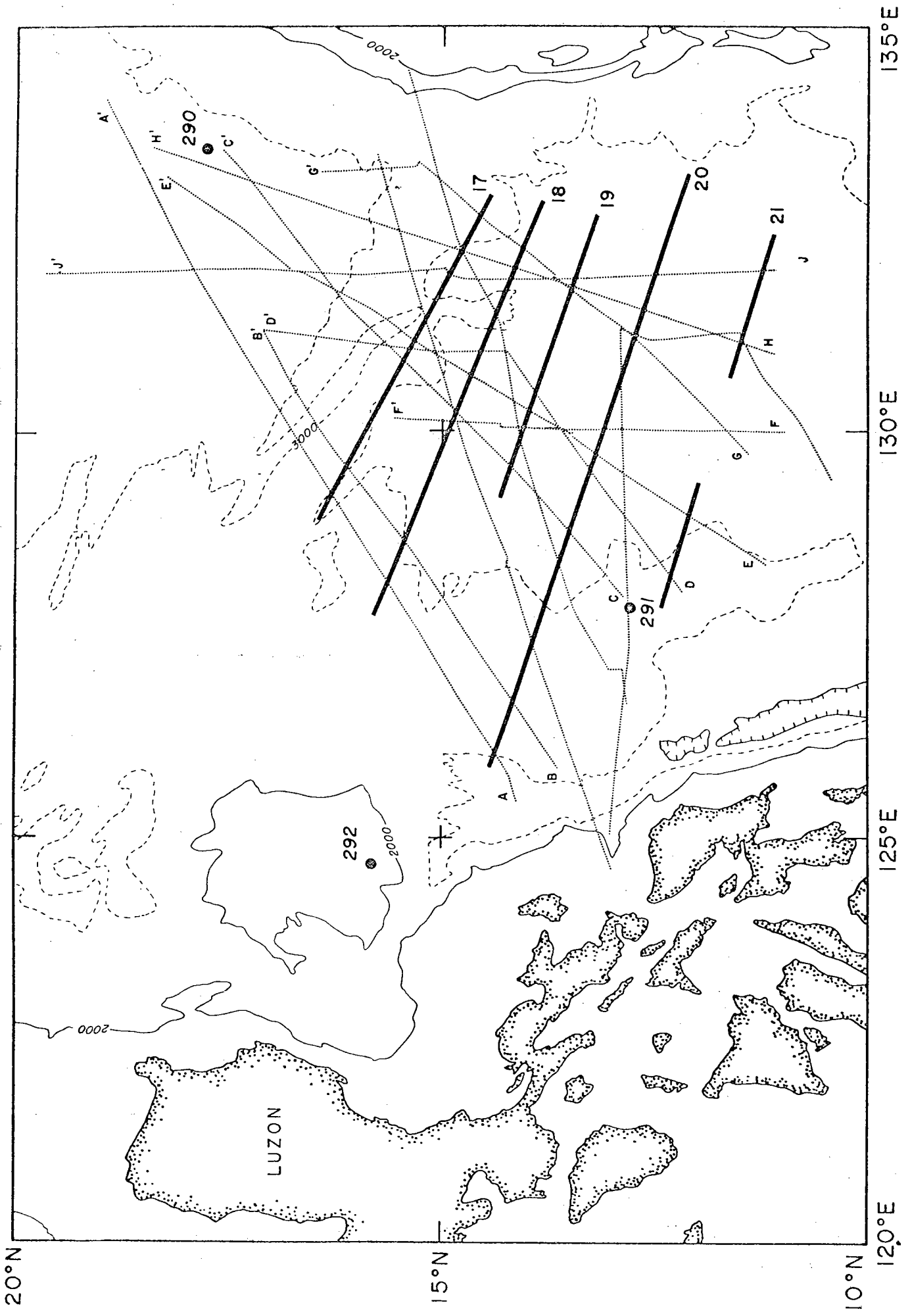


FIG 16

the paleontological dates from DSDP Site 291 [Ingle et al., 1975] and an origin south of the equator, which is supported by the paleomagnetic measurements described in this paper. The phase shifted anomalies and degree of phase shift are shown in Figure 17. These values are little changed from the previous estimates of Loudon [1976] and range from  $160^\circ$  to  $200^\circ$  for the various profiles. This range of  $\theta$  is plotted as a paleopole positions in figure 18a, and is consistent with the sediment paleolatitudes for DSDP sites 292, 294 and 290, which plot as small circles of paleopole positions in figure 18b. This means that the skewness of the magnetic anomalies is not greatly effected by tectonic effects such as can sometimes cause anomalous results [Cande, 1976], but is a true indicator of the remanent magnetic field. Unfortunately, we are still left a large uncertainty in determining a unique paleopole position.

The one remaining parameter which also depends on the remanent inclination and declination is the amplitude factor  $C$  in eq (6),

$$C = (\sin I \sin I_r) / (\sin I' \sin I'_r), \quad (8)$$

where  $\tan I'_r = \tan I_r / \sin \alpha_r$ . For the anomalies of figure 16 and 17,  $C$  can range from 0 to 0.94, depending on the value of the remanent strike ( $\alpha_r$ ). Values of  $C$  along  $\theta = 180^\circ$  are plotted in figure 18a. If we can assume that the magnetic

FIGURE 17

Phase shifted anomalies from tracks in Figure 16 projected along N 21°E. Degree of phase shift is given in parentheses next to the cruise identification for each track. Correlation of anomalies is extended from the original identifications of Louden (1976).

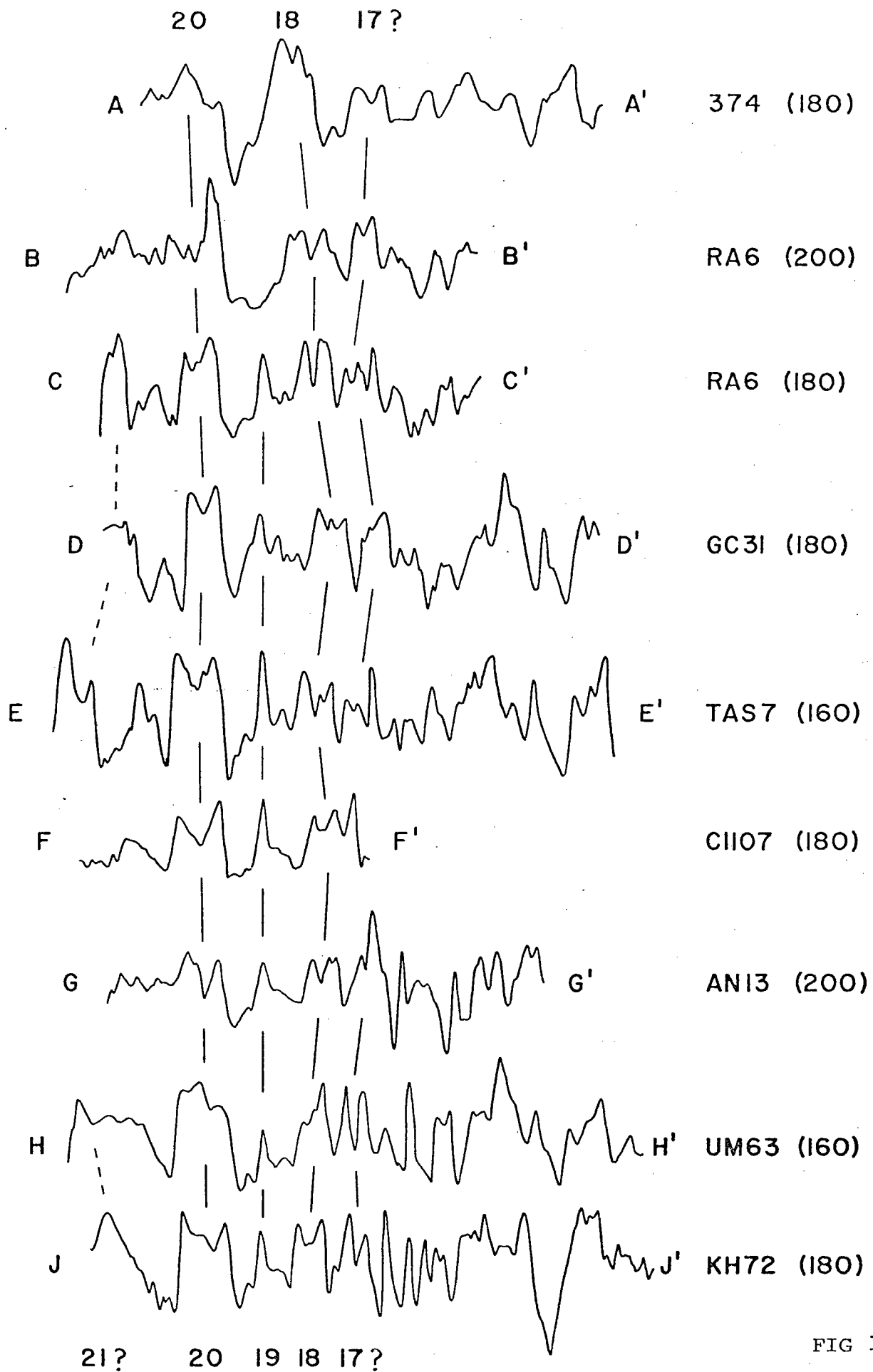


FIG 17

origins of our anomalies are the same as for anomalies in other ocean basins, then we can determine C by comparing the amplitude of our anomalies to those where C is known. This is done in Table V, where we list all of the major East-West anomalies. Excepting one anomaly sequence in the Indian Ocean, all amplitudes of East-West anomalies are large (700-1000γ); and the small amplitudes (200-300γ) of our nearly East-West Philippine Basin anomalies indicate that this plate must have undergone significant rotations in the past. Taking

$$0.2 < C < 0.6$$

limits our locus of pole positions in figure 18a to two areas at either extreme of our great circles, both of which are consistent with mean paleolatitudes for DSDP sites 292 and 294 (figure 18c).

We can compare our West Philippine Basin paleomagnetic pole positions to predictions based on known plate rotations. The rotation of the Philippine plate with respect to the spin axis is the vector sum of rotations of the Pacific plate relative to the spin axis and Philippine plate rotations with respect to the Pacific:

$$\omega_{\text{phil pole}} = \omega_{\text{phil pac}} + \omega_{\text{pac pole}} \tag{9}$$

Finite rotations of the Pacific with respect to the

TABLE V: Magnitude and Location of Major East-West Sea-Floor Spreading Magnetic Anomalies

ID	Anomaly No.	Site Lat	Site Long	Paleo-Lat	Peak to Peak Magnitude	C*	Reference
Phoenix	M1-M10	5°N	175°W	50°S	1000	0.9	Larson and Chase 1972
Wharton Basin	21	4°S	95°E	25°S	400	1.0	Sclater and Fisher 1974
	21	20°S	84°E	40°S	700	1.0	
North Pacific	27-32	50°N	170°W	30°N	1000	1.0	Cande 1976
Galapagos	3-5	3°N	87°W	1°N	700	1.0	Sclater and Klitgord 1973
Philippine Sea	20	15°N	130°E	5°S	200-300	?	Louden 1976

\*  $C = (\sin I \sin I_r) / (\sin I' \sin I'_r)$  where I is the inclination and I' is the effective inclination and r refers to remanent values.

FIGURE 18


Stereonet projections of paleomagnetic pole positions: (a) Small circles determined from paleolatitude measurements on sediments of DSDP sites 290 (dashed lines), 292 (solid lines) and 294 (dotted lines). Pairs of lines enclose areas of 95% confidence. (b) Great circles determined from the phase shifting of magnetic anomalies (Figure 17). Individual circles correspond to phase shifts of  $160^\circ$ ,  $180^\circ$  and  $200^\circ$ . Numbers correspond to magnetic anomaly amplitude factors of  $10.C$  as defined in the text. (c) Combined plots of mean paleolatitude data (lines) and phase shifting (shaded regions) where  $0.2 \leq C \leq 0.6$ . The amplitude constraint is determined from comparison with other east-west anomalies (Table V). (d) Shaded region includes region of (c) which falls between site 292 and 294 paleomagnetic data.  is the mean pole.



FIG. 18 (cont.)

position for Guam Miocene volcanics (Larson et al., 1974). Numbers indicate paleopole determinations from the motion of the Pacific and Philippine plates for various relative Phil-Pac poles:

(1)  $-5^{\circ}$ ,  $-52^{\circ}$  and (2)  $-3^{\circ}$ ,  $-63^{\circ}$  from Karig (1975);  
(3)  $26^{\circ}$ ,  $-37^{\circ}$  and (4)  $-1^{\circ}$ ,  $1^{\circ}$  from Fitch (1972);  
and (5)  $-7^{\circ}$ ,  $-38^{\circ}$  from Katsumata and Sykes (1969).

Each successive number represents an additional  $20^{\circ}$  of relative Phil-Pac rotation away from the Pacific paleopole for 50 m.y.b.p. (open circle) which is calculated from motions of Minster et al. (1974) and Clague and Jarrard (1973).

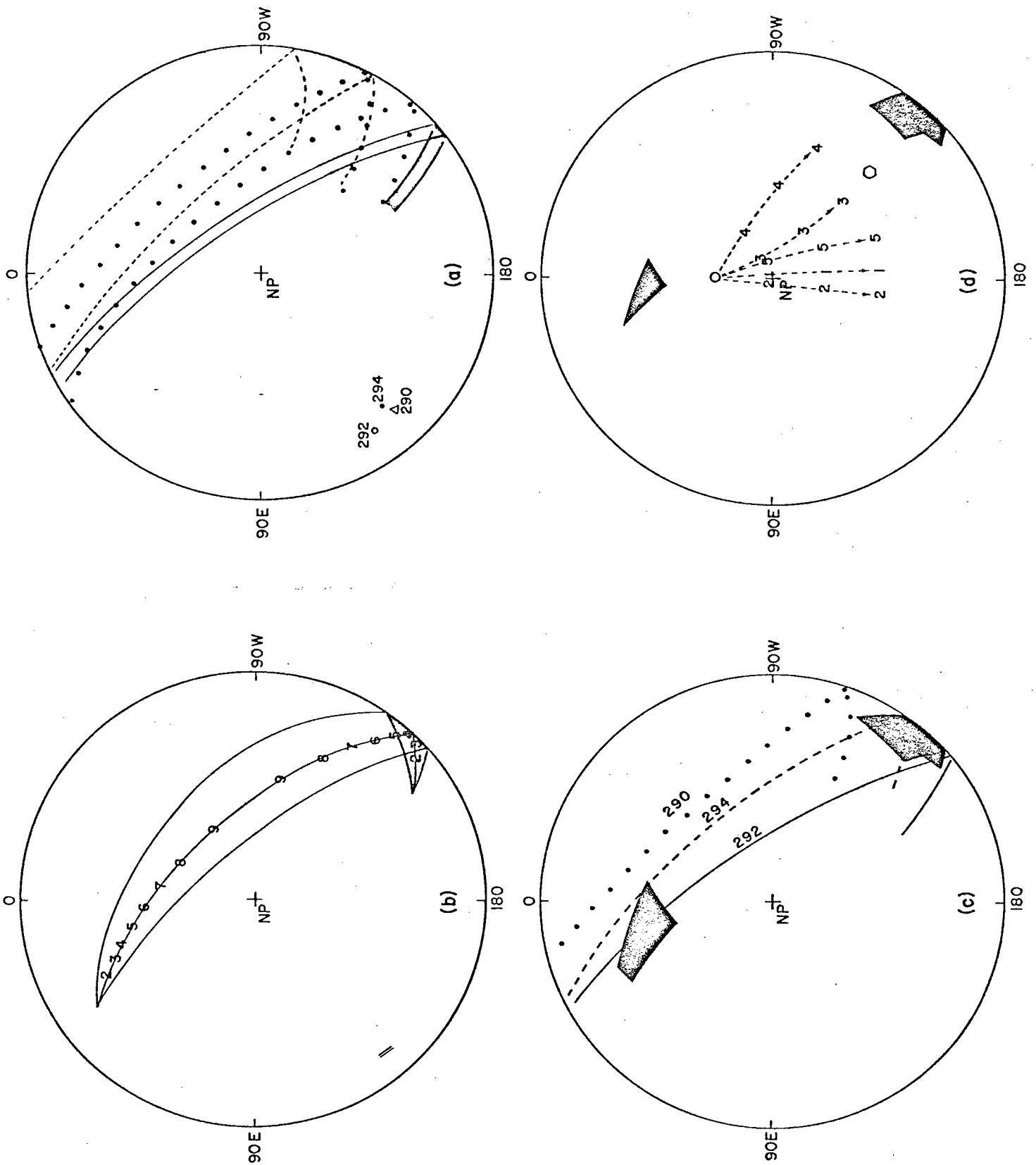


FIG 18

pole are recorded in bathymetric lineations such as the Emperor-Hawaiian seamount chain. Various determinations have been made of this pole for the past 40 m.y. (see Solomon and Sleep [1974] for a summary), but differences are not significant in this study. We use the Minster et al. [1974] pole ( $67^{\circ}\text{S}$ ,  $121^{\circ}\text{E}$ ) and rate ( $8 \cdot 10^{-7}$  deg/yr) over the past 40 m.y. Since our anomalies are as old as 50 m.y., we add to this a rotation of  $10^{\circ}$  ( $10^{-6}$  deg/yr over 10 m.y.) about the Clague and Jarrard [1973] pole ( $17^{\circ}\text{S}$ ,  $73^{\circ}\text{E}$ ) determined from the Emperor seamount chain.

The only information concerning  $\omega_{\text{phil pac}}$  comes from fitting fault plane solutions for recent earthquakes on the margins of the Philippine plate. This has been done by Katsumata and Sykes [1969] for earthquakes on the Izu-Bonin margin of the Pacific and Philippine plates, and by Fitch [1972] and Karig [1975] for earthquakes on the Japan-Ryuku margin of the Philippine and Eurasia plates. These latter solutions are then summed with Eurasia-Pacific rotations to get a Philippine-Pacific pole. These pole positions are not well constrained, are poorly located, and of course are only instantaneous and need not apply over a finite time period. We use them only as a point of reference.

Figure 18d shows the results of calculations based on equation 8. Paleomagnetic pole positions are calculated for varying amounts of rotation about the different Pacific-Philippine poles. Only one of our two possible pole positions

are compatible with these rotations. Plate motion between the Philippine and Pacific plates would have to have been the reverse of what it is today for the other paleopole to be possible. This is evidence for a large amount (  $60^\circ$  ) of clockwise rotation between the Pacific and Philippine plates, which conforms to the theory of Uyeda and Ben-Avraham [1972].

The same rotations can also explain the paleopole position for Miocene volcanic rocks on Guam [Larson et al., 1974]. This could mean that Guam has undergone the same rotations as has the West Philippine Basin, and that these rotations took place before Guam was separated from the West Philippine Basin by the opening of the Parece Vela Basin. In this case such a rotation would not be due to a deformation of the plate boundary alone but would be felt by the entire rigid plate. The other possibility is that these rotations have taken place continuously over the past 50 m.y. and that Guam and the West Philippine Basin have always remained fixed.

#### CONCLUSIONS

The following list is a summary of the major results of this study:

- (1) Sediment cores taken by the Deep Sea Drilling Project can yield well grouped paleolatitudes, especially when results of closely spaced samples can be averaged

together at several separate locations down core. These results can be of greater resolution than those from DSDP basalts, which may not contain enough flow units to adequately average out components of secular variation.

(2) The effect of viscous remanent magnetization in destabilizing the original remanent magnetization in DSDP basalts may be exaggerated by laboratory measurements of up to 1000 hours. Site 292 basalts have large viscosity coefficients while still retaining reversed polarities.

(3) Paleolatitudes for DSDP site 292 sediments indicate that this site formed south of the equator, and results for sites 290 and 294 can be explained in a similar manner. This adds independent justification for our inversion of magnetic anomalies in the West Philippine Basin with respect to North Pacific anomalies, which is necessary in order to identify them with world-wide Eocene reversals.

(4) Paleomagnetic pole positions determined from both the phase shifting of magnetic anomalies and the paleomagnetism of DSDP sites in the West Philippine Basin are consistent with each other, and can be restricted to two relatively well defined positions if the low amplitude of the anomalies is due to a rotation of the Philippine plate. One of these two positions, which includes a large amount of clockwise rotation of the Philippine plate with respect to the magnetic

pole, is consistent with the finite "absolute" rotations of the Pacific plate and present Philippine-Pacific plate rotations. It is also consistent with paleopole positions for Guam and may mean that both have undergone the same motions since the early Miocene.

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APPENDIX A

HEAT FLOW, DEPTH AND CRUSTAL THICKNESS OF  
THE MARGINAL BASINS OF THE SOUTH  
PHILIPPINE SEA

## Heat Flow, Depth, and Crustal Thickness of the Marginal Basins of the South Philippine Sea

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We present 51 heat flow measurements and two geophysical profiles across the West Philippine and Parece Vela basins. We show that both regions have a variable heat flow but that the scatter decreases markedly if we accept as reliable only measurements in areas of uniform sediment drape. Extending this argument to the deep ocean floor shows that the heat flow in these two marginal basins is not necessarily higher than that for deep ocean floor of the same age. On the other hand, the mean depth of both basins is greater and the oceanic crust thinner than the depth and crust of ocean floor of the same age. In the absence of a significant free air gravity anomaly over both basins, we suggest that the thinner crust may account for most of the increase in depth of the two basins. However, more refraction studies are needed to substantiate this difference before this explanation can be unreservedly accepted.

### INTRODUCTION

In the marginal basins of the western Pacific (Figure 1) there appears to be a simple relationship between increasing depth, decreasing heat flow, and increasing age [Karig, 1971]. This relationship is similar to that shown by the midocean ridges, although the heat flow in the younger marginal basins appears to be higher than that for oceanic crust of the same age [Sclater, 1972]. Apart from this difference in heat flow, the general similarity in the dredged rocks and other geological and geophysical parameters has led many to suggest that the intrusion process behind island arcs is similar to that now occurring at the crest of midocean ridges.

The southern half of the Philippine Sea exhibits the classic section of oceanic crust with both active and inactive basins (Figures 2 and 3). Immediately to the west of the volcanic chain of the Mariana arc system lies the Mariana Trough. This active basin has a shallow depth of 2500–3000 m, has outcropping fresh basalt, and is a region in which crustal extension has taken place at least until the recent past [Karig, 1971]. A remanent arc, the South Honshu Ridge, separates this active marginal basin from the inactive Parece Vela Basin. This inactive basin has an average depth of nearly 5 km except for a slightly shallower broad volcanoclastic apron flanking the western side of the South Honshu Ridge (Figure 3a). A zone of high-amplitude ridges and troughs lies close to the center of the basin. These features are best developed south of 20°N but appear to continue to the northern end of the basin. Near 17°N they veer southwestward and may continue as far south as 11°N as a central feature. Sparse calcareous nannofossils at the base of two Deep-Sea Drilling Project (DSDP) holes, 53 and 54, in the basin gave a middle Oligocene to early Miocene age for the lowermost sediment section near the center of the basin.

Apart from greater size and depth, the general structure of the Parece Vela Basin is similar to that of the Mariana Basin. On the basis of these similarities and the sediment ages from the DSDP holes, Karig [1971] has suggested that the Parece Vela Basin is an extensional feature formed by the intrusion of oceanic crust between the late Oligocene and early Miocene.

He has further suggested that the central ridge and trough feature marks the final position of the spreading center and that extension within this marginal basin is symmetrical.

The Palau-Kyushu Ridge, a supposed east-facing remanent arc associated with spreading in the Parece Vela Basin, separates this basin and the West Philippine Basin. The West Philippine Basin is by far the largest of the basins in the Philippine Sea and is the largest and the deepest (5500–6000 m) marginal basin in the world. A major system of troughs and ridges called the Central Basin Fault runs from northwest to southeast through the center of the basin. Deep-sea drilling in the West Philippine Basin yielded sediments no older than middle Eocene [Ingle *et al.*, 1975]. Careful analysis of the lineated magnetic anomalies in this basin, first observed by Ben-Avraham *et al.* [1972] and related to the magnetic time scale by Louden [1976], has shown that they are lineated N110°E. The anomalies are best matched by a symmetrical sequence on either side of the Central Basin Fault starting with anomaly 17 and ending with anomaly 22. On both the time scales of Heirtzler *et al.* [1968] and the modification of Sclater *et al.* [1974a] this sequence yields a middle to late Eocene age for the southwestern portion of the West Philippine Basin. These results are in excellent agreement with the age suggested from the analysis of piston cores in slumps [Karig, 1971] and from the provisional ages of the sediments at the base of the DSDP holes in the area [Ingle *et al.*, 1975]. North of the Central Basin Fault the anomalies are lineated, but identification is less certain. At present they are best matched by assuming symmetry about the fault. Unfortunately, the identification is sufficiently ill-defined so that it is possible that the anomalies could be a sequence younger than 17. Whichever interpretation is correct, we are confident that the basin is no older than Paleocene. The major question still unanswered is whether this crust was formed at a simple midocean ridge system [Uyeda and Ben-Avraham, 1972] or whether its tectonic environment was that of extension immediately behind an active subduction zone.

In this paper we present two each of bathymetric, seismic reflection, heat flow, and magnetic profiles and one gravity profile across the southern section of the Philippine Sea (Figure 3). We ran these profiles, concentrating on the surface heat flow, to investigate the similarities and differences between the West Philippine and Parece Vela basins and the ocean floor of the same age. We will show with a reanalysis of the relation of surface heat flow values and continuous uniform sediment cover that the high heat flow behind island arcs may not be

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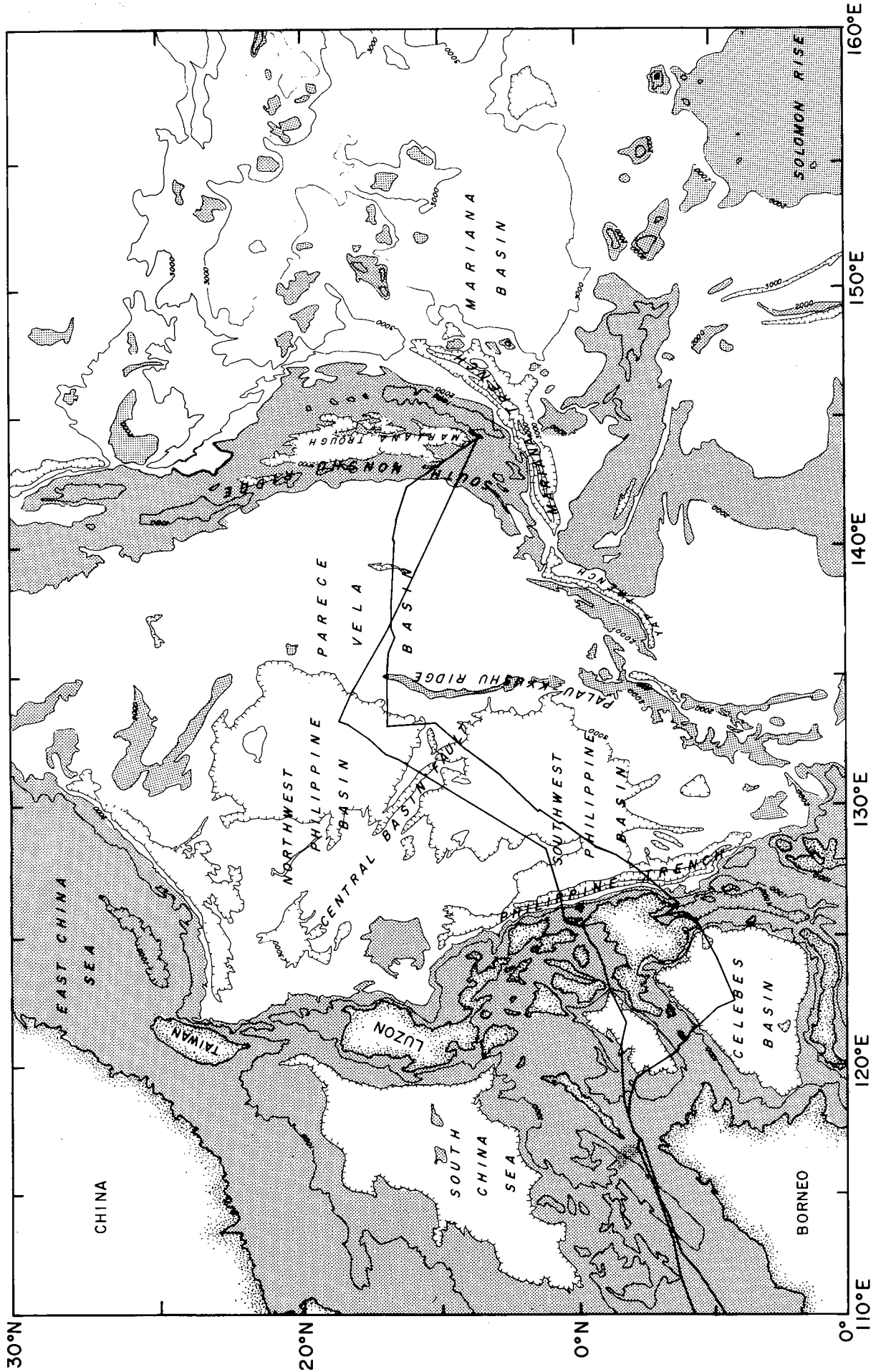


Fig. 1. A sketch of the Philippine Basin showing the principal topographic features and the track of the Antipode and Tasaday lines. The depths above 2000 fm (1 fm = 1.82 m) have been shaded.



significantly higher than that in normal oceanic crust of the same age. However, both marginal basins differ in two important features from oceanic crust of the same age. The basins are deeper and have a thinner oceanic crust. We suggest that these two features may be related.

#### HEAT FLOW MEASUREMENTS

In June and July 1971 we obtained 39 heat flow measurements in the marginal basins of the Indonesian region and

across the Philippine Sea from the R/V *Melville* on leg 13 of expedition Antipode of the Scripps Institution of Oceanography. Twelve further measurements in this area were added during July 1973 from the R/V *Thomas Washington* during leg 7 of Scripps Institution of Oceanography expedition Tasaday. We measured all the temperatures with a 2.5-m-long Bullard probe [Corry *et al.*, 1968]. On the first expedition we determined the thermal conductivities by two methods. In the first method we measured the conductivity of sediment recovered

TABLE 1. Heat Flow Stations in the Western Pacific

Station	Location		Corrected Depth, m	Gradient, °C/m	Conductivity,* 10 <sup>-3</sup> cal/°C cm s	Heat Flow, μcal/cm <sup>2</sup> s	Pene- tration†	Tilt, deg	Type of Environment
	N	E							
Antipode 13									
176	6°34.9'	113°50.5'	2350	0.92	1.68 (N)	1.55	f	0-15	B
177	6°46.3'	114°15.3'	2863	0.80	1.67 (A)	1.33	f	0-15	A
178	6°58.3'	114°39.7'	2882	0.92	1.66 (N)	1.52	f	0-15	A
179	7°07.7'	115°07.0'	2844	0.83	1.67 (A)	1.39	f	0-15	B
180	7°51.5'	119°19.5'	3641						
181	7°17.6'	119°45.9'	3801	1.08	1.74 (A)	1.90	f	0-15	A
182	6°41.0'	120°09.0'	3641	1.02	1.74 (N)	1.77	f	0-15	B-D
183	4°55.0'	121°53.0'	4790						
184	3°58.2'	122°14.0'	4965	0.90	1.66 (N)	1.49	f	0-15	A
185	4°39.7'	123°48.5'	4942	0.88	1.66 (A)	1.49	f	0-15	A
186	5°53.5'	125°48.5'	3007	0.24	1.66 (A)	0.41	f	0-15	D
188	8°04.0'	127°46.1'	5460	0.55	1.90 (A)	1.04	f	0-15	B-D
189	8°59.5'	128°18.0'	5827	0.78	2.00 (I)	1.56	f	0-15	A
190	9°41.9'	128°45.1'	5746	0.86	1.90 (I)	1.65	f	0-15	A
191	10°30.2'	129°11.6'	5766	0.81	2.00 (I)	1.62	f	0-15	A
192	11°23.2'	129°43.9'	5470	1.03	1.90 (A)	1.96	f	0-15	A-B
193	12°13.0'	130°27.2'	5754	0.88	1.96 (I)	1.73	f	0-15	B
194	12°57.5'	131°16.5'	5752	0.60	1.83 (I)	1.10	f	0-15	B-D
195	13°43.6'	131°55.2'	5900	2.02	1.98 (I)	3.99	f	0-15	C
196	13°59.5'	132°08.9'	5959	0.86	1.90 (A)	1.63	f	0-15	C
197	14°43.8'	132°46.7'	5343	...	1.93 (I)	0.95	75 cm	>30	C
199	15°33.7'	133°14.2'	5782	0.24	2.01 (I)	0.49	f	0-15	D
200	16°26.5'	133°12.8'	5790	0.61	2.09 (I)	1.28	f	0-15	C-D
201	17°03.6'	133°15.3'	5712	0.55	1.95 (I)	1.07	f	0-15	B
202	17°03.0'	134°02.0'	5493	0.10	2.05 (I)	0.20	f	0-15	A
203	17°05.8'	134°33.7'	5398	0.87	1.98 (I)	1.72	f	0-15	A
205	17°00.0'	135°22.1'	5073	0.40	1.82 (I)	0.73	f	0-15	D
206	16°57.1'	135°54.0'	5084	0.41	1.92 (I)	0.78	f	0-15	C
207	16°49.0'	136°35.5'	4716	0.28	1.91 (I)	0.54	f	0-15	C
208	16°55.0'	136°57.5'	4797?	0.24	1.91 (I)	0.45	f	0-15	C-D
209	16°53.1'	137°24.8'	5201	0.43	1.79 (I)	0.78	f	0-15	C
210	16°53.0'	137°53.0'	4586	2.68	1.83 (I)	4.91	f	0-15	C
211	16°52.3'	138°15.8'	4171						
214	16°46.6'	140°10.2'	4609	1.50	2.07 (I)	3.11	f	0-15	C
215	16°47.2'	140°44.6'	4816	0.28	1.85 (I)	0.53	f	0-15	A
216	16°39.7'	141°24.4'	4653	1.03	1.88 (I)	1.94	f	0-15	A
217	16°31.3'	141°49.0'	4506	1.14	1.90 (I)	2.16	f	0-15	A
218	16°22.9'	142°09.6'	4170	1.10	1.89 (I)	2.08	f	0-15	A
220	16°20.2'	142°33.8'	3672	0.23	1.98 (I)	0.46	f	0-15	B-D
Tasaday 7									
01	8°17.5'	120°50.2'	3932	1.03	1.90 (A)	1.96	f	0-15	?
02	8°25.8'	121°43.1'	4857	0.31	1.90 (A)	0.59	f	0-15	D
03	13°31.8'	129°58.2'	5673	0.94	1.90 (A)	1.79	f	0-15	A
04	15°28.6'	131°12.1'	5450	1.00	1.90 (A)	1.90	f	0-15	C
05	16°43.6'	132°06.9'	5761	0.49	1.90 (A)	0.93	f	0-15	C
06	17°23.9'	132°31.0'	5958	0.38	1.90 (A)	0.72	f	0-15	D
07	18°42.8'	133°24.8'	6132	1.28	1.90 (A)	2.43	f	0-15	A
08	17°34.7'	135°49.4'	4859	0.61	1.90 (A)	1.16	f	0-15	C
09	17°18.2'	136°22.2'	5053	0.19	1.90 (A)	0.36	f	0-15	D
10	17°00.6'	137°03.0'	5072	0.37	1.90 (A)	0.70	190 cm	>30	C-D
11	16°30.8'	138°03.9'	4883	0.29	1.90 (A)	0.55	f	0-15	C-D
12	16°08.6'	138°56.1'	5120	0.47	1.90 (A)	0.89	f	0-15	C-D
13	15°52.4'	139°34.3'	5206						C

Absence of data at station 180 was caused by the instrument catching fire and at stations 183 and 211 by the instruments falling over. At station 13 no data were recorded.

\* Here N indicates needle probe conductivity measurement, I indicates in situ conductivity measurement, and A indicates assumed conductivity.

† Here f indicates full penetration.

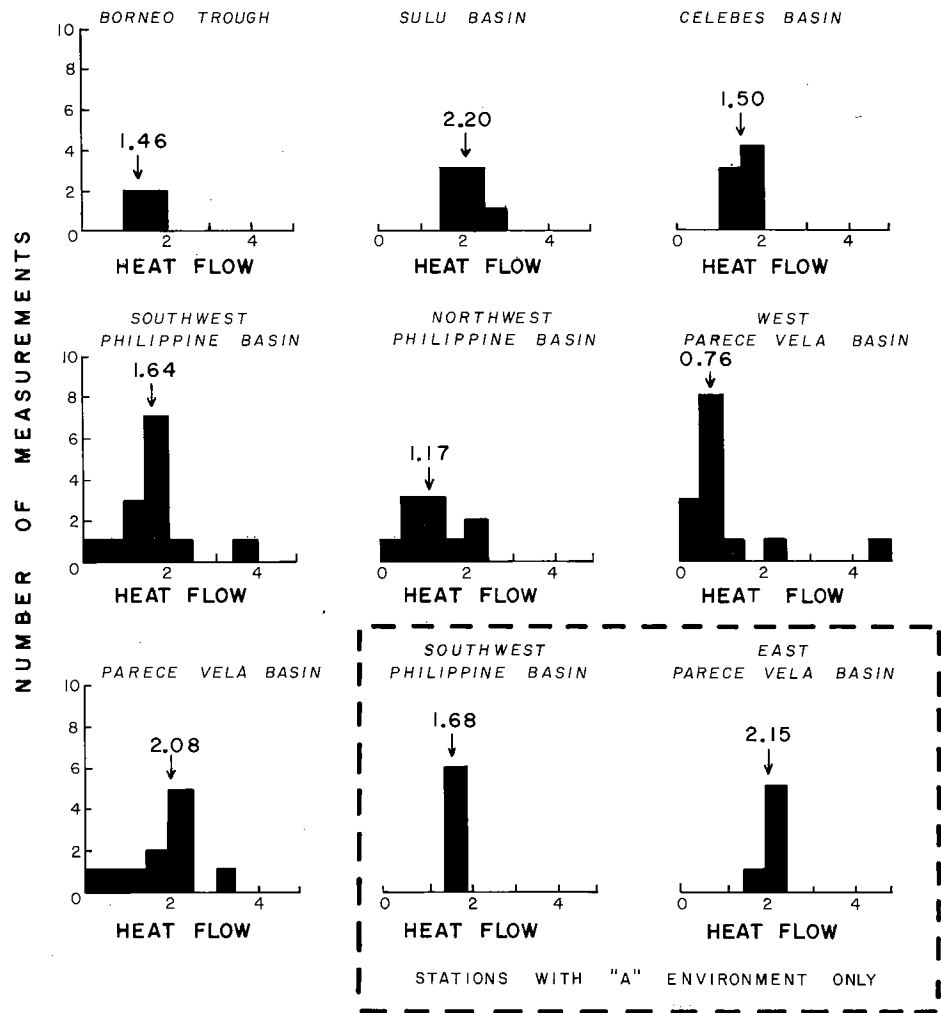


Fig. 4. Histogram of heat flow values in the tectonic provinces of the south Philippine Sea and marginal basins of the Indonesian area. The median values are presented above each histogram.

by deep-sea cores, using the needle probe technique described by von Herzen and Maxwell [1959]. In the second we determined the in situ conductivity while the Bullard probe was on the bottom during the temperature gradient measurement, utilizing the technique of Corry et al. [1968]. For the Tasaday

stations in the Philippine Sea we assumed a value of  $1.90 \times 10^{-3}$  cal/°C cm s, the mean value of the conductivities measured in the south Philippine Sea. For all the stations the heat flow was taken as the product of the temperature gradient and the conductivity.

TABLE 2. Heat Flow Statistics

Area	Number of Measurements	Heat Flow, $\mu\text{cal}/\text{cm}^2 \text{ s}$		
		Median	Mean	Standard Deviation
<i>Raw Data</i>				
Borneo Trough	4	1.46	1.45	0.12
Sulu Basin	7	2.20	2.13	0.30
Celebes Basin	7	1.50	1.51	0.23
Southwest Philippine Basin	14	1.64	1.59	0.86
Northeast Philippine Basin	10	1.17	1.24	0.69
Western Parece Vela Basin	14	0.76	1.10	1.19
Eastern Parece Vela Basin	11	2.08	1.80	0.81
All West Philippine Basin	24	1.48	1.44	0.80
All Parece Vela Basin	25	1.00	1.40	1.08
<i>Reliable Data</i>				
Southwest Philippine Basin	6	1.68	1.72	0.16
Eastern Parece Vela Basin	6	2.15	2.15	0.16

Reliable data are defined as measurements taken in an area of uniform sediment drape.

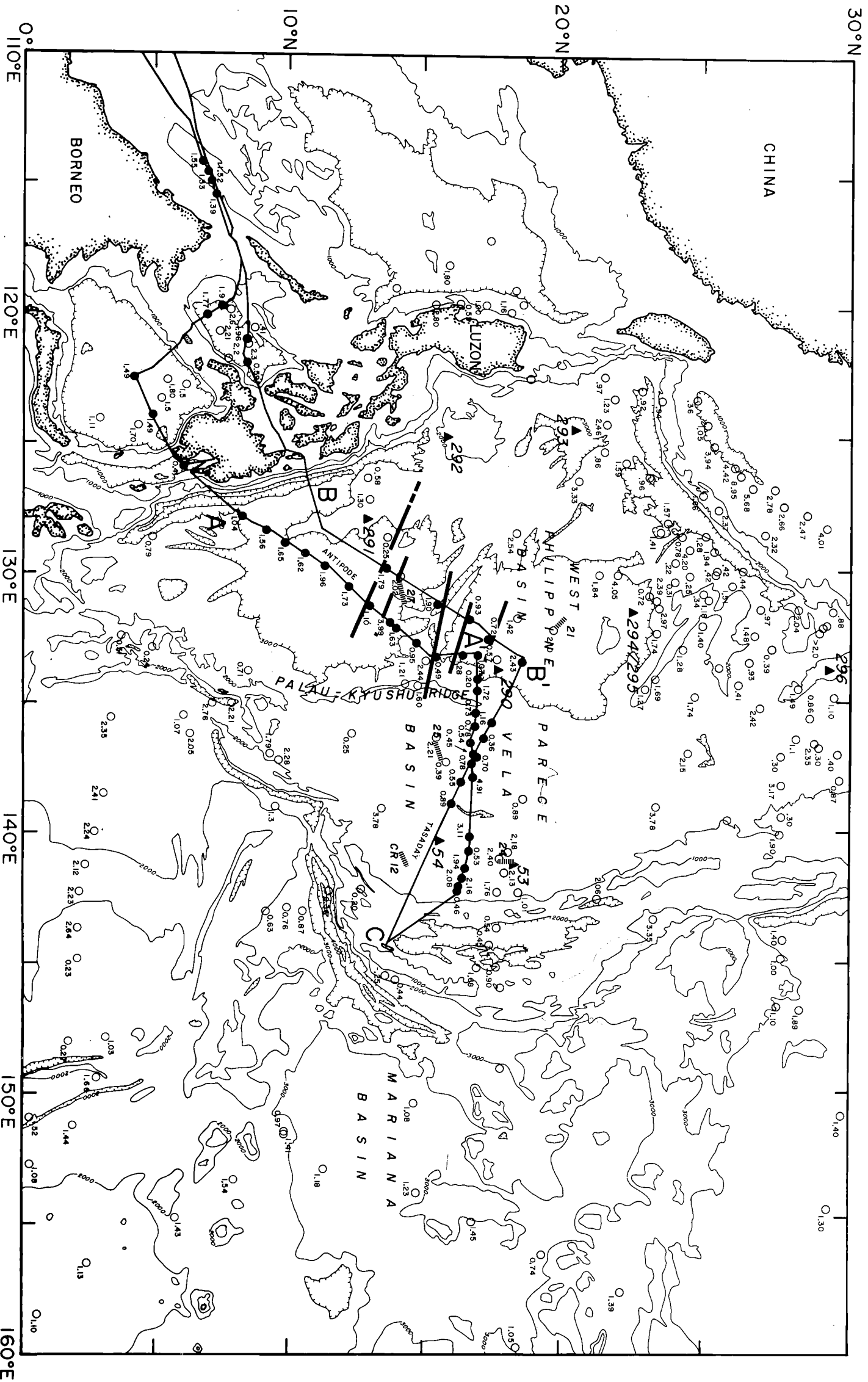
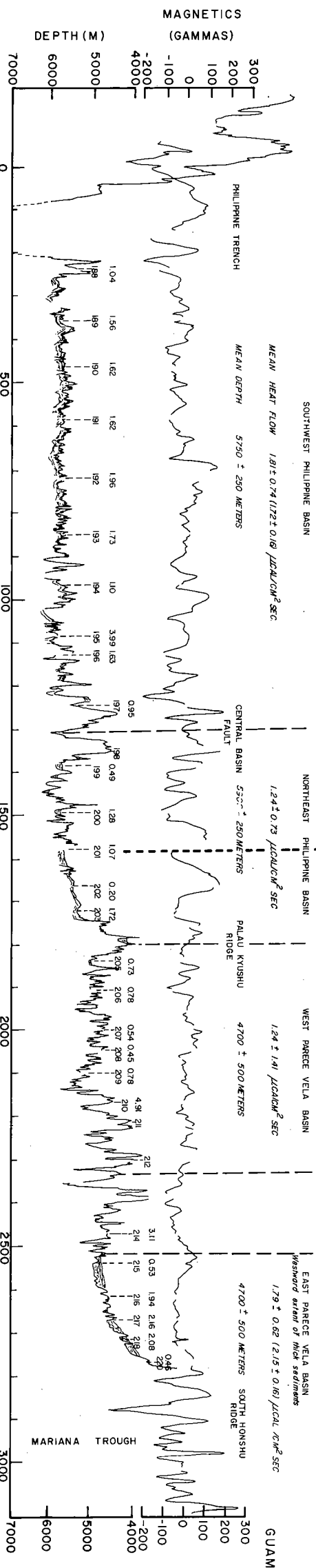


Fig. 2. Topographic chart of the Philippine Sea and marginal basins close to Indonesia. Heat flow stations, the track of Antipode 13 and Tassaday 7, and identified magnetic anomalies (after *Louden* [1976]) have been superimposed upon the chart. The solid circles are heat flow stations presented in this paper, and the solid triangles are Joides sites. Note the refraction stations.

A ANTIPODE



A'

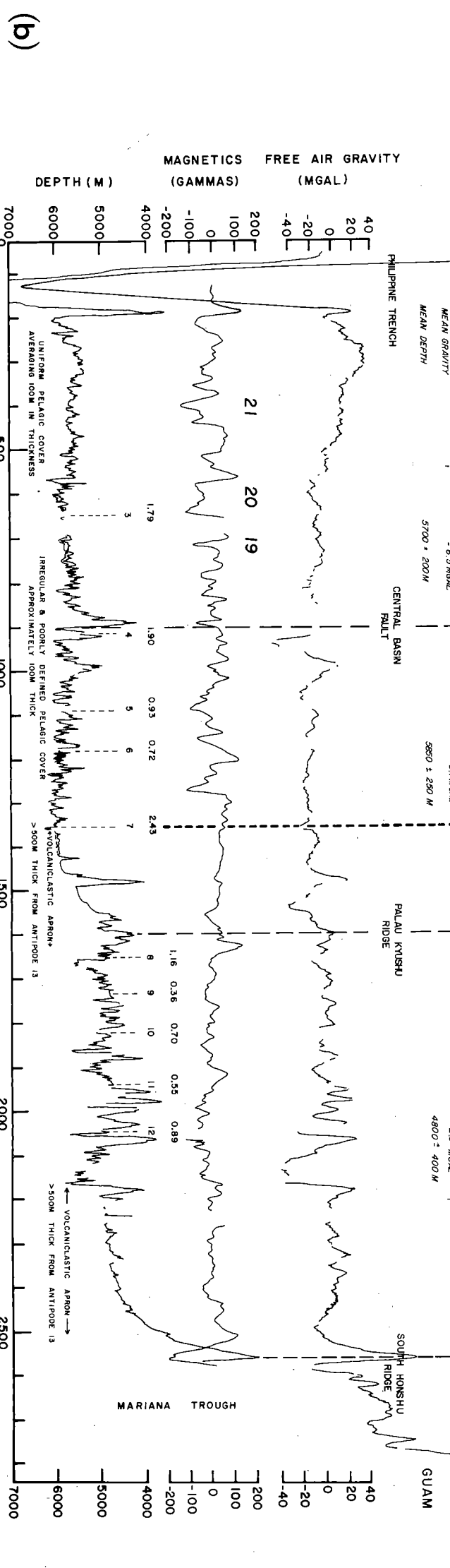
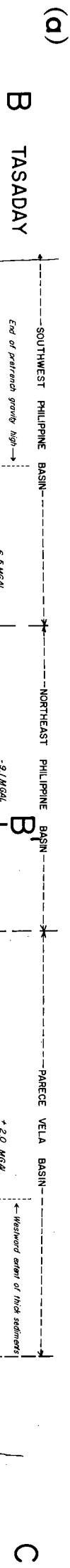


Fig. 3. (a) Bathymetric and line drawing of the Antipode 13 profile between the Philippine trench and the South Honshu Ridge. Heat flow measurements are shown along the track. Also presented are the mean heat flow and depth range for each of the separate tectonic provinces traversed by the profile. Because of instrument failure, only a magnetic profile was obtained across the Mariana Basin. (b) Same as Figure 3a, with free air gravity anomaly added for Tasaday 7 profile. (There is some evidence from cross-checking with an unpublished compilation of gravity data in the Philippine area (T. Watts, personal communication, 1975) that the gravity data west of the Central Basin Fault are 15 mGal low. The rest of the data agree with the unpublished compilation.) Also presented are mean gravity, mean depth, and range in depth for the separate tectonic provinces. Because of instrument failure, only a gravity profile was obtained across the Mariana Basin.

TABLE 3a. Depth and Age From Joides Deep-Sea Drilling Sites

Basin	Site	Location		Depth, m	Sediment, m		Topographic Correction, m	Corrected Depth, m	Biostratigraphic Age		Magnetic Anomaly	
		N	E		Thick-ness	Correc-tion			Period	Age, m.y.	Number	Age, m.y.
Parece Vela	53	18°02.0'	141°11.5'	4629	201	144	200*	~4830	Oligocene to early Miocene	32-18	...	...
	54	15°36.6'	140°18.1'	4990	294	212	...	5102	Oligocene to early Miocene	32-18	...	...
Philippine	290	17°44.9'	133°28.1'	6062	255	185	-150†	~5900	Late Eocene	45-38	<21	<53
	291	12°48.4'	127°49.9'	5217	127	93	+400†	~5600	Middle to late Eocene	50-38	21	53-49
	293	20°21.3'	124°05.6'	5599	563	405	?	~6004	?	?		
	294	22°34.7'	131°23.1'	5784	118	85	...	5869	Eocene?	54-45		

Parece Vela Basin sites are taken from *Fischer et al.* [1971], Philippine Basin sites from *Ingle et al.* [1975], and biostratigraphic ages from *Berggren and Van Couvering* [1974].

\* Correction from site 53 bathymetric chart [*Fischer et al.*, 1971].

† Correction from *Chase et al.* [1970].

The heat flow measurements were taken in a wide variety of places ranging from the enclosed Borneo, Sulu, and Celebes basins to the open and deep West Philippine Basin (Figure 2 and Table 1). On initial inspection the measurements, except for those of the enclosed basins, are not consistent (Figure 4 and Table 2). Even when the results are consistent, we do not know how to interpret them. For instance, the Borneo Basin has a uniform flow of  $1.45 \mu\text{cal}/\text{cm}^2 \text{ s}$ . The Sulu Trough has a higher mean heat flow,  $2.13 \mu\text{cal}/\text{cm}^2 \text{ s}$ , than the Celebes Basin,  $1.51 \mu\text{cal}/\text{cm}^2 \text{ s}$  (Table 2), but the reason for this difference is not known.

The variability in heat flow values starts in the West Philippine Basin and continues through the Parece Vela Basin. In both basins (Figures 3a and 3b) there is a reasonable correlation between lack of variability and a uniform sediment cover. The West Philippine Basin south of the Central Basin Fault has a uniform sediment drape, and the heat flow has little scatter. Within the fault and to the north, the topography is rough, the sediment drape less obvious, and the heat flow more variable. A spectacular example of such variability is found in the Parece Vela Basin. This basin is cut in two by a north-south-central region of peaks and troughs which is the presumed termination point of the active spreading center. Both sides of the basin are thought to be of the same age. Deep-sea drilling sites in the eastern basin gave an Oligocene to early Miocene age (Table 3). The only striking difference is that the eastern basin is covered by a uniform drape of sediments. The heat flow measurements in this well-sedimented area are high and fairly uniform ( $1.80 \pm 0.81 \mu\text{cal}/\text{cm}^2 \text{ s}$ ). On the other hand, the heat flow measurements in the western basin are generally low with a few very scattered high values ( $1.10 \pm 1.19 \mu\text{cal}/\text{cm}^2 \text{ s}$ ) which usually occur on top of topographic highs (stations 210 and 214 in Figure 3a).

This wide difference in heat flow variability in two supposedly similar tectonic basins is very disturbing. *Lister* [1972] and *Williams et al.* [1974] have suggested that because of cooling cracks and faulting the oceanic crust is permeable to fluid flow. *Sclater et al.* [1974b] have further suggested that if this effective permeability is never completely destroyed, then in any region of rough topography with outcropping basement highs, hydrothermal circulation will occur. These circulation patterns will cause much of the heat loss to the sea water to occur by mass transport through the permeable basement. Also, the patterns that they set up will tend to lower the conductive heat flow in the sediment ponds between the outcropping basement highs. As a consequence, conductive measurements in such regions, except in isolated cases on top of basement highs, will yield low values.

We believe that this model gives a good qualitative explanation of the heat flow across the Parece Vela Basin. In the western basin, where there is little sediment cover and a highly variable topography, most of the values lie below  $1.00 \mu\text{cal}/\text{cm}^2 \text{ s}$ . In fact, the median is  $0.76 \mu\text{cal}/\text{cm}^2 \text{ s}$ . However, in the eastern Parece Vela Basin just west of the South Honshu Ridge there is a thick sediment drape extending 200 km from the ridge (Figure 3). In this area, except for the two stations at the edge, the heat flow values show much less scatter (Figure 3a) and have a median of  $2.08 \mu\text{cal}/\text{cm}^2 \text{ s}$ .

On the basis of our qualitative explanation of the low values we reexamined the environment around each of the stations on the Tasaday, Antipode, and Scan [*Sclater et al.*, 1972] cruises in the area. We divided the environments into four groups: A, a uniform sediment drape with no nearby sediment outcrops; B, a uniform sediment drape with isolated nearby outcrops; C, very thin or no sediment drape; and D, a sediment pond in rough topography next to an outcropping basement high.

TABLE 3b. Depth and Age From Magnetic Anomalies and Topographic Profiles in the Philippine Basin

Area	Number of Measurements, m	Mean Depth, m	Estimated Range, m	Age Range, m.y.
Anomaly 17, Central Basin Fault	6	5050	$\pm 200$	43-39
Anomaly 21, southwest Philippine Basin	5	5750	$\pm 300$	53-49
Anomaly 21, northeast Philippine Basin	5	5950	$\pm 200$	53-49

Topographic profiles are taken from the present paper and *Louden* [1976], and magnetic anomaly identifications from *Louden* [1976].

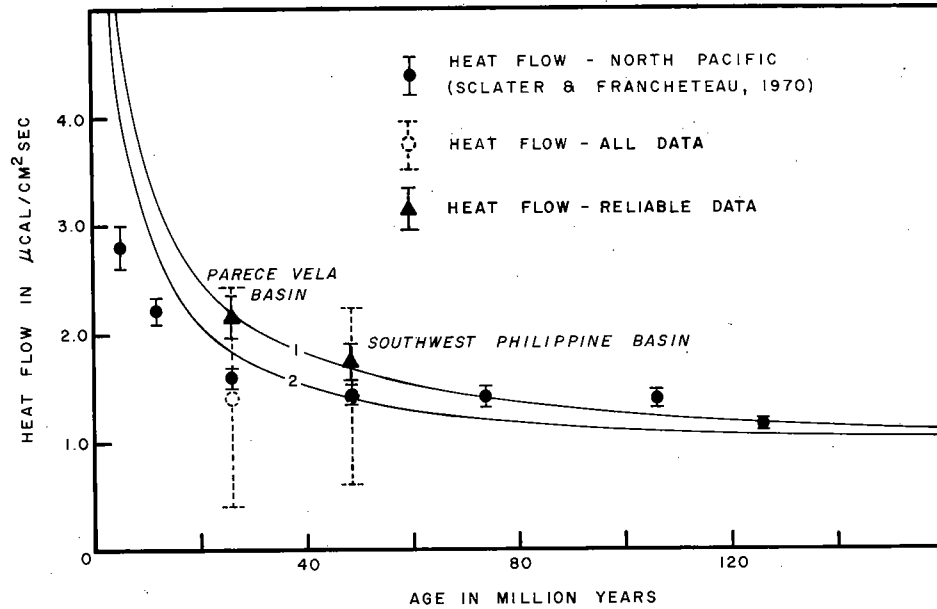


Fig. 5. The mean and standard deviation for all data and for selected reliable data from the southwest Philippine and Parece Vela basins superimposed upon a plot of mean heat flow and standard error for tectonic provinces in the North Pacific against age. Curves 1 and 2 represent the theoretical heat flow for 100-km- and 75-km-thick lithosphere plates, respectively (after Sclater and Francheteau [1970]).

We found six stations in the eastern Parece Vela Basin with an A environment. The median was 2.15 with a standard deviation of  $\pm 0.16 \mu\text{cal}/\text{cm}^2 \text{ s}$  (Figure 4). The fact that the scatter has been considerably reduced by using only A environment stations is significant. It indicates that only where the sediment cover (which is presumed to be impermeable) is thick enough and sufficiently wide to stop any regional loss of heat

by mass transport will reliable measurements of low scatter be obtained. We believe that the true heat flow from the eastern Parece Vela Basin is  $2.15 \pm 0.16 \mu\text{cal}/\text{cm}^2 \text{ s}$ . This is high and significantly different from that in the western basin. We carried out a similar exercise for the southwestern Philippine Basin. Again, when only the A stations are considered, the scatter is low, and we obtained a value of  $1.72 \pm 0.16 \mu\text{cal}/\text{cm}^2 \text{ s}$  (Figure 4).

This reinterpretation of data demonstrates an important and well-known fact; namely, much of the variability of oceanic heat flow measurement is a function of environment. However, if regions can be found with a uniform sediment drape of thickness greater than 100 m and an extent that is wider than that of probable hydrothermal convection cells (5-10 km), then heat flow measurements of surprisingly low scatter should be observed. If, in these areas, the effects of slumping and bottom water temperature variations can be shown to be small, then these heat flow values may be a true measure of the flow of depth beneath the oceanic crust.

COMPARISON OF HEAT FLOW AND DEPTH MEASUREMENTS

We compared the reliable heat flow average over the southern section of the West Philippine Basin and the eastern Parece Vela Basin with that over areas of the Pacific of the same age (Figure 5). Though both areas have higher heat flows than the North Pacific, the difference cannot be considered significant. In the North Pacific no account was taken of local environment, and as a consequence, the averages in the midocean ridges are almost certainly biased in the low direction. Recently, Sclater *et al.* [1975] have shown that if only heat flow values from well-sedimented areas are considered, the means when plotted against age are very close to the values predicted by a 100-km-thick lithosphere. Even the heat flow through the Parece Vela Basin is not significantly greater than the values observed over Oligocene crust in the North Pacific when the large number of unreliable low values used in determining the mean for ocean floor of the same age are removed.

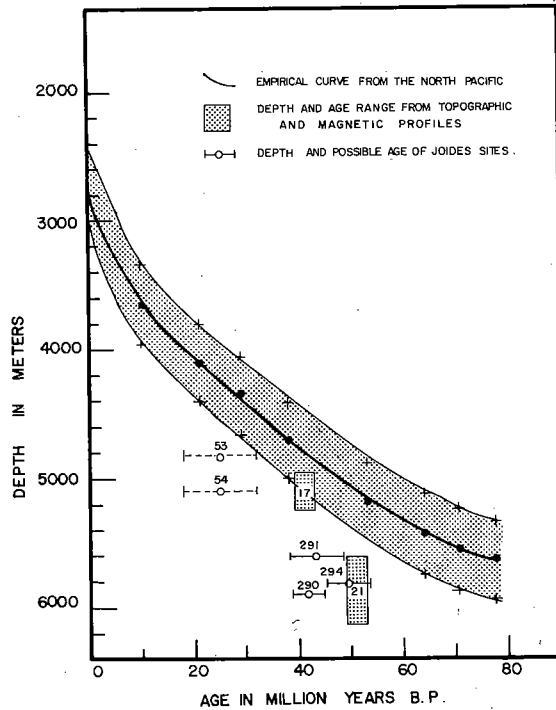


Fig. 6. Depth versus age data for the northeastern Pacific compared with similar data from the Philippine Sea. The stippled area represents a range of  $\pm 300 \text{ m}$  about the empirical curve for the northeastern Pacific. This is the expected range for an area with no free air gravity anomaly [Parsons and Sclater, 1975].

TABLE 4a. A Comparison of Velocity and Thickness of the Crustal Layers in the West Philippine and Parece Vela Basins and the North Pacific

Area	Layer	Velocity, km/s	Thickness, km	Assumed Density, g/m <sup>3</sup>	Data Source
West Philippine Basin	T	5.30 ±	0.92 ±	2.5	<i>Murauchi et al.</i> [1968] (stations 21 and 27)
	O	6.70 ±	4.03 ±	2.9	<i>Henry et al.</i> [1975] (station Tasaday 5-2)
Parece Vela Basin*	T	4.45	1.39	2.4	<i>Murauchi et al.</i> [1968] (stations 24N, 25W and 25E)
	O	6.81	3.00	2.9	<i>Gaskell et al.</i> [1959] (CR 12)
Western North Pacific	T	5.02 ± 0.63	2.05 ± 0.50	2.5	<i>Raitt</i> [1963]
	O	6.73 ± 0.25	5.08 ± 1.72	2.9	
North Pacific	T	5.19 ± 0.64	1.49 ± 0.98	2.5	<i>Shor et al.</i> [1970]
	O	6.81 ± 0.16	4.62 ± 1.30	2.9	
Hypothetical crust, 45 m.y. B.P.	T	5.1	1.66	2.5	<i>Le Pichon et al.</i> [1973]
	O	6.7	5.33	2.9	

T and O are the transition and oceanic layers, respectively, after *Shor et al.* [1970].

\* These data are not reliable; 24S was omitted because it had an unusually thick oceanic layer.

To investigate the tectonic similarities of the basins in the Philippine Sea further, we plotted their depth as a function of age and compared the basins with the North Pacific (Figure 6). In this case, both the Parece Vela and the West Philippine basins north and south of the Central Basin Fault have depths that are significantly greater than the depth predicted by the empirical curve for the North Pacific. A possible explanation of the differences for the Parece Vela Basin is that the lithospheric plate is thinner landward of the trench because of shear stresses along the bottom of the plate [*McKenzie*, 1969]. In this case the thinner plate cools more quickly, and by 30 m.y. B.P. it has reached equilibrium. The thinner lithosphere results in a high heat flow and a depth close to 4600 m [*Sclater*, 1972]. Though this mechanism can be invoked to account for the heat flow and elevation in the Parece Vela Basin, it cannot also work for the West Philippine Basin. In this case the depths are so deep that the residual elevation anomaly (actual depth minus predicted depth for crust of the same age) cannot be caused by thinning the plate. Any decrease in thickness of the lithosphere in equilibrium crust will increase the elevation, not decrease it [*Sclater*, 1972]. Thus some other explanation must be found for the deeper than normal depths in these two basins.

*Karig* [1971] pointed out that there was some evidence that the oceanic crust in the marginal basins of the Philippine Sea was thinner than that under the deep oceans and that this difference could account for the greater than normal depths [*Karig*, 1974]. We also observe that the mean free air gravity anomaly over both basins is less than  $\pm 10$  mGal (Figure 3b). Thus it appears that the basins are compensated and that the thinner crust could account for the deeper depths. To investigate this effect, we have compiled the reliable seismic data from *Murauchi et al.* [1968], *Henry et al.* [1975], and *Gaskell et al.* [1959] for the West Philippine Basin and the Parece Vela Basin. We separated the oceanic crust into three layers, sediment, transition, and oceanic, after the definition presented in Figure 1 of *Shor et al.* [1970], and we compared these with the compilations of *Raitt* [1963] and *Shor et al.* [1970] and with the theoretical northwestern Pacific crustal section of *Le Pichon et al.* [1973] (Table 4a). We included the three different compilations which represent the average for old ocean floor, the average for a fairly young ocean floor, and an empirical curve to give an estimate of the likely range of thickness of the two layers in normal oceanic crust. Both the West Philippine and the Parece Vela basins have a crustal thickness which is much less than that of any of the three

estimates for normal oceanic crust. We have calculated (Table 4b) that in the West Philippine Basin this decrease in crustal thickness will produce a residual depth anomaly of between 300 and 600 m. In the Parece Vela Basin the anomaly will be slightly less, 280–570 m. Clearly, much of the extra depth in these two basins can be accounted for by a thinner crust.

Before this explanation of the excess depths in the marginal basins can be unreservedly accepted, it is necessary to point out that there are problems with this mechanism. First, the seismic data coverage in both basins is poor, and more reversed lines are needed to substantiate the evidence of a thinner crust. As a consequence, the data have a high scatter, and the differences between the means for the basins and those for oceanic crust of the same age are only marginally significant. Second, there is no evidence from dredged rocks to indicate that the oceanic crustal layers may have a different composition under the marginal basins. Finally, the deep ocean floor in the North Pacific does not have such a large variation in crustal thickness [*Shor et al.*, 1970]. Thus some new special mechanism must be advanced to explain how and why this small section of the ocean floor should have a thinner crust than that created at an oceanic spreading center.

#### CONCLUSIONS

We have shown that both the Parece Vela and the West Philippine basins have mean depths that are deeper and an oceanic crust that is thinner than ocean floor of the same age. Both have higher than normal heat flow, but this heat flow is not necessarily greater than that of ocean floor of the same age. Both basins appear to have been formed by crustal extension originating at the center of the basin. However, in the case of the Parece Vela Basin this extension was east-west, while in the West Philippine Basin it was north-south. The West Philip-

TABLE 4b. Expected Increase In Depth Resulting From Thinner Crustal Layers

	Depth Increase, m
West Philippine Basin	
with respect to western North Pacific	~110
with respect to North Pacific	~320
with respect to hypothetical crust, 45 m.y. B.P.	~500
Parece Vela Basin	
with respect to western North Pacific	~570
with respect to North Pacific	~280
with respect to hypothetical crust, 30 m.y. B.P.	~360

pine Basin has clearly identifiable magnetic anomalies showing a spreading rate during extension of 6-cm/yr half rate. Only one other marginal basin, the Scotia spreading center [Barker, 1972], shows more easily identified magnetic lineations, and it has a half rate of 3 cm/yr. With present information it is clear that the West Philippine Basin is different from the other marginal basins in the western Pacific. However, it also has two significant differences from deep ocean floor of the same age. Hence it is not possible at the present moment to determine the environment in which this basin was formed.

In this view it is important to realize that explanations other than crustal variations exist to explain residual depth anomalies. It is possible that in such an area behind one downgoing slab and near another, complex dynamic forces could be acting to decrease the observed depth. A similar mechanism invoking dynamic forces in the upper mantle has already been advanced by Sclater et al. [1975] to explain the observed correlation between free air gravity anomalies and residual elevation anomalies in the North Atlantic. One of the disturbing features of accepting the thinner oceanic crust to explain the deeper depths in the marginal basins is that it adds another hypothesis to be considered when the worldwide correlation of residual depth and free air gravity anomalies is interpreted.

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APPENDIX B

PLOTS OF MAGNETIC ANOMALIES ALONG  
TRACK IN THE PHILIPPINE SEA

ABSTRACT

We present charts of the residual magnetic field along ship tracks in the Philippine Sea. A separate chart shows major bathymetric features together with a summary of identifiable magnetic anomalies in the West Philippine Basin.

## INTRODUCTION

The Philippine Sea represents one of the largest marginal basins in the West Pacific and is one of the best understood examples of basin formation by inner arc spreading (Karig, 1971). However, no comprehensive collection of magnetic anomalies exists for this region. This report is a first attempt at such a compilation and, hopefully, will serve as a data base for future work. Parts of this data have been used by Loudon (in press) for his identification of magnetic anomalies in the West Philippine Basin.

## DATA SOURCES AND REDUCTION

The data presented in the charts come from seven sources and represent four basic types of data reduction. In all cases the regional field has been removed by identical methods to those of Sclater, et al. (1971).

### Type A

1. Lamont-Doherty Geological Observatory cruises: VEMA 19, 21 and 24; CONRAD 11 and 12.
2. Scripps Institution of Oceanography cruises: CIRCE, ANTIPODE, LUSIAD ARGO, and TASADAY.
3. Deep Sea Drilling Project: Leg 31 of the R/V GLOMAR CHALLENGER.
4. Ocean Research Institute of the University of Tokyo: cruises of the R/V UMITAKE MARU (1963-67) and R/V

HAKUHO MARU (1972).

For the above sources, cards or tapes of magnetics and bathymetry vs. time and time vs. position were obtained from the original data library. These were then merged at five minute intervals.

Type B - U. S. Naval Oceanographic Office

1. Project Magnet lines: 301, 302B, 304, 307, 311, 313B, 318, 326, 356, 371, 372A, 374, 385, and 395.

2. Leg 6 of the R/V RUTH ANN.

Analogue records of magnetics vs. time were obtained on microfilm from NAVOCEANO. These were photographically enlarged and then digitized. The digitized output was merged (at five minute intervals for ship tracks or twenty second intervals for airplane lines) with position vs. time data, read either from plots (R/V RUTH ANN) or listings (Project Magnet).

Type C

U. S. Coast and Geodetic Survey: International Indian Ocean Expedition (1964) of the R/V PIONEER.

This data was punched from merged listings of position, bathymetry and magnetics at five minute intervals (Coast and Geodetic Survey, 1965).

Type D

National Ocean and Atmospheric Administration:

Trans-Pacific Survey 7201.

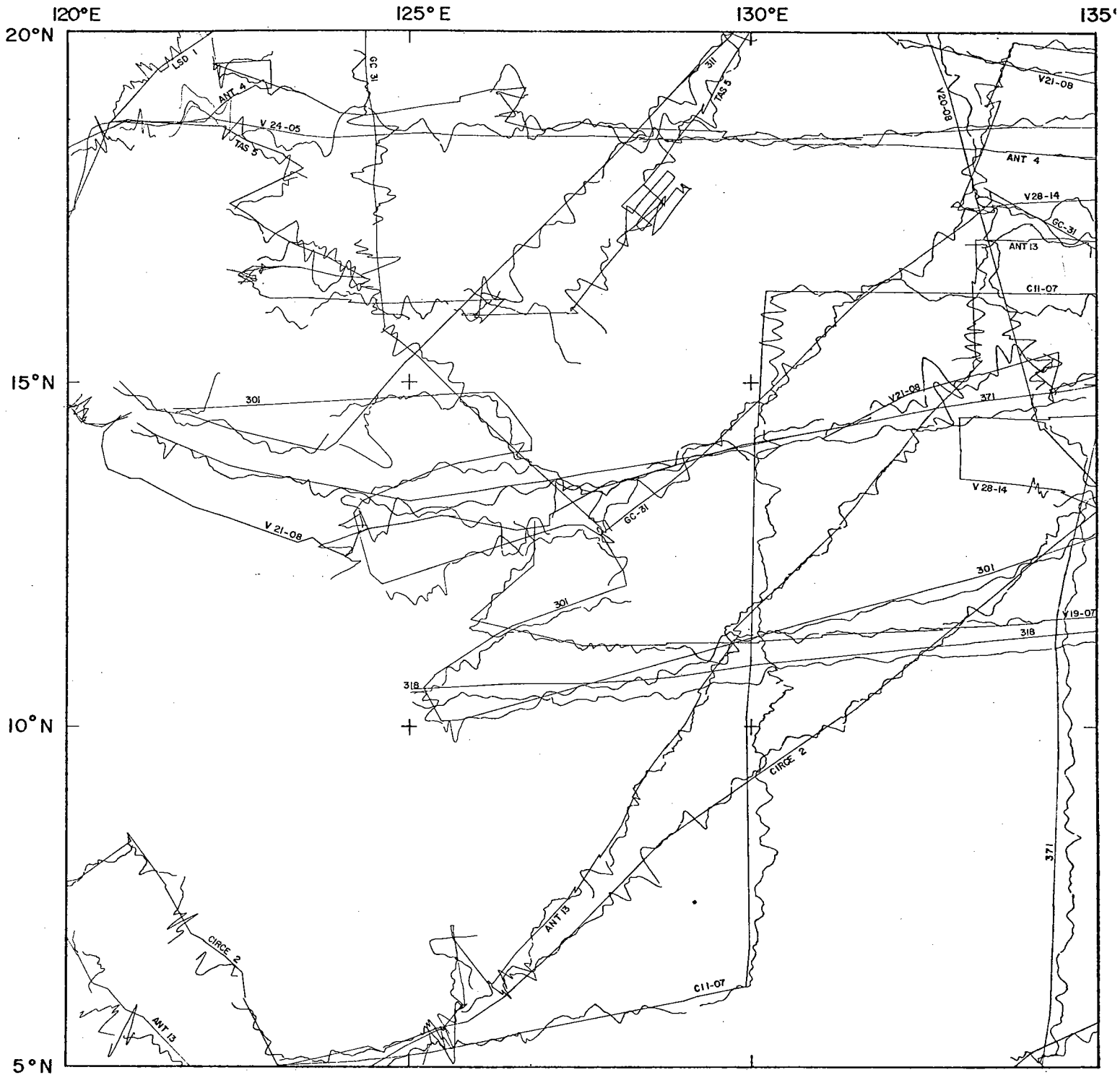
A tape of merged position, bathymetry, magnetics and gravity was obtained from NOAA. Readings were recorded every five minutes.

#### CHARTS

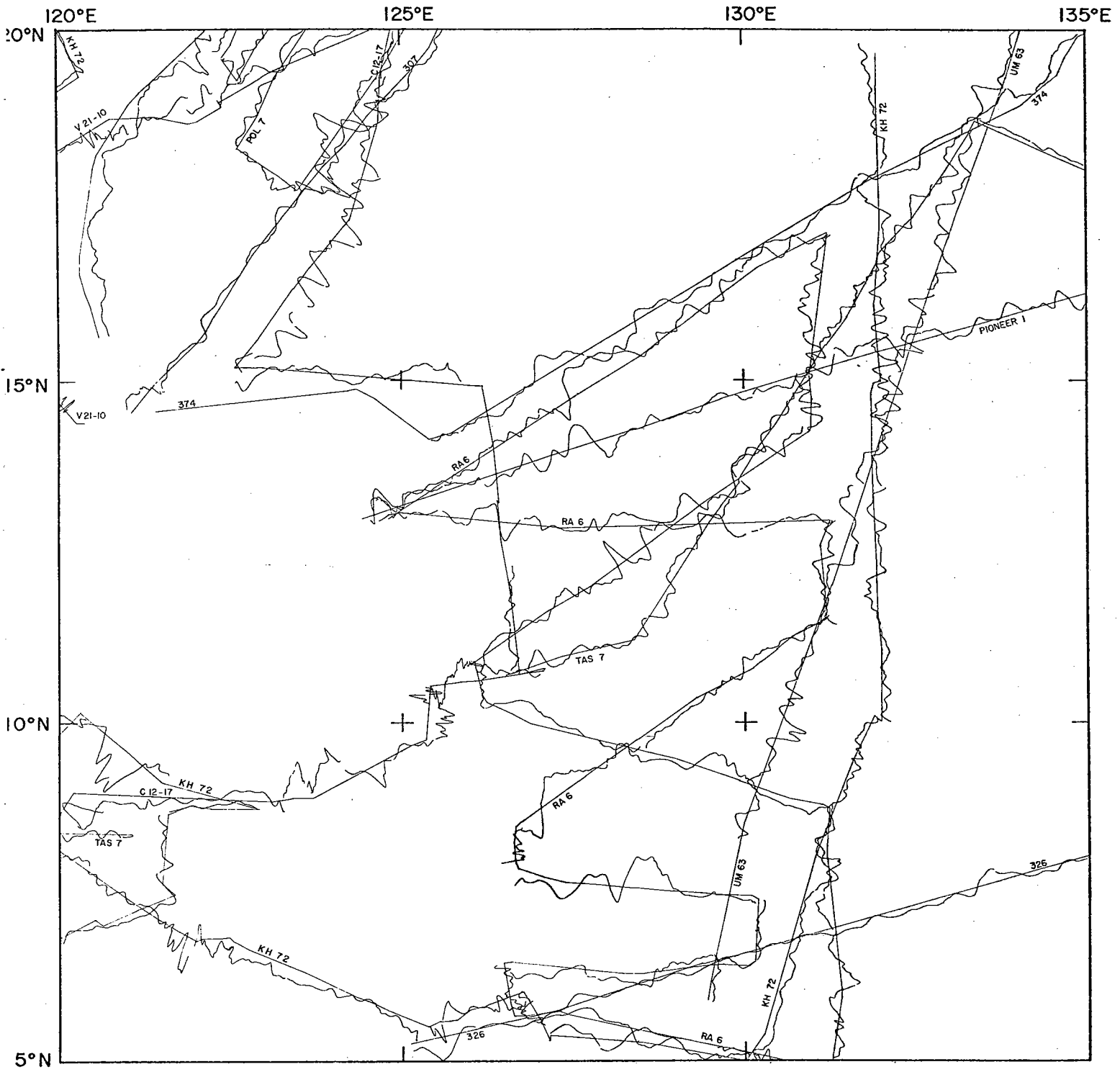
The above mentioned data was plotted at a scale of 1.8 inch/deg longitude and 333 gammas/inch in figures 1-8. In all cases positive anomalies are plotted above the ship track. On a North-South line they are plotted to the left.

FIGURES 1-8

Magnetic anomalies in the Philippine Sea plotted at right angles to track. Positive anomalies are plotted above the ship track except for a North-South line when they are plotted to the left.

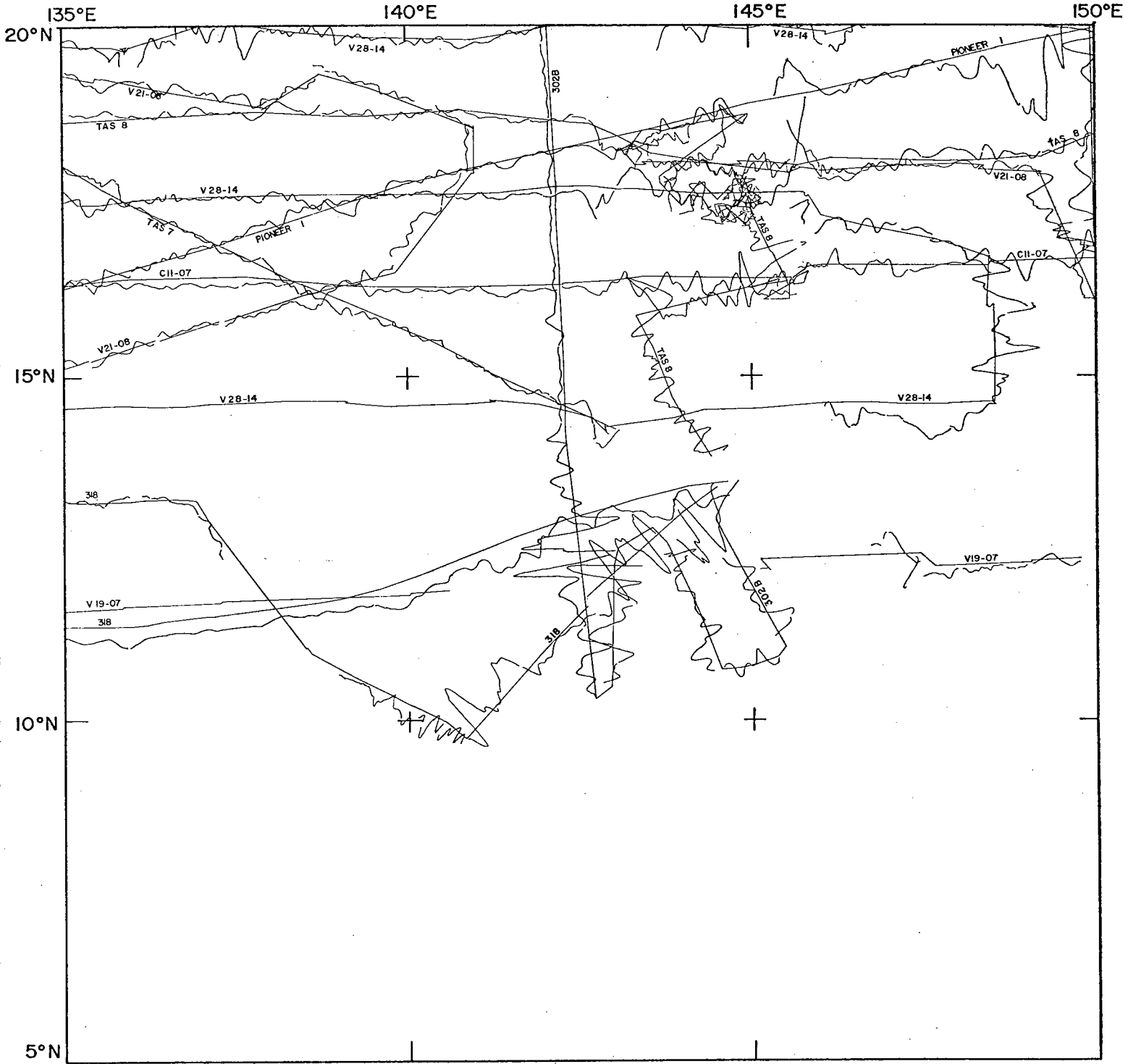


SCLATER & LOUDEN: FIG. 1A  
 MIT-0-75-160/PT 1-8-75

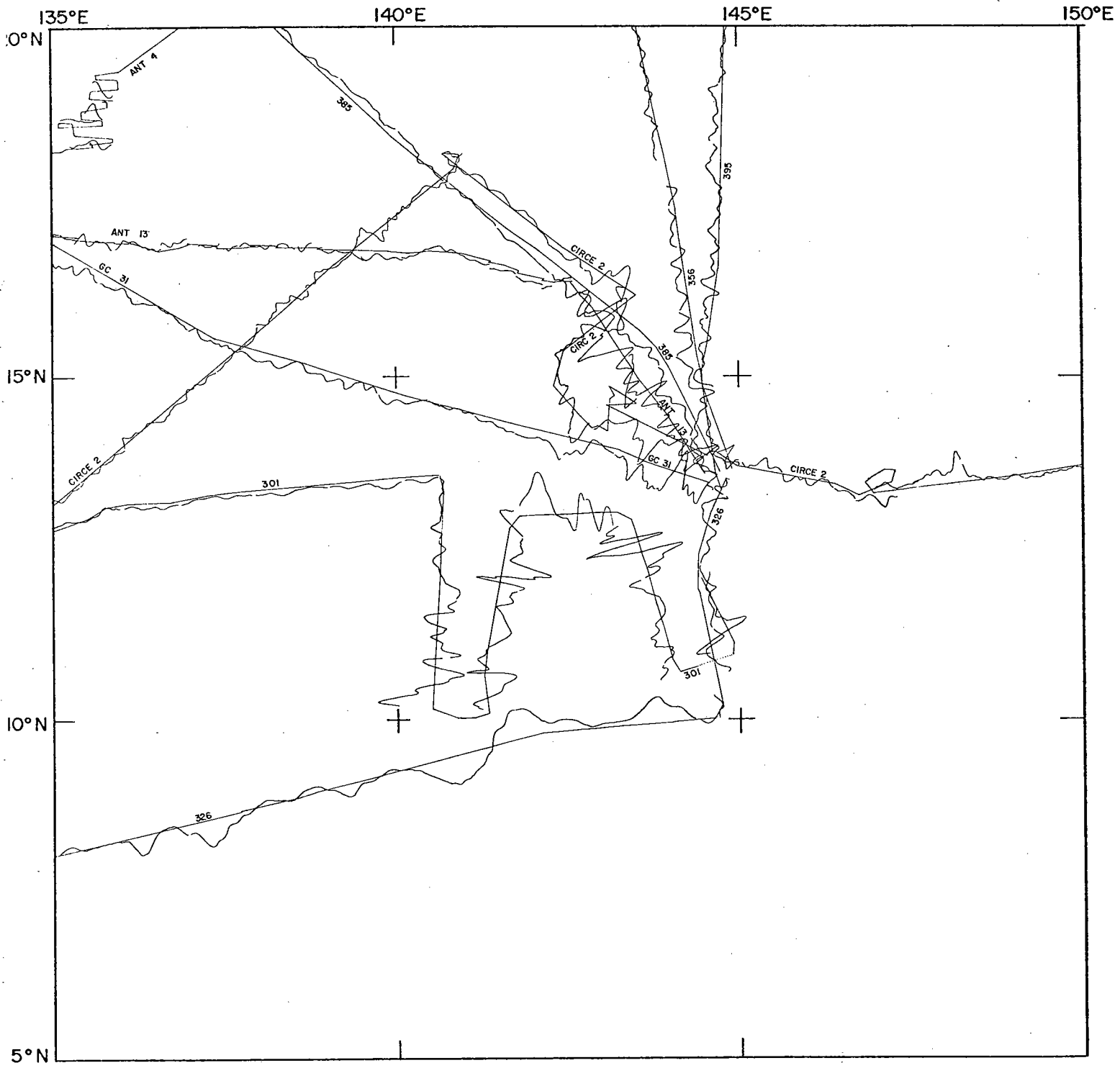


SCLATER & LOUDEN : FIG. 1B  
 MIT-O-75-161/PT 1-8-75

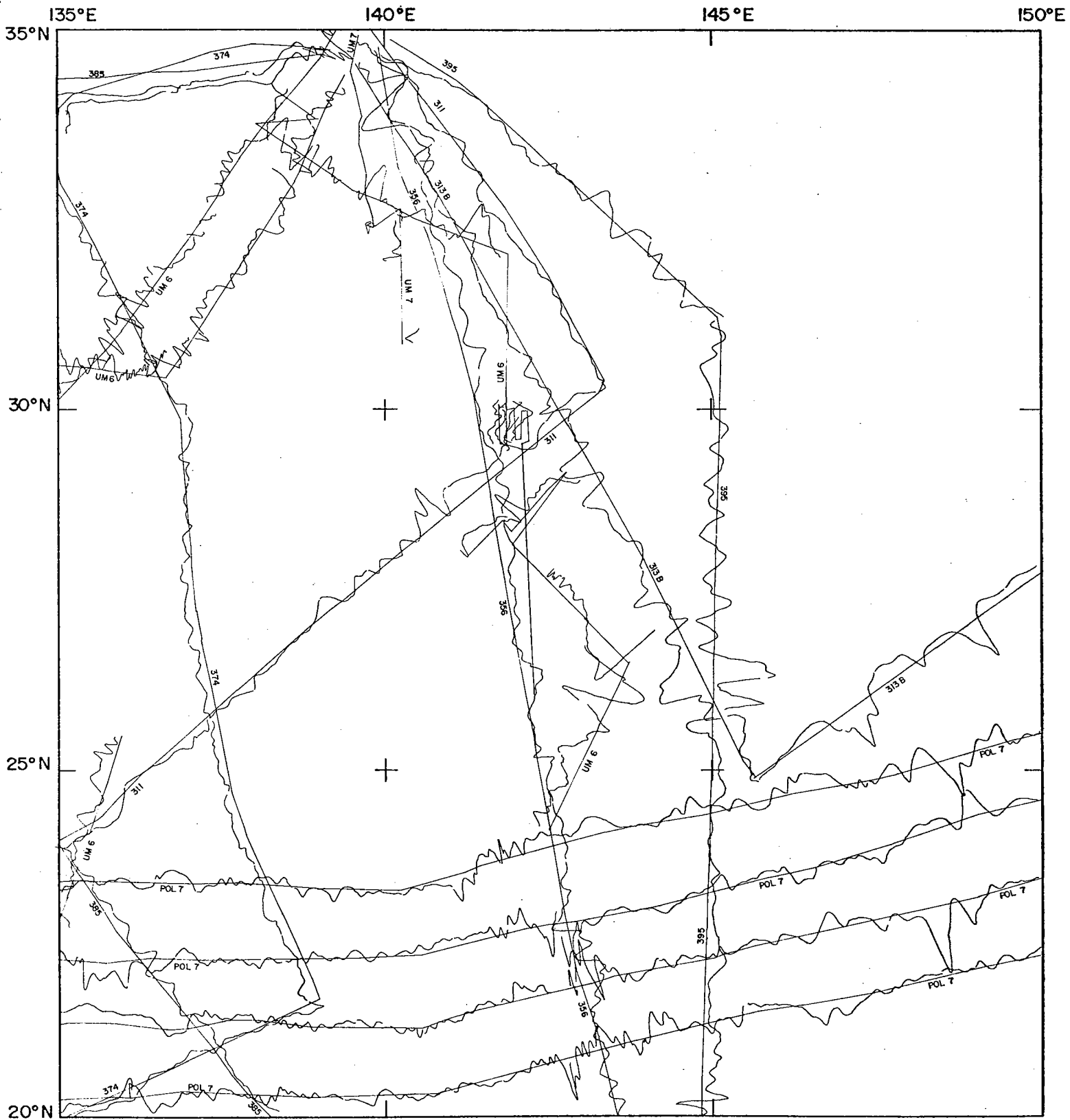




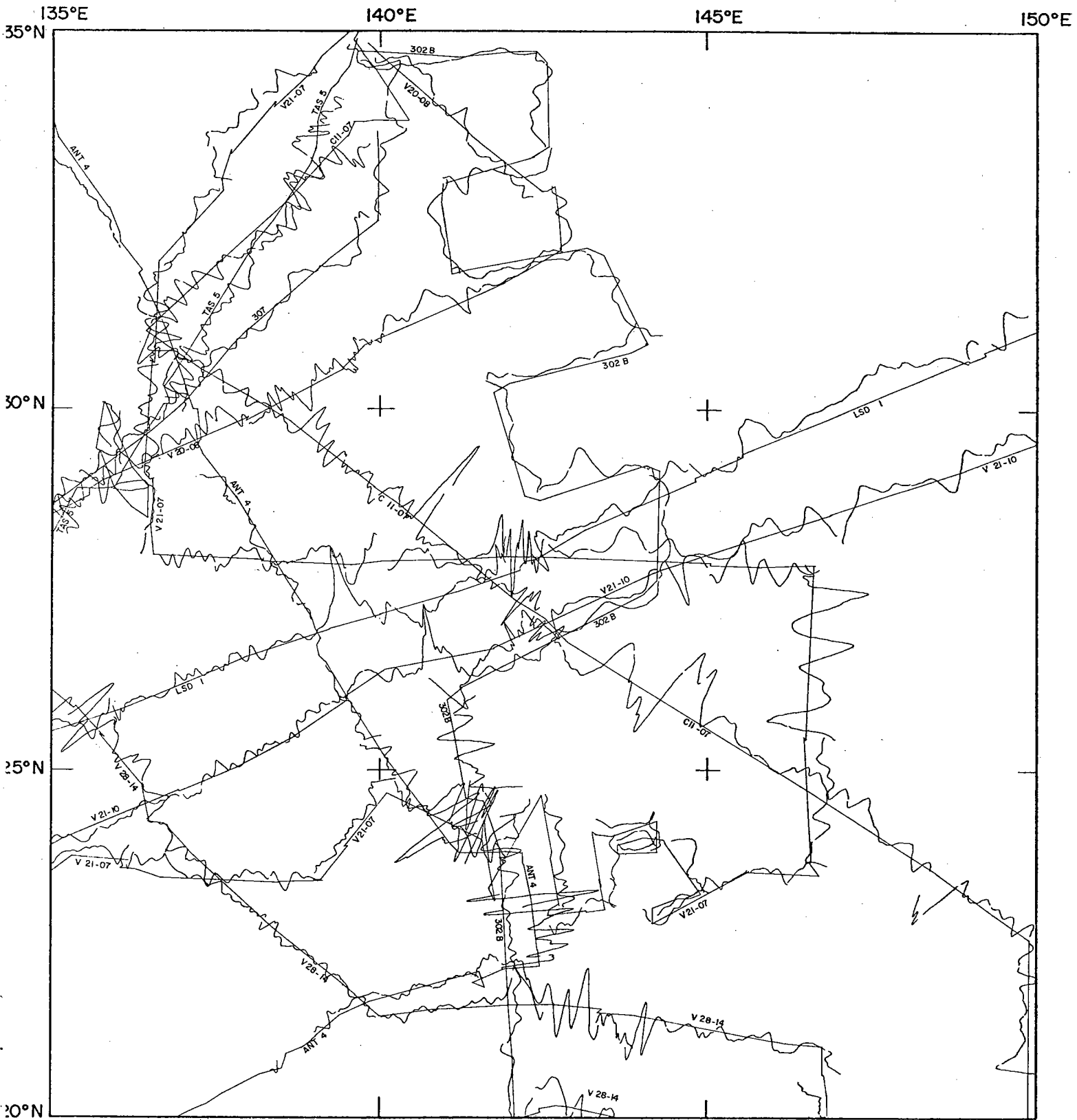
SCLATER & LOUDEN: FIG. 2A  
MIT-O-75-162/PT 1-8-75



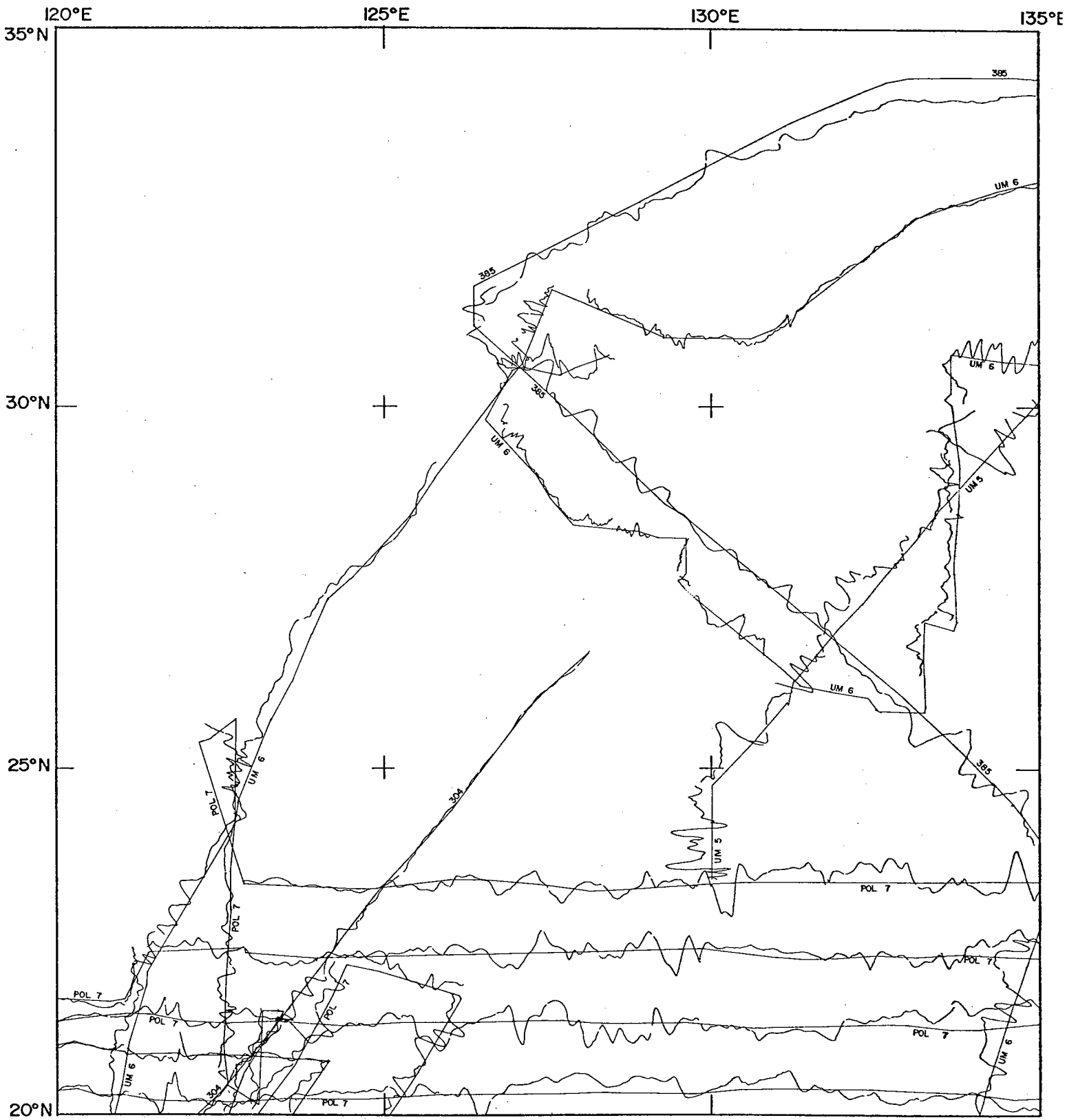
SCLATER & LOUDEN : FIG. 2B  
MIT-O-75-163/PT 1-8-75



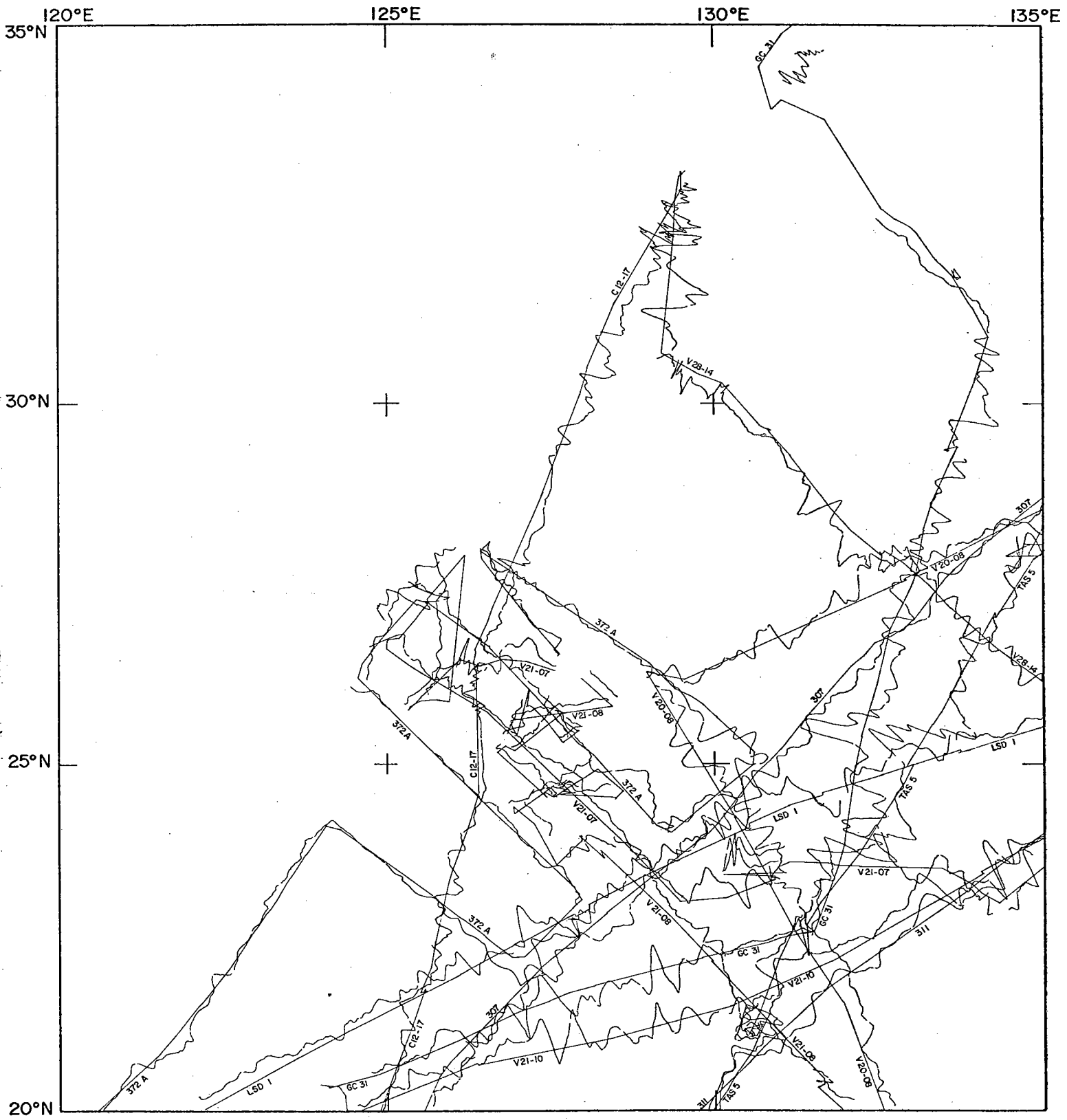
SCLATER & LOUDEN: FIG. 3A  
MIT-O-75-164/PT 1-8-75



SCLATER & LOUDEN: FIG. 3B  
 MIT-O-75-165/PT 1-8-75



SCLATER & LOUDEN · FIG. 4A  
 MIT-0-75-166/PT 1-8-75



SCLATER & LOUDEN: FIG. 4B  
 MIT-0-75-167/PT 1-8-75

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