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Geochemistry

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A STUDY OF THE HISTORY OF SEDIMENTATION IN THE RED SEA
BY MEANS OF ISOTOPIC AND GEOCHEMICAL METHODS

DISSERTATION

Presented in Partial Fulfillment of the Requirements for
the Degree Doctor of Philosophy in the Graduate
School of The Ohio State University

By
Phillip David Boger, B.S.

The Ohio State University
1976

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I would like to acknowledge the help, advice, guidance, and inspiration given to me by Dr. Gunter Faure. I sincerely hope that his faith in me and my abilities has been deserved.

I would like to thank my wife, Jane, for all her help and tolerance during the preparation of this manuscript and Willy Reiser, who introduced me to geology. Dr. David A. Ross of Woods Hole Oceanographic Institute provided the samples used in this study, Bob Wilkinson helped collect the samples, and Karen Tyler drafted some of the diagrams. Their help was greatly appreciated. The manuscript was critically reviewed by Dr. Kenneth Stanley and Dr. John Sutter, whose comments and suggestions were most helpful.

Funds for this research were granted by the National Science Foundation, Research Grant #GA39353.
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 CHAPTER I
INTRODUCTION

The possibility of using the isotopic composition of strontium to determine the geologic provenance of the detrital-silicate fraction of marine sediment was first suggested by Dasch (1969). He reported that the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of rocks are essentially unchanged by chemical weathering and that the strontium in the detrital-silicate fraction of sediment does not equilibrate isotopically with strontium in solution in the oceans. He determined the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the detrital-silicate fractions of sediment samples from core tops collected in the Atlantic Ocean and found regional variations which reflect the provenance of the sediment. High $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were found near the mouths of rivers draining old terrains whereas low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were observed near areas of younger rocks. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of detrital sediment may be used to calculate dates which may reflect the ages of the rocks of nearby land masses. However, such dates are ambiguous because they are based on mixtures of sediment derived from source rocks of differing ages. The objective of this research is to develop a method for the study of sedimentation based on the concentrations and isotopic compositions of strontium in the detrital fraction of sediment. This method provides precise information concerning varying proportions of sediment derived from different sources, such as old sialic rocks.
of the continental crust and young volcanic rocks of basaltic composition.

The isotopic compositions and concentrations of strontium in old sialic and young volcanic rocks differ greatly (Faure and Powell, 1972). Highly differentiated rocks such as granites tend to have relatively low strontium concentrations which average about 285 ppm (Faure and Powell, 1972). On the other hand, the average strontium concentration of basaltic rocks is about 465 ppm (Faure and Powell, 1972). Old sialic rocks have high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios because of the accumulation of radiogenic $^{87}\text{Sr}$ formed by decay of $^{87}\text{Rb}$. Faure and Powell (1972) calculate a $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.719 for 2.5 billion-year-old continental crust predominantly granitic in composition. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of young basaltic rocks typically lie in the band 0.704 ± 0.002 (Faure and Powell, 1972). Thus, old granitic rocks generally have high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and low strontium concentrations whereas young basaltic rocks have low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and high strontium concentrations.

Many depositional basins receive sediment derived from old sialic rocks as well as detritus associated with younger volcanic rocks of basaltic composition. A basin located between a continental land-mass and an island arc is an obvious example. This study is based on the hypothesis that the concentration and isotopic composition of Sr in the detrital-silicate fraction of sediment deposited in such a basin are uniquely determined by the relative proportions of volcanogenic detritus and weathering products derived from old sialic rocks. The equations describing such
binary mixtures have been derived by Pushkar et al. (1972); Faure and Boger (1976); and others. Shaffer (1974) has applied this technique to samples collected at the water-sediment interface in the Ross Sea and was able to obtain information regarding the regional distribution of volcanogenic detritus in the sediment of the Ross Sea. Neat (1975) has shown that even the detrital-silicate fractions of non-marine carbonates from the Flagstaff Formation of Utah can be treated in terms of this model.

This study is an attempt to determine time-dependent variations of the concentrations of volcanogenic sediment taken from a series of piston cores. The Red Sea basin was selected for this purpose because it receives sediment both from old sialic terrain and from young volcanic rocks. Moreover, previous work by Ikpeama, Boger, and Faure (1973) suggested that the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and Sr concentrations of detrital-silicate fractions of sediment collected in different depths in a piston core from the Red Sea could be accounted for by such a model. This project therefore expands upon these preliminary results and seeks to show that the systematic variations in the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and Sr concentrations of detrital sediment can be used to determine the depositional history of the Red Sea during the past 100,000 years.
CHAPTER II
GENERAL GEOLOGY OF THE RED SEA AREA

PHYSIOGRAPHY

The Red Sea is a narrow, elongate basin trending NNW-SSE for nearly 2,000 kilometers between Africa and the Arabian Peninsula (Figure 1). In the north the basin bifurcates, forming the Gulf of Suez and the Gulf of Aqaba. In the south, the Red Sea joins the Gulf of Aden through the Straits of Bab el Mandeb. The basin is approximately 175 kilometers wide in the northern portion, broadens to nearly 350 kilometers in the central region, and narrows to about 30 kilometers at the Straits of Bab el Mandeb. The bottom topography is dominated by a central valley which is approximately 60 kilometers wide and ranges in depth from 1,000 to more than 2,000 meters below sea level (Laughton, 1970). This median valley is surrounded by marginal shelves with an average water depth of about 200 meters. The Red Sea becomes progressively shallower toward the south, approaching a water depth of only 100 meters in the Straits of Bab el Mandeb.

The salinity of the Red Sea is higher than that of normal sea water because of the arid climate of the region and the restrictive nature of the Straits of Bab el Mandeb. There are no rivers flowing into the Red Sea and evaporation far exceeds precipitation, causing the surface salinities to increase (Siedler, 1969). The Straits of
Figure 1. Map of the Red Sea with 1000 m bathometric contours. Depths greater than 2000 m are shown in black.
Bab el Mandeb restrict the exchange of the more saline water of the Red Sea with the Gulf of Aden and further contributes to high salinities of the Red Sea, which range from 37 to 40 parts per thousand (Siedler, 1969).

The relationship between salinity and the exchange of water through the Straits of Bab el Mandeb suggests possible variation in salinity in the past due to fluctuations of sea level. This possibility is presented in greater detail in Chapter IX.

GEOLGY OF THE RED SEA AREA

The Red Sea is generally believed to be a proto-ocean resulting from the separation of the former African-Arabian Shield into two continental plates. The formation of the Red Sea basin occurred in two stages. Thinning and eventual break-up of the crust began in the Early Eocene and was followed by sea-floor spreading which continued to the Early Oligocene (Girdler and Styles, 1974). Associated with this first period of tectonic activity are the Upper Cretaceous to Oligocene volcanics of the Trapp Series located in the southwestern portion of the Red Sea area (Figure 2). The compositions of these volcanic rocks range from tholeiitic to andesitic with the predominant rock type being alkali basalt (Burek, 1969).

The first stage of tectonic activity in the Red Sea basin was followed by a period of inactivity lasting until the end of the Miocene. During this period, evaporite sequences, as much as 3.5 kilometers thick, were deposited over most of the Red Sea basin.
Legend

- Sedimentary Rocks of Phanerozoic Age
- Precambrian
- Cenozoic Volcanics

Generalized from ASGA-UNESCO Geological Map of Africa

Scale in kilometers

Figure 2. Generalized geologic map of the Red Sea area.
(Ross and Schlee, 1973). Because of similar evaporite deposits in the Mediterranean basin and because of the presence of Mediterranean fossil assemblages found in marine sediment interbedded with the Red Sea evaporites, Ross and Schlee (1973) concluded that the Red and Mediterranean Seas were connected during the Miocene. Near the end of the Miocene, a regressive sequence consisting of arkosic sandstones, shales, clays, and intraformational conglomerates was deposited over most of the Red Sea basin (Said, 1969).

Renewal of tectonic activity at the end of the Miocene isolated the Red Sea from the Mediterranean and the presence of marine sediment of Pleistocene age with fauna of Indian Ocean affinities indicates that the Red Sea and the Indian Ocean were connected at this time (Ross and Schlee, 1973). This second stage in the development of the Red Sea basin was accompanied by renewal of sea-floor spreading and the formation of an axial trough which is active at the present time (Young and Ross, 1974). The Miocene to Recent Aden Volcanics, which extend from the southern end of the Red Sea basin northward along the eastern margin of the basin into southeastern Turkey (Figure 2), are associated with this second phase of tectonic activity in the Red Sea area. The effusive rocks of the Aden Volcanic Belt are predominantly alkali basalts (Chase, 1969; Dickinson et al., 1969). Volcanic islands associated with the axial trough of the Red Sea and located near the Straits of Bab el Mandeb have been studied by Gass et al. (1973). These authors reported that the rocks in this group of islands are mainly alkali basalts, although they were able to identify rock types ranging from trachybasalt and trachyandesite
The rocks exposed in the Red Sea area can be classified in three general categories: Cenozoic volcanic rocks, Precambrian basement rocks, and sedimentary rocks of Phanerozoic age (Figure 2). The discussion of the volcanic rocks presented previously indicated that although their compositions are variable, they are predominantly alkali basalts. The geology of the Precambrian basement rocks is complicated and is discussed separately below.

The crystalline basement in western Arabia forms a shield area which has been separated from similar rocks in Africa as discussed previously. The tectonic history of the Arabian Shield has been provided by Dubertret (1970), Brown (1970), and Brown and Coleman (1972). More recently, Greenwood et al. (1976) and Fleck et al. (1976) have summarized the cratonic development and geochronology of the southwestern portion of the Arabian Shield. The following discussion is based on the information provided by these authors. The formation of the shield area occurred in at least three episodes that together constitute the Hijaz tectonic cycle. The first episode began with basalt to andesite to dacite volcanism and associated sedimentation and was terminated by the Aqiq orogeny. During the Aqiq orogeny, the previously deposited rocks were metamorphosed to greenschist facies and intruded by batholiths of gabbroic to quartz dioritic composition. The plutonic rocks have been dated at about 960 million years with initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of about 0.7029. The second episode involved the deposition of calc-alkaline andesitic to rhyodacitic volcanic rocks and volcanoclastics (Ablah Group) which
were subsequently metamorphosed and intruded by dioritic to granodioritic batholiths about 800 million years ago (initial $87\text{Sr}/86\text{Sr}$ ratio of 0.7028) during the Ranyah orogeny. Accompanying this orogeny was the deposition of clastics and volcanics (andesite to dacite) of the Halaban Group. These rocks, as well as the older rocks, were metamorphosed to greenschist facies and intruded by gabbroic to granitic plutons between 600 and 650 million years ago during the Yafikh orogeny. The third episode in the formation of the Arabian Shield began with the deposition of the Murdama Group which consists primarily of conglomerates overlain by siltstones and mudstones. The rocks of this group were metamorphosed to greenschist facies and intruded by granitic plutons at about 550 million years ago during the Bishah orogeny (Schmidt et al., 1973). The plutonic rocks which are associated with the three major orogenic episodes are currently exposed over extensive areas in the southern portion of the Arabian Shield. Although such detailed information is not available for the Precambrian rocks exposed along the western (African) flank of the Red Sea, it is assumed that they have similar compositions and ages as those of the southwestern portion of the Arabian Peninsula.

Most of the sedimentary rocks exposed in the Red Sea area are Miocene and younger. Sedimentary rocks of Pre-Miocene age have a patchy distribution in the area surrounding the Red Sea basin. Nubian-type sandstones of Cambrian and Ordovician age are present landward of the exposed rocks of the Precambrian basement complex in the north and extend southward where they are overlain by Permian
limestones (Brown, 1970). The Mesozoic era was almost entirely a time of non-deposition with the exception of the deposition of Middle to Late Jurassic marine clastics in the present-day Gulf of Aden area. Miocene and younger rocks are exposed along the coastal plain which lies between the fringing, crystalline highlands and the Red Sea. The basal unit is the Gebel Rusas Formation and consists of sandstones and gravels as well as coral limestones and calcareous sands (Said, 1969). The Evaporite Formation of Middle Miocene age, which overlies the Gebel Rusas Formation, is made up of discontinuous beds of gypsum having variable thicknesses. The Samh Formation overlies the evaporites and consists of arkosic sandstones, shales, clays, and intraformational conglomerates. The overlying Gabir Formation is an arenaceous series containing Pliocene fossils of Indian Ocean affinities (Brown, 1970). Deposits of Pleistocene age are predominantly alternating sequences of massive coral reefs and calcareous gravels. Said (1969) has reported alluvial deposits and gravel terraces of Recent age along the flanks of the crystalline highlands.

The Red Sea is almost entirely surrounded by crystalline rocks with the exception of the sedimentary rocks of the coastal plain (Figure 2). The area of exposed Precambrian igneous and metamorphic rocks shown in Figure 2 is more than twice that of the volcanic rocks. Furthermore, more than half of the exposed volcanic rocks are located in the extreme southwestern portion of the Red Sea area. If the clastic rocks of the coastal plain were derived from the crystalline highlands, then they are likely to contain rock and mineral fragments derived primarily from the Precambrian igneous
and metamorphic rocks.
CHAPTER III

THE TWO-COMPONENT MODEL

THE TWO-COMPONENT HYPOTHESIS FOR THE RED SEA

A study of the surficial geology of the Red Sea area indicates that there are two distinct sources for the detrital-silicate fraction of the sediment being deposited in the Red Sea basin:

(1) Sialic detritus consisting of the weathering products of the Precambrian igneous and metamorphic rocks as well as the associated Phanerozoic sedimentary rocks derived from them, and

(2) the weathering products and pyroclastic debris derived from the Cenozoic volcanics which are predominantly of basaltic composition.

These two components have characteristic and different $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and strontium concentrations. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of volcanic rocks from the Red Sea have been measured, and most of these values lie within the band $0.704 \pm 0.002$ (Dickinson et al., 1969; Dickinson, 1974) which is typical of young volcanic rocks (Faure and Powell, 1972). A value of 0.704 will be used in this study to represent the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the basaltic component.

The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of old, sialic rocks should be significantly higher than that of the basalt due to the accumulation of radiogenic $^{87}\text{Sr}$. However, it is difficult to specify the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of granitic rocks of the continental crust because there is such a great diversity in their compositions and ages. The strontium isotopic
composition of water draining continental crust can be used to estimate
the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of these rocks. Veizer and Compston (1974) calculated
an average value of $0.716 \pm 0.0033$ for previously published analyses of
fresh water primarily draining the Canadian Shield. Faure and Powell
(1972) estimate an average $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.719 for 2.5 billion-
year-old continental crust assuming an initial ratio of 0.701 and
a Rb/Sr ratio of 0.18. It is believed that the southern part of
the Arabian Shield was formed about 1050 to 550 million years ago
(Greenwood et al., 1976). These authors reported initial $^{87}\text{Sr}/^{86}\text{Sr}$
ratios ranging from 0.7029 to 0.7035 for 960 and 650 million-year-old
rocks from the Arabian Shield, respectively. Unfortunately, the data
used to calculate these ages and initial ratios were not published.
However, the present-day $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of these rocks can be
estimated from their reported initial ratios and ages by assuming
an average Rb/Sr ratio of 0.18 for continental crust as suggested by
Faure and Powell (1972). These estimated $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are 0.708
and 0.710 for the 650 and 960 million-year-old rocks of the Arabian
Shield, respectively. It has been suggested that the Arabian Shield
contains rocks which are older than 960 million years (Fleck et al.,
1973; Greenwood et al., 1976), and presumably their $^{87}\text{Sr}/^{86}\text{Sr}$ ratios
are greater than 0.710. Obviously, the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the sialic
rocks varies within narrow limits from place to place depending on
the Rb/Sr ratio and age of the rocks. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the
Arabian Shield and presumably the adjoining African Shield as well
should be less than 0.720 but greater than 0.710. For the purposes
of this study, a value of 0.715 will be used to represent the $^{87}\text{Sr}/^{86}\text{Sr}$
ratio of detritus derived from the sialic source. This value is consistent with measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of Precambrian rocks from the southern portion of the Arabian Shield (Fleck, personal communication to G. Faure).

The basaltic and sialic components have different strontium concentrations as well as different $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. The discussion of the geology of the Red Sea area presented in Chapter II indicated that the young volcanic rocks are predominantly alkali basalts. In a survey of the literature, Prinz (1967) calculated an average strontium concentration of 774 ppm for such rocks. The strontium concentration of detritus derived from the sialic source should be significantly lower than that of the volcanogenic material. Faure and Powell (1972) reported average strontium concentrations of 375 and 285 ppm for crustal rocks and granitic rocks, respectively. The strontium content of the detritus may be lower than these average values, because strontium is leached from the rocks during chemical weathering (Dasch, 1969). Therefore, the strontium concentration of the detritus derived from the older and more highly weathered sialic source may be as low as 100 ppm whereas the younger basaltic material is likely to remain relatively unchanged.

In summary, the working hypothesis of this study is that the sediment deposited in the Red Sea is a mixture of two components. The basaltic component is likely to have a $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.704 and a strontium concentration of nearly 800 ppm. The second component has been identified as detritus derived from a sialic source. Although more variable, 0.715 seems to be a reasonable
estimate of the $^{\text{87}}\text{Sr}/^{\text{86}}\text{Sr}$ ratio for this component. The strontium concentration of the detritus derived from the sialic source has been lowered due to extensive chemical weathering and may be as low as 100 ppm or less. Mixtures of detritus derived from these sources will have $^{\text{87}}\text{Sr}/^{\text{86}}\text{Sr}$ ratios and strontium concentrations that are intermediate in value between those of the components. Therefore, variations of the measured values of these parameters in the detrital fraction of the sediment should reflect changes in the proportions of volcanogenic or sialic detritus and, hence, changes in the sedimentation pattern of the Red Sea.

**THE TWO-COMPONENT MODEL**

Detritus derived from the basaltic and sialic sources becomes mechanically mixed during the depositional process. Figure 3 is a schematic representation of the mixing of these two components. The basaltic component has a strontium concentration represented by $\text{Sr}_B$ and a $^{\text{87}}\text{Sr}/^{\text{86}}\text{Sr}$ ratio which, in order to simplify the notation, is shown as $R_B$. Similarly, the strontium concentration of the sialic component is $\text{Sr}_S$, and the $^{\text{87}}\text{Sr}/^{\text{86}}\text{Sr}$ ratio is represented by $R_S$. If a factor, $f$, is defined as the fraction of the basaltic component contained in a given mixture, then the expression that relates the strontium concentration of the mixture to that of the end members is:

$$\text{Sr}_M = f \text{Sr}_B + (1 - f) \text{Sr}_S$$

(1)
Basaltic Component

\[ Sr_B, R_B \]

Sialic Component

\[ Sr_S, R_S \]

**Figure 3.** Schematic representation of the mixing of a basaltic and a sialic component.
where the subscripts M, B, and S refer to the mixture, basaltic component, and sialic component, respectively. In order to derive an equation for the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the mixture, it is necessary to obtain expressions for the numbers of atoms of each strontium isotope in a unit weight of the mixture. The number of $^{87}\text{Sr}$ atoms in a given weight can be expressed in terms of the Sr concentration as:

$$^{87}\text{Sr} = \frac{\text{Sr}^{87}\text{Ab} N}{W}$$  \hspace{1cm} (2)

where $^{87}\text{Ab}$ is the abundance of the $^{87}\text{Sr}$ isotope, $W$ is the atomic weight of strontium and $N$ is Avogadro's number. Then the total number of $^{87}\text{Sr}$ atoms in a given weight of a mixture of two components is given by the expression:

$$^{87}\text{Sr}_M = f \frac{\text{Sr}_B^{87}\text{Ab}_B N}{W_B} + (1 - f) \frac{\text{Sr}_S^{87}\text{Ab}_S N}{W_S}$$ \hspace{1cm} (3)

By similar reasoning, the number of $^{86}\text{Sr}$ atoms in a given weight of the mixture is:

$$^{86}\text{Sr}_M = f \frac{\text{Sr}_B^{86}\text{Ab}_B N}{W_B} + (1 - f) \frac{\text{Sr}_S^{86}\text{Ab}_S N}{W_S}$$ \hspace{1cm} (4)

Because the abundances of $^{87}\text{Sr}$ in the two components are different, the atomic weights of strontium and the abundances of $^{86}\text{Sr}$ are also different for these components. However, these differences are small
and amount to less than 1.0% in the case of the $^{86}\text{Sr}$ abundances and about 0.08% for the atomic weights when the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios vary between 0.700 and 0.800 (Faure and Boger, 1976). Therefore, negligible error will be introduced if the $^{86}\text{Sr}$ abundances and the strontium atomic weights are assumed to be equal. This simplifying assumption allows Equations (3) and (4) to be rewritten as:

$$87_{\text{Sr}} = f \frac{\text{Sr}_B \times 87_{\text{Ab}} N}{W} + (1 - f) \frac{\text{Sr}_S \times 87_{\text{Ab}} S}{W} \quad (5)$$

and as:

$$86_{\text{Sr}} = f \frac{\text{Sr}_B \times 86_{\text{Ab}} N}{W} + (1 - f) \frac{\text{Sr}_S \times 86_{\text{Ab}} N}{W} \quad (6)$$

Equations (5) and (6) are algebraically divided to yield:

$$R_M = \frac{f \text{Sr}_B \times 87_{\text{Ab}} B + (1 - f)(\text{Sr}_S \times 87_{\text{Ab}} S)}{f \text{Sr}_B \times 86_{\text{Ab}} B + (1 - f)(\text{Sr}_S \times 86_{\text{Ab}} S)} \quad (7)$$

All the terms in Equation (7) can be divided by $86_{\text{Ab}}$ which results in the expression:

$$R_M = \frac{f \text{Sr}_B \times (87_{\text{Ab}}/86_{\text{Ab}})_B + (1 - f)(\text{Sr}_S \times (87_{\text{Ab}}/86_{\text{Ab}})_S)}{f \text{Sr}_B + (1 - f) \text{Sr}_S} \quad (8)$$

Because

$$87_{\text{Ab}}/86_{\text{Ab}} = 87_{\text{Sr}}/86_{\text{Sr}} = R \quad (9)$$
and

\[ \text{Sr}_M = f \text{Sr}_B + (1 - f) \text{Sr}_S \]  

\hspace{1cm} (10)

Equation (8) reduces to:

\[ R_M = f \frac{\text{Sr}_B R_B}{\text{Sr}_M} + (1 - f) \frac{\text{Sr}_S R_S}{\text{Sr}_M} \]  

\hspace{1cm} (11)

Equations (10) and (11) relate the strontium concentration and isotope ratio of a mixture to that of the two components by means of the mixing factor, \( f \). These expressions are combined by eliminating \( f \). Equation (10) can be rewritten as:

\[ f = \frac{\text{Sr}_B - \text{Sr}_S}{\text{Sr}_M - \text{Sr}_S} \]  

\hspace{1cm} (12)

This expression is then substituted into Equation (11) and simplified to yield:

\[ R_M = \left[ \frac{\text{Sr}_B \text{Sr}_S (R_S - R_B)}{\text{Sr}_B - \text{Sr}_S} \right] \frac{1}{\text{Sr}_M} + \left[ \frac{\text{Sr}_B R_B - \text{Sr}_S R_B}{\text{Sr}_B - \text{Sr}_S} \right] \]  

\hspace{1cm} (13)

The strontium concentrations and isotope ratios of the end members are assumed to be constant for a particular mixing regime. Therefore, the terms in brackets in Equation (13) are constants, and the equation can be restated in a more general form:

\[ R_M = \frac{a}{\text{Sr}_M} + b \]  

\hspace{1cm} (14)
Equation (14) is the equation of a hyperbola when plotted in coordinates of $^{87}\text{Sr} / ^{86}\text{Sr}$ ratio and strontium concentration. It is useful to note that Equation (14) can also be regarded as a linear function when plotted in coordinates of $^{87}\text{Sr} / ^{86}\text{Sr}$ ratio and $1/\text{Sr}$ concentration. A more detailed discussion of the functional dependencies of Equation (14) is presented in Appendix A.

Many sedimentary processes involve the mixing of two or more components of contrasting chemical compositions. In many cases, these processes are dominated by two components and can be studied in terms of a two-component system. Equations (10) and (11) define a two-component model in terms of the concentration and composition of strontium. Implicit in this model are the conditions that the end members have characteristically different strontium concentrations and compositions and that the mixtures of these two components have remained closed systems after mixing has occurred. If these conditions are satisfied, then the two-component model can be applied to the study of sediment that results from the mixing of detritus derived from two different source regions.

The application of the two-component model to a particular geologic problem involves measuring the Sr concentrations and $^{87}\text{Sr} / ^{86}\text{Sr}$ ratios of a suite of samples believed to be the result of mixing. The observed $^{87}\text{Sr} / ^{86}\text{Sr}$ ratios are plotted as a function of the inverse Sr concentrations, and a line is fitted to these data points by least-squares regression. The goodness of fit of the regression line to the data points is indicative of the validity of the two-component hypothesis for these samples. The
The equation of the fitted line provides values for the constants, $a$ and $b$, of the mixing equation (14). In most cases, and particularly in this study, the strontium concentrations and isotope ratios of the end members cannot be measured directly. However, it is generally possible to make reasonable estimates of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of these components, as discussed previously. The Sr concentrations of the end members are calculated by substitution of their assumed $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the mixing equation derived from the analytical results. Once the Sr concentrations of the end members have been determined, Equation (10) is used to calculate the fraction, $f$, of the basaltic component contained in each sample for which Sr concentrations have been measured. The basaltic content of the detrital-silicate fraction of the sediment samples are plotted as a function of stratigraphic position. The systematic variation of the basalt concentration is used to define layers that are enriched in either the basaltic or sialic component. These litho-stratigraphies are then interpreted in terms of the depositional history of the sedimentary basin.

Equations can be derived that relate the concentration of strontium to that of other elements in a mixture of two components. The strontium concentration of a two-component mixture is related to concentrations of the end members by the expression:

$$\text{Sr}_M = f \text{Sr}_B + (1 - f) \text{Sr}_S$$  \hspace{1cm} (15)

This relationship applies to the concentrations of any element that
satisfies the conditions of the two-component model. If $Y$ is used to symbolize the concentration of an element, then Equation (15) becomes:

$$Y_M = f Y_B + (1 - f) Y_S$$  \hspace{1cm} (16)

These two expressions are combined by solving Equation (15) for $f$ and substituting into Equation (16):

$$Y_M = Sr_M \left[ \frac{Y_B - Y_S}{Sr_B - Sr_S} \right] + \left[ \frac{Y_S Sr_B - Y_B Sr_S}{Sr_B - Sr_S} \right]$$  \hspace{1cm} (17)

The terms in brackets in Equation (17) are constants, and the expression can be restated in a more general form:

$$Y_M = c Sr_M + d$$  \hspace{1cm} (18)

This is the equation of a straight line where $c$ is the slope and $d$, the $y$-intercept. Equation (18) implies that there should be a linear relationship between the concentrations of any two elements in a mixture of two components. However, systematic deviations will result if there has been selective addition or removal of one or both of the elements after mixing has occurred, or if a varying amount of a third component is present.
SYSTENATICS OF THE TWO-COMPONENT MODEL

Any two end members are uniquely defined by their characteristic $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and Sr concentrations. A mixture of these two components will have $^{87}\text{Sr}/^{86}\text{Sr}$ ratio and Sr concentration that has an intermediate value relative to those of the end members. The mixing hyperbola or line is the locus of all possible combinations of the two components. Estimates of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the mixing components for the Red Sea have been made, and the most likely values of these ratios are 0.704 and 0.715 for the basaltic and sialic components, respectively. Compared to the basaltic source, the sialic component has had a long and complex history, and this source may not have the same degree of homogeneity.

If the Sr concentration of the basaltic component is assumed to be constant, then a series of mixing hyperbolas and lines can be generated by varying the Sr concentration of the sialic end-member. In Figures 4 and 5 the Sr concentration of the basaltic component has been fixed at 800 ppm, and Equation (14) was used to construct mixing hyperbolas and lines for different values of the Sr concentration of the sialic component. Both figures have been contoured in terms of the $f$-values which were calculated by means of Equation (10) and have been expressed as weight percent basalt. Figures 4 and 5 are visual presentations of the functional relationships of the two-component model. For example, a detrital-silicate
Figure 4. Mixing hyperbolas generated by varying the Sr concentration of the sialic component from 30 to 800 ppm. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the sialic and basaltic components were assumed to be 0.715 and 0.704, respectively. The Sr concentration of the basaltic component was fixed at 800 ppm. The hyperbolas were contoured in 10% increments of the basaltic content.
Figure 4.
Sr Concentration, ppm
Figure 5. Mixing lines generated by varying the Sr concentration of the sialic component from 30 to 800 ppm. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the sialic and basaltic component were assumed to be 0.715 and 0.704, respectively. The Sr concentration of the basaltic component was fixed at 800 ppm. The lines were contoured in 10% increments of the basaltic content.
Figure 5.

\[ \frac{\text{Sr}}{\text{Sr}} \text{ Concentration, ppm} \times 10^{-3} \]

\[ 87^{\text{Sr}}/86^{\text{Sr}} \]

Concentration, ppm x 10^{-3}
sample that is a mixture of the basaltic component, as defined above, and a sialic component with a Sr concentration of 70 ppm will lie on the mixing hyperbola and mixing line designated 70 ppm in Figures 4 and 5. The position of the point on the mixing curves depends on the concentration of the basaltic component of the sample. If such a sample contained 30 percent by weight of the basaltic component, then it would plot at the intersection of the 70 ppm mixing curve and the 30 percent f-value contour.

Each of the mixing hyperbolas and lines in Figures 4 and 5, respectively, represent mixtures of two components that are uniquely defined in terms of their $^{87}\text{Sr} / ^{86}\text{Sr}$ ratios and Sr concentrations. In reality, these values may vary within narrow limits due to the heterogeneity of the source materials. Therefore, it would be more accurate to show the compositions of the end members in Figures 4 and 5 as areas rather than points. Mixtures of two such heterogeneous components should have $^{87}\text{Sr} / ^{86}\text{Sr}$ ratios and Sr concentrations that are randomly distributed with respect to a "best fit" mixing line or hyperbola. In order to facilitate the subsequent interpretation of the results, the end members are assumed to have unique values of their $^{87}\text{Sr} / ^{86}\text{Sr}$ ratios and Sr concentrations. This simplifying assumption does not significantly affect the qualitative interpretation of the results.

Systematic departures from the two-component model may arise because of the presence of additional components. These components can be classified into three general categories based on their strontium compositions:
(1) components that contain no strontium,

(2) components that contain strontium of known isotopic composition and concentration, or

(3) components for which neither the strontium composition nor the strontium concentration are known.

Examples of the first type include biogenic opal and chemical precipitates such as iron oxides, and sulfide minerals formed either during or after deposition. These components, which contain no strontium, act as diluents causing the measured strontium concentrations to be underestimated. The lowered Sr concentrations of samples that contain significant amounts of this type of material will fall to the left of the two-component hyperbola and to the right of the mixing line, since the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio is unchanged. A specific example of the presence of this type of an additional component will be discussed in detail in Chapter V.

The presence of additional components that contain strontium is a more serious complication. In some cases, the strontium composition and concentration of the additional component can be estimated, thereby simplifying the interpretation. In order to illustrate this type of mixing, assume that a sialic and basaltic component generate a 70 ppm mixing hyperbola and line as shown in Parts A and B of Figure 6, respectively. Furthermore, assume the presence of a third component which has the same $^{87}\text{Sr}/^{86}\text{Sr}$ ratio as the sialic end-member but a strontium concentration of 150 ppm. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and strontium concentrations of the mixtures of these three components will lie in the band defined by the 70 ppm line.
Figure 6. Example of three-component mixing where the third component has a different, but known, Sr concentration.
and 150 ppm mixing curves in Figure 6. One such mixture is labeled M. Due to the nature of the mixing diagrams, point M must lie on a secondary-mixing hyperbola (Part A) or line (Part B) which includes the point representing the third component. In Part B of Figure 6 this secondary-mixing line has been extended so as to intersect the 70 ppm mixing line at the point labeled I. This intersection gives the strontium composition of the sample before it was mixed with an unknown amount of the third component. The fraction of the third component contained in the sample can be calculated by means of the expression:

\[ \text{Sr}_M = k \text{Sr}_T + (1 - k) \text{Sr}_I \]  

(19)

where \( k \) is the fraction of the third component contained in the mixture, \( \text{Sr}_M \) is the measured Sr content, \( \text{Sr}_T \) is the concentration of the third component, and \( \text{Sr}_I \) is the original strontium concentration.

In the foregoing discussion, the additional component was assumed to have the same \(^{87}\text{Sr}/^{86}\text{Sr} \) ratio as one of the other end members in order to simplify the presentation. However, it is much more likely that both the \(^{87}\text{Sr}/^{86}\text{Sr} \) ratio and strontium concentration of the third component will be different from those of the other end members. Such additional components with different, but known, strontium compositions are authigenic minerals precipitated from sea water. The \(^{87}\text{Sr}/^{86}\text{Sr} \) ratio of the oceans is constant and averages 0.7091 at the present time (Faure and Powell, 1972).
Because strontium isotopes are not fractionated by natural processes, the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of a strontium-bearing mineral precipitated from sea water will be the same as that of the water. Calcium carbonate is an example of a possible third component of known strontium composition and concentration. However, calcium carbonate can be removed from the sediment by leaching with dilute HCl (Chapter IV), unless the objective of the study includes consideration of the mixing of carbonate and detrital fractions (Church, 1971).

More important are the authigenic phases which cannot be removed chemically such as zeolites or authigenic feldspars.

Phillipsite is a strontium-bearing zeolite associated with the alteration of volcanic debris, and the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of marine phillipsite has been shown to reflect the strontium composition of the sea water with which it has interacted (Pushkar and Peterson, 1967). Therefore, it is possible that the noncarbonate fraction of marine sediment could consist of three components having different compositions and concentrations of strontium: detrital-silicates derived from a sialic and basaltic source, as discussed previously, as well as zeolite alteration-products of the basaltic detritus. In order to illustrate such an occurrence, assume that the two detrital-silicate components generate a 70 ppm mixing hyperbola and line as before. If Recent sediment is being considered, then the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the phillipsite will be 0.7091, but the Sr concentration is unknown. Furthermore, if two or more samples having the same composition in terms of the detrital-silicate fractions contain variable amounts of the phillipsite
component, then these samples will define a secondary-mixing line that includes the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio and Sr concentration of the phillipsite. Two such mixtures are indicated by the points $M_3$ and $M_4$ in Figure 7. The strontium concentration of the phillipsite is determined by extending the secondary-mixing line in Part B to the point where the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio becomes 0.7091. The extension of the secondary-mixing line to intersect the 70 ppm line at the point labeled $I_2$ gives the strontium concentration of the sample before it was mixed with an unknown amount of the phillipsite. The fraction of the phillipsite contained in the mixtures can be determined by means of Equation (19).

The presence of additional components that contain strontium of unknown composition is the most general case. If these components are present in sufficient quantity and in a proper form to permit physical or chemical separation, then the composition and concentration of strontium can be determined analytically. When this is not possible, the interpretation of the presence of such additional components requires that specific conditions exist.

The first condition is the assumption that the two dominant components are pervasive throughout the length of time represented by the samples being analyzed. Secondly, the occurrence of an additional component is sporadic, and the concentration of this end member varies from zero to a maximum and back to zero over a relatively short interval of time. If the samples have been collected in intervals that are of the same order as the occurrence of the third component, then it is likely that those samples which contain variable
Figure 7. Example of three-component mixing where the third component has a different $^{87}\text{Sr}/^{86}\text{Sr}$ ratio and Sr concentration.
amounts of the third component will have similar concentrations of the dominant end-members. Therefore, samples which are variously mixed with the third component will tend to define secondary-mixing hyperbolas and lines as discussed previously. Furthermore, if there are two occurrences of the third component, then it is possible to generate two secondary-mixing lines. The intersection of these secondary-mixing lines will give the approximate $^{87}\text{Sr}/^{86}\text{Sr}$ ratio and Sr concentration of the third component, as shown in Part B of Figure 7. Once the strontium concentration of the additional component has been estimated, Equation (19) can be used to calculate the fraction of the third component contained in each sample, as before.

To this point in the discussion, all the end members used to illustrate the two-component model have had either high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and low Sr concentrations or low ratios and high concentrations. These combinations of values generate mixing hyperbolas and lines that are oriented as shown in Figures 4, 5, and 6. However, it is possible for a component to have a high $^{87}\text{Sr}/^{86}\text{Sr}$ ratio and a high Sr concentration. If this component is mixed with another component that has relatively low values of those parameters, then these end members will generate a mixing hyperbola and line that have different orientations. For example, in Figure 7 the point labeled $I_2$ is located on the mixing hyperbola and line defined by points $M_3$ and $M_4$ and containing the component labeled $T$. In this case, the $T$ component has a higher Sr concentration and $^{87}\text{Sr}/^{86}\text{Sr}$ ratio than $I_2$. Thus, the hyperbola is inverted with respect to the
70 ppm hyperbola, and the mixing line between $I_2$ and $T$ in Part B has a negative slope. The two-component model has been applied to such a mixing regime by Faure et al. (1974), and their results show that the interpretation is unaffected by the orientation of the mixing curves.
The samples used in this study were obtained with the cooperation of Dr. David A. Ross of the Woods Hole Oceanographic Institute. The sediment samples were collected serially along the lengths of piston cores raised in the Red Sea during R/V Chain cruises 61 and 100. The locations of the four cores sampled for this study are shown in Figure 8. The dry weight of the samples collected from these cores averaged about 10 grams each.

Core number 3P was collected at station 3 on leg 3 of R/V Chain cruise 100 in the median valley of the Red Sea at latitude 18°09'N and longitude 39°53'E. The water depth at this site was approximately 1390 meters. A total of 43 samples were taken along the 798 cm length of core 3P in 20 cm intervals, where possible. The lowermost 60 cm of the core were disturbed during the coring operation.

Core number 153P was collected during R/V Chain cruise 61 in the median valley at latitude 19°43'N and longitude 38°41'E. The water depth at this site was approximately 2700 meters. A total of 37 samples were taken in 20 cm intervals along the 755 cm length of this core. The lower 100 cm of core 153P were disturbed during coring.
Figure 8. Map of the Red Sea showing the locations of the piston cores sampled for this study as well as the location of the hot brine area.
Core number 4P was collected at station 5 of R/V Chain cruise 100 on the western flank of the median valley at latitude 19°05'N and longitude 39°59.5'E where the water depth was approximately 300 meters. This core was 577 cm long and 53 samples were taken in 10 cm intervals.

Core 154P was collected during R/V Chain cruise 61 in the median valley at latitude 19°34'N and longitude 38°59.5'E. The water depth at this site was 1276 meters. A total of 48 samples were taken from this 792 cm-long core. It was not possible to collect samples in regular intervals from core 154P due to its fragmented condition.

SAMPLE PREPARATION

All the samples were dispersed in doubly-distilled, de-ionized water and washed through a 100-mesh (149μm) Nylon screen in order to remove the tests of organisms for future study. More than 95% of the non-biogenic sediment passed through the screen. The remaining portion consisted primarily of semi-lithified material as well as mineral grains and fragments of volcanic glass. The sediment particles were disaggregated in an ultrasonic vibrator and resieved. The dry weights of the -100 mesh fractions were determined, and the samples were leached overnight with purified 2N HCl to remove the calcium carbonate fraction of the sediment. The non-carbonate fractions were dried and weighed, and the observed weight loss was used to calculate the carbonate content.
of each sample (Appendix B).

Each non-carbonate sample was formed into a pellet for X-ray analysis by combining approximately 3 grams of sample with 7 grams of boric acid crystals as a backing. The pellets were pressed under 12 tons per square inch pressure by means of a hydraulic press.

Aliquots of the non-carbonate samples were prepared for isotope ratio analysis by digesting 0.25 to 0.50 grams of the sample in a 1:5 solution of H$_2$SO$_4$ and HF. The resulting solutions were filtered using analytical grade filter paper (S and S #589) and $^{89}$Sr was added as a tracer. The strontium was separated by cation-exchange chromatography (Dowex 50 W-8X resin) using 2.25 HCl as the eluant. Precleaned, 15 ml Polyethylene beakers were used to collect the strontium fractions which were identified on the basis of their $^{89}$Sr activity. The two most active fractions were combined and evaporated to dryness in 5 ml Pyrex beakers. Perchloric acid was added to the samples to remove organic residue. The samples were redissolved in 2N HNO$_3$ and evaporated onto the pre-cleaned Ta filaments of the mass spectrometer. The liquid was held in contact with the filament by means of a Vycor needle mounted on a syringe, and evaporation was promoted by passing a current of about 1 Amp through the filaments.

**ELEMENT CONCENTRATIONS**

The elemental concentrations were determined by X-ray fluorescence. The abundances of Sr, Rb, Ca, and Fe were measured using an air-path
spectrometer (Diano Corporation, XRD-6). The concentration of Si was determined by means of a vacuum-path spectrometer (Diano Corporation, XRD-700). The operating conditions for these analyses are listed in Table 1. In general, the concentrations of these elements were determined by generating calibration curves for each element using U. S. Geological Survey rock standards (Flanagan, 1973). Matrix corrections were made by means of Compton-scattered radiation (Reynolds, 1963) of the target tube used for the analysis. All of the results were corrected to specific values of an intralaboratory standard (#48) in order to minimize time-dependent instrumental variations. Table 2 lists the concentrations of Sr and Rb as well as the abundances of the oxides of Ca, Si, and Fe used to construct the calibration curves.

**Silicon**

The peak intensities of silicon were measured at 108.03° (2Θ) using an EDT analyzing crystal. Because Chromium radiation was used in this determination, the intensity of the Chromium Compton-scattered radiation was used to correct for matrix effects. The intensities of the Compton-scattered radiation were measured at 109.20° (2Θ) based on the Compton-scattered radiation spectrum shown in Figure 9. The ratios of the Si to Cr-Compton peak intensities of 5 standards were measured four times each. All measurements were normalized to a value of 3.846 for the Si to Cr-Compton peak intensity ratio of intralaboratory standard #48 so as to reduce time-dependent variations. The averages of these corrected values were plotted as a function of the reported SiO₂ concentrations (Figure 10). Although calibration
Table 1. Operating conditions for X-ray fluorescence analysis.

<table>
<thead>
<tr>
<th>Element</th>
<th>Analyzing Crystal</th>
<th>X-ray Tube Target</th>
<th>kV</th>
<th>mA</th>
<th>Soller Slit</th>
<th>Detector Type</th>
<th>Setting</th>
<th>E</th>
<th>dE</th>
<th>Gain</th>
<th>Counting Time (sec.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sr</td>
<td>LiF(220)</td>
<td>Mo</td>
<td>65</td>
<td>55</td>
<td>0.01</td>
<td>Scint.</td>
<td>6.10</td>
<td>3</td>
<td>16</td>
<td>16</td>
<td>200</td>
</tr>
<tr>
<td>Rb</td>
<td>LiF(220)</td>
<td>Mo</td>
<td>65</td>
<td>55</td>
<td>0.01</td>
<td>Scint.</td>
<td>6.58</td>
<td>3</td>
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<td>16</td>
<td>200</td>
</tr>
<tr>
<td>Ca</td>
<td>LiF(220)</td>
<td>Mo</td>
<td>65</td>
<td>55</td>
<td>0.02</td>
<td>SPG-6</td>
<td>9.20</td>
<td>3</td>
<td>16</td>
<td>16</td>
<td>200</td>
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<tr>
<td>Fe</td>
<td>LiF(220)</td>
<td>Mo</td>
<td>65</td>
<td>55</td>
<td>0.01</td>
<td>Scint.</td>
<td>7.67</td>
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<td>16</td>
<td>16</td>
<td>100</td>
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<tr>
<td>Si</td>
<td>EDT</td>
<td>Cr</td>
<td>55</td>
<td>50</td>
<td>0.01</td>
<td>SPG-9C</td>
<td>1900</td>
<td>9</td>
<td>1</td>
<td>3</td>
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Table 2. Concentrations of standards.

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<tr>
<th>Standard</th>
<th>Sr ppm</th>
<th>Rb ppm</th>
<th>CaO %</th>
<th>SiO$_2$ %</th>
<th>FeO %</th>
<th>Fe$_2$O$_3$ %</th>
<th>Total Fe as Fe$_2$O$_3$ %</th>
</tr>
</thead>
<tbody>
<tr>
<td>G-2$^1$</td>
<td>479</td>
<td>168</td>
<td>1.94</td>
<td>69.11</td>
<td>1.45</td>
<td>1.08</td>
<td>2.69</td>
</tr>
<tr>
<td>AGV-1$^1$</td>
<td>657</td>
<td>67</td>
<td>4.90</td>
<td>59.00</td>
<td>2.05</td>
<td>4.51</td>
<td>6.79</td>
</tr>
<tr>
<td>BCR-1$^1$</td>
<td>330</td>
<td>46.6</td>
<td>6.92</td>
<td>54.50</td>
<td>8.80</td>
<td>3.68</td>
<td>13.46</td>
</tr>
<tr>
<td>GSP-1$^1$</td>
<td>233</td>
<td>254</td>
<td>2.02</td>
<td>67.38</td>
<td>2.31</td>
<td>1.77</td>
<td>4.34</td>
</tr>
<tr>
<td>48</td>
<td>199</td>
<td>109</td>
<td></td>
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<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Wyatt Fm.$^2$

|        |       |       |       |       |       |               |                           |
| 236     | 1.80  |       |       | 3.4   | 1.6   | 5.38          |                           |
| 61      | 1.8   |       |       | 2.2   | 1.1   | 3.54          |                           |
| 108     | 1.9   |       |       | 3.2   | 1.8   | 5.36          |                           |
| 129     | 2.0   |       |       | 1.8   | 1.1   | 3.10          |                           |

$^1$Flanagan, 1973

$^2$Ford, A. B., private communication to G. Faure.
Figure 9. Chromium $k_\alpha$, Compton-scattered radiation intensity spectrum measured using a LiF(220) analyzing crystal.
Figure 10. SiO$_2$ calibration diagram. The Si/Cr-Compton peak intensity ratios were measured using an EDT analyzing crystal and normalized to 3.846 for intra-laboratory standard #48. The coefficient of correlation is 0.998.
$\text{SiO}_2 = (12.40226)(\text{Si}/\text{Cr-C}) + 13.23$
curves are generally not linear over a wide range of concentrations, linearity is approximated in narrow intervals. Therefore, a straight line was fitted to the data points in Figure 10 by least-squares regression and has the equation:

$$\text{SiO}_2 = (12.40226)(\text{Si/Cr} - 0) + 13.23$$  \hspace{1cm} (20)

The coefficient of correlation of the fitted line is 0.998.

The silicon and Cr-Compton peak intensities of the non-carbonate samples were determined in duplicate. The ratios of these measurements were used in Equation (20) to calculate the SiO\textsubscript{2} concentration, and the averages of the duplicate determinations are tabulated in Appendix B.

Iron

The reported FeO and Fe\textsubscript{2}O\textsubscript{3} concentrations in Table 2 were recalculated in terms of the total Fe and expressed as Fe\textsubscript{2}O\textsubscript{3}, because this is the form most likely to be present in weathering products. The molecular weights of FeO and Fe\textsubscript{2}O\textsubscript{3} are 71.8464 and 159.6922, respectively. Therefore, FeO contains 77.7311% iron, and Fe\textsubscript{2}O\textsubscript{3} contains 69.9433% iron. The amount of Fe\textsubscript{2}O\textsubscript{3} that can be formed from FeO is given by the expression:

$$\text{Fe}_2\text{O}_3 = k \text{FeO}$$  \hspace{1cm} (21)

where \(k\) is the ratio of 77.7311% to 69.9433%, or 1.111344. The
Figure 11. $\text{Fe}_2\text{O}_3$ calibration diagram. The Fe/Mo-Compton peak intensity ratios were measured using a LiF(220) analyzing crystal and normalized to 0.5188 for the intralaboratory standard #48. The coefficient of correlation is 0.993.


\[ \text{Fe}_2\text{O}_3 = (11.4073)(\text{Fe/Mo-C}) + 0.94 \]

**Figure 11.**
reported concentrations of FeO were recalculated in terms of Fe$_2$O$_3$ by means of Equation (21). These recalculated values were added to the reported Fe$_2$O$_3$ concentrations to yield the total iron content expressed as Fe$_2$O$_3$ in Table 2.

The iron peak intensities of the standards were measured at 85.77° (2θ) and divided by the Mo-Compton peak intensities which were measured at 30.00° (2θ) (Gunner, 1971). All peak intensities were measured using a LiF (220) analyzing crystal, and the intensity ratios were normalized to 0.5188 for the intralaboratory standard #48. Three determinations of this ratio for ten standards were averaged, and these results were plotted as a function of the Fe$_2$O$_3$ concentrations in Figure 11. A line was fitted to these data points by least-squares regression and has the equation:

$$\text{Fe}_2\text{O}_3 = (11.4073)(\text{Fe}/\text{Mo} - C) + 0.9445$$  \hspace{1cm} (22)

The coefficient of correlation of the fitted line is 0.993.

The observed iron and Mo-Compton peak intensities of the non-carbonate samples were used in Equation (22) to calculate the Fe$_2$O$_3$ concentrations. The average of duplicate determinations are listed in Appendix B.

Calcium

The calcium peak intensities of 9 standards were measured four times and divided by the Mo-Compton intensities. The peak intensities were measured using a LiF (220) analyzing crystal, and the intensity
ratios were normalized to $4.199 \times 10^{-2}$ for the intralaboratory standard #48. The averages of the corrected ratios were plotted as a function of the reported CaO concentrations (Figure 12). A line was fitted to these data points by least-squares regression and has the equation:

$$\text{CaO} = (38.7307)(\text{Ca}/\text{Mo}_k - C) + 0.422978$$  \hspace{1cm} (23)

The coefficient of correlation of the fitted line is 0.989.

The observed calcium and Mo-Compton peak intensities of the non-carbonate samples were used in Equation (23) to calculate the CaO concentrations, and the average of duplicates are tabulated in Appendix B.

**Strontium**

The strontium peak intensities were measured at $35.85^0$ (20) and divided by the observed Mo-Compton intensities. The peak intensity ratios were normalized to 0.3151 for intralaboratory standard #48.

The average of twelve determinations of this ratio for each of four standards was plotted as a function of the reported Sr concentration in Figure 13. A line was fitted to these data points by least-squares regression resulting in the calibration equation:

$$\text{Sr} = 574.115 (\text{Sr}/\text{Mo}_k - C) - 1.02$$  \hspace{1cm} (24)

The coefficient of correlation of the fitted line is 0.999.
Figure 12. CaO calibration diagram. The Ca/Mo-Compton peak intensity ratios were measured using a LiF(220) analyzing crystal and normalized to $4.199 \times 10^{-2}$ for the intralaboratory standard #48. The coefficient of correlation is 0.989.
CaO$ = (38.7307)(\text{Ca/Mo-C}) + 0.42$

Figure 12.
Figure 13. Sr calibration diagram. The Sr/Mo-Compton peak intensities were measured using a LiF(220) analyzing crystal and normalized to 0.3151 for the intralaboratory standard #48. The coefficient of correlation is 0.999.
Figure 13. Sr/Mo-Compton Sr Concentration, ppm

Sr = 574.115 (Sr/Mo) - 1.02
The Sr to Mo-Compton peak intensity ratio was determined in duplicate for each of the non-carbonate samples and used in Equation (24) to calculate the Sr concentrations. The averages of these results are tabulated in Appendix B.

Rubidium

The rubidium peak intensities were measured in duplicate at 37.99±0.28 using a LiF (220) analyzing crystal and divided by the observed Mo-Compton intensities for each of four standards. The measured peak-intensity ratios were normalized to 0.1438 for the intralaboratory standard #48. The corrected averages of these ratios were plotted as a function of their reported rubidium concentrations as shown in Figure 14. A line was fitted to these data points by least-squares regression and has the equation:

$$Rb = 755.6 \left( \frac{Rb}{Mo} - 0 \right)$$  \hspace{1cm} (25)

The coefficient of correlation of the fitted line is 0.999.

The rubidium and Mo-Compton peak intensities of the non-carbonate samples were determined in duplicate, and the ratios of these values were used in Equation (25) to calculate the Rb concentrations. The averages of the determinations are listed in Appendix B.
Figure 14. Rb calibration diagram. The Rb/Mo-Compton peak intensities were measured using a LiF(220) analyzing crystal and normalized to 0.1438 for the intralaboratory standard #48. The coefficient of correlation is 0.999.
Slope = 755.6
REPRODUCIBILITY OF THE RESULTS

The reproducibility of the elemental concentrations was estimated by means of the expression:

\[
S = \sqrt{\frac{\sum (\text{deviation})^2}{\text{degrees of freedom}}}
\]  

(26)

where the deviation is the difference between the observed concentration and the average of replicate determinations, and the number of degrees of freedom is the difference between the number of measurements and the number of samples. All error estimates were recalculated to the 90% confidence level by means of Student's t-distribution. The calculated reproducibilities, stated in units of concentration, are: Sr ± 5 ppm, Rb ± 2 ppm, CaO ± 0.10%, Fe₂O₃ ± 0.20%, and SiO₂ ± 3%.

DEAD-TIME CORRECTION

The samples which exhibited the highest counting rates were used to test for loss in detector efficiency due to counter saturation. A dead-time correction equation was used:

\[
N_T = \frac{N_0}{(1 - N_0 \sigma)}
\]  

(27)

where \( N_0 \) is the observed counting rate expressed in counts per
second, $\sigma$ is a constant equal to $0.5 \times 10^{-6}$ seconds, and $N_T$ is the corrected counting rate. None of the samples showed significant changes in counting rates.

MINERAL CONCENTRATIONS

The absolute concentrations of quartz, feldspar, amphibole, illite, kaolinite, and chlorite were determined for selected samples by X-ray diffraction (Diano Corporation, XRD-6) using Ni-filtered, Cu k-alpha radiation. The minerals were identified from strip chart recordings (Biscaye, 1964 and 1965; Müller and Stoffers, 1974). The peak intensities were measured at the following 2θ angles: chlorite, 6.30°; illite, 8.84°; amphibole, 10.50°; kaolinite, 12.33°; and quartz, 26.66°. Glycolated samples showed insignificant amounts of montmorillonite. The feldspar intensities were measured as peak areas from 27.5° to 28.5° (2θ). Independent calibration curves were generated for each mineral by the addition method (quartz and feldspar) or the dilution method (amphibole, illite, kaolinite, and chlorite). Compton-scattered radiation intensities were used to correct for matrix effects (Cosgrove and Sulaiman, 1973).

The addition method has the advantage of minimizing matrix effects but requires sufficient material to make at least three pellets (i.e. about 10 grams). This technique involves combining known weights of a particular mineral with the sample being analyzed. The weight percent of the mineral added to the sample can be
calculated and plotted as a function of the observed peak intensity. For example, the observed feldspar peak intensity of an unaltered sample (no added feldspar) is plotted as having zero weight-percent feldspar added in Figure 15. Two other fractions of the sample were thoroughly mixed with 6.21 and 11.55 percent additional feldspar. The observed feldspar peak intensities for these samples were plotted in Figure 15. A line was fitted to these results by least-squares regression, and the y-intercept (expressed as a positive number) gives the feldspar content of the unaltered sample, in this case, 13.4 percent. The mixed samples contain the sum of the feldspar concentration of the unaltered sample plus the amount of feldspar added. A few samples were sufficiently large to permit the use of this technique, and they were used to generate the calibration curves for feldspar (Figure 16) and quartz (Figure 17).

The calibration curves for the other minerals were constructed by means of a dilution method. Known weights of each mineral were diluted with varying amounts of reagent-grade CaCO₃. The mineral and Mo-Compton peak intensities were measured, and the ratios were plotted as a function of the calculated mineral concentrations. Calibration lines were fitted to the results by least-squares regression. Calibrations for amphibole (Figure 18), illite (Figure 19), kaolinite (Figure 20), and chlorite (Figure 21) were obtained in this way. All the calibrations pass through the origin and are linear over the range of concentrations observed in this study. Replicate determinations of the mineral peak-intensities and Mo-Compton intensities were determined for selected samples. The
Figure 15: An example of the addition method used to determine the feldspar content of a sample.
Figure 16. Feldspar calibration diagram. Coefficient of correlation is 0.855.
Figure 17. Quartz calibration diagram. The coefficient of correlation is 0.860.
Figure 18. Amphibole calibration diagram. The coefficient of correlation is 0.998.

Amphibole, wt. %

slope = 0.005501

Intensity / Mo-Compton x10^6

100 200 300 400 500

30 25 20 15 10 5
slope = 0.03688

Figure 19. Illite calibration diagram. The coefficient of correlation of the fitted line is 0.978.
Figure 20. Kaolinite calibration diagram. The coefficient of correlation is 0.954.
Figure 21. Chlorite calibration diagram. The coefficient of correlation is 0.989.
ratios of these observed values were used in conjunction with the slopes of the respective calibration lines to determine the mineral concentrations tabulated in Appendix B.

Replicate determinations of the mineral concentrations for each sample were used to calculate standard deviations. These standard deviations were divided by the average mineral concentrations to obtain the percent error for each analysis. The percent errors calculated in this manner for all the samples were averaged for each mineral, and these average values are: quartz, ± 10%; feldspar, ± 10%; amphibole, ± 14%; illite, ± 4%; kaolinite, ± 7%; and chlorite, ± 6%.

STRONTIUM ISOTOPE RATIO ANALYSIS

The isotopic composition of strontium was determined on a solid-source, triple-filament mass spectrometer (Nuclide Corporation, Model 12-90-S) equipped with a computerized data-acquisition system. All of the measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were corrected for fractionation to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ by means of a factor:

$$f = \frac{2(\text{Sr}^{86}/\text{Sr}^{88})}{(\text{Sr}^{86}/\text{Sr}^{88}) + 0.1194}$$  \hspace{1cm} (28)$$

where $^{86}\text{Sr}/^{88}\text{Sr}$ is the observed value of the ratio. Twelve determinations of the Elmer and Amend interlaboratory standard are listed in Table 3. The average of these measured values is
0.70792 ± 0.00006 at the 90% confidence level. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of selected samples were measured, and the values tabulated in Appendix B represent the average of approximately 600 scans of the mass spectrum.
Table 3. Determinations of the Elmer and Amend SrCO₃ standard.

<table>
<thead>
<tr>
<th>Run</th>
<th>⁸⁷Sr/⁸⁶Sr</th>
<th>⁸⁶Sr/⁸⁸Sr</th>
<th>Corrected ⁸⁷Sr/⁸⁶Sr</th>
<th>Date</th>
</tr>
</thead>
<tbody>
<tr>
<td>2430</td>
<td>0.70732</td>
<td>0.11962</td>
<td>0.70805</td>
<td>10/11/71</td>
</tr>
<tr>
<td>2448</td>
<td>0.70740</td>
<td>0.11961</td>
<td>0.70801</td>
<td>3/23/72</td>
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<tr>
<td>2479</td>
<td>0.70480</td>
<td>0.12050</td>
<td>0.70804</td>
<td>4/27/72</td>
</tr>
<tr>
<td>2496</td>
<td>0.70440</td>
<td>0.12064</td>
<td>0.70806</td>
<td>6/30/72</td>
</tr>
<tr>
<td>3</td>
<td>0.70794</td>
<td>0.11941</td>
<td>0.70797</td>
<td>9/4/72</td>
</tr>
<tr>
<td>15</td>
<td>0.70723</td>
<td>0.11959</td>
<td>0.70780</td>
<td>12/19/72</td>
</tr>
<tr>
<td>22</td>
<td>0.70612</td>
<td>0.12008</td>
<td>0.70808</td>
<td>1/31/73</td>
</tr>
<tr>
<td>29</td>
<td>0.70735</td>
<td>0.11952</td>
<td>0.70774</td>
<td>2/28/73</td>
</tr>
<tr>
<td>49</td>
<td>0.70744</td>
<td>0.11950</td>
<td>0.70776</td>
<td>6/19/73</td>
</tr>
<tr>
<td>89</td>
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<td>0.70804</td>
<td>10/23/73</td>
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<tr>
<td>146</td>
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<td>0.11927</td>
<td>0.70765</td>
<td>11/11/74</td>
</tr>
<tr>
<td>153</td>
<td>0.70583</td>
<td>0.12009</td>
<td>0.70789</td>
<td>6/18/75</td>
</tr>
</tbody>
</table>

Average corrected ⁸⁷Sr/⁸⁶Sr = 0.70792 ± 0.00006.
CHAPTER V

INTERPRETATION OF THE ELEMENTAL CONCENTRATIONS
AND ISOTOPE RATIOS

INTRODUCTION

The hypothesis that the non-carbonate fraction of the sediment from the Red Sea can be regarded as a mixture of two components has been discussed in Chapter III. A two-component model was developed which predicts that the measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of mixtures will generate a linear array when plotted as a function of their inverse strontium concentrations. Furthermore, the two-component model predicts a linear relationship between the concentrations of strontium and those of any other element that satisfies the assumptions of the model. Such linear patterns provide convincing evidence that the non-carbonate fractions of the sediment from the Red Sea can be regarded as a mixture of two components of contrasting chemical compositions, at least to a first approximation.

In order to test this hypothesis, the concentrations of strontium, CaO, Fe$_2$O$_3$, and SiO$_2$ were measured for the non-carbonate fractions of the sediment from cores 3P, 153P, and 4P (Appendix B). The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were determined for selected samples which represent the range of strontium concentrations available in each core. A discussion of the results of these analyses for each core is presented below.
Figure 22. Core 3P: Two-component line fitted to the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and $1/\text{Sr}$ concentrations. The coefficient of correlation is 0.979.
Figure 23. Core JP: Two-component hyperbola fitted to the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and Sr concentrations.
RESULTS FOR CORE 3P

The strontium concentrations of the non-carbonate fractions of the sediment from core 3P range from about 100 to 300 ppm. Seventeen samples were selected for \(^{87}\text{Sr}/^{86}\text{Sr}\) ratio analysis on the basis of their strontium content. The measured \(^{87}\text{Sr}/^{86}\text{Sr}\) ratios of the non-carbonate samples from core 3P range from about 0.7056 to nearly 0.710. The measured \(^{87}\text{Sr}/^{86}\text{Sr}\) ratios were plotted as a function of their inverse strontium concentrations in Figure 22, and a line was fitted to these data points by least-squares regression. The fit of the line to the data points is satisfactory and supports the two-component interpretation for these samples. Therefore, the equation of the fitted line is the mixing equation for core 3P:

\[
\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_M = 0.8158 \left(\frac{1}{\text{Sr}}\right)_M + 0.70294
\]  

(29)

The coefficient of correlation of the fitted line is 0.979. The mixing equation was used to plot the hyperbola in Figure 23.

The mixing line represents all possible mixtures of the silicic and basaltic components. Therefore, the \(^{87}\text{Sr}/^{86}\text{Sr}\) ratios and strontium concentrations of these end members must satisfy Equation (29). The discussion presented in Chapter III indicated that the \(^{87}\text{Sr}/^{86}\text{Sr}\) ratio of the basaltic component varies between 0.702 and 0.706. These values can be substituted into Equation (29) to
obtain the associated strontium concentrations. However, substitution of 0.702 into Equation (29) leads to a negative strontium content because this value is less than the y-intercept of the fitted line. Therefore, the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the basaltic component cannot be less than 0.7029 which corresponds to an infinite strontium concentration. Substituting a value of 0.706 for the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the basaltic component into Equation (29) yields a strontium concentration of 266 ppm. Small changes in the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the basaltic component results in large variations in the associated strontium content due to the hyperbolic relationship between these parameters as shown in Figure 23. This difficulty is considerably lessened in the case of the basaltic component because the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of young volcanic rocks of basaltic composition are relatively uniform and average $0.7037 \pm 0.0001$ (Faure and Powell, 1972). For the purposes of this study, 0.704 seems to be a reasonable estimate for the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the basaltic component. Substituting this value into Equation (29) yields a strontium concentration of 772 ppm for the basaltic component. This calculated strontium concentration is in excellent agreement with average values for alkali basalts (774 ppm) as compiled by Prinz (1967), and the position of this end member is shown in Figures 22 and 23.

The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the sialic component should lie between 0.720 and 0.710 (Chapter III). Substitution of these values into Equation (29) yields strontium concentrations of 48 and 115 ppm. In the case of the sialic component, large variations in the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio results in small changes in the strontium concentration because the mixing hyperbola is asymptotic to the $^{87}\text{Sr}/^{86}\text{Sr}$ axis (Figure 23).
Although the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the sialic component is variable, 0.715 seems to be a reasonable estimate of this parameter for the study area (Chapter III). Substitution of this value into Equation (29) results in a strontium concentration of 68 ppm for the sialic component, and the position of this end member is shown in Figures 22 and 23.

The calculated strontium concentration of the basaltic component is consistent with published values whereas that of the sialic component is lower than expected for rocks of granitic composition (Chapter III). This result suggests that the sialic fraction consists primarily of clay minerals as well as oxides of iron, aluminum, and manganese which are the weathering products of granitic rocks. On the other hand, the volcanic fraction consists of relatively unweathered detritus. This implied difference in the degree of weathering of the two source rocks is discussed with reference to the observed mineral concentrations presented in Chapter VII.

The results of the strontium concentrations and isotope ratios shown in Figure 22 suggest that the non-carbonate fraction of the sediment from core 3P is a mixture of two components. If this is true, then the two-component model predicts a linear relationship between the concentrations of strontium and those of other elements provided the system has remained closed. In order to test this hypothesis, the concentrations of CaO, Fe$_2$O$_3$, and SiO$_2$ were determined as discussed in Chapter IV. The measured CaO concentrations range from about 2 to 3 percent and average 2.5 percent of the non-carbonate sediment. The Fe$_2$O$_3$ content varies between 6 and 9 percent
and averages about 7.5 percent. The range of values of the SiO$_2$ concentrations is from about 60 to 74 percent with an average SiO$_2$ content of 65 percent. The observed CaO, Fe$_2$O$_3$, and SiO$_2$ concentrations were averaged in 20 ppm increments of their strontium concentrations so as to reduce the scatter of the results. These average values are plotted as a function of the average strontium content of each increment in Figure 24. Straight lines were fitted to these data points by least-squares regression:

\[
(CaO)_M = (7.266 \times 10^{-3}) Sr_M + 1.01 \\
(Fe_2O_3)_M = (-8.156 \times 10^{-3}) Sr_M + 9.05 \\
(SiO_2)_M = (2.081 \times 10^{-2}) Sr_M + 60.37
\]

The lines in Figure 24 emphasize the trends of these data and do not necessarily represent statistically valid, "best-fit" regression lines. Furthermore, Equations (30), (31), and (32) are stated for comparison purposes only.

The trends of these data are consistent with the hypothesis that the non-carbonate sediment from this location in the Red Sea is a mixture of detritus derived from two sources of contrasting chemical compositions.

The chemical compositions of 36 basaltic rocks from volcanic islands in the Red Sea were reported by Gass et al. (1973), and the average CaO, Fe$_2$O$_3$, and SiO$_2$ concentrations are listed in Table 4. The averages of 20 analyses of basement rocks from the
Figure 24. Core 3P: Two-component lines fitted to the CaO, Fe$_2$O$_3$, and SiO$_2$ concentrations vs Sr content.
Table 4. Average CaO, Fe$_2$O$_3$, and SiO$_2$ concentrations of rocks of sialic and basaltic composition from the Red Sea area.

<table>
<thead>
<tr>
<th></th>
<th>Sialic Rocks$^1$</th>
<th>Basaltic Rocks$^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>CaO, %</td>
<td>2.2</td>
<td>9.2</td>
</tr>
<tr>
<td>Fe$_2$O$_3$, %</td>
<td>5.6</td>
<td>11.0</td>
</tr>
<tr>
<td>SiO$_2$, %</td>
<td>67.0</td>
<td>50.0</td>
</tr>
</tbody>
</table>

$^1$Mohr, 1970.

Red Sea region (Mohr, 1970) are listed in Table 4. The trend of the CaO concentrations in Figure 24 is consistent with the assumption that this oxide is associated with detritus of basaltic composition. However, the trends of the Fe$_2$O$_3$ and SiO$_2$ concentrations are not as expected. The Fe$_2$O$_3$ concentration appears to decrease as the sediment becomes enriched in volcanogenic detritus, i.e., higher strontium concentration. On the other hand, the SiO$_2$ concentration increases with increasing input of the basaltic component. These apparently anomalous relationships are discussed with reference to the observed mineral concentrations presented in Chapter VII.

Rubidium concentrations of 17 samples range from 42 to 75 ppm with an average concentration of about 60 ppm. The measured rubidium concentrations were plotted as a function of their strontium content in Figure 25, and a line was fitted to these data points by least-squares regression:

$$\text{Rb} = (-0.1425) \text{Sr} + 87.8$$  \hspace{1cm} (33)

The rubidium content of the end members can be estimated by substitution of their previously determined strontium concentrations in Equation (33). However, this calculation leads to a negative rubidium concentration for the basaltic component. This anomalous result is caused by over-steepening of the slope of the fitted line in Figure 25 due to preferential uptake of rubidium dissolved in sea water by clay minerals of the sialic fraction. A similar example of rubidium fixation during deposition of marine sediment
Figure 25. Core JP: Anomalous two-component line fitted to the Rb and Sr concentrations.
was observed by Shaffer (1974).

Linear patterns generated by element concentrations do not provide unequivocal evidence of mixing, since such patterns may be the result of selective addition or removal of elements or minerals from the sediment. However, the fit of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios vs $1/\text{Sr}$ concentrations results to a straight line in combination with the linear relationships between the concentrations of strontium and those of the major elements provide convincing evidence that, at least to a first approximation, the non-carbonate fractions of the sediment from core 3P satisfy the two-component model.

RESULTS FOR CORE 153P

The strontium concentrations of all the non-carbonate fractions of the sediment from core 153P range from about 150 to 270 ppm, with an average concentration of nearly 200 ppm. $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were determined for 12 non-carbonate samples, selected so as to represent the entire range of strontium concentrations available in this core. The measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of these samples range from about 0.7057 to 0.7075 and are plotted as a function of their inverse strontium concentrations in Figure 26. Although these data points are more dispersed than the results of core 3P, the majority of the points are co-linear, indicating that the two-component hypothesis is satisfied by most of these samples. However, some data points deviate from the predicted pattern, and the presence of additional components must be considered. Information
Figure 26. Core 153P: Two-component line fitted to the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and $1/\text{Sr}$ concentrations.
concerning the dispersion of these results is provided by their Fe$_2$O$_3$ concentrations as shown below.

The Fe$_2$O$_3$ concentrations of the non-carbonate samples from core 153P range from nearly 8 to 28 percent and average about 12 percent of the sediment. Although most of the samples from core 153P have Fe$_2$O$_3$ concentrations that are comparable to those of core 3P, some of the samples have anomalously high Fe$_2$O$_3$ contents (Appendix B). Furthermore, those samples that deviate from the predicted two-component pattern in Figure 26 also contain excessive amounts of Fe$_2$O$_3$. This suggests the possibility that at least one of the additional components causing this dispersion may be excess Fe$_2$O$_3$. In this case, Fe$_2$O$_3$ contains no strontium and acts as a diluant causing the measured strontium concentrations to be lowered. This is consistent with the fact that these samples plot to the right of the mixing line in Figure 26. If the amount of excess Fe$_2$O$_3$ can be determined, then the strontium concentrations can be recalculated. These refined values should cause the deviant points to conform more closely to the predicted two-component pattern as shown by arrows in Figure 26.

The amount of excess Fe$_2$O$_3$ is determined by averaging the Fe$_2$O$_3$ concentrations in 10 ppm increments of their strontium contents. These values are then plotted as a function of the average strontium content of each increment as shown in Figure 27. Four of the data points appear to be co-linear and a least-squares
Figure 27. Core 153P: Plot of the $\text{Fe}_2\text{O}_3$ vs Sr concentrations showing the secondary-mixing lines of those samples that contain excessive amounts of $\text{Fe}_2\text{O}_3$. 
The regression line is fitted to these points:

$$\left( \text{Fe}_2\text{O}_3 \right)_M = (-9.132 \times 10^{-3}) \text{Sr}_M + 10.97$$  \hspace{1cm} (34)

This line is in good agreement with the results obtained for core 3P which is consistent with the assumption that these samples satisfy the two-component model. The data points which do not fit the two-component line represent single samples that contain varying amounts of excess Fe$_2$O$_3$. Therefore, a secondary-mixing line is generated between each deviant point and a point on the diagram representing 100 percent Fe$_2$O$_3$ and 0 ppm strontium. The intersection of these secondary-mixing lines and the fitted line provides an estimate of the Fe$_2$O$_3$ content of each sample before the excess Fe$_2$O$_3$ was added. The amount of excess Fe$_2$O$_3$ can then be determined by subtracting the estimated concentration from the measured concentration. For example, the measured Fe$_2$O$_3$ content of sample number 330 was about 23 percent (Appendix B), and the intersection of the secondary-mixing line with the two-component line provides an estimated Fe$_2$O$_3$ content of less than 10 percent (Table 5). Therefore, sample 330 contains about 13 percent excess Fe$_2$O$_3$.

Because the two-component model assumes that all the non-carbonate sediment was derived from the basaltic and sialic sources, the measured Sr, CaO, and SiO$_2$ must be multiplied by a factor equal to 1/0.87 so as to correct for the dilution effect of the third component. These recalculated concentrations for sample 330 are listed in Table 5, as are the recalculated values for the other
Table 5. Recalculated element concentrations for core 153P after removal of excess Fe$_2$O$_3$.

<table>
<thead>
<tr>
<th>Sample Depth (cm)</th>
<th>Fe$_2$O$_3$ Concentration</th>
<th>Recalculated Concentrations</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Estimated %</td>
<td>Excess %</td>
</tr>
<tr>
<td>150</td>
<td>9.48</td>
<td>3.7</td>
</tr>
<tr>
<td>190</td>
<td>9.29</td>
<td>18.3</td>
</tr>
<tr>
<td>210</td>
<td>9.30</td>
<td>5.4</td>
</tr>
<tr>
<td>290</td>
<td>9.10</td>
<td>12.6</td>
</tr>
<tr>
<td>330</td>
<td>9.62</td>
<td>13.3</td>
</tr>
</tbody>
</table>
samples that contain significant amounts of excess Fe$_2$O$_3$. These recalculated values can now be used to refine the Sr mixing line in Figure 26.

The recalculated strontium concentrations of the five samples which contained excess Fe$_2$O$_3$ as well as the seven unaffected samples were used to fit a least-squares regression line to the data points in Figure 26:

$$(^{87}\text{Sr}/^{86}\text{Sr})_M = 0.7216 (1/\text{Sr})_M + 0.70310 \quad (35)$$

The coefficient of correlation of the fitted line is 0.981. Equation (35) is the refined mixing equation for core 153P and was used to plot the hyperbola in Figure 28.

Recalculation of the strontium concentrations causes the divergent points to move closer to the mixing hyperbola as shown by the insert in Figure 28. The fit of the refined results to the mixing line and hyperbola indicates that the excess Fe$_2$O$_3$ was the primary cause of the deviation of the original values.

The strontium concentrations of the end members can be calculated by substitution of the assumed $^{87}$Sr/$^{86}$Sr ratios in Equation (35). This calculation yields strontium concentrations of 61 and 800 ppm for the sialic and basaltic components, respectively. The positions of these components are indicated in Figures 26 and 28. The estimated strontium concentrations obtained for the end members of core 153P are in very good agreement with those of core 3P.

The CaO contents of the non-carbonate samples from core 153P
Figure 28. Core 153P: Two-component hyperbola fitted to the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and Sr concentrations showing the refinement of the Sr concentrations.
range from more than 2 to nearly 4 percent and average about 2.5 percent, whereas the SiO$_2$ concentrations vary between 47 and 65 percent and average about 59 percent. The CaO and SiO$_2$ concentrations were averaged in 10 ppm increments of their strontium content, and are plotted in Figure 29. The refinement of these results is shown by the arrows. Those samples which contain excessive amounts of Fe$_2$O$_3$ also appear to have anomalously high CaO concentrations. The most deviant sample was used to calculate the amount of excess CaO, and the other concentrations were recalculated as before. Because the amount of excess CaO is 1 percent or less, this calculation results in insignificant changes for the concentrations of the other elements. The six data points that do not contain excess CaO appear to be co-linear, and a least-squares regression line is fitted to these points so as to emphasize the trend of these results:

\[
(CaO)_H = (7.95 \times 10^{-3}) \, Sr_M + 0.88 \tag{36}
\]

The refined SiO$_2$ concentrations tend to improve the fit of these results, and a line is fitted to these data points to emphasize their trend:

\[
(SiO_2)_H = (4.75 \times 10^{-2}) \, Sr_M + 53.6 \tag{37}
\]

The patterns of the refined CaO, Fe$_2$O$_3$, and SiO$_2$ concentrations are consistent with the two-component model and are in good agreement
Figure 29. Core 153P: Two-component lines fitted to the CaO and SiO$_2$ concentrations showing the refinement of these results.
with the results of core 3P.

The recalculated rubidium concentrations are plotted as a function of their strontium contents in Figure 30, and a line is fitted to these data points:

\[ \text{Rb} = (-0.1028) \text{Sr} + 76.9 \]  \hspace{1cm} (38)

The substitution of the estimated strontium concentration of the basaltic component into Equation (38) results in a negative rubidium content for this end member. These results suggest the same pattern of selective addition of rubidium to the finer-grained sialic detritus as was observed for core 3P.

The presence of excessive amounts of iron in some samples from core 153P is not unexpected. The authigenic precipitation of iron-rich minerals from brine pools in the Red Sea is well-documented (Bischoff, 1969). There is evidence that the overflow of these brine pools in the past resulted in the accumulation of brines in adjoining depressions (Ross, 1972; Brewer et al., 1971). Core 153P is located nearly 200 km from the major Red Sea brine area, and it is unlikely that these brines are the source of the excess Fe₂O₃ observed in the samples from this core. However, recent exploration has led to the discovery of other brine areas (Ross et al., 1973). Because the samples with excess Fe₂O₃ are contiguous (150 to 210 cm and 290 to 330 cm), it is possible that there may have been a brine pool near the site of core 153P during the time represented by these samples. It is likely, therefore, that the
Figure 30. Core 153P: Anomalous two-component line fitted to the Rb vs Sr concentrations.
excess Fe$_2$O$_3$ is authigenic in origin. This study is concerned with the detrital fraction of the sediment and the removal of the authigenic fraction of the iron and subsequent interpretation in terms of a two-component system is justified. The removal of the excess Fe$_2$O$_3$ results in a refinement of the Sr concentrations used to derive the mixing equation and also improves the fit of the data points in the diagram showing the linear relationships between the concentrations of Sr, CaO, and SiO$_2$ as predicted by the two-component model.

Cores 3P and 153P were collected from similar depositional environments in the median valley of the Red Sea (Chapter IV). The analytical results for these cores show similar patterns that are consistent with the two-component model. The next step is to apply the same technique to a core from a different depositional environment such as core 4P which was collected in relatively shallow water on the flank of the continental shelf.

RESULTS FOR CORE 4P

The strontium concentrations of the non-carbonate fractions of the sediment samples from core 4P range from about 100 to 250 ppm and average about 200 ppm. The $^{87}$Sr/$^{86}$Sr ratios were determined for 10 samples selected so as to represent the range of strontium concentrations. The $^{87}$Sr/$^{86}$Sr ratios of these samples vary between about 0.7062 and 0.7080 and are plotted as a function of their inverse strontium concentrations in Figure 31. Eight of the data
Figure 31. Core 4P: Two-component line fitted to the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and $1/\text{Sr}$ concentrations.
Figure 31.

$87\text{Sr}/86\text{Sr}$ vs. $1/\text{Sr}$ Concentration, ppm$^{-1}$ (x $10^{-3}$)

- Basaltic Component
- Sialic Component
points are co-linear, and a line has been fitted to these points by least-squares regression:

\[
\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_M = (0.75084)(1/\text{Sr})_M + 0.70321 \tag{39}
\]

The coefficient of correlation of the fitted line is 0.962. This equation was used to plot the hyperbola in Figure 32.

Two of the data points do not appear to satisfy the two-component model. The sample with the greatest deviation was re-analyzed to insure that the variance from the predicted pattern was not due to analytical error. The average of replicate determinations of the strontium concentrations and \(^{87}\text{Sr}/^{86}\text{Sr}\) ratios are 99 ± 5 ppm and 0.7081 ± 0.0004, respectively. These results indicate that the deviation is not attributable to analytical error.

The two samples that do not fit the mixing line show the same pattern of variance as was observed for core 153P. They are contiguous (330 and 340 cm) and plot to the right of the mixing line. This suggests the possibility that the samples contain an additional component. Based on the results of core 153P, it is assumed that the third component contains no strontium and acts as a diluant, causing the measured strontium concentrations to be too low. The strontium concentrations of the divergent samples can be recalculated by substituting the measured \(^{87}\text{Sr}/^{86}\text{Sr}\) ratios in Equation (39). The values of the recalculated strontium concentrations are listed in Table 6.

The estimated strontium concentrations of the end members
Figure 32. Core 4P: Two-component hyperbola fitted to the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and Sr concentrations.
Table 6. Recalculated element concentrations for core 4P after removal of a third component which was assumed to contain no Sr.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Sr (ppm)</th>
<th>CaO (%)</th>
<th>Fe₂O₃ (%)</th>
<th>SiO₂ (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>330</td>
<td>153</td>
<td>3.0</td>
<td>6.5</td>
<td>90.9</td>
</tr>
<tr>
<td>340</td>
<td>158</td>
<td>7.5</td>
<td>6.1</td>
<td>63.9</td>
</tr>
</tbody>
</table>
were determined by substitution of their assumed $^{87}\text{Sr}/^{86}\text{Sr}$ ratios into Equation (39). This calculation resulted in estimates of 64 and 995 ppm for the sialic and basaltic components, respectively. The positions of these end members are shown on the mixing line and hyperbola in Figures 31 and 32. The estimated strontium concentrations of the end members for core 4P are in very good agreement with those previously obtained for cores 3P and 153P.

The CaO, Fe$_2$O$_3$, and SiO$_2$ concentrations of the 10 non-carbonate samples for which $^{87}\text{Sr}/^{86}\text{Sr}$ ratios had been determined are listed in Appendix B. The CaO concentrations range from nearly 2 to 3 percent and average about 2.5 percent, whereas the SiO$_2$ contents vary between nearly 59 to 70 percent and average about 64 percent of the non-carbonate fractions. These results are not substantially different from those obtained for cores 3P and 153P. The Fe$_2$O$_3$ concentrations range from 4 to less than 7 percent of the non-carbonate sediment. The average Fe$_2$O$_3$ content of core 4P is about 5.8 percent which is significantly lower than that of cores 3P and 153P (7.5 and 12 percent, respectively). Cores 3P and 153P are from the median valley of the Red Sea whereas the core 4P location is on the flank of the continental shelf in a shallow-water environment. Therefore, the variation in the average Fe$_2$O$_3$ contents may reflect the different depositional environments of these core locations.

The measured CaO, Fe$_2$O$_3$, and SiO$_2$ concentrations of the 10 samples from core 4P are plotted as a function of their strontium contents in Figure 33. The oxide concentrations of the two deviant
Figure 33. Core 4P: Two-component line fitted to the CaO, Fe$_2$O$_3$, and SiO$_2$ concentrations vs Sr concentrations showing the refinement of these results by removal of a third component which was assumed to contain no Sr.
samples (330 and 340) were recalculated by using the percentage of change of the refined strontium results, and these values are listed in Table 6. The CaO concentrations suggest the presence of excessive amounts of this oxide, as shown in Figure 33. However, as was the case for core 153P, the amount of excess CaO is not sufficient to cause significant changes in the concentrations of the other elements. The 7 samples that do not contain excess CaO appear to be co-linear, and a least-squares regression line is fitted to these data points in order to emphasize the trend of these results:

\[(\text{CaO})_M = (5.77 \times 10^{-3}) \text{Sr}_M + 1.28\]  

(40)

The fit of the Fe$_2$O$_3$ concentrations in Figure 33 is improved by the recalculation of the strontium and Fe$_2$O$_3$ concentrations of the two deviant samples, and a line has been fitted to these points to emphasize their trend:

\[(\text{Fe}_2\text{O}_3)_M = (-3.54 \times 10^{-3}) \text{Sr}_M + 6.64\]  

(41)

Although the data points are rather dispersed with respect to the fitted line, there are no systematic trends that can be interpreted in terms of enrichment or depletion of Fe$_2$O$_3$ in these samples. Furthermore, each data point in Figure 33 represents a single sample whereas those of cores 3P and 153P are averages of several samples. Therefore, the dispersion of these results is assumed to be a
measure of the inhomogeneity of the samples rather than indicating the presence of anomalous amounts of Fe₂O₃.

The recalculated SiO₂ concentrations of the deviant samples from core 4P indicate that sample 330 contains nearly 91 percent of this oxide (Table 6). This is not consistent with the SiO₂ concentration of the other samples, as shown in Figure 33. The unusually high SiO₂ content of this sample is not reflected in the observed concentrations of the silicate minerals which are abnormally low (Appendix B). It is possible that the high SiO₂ concentration is due to the presence of X-ray amorphous material such as volcanic glass or siliceous fossil-material. However, microscopic analysis has not substantiated this possibility. Because the SiO₂ concentration of sample 340 is consistent with the other results, the measured SiO₂ content of sample 330 is assumed to be erroneous and is ignored in the fitting of the line to the data points in Figure 33. The equation of the least-squares regression line which was drawn to emphasize the trend of these results is:

\[
(SiO_2)_M = (3.51 \times 10^{-2}) \text{Sr}_M + 57.9 \quad (42)
\]

In most cases, the recalculated CaO, Fe₂O₃, and SiO₂ concentrations improve the fit of these results to the linear pattern predicted by the two-component model.

The \(^{87}\text{Sr}/^{86}\text{Sr}\) ratios plotted as a function of the inverse strontium concentrations generate linear arrays that are consistent with the two-component interpretation for the non-carbonate
fractions of the sediment from cores 3P, 153P, and 4P. Mixing equations can be derived from the analytical results for each core and are used to estimate the strontium concentrations of the hypothetical end members. The estimated strontium contents of the sialic component range from 61 to 68 ppm, whereas those of the basaltic component vary between 772 and 995 ppm. Although the cores are separated by as much as 200 km and represent different depositional environments, the close agreement of the independently-determined strontium concentrations of the end members indicates that these sites have been receiving detritus of similar composition. Furthermore, the strontium concentration of the basaltic component is in excellent agreement with published average values for alkali basalts, whereas that of the sialic component is significantly lower than expected. This discrepancy leads to the conclusion that the sialic detritus is composed of clay minerals as well as oxides of Fe, Al, and Mn, which are the weathering products of granitic rocks, and that the volcanogenic fraction consists of much less weathered detritus. The two-component interpretation is further substantiated by the linear relationships between the major element concentrations and the strontium contents of the non-carbonate samples from cores 3P, 153P, and 4P. The CaO, Fe$_2$O$_3$, and SiO$_2$ vs strontium plots have similar systematic trends for the three cores which indicates that the composition of the detritus is similar at these locations. The apparently anomalous trends of the Fe$_2$O$_3$ and SiO$_2$ concentrations will be used in Chapter VII to interpret the mineral concentration results.
CHAPTER VI
DETERMINATION OF THE CONCENTRATION OF THE VOLCANOGENIC DETRITUS
AND CORRELATION OF THE LITHO-STRATIGRAPHEIS
FOR CORES 3P, 153P, AND 4P

INTRODUCTION

The next step is to calculate the fraction of the basaltic component, $f$, contained in each mixture. This calculation involves substitution of the estimated strontium concentrations of the end members and the measured strontium concentration of the samples into Equation (10). The $f$-values calculated in this manner are clearly dependent upon estimated strontium concentrations of the end members which vary within narrow limits due to the heterogeneity of the source materials (Chapter III and Chapter V). However, in order to facilitate the interpretation of the results, the end members are assumed to have unique values of their $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and, therefore, their Sr concentrations. This simplifying assumption does not significantly affect the qualitative interpretation of the results. Furthermore, the calculation of the $f$-values of each mixture requires the additional assumption that all the non-carbonate samples for a particular core satisfy their respective mixing equations, although these equations are derived from the results of selected samples only. This assumption is not likely to introduce significant error, because the samples were
selected solely on the basis of their measured strontium concentrations and were taken throughout the entire length of each core. The calculated f-values are expressed as weight percent of the basaltic component for each non-carbonate sample and plotted as a function of the sample depth in the cores. The systematic variations in the concentrations of the basaltic component define layers, thus providing litho-stratigraphies for each core. The lateral equivalence of sediment layers is used to correlate the stratigraphies by a graphical technique. This method provides correlation equations which permit the sample depths of each core to be recalculated in equivalent units. An approximate time-scale extending back to about 20,000 years in the past is provided by the dates of the calcium carbonate fractions of the cores.

CALCULATION OF THE CONCENTRATION OF VOLCANOGENIC DETRITUS IN THE SEDIMENT

The concentration of volcanogenic detritus in the sediment samples from core 3P varies from 6 to 34 percent, whereas those of cores 153P and 4P range from 12 to 28 percent, respectively. The variations in the concentration of the basaltic component result from increasing or decreasing input of detritus derived from this source. Because these samples are mixtures of two components, decreases in the concentration of the volcanogenic material can be interpreted as increased input of detritus derived
from the sialic source. However, this interpretation can be ambiguous in that fictitious sialic or volcanic events can be generated by the lack of deposition of detritus derived from the basaltic or sialic sources, respectively. In order to facilitate the ongoing discussion, it is assumed that there exists a steady-state mixing condition and that the systematic variations in the basaltic composition of adjacent samples define layers of sediment enriched in detritus derived from either the basaltic or sialic sources. For example, over the interval from 100 to 270 cm in the sediment profile of core 153P, the basaltic composition of the samples decreases from about 24 percent to less than 14 percent, then increases to more than 25 percent. This systematic variation in the composition of the sediment defines a layer that is relatively depleted in volcanogenic detritus, i.e., a layer enriched in sialic detritus. Other layers that contain sediment enriched in either the basaltic or sialic component were identified in this manner, and these sediment layers define a litho-stratigraphy for each core.

The apparent lateral equivalence of sediment layers suggests the possibility that the stratigraphies can be correlated. Cores 3P and 153P were collected from similar depositional environments in the median valley and contain pelagic sediment which was transported either in suspension or by wind. Therefore, laterally-equivalent sediment layers identified in these cores are assumed to represent events which are correlative. The different depositional environment of core 4P implies that the correlation of this core with the
Figure 34. Systematic variation of the concentration of the basaltic component in cores 153P, 3P, and 4P as a function of depth. Sample depths in cm for core 153P are shown in the scale on the left. The scale on the right shows the sample depths in cm for cores 3P and 4P. Lines have been drawn to suggest the correlation of sediment layers defined on the basis of their basaltic content.
Figure 34.
other cores may be more difficult as discussed below.

CORRELATION OF THE LITHO-STRATIGRAPHIES

Cores 153P and 3P were correlated on the basis of the relative positions and basaltic compositions of sediment layers. The correlation of core 4P with the other cores was less obvious and will be discussed separately. A sialic-rich layer in core 153P (100 to 200 cm) was matched with a similar sialic-rich layer in core 3P (40 to 100 cm). By aligning these sediment layers as shown in Figure 34, a point-by-point correlation between these cores was established using a graphical correlation technique first proposed by Shaw (1964). This method involves plotting the depth of each sample in core 153P as a function of the depth to the correlative sample in core 3P. Such a correlation diagram is shown in Figure 35. A line was fitted to these data points by least-squares regression:

\[ 153P = 1.13 (3P) + 100.59 \] (43)

where 153P and 3P are the depths in centimeters of the samples in the respective cores. The coefficient of correlation of the fitted line is 0.990.

The fit of the data points to the line substantiates the correlation of these cores. The slope of the correlation line implies that the average rate of sediment accumulation in core 153P is approximately 13% greater than that of core 3P. The core tops are not correlative, and the numerical value of the constant in Equation (43) indicates that about 100 cm of sediment are missing from the top of core 3P relative to core 153P. The sediment that appears to be missing from
Figure 35. Point-by-point correlation of cores 153P and 3P. The coefficient of correlation of the fitted line is 0.990.
the top of core 3P could have been lost in the coring operation or may not be present due to non-deposition or erosion. The fit of the data points to the correlation line implies that the rate of sediment accumulation was relatively constant and continuous throughout the time represented by the correlation.

Once cores 153P and 3P have been correlated, the next step is to correlate them with core 4P. This core is located on the edge of the continental shelf in relatively shallow water, whereas cores 153P and 3P are from deep water in the median valley (Chapter IV). The layers identified on the basis of the systematic variations in the basaltic composition of the sediment from core 4P are less well-defined than those of cores 153P and 3P; presumably due to the different depositional environments of the cores. Although this discrepancy makes correlation difficult, there are sufficient similarities in the sediment profiles of these cores to suggest that a correlation is possible. One such correlation is indicated by the lines in Figure 34. This match of the cores leads to the identification of 14 correlative samples between cores 4P and 153P. These sample depths were plotted on a correlation diagram as before (Figure 36), and a least-squares regression line was fitted to these data points:

\[ 153P = 2.17(4P) + 90.24 \]  \hspace{1cm} (44)

where 153P and 4P represent the sample depths in those cores. The coefficient of correlation of the fitted line is 0.999. Because the correlation of core 4P with the others is less obvious than that of core 3P with core 153P, other correlations were attempted. However, these correlations do not produce linear arrays of comparable quality
Figure 36. Point-by-point correlation of cores 153P and 4P. The coefficient of correlation is 0.994.
as that in Figure 36. The use of the linearity of the correlation
plot to test the correlation of the cores assumes that the rate of
sedimentation of core 4P has been fairly constant and continuous
throughout the time represented by the samples. This assumption is
consistent with the results of the correlation of cores 153P and 3P.
Therefore, the correlation indicated by the lines in Figure 34 is
assumed to be valid, and Equation (44) is the correlation equation for
core 4P. The slope of Equation (44) implies that the relative rate
of sediment accumulation in core 4P is less than half that of core 153P.
Furthermore, the numerical value of the constant in Equation (44)
indicates that about 90 cm of sediment are missing from the top of
core 4P relative to core 3P. As in the case of core 3P, the sediment
missing from the top of core 4P could have been lost in the coring
operation or may not have been present due to non-deposition or erosion.

$^{14}$C DATES OF THE CARBONATE FRACTION

Six $^{14}$C dates of calcium carbonate from cores 153P and 154P have
been reported by Ku et al. (1969). Four additional dates were deter-
mined for samples from cores 3P and 4P by Geochron Laboratories for
this study. The $^{14}$C dates and depths of the samples are listed in
Table 7. These results indicate that sediment in the interval 85 to
100 cm in core 153P was deposited about 5,000 years ago, whereas
sediment in the interval 75 to 85 cm from core 3P was deposited about
21,000 years ago. Furthermore, sediment near the top of core 4P is
also significantly older than that of core 153P. These differences
in the reported $^{14}$C dates are consistent with the conclusion that
Table 7. $^{14}$C dates of carbonate fractions from cores 153P, 154P, 3P, and 4P.

<table>
<thead>
<tr>
<th>Core</th>
<th>Depth, cm</th>
<th>$^{14}$C date, yrs.</th>
<th>Equivalent depth in core 153P, cm</th>
</tr>
</thead>
<tbody>
<tr>
<td>153P$^1$</td>
<td>85-100</td>
<td>5,150 ± 500</td>
<td>-</td>
</tr>
<tr>
<td>153P$^1$</td>
<td>170-185</td>
<td>16,400 ± 1,600</td>
<td>-</td>
</tr>
<tr>
<td>153P$^1$</td>
<td>237-259</td>
<td>18,900 ± 1,100</td>
<td>-</td>
</tr>
<tr>
<td>154P$^1$</td>
<td>130-140</td>
<td>13,450 ± 450</td>
<td>-</td>
</tr>
<tr>
<td>154P$^1$</td>
<td>185-200</td>
<td>17,900 ± 600</td>
<td>-</td>
</tr>
<tr>
<td>154P$^1$</td>
<td>360-375</td>
<td>&gt;27,000</td>
<td>-</td>
</tr>
<tr>
<td>3P$^2$</td>
<td>75-85</td>
<td>21,150 ± 1,000</td>
<td>185-196</td>
</tr>
<tr>
<td>3P$^2$</td>
<td>185-195</td>
<td>&gt;32,000</td>
<td>310-320</td>
</tr>
<tr>
<td>4P$^2$</td>
<td>60-70</td>
<td>&gt;32,000</td>
<td>220-242</td>
</tr>
<tr>
<td>4P$^2$</td>
<td>190-200</td>
<td>36,900 ± 2,700</td>
<td>502-523</td>
</tr>
</tbody>
</table>

$^1$Ku et al., 1969

$^2$Geochron laboratories, $^{13}$C correction included.
Figure 37. $^{14}C$ dates of calcium carbonate from cores 153P, 154P, 3P, and 4P plotted as a function of equivalent depth in core 153P.
Figure 37.

$14^C$ Date x $10^3$, years

Depth in Core 153P units, cm
about one meter of sediment is missing from the tops of cores 3P and 4P relative to core 153P. Correlation equations (43) and (44) were used to recalculate the depths of samples from cores 3P and 4P in terms of equivalent depths in core 153P (Table 7). The $^{14}$C dates were plotted as a function of depth in core 153P in Figure 37. Comparison of the $^{14}$C dates for cores 153P and 154P indicates that these cores have correlative tops and similar sedimentation rates. Therefore, it is not necessary to recalculate the core 154P sample depths in order to plot their $^{14}$C dates in Figure 37. The data points do not fit a single line, and for this reason, two limiting lines were drawn. The median line, drawn between the limiting lines, provides an approximate chronology for these cores extending from the Present to about 20,000 years in the past. Figure 37 was used to estimate an average rate of accumulation of approximately 10 cm per 1000 years for cores 153P and 154P. Sedimentation rates of 9 and 5 cm per 1000 years were obtained for cores 3P and 4P, respectively, by means of the slopes of their correlation equations (Equations (43) and (44)). These values are consistent with accumulation rates for the Red Sea reported by other workers (Ku et al., 1969; Milliman et al., 1969).

The $^{14}$C dates for the carbonate fractions from cores 153P, 3P, and 4P tend to confirm the correlations established in the previous section. Therefore, the correlation equations were used to recalculate the depths of the samples from cores 3P and 4P in terms of equivalent depths in core 153P. The concentrations of the basaltic component of the samples from cores 3P and 4P were then replotted as a function
of their equivalent core 153P depths as shown in Figure 3B. Plotting
the sedimentary profiles in this manner removes the effects of the
different sedimentation rates and thereby facilitates the inter-
pretation of the results.

INTERPRETATION OF THE LITHO-STRATIGRAPHIES

The process of sedimentation in the Red Sea has apparently
operated in such a way that the concentration of volcanogenic
detritus varies from less than 10 to more than 30 percent of the non-
carbonate fraction of the sediment. There are two layers that are
significantly enriched in volcanogenic detritus occurring at depths
of about 270 and 390 cm. Five sediment layers that appear to be
enriched in sialic detritus occur at depths of about 150, 310, 430,
640, and 840 cm. The two older sialic-rich layers are not present
in core 153P because the sediment below 550 cm in this core was
disturbed in coring.

The concentration of the basaltic component appears to increase
toward the tops of cores 3P and 4P, with a possible transition at
a depth of about 500 cm. A similar transition may be present in
core 153P, but is obscured due to the disturbance of the sediment.
This suggests a trend toward increasing input of volcanogenic
detritus in more recent times.

The average basaltic concentration of the upper 300 cm of core
3P is about 25 percent, whereas cores 153P and 4P average about 18
and 19 percent, respectively. Core 153P is the northernmost core
and core 3P is the southernmost. This apparent regional trend in the average basaltic composition of the upper 300 cm of the cores suggests that large amounts of volcanogenic detritus may be derived from volcanic islands located at the southern end of the Red Sea and from volcanic rocks exposed in the Afar region of Ethiopia as well as the southern Arabian peninsula (Chapter II).

Comparison of the general patterns of the sediment profiles in Figure 38 indicates that the sediment layers of core 4P are less well defined than those of cores 153P and 3P. This difference is attributed to the different depositional environments of these cores. Cores 153P and 3P are from the median valley and are separated by more than 200 kilometers. Core 4P is located between the other cores in a shallow-water environment. The differences in the sediment profiles suggest that, for correlation purposes, the environment of deposition is a more important consideration than the distance between the cores.

Litho-stratigraphies were defined on the basis of the systematic variations in the concentration of the basaltic component contained in the non-carbonate fractions of the sediment. The basaltic content was derived from the isotopic compositions and concentrations of strontium by means of the two-component mixing model which assumes the existence of two components of different composition. If the litho-stratigraphies defined in this manner are valid, then there should be systematic variations in the calculated basaltic compositions of the sediment samples as a function of their mineral concentrations. This hypothesis is examined in the following chapter.
Figure 38. Systematic variation of the concentration of the basaltic component in cores 153P, 3P, and 4P as a function of equivalent depth in core 153P.
Figure 38.
CHAPTER VII

INTERPRETATION OF THE MINERAL CONCENTRATIONS

Mineral concentrations were determined by X-ray diffraction for selected samples which were chosen so as to represent the range of basaltic concentrations available in each core. The measured concentrations of quartz, feldspar, amphibole, illite, kaolinite, and chlorite were averaged with reference to the incremental scale of basalt concentration defined in Chapter V. The average concentrations of these minerals are plotted as a function of the average basalt content of each increment in Figures 39 and 40. The amount of material which was not directly detectable by X-ray diffraction (X-ray amorphous) was calculated by subtracting the sum of the observed mineral concentration of each sample from 100 percent. The concentrations of the X-ray amorphous material were averaged and plotted with reference to the incremental scale of the basaltic composition in Figure 41. Lines were drawn through the data points in these diagrams so as to emphasize the trends of these results. The interpretation of the mineral concentrations provides important information concerning the deposition of sediment in the Red Sea as presented in the following discussion.

The measured amphibole contents of the samples vary from less than one to more than three percent and average about two percent.
of the non-carbonate fractions of the sediment from cores 3P, 153P, and 4P. The trends in Figure 39 indicate that the concentration of this mineral increases with increasing content of volcanogenic detritus for all three cores. This pattern is consistent with the assumption that the amphibole is primarily derived from basic volcanic rocks.

Cores 3P and 153P have similar feldspar contents which range from 6 to 14 percent and average about 10 percent of the non-carbonate sediment. The feldspar concentrations of core 4P are consistently higher and vary from nearly 8 to 38 percent, with an average concentration of about 24 percent. Furthermore, the trends of the feldspar concentrations of cores 3P and 153P (Figure 39) increase with increasing basaltic composition, whereas core 4P displays an inverse relationship between these parameters. The difference in the feldspar concentrations may be attributed to the different depositional environments of these cores. In the case of core 4P, the inverse relationship between the feldspar content and the basaltic composition of the samples as well as their consistently higher feldspar concentrations presumably result from the input of significant amounts of detrital-feldspar grains derived directly from the sialic source. For cores 3P and 153P, the apparent low concentrations of the feldspar in the sialic-rich sediment can be attributed to the removal from suspension in seawater of the relatively coarse detrital-feldspar fraction in the near-shore environment. Moreover, most of the feldspar associated with the sialic source has been converted to fine-grained weathering
Figure 39. Variation of the absolute quartz, feldspar, and amphibole concentrations vs basaltic composition.
products such as clay minerals which is suggested by the low strontium concentration of this component (Chapter V). The high feldspar content of sediment enriched in volcanogenic detritus reflects the relatively unweathered condition of this component, consistent with its high strontium concentration (Chapter V).

The measured quartz concentrations range from nearly 10 to 20 percent and average 14 percent of the non-carbonate sediment. The quartz content of the cores increases directly with the basaltic composition of the sediment (Figure 39). Although the association of a high quartz content with sediment enriched in volcanogenic detritus was not expected, this pattern is consistent with the silica contents of the samples presented in Chapter V. It is possible that quartz and most of the volcanogenic detritus are transported together by wind and, therefore, become associated. This accidental association may be enhanced by the removal from suspension in seawater of the relatively coarse quartz fraction of the sialic component from the clay minerals of which the sialic detritus is primarily composed.

About 40 percent of the non-carbonate sediment consists of clay minerals of which illite is the most abundant with measured concentrations ranging from nearly 13 to 43 percent and averaging about 25 percent of the sediment. Kaolinite concentrations vary between 6 and 19 percent and average about 13 percent of the non-carbonate fraction. Chlorite is the least abundant clay mineral with concentrations ranging from less than 1 to 2.5 percent and averaging about 1.5 percent of the sediment. The average clay-mineral content of cores 3P and 153P is more than 42 percent, whereas that of core 4P
Figure 40. Variation of the absolute illite, kaolinite, and chlorite concentrations vs basaltic composition.
is less than 35 percent. The differences in the clay-mineral contents of these cores is consistent with the assumption that cores 3P and 153P represent a depositional environment which is receiving more fine-grained detritus than is core 4P. This interpretation is compatible with a similar conclusion based on the feldspar concentration results. The measured illite, kaolinite, and chlorite concentrations were averaged and plotted with reference to the incremental scale of the basaltic composition in Figure 40. The concentrations of the clay minerals in core 3P and, to a lesser extent, in the other cores have maxima in increment number 4 (15 to 20 percent basaltic content). The explanation is that the clay-mineral concentrations decrease with increasing volcanogenic content due to the unweathered nature of this component. The clay-mineral content of the sialic detritus is lowered presumably because a significant fraction of the weathering products are poorly ordered clays and amorphous oxides of iron, aluminum, and manganese which are not directly detectable by X-ray diffraction. The presence of iron oxides which are associated with the sialic-rich sediment is consistent with the Fe$_2$O$_3$ concentrations presented in Chapter V.

The variations of the clay-mineral concentrations of cores 153P and 4P are similar but not as definitive as those of core 3P due to the narrower range of basaltic composition of these cores.

The content of X-ray amorphous material ranges from 3 to more than 50 percent of the non-carbonate fraction of the sediment. The patterns of the X-ray amorphous concentrations are similar (Figure 41) in that all three cores have minima in increment number 4 (15 to
Figure 41. Variation of the X-ray amorphous content vs basaltic composition.
20 percent basaltic content) or number 5 (20 to 25 percent basaltic content). The high X-ray amorphous content of the sialic-rich sediment is consistent with the assumption that the weathering products derived from this source contain significant amounts of disordered clays and amorphous oxides of iron, aluminum, and manganese. The abundance of X-ray amorphous material in samples containing high concentrations of volcanogenic detritus reflects the presence of volcanic glass which was observed in the +100 mesh fractions of the sediment. The mineral compositions of the sediment samples vary systematically as a function of the basaltic concentrations which was derived from the isotopic compositions and concentrations of strontium by means of the two-component model. Therefore, although these results were determined independently, they are internally consistent. This implies that the litho-stratigraphies of the cores defined on the basis of isotopic criteria are valid.
CHAPTER VIII

INTERPRETATION OF THE CALCIUM CARBONATE
CONCENTRATIONS AND $^{18}$O VALUES

The concentrations of calcium carbonate of the -100 mesh fractions were determined from the weight loss of the samples after leaching with dilute HCl (Chapter IV). Carbonate concentrations measured in this manner range from about 19 to 87 percent and average more than 50 percent of the sediment. The high carbonate content of the sediment samples is consistent with the lithologic logs of the piston cores used in this study wherein the sediment is described as silty calcareous-ooze. Furthermore, these results are consistent with calcium carbonate concentrations reported by Milliman et al. (1969) and Ku et al. (1969).

Core 153P is the northernmost core and has the lowest average carbonate content (49%). Core 3P has the highest average concentration (70%) and is the southernmost site. Cores 4P and 154P have carbonate concentrations that average 53 and 56 percent, respectively, and are located between the other cores (Figure 8). This systematic increase of the carbonate concentrations in a southerly direction is consistent with previously reported results (Milliman et al., 1969). Furthermore, this trend of the carbonate concentrations may be due to climatic factors which have affected the pattern of carbonate deposition in the Red Sea. In order to
test for a possible relationship, the calcium carbonate concentrations and basaltic compositions of the sediment samples were compared as discussed below.

The carbonate concentrations were averaged with reference to the incremental scale of the concentrations of the basaltic component (Chapter VI) and plotted in Figure 42 as a function of the average basaltic composition of each increment. Lines were drawn through the data points to emphasize the trends of the results. The point in increment number 7 of core 3P was disregarded because it represents a single sample. These results indicate that the rate of carbonate deposition in the Red Sea increased at times when the rate of deposition of sediment derived from old sialic rocks was also increasing. This association suggests that both processes may have been controlled by the same mechanism. Additional information concerning this phenomenon is provided by the δ¹⁸O values of foraminifera tests from core 154P which is located only 18 km southeast of core 153P.

Deuser and Degens (1969) report that the δ¹⁸O values of foraminifera tests from core 154P show a pattern of gradual increase followed by rapid decrease which is repeated four times over the length of the core (Figure 43). The observed variations of this parameter were attributed to fluctuations in the salinity of the Red Sea. Increases in δ¹⁸O values represent high salinity due to excessive evaporation whereas decreases in the δ¹⁸O values are associated with lowered salinity due to "freshening" of the Red Sea. The
Figure 42. Plot of the calcium carbonate concentrations of the -100 mesh fractions vs basaltic composition for cores 3P, 153P, and 4P.
Figure 43. Core 154P: $\delta^{18}O$ values of foraminifera tests made available by W. C. Deuser.
exchange of water between the Red Sea and the Indian Ocean is limited by the restrictive nature of the Straits of Bab el Mandeb (Chapter II). Gradual lowering of sea level in response to the formation of continental glaciers in the northern hemisphere further inhibited circulation, causing the salinity of the Red Sea to increase as indicated by high $\delta^{18}O$ values. Re-establishment of more effective circulation with the Indian Ocean as sea level rose due to melting of continental glaciers resulted in "freshening" of the Red Sea as indicated by the decrease in the $\delta^{18}O$ values.

In order to test for a possible relationship between the variations of the $\delta^{18}O$ values of foraminifera tests reported by Deuser and Degens (1969) and the concentration of sediment derived from old sialic rocks, the concentrations of calcium carbonate and strontium were determined for 50 samples from core 154P (Chapter IV). The basaltic compositions of these samples were calculated from their measured strontium concentrations by means of the mixing equation derived previously for core 153P (Chapter V, Equation (35)). This calculation required the assumption that the samples from core 154P satisfied the mixing equation of core 153P even though they were not used in the derivation of this equation. This assumption is likely to introduce negligible error, because these cores were collected from similar depositional environments in the median valley separated by only 18 km. Furthermore, mixing equations derived from the analytical results of cores separated by more than 200 km are not significantly different (Chapter V).
The calcium carbonate concentrations were plotted, as before, with reference to the incremental scale in Figure 44. The trend of the data points indicates that the rate of carbonate deposition in core 154P increased with increasing concentration of sediment derived from old sialic rocks. These results are identical to those described earlier for cores 153P, 3P, and 4P.

The $\delta^{18}O$ values of the samples from core 154P used in this study were interpolated from results made available by W. C. Deuser (Figure 43). The interpolated $\delta^{18}O$ values were averaged with reference to the incremental scale of basaltic concentrations and plotted as a function of the average basaltic composition of each increment in Figure 44. The data points fit a straight line having a positive slope, indicating that increasing input of sediment derived from old sialic rocks coincided with decreasing $\delta^{18}O$ values of the foraminifera tests in response to "freshening" of the Red Sea.

These results suggest that the input of sediment from old sialic rocks and deposition of carbonate sediment in the Red Sea were both related to decreasing salinity as indicated by the $\delta^{18}O$ values of foraminifera tests. Deuser and Degens (1969) related the salinity fluctuations to eustatic changes in sea level reflecting the waxing and waning of continental glaciers in the northern hemisphere. Therefore, the input of sediment of sialic composition appears to be controlled by climatic factors associated with the melting of ice masses in the northern hemisphere.
Figure 44. Core 154P: Calcium carbonate concentrations of the -100 mesh fractions and $^{18}O$ values plotted as a function of their basaltic composition.
Figure 44.
CHAPTER IX

TENTATIVE CHRONOLOGY OF QUATERNARY SEDIMENTATION

The discussion presented in the previous chapter suggests that sedimentation in the Red Sea was indirectly influenced by the accumulation and melting of continental ice masses in the northern hemisphere. Therefore, it may be possible to relate the sedimentary profiles derived in Chapter V to the Quaternary history of the northern hemisphere. In order to test this hypothesis, the chronology developed in Chapter V was used to plot the concentrations of the volcanogenic detritus in cores 153P, 3P, and 4P as a function of their age in Figure 45. The time-scale for the last 20,000 years was taken directly from Figure 37 (Chapter VI). The ages of the sediment deposited between 20,000 and 90,000 years ago were determined from a linear extrapolation of the \(^{14}C\) dates for cores 153P, 154P, 3P, and 4P. Also shown in Figure 45 are mean annual temperatures and Quaternary chronostratigraphy compiled by van der Hammen et al. (1967).

Comparison of the sedimentary profiles with the chronostratigraphy in Figure 45 indicates that sediment layers enriched in the sialic component coincide rather well with interstadials representing brief warming trends during the Würm glaciation. The depth intervals and extrapolated ages of the sialic-rich layers are listed in Table 8 along with Interstadial ages reported by van der Hammen et al. (1967).
Table 8. Tentative correlation and ages of sialic-rich layers identified in cores 153P, 3P, and 4P with interstadials during the Würm Glaciation.

<table>
<thead>
<tr>
<th>Depth interval in core 153P units, cm.</th>
<th>Extrapolated $^{14}$C age, years</th>
<th>Interstadial (van der Hammen, et al., 1967)</th>
<th>Age, years B. P.</th>
</tr>
</thead>
<tbody>
<tr>
<td>140-216</td>
<td>10,000-20,000</td>
<td>Allerød</td>
<td>11,000-11,800</td>
</tr>
<tr>
<td>300-340</td>
<td>29,000-33,000</td>
<td>Denekamp</td>
<td>29,000-32,000</td>
</tr>
<tr>
<td>400-460</td>
<td>38,500-44,000</td>
<td>Hengelo</td>
<td>37,000-39,000</td>
</tr>
<tr>
<td>618-680</td>
<td>59,000-65,000</td>
<td>Brézup</td>
<td>about 59,000</td>
</tr>
<tr>
<td>780-870</td>
<td>78,000-85,000</td>
<td>Amersfoort</td>
<td>about 63,500</td>
</tr>
<tr>
<td>780-870</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Figure 45. Apparent coincidence of sediment layers enriched in the sialic component with episodes of warmer climate and interstadials during the Würm Glaciation. The temperature curve and chronostratigraphy are from van der Hammen et al. (1967).
Figure 45.
The youngest sialic-rich layer (140-216 cm) was deposited between 10,000 and 20,000 years ago and may be representative of the Allerød and Bølling Interstadials which occurred from 11,000 to 11,800 years B. P. and 12,000 and 12,400 years B. P., respectively. The second layer between 300 and 340 cm (29,000 to 33,000 years B. P.) correlates well with the Denekamp Interstadial between 29,000 and 32,000 years B. P. The third sialic-rich layer (400-460 cm) was deposited between 38,500 and 44,000 years ago and appears to predate the Hengelo Interstadial (37,000 to 39,000 years B. P.) by a few thousand years. This discrepancy may be due to interference from the sediment layer enriched in volcanogenic detritus deposited immediately above the sialic-rich layer. The Brørup (about 59,000 years B. P.) and Amersfoort (about 63,500 years B. P.) Interstadials appear to be merged in one sialic-rich layer between 618 and 680 cm (59,000 to 65,000 years B. P.) in core 3P and to a lesser extent in core 4P. This layer was not identified in core 153P because the sediment deposited before about 50,000 years ago in this core was disturbed during coring. Finally, cores 3P and 4P record a well-defined sialic-rich layer from 800 to 880 cm dated between 78,000 and 85,000 years ago. This layer was deposited during the Interglacial between the Riss and Würm Glacial periods.

The apparent coincidence of sediment layers enriched in the sialic component with interstadials during the Würm glaciation suggests the possibility that the input of sialic detritus may be controlled by changing climatic conditions. If precipitation
and continental runoff were increased during warming trends represented by the interstadials, then the proportion of detritus derived from old sialic rocks would have increased because these rocks are exposed over a much larger area than are the volcanic rocks in the Red Sea region. Several workers have argued for and against the relationship between high-latitude glacial and interglacial periods with low-latitude pluvial and interpluvial stages (Flint, 1959; Butzer et al., 1972; Parmenter and Folger, 1974; Newell et al., 1975). The results of these studies are ambiguous in that some authors suggest a correspondence between glacial and pluvial periods, whereas others argue for an interglacial-pluvial relationship. General agreement may not be possible because of the shift of climatic belts during the transition from glacial to interglacial periods which may result in wet conditions at one location while dry conditions prevail in other areas. Therefore, if a relationship between high-latitude glacial or interglacial periods and low-latitude pluvial or interpluvial stages does exist, then this correspondence may best be studied in isolated locations. In the case of the Red Sea region, Deuser et al. (1976) suggest that high-latitude interglacial periods correspond with pluvial stages. Furthermore, a study of the sediment collected during the Deep Sea Drilling Project at Site 228 in the Red Sea (latitude 19°05.16'N, longitude 39°00.20' E) indicated that considerable amounts of detritus were derived from a terrigeneous source as a result of pluvial periods in the Pleistocene (Whitmarsh et al., 1974). According to the authors, these pluvial periods resulted in increased
continental runoff, and detritus derived from the old, sialic rocks of the Sudan was deposited along the western continental-shelf of the Red Sea. If high-latitude interglacial periods correspond to pluvial periods in the Red Sea area, then it is possible that high-latitude interstadials may correspond with wetter climatic conditions in the Red Sea area as well, although this association may be less pronounced. If precipitation were increased during warming trends represented by the interstadials, then the proportion of detritus derived from old sialic rocks would have increased because of the large exposures of these rocks. This interpretation is consistent with the coincidence of the sialic-rich sediment layers with the interstadials during the Warm Glacial period as shown in Figure 45.
CHAPTER X
HISTORY OF SEDIMENTATION IN THE RED SEA
DURING THE PAST 90,000 YEARS

It is now possible to present a detailed history of sedimentation for the Red Sea based on the sedimentary record of cores 153P, 3P, and 4P. The sediment deposited before about 50,000 years ago in core 153P was disturbed in coring. However, cores 3P and 4P contain undisturbed sediment that was deposited about 90,000 years ago (Figure 45). At this time, detritus derived from young volcanic rocks was contributing about 20 percent of the sediment in core 3P and about 15 percent in core 4P. The difference in the basaltic composition of the sediment in these cores implies that core 3P was closer to the source of the volcanogenic detritus as discussed in Chapter V. Between about 85,000 and 78,000 years ago, the abundance of volcanic detritus systematically decreased to 6 percent in core 3P and 10 percent in core 4P, presumably due to the increased input of sediment derived from old sialic rocks. This sialic-rich layer may represent changing climatic conditions in the Red Sea in response to the beginning of the Würm Glacial period.

From about 78,000 to 66,000 years ago, the basaltic composition of the sediment gradually increased from 15 to 18 percent in core 3P and from 12 to 16 percent in core 4P. However, during the next 5,000
years, the warming trends associated with the Amersfoort and Brörup
Interstadials apparently resulted in more humid conditions in the
Red Sea area causing increased continental runoff and subsequent
enrichment of the sediment in detritus derived from old sialic rocks,
thereby increasing the concentration of sialic detritus about 5 percent
in core 3P.

Between 59,000 and 45,000 years ago, the basaltic composition
of the sediment systematically increased from about 18 to 22 percent
in cores 3P and 4P. Beginning about 45,000 years ago, the concentra-
tion of sialic detritus was increased by about 10 percent in cores
3P and 4P, and to a lesser extent in core 153P. The enrichment of
the sediment in the sialic component was presumably a response to
climatic changes associated with the Hengelo Interstadial. However,
this sialic-rich layer was abruptly terminated by a volcanic event
about 37,000 years ago. The cause of the basaltic-rich layer is pro-
bably the eruption of volcanoes within the Red Sea basin or adjoining
areas and therefore, has no relationship with changing climatic
conditions. This event systematically increased the basaltic compo-
sition of the sediment from 11 to 34 percent in core 3P, from 17 to
27 percent in core 153P, and from 14 to 21 percent in core 4P.
Comparison of these changes in the basaltic composition of the
sediment with the locations of the cores (Figure 8) indicates that
the source of the volcanogenic detritus may have been west and
slightly north of core 3P.

From about 33,000 to 29,000 years ago, the sediment was enriched
in detritus from old sialic rocks, apparently resulting from increased
precipitation and continental runoff associated with the warming trend during the Denekamp Interstadiial. The first iron-rich layer was deposited in core 153P at this time, increasing the Fe$_2$O$_3$ concentration of the sediment about 13 percent (Chapter V). A volcanic event about 26,000 years ago terminated the sialic-rich layer and increased the concentration of the basaltic component in the sediment from about 14 to 26 percent in core 153P, from 21 to 27 percent in core 3P, and from 18 to 21 percent in core 4P. These differences in the basaltic composition of the sediment indicate that the volcanic source may have been west and slightly south of core 153P.

Beginning about 20,000 years ago and continuing for the next 10,000 years, the sialic composition of the sediment gradually increased, presumably as a result of changing climatic conditions near the end of the Würm Glacial period. Also during this time, the most recent iron-rich layer was deposited in core 153P, increasing the Fe$_2$O$_3$ concentration of the sediment by about 18 percent. Sediment deposited more recently than about 10,000 years ago is missing from cores 3P and 4P, and consequently, the sediment record of these cores ends at this time. However, the sediment profile of core 153P indicates the presence of a well-defined volcanic event about 8,000 years ago which increased the concentration of the basaltic component in the sediment from 13 to 23 percent. This event is followed by an apparent sialic-rich layer (about 4,000 years ago) which is terminated by a volcanic event (2,500 years ago). However,
these events may be spurious due to the disturbance of the upper sediment layer during the coring operation.

The objective of this research was to use isotopic and geochemical methods to develop a technique for the study of sedimentary processes and to interpret changing sedimentary patterns in terms of the depositional history of a basin. A technique based on the isotopic compositions and concentrations of strontium was developed and applied to the detrital, noncarbonate fractions of sediment samples taken serially from piston cores collected in the Red Sea. It was shown that this technique provided precise information concerning the provenance of the sediment, and led to the development of correlatable stratigraphies which were used to establish a history of sedimentation in the Red Sea basin.

This approach to the study of sedimentary processes should be applicable to marine or nonmarine sedimentary basins of any age, provided that these basins are receiving detritus derived from sources of contrasting chemical composition. In the case of sediment deposited in the past, the age of deposition must be determined, and the measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios recalculated to correct for the amount of radiogenic $^{87}\text{Sr}$ which has accumulated since deposition of the sediment. Once this has been accomplished, the mixing model can be applied as though the sediment were of Recent age. Furthermore, this approach to the study of sedimentary processes should be equally applicable to mixtures of components having different isotopic compositions and concentrations of lead as well as strontium. The isotopic composition of lead is changed by the
decay of $^{238}\text{U}$ to $^{206}\text{Pb}$, $^{235}\text{U}$ to $^{207}\text{Pb}$, and $^{232}\text{Th}$ to $^{208}\text{Pb}$. Mixing equations relating the lead isotopic compositions and concentrations of the components to those of the mixtures can be defined in a manner similar to the strontium mixing equations derived in Chapter III.

The application of this alternative approach to the study of sedimentary processes could provide independent corroboration of the mixing interpretation for a particular depositional regime.
REFERENCES


In order to show that Equation (14) is hyperbolic, it is necessary to begin with the general equation that describes all quadratic curves of the second order:

\[ Ax^2 + Bxy + Cy^2 + Dx + Ey + F = 0 \]  

(A1)

Rewriting Equation (14) in this form yields:

\[ R_M S_{r_M} - b S_{r_M} - a = 0 \]  

(A2)

If \( S_{r_M} = x \) and \( R_M = y \), then Equations (A1) and (A2) can be compared. In this case, \( A = C = E = 0; B = 1; D = -b; \) and \( F = -a \). The presence of the mixed term, \( Bxy \), indicates that the coordinate axes are not symmetry axes. This is true for any curve generated by Equation (A2), because \( R_M \) and \( S_{r_M} \) are always positive numbers. However, the standard equation of a quadratic curve is stated so that one of the coordinate axes is a symmetry axis. Therefore, it is necessary to eliminate the mixed term in Equation (A2) by redefining the coordinate axes. This is accomplished by rotating the coordinate axes through an angle, \( \alpha \),
which is determined by the values of the constants $A$, $B$, and $C$:

\[
\cot 2 \alpha = \frac{(A - C)}{B} \quad (A3)
\]

The value of $\alpha$ must be $45^\circ$ because $A = C = 0$. $R_M$ and $Sr_M$ can now be restated in terms of the new coordinates which have been rotated by $45^\circ$:

\[
Sr'_M = Sr_M \cos \theta - R_M \sin \theta = \left(\frac{1}{2}\right)(Sr_M - R_M) \quad (A4)
\]

\[
R'_M = Sr_M \sin \theta + R_M \cos \theta = \left(\frac{1}{2}\right)(Sr_M + R_M) \quad (A5)
\]

Substitution of Equations (A4) and (A5) in Equation (A2) yields:

\[
\frac{1}{2}Sr'_M^2 - \frac{1}{2}R'_M^2 - \left(\frac{b}{2}\right) Sr_M + \left(\frac{b}{2}\right) R_M - a = 0 \quad (A6)
\]

Equation (A6) can now be rewritten in a standard quadratic form by completing the square and rearranging the terms:

\[
\frac{(Sr_M - (b/2))^2}{2a} - \frac{(R_M - (b/2))^2}{2a} = 1 \quad (A7)
\]

The standard equation for a hyperbola is:

\[
\frac{(x - h)^2}{c^2} - \frac{(y - k)^2}{d^2} = 1 \quad (A8)
\]
Comparison of Equations (A7) and (A8) shows that these equations have the same form where:

\[ h = k = \frac{b}{2} \]  \hspace{1cm} (A9)

and

\[ c = d = -\sqrt{2a^3} \]  \hspace{1cm} (A10)

Therefore, any curve generated by Equation (A2) will be a hyperbola. The asymptotes of the hyperbola will meet at right angles, because \( c = d \). Because \( h \) and \( k \) are not zero, the center of the hyperbola will be displaced from the origin. However, the transverse axis of the hyperbola will pass through the origin at an angle of 45° with respect to the original coordinate axes. This property provides sufficient symmetry to enable any hyperbola generated by Equation (A2) to be transformed into a straight line by redefining one of the coordinate axes:

\[ \beta = 1/ \sqrt{\gamma M} \]  \hspace{1cm} (A11)

Then Equation (A2) can be rewritten as:

\[ R_M = a \sqrt{\beta} + b \]  \hspace{1cm} (A12)

This is the equation of a straight line in coordinates of \(^{87}\text{Sr}/^{86}\text{Sr} \)
ratio and $\beta$, the inverse of the strontium concentration.
Appendix B: Summary of the Analytical Results.

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Core 153P

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