Title: MAGNETIC AND GRAVITY INTERPRETATION OF YALOC-69 DATA FROM THE COCOS PLATE AREA

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Magnetic, gravity and bathymetry data were collected on an extended cruise of the R/V Yaquina in 1969. The last set of data was obtained from those track lines leaving the Panama Basin. The area covered is mainly the Cocos plate (Molnar and Sykes, 1969). The data is analyzed and compared with results of previous workers and the geophysical implications considered.

Generally speaking, from the magnetic part of the data, both direct and indirect methods show support of Vine and Matthew's (1963) hypothesis of sea-floor spreading and the subsequent principles of new global tectonics. The most northern magnetic anomaly profile across the East Pacific rise (at 18.3°N) shows a spreading rate about 3 cm/yr. and the most southern one (at 12.8°N) shows a rate about 5.2 cm/yr. The Cocos plate has been assumed to move in a northeast-southwest direction (N30°E to N45°E), and rotate with respect to the
Pacific plate about a pole at $40^\circ$N, $110^\circ$W with an angular velocity of $19.6 \times 10^{-7}$ deg./yr. (Larson and Chase, 1970). New material comes up from the west boundary - the East Pacific rise, and the south boundary - the Galapagos rift, causing the Cocos plate to underthrust the Americas plate at the middle American arc. Some of the points of new global tectonics can not be detected from this set of data; further detailed study of more track lines and sea bottom core samples are needed. The results of both analytical methods for determining the magnitude of induced and remanent magnetization in the second layer shows some consistence with the work of Schaeffer and Schwarz (1970), and Irving et al. (1970) at the Mid-Atlantic ridge near $45^\circ$N, in which a thinner magnetization layer at the ridge and the attenuation of magnetization intensity away from the ridge axis are suggested. Free-air gravity anomaly profiles have been employed to determine the crustal structure of two sections; a ridge section at $12.8^\circ$N and a trench section at $14^\circ$N. For the ridge section, if the anomalous mantle was converted from normal mantle, the extension of anomalous mantle into the normal mantle requires some uplift and/or lateral expansion in the rise crest area. The tensional configuration suggested in the trench crustal section agrees with the model proposed by Elsasser (1968) for the differential movement between two lithospheric blocks.

This work gives some speculations that evidence which supports the present new global tectonics theory is limited to a certain degree
of accuracy. Further study of the theory based upon physics, its mechanism, and measurement techniques that would give more reliable evidence have to be developed before it can be ascertained what really happens beneath this wild, wild world.
Magnetic and Gravity Interpretation of Yaloc-69 Data from the Cocos Plate Area

by

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# TABLE OF CONTENTS

**INTRODUCTION**

<table>
<thead>
<tr>
<th>Topic</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sea-Floor Structure and Continental Drift Hypothesis -- A Synthesis</td>
<td>1</td>
</tr>
<tr>
<td>Description of Cocos Plate and Regional Marine Geology</td>
<td>7</td>
</tr>
</tbody>
</table>

**MAGNETICS**

<table>
<thead>
<tr>
<th>Topic</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Anomalous Field Introduction</td>
<td>14</td>
</tr>
<tr>
<td>Data and Its Reduction</td>
<td>14</td>
</tr>
<tr>
<td>Anomaly Working Equations, Derivation of Applicable Computer Programs</td>
<td>18</td>
</tr>
<tr>
<td>Model Calculation - Ridge Sections</td>
<td>25</td>
</tr>
<tr>
<td>Indirect Method</td>
<td>25</td>
</tr>
<tr>
<td>Direct Method</td>
<td>32</td>
</tr>
<tr>
<td>Comparison and Evaluation of Methods</td>
<td>50</td>
</tr>
<tr>
<td>Model Calculation - Trench Sections</td>
<td>51</td>
</tr>
<tr>
<td>Indirect Method</td>
<td>51</td>
</tr>
<tr>
<td>Conclusion</td>
<td>61</td>
</tr>
</tbody>
</table>

**GRAVITY**

<table>
<thead>
<tr>
<th>Topic</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Anomalous Field Introduction</td>
<td>62</td>
</tr>
<tr>
<td>Data and Its Reduction</td>
<td>62</td>
</tr>
<tr>
<td>Anomaly Working Equation</td>
<td>64</td>
</tr>
<tr>
<td>Free-Air Anomaly Interpretation</td>
<td>67</td>
</tr>
<tr>
<td>Crustal Sections</td>
<td>71</td>
</tr>
<tr>
<td>Introduction</td>
<td>71</td>
</tr>
<tr>
<td>Ridge Section and Its Geophysical Implication</td>
<td>73</td>
</tr>
<tr>
<td>Trench Section and Its Geophysical Implication</td>
<td>79</td>
</tr>
</tbody>
</table>

**CONCLUSIONS AND DISCUSSION**                                              | 83   |

**BIBLIOGRAPHY**                                                            | 86   |

**APPENDIX**                                                               | 94   |
<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Boundaries and relative motion of lithospheric plates (adopted from Molnar and Sykes, 1969).</td>
<td>9</td>
</tr>
<tr>
<td>2</td>
<td>The Cocos plate and its boundaries with the Americas and Caribbean plates showing bathymetric contours and ship track lines.</td>
<td>12</td>
</tr>
<tr>
<td>3</td>
<td>Magnetic, gravity and bathymetric profiles 2-3 and 4-5 over East Pacific rise.</td>
<td>26</td>
</tr>
<tr>
<td>4</td>
<td>Magnetic, gravity and bathymetric profiles 1-2 and 4-3.</td>
<td>27</td>
</tr>
<tr>
<td>5</td>
<td>Magnetic profiles across ridge area. Theoretical profiles calculated from the models of Heirtzler et al. (1968) and Cox (1969) time scale of normally (black) and reversely (white) magnetized blocks.</td>
<td>30</td>
</tr>
<tr>
<td>6a</td>
<td>Parameters used in the working model and computer program (from Emilia, 1969).</td>
<td>33</td>
</tr>
<tr>
<td>6b</td>
<td>Parameters describing the direction of the regional geomagnetic field and the direction of the magnetization (from Emilia, 1969).</td>
<td>33</td>
</tr>
<tr>
<td>7</td>
<td>Results of application of the direct method to the 2-3 magnetic anomaly profile.</td>
<td>34</td>
</tr>
<tr>
<td>8</td>
<td>Results of application of the direct method to the 4-5 magnetic anomaly profile.</td>
<td>35</td>
</tr>
<tr>
<td>9</td>
<td>Spreading rate curve and associated least square fit of the magnetic profile shown in Figure 7.</td>
<td>44</td>
</tr>
<tr>
<td>10</td>
<td>Spreading rate curve and associated least square fit of the magnetic profile shown in Figure 8.</td>
<td>45</td>
</tr>
<tr>
<td>11</td>
<td>Projected magnetic, gravity, and bathymetric profiles 5-6 to 11-S over Middle America trench.</td>
<td>52</td>
</tr>
<tr>
<td>Figure</td>
<td>Description</td>
<td>Page</td>
</tr>
<tr>
<td>--------</td>
<td>-----------------------------------------------------------------------------</td>
<td>------</td>
</tr>
<tr>
<td>12</td>
<td>Magnetic anomaly profiles 7-8 and 9-8 across trench area compared with theoretical calculations by indirect method.</td>
<td>53</td>
</tr>
<tr>
<td>13</td>
<td>Deduced crustal section along the track line 4-5 by seismic control (from Raitt, 1956) and observed gravity anomaly.</td>
<td>77</td>
</tr>
<tr>
<td>14</td>
<td>Deduced crustal section along the track line 9-8 by seismic control (from Shor and Fisher, 1969) and observed gravity anomaly.</td>
<td>80</td>
</tr>
<tr>
<td>Table</td>
<td>Description</td>
<td>Page</td>
</tr>
<tr>
<td>-------</td>
<td>-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------</td>
<td>------</td>
</tr>
<tr>
<td>1</td>
<td>Mass per unit area to the depth of compensation from seismic control and from layering adjusted by gravity anomaly over ridge and trench area. (Refer to Figure 13 and 14 for station number and locations.)</td>
<td>74</td>
</tr>
</tbody>
</table>
MAGNETIC AND GRAVITY INTERPRETATION OF
YALOC-69 DATA FROM THE COCOS
PLATE AREA

INTRODUCTION

Sea-Floor Structure and Continental Drift Hypothesis -- A Synthesis

A large amount of supporting evidence has been provided since Wegener (1929) hypothesized that the present continents split from one super continent and subsequently moved to their present positions. The hypothesis that the present continents were once part of a single continental mass is still somewhat controversial. However, the second conjecture, that the continents have "drifted", is almost universally accepted. Early arguments in favor of continental drift were based primarily on geometrical fit, paleomagnetic poles, geological structure, and biological similarities (e.g., Runcorn, 1966) and were not generally accepted. The discovery of the correlation of the sea-floor magnetic anomaly pattern with the paleomagnetic time scale led to belief. In 1963, Vine and Matthews hypothesized that the sea floor is forced apart at oceanic ridges by upward rising large thermal convection currents. New material (basalt) rises to the surface, cools down through the Curie temperature and acquires a TRM (thermal remanent magnetization) in the direction of the ambient geomagnetic field. Due to changes in polarity of the magnetic field and transportation of material away from the axis, alternate bands of normal and reversely magnetized material are produced, which in turn generate
positive and negative magnetic anomalies. As the material moves away from the axis on both sides, a symmetric linear magnetic anomaly pattern is generated parallel to the ridge. Using K-Ar dates on lava flows with measured magnetic polarities, paleomagnetic time scales were constructed for about the last 4.5 million years (Doell et al., 1966; Doell and Dalrymple, 1966). Vine (1966), and many later papers, demonstrated a strong correlation between the observed patterns of oceanic magnetic anomalies and the polarity changes of the geomagnetic field. Heirtzler et al. (1968) extrapolated the existing time scale to about 80 m.y. ago by assuming constant spreading rates and using the magnetic anomaly patterns from the South Atlantic as a standard. From his absolute time scale and anomaly pattern, Heirtzler found the pattern of opening of the oceans is in good agreement with continental drift, in particular with the history of the breakup of Gondwanaland.

Dietz (1961) and Hess (1962) first proposed the concept of seafloor spreading. Following Vine, Matthews, and Heirtzler's analyses of magnetic data, Morgan (1968) and Le Pichon (1968) made one further advancement to the understanding of continental drift - the new global tectonics. They suggested that the surface of the earth can be approximated by a small number of rigid blocks in relative motion with respect to each other. Morgan derived plate theory independently of Le Pichon; however Le Pichon's treatment is more generalized and
global. Morgan assumed 20 blocks and did not use statistical methods to analyze the data. Le Pichon (1968) used six big rigid plates (lithosphere plates) and determined possible centers of rotation for the plates through a least square fit of spreading rates from oceanic rises and azimuths of transform faults at the intersections with ridge axes. From this pattern of global surface motion Le Pichon reconstructed the history of sea-floor spreading during late Cenozoic, early Cenozoic and late Mesozoic times. This history shows support of the sediment distribution because of episodic spreading of Ewing and Ewing (1967). Isacks et al. (1968) provided seismic evidence, earthquake epicenter and focal mechanism studies, to support the new tectonics suggested by a synthesis of continental drift, sea-floor spreading, transform faults, and underthrusting of the lithosphere at island arcs. They suggested the mechanism of sinking of the whole lithosphere which retains its identity to a depth as great as 700 km at trenches because of thermal inertia. Furthermore, the lithosphere (tectosphere) cannot penetrate deeper because of the stiffness of the mantle under the continents. Spreading at the mid-ocean ridges which corresponds to separating of two plates in response to the stress pattern will stop when the sinking edge of the plate reaches 700 km. A readjustment of the stress patterns will initiate a new spreading cycle and a new rise-trench system will be formed. If the rate of underthrusting of the oceanic crust is of the order of 6 cm/yr., the average length of the
active part of a cycle of spreading is unlikely to exceed 30 m.y., which corresponds to 1800 km of crust. This length roughly corresponds to the length of a Gutenberg fault zone having a dip of 30° and extending to a depth of 800 km. This process is referred to as the episodicity of sea floor spreading. As the spreading rates increase, convergence of blocks shows compression features on the surface until a certain maximum is reached. At spreading rates greater than 5-6 cm/yr., lithosphere sinks on its own along an active trench and the associated surface features are tensional. This is consistent with Le Pichon's (1968) results of narrow range (6-9 cm/yr) of rates of differential movement associated with trenches where active thrusting of tectosphere occurs.

Although the pattern of surface motions and deformations is reasonably well known, the mechanism by which mechanical energy for the surface motion is provided is still unsolved. The majority of investigators favor some type of mantle thermal convection currents. However, Knopoff (1964) and MacDonald (1965) argue against the possibility of the existence of large scale convection current cells using various reasons such as the characteristics of the Bullen C zone (rigidity, inhomogeneity, phase transition or chemical inhomogeneity), heat flow and gravity data. The pattern, size and location of the convection cells are still problems to be solved before the validity of thermal convection currents as the driving mechanism can be determined.
McKenzie (1969) does show that it is possible to combine the plate theory and convection current hypothesis. He uses the lithosphere as the boundary condition of a convection cell, calculates the temperature distribution in the lithosphere thrust beneath island arcs, and determines the flow and stress elsewhere in the mantle. He then demonstrates that earthquakes are restricted to those regions of the mantle which are colder than a particular temperature (about 680° C). The flow and stress heating in the mantle can maintain the high heat flow anomalies observed behind island arcs. Convection is controlled by the input of cold material (sinking lithosphere), which in turn causes general downward movements in the mantle near island arcs. Input of cold lithosphere must cease when the island arc tries to consume a continent, since the light continental crust cannot sink through the denser mantle. Attempts to assimilate continental crust in this way can produce fold mountains, and also permit a rearrangement of convection cells. The cold lithosphere input and the convection currents are the cause and effect of each other.

The ideas of a minority group, the expansionists, cannot be excluded as possibilities for the driving mechanism. Egyed (1956), Carey (1958), Heezen (1959), and Jordan (1962), all proposed the possibility that continental separation was caused by an expansion of the interior of the earth. In 1933, Hilgenberg suggested that the original earth crust split apart due to expansion of the earth's interior,
and the ocean formed in the void between the continents. A decrease of the gravitational constant was considered as the cause of expansion by some geophysicists, while others, e.g., Dirac (1937), suggested other reasons. Meservey (1969) reported a strong argument that there is no topologically possible transformation of the continents on an earth of the present size from their present position to even approximately the position of the universally accepted reconstruction. He introduced constraints on the perimeter of the Pacific. From modern knowledge of sea floor spreading, it is quite unambiguous that the links which connect the circum Pacific continents and the area enclosed by the perimeter decrease as we go backward in time, at least from the present to 80 million years ago. However, from the equi-distant projection of the earth's surface (with origin of the projection in the central Pacific on equator at 165° W), if the reconstruction due to Bullard and Hurley is to be achieved, the perimeter of the Pacific must have at some time enclosed at least half of the earth's surface in order for it to pass over the earth's circumference (going back in time) and be assembled on the opposite side of the earth. That is topologically impossible. The author suggested that the expansion hypothesis can resolve the problem. Both possibilities of the mechanism mentioned above still have some problems unsolved. In addition to the expansion hypothesis, other alternatives to the plate hypothesis have been suggested. van Andel (1968) suggested that the
magnetic lineations may result by arching of a pre-existing crust from thermal expansion or hydration of the underlying mantle. A combination of tensional fractures together with a gradually contracting volcanic zone (model 2 in his paper) could yield a symmetric pattern of positive and negative magnetic anomalies. It must be concluded that the driving mechanism of sea-floor spreading, plate tectonics, and continental drift is still undecided.

**Descriptions of Cocos Plate and Regional Marine Geology**

Le Pichon (1968), investigated the relative motion of the Pacific plate and the Americas plate together with the absorption by the Caribbean block of whatever differential movement may exist between the North and South American blocks. He postulated a filled Panama trench, similar to a filled trench east of Juan de Fuca ridge proposed by Ewing et al., (1968), and suggested that these areas are inactive sections of a once continuous trench system along the west coast of North America. They were abandoned and filled at the time of the change in the pattern of spreading between 10 and 5 m. y. ago (Vine, 1966). Spreading along the Galapagos rift zone and East Pacific rise is combined to produce movement in a southwest-northeast direction. He found some confusion and discrepancies in the relative plate motion for his reconstruction of the global surface pattern in the mid-America area but did not suggest any further explanation. Molnar and Sykes
(1969) redetermined more than 600 hypocenters in the Middle Ameri-
can region. They defined, and named, two small plates by epicenter
locations and earthquake focal mechanisms together with an evaluation
of previous work. The Cocos plate is bordered by the East Pacific
rise, the Galapagos rift zone, the north-trending Panama fracture
zone near 82° W, and the Middle American arc; the Caribbean plate
underlies the Caribbean Sea and is bounded by the Middle American
arc, the Cayman trough, the West Indies arc, and the seismic zone
through northern South America (Figure 1). From focal mechanisms
of 70 earthquakes in these regions, they have ascertained the relative
motion of these two plates with respect to the surrounding regions or
plates. Their results do not conflict with those prior geologic and
geophysical studies of this area (Hess, 1933, 1938; Bucher, 1952;
Hess and Maxwell, 1953; Gunn, 1943; and Wilson, 1966).

The motion of the Cocos plate can be summarized as follows:
a. spreading from the East Pacific rise and at the Galapagos
   rift zone,
b. underthrusting at the Mid-American trench off Mexico and
   Guatemala in a north-easterly direction and in a more
   north-northeasterly direction off the rest of Central Mexico
   north of the Tehuantepec ridge.
c. right-lateral strike-slip motion along the Panama fracture
   zone (a transition fault connecting the Galapagos rift zone
Figure 1. Boundaries and relative motion of lithospheric plates adopted from Molnar and Sykes, 1969). Names of plates are given by large letters, while the geological structures that define plate boundaries are denoted by smaller letters. Dashed lines across the rise show fracture zones, others show unclearly defined plate boundaries.
and the Middle American arc).

The motion of the Caribbean Plate can be summarized as follows:

a. is moving easterly with respect to the Americas plate, which is here taken to include both North and South America and the Western Atlantic.

b. It is being underthrust by the Americas Plate in a westerly direction at the Lesser Antilles and near Puerto Rico. The relative motion is perpendicular to the trend of the Lesser Antilles. Unlike the Lesser Antilles, however, motion at present along the Puerto Rico trench is almost parallel to the trench on nearly horizontal fault planes.

c. undergoes left-lateral strike-slip motion along steeply dipping fault planes on the Cayman trough.

Computation of rates of motion indicates that underthrusting of the Cocos plate is proceeding at a higher rate in southeastern Mexico and Guatemala than in western Mexico and that the Caribbean plate is moving at a lower rate relative to North America than is the Cocos plate (Molnar and Sykes, 1969).

The Cocos plate is just on the continental side of the Andesite Line (Menard, 1964; MacDonald, 1949), which is at the edge of the basin formed by the Tertiary folds and the East Pacific rise, and separates the region of olivine basalt lavas from the surrounding zone of andesite and acid and calc-alkali lavas. Except for shocks
associated with volcanic activity, no well defined epicenters are known within the Pacific Basin. The interior of the Cocos plate is aseismic in contrast to the high degree of seismicity on the edge of the plate. This implies that the Cocos plate may indeed be approximated as a rigid plate, which undergoes very little deformation except along its edges. This high activity along the edges presumably results from an interaction of the Cocos plate with the other rigid plates (Figure 1). There are no abyssal hills and archipelagic aprons in this area.

The plate topography includes two ridges (the Cocos and Tehuantepec) and scattered seamounts (Figure 2). The Clarion and Clipperton fracture zones approximate the boundary of the Middle America Trench (Menard, 1955), and a profound change in the topography, trend, and crustal structure of the trench occurs where the Tehuantepec ridge intersects it (Menard and Fisher, 1958; Fisher, 1961). Northwest of the Tehuantepec ridge the trench is characterized by a U-shape cross section, a steeper shoreward flank and a flat bottom suggesting sedimentary fill. The trench shoals west of the Gulf of Tehuantepec, deepens abruptly to a maximum off western Guatemala and shoals gradually again to merge into the sea floor off Costa Rica. Southeast of the Tehuantepec ridge the trench is V-shaped in cross section with an irregular bottom in the Guatemala basin area. A thicker crustal layer and little sedimentary fill (which suggests a younger age),
Figure 2. The Cocos plate and its boundaries with the Americas and Caribbean plates showing bathymetry contours and ship track lines.
borders a volcanically active coast (Fisher, 1961).

Menard (1964) proposed the Cocos and Tehuantepec ridges formed as follows: first, oceanic crust is pulled against the continent by a sharp bend in the coast and with differential movement on an east-west fracture zone; second, the angle in the coast produces cracks in the oceanic crust, the volcanic ridge develops along the cracks; and third, the crust southeast of the ridge is pulled away to form the downbuckled deep trench. The crustal blocks northwest of the ridge and south of the fracture zone are pulled perpendicular to the coast and arch upward, while the detached southeast block is passive. There are, however, other explanations for the origin of this kind of aseismic ridge. For example Wilson (1963) considered the source of lava flow of the Hawaiian Islands to be in the lower moving part of convection currents, deeper than that of the mid-ocean ridge. The differential motion carries old volcanos away from the source, while new volcanos form over the source. The length of the island chain, or aseismic ridge, depends on how long the source has been active. The origin of aseismic ridges is not settled and additional investigations are needed before we can decide what process is most likely.
MAGNETICS

Anomalous Field Introduction

Data and Its Reduction

On the return of the R/V Yaquina from the Yaloc-69 cruise to the Panama Basin - Galapagos Rift area, geophysical measurements were made over the Cocos plate and its boundaries with the Caribbean and Americas plates. Magnetic, gravity, and seismic reflection data has been collected. The track lines include eight crossings of the middle America trench and three of the East Pacific rise. Magnetic, gravity and bathymetric measurements, obtained on this leg of the cruise are presented and interpreted in this thesis. The area covered (Figure 2), from $9^\circ$ to $20^\circ$N, $85^\circ$W to $112^\circ$W, along the coastline is very large and the track lines are very sparse. The data collected is not extensive enough for detailed analysis of the whole Cocos plate and East Pacific rise. However, this study can draw some inferences about the structure of this area; based on the previous work and the data derived from these track lines.

Total magnetic field intensities were recorded continuously along track lines using a proton precession magnetometer constructed at O.S. U. (Emilia et al., 1968) patterned after a Packard-Varian unit. The magnetometer has a resolution of ±5 gammas. The total magnetic
field intensity value is recorded in masons (5.0 gamma, 1 gamma = \(10^{-5}\) oersted) by the magnetometer. Strip chart records were digitized every 5 minutes, together with all high and low peak values. After removal of the regional geomagnetic field the magnetic anomaly values in gammas were plotted against distance (units in nautical miles) by the plotter of the OS-3 system in the computer center of Oregon State University. Bathymetry data was digitized in fathoms, but has been converted to meters for the plots.

The regional magnetic field was removed in the following manner. Assuming that the earth is a sphere of radius \(a\) and that there is no magnetic material near the ground, the earth's magnetic field can be derived from a potential function \(V\) which satisfies Laplace's equation and can be represented as a series of spherical harmonics in the form

\[
V = \sum_{n=0}^{\infty} \sum_{m=0}^{\infty} P_n^m(\theta) \{ c_n^m (r/a)^n + c_n^m (a/r)^{n+1} \} A_n^m \cos m\phi \\
+ [s_n^m (r/a)^n + (1-s_n^m)(a/r)^{n+1}] B_n^m \sin m\phi
\]

Since it is a scalar potential (see next section about the anomaly working equation) \(\vec{F} = \vec{V} V\) is used to calculate the field intensity and the \(X, Y, Z\) components of points \((r, \theta, \phi)\) around the surface of the
earth.

\[ Z_{r=a} = (\frac{\partial V}{\partial r})_{r=a} = \sum_{n=0}^{\infty} \sum_{m=0}^{n} \frac{P_n^m(\cos \theta)}{c_n^m - (n+1)(1 - c_n^m)} [n c_n^m - (n+1)(1 - c_n^m)] \]

\[ A_n^m \cos m\phi + [n s_n^m - (n+1)(1 - s_n^m)] B_n^m \sin m\phi \]

\[ Y_{r=a} = (\frac{-\partial V}{r \sin \theta \partial \phi})_{r=a} = \frac{1}{\sin \theta} \sum_{n=0}^{\infty} \sum_{m=0}^{n} \frac{P_n^m}{m A_n^m \sin m\phi - m B_n^m \cos m\phi} \]

\[ X_{r=a} = (\frac{\partial V}{r \partial \theta})_{r=a} = \sum_{n=0}^{\infty} \sum_{m=0}^{n} \frac{d P_n^m}{d \theta} \{ A_n^m \cos m\phi + B_n^m \sin m\phi \} \]

Let \( \alpha_n^m = \{ n c_n^m - (n+1)(1 - c_n^m) \} A_n^m \) and \( \beta_n^m = \{ n s_n^m - (n+1)(1 - s_n^m) \} B_n^m \)

then we can write

\[ Z_{r=a} = \sum_{m=0}^{\infty} \{ a_m(\theta) \cos m\phi + b_m(\theta) \sin m\phi \} \]

with

\[ a_o(\theta) = \sum \gamma_n P_n(\theta), \quad a_m(\theta) = \sum \alpha_n^m P_n^m(\theta), \quad b_m(\theta) = \sum \beta_n^m P_n^m(\theta) \]

The coefficients \( \alpha_n^m, \beta_n^m, \gamma_n^m \), are determined from known \( a_m \)

values around the world by the least square method.

The coefficients from Cain and Cain, 1968; Cain, et al., 1968

were used with secular variation taken into consideration. After sub-
traction of these regional field values from the measured values the
anomaly values were obtained. All of these processes were handled by computer programs (Gemperle and Keeling, 1970). However, since the coefficient $\alpha_n^m$, $\beta_n^m$ are only taken to the eighth order ($n = 0 - 8$, $m = 0 - 8$) (Gemperle and Keeling, 1970), the deviation of $F$ from the observed magnetic field values may still be very large. The lengths of the track lines are only a fraction of the shortest wave length ($1/8$ of earth perimeter) of the harmonic representation. Therefore in the following section on direct and indirect methods, the anomalies used are the ones derived after subtraction of the average anomaly value from the anomalies obtained above. This method is also used by other workers (e.g., Chase et al., 1970), i.e., adjustment to an arbitrary base line. Nettleton's (1954) paper shows some of the ambiguity in this step:

... the regional is what you take out to make what's left look like the structure. A critical review of various methods for determining regional effects indicates that neither the geophysical systems nor the various numerical and mathematical systems provides a philosophically sound basis for separating a potential field into residual and regional components... former depends on judgment of the operator, and the latter on assumptions or empirical steps in the mathematical deviation of numerical factors or coefficients used in the calculation. It is concluded that the lack of an objective criterion for the isolation of anomalies is due to the inherent ambiguity in the source of potential fields, but that in spite of the empirical nature of any regional method, the proper application of a system appropriate to the particular problem encountered can give very useful results.

Because of the ambiguity in the removal of regional geomagnetic field, some apparent results in the following sections are not entirely in
accordance with those predicted by theory. After interpretation, it is seen that reasonable results are obtained.

Anomaly Working Equation, Derivation of Applicable Computer Programs

This section shows how the important equations are derived. For more details and rigorous discussion of magnetic anomaly calculations see Grant and West (1965) and other sources. We assume that source of the magnetic anomalies lies in the second oceanic layer and that the anomaly has a dipole character. For detailed analysis of the magnetic field produced by magnetized material, refer to 10-2 of Reitz and Milford (1967). Only the result is stated here. The magnetic field at a point \( 0 \) outside the magnetized body can be written as:

\[
\mathbf{B}(\mathbf{r}) = -\nabla A(\mathbf{r}) + 4\pi \mathbf{M}_l(\mathbf{r})
\]

For the present case, local magnetization \( M_l = 0 \) where

\[
A(\mathbf{r}) = \int_{V_0} \mathbf{m}(\mathbf{r}') \cdot \frac{(\mathbf{r} - \mathbf{r}')}{|\mathbf{r} - \mathbf{r}'|^3} \, dv'
\]  

(1)
is the scalar magnetic potential, \( \vec{B} \) is the magnetic induction and \( \vec{m} \) is the magnetic moment per unit volume. Notice that the formula for \( \vec{B} \) is derived from \( \vec{B} = \vec{\nabla} \times \vec{A}'(\vec{r}) \) where

\[
\vec{A}'(\vec{r}) = \int_{V_o} \vec{m}(\vec{r}') \times \frac{(\vec{r} - \vec{r}')}{|\vec{r} - \vec{r}'|^3} \, dv'
\]

is the vector magnetic potential.

Now, the magnetic intensity \( \vec{H} = \vec{B} \) at point \( \vec{r} \) due to magnetized material is

\[
\vec{H} = -\vec{\nabla} \vec{A} 
\]

Assuming that the anomaly field is much less than the regional geomagnetic field, then the component of the anomaly field measured by a total-field magnetometer is that which lies in the direction of the regional geomagnetic field.

\[
\text{anomaly } \Delta \vec{T}(\vec{r}) = -\vec{\nabla}_g \vec{A} \tag{2}
\]

\(\vec{\nabla}_g\): gradient in direction of geomagnetic field.

The solution to (2) is obtained using numerical method. The numerical representation of (2) can be written as:

\[
\mathbf{G} \mathbf{M} = \Delta \mathbf{T} \tag{3}
\]

\(\mathbf{G}\): Working equation, a \( n \times n \) matrix

\(\mathbf{M}\): Magnetization (vector of \( n \)-dimension)
\[ \Delta \mathbf{T} \]: Anomaly (vector of \( n \)-dimension)

The following sections derive indirect and direct method solutions to Equation (3).

**Indirect method:** (Talwani and Heirtzler, 1964). To calculate \( \Delta \mathbf{T} \), we must figure out each element of \( G \), i.e., \( \Delta \mathbf{T} \) due to each magnetized block of rock. Assume a two dimensional structure for each block with the \( y \) dimension perpendicular to \( x \) and \( z \) directions, extending from \(-\infty\) to \( \infty \). From (1) if \( \mathbf{m} = J \Delta x \Delta y \Delta z \),

\[
\mathbf{J} = J \mathbf{i}_x + J \mathbf{j}_y + J \mathbf{k}_z
\]

then

\[
A_s = \frac{\mathbf{m} \cdot \mathbf{R}}{R^3} = \frac{J x_j + J y_j + J z_j}{(x^2 + y^2 + z^2)} \Delta x \Delta y \Delta z.
\]

\[
A = \int \int_{x=-\infty}^{\infty} \int_{y=-\infty}^{\infty} \int_{z=-\infty}^{\infty} \frac{J x_j + J y_j + J z_j}{x^2 + y^2 + z^2} \, dx dy dz
\]

\[
= 2 \int \int_{x,z} \frac{J x_j + J z_j}{x^2 + z^2} \, dx dz
\]

\[
\therefore \quad \Delta \mathbf{T} = -\nabla A
\]
\[
\Delta T_v = -\frac{\partial A}{\partial z} = 2 \int_{z_1}^{z_2} \int_x^\infty \frac{2xzJ_x - J_z(x^2 - z^2)}{(x^2 + z^2)^2} \, dx \, dz
\]

\[
= 2 \int_{z_1}^{z_2} \frac{J_x - J_z}{x^2 + z^2} \, dz
\]

\[
= 2 \sin \phi \left[ J_x \{ (\theta_2 - \theta_1) \cos \phi + \sin \phi \log \frac{r_2}{r_1} \} \right.

\left. - J_z \{ (\theta_2 - \theta_1) \sin \phi - \cos \phi \log \frac{r_2}{r_1} \} \right]
\]

Similarly,

\[
\Delta T_h = -\frac{\partial A}{\partial x} = 2 \sin \phi \left[ J_x \{ (\theta_2 - \theta_1) \sin \phi - \cos \phi \log \frac{r_2}{r_1} \} \right.

\left. + J_z \{ (\theta_2 - \theta_1) \cos \phi + \sin \phi \log \frac{r_2}{r_1} \} \right]
\]

From the figure above, we see

\[
J_x = J \cos A \cos (C-B)
\]

\[
J_z = J \sin A
\]
If direction of \( J \) is in the direction of the total magnetic intensity then

\[
A = I \text{ (inclination)} \\
B = D \text{ (declination)} \\
J = KF \\
K = \text{susceptibility} \\
F = \text{total magnetic intensity}.
\]

The total intensity anomaly is given by

\[
\Delta T = \Delta T_v \sin I + \Delta T_h \cos I \cos (C-D) \tag{4}
\]

For different angles of \( A, B, \) and susceptibility \( K, \) (4) is used separately to calculate the anomaly caused by the induced (present) and remanent (ancient) fields. This is the basic equation used in the first computer program for the indirect method (Appendix I). The attached figure shows how the superposition of integrals from Equation (4) are used to compute the anomaly of a two dimensional block. The results from integration along the closed path \( mnpq \) cancel in the dotted and shaded areas and only the area inside \( mnpq \) contributes to the solution. A case for a rectangular block is as follows:
Insert \( \phi = 90^\circ \) in equations for \( \Delta T_v \) and \( \Delta T_h \) derived above, and obtain the following equations.

\[
\Delta T_v = J \log \frac{x_2 + z_2}{x_1 + z_1} - 2J z \left( \tan^{-1} \frac{z_2}{x_2} - \tan^{-1} \frac{z_1}{x_1} \right)
\]

\[
\Delta T_h = 2J \left( \tan^{-1} \frac{z_2}{x_2} - \tan^{-1} \frac{z_1}{x_1} \right) + J \log \frac{x_2^2 + z_2^2}{x_1^2 + z_1^2}
\]

Again for the total intensity anomaly,

\[
\Delta T = \Delta T_v \sin I + \Delta T_h \cos I \cos (C-D)
\]

This is the basic equation used in the second program of the indirect method. For more detailed demonstration and algorithms refer to Appendix I.


The purpose of the direct method is to find the magnetization distribution \( \vec{M} \) from \( \Delta \vec{T} \) which is the total intensity magnetic anomaly, i.e. \( \vec{M} = G^{-1} \Delta \vec{T} \). We must therefore find the inverse of the matrix \( G \) (ref. Equation 3). For a detailed discussion of the stability of solutions, refer to Emilia (1969). The solutions are not forced if the position of the center of each magnetic volume element is chosen to correspond closely to the position of a field point. This procedure is used in the following sections. The solution may be ill-conditioned if \( \text{Cond} (G) = \mu_1 / \mu_n \geq 1 \) where
\[ \mu_1 = \text{largest eigenvalue of } G. \]
\[ \mu_n = \text{smallest} \]

If \( \mu_n \) is very small it may cause an unstable case. However, in the present case of the spreading model Cond \((G)\) does not violate the stable condition. Equation (3) can be written in component form as

\[ \Delta T_j = \sum_{k=1}^{n} G_{jk} M_k \quad (j=1, \cdots, n) \quad (5) \]

This is a \( n \times n \) system of linear algebraic equations. The two computer programs used to solve Equation (5) are from Emilia (1969).

a. exact solution program (Gaussian reduction)
b. approximate solution program (iteration)

Reduced from (3),

\[ (I-L-U)\bar{M} = \bar{S} = \Delta \bar{T} \]

I = unit matrix

L = lower triangular matrix with zero elements along the diagonal

U = upper triangular matrix

\[ M_k^i = \sum_{j=1}^{k-1} C_{kj} M_j^{(i)} + \sum_{j=k}^{n} C_{kj} M_j^{(i-1)} + S_k \]

where \( C = L + U \) with \( \bar{M}^{(0)} = \bar{S} \), the first approximation. The two conditions used to terminate the iteration are:
\[(i) \quad ||M^{(k)} - M^{(k-1)}||_e < c \quad ||M^{(k-1)}||_e\]

where \( || \cdot ||_e \) means Euclidean norm.

\[(ii) \quad \max \sum_{j=1}^{n} G_{jk}M_k < E \quad (j = 1, \ldots , n)\]

\(E = \) preset physical limit

\(C = \) ambiguous constant determined mathematically.

Both \(a\) and \(b\) are to be used in the following section and some comparisons have been made. Some comparisons with the indirect method will also be included.

**Model Calculations - Ridge Sections**

**Indirect Method**

Profiles 4-5, 2-3, and 4-3 (Figures 3 and 4) are the three profiles obtained from the track lines across the East Pacific rise. The magnetic anomalies are employed for interpretation by indirect and direct methods. The block models drawn on the topography profiles are used in the direct method. The gravity anomalies will be used in the gravity section later.

The simplified program for rectangular blocks (Appendix I and page 23) was used to calculate theoretical profiles (Figure 5) for specified models. In these profiles, we assume an east-west spreading rate of 2.8 cm/yr. and 5.5 cm/yr. The geomagnetic time scale used to define the blocks of normal and reversed magnetic polarities
**Figure 3.** Magnetic, gravity and bathymetric profiles 2-3 and 4-5 over East Pacific Rise. See Figure 2 for locations. Block models drawn on topography profiles are used in the direct method calculation (refer to Figure 6, 7 and 8). Dashed line represents missing data.
Figure 4. Magnetic, gravity and bathymetric profiles 1-2 and 4-3. Profile 4-3 over East Pacific Rise. See Figure 2 in locations.
was adopted from Cox (1969) after 4.5 m.y. and from Heirtzler et al. (1968) from 4.5-76.33 m.y.

The depth range of the magnetized block is 3.3-5.0 km (Vine, 1966) with the following additional parameters (Appendix I):

original azimuth = $90^\circ$ (original magnetization vector, an axial dipole)
original latitude = $12.8^\circ$

(latitude of track line 4-5 in Figure 2)

present azimuth = $81^\circ$ (present geomagnetic field, inclined dipole, formula used, in Appendix I)
present latitude = $19.4^\circ$

The average thickness of the second layer of East Pacific rise is about 1 km (Menard, 1960) and varies from 2.8 to 4 km below the sea surface (see gravity section).

The center block is of double magnetization.

The results derived from the comparison of some observed profiles from track lines 1-3, 4-3, and 4-6 (Figure 2) with the theoretical profiles are as follows:

(i) The plotted profiles are theoretical sea floor spreading profiles of the East Pacific rise at $12.8^\circ$ latitude. The change in shape of theoretical profile from $12.8^\circ$ to $18.3^\circ$ latitude is minimal. The amplitude of the theoretical profile and its changes with the distance should not be considered very precise or important due to simplified assumption about the geomagnetic field and model body. What is
meaningful is the shape of the profile.

(ii) The three magnetic anomaly profiles that cross the East Pacific rise are plotted in Figure 5. Referring to Figure 2 and 5, the 1-2 profile has been projected (through $50^\circ$) to azimuth $270^\circ$, parallel to the 2-3 and 4-5 profiles; the 5-6 profile has also been projected (through $56^\circ$) to azimuth $270^\circ$; the 4-3 profile has been projected through $66^\circ$. For the projected sections, only a rough comparison can be made. Since distance has been shortened the anomalies have been squeezed and the frequency of oscillation increased; however, the true, non-projected case would not oscillate so much.

(iii) The comparison of these observed profiles with the theoretical ones suggests that the spreading rate is approximately 3 cm/yr at the northern profile 2-3 and 5.5 cm/yr at the southern profile 4-5. This is consistent with results of Larson and Chase (1970). It is difficult to determine from this comparison the variations in spreading rate from the ridge; however, they are not large.

(iv) Although the theoretical profile assumes east-west spreading, the good correlation does not imply such a spreading direction since a little change in spreading angle would not affect the shape of the anomalies detected.
Figure 5. Magnetic anomaly profiles across ridge area. Theoretical profiles calculated from the models shown below of Heirtzler et al. (1968) and Cox's (1969) paleomagnetic time scale of normally (black) and reversely (white) magnetized blocks assuming constant rate of spreading. Number marked at the beginning and end of each profile corresponds to the beginning and end of a specific track line shown in the inset map. Dashed vertical lines connect similarly shaped anomalies. Some specific anomalies are identified by the numbers at the top of the theoretical profiles.
(v) The amplitude of theoretical anomalies also suggests the susceptibility of the second layer in this area is of the order of $10^{-2}$ emu/cm$^3$ with higher values in the center block (more detailed discussion will be stated in the section on the direct method).

(vi) The large anomaly in 4-3 section is probably due to a seamount as can be seen from the topography (Figure 4). The layer of magnetic material thickens sharply at this point.

(vii) As shown at the beginning of the 4-5 profile of Figure 5, there is an alternate explanation of the correlation, i.e., as the track line passes the fracture zone the magnetic anomaly lineation pattern may be repeated. For a clearer picture, refer to Heirtzler (1968), p. 5. The topography profile observed at this section is almost the same as Heirtzler (1968).

(viii) The flatness of nearly constant anomalies at about 10 m.y. in profile 2-3 may suggest support of a halt in sea floor spreading, as mentioned by Ewing and Ewing (1967).

(ix) The center of spreading determined by the magnetic anomaly (by the direct and indirect method) is not right above the topographically defined ridge (refer to Figures 3, 7, and 8). This feature may be explained by the characteristics of the inclined dipole field.
Direct Method

Both exact (Gaussian reduction) and iteration programs (p. 24) are used to calculate two direct method profiles across the East Pacific rise (profiles 2-3 and 4-5). The two solutions are compared and evaluated for their significance in interpreting the sea-floor spreading model.

Some explanation and description of Figure 7 (profile 2-3) and Figure 8 (profile 4-5) are as follows:

(i) The magnetic field and magnetization parameters used in the direct calculations (Figure 6) are from Emilia (1969). Before the program is run and the magnetization distribution is found, it is necessary to set up the dimensions of the second layer (Figure 6) under the anomaly profile. The models are as shown in Figure 3. Note that the model is quite simplified compared with the gravity and seismic sections described in the later section. If every topography change is taken into account, it would be too tedious and complicated to use the computer program adopted from Emilia (1969). However, the simplified model is close to the real condition as far as the magnetization distribution is concerned with the thickness of the first layer assumed to be of $10^{-1}$ km, the second layer thickness assumed to be 1.5 km, and an assumption of reasonable topography sloping away from the ridge axis.
Figure 6a. Parameters used in the working model and computer program (from Emilia, 1969).

\[ \text{TOTAL MAGNETIC FIELD ANOMALY} \]

\[ \text{SLOPE} + S_1 \]

\[ \text{SLOPE} + S_2 \]

\[ \text{TOTAL} \]

\[ \text{MAGNETIC FIELD} \]

\[ \text{(os, st, ds)} \]

Figure 6b. Parameters describing the direction of the regional geomagnetic field (left) and describing the direction of the magnetization (right) (from Emilia, 1969). Notice some of the parameters are not referred to in the text.

\[ \bar{i} = \text{UNIT VECTOR IN DIRECTION OF REGIONAL GEOMAGNETIC FIELD} \]

\[ D = \text{FIELD DECLINATION} \]

\[ I = \text{FIELD INCLINATION} \]

\[ \bar{t} = \text{MAGNETIZATION INCLINATION} \]

\[ \bar{g} = \text{UNIT VECTOR IN DIRECTION OF MAGNETIZATION} \]
Figure 7. Top to bottom: Digitized version of 2-3 magnetic anomaly profile (solid line) and the magnetic anomaly regenerated from direct method iteration solution (dashed line). Magnetization distribution in the source body from the exact solution (solid line) and the iteration solution (dashed line). The outline of the source body with normal (reversed) magnetization in black (white) respectively. Parameters used in the computer program for both solutions.
Figure 8. Top to bottom: Digitized version of 4-5 magnetic anomaly profile. Magnetization distribution in the source body from the exact solution of the direct method. Outline of the source body with sections of normal (reversed) magnetization in black (white) respectively. Parameters used in the computer program.
(ii) Check of digitization frequency $W_c$. In this work the digitization frequency is taken to be the distance covered by the ship in ten minutes. When the ship's speed is slower the frequency will be higher. In Figure 7, $W_c = 1/2.16$ and $1/3.12$ in Figure 8. $W_c = 1/2.88$. The smallest frequency is $1/3.12 = 0.32$. Maximum amplitude (A) of anomalies is about 350 $\gamma$. Therefore, from p. 39 of Emilia (1969) the criterion gradient of anomaly is $|\Delta T'(y)| \leq W_c A = 112 \gamma$/km; a wide range of field gradients is allowed with these parameters. This requirement can easily be seen satisfied for the two profiles run in Figures 7 and 8. The digitization interval is small enough to allow the smallest section of interest to be identified.

(iii) The iteration process in the first run (Figure 7) is stopped when $E = 50$, or $C = 0.1$ (refer to page 25). After the program was run it was found that the criterion set up by $E$ is the one terminated in the iteration, i.e., when the magnitude of the largest component of the residual field is less than 50$\gamma$. Although the accuracy of the proton precession magnetometer is 5-10$\gamma$, a 50$\gamma$ residual was used because the diurnal variation which is of the order of 35$\gamma$ was not taken into account, and the adjustment of the zero line (base line) of magnetic anomaly is ambiguous.

(iv) Because the very large number of equations involved in each profile exceeded the computer storage capacity, the first profile was run in three sections and the second in four sections with certain
anomaly points overlapping with each other. The magnetization distribution thus obtained by the iteration process is different at the overlapping area with different runs. The difference may be as much as 50\(\gamma\) when magnetic anomaly values are recalculated. The average values are taken as the final results. In the exact solution, although the recalculated magnetic anomaly values are the same as the original values, the magnetic distributions (susceptibility of the block) differ from each other. The differences are large near the end (overlapping sides) of each run. Also, and more importantly, (Emilia, 1969) the edge effect has been neglected. Both of these factors makes the magnetization distribution at or near the end of each run less reliable for interpretation and less indicative of reality. However, this is not particularly detrimental to the interpretation, because it will be shown in the next paragraph that what is important in the determination of normal or reversed magnetization is the relative magnetization, or the magnetization contrast.

(v) Due to the ambiguity of the regional geomagnetic field removal (discussed in page 15-17), and the variation of real geomagnetic field intensity at the time the material extruded, the positive (negative) magnetization value does not imply positive (negative) geomagnetic field values (Emilia, 1969). Therefore, the relative magnetization was employed to determine the normal and reversed geomagnetic field boundaries. This is why Bott (1967) hesitated to use the direct method
on the anomalies to interpret normal or reversed geomagnetic field:

The models for layer two show a conspicuous symmetry about the center of the Juan de Fuca Ridge, supporting the hypothesis of Vine and Matthews (1963) that the anomalies are caused during the process of ocean floor spreading in both directions from the center. A layer of uniform magnetization could be added to the model without affecting the calculated anomalies, it is therefore not possible to decide from these anomalies whether the magnetization is alternatively normal and reversed, or alternative strong and weak.

Due to the above facts, the later result and interpretation can not be overlooked.

(vi) The zero line of magnetic anomaly employed for each profile is the average value of the anomaly values over the entire profile. The base line was adjusted to make the magnetization distribution more easily identified with the normal and reversed geomagnetic field epoch. This adjustment can be seen from comparison of the anomaly patterns in Figures 7, 8, and 3. This adjustment seems reasonable with the consideration of the inaccuracy of the geomagnetic field removal.

(vii) The result obtained from both programs of this method is the apparent susceptibility distribution (induced magnetization neglected), assuming an axial dipole for the paleomagnetic field.

The results and discussion of the method are summarized as follows.

(i) The magnetization distribution obtained is close to the magnetic anomaly in pattern. Therefore, in the iteration process, the
starting input of anomaly values is a good first approximation of the magnetization distribution.

(ii) The magnitude of magnetization distribution (susceptibility) is of the order $10^{-3}$ to $10^{-2}$ emu/cm$^3$ which is the same as the previous results of other workers. As shown in Figures 7 and 8, the general view of the magnitude is that the one from northern profile of the rise is smaller than the profile farther south. This might be due to the proximity of northern one to the continent and/or the topography. The changes of the characteristics of remanant magnetization and chemical composition are the other possible reasons.

(iii) In their theoretical models, Pitman and Heirtzler (1966), Vine and Wilson (1965) and Vine (1966) showed that the central block (material injected from the crust to the bottom of the oceanic ridge) needed to be double the magnetization of the neighboring blocks in order to reproduce the magnitude of the observed anomalies at the center oceanic ridges. This may be because the new material has not been contaminated by the alternative reversal of geomagnetic field and by the neighboring material. Recently, after examining samples collected from the Median Valley of the Mid-Atlantic Ridge at about $45^\circ$N, Irving et al., (1970) have reported that the remanence of basalt type rock varies with distance from the axis, samples from the Median Valley (mean value $574 \times 10^{-4}$ cgs) being ten times more magnetic than samples at a greater distance. Most of this decrease in intensity
occurs within a few kilometers of the central axis and in the zone of active volcanism. They suggested also that the dramatic drop in intensity is caused by viscous decay enhanced by thermal cycling or by a chemical change in the narrow volcanic zone. This has also been examined more carefully by Schaeffer and Schwarz (1970). Their results are quoted here:

Heating up to 550°C of the specimens produces an increase in Curie temperature and saturation magnetization of titanomagnetite, which oxidized to titanomaghemite, further heating causes a decrease in saturation magnetization, Curie point to -20°C. . . . subsequent transition of titanomaghemite to hemeite . . . The thermomagnetic properties of the samples do not seem to be related to the polarity of the magnetic anomaly over their collecting sites. They may have a bearing in the intensities. . . . The gradual increase in Curie point with distance of the sampling sites from the ridge axis can explain the progression attenuation of the intensity of the magnetic anomalies. The original TRM was likely replaced by CRM components of low intensity and of different directions.

Irving et al. (1970) and also Carmichael (1970) suggested that the second layer which causes the magnetic anomaly should be a very thin layer (few hundred meters) of fine grain basalt. Underneath this layer, the coarse grained rocks were relatively weakly magnetized, which contributes little to the magnetic anomalies. Below the second layer the third layer is composed predominantly of gabbro and metamorphosed basalt (Greenstones, diabases. . . . etc., weakly magnetized) rather than serpentinized peridotite which may be strongly magnetic.

In this work, from Figure 7 and 8, it is also found from the magnetization distribution pattern that the susceptibility due to the ancient field
decreases from the ridge axis. However, the material under the ridge areas cannot be recognized as about ten times or even twice magnetic than those neighboring materials. This is the same result as Bott (1961) and Emilia (1969) obtained after they used the direct method and tried to make some interpretation.

However, from the above discussion, the characteristics of the direct method and the dipole character, it is obvious that if the magnetized layer is much thinner under the ridge than that away from it then the susceptibilities of the central blocks would be much higher than those of the neighboring blocks (i.e., a reduction of the thickness of the second layer in the ridge area from 1.5 km to 0.3 km). The results of Schaeffer and Schwarz (1970) and Irving et al. (1970) are again identified by Taiwani et al., (1971) in the study of Reykjanes Ridge crest.

(iv) The abrupt changes from the positive susceptibility to negative susceptibility or vice versa are evident on the pattern of the magnetization. This narrow transition zone supports the dike injection hypothesis as is proposed by Loncarevic et al. (1966).

(v) From the magnetization contrast as discussed in (v) of p. 37-38, it is possible to determine approximately the periods of the alternatively normal and reversed geomagnetic field. This was shown both in Figure 7 and 8, where black blocks described the normal geomagnetic field, white blocks reversed. This time scale is set up
by the careful correlation of the magnetization contrast in Figures 7 and 8 with the results of Cox (1969) (before 4.5 to 76.33 m.y. ago by the extrapolation of the dating result assuming a constant rate of spreading, Heirtzler et al., 1968). Some of the events cannot be identified. After the width of the alternatively normal and reversed magnetization blocks has been set up, it is plotted in terms of the time scale of Cox (1969) and Heirtzler et al., (1968) (Figures 9 and 10). The boundary between each change from the normal to reversed or reversed to normal is known with much more confidence than the paleomagnetic date. However, there are still some ambiguities in deciding the boundary, i.e., the boundary can vary due to human judgment. From the statistical analysis result of K-Ar dating by Cox and Dalrymple (1967) the uncertainty of the paleomagnetic date after 3.6 m.y. ago is as follows, from 0 to 1.0 m.y. ago, the standard deviation (σ) is ±0.041 m.y., from 1 to 3.0 m.y. ago, σ=±0.076 m.y. and from 3.0 to 3.6 m.y. ago, σ=±0.097 m.y. There is no information before this date, the deviation should be larger due to the dating technique and the hypothesis of assuming the constant rate of spreading. In this work, σ=±0.097 m.y. is assumed for period before 3.6 m.y. These are the lengths of the line segments shown in Figure 9 and 10. It is shown in Figure 7, the results from the iteration solution (dashed line) and from the exact solution (solid line, Gaussian reduction) are about the same in determining the periods of the normal
and reversed magnetizations so are in determining the paleomagnetic time scales. However, for the magnetization distribution and the anomalies reproduced by these magnetized blocks, the result from the iteration solution is a little bit more smooth. The iteration solution provides a better way for interpretation since in the real physical case the diurnal variation, the deviation due to instrument, navigation and man made mistakes, should be taken into account. The center point of each line segment of the time scale and the corresponding distance have been put together trying to make a least square linear fit with each other. The computer program used was obtained from the Department of Statistics of Oregon State University.

A brief review of the principle of the least square linear fit is as follows (Brunk, 1960):

Let the distance scale be $y$ and time scale be $x$; assume that they can be correlated by a linear segment

$$y = mx + b.$$  

Then the least square method is to make

$$\sum_{i=1}^{n} (y_i - mx_i - b)^2$$

minimum;

$\Leftrightarrow$ finding the line of the best fit to the point $(x_i, y_i)$

$\Leftrightarrow$ finding the regression line of fictitious random variable $y$

on the fictitious random variable $x$

$\Rightarrow$ the regression line obtained is

$$\frac{y - \mu_y}{\sigma_y} = \frac{x - \mu_x}{\sigma_x}$$
Figure 9. Spreading rate plot and the associated least square linear fit (dashed line) of the magnetization distribution from Figure 7. Right and left refer to the sides of the profile away from the ridge axis of Figure 7.
Figure 10. Spreading rate plot and the associated least square linear fit (dashed line) of the magnetization distribution from Figure 8. Right and left refer to the sides of the profile away from the ridge axis of Figure 8.
where,

\[ \mu_x = \frac{1}{n} \sum_{i=1}^{n} x_i = \overline{x}, \quad \mu_y = \frac{1}{n} \sum_{i=1}^{n} y_i = \overline{y} \]

\[ \sigma_x^2 = \frac{1}{n} \sum_{i=1}^{n} (x_i - \overline{x})^2 = (\frac{1}{n} \sum_{i=1}^{n} x_i^2) - \overline{x}^2 = S_x^2 \]

\[ \sigma_y^2 = \frac{1}{n} \sum_{i=1}^{n} (y_i - \overline{y})^2 = (\frac{1}{n} \sum_{i=1}^{n} y_i^2) - \overline{y}^2 = S_y^2 \]

\[ \tilde{\rho} = \text{correlation coefficient} = \frac{1}{n} \sum_{i=1}^{n} (x_i - \overline{x})(y_i - \overline{y})/S_xS_y \]

\[ = [(\frac{1}{n} \sum_{i=1}^{n} x_iy_i) - \overline{x}\overline{y}]/S_xS_y = r \]

Finally, after simplifying the least square line of best fit is shown in the following equation

\[ u - \overline{y} = \frac{\sum_{i=1}^{n} x_i y_i - n \overline{x}\overline{y}}{\sum_{i=1}^{n} x_i^2 - nx^2} (x - \overline{x}) \]

with the correlation coefficient,

\[ r = \tilde{\rho} = 1 \text{ if and only if every } (x_i, y_i) \text{ lie on the regression} \]
line of positive slope

\[ r = \rho = -1 \] if and only if every \((x_i, y_i)\) lies on the regression line of negative slope.

This program is employed to run the data of both sides from the ridge of two magnetization distribution from two profiles across the East Pacific rise. It is found that the slope of each linear fit (spreading velocity) and the correlation coefficient are:

<table>
<thead>
<tr>
<th>right side</th>
<th>left side</th>
</tr>
</thead>
<tbody>
<tr>
<td>Spreading Velocity</td>
<td>cm/yr.</td>
</tr>
<tr>
<td>Figure 9</td>
<td>2.89</td>
</tr>
<tr>
<td>Figure 10</td>
<td>5.10</td>
</tr>
</tbody>
</table>

We see the \( r \) is close to \(+1\). This means that the linear fit of \( x \) and \( y \) is a very good approximation, i.e., approximation of the constant rate of spreading. This supports the assumption by Heirtzler et al. (1968) of constant spreading rate.

Generally speaking, the spreading velocity is close to the results of the indirect method which predicts it is around 3.0 and 5.5 cm/yr. It is also noted that the spreading velocities of left and right side of the ridge are not equal. The right side has a smaller velocity in the northern profile 2-3. This may be interpreted as because the right side of the ridge is spreading toward the continent. According to the theory of plate tectonics, the island arc or trench area is the place
where the lithosphere sinks. Due to the stiffness of the continental crust, the sink will be stopped by the continents. Sometimes the continents would override the trench area (fill-up) and make the trench area disappear. Until the stress pattern changes and a new trench forms the spreading and the rotation of the whole plate will start in a new cycle.

It is clear that the velocity of Figure 8 (southern profile) is larger than that of Figure 7 (northern profile) from the results of the least square fit. The result supports the result of Larson and Chase (1970) who determine the Cocos-Pacific rotational pole as lying at 40°N, 110°W by the spreading velocities from the magnetic anomalies and the azimuths (strikes) of the Orozco and Siquieros Fracture Zones. They also indicate that the pole has an angular rate of $19.6 \times 10^{-7}$ deg./yr. The latitude of the pole determined by Molnar and Sykes (1969) seems to be too low. With the same angular rate, the farther away from the rotating center the higher will be the spreading velocity.

Note also the same pattern of spreading shown in Figures 9 and 10, i.e., the right (or left) side patterns of both profiles are the same. This suggests the same spreading history occurred in this area.

The line of the least square fit should pass through the origin where the last record of $y$ is present. The straight lines from this work do not, they must be shifted a little bit to pass through the origin.
If constrained to pass through the origin, the least square fit parameters (correlation coefficient and spreading velocity) would change slightly, but not significantly.

Finally, Emilia and Heinrichs (1969), show it is not possible to assume a constant rate (or nearly constant) of spreading without associating the "Ol duvâi" magnetic anomaly with the Gilsa event (-1.65 m.y.). The Ol duvâi event (1.96 m.y.) can be correlated with a consistently appearing minor anomaly called $W$, which has an average age of 2 m.y. (s.d., standard deviation, 0.06 m.y.). In this work the time scale of Cox (1969) after 4.5 m.y. and Heirtzier et al. (1968) before 4.5 m.y. are used. In Cox's (1969) scale, the $W$ event is included (called Ol duvia event; too, 2.11 - 2.13 m.y.). After checking the least square fits of the two profiles in Figures 9 and 10 it was found that they are very well approximated as constant rates of spreading. This fact provides the support of the validity of the scale of both Cox (1969) and Heirtzier et al. (1968). The four straight lines shown are also close to that required by the two assumptions by Emilia and Heinrichs (1969) of constant rate spreading and the fixed origin, start of Brunhes, center of Jaramillo and start of Gauss so that the result of this work supports the results of Emilia and Heinrichs (1969); except the $X$ event has not been considered because it cannot be identified in the anomaly patterns.
Comparison and Evaluation of Methods

It is not hard to see that both methods have advantages and disadvantages. First the direct method seems to give more detailed information because from it the magnetization distribution is obtained. While in order to get the same type of information from indirect method a lot of times must be gone through (if it is a long profile, it would be too tedious to work on) for varying the susceptibility, and the dimension of the second layer, trying to arrive finally at the observed anomalies. The weak point, however, of the direct method is that the model of the second layer has to be assumed to be very simple.

The program used for the direct method is a kind of linear programming to find the inverse of a matrix. Due to this nature, some statistical analysis can be derived from chosen program (iteration process) which makes sense physically and the calculation can be simplified by reducing the number of blocks while keeping the number of the anomaly field points the same as original block number. But it takes more computer time too. After the magnetization distribution has been determined the corresponding paleomagnetic time scale can be analyzed, and some interpretations can be made too. However, the mathematics involved in the method is far more complicated than the indirect method. If the linear equations involved in the direct method are too many, which would exceed the storage capacity of the computer,
the whole profile must be separated into several parts with some overlappings. These overlappings cause some trouble already mentioned in the previous section. The edge effect has been neglected; this makes the magnetization of the boundary blocks less reliable. This is another weak point. The program has to be revised to take the edge effect into consideration which will be some more numerical approximation involved. As also been noted, the magnetization distribution obtained from the direct method gives us the good criterion to decide the places of magnetization polarity changes only based upon the relative sense. The absolute values of susceptibilities are of questionable meaning.

Model Calculation - Trench Sections

Indirect Method

Figure 11 shows the projected magnetic, gravity, and bathymetric profiles of track lines across the Middle America trench. The gravity anomalies will be used later. Magnetic anomalies are interpreted here. The computer program for the general case (p. 20-22 and Appendix I) has been employed to interpret the trench section (Figures 11 and 12) using both induced and remanent magnetization. Because trench and ridge sections are good two-dimensional structures, the Talwani-Heirtzler (1964) method can be used. An axial dipole is
Figure 11. Projected magnetic, gravity and bathymetric profiles of trench crossings (5-6 to 11-S). See Figure 2 for locations. Dashed lines show the trend (with a certain degree of confidence) of the profile where data is missing.
Figure 12. Magnetic anomaly profiles 7-8 and 9-8 across trench area. See Figure 2 for locations. Plus and minus signs describe the change of the direction of magnetization of the corresponding block when theoretical profiles are generated. Refer to the text for description.
assumed for the remanent field. The equation:

\[ F = 0.312 (1 + 3 \cos^2 \theta)^{1/2} \times 10^5 \times X \]

with

Declination = 0

Inclination = I, \( \tan I = 2 \cot \theta \)

\( \theta = \text{geomagnetic or geographic co-latitude} \)

\( X = \text{susceptibility} \)

is used to calculate the magnetized block intensity, i.e., \( J'_x, J'_z \) (Appendix I). Magnetic field parameters (i.e., declination, inclination and total field intensity) for the induced magnetization contribution, were taken from U.S. Navel Oceanographic office charts No. 1700, 1703, 1706. From these values and the assumed induced susceptibility, the induced magnetization of the block, i.e., \( J'_x, J'_z \) (Appendix I) was obtained. Formula (4), p. 22, was used to calculate the anomalies due to remanent and induced magnetization.

In both sections interpreted below, the second layer dimension is taken from the gravity cross sections derived in a later section of the text. These sections are approximately correlated to the seismic results (Fisher, 1961; Shor and Fisher, 1961). A 1-km first layer and a 1.85 km second layer were assumed. The second layer thickens toward the continent, where the boundary between the second and third layer is about 11 km below sea surface (Figure 12). The magnetic
intensity, inclination and declination are different in different sections and blocks. The susceptibility and direction of magnetization are varied in an attempt to correlate the calculated anomalies from the assumed models with the observed results in pattern and amplitude.

The results of the two different sections calculated and some general views of all the trench profiles are as follows:

(i) For the section 2, the profile is from track segment 7-8 north of Tehauntepec ridge. The profile has been projected to $204^\circ$ to make it almost perpendicular to the coast. Then the indirect method is applied on this projected profile. The general parameters used in the calculation of this section are: induced susceptibility, $0.002 \text{ emu/cm}^3$; present magnetic field intensity, $41,000$ gammas; inclination changes in the blocks from right to left gradually from $42^\circ$ to $40^\circ N$; declination changes in the same direction from $7.6^\circ$ to $8.4^\circ E$.

For the remanent field contribution, the inclination changes gradually in the same way as above from $29^\circ$ to $26.5^\circ N$, the declination is always zero, the ancient geomagnetic field intensity is $34,200$ gammas (from $F = 0.312 (1 + 3 \cos \theta)^{1/2}$). The apparent susceptibility giving the remanent magnetization varies from model S2A to S2C. For S2A, the susceptibility is $0.0059 \text{ emu/cm}^3$ and the whole section is normally magnetized (i.e., in the same direction as the present field). It is noticed that in the left part (the ocean side of trench), the anomalies thus generated are very small in magnitude compared with that of the
observed anomalies. Also, the pattern is not consistent with the observed field. This is true even if the susceptibility is raised to \( 0.018 \text{ emu/cm}^3 \) (refer to S2B). After a few tries it was found that the only possibility of generating the amplitude of the observed anomalies (S20) rested upon the alternative change of direction of magnetization (i.e. from normal to reversed to normal). The final result is shown in S2C with the susceptibility due to the ancient field equal to \( \pm 0.0059 \text{ emu/cm}^3 \). (The sign is alternatively changed.) Compared with the (S20) observed anomalies, the good correlations are evident. These correlations support the theory of sea floor spreading, extension of the Vine-Matthews (1963) hypothesis, (Heirtzler et al., 1968) the transportation from ridges of the whole lithosphere, and the trenches are the places where lithosphere sinks and underthrusts. (The magnetic quiet zone will be discussed later.) The magnetic record shows the expected sequence of normal and reversed magnetized bodies approaching the trench, but there is no evidence of the next body, i.e., in the trench. The anomalies between the trench and the continent will also be discussed later. The magnetic anomalies from the two-dimensional bodies assumed thus also suggest that the direction of spreading is east-west to northeast-southwest. Due to the shortness of the anomaly profile, (it does not extend to the ridge) a correlation of the time scale set up by Heirtzler et al., (1968) was not attempted.
Profiles 5-6, 8-7, 9-10, 11-S have the same character, so their interpretations are the same. They support sea floor spreading and approximately east-west, northeast-southwest spreading. Particularly those on the 9-10 and 5-6 profiles, in such a long distance, anomalies contain only one large positive peak. This implies that spreading is more likely to be in the northeast-southwest direction. Profile 7-6 is the only one which appears to be perpendicular to the anomalies in the direction of increasing age. The anomalies should not oscillate too much, but do. This may be due to the effect of topography, shape, and dimension of the second layer.

(ii) For section 1, the profile is from track segment 9-8. The magnetic anomaly does not show great changes and oscillation. The topography structure shows that the 2-dimensional model is a good approximation. The profile also has been projected to 204° and then the indirect method has been applied. The general parameters used are: induced susceptibility, 0.002 emu/cm³; present field declination, 7.6°E for all these blocks; present geomagnetic field intensity changes in the blocks from right to left gradually from 42,000 to 41,000 gammas; inclination changes in the same way from 42° to 38°N. Again, for the remanent field contribution; inclination changes in the same manner as above from 29° to 24°N; declination remains always zero; remanent geomagnetic field intensity, 34,200 gammas; and apparent susceptibility of blocks giving the remanent field varies.
For S1A the susceptibility value is 0.0059 emu/cm^3 positive (i.e., normal field contribution, same direction as the present field). This result from indirect method (2-dimensional model) is completely different from the S10 (observed anomaly), even for the anomaly value in the trench area. Next for S1B the direction of magnetization is reversed and the susceptibility is set at -0.018 emu/cm^3 with results which show that the pattern is similar to S10 but that the amplitude is not. Therefore, in S1C a susceptibility of -0.0118 emu/cm^3 is assumed and the anomalies then generated became similar to S10 both in shape and magnitude. However, if looked more carefully, the detailed pattern is different, especially in the ocean site.

The small amplitude anomalies may be due to a small angle between the track line and the normal to the anomalies in the direction of increasing age (diagram above). If so, the 2-dimensional structure assumption is wrong. It is not possible to derive the apparent susceptibility from this method. An even smaller amplitude anomaly
is found in profile 11-10, which means that the angle $\theta$ is even smaller than previously and close to zero. Thus from these two anomaly patterns, a northeast-southwest spreading in the area, as well as a northeast-southwest direction of movement of the whole Cocos plate are suggested.

(iii) The magnetic quiet zone appears almost in every profile of Figure 11 above the trench area. It was noted by Heirtzler et al. (1968),

The magnetic evidence for what happens at the trench is ambiguous, however, the sea floor magnetic pattern is altered suddenly... There is no sign of a magnetic body that should (on the basis of measurements made elsewhere) be located about three kilometers below the trench floor. Its absence might be explained by heating or mechanical deformation, but neither sufficiently high temperatures nor enough seismic activity to cause deformation is indicated so close to the trench floor...

This is a problem in the sea floor spreading and continental drift. However, the quiet zone does define the boundary of the plates (Le Pichon, 1968).

(iv) Just before the quiet zone on the ocean side of the trench in the first four profiles of Figure 11, the magnetic anomaly is positive (due to the normal magnetization block), while in the last three profiles it is negative (due to the reversed magnetization block). This means that the ages of the normal and reversed magnetized segments of the oceanic crust adjacent to (ocean side) the trench are different. This supports Fisher's result (1961) that the Gulf of Tehuantepec
separates the Middle America trench into two categories according to the topography and structure, Northwest of the Gulf the trench is U shaped trench with a shallower M-discontinuity and an older crust, while southeast of the Gulf the trench is V shaped with a deeper M-discontinuity and younger crust. (The fifth profile of Figure 11 is located just across the Tehuantepec ridge.) The first four track lines are in the Guatemala Basin. The above interpretation is also consistent with the sequences of formation of East Pacific rise hypothesized by Menard (1964).

(v) In the model calculation by indirect method of these two trench sections, an identical susceptibility of the second layer for the continental shelf area and the ocean side was used. Since the susceptibility changes, the theoretical anomaly should be smaller above the continental shelf area. After the Andesite Line, the material in the second layer of oceanic crust changes from the olivine basalt lava to andesite lava and after trench, going inward to continent, the andesite lava changes gradually to granite of typical continental crust. The ferromagnesian mineral content becomes progressively smaller with distance from ocean basins to continental areas; therefore, the susceptibility of the blocks and the magnitude of the theoretical anomaly should be smaller above the continental shelf than those of S1C and S2C.
Conclusion

After all, from both the indirect and direct method, in the general point of view, except some of the prominent deficiencies in the theory of sea-floor spreading and plate tectonics itself, the result from the analysis of this set of data of magnetics supports the idea of new global tectonics.
GRAVITY

Anomalous Field Introduction

Data and Its Reduction

Gravity measurements were made with LaCoste-Romberg gimbal-suspended surface-ship gravity meter S-9. This type of gravimeter operates on moving platforms and measures relative gravity values. A brief review of the reduction of the gravity value is given, see Couch (1969) for additional details.

In order to determine absolute gravity values the relative values must be tied to a base station on land where the absolute gravity value is known. In this work the base stations are at Talara, Peru and Puntarenas C. R. The estimated accuracy of the measurements is \( \pm 5 \) mgal based on the land ties and track intersections on preceding legs of the cruise. Analysis of the long term drift rate of meter S-9, however, shows a positive offset for both base ties and the possibility exists that all the measurements are high by \( \pm 12 \) mgal (Gemperle, 1971). Since the average value of all the data is 2.2 mgal, the meter drift appears to be real and it is believed that true gravity values are
presented. Since the gravity meter is operated on a moving platform and the ship is moving on the earth's surface, corrections must be made to the recordings of the gravity meter before they can be reduced to free-air anomaly values.

The Browne correction is due to horizontal ship accelerations (\(\vec{x}\), see figure of last page) which in turn accelerate the sensing element of the gravity meter. As mentioned by Couch (1969), the measurements can be accurate to within \(\pm 3\) mgal at Browne corrections up to 300 mgal for meter S-9. The errors systematically increase with larger Browne corrections. No correction or adjustment of gravity measurements was made to correct errors due to the inaccurate horizontal acceleration determination since few data points had a Browne effect over 400 mgal.

The vertical acceleration of the ship were removed by electronic filtering. However, the filter response is not linear and the limit for vertical acceleration of the ship allowed by the design gravity meter S-9 may be exceeded at sea. Vertical acceleration corrections were calculated using meter responses determined by Couch (1969). For more information, one is referred to Appendix I of Couch (1969) and Dehlinger et al. (1966).

A third correction arises from changes in centrifugal acceleration as the ship moves on the earth surface. This effect decreases the true gravitational acceleration, and the correction to this is called the
Eötvös correction, \( \Delta g_e = wV_e \cos \theta + (V_e^2 + V_n^2)/R \) with \( w \) = angular velocity of earth, \( V_e \), \( V_n \) = east-west, north-south earth surface velocity, \( \theta \) = latitude, \( R \) = earth radius, with the estimated uncertainty to be ± 1 mgal.

In order to get free-air anomalies, theoretical gravity values (latitude effect) must be subtracted from the corrected gravity values. That is, F. A. = free-air anomalies = \( g_{all} - g_t \) where

- \( g_{all} \) = gravity values after all the corrections have been made
- \( g_t \) = the theoretical gravity values which are obtained from the 1930 International Gravity Formula.

All of the processes involved above have been taken care of by the CDC3300 computer and free-air anomalies were obtained. The detailed programs are listed in Gemperle and Keeling (1970).

**Anomaly Working Equation**

At this time only the indirect method is available. The following derivation of the working equation is only an introduction, for more detailed discussion see Grant and West (1965).

The magnetic field is defined by a vector potential. The dipole field, however, can be considered as a scalar potential because of the dipole character. On the other hand the gravity field is derived from a scalar potential directly.
From the Newton's law of universal gravitation, \( F(r) = G \frac{m_1 m_2}{r^2} \), acts in direction of \( r \) where

\[ G = \text{gravitational constant} \]

\( m_1, m_2 = \text{two mass units} \)

\( r = \text{the distance between} \ m_1 \text{ and } m_2 \).

Now the gravitational field is \( \vec{F}(r) = -G \frac{m_1}{r} \hat{r} = -\nabla U(r) \) with \( U \) assumed to be a scalar potential.

Since potentials in free space are additive, the gravitational potential due to a continuous distribution of matter may be calculated at external points by integration. Thus if the mass is distributed continuously with the density \( \rho(r_0) \) through the volume \( V \), the gravitational potential at an exterior point \( p(r) \) as shown in the figure

\[
U_{p}(\vec{r}) = -\int_{V} \frac{G \rho(r_0) d^3r_0}{|\vec{r} - \vec{r}_0|} = -G \int_{V} \frac{\rho(\vec{r}_0) d^3r_0}{|\vec{r} - \vec{r}_0|} \tag{6}
\]
with

$$|\mathbf{r} - \mathbf{r}_0| = (r^2 + r_0^2 - 2r_0 \cos \theta)^{1/2}.$$  

This equation can be compared with (1) on page 18 where the magnetic scalar potential formula is given.

The principle of the computer program used is introduced briefly.

The idea is adopted from Talwani, Worzel and Landisman (1959) for two dimensional bodies.

From (6) assume that the y dimension goes from the plus infinity to minus infinity and $\rho$ is independent of $y$, therefore,

$$U(x, z) = -G \int_S \rho(x_0, z_0) \, dx \, dz \int_{-\infty}^{\infty} \left[ (x-x_0)^2 + (y-y_0)^2 + (z-z_0)^2 \right]^{-1/2} dy_0$$

$$= 2G \int_S \rho(x_0, z_0) \log R \, dx_0 \, dz_0$$  

(7)

with

$$R = \left[ (x-x_0)^2 + (z-z_0)^2 \right]^{1/2}$$

$$S = \text{cross section} \ (x, z \ \text{dimension}) \ \text{of} \ \text{the} \ \text{body}.$$  

Now assume $\rho$ is a constant and $x = z = 0$, from (7) and $x_0 = \xi$, $z_0 = \zeta$, we obtained

$$\Delta g(0) = \frac{\partial U(x, z)}{\partial z} = 2G\Delta \rho \int_S \int \frac{\xi d\xi \, d\zeta}{(\xi^2 + \zeta^2)^{3/2}}.$$
Set \( \xi = a_k \xi + b_k \), where
\[
a_k = \frac{x_{k+1} - x_k}{z_{k+1} - z_k}, \quad b_k = \frac{x_k z_{k+1} - x_{k+1} z_k}{z_{k+1} - z_k}
\]
thus the gravity effect at 0 is given by
\[
\Delta g(0) = 2G\Delta \rho \int \tan^{-1} \left( \frac{\xi}{\xi_0^*} \right) d\xi \approx 2G\Delta \rho \sum_{k=1}^{n} \int \frac{z_{k+1}}{z_k} \tan^{-1} \left( a_k + \frac{b_k}{\xi} \right) d\xi
\]
\[
= 2G\Delta \rho \sum_{k=1}^{n} \frac{b_k}{1 + a_k} \left[ \frac{1}{2} \ln \left( \frac{x_{k+1}^2 + z_{k+1}^2}{x_k^2 + z_k^2} \right) + a_k \left( \tan^{-1} \frac{x_{k+1} z_k}{z_{k+1}^2} - \tan^{-1} \frac{x_k z_{k+1}}{z_k^2} \right) \right]
\]

This is the formula used in the computer program to run the two-dimensional structure in the following section. The program available here was initially written by Mr. W. A. Rinehart and extended and modified by Dr. E. F. Chiburis and Mr. M. Gemperle. For a listing of the program and detailed illustrations refer to Couch (1969), Appendix III.

**Free-Air Anomaly Interpretation**

Before discussing the calculated crustal structure models, a general view of the free-air anomalies obtained from this work is
presented. Since many of the ship track lines were made at angles other than $90^\circ$ to the trend of the topography, it is hard to compare them with each other. These profiles have been projected to $204^\circ$ clockwise from the geographic north. This makes each profile roughly perpendicular to the trend of the topography to suit the 2-dimensional structure for the later calculation. All the free-air anomaly as well as the magnetic and topographic profiles across the trenches (Figure 11) have been projected. However, the profiles across the ridge area and some others have not been projected because they are already almost perpendicular to the trend of the topography or there is no special value to make the projection (see Figures 3 and 4). If the projection angle is too large (as 11-S profile) it should not have much confidence put on it while making interpretation. Numbers of each profile shown in the figures are from the track line map (Figure 2).

The trench axis is chosen by a point near the center of the trench floor. It should be noted that this axis only provides a reference line for comparing profiles. No genetic significance is implied. Dashed lines in the profiles indicate the possible trend of the profile with some degree of confidence where data is missing.

From these profiles it is clear that the trench is deepest in the 9-8 profile which is obtained just to the south of the Tehuantepec ridge and shoals to the both sides of it. The anomaly value at the
trench area is most negative near the Tehuantepec ridge and becomes less negative on both sides (Figure 11). The less negative anomalies here imply an addition of mass, probably associated with the formation of the ridges (East Pacific rise and Cocos ridge).

Some of the free-air gravity anomaly profiles (as 7-8) are saddle-shaped, and are not clearly in ratio to the topography. This suggests that the low-density sediment fill in this area is thicker than elsewhere.

The average value of the entire free-air anomaly profile of both ridge crossing (profile 4-5 and 4-3), is very close to zero. This suggests isostatic equilibrium (compensated feature) in the ridge area. The average value of the whole set of data of this work is 2.2 mgal, which shows a good compensated area too.

As mentioned in the introduction, Fisher (1961) suggested that the Gulf of Tehuantepec marks a major change in the trench configuration and possibly in age. Northwest it is generally U-shaped in cross section with a flat bottom suggesting sediment fill and an older age, in comparison to the V-shaped trench southeast of the Gulf, which borders a volcanically active coast (refer to Figure 5 of Molar and Sykes, 1969 for volcanic activity). The seismic refraction results of Shor and Fisher (1961) also show differences between the two areas. The depth to the M-discontinuity is shallower and the crust below sediment is thinner northwest of the Gulf of Tehuantepec.
From the topography and gravity anomaly profiles we can see support of Fisher’s ideas. The suggestion that the age of the U-shaped trench is older northwest of the Gulf of Tehuantepec seems correlated with the types of rocks outcropping along the coast. Along the coast southeast of the Gulf the rock is mainly of Quaternary continental types including Bolson deposits, alluvium, valley fill, lake beds, windblown sand, glacial deposits, some tuffs and lava, flood plains and coastal terrace deposits, beach sands and some limestone residues. Northwest of the Gulf, except for minor amounts of the Quaternary continental type, the rocks are mainly Paleozoic metamorphic and igneous types and Mesozoic and Precambrian metamorphic rocks. However as suggested previously by Fisher (1961) there seems to be no good correlation between the types of rock outcropping on the coast or their resistance to erosion and the sediment accumulation. It is noticed that (except 5-6 profile) the shoreward flank of the trench northwest of the Gulf of Tehuantepec (profile 7-8, 7-6) is rough. This suggests that the topography must not have a sufficient extent parallel to the trench or closure to serve as effective sediment traps (Menard, 1964). Detritus coming off the continent would then be able to reach the trench floor. In contrast, the shoreward flank of the trench is smooth southeast of the Gulf. This implies that the sediment has been trapped on the shelf or slope and has been unable to reach the trench floor. One is referred to Ross and Shor (1965) for
a seismic reflection study of Middle America Trench, which shows the sediment traps parallel to the trench on the shelf and slope. Thus the U or V shape of trench may not be related to age but rather be a consequence of different sedimentation regimes. If so explained the result of this work would give some support to the previous results of Fisher (1961) and is similar to the explanation of Hayes (1966) of the Peru-Chile trench study. The Tehuantepec ridge actually becomes a wall to prevent the flow of sediments along the length of the trench. For a better and complete explanation, seismic reflection and refraction data are required.

**Crustal Sections**

**Introduction**

Profiles 4-5 and 9-8 have been studied in some detail. The reasons that these two profiles were chosen are because profile 4-5 crosses the ridge perpendicular to the topography trend, which has been studied carefully before in the magnetics section and profile 9-8 is obtained above a typical trench section which is very close in position to one of the seismic refraction lines of Shor and Fisher (1961).

In Figures 13 and 14, the solid curve represents the observed free-air gravity anomaly and the circles are the associated anomalies calculated from a two-dimensional structure and the topographic section
shown below. The seismic refraction control points are from Shor and Fisher (1961) for the Middle America trench section and from Raitt (1956) for the East Pacific rise section. The configuration of the layers and the depth of the M-discontinuity were determined by fitting the calculated free-air gravity anomaly values to the observed data.

It was pointed out by Taiwani (1964) that in order to arrive at the free-air anomalies from the computed section, one has to subtract a constant based on the standard section (nearly $= 2\pi \gamma p h$), from the calculated total attraction. The layering of the standard section is unimportant, the important thing is the total mass per unit area down to a certain depth, below which a uniform density with no horizontal variation is assumed. The mass has to be the same everywhere over the surface of the earth after the gravity anomalies have been taken into account. The total mass is related to the total attraction from the seismically determined section. Seismic control of each layer of both sections is shown in the figure with the black block, and also in Table I which also includes the total mass per unit area of each seismic section and the adjusted total mass per unit area for the observed anomalies.

A depth of compensation of 50 km is used for the assumed section. It is not very clear whether there are still some horizontal variations of density beneath this depth. However, it is a reasonable assumption that the compensation depth is within 50 km (MacDonald,
The mass per unit area derived from the seismic section control is close to that of Talwani (1959) for Puerto Rico Trench studies, that of Talwani et al. (1965) for mid-ocean ridges, and the results of Hayes (1966) for Peru-Chile Trench, i.e., \(2.87 \times 32 \times 10^5 \times 10^5 \text{gm/cm}^2\). This average density, \(2.87 \text{gm/cm}^3\), of crust to 32 km depth, and \(3.46 \text{gm/cm}^3\) below 32 km are the densities used in the calculation of the total mass per unit area with the assumed section which was first defined by Worzel and Shurbet (1955). The result is \((91.84 + 3.46 \times 18) \times 10^5 = 154.12 \times 10^5 \text{gm/cm}^2\). The density assumed for each section was obtained from the Nafe-Drake curves (Talwani et al., 1959). It has been found that the subcrustal density has to be varied in order to bring the result of the calculation for the assumed (from seismic control) cross section to be consistent with the observed anomalies.

**Ridge Section and Its Geophysical Implication**

The ridge cross section has been derived from seismic control points of the averaged East Pacific rise from Raitt (1956) and the free-air gravity anomalies obtained during the cruise. The observed and calculated anomalies show good consistency with each other. The final result shows good consistency with Menard (1960) and Talwani et al., (1956). From Raitt's seismic refraction studies, the crust under the East Pacific rise (as well as other oceanic ridges, such as
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<th>Mass/unit area ( \times 10^5 ) gm/cm(^2)</th>
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the Mid-Atlantic ridge) is thinner than that under the rise flanks and
the rest of the ocean basin. Also, the mantle seismic velocity is
lower under the rise than that under the rise flank, where the normal
mantle velocity appears.

The reasoning used in deriving the boundaries between density
variations in the mantle and its geophysical implication on new global
tectonics follows. Over the Mid-Atlantic ridge Talwani et al. (1961,
1965) show that it is unlikely that a major part of the isostatic com-
pensation for the crestal provinces occurs at a depth greater than
25 km. This is because in this area topography and/or subsurface
layers have steep slopes and the slope of the anomaly requires a max-
imum depth to which the compensation should be made. Although a
small part of the compensation may go deeper, the major part of it
must be contributed from shallow depth to prevent much sharper
boundary discontinuities in the deeper part of the crust. However,
since the slopes of the topography and the Bouguer anomalies on the
flanks of the East Pacific rise (Figure 13) are not as steep as those
of the Mid-Atlantic ridge, it is not possible to say that most of the
compensation lies above a depth of 30 km. In view of this, some
speculation on the possible deviation of the derived crustal section
follows. A lateral variation of density in the subcrustal layer has
been used to try to tie the gravity data in with the seismic data. As
mentioned by Talwani et al. (1965), conflict does occur where the
Figure 13. Deduced crustal and subcrustal cross section along the track line 4-5 (refer to Figure 2) across the East Pacific rise by the seismic control (Raitt, 1956) and observed gravity anomaly.
water depth decreases along the ridge flanks to the ridge crest while the crustal thickness does not increase as required by Airy isostasy to maintain the mass balance. In order to arrive at the observed gravity anomalies from the assumed seismic sections, the anomalous mantle underneath the rise crest has to extend underneath normal mantle on the rise flanks. The boundaries of the density discontinuity may not be as sharp as that shown in Figure 13, but the result would change little. This type of boundary may be a kind of phase change boundary. This extension of anomalous mantle into the normal mantle has some implication on the present global tectonics theory. As was suggested by Taiwani et al. (1965), if the anomalous mantle was converted from normal mantle, it requires some uplift and/or lateral expansion in the rise crest area. This idea gives very good gravity and seismic support to the hypothesis that the material rises near the rise crest and spreads apart, causing the movement of the whole lithosphere.

At the left of the rise crustal section, there appears a seamount, which is a three dimensional figure. Therefore, the derivation of the crustal section by a calculation assuming the two dimensional structure is not reasonable. One can not put too much confidence on the crustal section layering underneath it. The possibility that it might be the locality of occurrence of a fracture zone (as discussed in the magnetics section) cannot be ruled out.
Trench Section and Its Geophysical Implications

The derived crustal section is in good agreement with seismic and gravity observations. It is obvious from Figure 14 that the crust is thinner under the trench area and that the thickness lies between the crustal thickness of normal continental and oceanic areas. This is characteristic of a transition zone. The most striking characteristic is that of the thinning of crust (even thinner than that directly beneath the trench) beneath the seaward flank of the trench. This configuration of the thinning of the crust may be replaced by one which assumes a higher density material in the crust, but too high a density would be required. The thinning of the crust with high density material interpreted as being mantle is more reasonable. This result is the same as those of Shor and Fisher (1961) for Middle America Trench, Talwani (1959) for Puerto Rico Trench and Hayes (1966) for Peru-Chile Trench section (near Antofagasta and Callao). From this thinning of crust and the uniform standard section set up for the calculation, a deficit of mass associated with the trench is expected. According to Talwani (1964), the average mass anomaly depends upon the area extent considered, and the local deficit of mass requires compressive forces to maintain dynamic balance. When the entire set of data (from this work) is considered assuming isostatic equilibrium over a large area, it is not necessary to invoke compressive
Figure 14. Deduced crustal and subcrustal cross section along the track line 9-8 (refer to Figure 2) across the Middle America trench by the seismic control (Shor and Fisher, 1961) and observed gravity anomaly. Dashed line, which is less reliable, indicates a gradual density transition.
force to explain the mass imbalance which occurs. The average value of free-air gravity anomalies of all of the data is 2.2 mgal. Thinning of the crust on the offshore flank of the trench, also suggests the absence of compressive forces in the mechanism of formation of the feature. In the opinion of Worzel (1965), the topography and negative gravity anomalies associated with trenches can be produced by graben type faulting or by a combination of faulting and flexure in the transition region of the ocean-continent margin. As a further step, he proposed that the crustal stresses must be lateral extensional. According to Hayes, (1966) the Peru-Chile trench could be produced by high-angle normal faulting near the base of the continental slope accompanied by a downward flexure further offshore. The thinning of the crust of the offshore flank existed prior to faulting and served to determine the hinge-line of the flexure. He favored also the extensional rather than compressional stress idea. As long ago as 1955, Ewing and Heezen pointed out that the tensional rather than compressional stress is necessary to explain a local crustal thinning. The writer here suggests the same type of ideas as these author's (normal faulting and extensional), which in turn does not conflict with the principle of new global tectonics. The focal mechanism study of earthquakes in this area, which provided evidence for the lithosphere sinking in the trench zone, has been performed by Molnar and Sykes (1969). Also it is concluded by Le Pichon (1968) and suggested by
Elasasser (1968) that,

Regions in which earth's surface is shortened or destroyed are called regions of compression. Most geologists would agree that the Alpine-Himalayan belt is indeed a region of compression. It has been argued convincingly however, that deep sea trenches are regions of tensile stress. Elasasser has suggested that the tectosphere becomes denser as it slides down the Gutenberg fault zone and acquires a sufficiently greater density than surrounding mantle material to sink on its own. "The motion... then leads to a tensile pull in the adjacent part of the tectosphere."... As the differential movement of compression between two blocks increases, the associated surface compressional features apparently become larger and reach a maximum for a rate of movement of about 5-6 cm/yr (Himalayas). At larger rates, the lithosphere sinks along an active trench, and the associated surface features are tensional instead of compressional... As seen from the free-air gravity anomaly profiles of the trenches, the very steep gradient of anomaly value near the trench, along with an even steeper gradient of the Bouguer anomaly, implies that the models which provide compensation at greater depth under the trench area are ruled out. This has been discussed more in the former section where the ridge cross section is demonstrated.

The Tehuantepec ridge is a prominent submarine feature transverse to the Middle America Trench. It strongly alters the character of the trench. For this reason, in the evolution of the structure of this area the ridge is presumed younger than or contemporaneous with the age of the trench.
CONCLUSIONS AND DISCUSSION

The principles of sea-floor spreading, the new global tectonics and the associated possible mechanisms have been re-examined by the geophysical data obtained from Yaloc-69. The data analyzed in this work includes bathymetry, gravity and magnetics measurements in the Cocos plate area. Although the data collected are not so extensive that precise analysis could be made, some results do have some geophysical implications which show support both for and against the plate tectonics theory and some results of previous work. Indirect and direct methods of magnetics, and the indirect method of gravity have been employed. The indirect method of magnetics shows (1) there is good correlation between the sea-floor spreading magnetic pattern and Heirtzler et al., (1968) and Cox's (1969) paleomagnetic time scale. This supports Vine and Matthew's (1963) hypothesis. (2) In the trench province the magnetic anomalies have to be generated by alternative positive and negative remanent magnetizations of second layer, which can not be due to induced magnetization alone (1/10 of remanent magnetization in magnitude). This favors the sea-floor spreading hypothesis and also the idea of movement of the lithosphere of plate theory. (3) The direction of movement of Cocos plate is northeast and southwest, which supports the results of Molnar and Sykes (1969). However, their result of the readjustment in the motion of the Cocos plate which
supposedly occurred about 4 m. y. ago instead of 10 m. y., has not been found. The ten m. y. ago event of episodic spreading is preferred (refer to Ewing and Ewing, 1967). (4) The greater age of the trench floor northwest of the Tehuantepec ridge is suggested by the different sign of the magnetic anomaly just offshore the trench. The direct method of magnetics shows (1) support of the paleomagnetic time scale by Cox (1969) after 4.5 m. y. and Heirtzler et al., (1968) from 4.5 to 76.33 m. y. ago, assuming a constant rate of spreading with spreading velocity between 2.89 and 5.10 cm/yr. This is consistent with the results of Isacks et al. (1968) derived from the length of seismic zone underthrust down dip and Larson and Chase (1970). This favors sea-floor spreading and plate tectonics. (2) The magnetization of the second layer decreases from the ridge crest toward the flanks. This is the same result found by Schaeffer and Schwarz (1970) who explained that this decrease is due to the TRM being replaced by the CRM components. The gravity results show, (1) thinning of the crust underneath the rise, trench and just offshore of the trench and imply that material rises from underneath the ridge, expands apart from the ridge, and sinks at the trench area. At the trench area where the lithosphere sinks, the structure shows tensional instead of compressional tectonics. This result favors the principles of sea-floor spreading and plate tectonics as has been mentioned by Elsasser (1968). (2) The Tehuantepec ridge divides the trench province into two
separate parts which can be distinguished by the free-air anomaly pattern.

In general the data from this work support sea-floor spreading and plate tectonic theories. If further study of the theory in more detail is needed, it is necessary to have more and various kinds of data such as those from coring, seismic, magnetic measurement of the coring samples... etc. So that although this set of theory is really attractive, other propositions of the continental drift, such as expansion hypothesis mentioned in the introduction of this writing still cannot be excluded so far.
BIBLIOGRAPHY


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Hurley, P. M. 1968. The confirmation of continental drift. Scientific American v. 218, no. 4 pp. 53-64.


APPENDICES
General Case (listing of program - TAWAHETZ attached)

As indicated on page 22, Equation (4) is used for the most general computation:

\[ \Delta T_v = \Delta T_T \sin I + \Delta T_h \cos I \cos (C-D) \]

First, \( \Delta T_v \) and \( \Delta T_h \) can be rewritten as

\[ \Delta T_v = 2(JQ - JP) \]
\[ \Delta T_h = 2(JP - JQ) \]

where

\[ P = \frac{z_{21}^2}{z_{21}^2 + x_{12}^2} (\theta_1 - \theta_2) + \frac{z_{21} x_{12}}{z_{21}^2 + x_{12}^2} \log \frac{r_2}{r_1} \]
\[ Q = \frac{z_{21} x_{12}}{z_{21}^2 + x_{12}^2} (\theta_1 - \theta_2) - \frac{2}{z_{21}^2 + x_{12}^2} \log \frac{r_2}{r_1} \]

and

\[ x_{12} = x_1 - x_2, \quad z_{21} = z_2 - z_1, \quad r_1 = (x_1^2 + z_1^2)^{1/2}, \quad r_2 = (x_2^2 + z_2^2)^{1/2} \]
To obtain the total intensity anomaly value at a field point, the field due to each individual block must be superposed.

For each block the following items have to be defined:

(1) total number of field points where anomalies are to be calculated

(2) $x$ coordinate of the first field point where the anomaly is to be calculated

(3) separation of the field points

(4) total number of corners of the block

(5) angle between the positive $x$ axis and the geographic north, measured clockwise from geographic north, in degrees

(6) coordinates of corners of the block

(7) magnitude and direction of the induced magnetization (present geomagnetic field parameters of an inclined dipole)
(8) magnitude and direction of the remanent magnetization

(ancient field parameters of an axial dipole averaged characteristic of the ancient field)

The source program is from Talwani and Heirtzler (1964), and has been modified by the author. It is written in FORTRAN IV and was used on a CDC3300 computer. After the calculation, the superposed $\Delta T_h, \Delta T_v$, and $\Delta T$ are obtained in gammas, and a second program is used to plot the $\Delta T$ vs distance. For a detailed illustration refer to the program list attached.

Special Case (listing of program - VINEMATT attached)

This program was obtained from the Scripps Institute of Oceanography and has been modified by Mr. M. Gemperle and the author. In this program the following items are defined:

(1) constant rate of spreading in cm/yr.

(2) paleomagnetic time scale (Heirtzler et al., 1968, Cox, 1969)

(3) present and original latitudes of anomalies in the direction of increasing age (latitude = $90^\circ - \theta, 90^\circ - \theta'$)

(4) present and original angles between the line normal to the anomalies in the direction of increasing age and true north (azimuth = $\phi, \phi'$)

(5) depth and thickness of the magnetized block (second layer)
As mentioned in page 20, a two dimensional model is assumed. From the items (1) and (2) above the dimension (x and z coordinates of each rectangular block) and the character (normal or reversed magnetization) of each block are defined. Then the Equation (4) of page 23 is used to calculate the superposed anomaly field by the items (3) and (4) above of magnetization parameters.

In this program a one thousand-kilometer block at each side of the model is added to take care of the edge effect. In the whole process of calculation, the earth present field is evaluated from the dipole (inclined dipole) assumption, i.e., 
\[ F = \frac{M}{a^3} (1 + 3 \cos \theta)^{1/2} \]  
with the associated formulas: 
\[ \tan \theta = 2 \cot \phi \]
\[ J_x = F \sin \theta \cos \phi \]
\[ J_y = F \sin \theta \sin \phi \]
\[ J_z = F \cos \theta \]
where $\theta = \text{co-latitude (geomagnetic coordinate)}$

$$\frac{M}{a^3} = 0.312 \text{ oersted (regular part of the averaged earth field before 1945).}$$

For the remanent magnetization contribution, an axial dipole is assumed and a susceptibility of $0.01 \text{ emu/cm}^3$ is used for all black except the central block, which has a susceptibility of $0.02 \text{ emu/cm}^3$. The dipole formula of the last page is again employed. In this program the minor effect, due to the induced magnetization, is not considered because only the approximate shape and magnitude of the anomaly caused by the theoretical spreading model are required. After the superposition of the contribution of each block due to $J_x', J_y'$, $(\text{We defined them as } R_x, R_z; \text{ where } J_x' = R \sin l' \cos \phi', J_y' = R \cos l', \phi' = 0^\circ, \tan l' = 2 \cot \theta', R = 0.01 \frac{M}{a^3} (1 + 3 \cos^2 \theta')^{1/2}).$ the total intensity anomaly is found by

$$\Delta T = \left[ (J_x + R_x)^2 + (J_y + R_y)^2 + J_y^2 \right]^{1/2} - F.$$
PROGRAM TAMAHETZ

C MODIFIED TALWANI-HEIRTZLER MAGNETIC ANOMALY 2D

DIMENSION FX(500),EXX(200),ZEE(200),PSUM(500),QSUM(500),H(500),V(5100),T(500),X(500),HX(500),VX(500)

801 FORMAT(I5,F8.3)
802 FORMAT(F8.3,F8.3)
804 FORMAT(3F15.6)
805 FORMAT(F10.5)
806 FORMAT(*- MAG DUE INDUCED SUS=F7.4,F7.0,F0=*,F5.1,*DIP=F4.1)

110 READ(1,805) S1.
120 IF EOF(1)) WRITE(61, 813)
130 READ(1,801)KTOT,ZO
140 IF(EOF(1) )WRITE(61, 813)
150 READ(1,802)EXX(J),ZEE(J)
160 IF (EOF (1)) WRITE(61, 813)
170 PSUM(K)=0
QSUM(K)=0
X1=EXX(J)-FX(K)
Z1=ZEE(J)-FZ
RSQ1=X1**2+Z1**2
IF(X1) 110,11.0,180
110 IF(Z1)120,130,130
120 THETA=ATANF(Z1/X1)-3.1415927
GO TO 200
130 THETA=ATANF(Z1/X1)+3.1415927
GO TO 200
140 IF(Z1)150,160,170
150 THETA=-1.5707963
GO TO 200
160 THETA=0
GO TO 200
170 THETA=1.5707963
GO TO 200
180 THETA=ATANF(Z1/X1)
200 J=2
210 X2=EXX(J)-FX(K)
Z2=ZEE(J)-FZ
RSQ2=X2**2+Z2**2
IF(X2)210,240,280
210 IF(Z2)220,230,230
220 THETA=ATANF(Z2/X2)+3.1415927
GO TO 300
230 \text{THETB=ATANF(Z2/X2)+3.1415927}
\text{GO TO 300}
240 \text{IF(Z2)250,260,270}
250 \text{THETB=-1.5707963}
\text{GO TO 300}
260 \text{THETB=0.0}
\text{GO TO 300}
270 \text{THETB=1.5707963}
\text{GO TO 300}
280 \text{THETB=ATANF(Z2/X2)}
300 \text{IF(Z1-Z2)320,31,320}
31 \text{P=0.}
\text{Q=0.}
\text{GO TO 32}
320 \text{OMEGA=THETA-THETB}
\text{IF(OMEGA)3201,3202,3202}
3201 \text{IF(OMEGA+3.1415927)320,330,330}
3202 \text{IF(OMEGA-3.1415927)330,330,340}
330 \text{THETD=OMEGA}
\text{GO TO 370}
340 \text{IF(OMEGA)350,360,360}
350 \text{THETD=OMEGA+6.2831853}
\text{GO TO 370}
360 \text{THETD=OMEGA-6.2831853}
370 \text{X12=X1-X2}
\text{Z21=Z2-Z1}
\text{XSQ=X12**2}
\text{ZSQ=Z21**2}
\text{XZ=Z21*X12}
\text{GL=0.5*LOGF(RSQ2/RSQ1)}
\text{P=((ZSQ/(XSQ+ZSQ))*THETD)+((XZ/(XSQ+ZSQ))*GL)}
\text{Q=(THETD*(XZ/(XSQ+ZSQ)))-(GL*(ZSQ/(XSQ+ZSQ)))}
32 \text{IF(S4)33,34}
33 \text{WRITE(5,803)K,J,X1,X2,Z1,Z2,THETA,THETB,THETD}
\text{WRITE(5,804)GL,P,Q}
34 \text{PSUM(K)=PSUM(K)+P}
\text{QSUM(K)=QSUM(K)+Q}
\text{X1=X2}
\text{Z1=Z2}
\text{RSQ1=RSQ2}
\text{THETA=THETB}
\text{J=J+1}
\text{JR=J-1}
\text{IF(JR=JTOT)201,36,36}
36 \text{CONTINUE}
\text{READ(1,808)D,DIP,F}
\text{IF(EOF(1))WRITE(61,813)}
\text{CDIPD=COSF(.0174533*DIP)}
\text{SDIPD=SINF(.0174533*DIP)}
\text{SDD=COSF(.0174533*(C-D))}
\text{READ(1,809)SUS}
\text{IF(EOF(1))WRITE(61,813)}
\text{WRITE(6,806)SUS,F,D,DIP}
\text{AN=SUS*F}
\text{CDJP=CDIPD}
\text{SDJP=SDIPD}
\text{SE=SDD}
\text{READ(1,808)AM,B,A}
\text{IF(EOF(1))WRITE(61,813)}
\text{WRITE(6,809)AM,B,A}
WRITE (7, 807)
CDIP = COSF (.0174533*A)
SDIP = SINF (.0174533*A)
SD = COSF (.0174533*(C-B))
DO 4321 K = 1, KTOT
HX(K) = 2.*AM*( CDIP*SD*PSUM(K) ) + (SDIP*QSUM(K) )
VX(K) = 2.*AM*( CDIP*SD*QSUM(K) ) - (SDIP*PSUM(K) )
HX(K) = 2.*AM*( CDIP*SE*PSUM(K) ) + (SDIP*QSUM(K) ) + HX(K)
VX(K) = VX(K) + 2.*AM*( CDIP*SE*QSUM(K) ) - (SDIP*PSUM(K) )
H(K) = H(K) + HX(K)
V(K) = V(K) + VX(K)
X(K) = HX(K)*CDIP*SD+VX(K)*SDIP+T(K) = T(K)+X(K)
4321 CONTINUE
READ (1, 805) S4
IF (EOF(1)) 600, 480
600 DO 610 K = 1, KTOT
610 WRITE (7, 810) K, FX(K), H(K), V(K), T(K)
READ (2, 805) SCALX
IF (EOF(2)) CALL EXIT
M = KTOT
EX = FX(M)
IF (SCALX < 0.000001) 510, 510, 520
510 SCALX = 10./EX
520 LY = SCALX*EX+2.
EX = LY/SCALX
IF (AXIS(1, 10, LY, 100., 2000., EX, -1000., 0., 0., 0., 20., 4)) 530, 540
540 CALL EXIT
530 XW = -T(1)
Y = FX(1)
CALL PLOTXY(XW, Y, 0, 0)
DO 550 K = 1, KTOT
XW = -T(K)
Y = FX(K)
550 CALL PLOTXY(XW, Y, 1, 9)
CALL PLOTXY(-1000., EX, 0, 0)
END

NO ERRORS FOR TAWAHETZ
SYNTHETIC MAGNETIC ANOMALIES ARE CALCULATED FROM THE VINE MATTHEWS HYPOTHESIS.

THE REVERSAL TIME SCALE IS FIRST READ IN IN FORMAT 501 AND MUST BE
ARRANGED TO START AT THE YOUNGEST REVERSAL. TA(1) IS THE TIME OF THE
NEXT REVERSAL IF [SIGN] IS +1. IF THE CHANGE WAS FROM REVERSED TO NORMAL, -1.
THE OPPOSITE. THIS 9.2 +1,10 = -1 IS A REVERSED PERIOD FROM 9.2 TO 7.8 YEARS.
THE TIMES MUST BE IN MILLIONS OF YEARS. THE TIME SCALE MUST BEGIN WITH A CARD.
GIVING SIGN OF THE PHASE WHICH FOLLOWED TA(1) AND END WITH A BLANK CARD.

CLAT IS THE LATITUDE AT WHICH THE FLOOR WAS FORMED. CONFORM THE ANGLE BETWEEN THE
NORMAL TO THE ANOMALIES IN THE DIRECTION OF INCREASING AG AND TRUE NORTH.
TAKE'S FROM 0 TO 360.

VEL IS THE SPREADING VELOCITY IN CM/YEAR AND PHRM PRESENT LATITUDE A
AND ORIENTATION. ALL THESE VARIABLES MUST BE SET.

THE FOLLOWING ARE OPTIONAL.

IF [SYM]=+1 THE ANOMALY FROM A SYMMETRIC SPREADING RIDGE IS GENERATED.
IF DEEP=1, THE MAGNETIC CRUST IS BETWEEN 5.5 AND 7.5 KM DEEP, OTHERWISE IT IS
TAKEN TO BE FROM 3.3 TO 5.5 KM.

IF SCALX=0, THE HORIZONTAL SCALE IS ADJUSTED TO FILL A 10 IN PLT. OTHERWISE IT
GIVES THE LENGTH IN INCHES WHICH REPRESENTS 1 KM.

CARDs OF FORMAT 500 MAY FOLLOW EACH OTHER INITIAL EXECUTION HAVING 502,
BY A BLANK CARD. IF VEL IS NEGATIVE A NEW TIME SCALE IS READ IN. A THROWN BY A
BLANK CARD AS BEFORE FOLLOWED BY CARDS IN FORMAT 501.

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CONTINUE
DO=3,1,15/1AU*.
THEN0°(90°-CLAT)
CA=ATAN(SIN(TH)/(2*COS(TH)))
IF(CNORM=140.)*10,5,5
CA=0°
10 CNORM=10CNORM
CA=SQR((1+3*COS(TH)**2-COS(TH))/300.)
CM=SIN(45°)*COS(CNORM)
CM=SIN(CA)*COS(CNORM)
CM=PH/6.5
TM=X(N)*100°-PLAT)
CA=ATAN(SIN(TH)/(2*COS(TH)))
IF(PNCRN=180.)*15,12,12
PA=PA*
15 PNCRN=4PNCRN
PA=SQR((1+3*COS(TH)**2-COS(TH))/300.)
CM=PHI*SIN(PA)*COS(PNCRN)
CM=PHI*SIN(PA)*SIN(PNCRN)
CM=PHI*COS(PA)
CM=-1
IF(ISYM=1.)*30,50,30
30 TC=10,J=1+NN
XA(J)=XJ(1)*100*VEL*(TA(J)-TA(J))
40 CONTINUE
GC TC 70
50 J=1
51 J=J+1
55 LI=N-NJ+2
L2=N-NJ+1
XA(J)=XJ(J+1)+100*VEL*(TA(L1)-TA(L2))
GC TC 51
60 XA(N)=XN(N+1)+200*VEL*(TA(2)-TA(1))
GC TC 51
65 L1=N-NP
L2=N-1
XA(J)=XJ(J-1)+100*VEL*(TA(L1)-TA(L2))
IF(J-2*N+2)*51,66,66
66 IF(N-NRN)*61,69,61
61 NRNN
NR=N-2
ISIGN=ISIGN(N)
67 IS=J+1+NR
L1=J+1
67 CONTINUE
ISIGN(2*N-2)=ISIGN
ISIGN(N-1)=ISIGN(N)
NR=2
6C 68 J=1+NR
L1=2*N-7
CS3 FORTRAN VERSION 2.1 VINE 09/09/70 1729

ISIGN(J) = ISIGN(N)

CONTINUE

N = P + (N-1)

NN = N - 1

M = X(I(I))/2.

IF (M > 2500) 75, 75, 71

WRITE (61, 520)

RETURN

520 FORMAT('IH, N IS GREATER THAN 2500')

55 \*I = 0

DO 10 I = 1, MM

X(I) = 2.0*(I-1)

N(I) = 0.

H7(I) = I.

4N = I - 1

DO 40 J = 1, NN

X(N) = XA(J) - XP(I)

IF (ABS(X(N)) > 500.0) 82, 72, 70

K = 1

82

X(N) = YA(J-1) - XP(I)

IF (ABS(X(N)) > 500.0) 85, 85, 83

K = K + 1

85

IF (K-2) 85, 90, 90

SM = M*ISIGN(J)

SMZ = MZ*ISIGN(J)

IF (ISYM = 1) H4, 81, 84

81

IF (J-N/2) 84, 86, 84

SM = P*SMX

SMZ = P*SMZ

CONTINUE

CALL FIELD

H4 = HX + UPHX

H7 = HZ + UPHZ

CONTINUE

X(N) = -XP(I)

Y(N) = X(N) - 1000.

SM = MX*ISIGN(J)

SMZ = MZ*ISIGN(J)

CALL FIELD

H4 = HX + UPHX

H7 = HZ + UPHZ

X(N) = XA(N) - XP(I)

X(N) = X(N) + 1000.

SM = MX*ISIGN(N)

SMZ = MZ*ISIGN(N)

CALL FIELD

H4 = HX + UPHX

H7 = HZ + UPHZ

H4 = SQRT((PHX*HX)*(PHX*HX) + (PHZ*HZ)*(PHZ*HZ) + PHY*PHY)

H4(1) = H4

CONTINUE

4M = 1

EX = XM(MM)

S = 0.

IF (SCALX - 0.000001) 701, 701, 702

701

SCALX = 1.0/EX

702

LY = SCALX*EX + 2.

FX = LY/SCALX

IF (AXISXY(2, 10.0) LY, 100.0, 2000.0, EX = -1000.0, SX, U, 0.0, 100.0, 100.0)) 703, 704

704

GRTSTY6
CS3 FORTRAN VERSION 2.1  VINE  09/09/70  1724

703  XW=-HA(I)
     Y=XP(1)
     CALL PLOTXY(XW,Y,0,0)
    DC 705  I=1,MM
     XW=-HA(I)
     Y=XP(1)
705  CALL PLOTXY(XW,Y,1,9)
     CALL PLOTXY(900.,0.,0.,0.)
    NR=I-1
     XW=400.
    DC 200  J=1,NK
     IE=I14H(J)
     IF(I*L1+0) I=0
     Y=XA(J+1)
200  CALL PLOTXY(XW,Y,I,32)
     CALL PLOTXY(1000.,XP(2)*0.,0.)
     CALL LABEL(144.1,3,LAR)
     CALL PLOTXY(-1000.,EX,0.,0)
    WRITF(5*7(7) XP(I)*HA(I)*I=1,MM)
706  FORMAT(6,2F9.1,3X))
    RETURN
END

NC ERRORS FOR VINE

P 01465  C 12466  D 00044

CS3 FORTRAN VERSION 2.1  09/09/70  1724

COMMON /DA=CLAT,SCALX,SIZM,BHX,PIA(400),SIGN(400),NN*7(2)
11 SCALEX=SMX,SMZ,PHX,PHZ,XA(100),HA(100),XP(100),X(2)
COMMON /DATA/LAH*(3B)
PHX=0.,
PHZ=0.,
DC 5  I=1,2
A=X(1)*X(I)
DC 5  J=1,2
ATN=ATN(A(I)/Z(J))
ALG=ALG(A(I)/Z(J))
DEX=2.*SMX*ATN*SMZ*ALG
DEZ=2.*SMX*ATN*SMZ*ALG
IF(I,J=2*(I+J)/2)15,11,15
10  PHX=DEX*PHX
     PHZ=DEZ*PHZ
     TC 5
15  PHX=DEF*PHX
     PHZ=DEF*PHZ
     CONTINUE
    RETURN
END

NC ERRORS FOR FIELD

P 00211  C 12466  D 00044