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THE PETROGENESIS AND TECTONIC SETTING
OF THE EGYPTIAN ALKALINE COMPLEXES

By

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ABSTRACT

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The Egyptian alkaline complexes are similar in composition to many alkaline suites in other areas, and are considered to represent highly differentiated, mantle-derived alkaline basaltic magmas. Low-pressure crystal fractionation is considered to have produced the initial bulk salic compositions; magma zoning in the remaining halogen-volatile-rich alkaline melt is considered to account for the radial arrangement of silica-undersaturated and oversaturated rock-types and the associated isotopic (Lutz, 1979) and chemical gradients.

The Nubian shield provided a stable crustal setting which allowed the long-term emplacement and similar quiescent differentiation of these melts along the major Pan-African lineaments throughout portions of the Phanerozoic eon: 554, 404, 351, 224, 140, and 90 m.y. B.P. These melts are considered to have been generated in the asthenosphere in response to shear-heating between the lithosphere and the mesosphere during periods of major plate accelerations.

Dedicated to
Jacqueline and
Matthew Robert Allan

ACKNOWLEDGMENTS

I thank Tom for his patience, help and friendship-- particularly considering the time and distance associated with the following de Gruyter space. I regret the delay and the freeze frame of friends I left behind. Those frames never come back. I do remember one particular frame of a different sort that turned up in a slide presentation of mine (who's chest was that again?). T² and Mongobongowhere-didyoutgetyourcongo-congolas? (This is the only type of acknowledgments anyone expected from me.)

So Tom...with tipped Stellas, here's to Eastern Desert days and nights, Aswan gozas, the infamous ascent of Nigrub El Fogani, the House that Jack Built, Ocum's Razor (heh-heh) and those power spots that persist in Desert rock and sand.

I wished for one more quarter's worth of more leisurely talking, writing and enjoying. Mongo and I would have surely walked at least once across the expanse of the mantle.

To T² who helped me by example more than he realized (although I learn slowly). Thanks for friendship and for finding a woman who will finally make something out of you.

I want to extend special thanks across the many miles to Ahmed Abdallah Abdel Megid for Desert assistance, smiles and a open friendliness to a "hkawaga".

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Thanks does not cover the patience and hard work that Jackie has given me: sample sawing and grinding, help with diagrams, enormous patience in my on/off writing for this thesis and attempts at articles, and most of all for listening to my ramblings, and keeping the warmth in our home--always. Asequim Acundúa.

If there is light in this, it is solely due to His faithfulness to me when I was not to Him.

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GENERAL INTRODUCTION

This thesis presents research pertaining to the geochemistry and tectonic setting of alkaline complexes located in the Eastern Desert of Egypt. The text is divided into two parts: Part I deals with the geochemistry and petrogenesis--primarily based on the rare earth elements and thorium; Part II discusses the tectonic setting of alkaline magmatism in general, and relates the Egyptian alkaline province to the tectonic development of Afro-Arabia.

The field work was conducted in the spring of 1978. Approximately 290 samples were collected from the four alkaline complexes visited by this study: Gabal Abu Khrug, Gabal El Kahfa, Gabal El Naga, and Gabal Nigrub El Fogani. The radiometric dates for these complexes were not known at the time of the field work, but were later supplied by workers at the University of Pennsylvania (see Lutz, 1979; Serencsits et al., 1981).

About 200 thin sections were prepared and studied. Two hundred and eleven samples were analyzed for rare earth elements (REE) and thorium. Forty-five samples were analyzed out of house for major elements and 15 trace elements. This study also used major element data from El Ramly et al., 1969a, 1969b, and 1971, as well as Rb-Sr data collected by Lutz, 1979.

PART I

THE GEOCHEMISTRY AND PETROGENESIS OF THE EGYPTIAN ALKALINE COMPLEXES

Introduction

The first objective of this study is to investigate the petrogenesis of some of the alkaline complexes in the Eastern Desert of Egypt. Chapter One introduces the general setting of the Egyptian alkaline province, with a brief description of the petrography and general structure of the complex which was selected for more detailed study (Gabal Abu Khruq). A more complete description of the four alkaline complexes sampled by this study is presented in Appendix I. The second chapter consists of a discussion of the models proposed for the origin of alkaline magmas. The third chapter presents the chemical data on the Egyptian alkaline complexes. The fourth chapter involves a discussion of the petrogenesis of these alkaline complexes, focussing on Gabal Abu Khruq. This involves the REE geochemistry of these alkaline complexes, as well as a comparison to alkaline suites in other areas.

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CHAPTER 1

THE GENERAL SETTING AND NATURE OF THE EGYPTIAN ALKALINE COMPLEXES

The Egyptian alkaline province is located between the Red Sea and the Nile River, predominantly in the southern portion of the Egyptian Eastern Desert, Figure 1. This area is part of the exposed crystalline basement of the Nubian shield, consisting of Precambrian amphibolitic and granitic gneisses, metavolcanics, metasediments, and an abundance of Pan-African salic intrusions. Considering the profusion of alkaline magmatic centers in Egypt, as well as Afro-Arabia in general, this area may represent one of the largest alkaline provinces in the world (Vail, 1970, 1976, 1978; El Ramly et al., 1971; Baker et al., 1973; Gass, 1977; Almond, 1977; Shimron, 1975; Hahn et al., 1976).

The Egyptian alkaline complexes are located along the major lineaments of the Nubian shield (Garson and Krs, 1976; El Ramly et al., 1969a, 1969b, 1971; Lutz, 1979). (See Part II). The radiometric dates of these complexes range over a period of more than 450 m.y., from 554 to 89 m.y. B.P. (Serencsits et al., 1981; Hashad and El Reedy, 1979; Lutz, 1979). In spite of their ages spanning most of the Phanerozoic eon, they are chemically and mineralogically similar to each other as well as numerous alkaline suites in other areas. The strontium and oxygen isotopic data from ten alkaline complexes are consistent with their

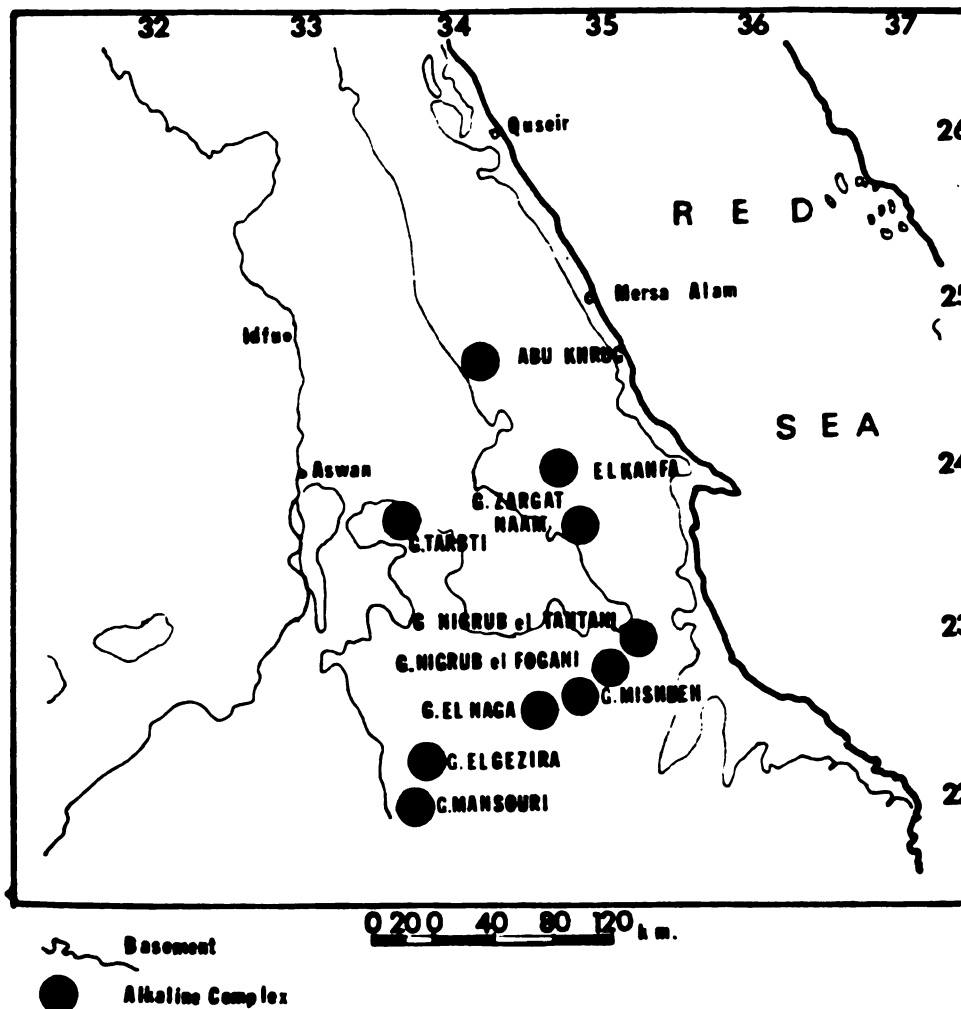


Figure 1.

Some of the major alkaline complexes in the Eastern Desert of Egypt (after El Ramly et al., 1971).

* Refer to Figures 44 and 45 for tectonic map and relationship of the study area to the African continent.

origin as mantle-derived melts with no significant crustal contamination (Lutz, 1979; e.g. initial $\text{Sr}^{87/86} = .703$).

The Egyptian alkaline complexes may be generally described as excavated, volcanic structures with a wide variety of sodic-alkaline plutonic and effusive rock-types. The radiometric dates and structural relationships of these different rock-types indicate that each complex represents a single discrete magmatic event and the rock-types do not represent separate and/or unrelated magmatic events. The complexes sampled by this study (Gabal El Naga, Gabal Nigrub El Fogani, Gabal El Kahfa, and Gabal Abu Khruq) are dominated by sodic, miarolitic, silica-undersaturated and oversaturated syenites which are compositionally near the agpaitic-miaskitic boundary. Other rock-types include gabbros, alkaline trachytes and rhyolites, and a wide variety of dike rocks. These four alkaline complexes represent part of the two youngest episodes of alkaline magmatism in Egypt which are known to have formed major alkaline complexes. The older episode (G. El Naga and G. Nigrub El Fogani) occurred during the Late Jurassic time; the younger episode (G. El Kahfa and G. Abu Khruq) occurred during the Late Cretaceous time (Serencsits et al., 1981; Lutz, 1979). All four complexes contain both silica-undersaturated and oversaturated rock-types. The older complexes which were not sampled by this study, tend to consist of only silica-oversaturated rock-types (Lutz, 1979).

All four complexes of this study are well-developed ring structures, rather than isometric masses or poorly-developed ring structures. Consistently, the silica-over-saturated syenites form the outer portions of the complexes, while the silica-undersaturated syenites are found in the central portions. The remains of the overlying volcanic cones tend to consist of evolved, siliceous and altered flows. Based on limited exposures and comparisons to similar alkaline complexes in other areas, gabbros and ultramafic cumulates are inferred to form an underlying substratum to these salic ring structures. Throughout these alkaline complexes there are examples of gradational boundaries between some of the coarse-grained rock-types with their corresponding chemical and mineralogic gradients. There are also many examples in which the boundaries between rock-types are obscured by faults, alteration zones, dikes, erosion, or colluvium. Overall, however, there is a scarcity of sharp contacts between the coarse-grained rock-types. As will be discussed, these complexes may represent a magma (or pulses of a magma) which contained a radial compositional gradient (i.e. liquid zoning). Such an approach is further supported by similar alkaline complexes in other areas which display more intact, unfaulted gradational zonation (e.g. Pankhurst et al., 1976).

Gabal Abu Khruq represents the largest alkaline complex of this study. It is the most studied alkaline complex in Egypt due to its economic potential, degree of preservation,

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size, and wide range of rock-types (El Ramly et al., 1969a, 1971; Serencsits et al., 1981; Lutz, 1979). This complex is typical for many of the Egyptian alkaline complexes, and serves as an example of the type of structure and mineralogy of these alkaline ring complexes.

G. Abu Khruq is a sodic silica-undersaturated/oversaturated, gabbro-syenite-volcanic alkaline complex, Figure 2. The major rock-types are 1) olivine-rich cumulates (wehrlite), 2) labradorite-rich gabbros, 3) pulaskite syenites (silica-undersaturated, small proportion of modal nepheline), 4) alkaline syenites (silica-oversaturated, little or no modal quartz), 5) quartz syenites (5-25% modal quartz), 6) alkaline trachytes and rhyolites (silica-oversaturated), 7) foyaite syenites (strongly silica-undersaturated, nepheline and sodalite-rich). The overall mineralogy of these rock-types is characterized by a predominance of euhedral laths of alkali feldspar in the syenites and volcanics, and labradorite in the gabbros. This is the "agpaitic order of crystallization" common for many alkaline suites (see Kogarko et al., 1978). Pyroxene is the dominant ferromagnesian mineral (aegerine-augite in the syenites and volcanics; diopside in the gabbros). Nepheline and sodalite are abundant in the central silica-undersaturated syenites (foyaite); quartz (both primary and secondary) becomes progressively more abundant in the outer perimeters of the complex. The syenites are dominantly agpaitic, but also range into the miaskitic compositions (see Sørensen, 1974, p. 6).

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El Ramly et al. (1969a) suggest that the original size of the volcanic cone at Abu Khruq was about 9 kilometers in diameter and 1.5 kilometers in height. The present complex consists of a large discontinuous outer ring and an inner ring with poorly-defined stock-like bodies of syenite. Two major fault systems dissect the complex: one runs NNW through the western portion of the complex, the other runs approximately NW-SE through the inner and outer rings of the complex, Figure 2. These fault systems are inferred to be part of the major lineaments of the Nubian shield (see Part II). Conical faults are inferred to exist in the low-lying, flat wadis between the rings. These local collapse surfaces probably occurred during and after the latter stages of crystallization of the melt. The surrounding country rock is made up of amphibole schists and gneisses, with a large granitic body directly against the eastern side of the complex.

The increasing silica content from the center of the complex to the margin, is one of the most salient features of Abu Khruq, as well as the other complexes visited. The central foyaites are highly silica-deficient; the syenites surrounding the foyaites tend to be slightly silica-under-saturated (pulaskites). The southeastern portion of the outer ring is also composed of pulaskites and foyaitic syenites. The slightly silica-oversaturated syenites (alkaline syenites) tend to be found in the outer portions of the inner ring and portions of the outer ring. Other

Figure 2.

Geologic map of Gabal Abu Khruq (revised from El Ramly et. al., 1969a and 1971). Numbers denote sample locations. Sample numbers with prime (') symbol are the "Eg-" series; sample numbers without the prime symbol are the "A-" series.



Wadi colluvium



Foyaites



Gabbros



Volcanics



Pukaskites & Alkaline Syenites



Amphibole schists & gneisses



Quartz Syenites



Granite gneiss



Faults

Scale: 1:50,000

1 mile

Figure 2.

Amphibole schists & gneisses

[illegible]

than the southeastern portion of the outer ring, the quartz syenites form the bulk of the outer ring with modal quartz increasing toward the perimeters. There are gradational boundaries between the foyaïtes and pulaskites, between the alkaline syenites and quartz syenites, and although further sampling should be done, slightly silica-undersaturated and slightly silica-oversaturated samples were collected from apparently the same syenite bodies. There is also a lack of clear, sharp contacts between the syenites and the underlying gabbros at Abu Khruq as well as some of the other complexes visited. Remnants of the volcanic cone cap parts of both the outer and inner rings; a highly altered chimney facies is exposed on the Abu Khruq peak in the central ring. Almost the entire volume of these volcanics appear to be siliceous, autometasomatized flows.

It is likely that Abu Khruq and at least some of the other Egyptian alkaline complexes are underlain by a silica-undersaturated ultramafic substratum (olivine and clinopyroxene cumulate phases). Although further investigation is needed (drilling or geophysical detection), the existence of such a substratum is indicated by one sample of an olivine-rich rock-type at Abu Khruq (wehrlite), olivine-clinopyroxene xenoliths in possible diatremes at El Kahfa, and a comparison to similar alkaline complexes in other areas (Upton, 1974; Bridgewater and Harry, 1968; Gastesi, 1969; Munoz, 1969; Ludden, 1977; Fernandez, 1980).

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Due to the gradational boundaries, the partially concealed substratum, and the faulted pattern of the complex, a simple order of crystallization/intrusion is rather obscure. El Ramly et al. (1969a) suggest that the order of occurrence at Abu Khruq was: volcanics, gabbros, alkaline syenites (which include both silica-undersaturated and oversaturated varieties), quartz syenites, and finally, foyaïtes (in order of older to younger). Such an order cannot be explained by a simple crystal fractionation model. Although this order may not be correct, the appearance of such field relationships is important to note.

For a more complete description of the four alkaline complexes sampled by this study, refer to Appendix I.

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CHAPTER 2

PETROGENETIC MODELS FOR PRIMITIVE ALKALINE MAGMAS

1. Origin of Primitive Alkaline Magmas

The primitive initial strontium ratios (average .703) and the chemical characteristics of the Egyptian alkaline complexes are consistent with a mantle origin (see Lutz, 1979). Alkaline basaltic melts, compared to tholeiitic types, are characteristically enriched in alkalies, minor, and incompatible trace elements. This enrichment is a fundamental problem in explaining the origin of primitive alkaline magmas. Primitive alkaline basalts and their differentiates from a wide variety of crustal settings have similar compositional trends (Bowen, 1928; Schwarzer and Rogers, 1974; Tilley, 1958; McBirney and Aoki, 1968; Kay and Gast, 1973; Kogarko, 1977; Miyashiro, 1978; Frey et al., 1978; Allegre, et al., 1981).

McBirney and Aoki (1968) emphasized the similarities between the alkaline oceanic suite of Tahiti and the continental alkaline suites of the Monteregian Hills in Canada. Kay and Gast (1973) showed that alkaline basaltic rock-types often have similar major element, REE, Sr, and Ba patterns. Miyashiro (1978) reviewed the major element and trace element characteristics of alkaline volcanics and noted that similar types occur in all types of crustal settings. In an integrated trace element and major element model on the production of basalts ranging from tholeiites to highly

alkaline basaltic varieties, Frey et al. (1978) noted the similarity of many alkaline basalts within different tectonic settings--as opposed to the variability of the tholeiitic varieties. Schwarzer and Rogers (1974) compiled a large number of major element data on alkaline olivine basalts and their differentiates. They concluded that there were no systematic differences in the major element characteristics of alkaline suites from continental, oceanic, and island-arc tectonic settings. This consistency in the chemistry of alkaline suites may represent a fairly regular alkaline magmatic process on a global basis. It is likely to involve a similar source under both continental and oceanic crust and/or a similar mechanism of enrichment (see Allegre, et al., 1981).

Experimental and natural data has been used to demonstrate that primitive alkaline magmas are generated by small degrees of partial melting of mantle material (Gast, 1968; Allegre et al., 1977; Frey et al., 1978) at greater depths than the source regions for tholeiitic magmas (Green and Ringwood, 1967; Kushiro et al., 1968; Green, 1970; Wilkinson, 1974; Frey and Green, 1974; Maaloe and Aoki, 1977). Although small degrees of partial melting of mantle material (e.g. 2-15%, Frey et al., 1978) can account for a certain amount of their enriched character, petrogenetic modelling has caused workers to conclude that the source regions must also be initially enriched in certain components (e.g. Kay and Gast, 1973; Shimizu and Arculus,

1975; Sun and Hanson, 1975; Frey et al., 1978). These enriched zones in the mantle must be explained in order to truly understand the origin of alkaline magmas.

Recent work has implemented volatile processes within the mantle in order to account for this enrichment. The role of volatiles in highly-evolved alkaline magmas has been emphasized for a long time (Symth, 1913; Kogarko, 1974). However, their presence and activity in the mantle is of current debate and has direct bearing on the origin of alkaline magmas in the more recent models. One approach invokes volatiles bound in the mantle's crystallized phases (e.g. Eggler, 1976; Newton and Sharp, 1975; Wyllie, 1979), and another approach suggests that they exist as vapor phases (e.g. Mysen and Holloway, 1977; Mysen, 1977b). Regardless of its form of residence, once heating begins, the result is some type of metasomatic process. In a comprehensive model on basalt petrogenesis, Frey et al. (1978) use a melt or fluid that is possibly supercritically enriched with CO_2 and H_2O to selectively transport the elements which are most strongly incompatible with the mantle phases into concentrations within the low-velocity zone and the base of the lithosphere. This, Frey et al. (1978) contend, will lead to the formation of melts with the observed enriched characteristics of alkaline magmas.

However, this enrichment mechanism is still uncertain. A detailed investigation into the degree of consistency in the enrichment characteristics of alkaline rocks from

different areas of the world would aid in determining whether a regular, reproducible enrichment mechanism seems likely, or whether several and/or erratic mechanisms cause different levels of enrichment. This may have great importance in many alkaline suites--including the Egyptian alkaline complexes. It is clear that volatile phases exert a considerable amount of influence on the composition of the mafic, undifferentiated alkaline melts, and this influence would tend to become even more critical as the melt evolves and becomes more concentrated in volatile and alkali components.

The role of partial melting of a similar, enriched mantle source region provides a regular process for the initial generation of alkaline basaltic magmas. However, not only are mafic alkaline rock-types often similar throughout the world, but given the initial degree of silica-saturation and alkalinity, their differentiates commonly follow similar and very specific compositional trends (e.g. Schwarzer and Rogers, 1974).

2. The Differentiation of Primitive Alkaline Magmas

The majority of the exposed rock-types in the Egyptian alkaline province are highly-differentiated salic varieties. The predominance of salic varieties is characteristic of continental alkaline provinces, and is the subject of considerable debate. Four fundamental models have been proposed for the origin of salic alkaline magmas. One of these involves a crustal source and is unlikely to yield

alkaline magmas with primitive characteristics similar to the Egyptian suites (crustal anatexis and metasomatism: Bailey, 1964, 1970, 1974; Williams, 1970; MacDonald et al., 1970). The three remaining models are:

1. Extraction of salic magmas directly from the mantle (Wright, 1970; Martin, 1974).
2. Fractional crystallization of mantle-derived basaltic magmas (Bowen, 1937, 1945; Saggerson, 1970; Weaver et al., 1972; Barberi et al., 1975; Baker et al., 1977; Baker and Henage, 1977).
3. Liquid fractionation:
 - a) Liquid immiscibility (Philpotts, 1971, 1972, 1974, 1976; Freestone, 1978; Eby, 1979).
 - b) Thermogravitational diffusion or magma zoning (Hildreth, 1977, 1979, 1981; Kogarko et al., 1974; Shaw, 1974; Shaw H.R., et al., 1976).

The presence of mafic end-members in the Egyptian alkaline province, and the continuous compositional trends between the mafic and salic rock-types may be used to discount the model involving the direct extraction of salic alkaline magmas from the mantle for the origin of the Egyptian alkaline complexes.

As will be discussed in the following chapter, there is considerable amount of evidence that fractional crystallization of a mantle-derived alkaline basaltic parent magma was the major mechanism responsible for the development of the salic compositions of the Egyptian alkaline suites.

This is based on isotopic data (Lutz, 1979), field relationships, chemical trends, and a comparison of these trends to well-studied alkaline suites in which fractional crystallization has been demonstrated.

Since the time of Bowen (1937), fractional crystallization in alkaline magmas has often been suggested as the dominant process of differentiation. Later work by Coombs (1963), Kuno (1968), and Coombs and Wilkinson (1969) demonstrated that the compositional variation of many alkaline suites can be explained by the regular crystal fractionation of an alkaline basaltic melt. Trace element data has become widely used for petrogenetic modelling, and many alkaline suites have been shown to be dominated by fractional crystallization (e.g. Zielinski and Frey, 1970; Sigurdsson et al., 1973; Zielinski, 1975; Price and Taylor, 1973; Sun and Hanson, 1976; Rock, 1978; Baker et al., 1977; Baker and Henage, 1977; White et al., 1979). Regular fractional crystallization of alkaline basaltic magmas derived from mantle sources of similar composition and/or similar enrichment mechanisms, might account for much of the similarity in the compositional variation of the salic alkaline rock-types in different parts of the world.

Experimental data and natural studies indicate that alkaline melt systems in particular, demonstrate chemical characteristics well-suited for immiscible liquid relationships. However, this process is presently difficult to detect and verify in slow-cooling magma systems. Although

this study does not discount the operation of this mechanism in the development of the Egyptian alkaline complexes, there is little data to substantiate it as the dominant process of differentiation. There is some evidence that indicates that it may have played a part in attendant mechanisms. There are net-vein complexes at G. Nigrub El Fogani, for example, which contain apparent liquid-liquid contacts between mafic and salic rock-types. The REE relationships between these rock-types are consistent with some of the experimental results of liquid immiscible systems (see Appendix III; Watson, 1975, 1976; Watson and Natlund, 1977; Ryerson and Hess, 1978).

Although fractional crystallization of an alkaline basaltic parent melt is considered to have provided the development of the bulk salic compositions in the Egyptian alkaline complexes, it cannot account for the relationship between the different types of salic rocks. The most evident relationship which cannot be explained by simple crystal fractionation is the presence of both silica-undersaturated and oversaturated salic rock-types within many of the Egyptian alkaline complexes. This study considers that thermogravitational diffusion and its attendant fluid effects are responsible for the development of the silica-oversaturated salic compositions from a silica-undersaturated melt system. This thermogravitational diffusion model is evaluated in the following section and is supported by field relationships and chemical data. In the next sections

it will be argued that the origin of the Egyptian alkaline complexes consisted of efficient crystal fractionation of a primitive alkaline basaltic parent magma with the gradual development of a thermogravitational diffusion mechanism.

CHAPTER 3

THE GEOCHEMISTRY OF THE EGYPTIAN ALKALINE COMPLEXES

Introduction

This chapter will present major element and trace element data on the Egyptian alkaline complexes. This study sampled four of the complexes--all of which are younger complexes composed of both silica-undersaturated and oversaturated rock-types. Approximately 290 samples were collected from the four alkaline complexes visited by this study. About 200 thin sections were prepared; 145 of these were from G. Abu Khruq. Two hundred and eleven samples were analyzed for the REE and Th; 100 of these were from G. Abu Khruq. In addition, 45 samples were analyzed out of house for major elements and 15 trace elements (Barringer Magenta Limited); 22 of these samples were from G. Abu Khruq. This study also uses major element and chlorine data collected by El Ramly et al. (1969a, 1969b, 1971), and the Rb-Sr data collected by Lutz (1979).

The rare earth elements and thorium were analyzed via non-destructive gamma-ray spectrometry using instrumental neutron activation analysis (INAA) as described by Gordon et al. (1968). An Ortec MCA 6240A coaxial Ge-(Li) low energy photon detector was used, with a resolution of 2.08 KeV (FWHM) at 1.33 MeV and a 14.6% efficiency. The samples were analyzed with about a 5-10% analytical precision. The USGS and Canadian (Centres for Mineral and Energy

Technology) standards were used. Values for the concentrations of the standards were obtained from Abbey (1977), Flannagan (1973) and Duffield et al. (1977). The most significant problem in the REE data may be that of Yb. As can be seen in the normalized REE diagrams, Yb concentrations tend to be anomalous. This may be due to 1) analytical problems of which peak interference in certain rock-types may be responsible, or 2) a real Yb anomaly which may be characteristic of these types of salic alkaline rocks. Very similar REE distributions with this anomalous Yb pattern were noted in similar alkaline rock-types from other areas (e.g. Canary Islands, De Paeppe et al., 1971).

1) Major Elements .

A standard way of demonstrating the alkalinity and course of differentiation of alkaline suites is with the use of silica-alkali diagrams (MacDonald and Katsura, 1964; see Miyashiro, 1978). Figure 3 demonstrates the alkaline nature of the Egyptian alkaline complexes with reference to their respective ages. These trends are typical of many alkaline suites in which crystal fractionation is thought to have been a dominant mechanism of differentiation (MacDonald, 1974). It is important to point out the changes in alkalinity and silica-saturation with the ages of the alkaline complexes. The older complexes are composed of only silica-oversaturated rock-types of moderate levels of alkalinity. The younger complexes, on the other hand, are dominated by silica-undersaturated rock-types, but also

Figure 3.

Silica-alkali diagrams for several Egyptian alkaline complexes showing progressive changes in alkalinity and silica-saturation with time. Curved lines represent the boundaries between sub-alkalic and alkalic rocks (after Miyashiro, 1978). The chemical data are from this study and El Ramly et al., 1969a, 1969b, 1971; the radiometric dates are from Serencsits et al., 1981 and Lutz, 1979.

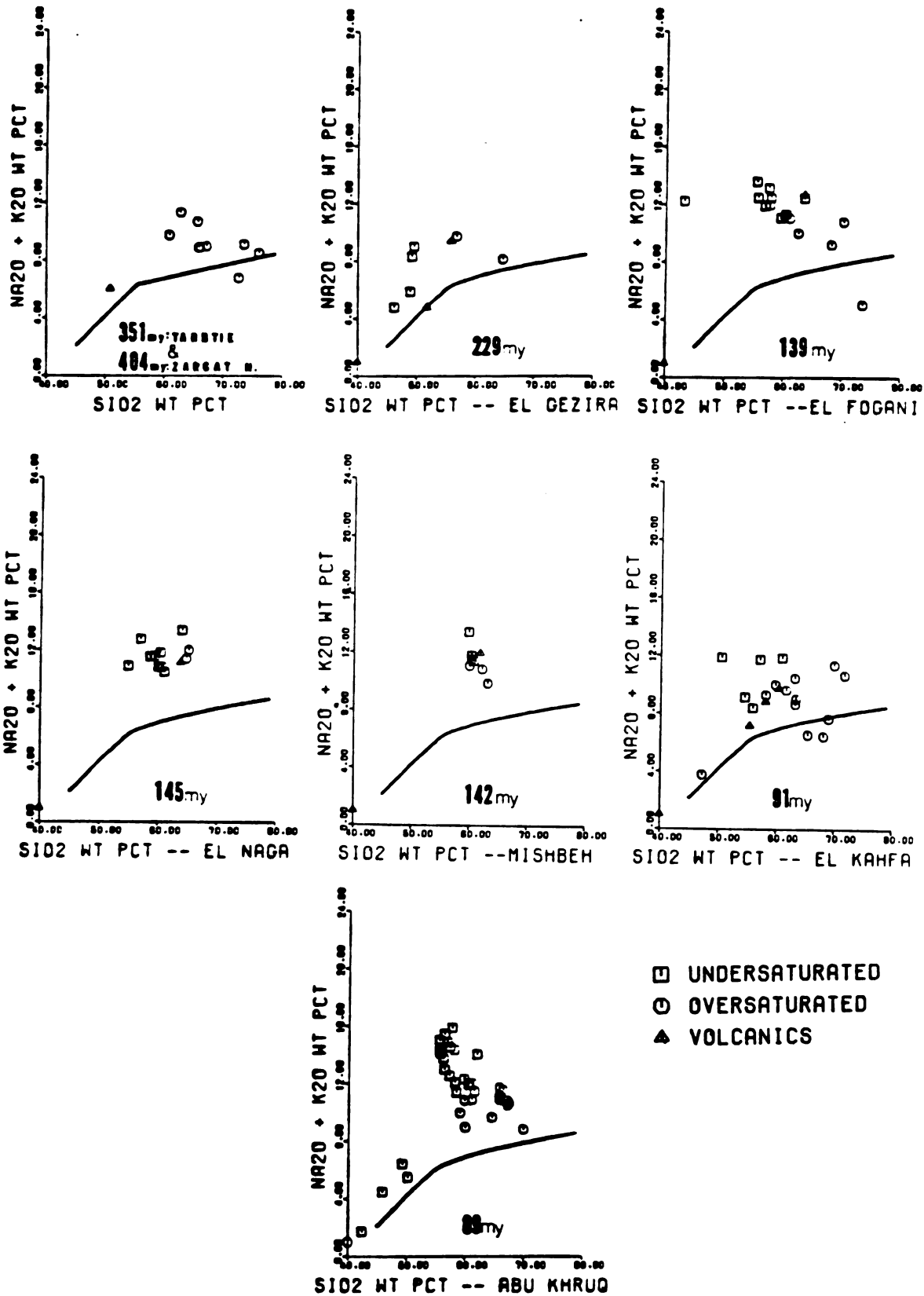


Figure 3.

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contain well-differentiated silica-oversaturated rock members. It appears that there has been a steady increase in alkalinity and an increasing predominance of silica-undersaturation through time.

Figures 4 and 5 show the general differentiation trends for the silica-oversaturated complexes and the silica-undersaturated complexes, respectively, in terms of AFM ternary diagrams. The older silica-oversaturated complexes form a more pronounced iron enrichment trend than the alkali-enriched, younger silica-undersaturated/oversaturated complexes. However, both trends are similar to those of other alkaline suites evolving by fractional crystallization (see Schwarzer and Rogers, 1974 and Coombs, 1963).

Figure 6 demonstrates the calcic-alkali trend for the Egyptian alkaline complexes in terms of the K_2O-Na_2O-CaO ternary diagram. The CaO -rich plagioclase-dominated trend leading to the high proportion of alkali-rich rock members is characteristic of the fractional crystallization trend of alkaline suites in general.

Figure 7 shows the CaO/Na_2O+K_2O ratios versus SiO_2 for many of the Egyptian alkaline complexes. This demonstrates that the general alkali-lime index (ALI) is about 51-52--within the range of typical, moderately alkaline suites (e.g. Petro, Vogel and Wilband, 1979).

A. Major Element Variation Diagrams - Gabal Abu Khruq

The changes in the concentrations of the individual

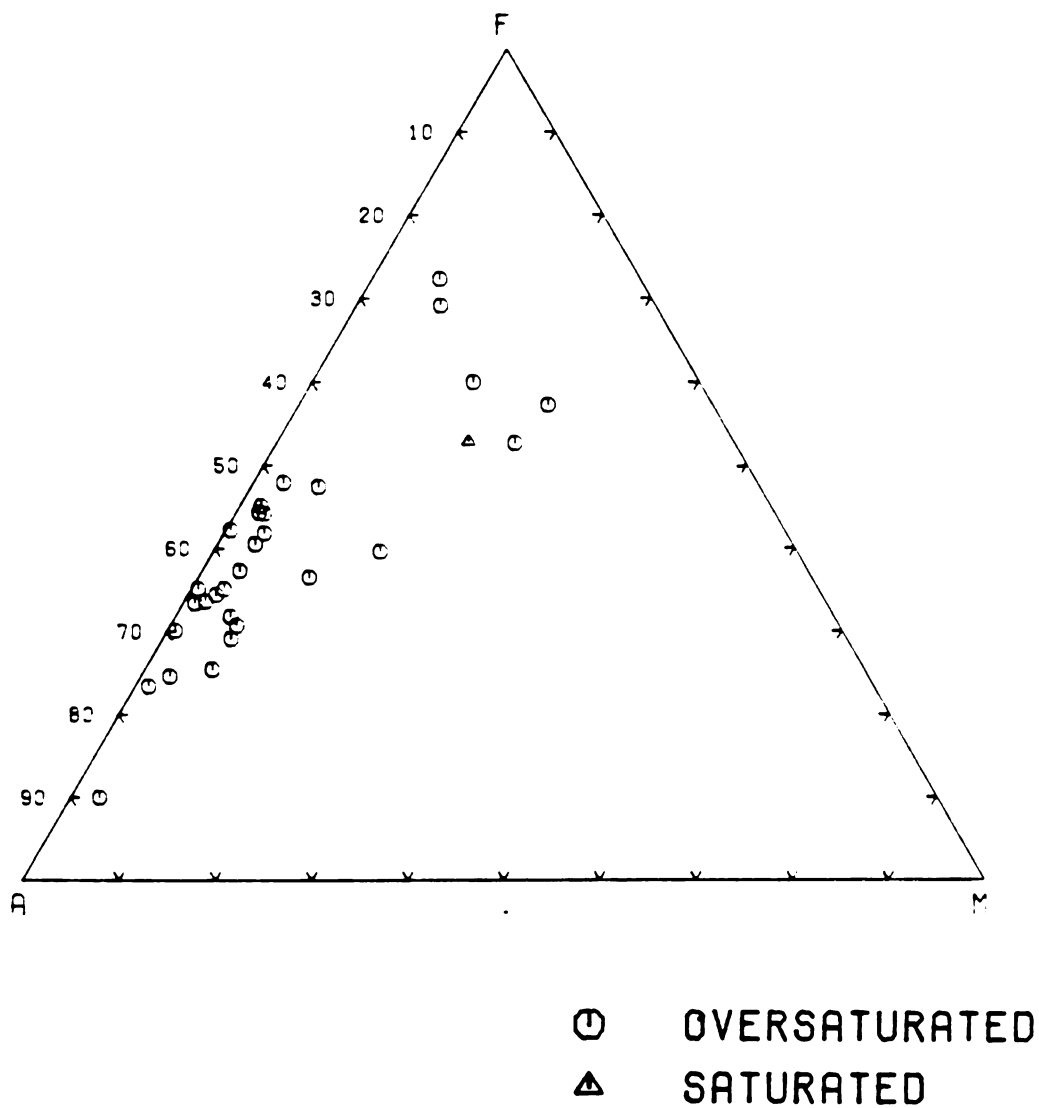


Figure 4.

AFM ternary diagram of data from Egyptian alkaline complexes containing only silica-oversaturated rock-types (data from El Ramly et al., 1971).

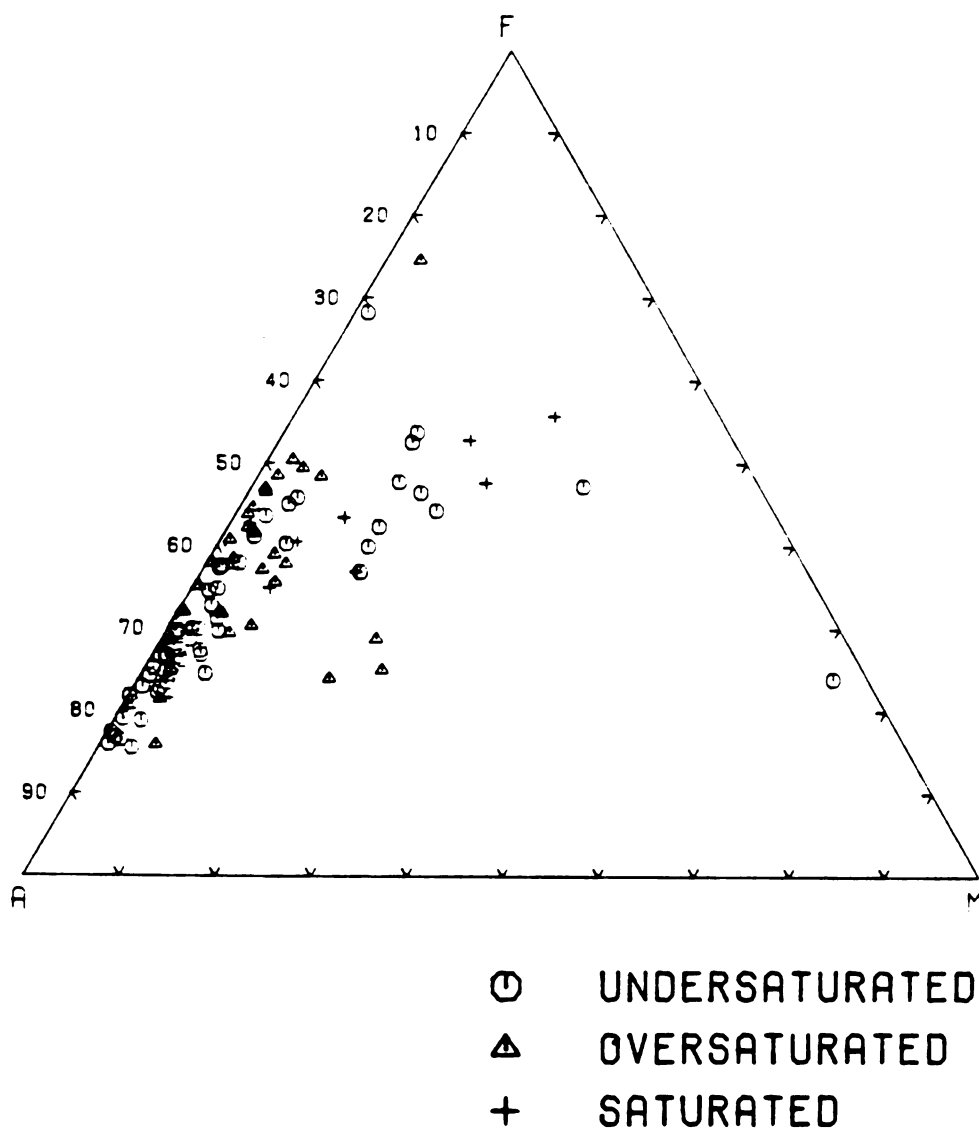


Figure 5.

AFM ternary diagram of data from Egyptian alkaline complexes containing both silica-undersaturated and oversaturated rock-types (data from this study and El Ramly et al., 1969a, 1969b, 1971).

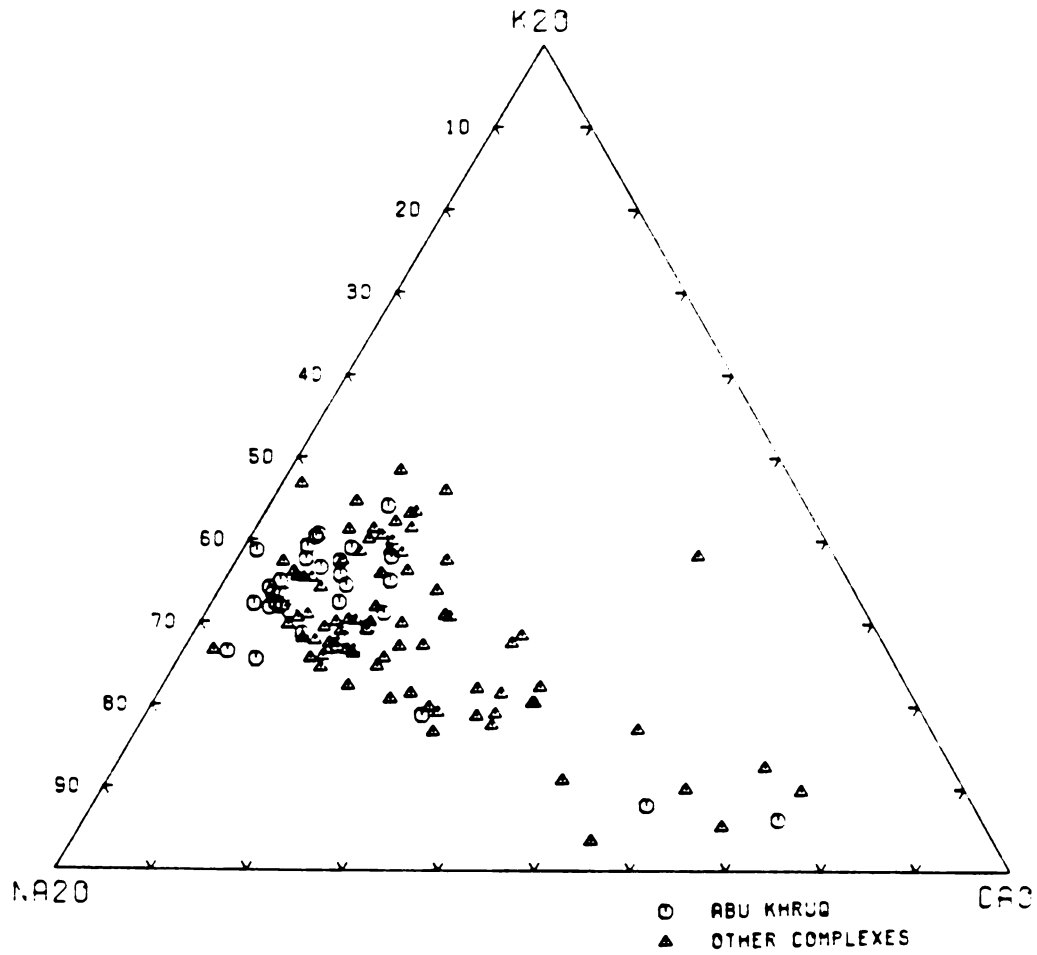


Figure 6.
 Calcic-alkali ternary diagram for the data from the
 Egyptian alkaline complexes (G. Abu Khruq is denoted
 by octagons).

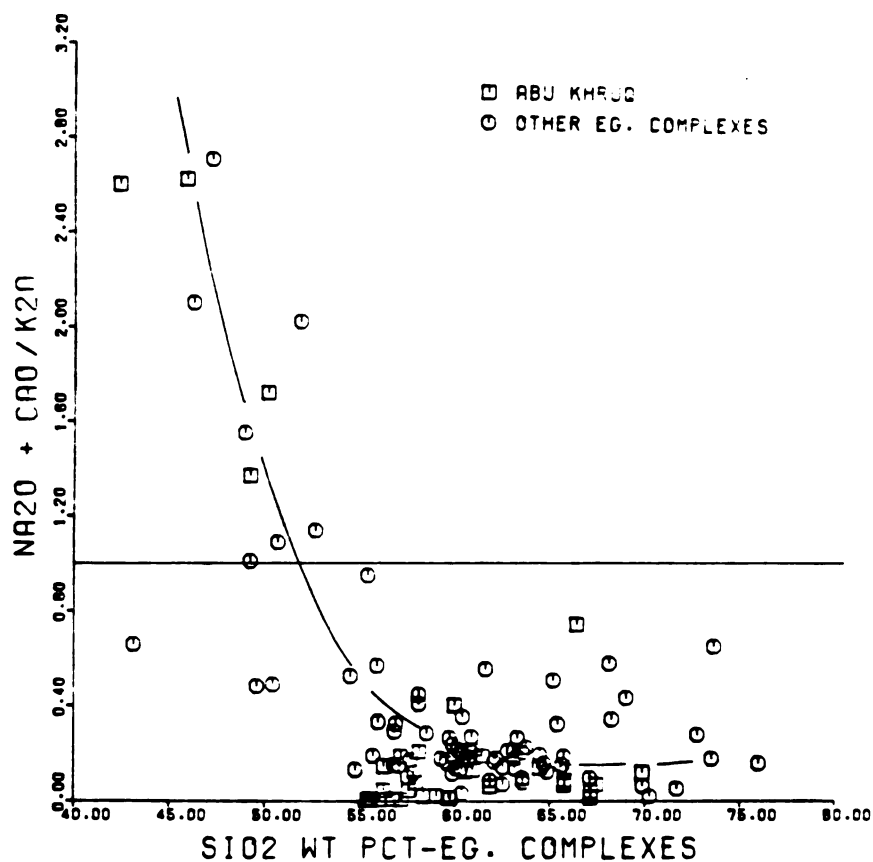


Figure 7.

Alkali-lime index diagram for the data from the Egyptian alkaline complexes. Alkali-lime index (ALI) is read as the SiO_2 value at the intersection of the data trend and the line where $\text{Na}_2\text{O} + \text{CaO}/\text{K}_2\text{O}$ is equal to one. The ALI from this diagram is about 51-52.

major elements throughout the alkaline rock series at G. Abu Khruq are shown in Figure 8. Even though these diagrams provide only very general information on the chemical trends of the major elements at Abu Khruq, they are useful in illustrating the highly-evolved nature of the majority of the exposed rock-types, as well as the relationship between the silica-undersaturated rock-types (gabbros, pulaskites, and foyaïtes), and the silica-oversaturated rock-types (alkaline syenites, quartz syenites, and volcanics).

The major element oxides which are incorporated into the more mafic rocks (TiO_2 , $\text{FeO}_{\text{total}}$, CaO , MnO , and P_2O_5) display depletion trends which are consistent with crystal fractionation. The major element oxides which become more important in the salic stages of differentiation (SiO_2 , Al_2O_3 , Na_2O , and K_2O) display more complicated trends. These trends consist of diverging patterns in which the silica-undersaturated syenites are enriched in Al_2O_3 and Na_2O relative to the silica-oversaturated rock-types which are comparatively enriched in SiO_2 and $\text{FeO}_{\text{total}}$, as well as lower $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios. The sharp inflections in these trends are considered to reflect the crystallization of the sodic-alumina phases in the salic stages of differentiation (alkali feldspar, nepheline, and sodalite). The volcanics (silica-oversaturated) form well-defined trends which plot along concentration ranges similar to the silica-oversaturated syenites. The greater scatter in the syenite trends (both silica-oversaturated and undersaturated) is

Figure 8.

Major element variation diagrams for the data from Gabal Abu Khruq (data are from this study and El Ramly et al., 1969a). MgO is used as the index of crystallinity.

Solid squares = gabbros (always found on the right-hand side of the variation diagrams).

Open squares = silica-undersaturated syenites (pulaskites and foyaïtes).

Hexagons = silica-oversaturated syenites (alkaline syenites).

Triangles = volcanics.

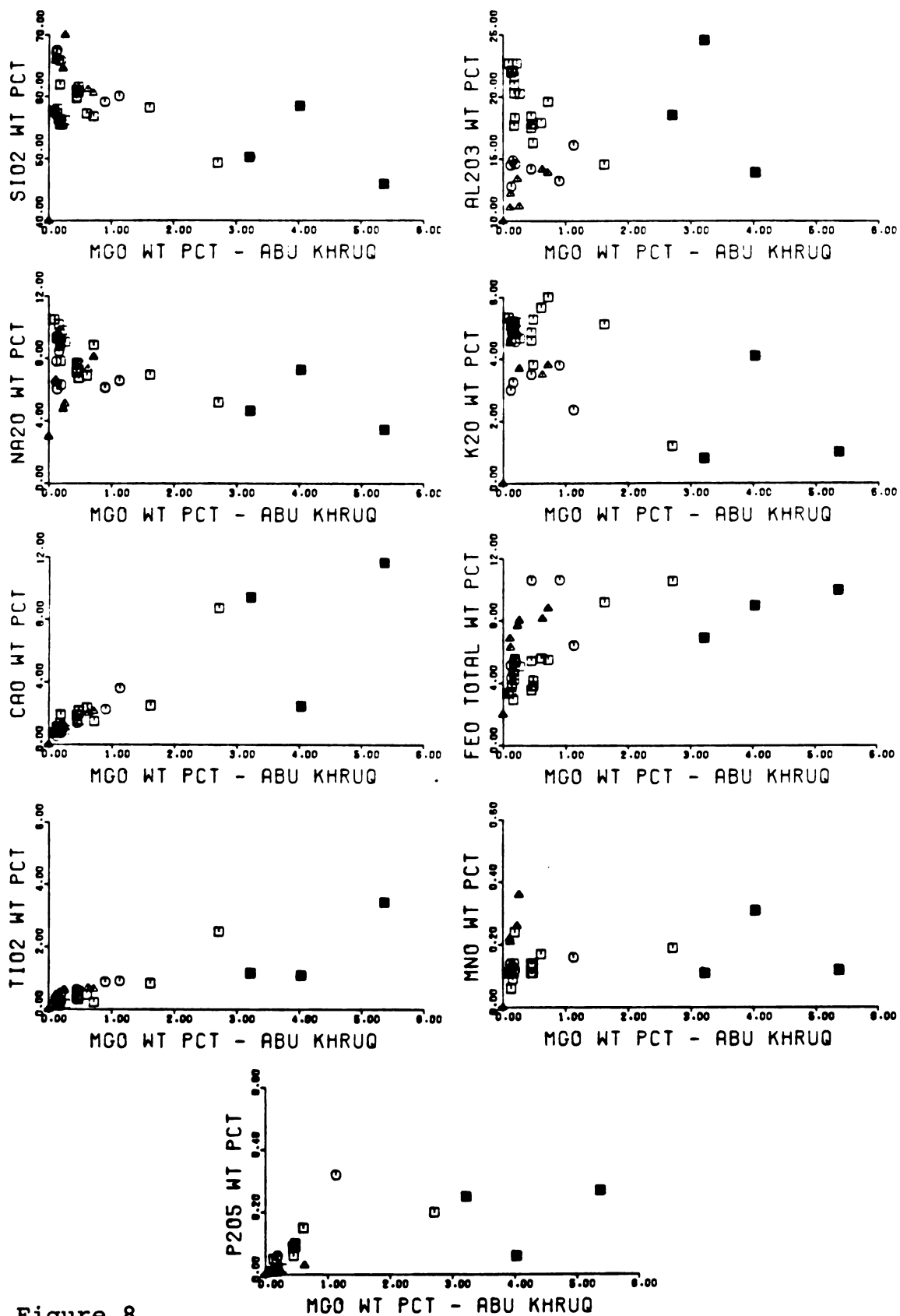


Figure 8.

attributed to the individual unique compositions often found in coarse-grained samples. The foyaïtes (strongly silica-undersaturated) formed the last stage of crystallization at Abu Khruq and plot as a dense cluster of similar compositions. The large separation between the gabbros and the more mafic (silica-undersaturated) syenites reflects the presence of large quantities of alkali feldspar in the syenites, and the cumulate, labradorite-rich nature of the gabbros (especially rich in CaO and Al_2O_3). The variable $\text{FeO}_{\text{total}}$ and TiO_2 concentrations of the more mafic syenites and gabbros are considered to reflect the larger proportions of the iron-titanium oxides in individual samples. The gabbros and mafic syenites are rich in P_2O_5 because they contain most of the apatite observed in these types of alkaline suites.

2) Trace Elements

The trace element data consists of predominantly the REE and thorium concentrations from the four silica-undersaturated/oversaturated complexes sampled by this study. There is a limited amount of compatible and other incompatible trace element data, as well as Rb-Sr data from Lutz (1979).

A. Rare Earth Elements

Table 1 is a compilation of the rare earth element and thorium data determined in this study for the four alkaline complexes investigated. Figures 9-13 display the normalized REE patterns for the rocks from G. Abu Khruq.

Table 1. Rare Earth Element and Thorium Data
for the Egyptian Alkaline Complexes:
G. Abu Khruq; G. El Kahfa;
G. El Naga; G. Nigrub El Fogani.

*First value in ppm.

*Second value is chondrite normalized for the REE.

I. Gabal Abu Khruq

	La	Ce	Sm	Eu	Tb	Yb	Lu	Th	
<u>Gabbros</u>									
A-1	22.61 (68.51)	55.77 (63.37)	7.28 (40.20)	3.79 (54.87)	1.32 (28.09)	3.40 (17.01)	0.55 (16.15)	3.71	
A-5b	42.74 (129.51)	84.31 (95.80)	9.34 (51.62)	6.35 (92.03)	1.23 (26.22)	2.43 (12.15)	0.29 (8.54)	2.54	
A-5c	38.09 (115.42)	80.21 (91.14)	9.08 (50.18)	5.98 (86.72)	0.97 (20.70)	2.87 (14.34)	0.47 (13.68)	3.15	
A-76 (altered)	7.11 (21.54)	25.85 (29.38)	1.84 (10.19)	2.10 (30.50)	0.38 (8.14)	0.49 (2.45)	0.14 (4.18)	1.40	
A-82	48.95 (148.34)	89.61 (101.83)	11.37 (62.83)	6.55 (94.87)	1.57 (33.33)	3.43 (17.14)	0.44 (12.99)	2.77	
A-83	11.02 (33.38)	34.86 (39.62)	3.46 (19.10)	1.73 (25.04)	0.66 (13.95)	1.49 (7.47)	0.19 (5.69)	0.44	
A-91a	18.82 (57.01)	47.36 (53.81)	4.11 (22.67)	2.61 (37.81)	0.54 (11.52)	1.26 (6.31)	0.22 (6.34)	1.83	
A-91b	19.93 (60.38)	32.76 (37.23)	4.73 (26.15)	2.87 (41.53)	0.66 (14.00)	1.47 (7.32)	0.17 (4.92)	1.21	
A-103	23.73 (71.90)	48.42 (55.02)	5.25 (28.99)	2.68 (38.79)	0.56 (11.89)	1.26 (6.28)	0.30 (8.85)	2.59	
<u>Silica-Undersaturated Syenites (F.=foyaite; P.=pulaskite)</u>									
A-7	75.29 (228.15)	131.41 (149.33)	8.12 (44.87)	0.65 (9.39)	1.09 (23.25)	4.96 (24.79)	0.51 (15.00)	15.50	F.
A-9	82.08 (248.72)	153.38 (174.29)	11.29 (62.40)	0.75 (10.88)	1.57 (33.31)	10.22 (51.12)	0.94 (27.74)	23.47	F.
A-10a	77.08 (233.58)	149.86 (170.29)	9.76 (53.95)	0.88 (12.83)	1.65 (35.02)	5.98 (29.88)	0.65 (19.12)	15.50	F.
A-11	86.19 (261.17)	142.41 (161.83)	13.92 (76.91)	3.69 (53.44)	1.51 (32.03)	6.10 (30.51)	0.50 (14.60)	9.21	P.
A-16	47.30 (143.34)	114.40 (130.00)	11.23 (62.04)	2.97 (42.98)	2.22 (47.14)	8.86 (44.32)	0.72 (21.04)	10.65	P.
A-19a	132.14 (400.44)	199.69 (226.92)	12.06 (66.66)	0.70 (10.15)	1.38 (29.39)	8.31 (41.53)	0.82 (24.13)	38.00	F.
A-20	58.00 (175.75)	91.43 (103.89)	7.72 (42.65)	0.67 (9.78)	0.87 (18.42)	4.58 (22.90)	0.50 (14.81)	10.41	F.
A-21	72.37 (219.31)	149.82 (170.25)	10.37 (57.30)	0.82 (11.86)	1.43 (30.50)	4.36 (21.82)	0.44 (13.08)	10.75	F.
A-30	49.24 (149.21)	88.35 (100.40)	7.31 (40.40)	0.66 (9.58)	1.18 (25.03)	2.14 (10.72)	0.40 (11.66)	8.54	F.
A-32a	125.67 (380.82)	220.87 (250.99)	13.91 (76.86)	1.06 (15.36)	1.93 (41.12)	10.64 (53.18)	0.92 (27.09)	37.29	F.
A-32c	105.51 (319.74)	172.00 (195.45)	13.42 (74.16)	1.24 (18.00)	1.79 (38.13)	6.58 (32.90)	1.00 (29.30)	32.53	F.
A-35	110.40 (334.54)	176.79 (200.90)	15.90 (87.87)	1.97 (28.56)	2.06 (43.80)	4.72 (23.61)	0.67 (19.74)	18.47	P.
A-45	23.64 (71.63)	60.62 (68.89)	4.95 (27.35)	0.51 (7.33)	0.94 (19.94)	2.06 (10.28)	0.38 (11.09)	6.32	P.
A-50	53.84 (163.17)	96.53 (109.69)	6.91 (38.15)	0.68 (9.93)	0.73 (15.57)	4.00 (19.98)	0.44 (13.07)	11.44	P.
A-51	43.60 (132.13)	71.27 (80.99)	6.14 (33.93)	0.78 (11.25)	0.73 (15.48)	3.27 (16.34)	0.28 (8.17)	5.60	P.
A-56a	113.82 (344.91)	221.69 (251.92)	15.18 (83.85)	1.03 (14.95)	1.78 (37.98)	8.24 (41.19)	0.78 (23.02)	24.15	P.

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Table 1 (cont'd).

	La	Ce	Sm	Eu	Tb	Yb	Lu	Th	
A-56b	131.67 (399.01)	243.54 (276.75)	16.36 (90.37)	1.12 (16.20)	1.96 (41.78)	8.85 (44.26)	0.86 (25.34)	23.13	P.
A-59a	85.69 (259.67)	144.02 (163.66)	9.52 (52.57)	1.72 (24.86)	1.26 (26.89)	4.35 (21.76)	0.45 (13.20)	12.11	P.
A-59b	75.73 (229.47)	133.63 (151.85)	8.33 (46.05)	1.62 (23.43)	0.87 (18.49)	5.87 (29.35)	0.55 (16.05)	19.77	P.
A-60a	44.29 (134.21)	62.33 (70.83)	4.70 (25.93)	3.29 (47.72)	0.62 (13.15)	2.11 (10.53)	0.28 (8.08)	8.70	P.
A-62	84.83 (257.05)	153.81 (174.78)	10.06 (55.60)	0.90 (13.00)	1.35 (28.72)	5.88 (29.38)	0.52 (15.39)	13.28	P.
A-65	68.32 (207.03)	124.63 (141.63)	10.15 (56.05)	0.99 (14.37)	1.83 (38.89)	4.08 (20.38)	0.64 (18.70)	14.73	P.
A-66a	80.31 (243.37)	141.33 (160.60)	12.72 (70.26)	1.47 (21.32)	1.66 (35.42)	3.84 (19.22)	0.52 (15.38)	8.49	P.
A-86	107.80 (324.50)	187.83 (213.33)	13.04 (72.05)	1.36 (19.65)	1.80 (38.37)	7.37 (36.83)	0.51 (15.09)	21.07	F.
A-87	90.39 (275.23)	175.34 (199.24)	12.59 (69.56)	0.87 (12.59)	1.64 (34.86)	6.87 (29.57)	0.69 (19.99)	16.58	F.
A-88	92.95 (281.66)	129.25 (146.87)	12.55 (69.34)	0.97 (13.99)	1.60 (33.95)	6.52 (32.62)	0.73 (21.34)	17.55	F.
Eg-22	91.48 (277.22)	138.42 (157.29)	11.26 (62.23)	1.59 (23.04)	1.65 (35.10)	4.64 (23.21)	0.59 (17.48)	14.81	F.
<u>Silica-Oversaturated Syenites (A.S.=alkaline syenite; Q.S.=quartz syenite)</u>									
A-66b	79.11 (239.74)	141.10 (160.34)	14.43 (79.75)	2.98 (43.26)	2.67 (56.85)	6.73 (33.66)	1.01 (29.61)	11.48	A.S.
A-69	47.77 (144.74)	83.17 (94.51)	10.72 (59.20)	3.97 (57.54)	1.46 (31.16)	3.51 (17.56)	0.73 (21.46)	7.53	A.S.
A-70	88.29 (267.56)	170.61 (193.88)	14.57 (80.52)	1.96 (28.45)	1.92 (40.91)	7.47 (37.34)	1.01 (29.59)	14.79	Q.S.
A-72	78.53 (237.98)	158.61 (180.24)	14.03 (77.51)	1.83 (26.57)	2.31 (49.06)	9.64 (48.18)	0.98 (28.93)	16.04	Q.S.
A-75	71.33 (216.17)	142.98 (162.48)	13.30 (73.49)	2.00 (28.98)	1.94 (41.36)	8.64 (43.18)	0.86 (25.17)	14.46	Q.S.
A-77	60.19 (182.40)	120.07 (136.45)	9.50 (52.51)	1.86 (26.98)	1.49 (31.60)	5.98 (29.88)	0.73 (21.53)	11.73	Q.S.
A-79a	46.03 (139.48)	91.33 (103.78)	9.70 (53.60)	3.77 (54.62)	1.03 (21.96)	4.86 (24.30)	0.59 (17.29)	5.46	A.S.
A-79b	154.17 (467.19)	275.78 (313.39)	24.94 (137.81)	4.52 (65.51)	3.50 (74.57)	17.47 (87.35)	1.86 (54.64)	27.72	Q.S.
A-81	129.33 (391.92)	233.43 (265.26)	20.20 (111.59)	3.70 (53.66)	3.02 (64.21)	11.50 (57.49)	1.21 (35.67)	8.72	Q.S.
A-85	82.20 (249.09)	158.19 (179.76)	17.06 (94.28)	4.70 (68.13)	4.26 (90.55)	11.70 (58.48)	1.14 (33.52)	15.74	Q.S.
Eg-5	98.35 (298.03)	153.34 (174.25)	12.62 (69.74)	1.33 (19.22)	1.40 (29.78)	2.37 (11.83)	0.79 (23.14)	12.31	A.S.
Eg-23	77.17 (233.85)	147.23 (167.31)	14.64 (80.88)	3.62 (52.42)	1.55 (33.04)	2.50 (12.49)	0.56 (16.36)	7.77	Q.S.
Eg-25	86.97 (263.54)	137.96 (156.77)	10.90 (60.23)	1.49 (21.56)	1.10 (23.41)	2.50 (12.48)	0.71 (21.02)	8.75	A.S.
Eg-27	53.94 (163.46)	124.34 (141.30)	11.45 (63.28)	1.69 (24.43)	2.76 (58.70)	9.81 (49.03)	0.76 (22.48)	15.30	A.S.
Eg-32	31.13 (94.35)	56.03 (63.67)	8.34 (46.06)	4.97 (72.10)	0.84 (17.80)	2.03 (10.16)	0.39 (11.52)	4.70	A.S.
A-44	62.90 (109.60)	177.95 (202.21)	19.20 (106.07)	4.43 (64.14)	5.16 (109.81)	13.36 (66.78)	1.04 (30.46)	16.86	Q.S.
A-92a	51.45 (155.90)	89.21 (101.38)	10.17 (56.18)	2.91 (42.18)	1.47 (31.27)	4.41 (22.05)	0.73 (21.50)	7.89	A.S.
A-93	32.35 (98.04)	122.45 (139.15)	6.99 (38.64)	0.68 (9.93)	1.61 (34.18)	6.74 (33.72)	0.81 (23.76)	11.44	Q.S. ...very altered
A-96a	147.12 (445.83)	260.77 (296.33)	26.74 (147.76)	5.13 (74.32)	4.07 (86.67)	16.53 (82.63)	1.58 (46.50)	20.17	Q.S.
A-96c	217.15 (658.02)	343.14 (389.93)	33.72 (186.30)	5.86 (84.93)	3.76 (80.03)	17.13 (85.66)	1.57 (46.12)	44.69	Q.S.

Table 1 (cont'd).

	La	Ce	Sm	Eu	Tb	Yb	Lu	Th	
A-98	106.88 (323.88)	182.06 (206.88)	18.32 (101.23)	3.83 (55.47)	2.78 (59.12)	10.64 (53.21)	1.01 (29.69)	17.59	Q.S.
A-102	99.10 (300.30)	167.98 (190.89)	18.95 (104.72)	4.96 (71.85)	2.37 (50.46)	9.70 (48.52)	1.23 (36.10)	9.92	Q.S.
<u>Volcanics</u> (Rhyolites/Trachytes)									
A-37	121.91 (369.43)	224.14 (254.71)	18.75 (103.57)	3.71 (53.71)	2.40 (51.02)	11.15 (55.76)	1.07 (31.43)	17.21	
A-41a	92.05 (278.93)	179.73 (204.24)	17.07 (94.29)	3.98 (57.64)	2.08 (44.24)	8.85 (44.25)	0.84 (24.70)	12.81	
A-42	111.14 (336.78)	207.94 (236.30)	18.67 (103.15)	3.76 (54.51)	2.28 (48.49)	9.53 (47.64)	1.02 (30.01)	14.44	
A-84	147.55 (447.13)	256.11 (291.03)	25.37 (140.17)	5.38 (77.91)	3.63 (77.20)	14.64 (73.20)	1.48 (43.45)	18.14	
A-90	135.50 (410.60)	197.71 (224.67)	21.24 (117.34)	3.78 (54.75)	2.51 (53.46)	11.20 (56.02)	1.08 (31.77)	19.34	
A-95	70.65 (214.08)	124.59 (141.58)	12.44 (68.73)	4.58 (66.44)	1.29 (27.45)	5.14 (25.72)	0.46 (13.57)	6.88...very altered	
A-97a	134.45 (407.43)	254.71 (289.44)	20.98 (115.92)	2.66 (38.49)	2.90 (61.61)	13.79 (68.93)	1.26 (37.20)	25.29	
A-99	100.15 (303.47)	178.60 (202.96)	16.75 (92.54)	3.63 (52.58)	2.18 (46.35)	8.64 (43.18)	0.88 (25.95)	13.55	
Eg-14	98.53 (298.53)	184.85 (210.05)	17.20 (95.04)	3.17 (45.96)	2.02 (43.04)	2.98 (14.88)	1.04 (30.53)	12.60	
Eg-16	179.27 (543.25)	271.78 (308.84)	28.09 (155.20)	6.04 (87.54)	3.68 (78.38)	3.59 (17.93)	1.28 (37.62)	19.10	
<u>Dike Rocks</u>									
<u>Silica-Undersaturated:</u>									
A-6	36.26 (109.87)	63.04 (71.69)	5.26 (29.06)	0.77 (11.11)	0.69 (14.74)	2.95 (14.77)	0.32 (9.19)	5.66...pegmatite	
A-19b	73.67 (223.24)	119.47 (135.76)	7.86 (43.41)	0.83 (11.97)	0.73 (15.50)	3.69 (18.47)	0.45 (13.38)	9.27	
A-24	82.50 (250.00)	135.59 (154.08)	9.63 (53.19)	1.06 (15.42)	0.92 (19.52)	4.46 (22.28)	0.53 (15.57)	11.67	
A-28	89.10 (269.99)	168.23 (191.17)	10.59 (58.49)	1.28 (18.52)	1.47 (31.29)	6.45 (32.24)	0.62 (18.19)	20.86	
A-49	105.77 (320.52)	202.74 (230.39)	12.15 (67.11)	0.82 (11.87)	1.18 (25.04)	6.35 (31.77)	0.67 (19.85)	18.86	
A-55a	17.91 (54.26)	21.26 (24.16)	2.04 (11.28)	0.45 (6.45)	0.38 (8.04)	1.39 (6.94)	0.30 (8.76)	7.23...pegmatite	
A-63	67.55 (204.71)	112.79 (128.17)	7.55 (41.74)	0.53 (7.66)	0.93 (19.81)	3.29 (16.44)	0.57 (16.83)	13.47	
A-68	49.52 (150.06)	91.47 (103.94)	8.48 (46.85)	2.33 (33.75)	1.27 (26.94)	4.87 (24.35)	0.49 (14.49)	5.49	
<u>Silica-Oversaturated:</u>									
A-2	94.22 (285.52)	176.13 (200.15)	15.71 (86.79)	2.40 (34.75)	2.22 (47.13)	8.57 (42.83)	0.83 (24.50)	11.54	
A-83b	93.00 (281.82)	155.93 (177.19)	14.12 (77.98)	1.18 (17.13)	2.32 (49.43)	5.73 (28.65)	0.94 (27.56)	38.54	
A-91c	78.20 (236.98)	128.50 (146.02)	13.93 (76.96)	2.59 (37.55)	2.59 (55.03)	5.50 (27.49)	0.87 (25.55)	6.98	
A-92b	100.65 (305.01)	178.88 (203.27)	18.80 (103.87)	1.63 (23.57)	3.13 (66.51)	7.84 (39.16)	1.11 (32.55)	18.62	

Table 1 (cont'd).

	La	Ce	Sm	Eu	Tb	Yb	Lu	Th
<u>Xenoliths</u>								
A-10b	91.48 (277.22)	166.79 (189.53)	11.63 (64.25)	0.66 (9.62)	1.20 (25.48)	4.91 (24.53)	0.51 (14.94)	17.03
A-33a	24.64 (74.65)	36.79 (41.81)	3.32 (18.31)	0.16 (2.35)	0.55 (11.65)	4.45 (22.25)	0.60 (17.62)	5.46
A-33b	45.69 (138.47)	81.32 (92.41)	5.95 (32.86)	0.20 (2.94)	0.85 (18.12)	10.28 (51.38)	0.97 (28.59)	14.20
A-34	18.15 (55.00)	43.22 (49.12)	3.24 (17.91)	0.18 (2.60)	0.72 (15.27)	8.67 (43.37)	0.77 (22.72)	5.28
A-48	78.99 (239.35)	153.40 (174.32)	10.71 (59.16)	2.09 (30.26)	1.18 (25.09)	4.50 (22.52)	0.50 (14.74)	7.43
<u>Wehrlite--Olivine-rich Ultramafic Cumulate</u>								
A-94	10.93 (33.12)	35.58 (40.43)	2.10 (11.63)	0.54 (7.89)	0.46 (9.79)	0.61 (3.07)	0.19 (5.67)	1.48 (serpentized)
<u>Country Rock</u>								
A-57	8.54 (25.88)	26.23 (29.81)	3.16 (17.46)	0.65 (9.40)	0.37 (7.97)	1.34 (6.68)	0.17 (5.01)	2.76 (granite gneiss)
A-64	9.18 (27.83)	15.06 (17.12)	1.06 (5.83)	0.69 (9.93)	0.41 (8.74)	1.18 (5.90)	0.12 (3.49)	0.00 (altered amph. gneiss)
A-67	2.81 (8.50)	10.68 (12.14)	1.13 (6.26)	0.82 (11.84)	0.13 (2.78)	1.07 (5.37)	0.31 (9.24)	0.00 (amph. gneiss)
A-73	82.19 (249.07)	137.17 (155.87)	12.52 (69.19)	1.90 (27.56)	1.98 (42.21)	9.24 (46.21)	0.93 (27.23)	13.80 (hornfels xenolith)
II. <u>Gabal El Kahfa</u>								
<u>Gabbros</u>								
K-1	40.40 (122.43)	66.88 (75.99)	8.20 (45.30)	2.87 (41.62)	1.04 (22.15)	0.06 (0.30)	0.01 (0.22)	2.90
K-9	60.66 (183.83)	106.39 (120.90)	10.24 (56.58)	4.95 (71.73)	1.24 (26.38)	3.11 (15.53)	0.54 (16.01)	6.77
K-10c	57.81 (175.17)	99.55 (113.13)	11.25 (62.15)	4.45 (64.50)	1.05 (22.30)	3.66 (18.28)	0.76 (22.47)	7.21
K-11a (altered)	6.60 (20.01)	22.38 (25.43)	2.93 (16.17)	1.05 (15.16)	1.13 (23.95)	2.57 (12.86)	0.51 (14.98)	0.64
K-11b (altered)	8.38 (25.38)	29.11 (33.08)	4.81 (26.60)	1.43 (20.67)	0.83 (17.60)	1.78 (8.91)	0.67 (19.72)	0.90
K-37a	53.26 (161.39)	79.65 (90.51)	9.02 (49.86)	3.88 (56.18)	1.47 (31.26)	3.28 (16.39)	0.30 (8.88)	6.39
K-37b	51.59 (156.34)	81.35 (92.45)	8.05 (44.48)	2.28 (33.05)	1.74 (37.06)	1.94 (9.72)	0.28 (8.09)	9.32
K-39	40.41 (122.44)	71.91 (81.71)	7.26 (40.10)	2.83 (41.08)	1.47 (31.26)	2.32 (11.61)	0.37 (10.91)	6.01
<u>Syenites</u>								
K-3	57.62 (174.61)	93.30 (106.02)	8.07 (44.61)	2.44 (35.37)	0.86 (18.20)	1.79 (8.96)	0.37 (10.78)	9.00
K-4	54.91 (166.41)	101.17 (114.97)	9.67 (53.41)	3.18 (46.08)	1.10 (23.43)	2.27 (11.35)	0.32 (9.42)	7.77
K-5a	106.96 (324.13)	181.60 (206.37)	16.26 (89.85)	2.97 (40.39)	1.84 (39.14)	2.56 (12.80)	0.69 (20.23)	10.31
K-5b	56.19 (170.28)	94.37 (107.24)	9.60 (53.04)	3.53 (51.22)	1.05 (22.34)	1.86 (9.32)	0.49 (14.54)	7.88
K-6	51.46 (155.94)	81.18 (92.25)	9.21 (50.90)	3.93 (56.98)	0.87 (18.56)	2.46 (12.28)	0.32 (9.50)	6.91
K-7	68.68 (208.13)	116.17 (132.01)	10.50 (58.04)	2.54 (36.88)	1.44 (30.73)	2.38 (11.91)	0.81 (23.68)	8.49

Table 1 (cont'd).

	La	Ce	Sm	Eu	Tb	Yb	Lu	Th
K-13	87.26 (264.41)	132.40 (150.46)	10.09 (55.74)	2.35 (34.03)	1.60 (34.03)	1.95 (9.74)	0.41 (11.99)	13.41
K-14	71.25 (215.92)	107.44 (122.09)	8.36 (46.17)	3.03 (43.97)	0.94 (19.93)	3.06 (15.29)	0.32 (9.31)	9.58
K-15	80.00 (242.42)	124.22 (141.16)	11.35 (62.69)	3.34 (48.47)	0.99 (21.11)	3.41 (17.07)	0.43 (12.60)	8.72
K-16b	73.40 (222.41)	117.46 (133.48)	10.68 (59.01)	3.29 (47.63)	0.92 (19.60)	3.20 (16.01)	0.25 (7.34)	8.13
K-17	51.54 (156.19)	82.04 (93.23)	9.38 (51.82)	3.29 (47.75)	0.97 (20.62)	4.08 (20.40)	0.45 (13.23)	5.71
K-20	61.96 (187.76)	97.14 (110.38)	6.67 (36.88)	1.72 (24.97)	0.83 (17.77)	1.52 (7.60)	0.20 (5.82)	8.12
K-23	79.58 (241.14)	126.72 (144.00)	11.79 (65.16)	2.48 (35.96)	1.50 (31.89)	1.69 (8.47)	0.64 (18.69)	11.43
K-24	60.42 (183.10)	89.56 (101.77)	8.01 (44.27)	0.71 (10.36)	0.88 (18.75)	1.44 (7.20)	0.56 (16.40)	6.38
K-25	80.05 (242.56)	128.39 (145.90)	13.32 (73.60)	3.09 (44.81)	1.80 (38.21)	2.28 (11.42)	0.74 (21.91)	9.19
K-26	33.65 (101.98)	52.96 (60.18)	6.45 (35.63)	1.11 (16.08)	0.50 (10.61)	1.96 (9.82)	0.27 (7.86)	1.59
K-27	117.91 (357.30)	191.66 (217.80)	18.77 (103.69)	2.93 (42.44)	2.51 (53.34)	2.68 (13.41)	0.98 (28.69)	9.24
K-28	101.47 (307.49)	147.62 (167.75)	13.84 (76.49)	3.71 (53.81)	2.47 (52.61)	5.77 (28.85)	0.74 (21.64)	9.66
K-29	30.56 (92.59)	54.80 (62.28)	5.76 (31.85)	1.64 (23.79)	0.95 (20.30)	2.13 (10.63)	0.37 (10.88)	12.86
K-32a	27.09 (82.10)	41.08 (46.69)	3.61 (19.94)	0.54 (7.79)	0.50 (10.69)	1.71 (8.57)	0.35 (10.19)	0.86
K-32b	60.48 (183.28)	109.86 (124.84)	10.93 (60.36)	2.09 (30.34)	2.21 (47.11)	4.55 (22.74)	0.71 (20.92)	8.35
K-32c	33.61 (101.84)	66.36 (75.41)	6.20 (34.27)	4.98 (72.20)	1.39 (29.54)	1.39 (4.23)	0.85 (9.20)	4.42
K-33	23.17 (70.20)	35.38 (40.20)	2.92 (16.14)	0.34 (4.93)	0.31 (6.50)	1.53 (7.64)	0.35 (10.17)	1.00
K-34	19.08 (57.81)	33.60 (38.18)	3.36 (18.55)	0.35 (5.08)	0.56 (11.82)	1.01 (5.07)	0.26 (7.65)	1.20
K-35	76.49 (231.79)	129.10 (146.70)	14.42 (79.65)	4.67 (67.63)	2.65 (56.30)	5.48 (27.39)	0.72 (21.10)	18.06
K-36	76.47 (231.74)	131.07 (148.94)	11.88 (65.66)	4.44 (64.38)	1.97 (41.84)	4.40 (21.99)	0.67 (19.80)	12.33
K-38	77.25 (234.10)	132.05 (150.06)	12.26 (67.72)	3.91 (56.72)	2.75 (58.44)	4.97 (24.84)	0.43 (12.60)	15.52
K-41	83.04 (251.64)	124.97 (142.01)	10.43 (57.61)	4.61 (66.81)	1.75 (37.32)	3.11 (15.55)	0.36 (10.63)	21.59
K-42	79.74 (241.65)	132.39 (150.44)	13.17 (72.74)	3.41 (49.43)	1.10 (23.51)	4.51 (22.53)	0.51 (14.97)	7.36
K-43c	98.18 (297.51)	143.05 (162.56)	15.26 (84.30)	3.94 (57.12)	2.67 (56.82)	4.21 (21.04)	0.64 (18.71)	14.25
K-44	26.12 (79.16)	37.88 (43.05)	4.24 (23.42)	0.78 (11.34)	1.00 (21.28)	2.18 (10.91)	0.21 (6.28)	4.71
K-52	15.49 (46.92)	35.79 (40.67)	5.15 (28.45)	1.16 (16.80)	1.13 (24.05)	2.91 (14.55)	0.31 (9.08)	6.10
K-56a	79.81 (241.86)	112.71 (128.08)	9.93 (54.86)	3.57 (51.68)	1.87 (39.80)	1.57 (7.85)	0.34 (9.95)	10.91
K-60	30.13 (91.30)	55.39 (62.94)	6.67 (36.86)	1.32 (19.14)	1.21 (25.70)	4.00 (19.98)	0.45 (13.14)	4.81

Table 1 (cont'd).

	La	Ce	Sm	Eu	Tb	Yb	Lu	Th
<u>Volcanics</u>								
K-18a	58.15 (176.21)	95.31 (108.31)	9.83 (54.33)	3.23 (46.80)	1.21 (25.71)	4.20 (21.02)	0.48 (14.08)	5.79
K-19	82.27 (249.32)	129.18 (146.79)	10.76 (59.46)	3.36 (48.65)	1.26 (26.91)	3.18 (15.91)	0.30 (8.71)	8.63
K-59b	73.16 (221.71)	104.67 (118.95)	12.35 (68.24)	4.03 (58.44)	1.82 (38.70)	4.08 (20.39)	0.37 (10.92)	8.36
K-62	3.52 (10.68)	16.36 (18.59)	2.11 (11.68)	0.91 (13.15)	0.43 (9.08)	0.72 (15.29)	0.38 (11.10)	0.00
<u>Dikes</u>								
K-8b	59.76 (181.09)	75.30 (85.57)	6.57 (36.28)	1.03 (14.92)	0.55 (11.63)	2.54 (12.69)	0.74 (21.68)	1.15
K-22	104.36 (316.23)	150.46 (170.98)	12.28 (67.87)	2.50 (36.22)	1.47 (31.21)	2.14 (10.72)	0.56 (16.36)	13.05
K-40	125.37 (379.91)	182.23 (207.07)	9.77 (53.96)	0.54 (7.82)	2.88 (61.23)	6.52 (32.58)	0.75 (22.05)	40.80
K-45	72.83 (220.71)	115.43 (131.17)	9.89 (54.65)	1.67 (24.23)	2.44 (52.01)	8.20 (40.98)	0.94 (27.70)	19.29
K-47a	66.57 (201.72)	102.42 (116.38)	11.46 (63.33)	3.74 (54.18)	1.88 (39.97)	4.80 (23.99)	0.67 (19.58)	6.02
K-47b	119.88 (363.26)	187.67 (213.26)	16.19 (89.47)	4.24 (61.41)	2.28 (48.51)	9.10 (45.51)	0.85 (25.01)	16.22
K-50	114.56 (347.15)	168.66 (191.66)	17.27 (95.42)	2.83 (41.08)	3.86 (82.10)	7.25 (36.25)	1.07 (31.42)	17.61
K-51a	86.66 (262.60)	138.44 (157.32)	15.16 (83.75)	1.98 (28.68)	2.62 (55.70)	5.32 (26.58)	0.60 (17.68)	5.03
K-51b	7.51 (22.76)	14.06 (15.97)	3.15 (17.38)	1.31 (18.99)	0.14 (3.08)	3.27 (16.33)	0.33 (9.84)	0.33
K-55a	167.57 (507.80)	220.34 (250.39)	14.06 (77.70)	2.35 (34.12)	2.83 (60.16)	4.47 (22.37)	0.48 (14.22)	59.61
K-55b	93.42 (283.08)	123.16 (139.95)	9.19 (50.80)	2.32 (33.59)	1.80 (38.39)	3.44 (17.19)	0.41 (11.96)	17.52
K-61	113.29 (343.29)	183.41 (208.42)	14.82 (81.90)	1.64 (23.82)	3.66 (77.86)	7.76 (38.80)	0.75 (22.19)	12.53
K-63b	25.49 (77.25)	42.85 (48.69)	5.29 (29.24)	1.23 (17.89)	1.00 (21.30)	2.28 (11.41)	0.43 (12.61)	3.47
<u>Country Rock</u>								
K-12b	3.88 (11.77)	18.38 (20.88)	2.17 (11.99)	0.91 (13.20)	0.08 (1.72)	2.21 (11.07)	0.31 (9.18)	0.85
K-46	252.75 (765.91)	406.41 (461.82)	31.34 (173.13)	4.56 (66.13)	2.47 (52.56)	6.09 (30.44)	0.89 (26.31)	9.78 (femitized)
K-48a	9.14 (27.69)	21.19 (24.08)	2.62 (14.50)	0.69 (9.96)	0.52 (10.98)	0.47 (2.33)	0.22 (6.49)	2.28
K-59a	12.95 (39.25)	29.11 (33.08)	5.66 (31.25)	1.41 (20.42)	1.03 (21.90)	1.76 (8.82)	0.38 (11.09)	0.46
K-63a	12.44 (37.70)	21.10 (23.98)	3.60 (19.90)	1.28 (18.57)	0.78 (16.64)	0.54 (2.64)	0.19 (5.50)	0.29

Table 1 (cont'd).

III. Gabal El Naga

	La	Ce	Sm	Eu	Tb	Yb	Lu	Th
<u>Syenites</u>								
N-5b	76.78 (232.66)	125.93 (143.10)	10.95 (60.50)	4.79 (69.39)	1.34 (28.52)	2.77 (13.85)	0.49 (14.35)	10.08
N-5c	74.73 (226.46)	124.51 (141.49)	10.63 (58.75)	4.44 (64.29)	1.21 (25.72)	2.71 (13.53)	0.47 (13.95)	9.76
N-7a	66.18 (200.54)	120.99 (137.49)	8.13 (44.94)	3.85 (55.83)	0.94 (20.00)	4.00 (20.02)	0.64 (18.97)	11.88
N-10	54.77 (165.96)	93.88 (106.69)	5.60 (30.93)	5.88 (85.18)	0.72 (15.37)	2.25 (11.24)	0.37 (11.03)	5.10
N-14	71.64 (217.08)	128.63 (146.17)	10.49 (57.94)	3.35 (48.54)	1.10 (23.50)	3.77 (18.86)	0.54 (15.91)	10.16
N-17	98.99 (299.97)	173.26 (196.88)	15.18 (83.89)	4.50 (65.26)	2.18 (46.38)	5.06 (25.31)	0.81 (23.82)	15.74
N-22	57.88 (175.39)	121.35 (137.90)	9.11 (50.36)	4.21 (61.02)	0.78 (16.57)	2.73 (13.66)	0.57 (16.70)	5.47
N-23	42.06 (127.47)	89.99 (102.26)	5.31 (29.36)	4.84 (70.11)	0.52 (10.99)	3.00 (15.01)	0.48 (14.05)	5.77
N-25	114.92 (348.24)	200.43 (227.76)	12.49 (69.02)	3.59 (52.06)	1.43 (30.48)	6.23 (31.15)	1.01 (29.57)	17.19
N-32	79.58 (241.15)	134.94 (153.34)	11.81 (65.24)	5.35 (77.49)	1.39 (29.59)	2.36 (11.81)	0.53 (15.46)	4.41
N-33	56.71 (171.86)	111.59 (126.81)	5.76 (31.84)	6.68 (96.76)	0.83 (17.76)	3.70 (18.49)	0.55 (16.07)	7.59
N-34	51.24 (155.26)	92.58 (105.20)	5.41 (29.87)	6.83 (98.99)	0.61 (13.04)	3.14 (15.72)	0.43 (12.60)	6.62
N-36	68.03 (206.14)	139.98 (159.07)	6.96 (38.46)	3.80 (55.11)	0.80 (17.09)	2.84 (14.20)	0.56 (16.34)	12.82
N-40	48.55 (147.13)	98.91 (112.39)	5.71 (31.52)	5.39 (78.10)	0.79 (16.83)	2.06 (10.28)	0.58 (17.17)	7.34
N-47	66.07 (200.22)	118.01 (134.10)	8.45 (46.70)	6.47 (93.77)	1.19 (25.22)	2.40 (12.02)	0.50 (14.82)	10.87
N-48a	44.23 (134.03)	97.02 (110.25)	6.78 (37.46)	1.33 (19.30)	0.54 (11.59)	2.51 (12.57)	0.53 (15.56)	3.80
N-51	65.54 (198.60)	142.92 (162.41)	9.19 (50.80)	4.17 (60.43)	0.83 (17.61)	3.41 (17.06)	0.60 (17.68)	7.03
N-52	69.33 (210.09)	122.47 (139.17)	7.50 (41.46)	4.48 (64.88)	0.74 (15.76)	1.89 (9.46)	0.56 (16.38)	5.59

Volcanics

N-1a	87.25 (264.39)	134.69 (153.06)	11.70 (64.66)	4.44 (64.39)	1.57 (33.47)	3.10 (15.51)	0.65 (19.20)	11.49
N-4	111.03 (336.45)	234.90 (266.93)	11.09 (61.29)	1.51 (21.83)	1.16 (24.78)	6.12 (30.61)	1.11 (32.78)	19.18
N-27	80.90 (245.14)	143.88 (163.50)	12.06 (66.64)	3.82 (55.41)	1.54 (32.77)	3.96 (19.81)	0.56 (16.58)	10.82
N-28	80.39 (243.62)	131.79 (149.77)	11.99 (61.80)	4.51 (65.36)	1.39 (29.55)	2.89 (14.43)	0.47 (13.75)	11.00
N-30	93.63 (283.74)	153.38 (174.29)	11.47 (63.39)	3.16 (45.80)	1.47 (31.25)	4.35 (21.73)	0.70 (20.62)	12.77

Dikes

N-20	89.45 (271.05)	145.98 (165.88)	12.70 (70.14)	3.58 (51.95)	1.52 (32.28)	3.78 (18.92)	0.62 (18.17)	9.48
N-38	53.33 (161.61)	109.02 (123.88)	7.63 (42.13)	1.90 (27.57)	0.78 (16.61)	2.46 (12.31)	0.42 (12.44)	4.30
N-39	185.75 (562.88)	341.65 (388.23)	17.98 (99.36)	1.20 (17.44)	2.00 (42.57)	10.78 (53.91)	1.67 (49.26)	33.45

Country Rock

N-45	44.17 (133.85)	94.34 (107.20)	7.11 (39.28)	1.54 (22.38)	0.83 (17.74)	2.47 (12.37)	0.58 (17.03)	8.47
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Table 1 (cont'd).

IV. Gabal Nigrub El Fogani

<u>Syenites</u>	<u>La</u>	<u>Ce</u>	<u>Sm</u>	<u>Eu</u>	<u>Tb</u>	<u>Yb</u>	<u>Lu</u>	<u>Th</u>
F-1	52.12 (157.94)	113.61 (129.11)	8.79 (48.55)	3.23 (46.78)	0.94 (19.99)	3.19 (15.97)	0.53 (15.62)	7.28
F-5	28.49 (86.35)	72.46 (82.34)	3.71 (20.47)	0.78 (11.36)	0.27 (5.71)	2.28 (11.39)	0.36 (10.44)	8.96
F-9	56.33 (170.69)	114.46 (130.07)	9.67 (53.40)	3.33 (48.19)	1.08 (22.91)	3.27 (16.36)	0.50 (14.79)	9.56
F-10	51.09 (154.83)	114.64 (130.27)	7.32 (40.43)	0.93 (13.44)	0.92 (19.55)	2.14 (10.72)	0.49 (14.52)	5.55
F-12	50.27 (152.33)	105.74 (120.16)	6.30 (34.83)	1.30 (18.84)	0.60 (12.66)	2.69 (13.45)	0.48 (14.21)	6.70
F-14	91.22 (276.41)	162.82 (185.02)	15.33 (84.70)	4.44 (64.33)	1.45 (30.90)	6.02 (30.12)	0.52 (15.40)	19.58 (net-vein mixed rock)
F-15	48.51 (147.00)	94.80 (107.73)	6.09 (33.66)	1.49 (21.65)	0.63 (13.40)	2.83 (14.15)	0.37 (10.87)	8.03 (net-vein leucosyenite)
F-16	93.91 (284.57)	161.72 (183.77)	17.48 (96.55)	5.76 (83.50)	1.76 (37.36)	4.51 (22.57)	0.38 (11.05)	5.87 (net-vein melanosyenite)
F-19b	77.46 (234.73)	170.66 (193.93)	9.93 (54.89)	1.62 (23.52)	0.94 (20.07)	2.85 (14.27)	0.71 (20.78)	7.70
F-21	43.97 (133.24)	83.95 (95.39)	7.20 (39.79)	9.85 (142.69)	0.91 (19.29)	0.94 (4.72)	0.30 (8.76)	3.93
F-22	88.53 (268.26)	179.12 (203.55)	10.09 (55.73)	1.36 (19.72)	1.14 (24.17)	5.72 (28.59)	0.98 (28.74)	20.03
F-23	21.25 (64.41)	40.47 (45.98)	2.03 (11.20)	0.21 (3.01)	0.35 (7.43)	4.83 (24.14)	0.86 (25.35)	7.49
F-24	93.34 (282.86)	196.46 (223.25)	9.74 (53.81)	1.26 (18.23)	0.91 (19.37)	3.93 (19.63)	0.56 (16.46)	17.36
F-31	56.96 (172.62)	109.56 (124.50)	9.88 (54.60)	5.15 (74.64)	1.08 (22.88)	5.21 (26.06)	0.92 (27.08)	4.79
<u>Volcanic</u>								
F-17	94.44 (286.19)	206.25 (234.38)	14.34 (79.23)	2.64 (38.33)	2.25 (47.79)	9.72 (48.58)	1.70 (50.02)	13.95
<u>Dike</u>								
F-7	10.47 (31.73)	30.41 (34.56)	3.99 (22.06)	0.23 (3.39)	0.50 (10.56)	2.66 (13.32)	0.57 (16.77)	12.21
<u>Country Rock</u>								
F-4	8.23 (24.95)	33.55 (38.13)	1.71 (9.44)	0.61 (8.84)	0.18 (3.83)	0.72 (3.59)	0.17 (5.06)	2.26
F-30a	21.24 (64.37)	49.80 (56.59)	1.90 (10.51)	0.62 (8.94)	0.26 (5.61)	0.51 (2.53)	0.20 (5.83)	3.45

Table 1-A.

Average REE and Thorium Concentrations for the
Major Rock Types at Gabal Abu Khruq

Rock Type	Ce	Sm	Lu	Th
Foyaïtes n=18	159.47	11.23	0.64	19.78
Pulaskites n=7	115.01	9.90	0.49	10.37
Quartz Syenites n=13	195.29	18.61	1.12	17.40
Alkaline Syenites n=6	115.99	11.62	0.74	10.01
Volcanics n=9	210.04	19.65	1.06	16.94
Gabbros n=9	55.46	6.07	0.31	2.18
Gabbros* n=2	38.33	4.03	0.20	1.16

* Two outcrops of gabbro represent more
melanocratic varieties.

Figure 9.

Normalized REE patterns for the gabbros from Gabal Abu Khruq. Note large positive Eu anomalies and overall similar patterns except for one highly altered sample, A-76.

- a) More mafic, lower REE samples.
- b) All other gabbro samples.

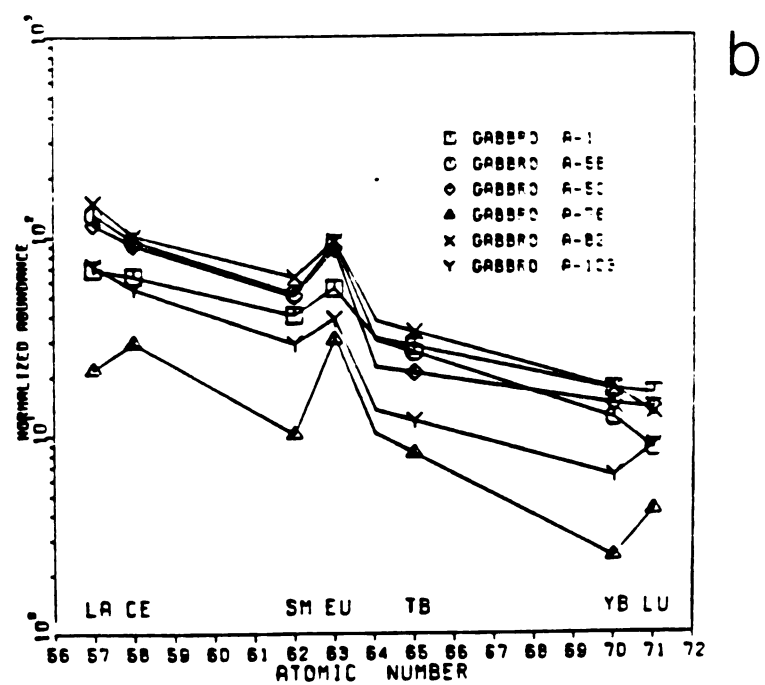
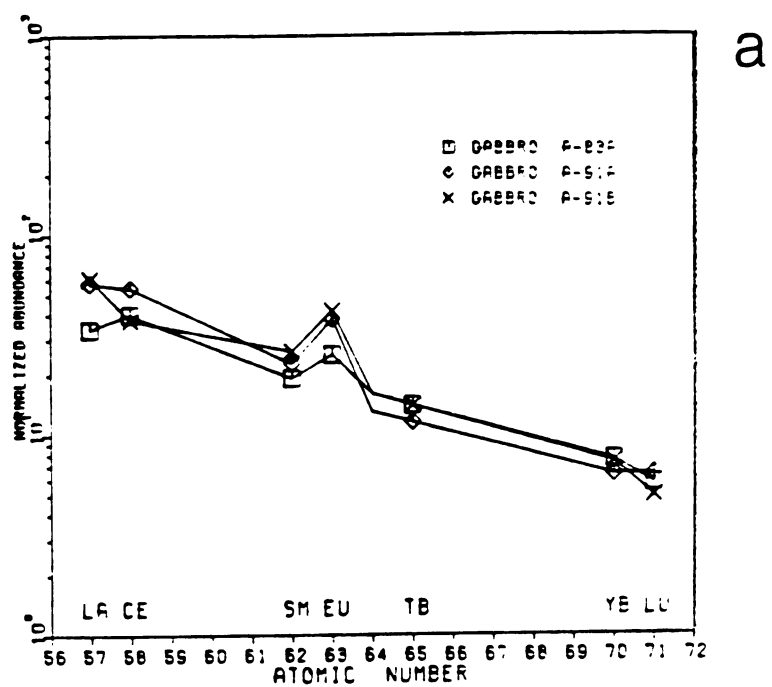


Figure 9.

Figure 10.

Normalized REE patterns for the low-REE syenites from Gabal Abu Khruq.

- a) Silica-undersaturated syenites which contain the lowest REE concentrations.
- b) Silica-undersaturated syenites with slightly higher REE concentrations.
- c) Silica-oversaturated syenites with REE concentrations comparable to the silica-undersaturated syenites in diagram (b).
- d) Pegmatoidal silica-undersaturated samples.

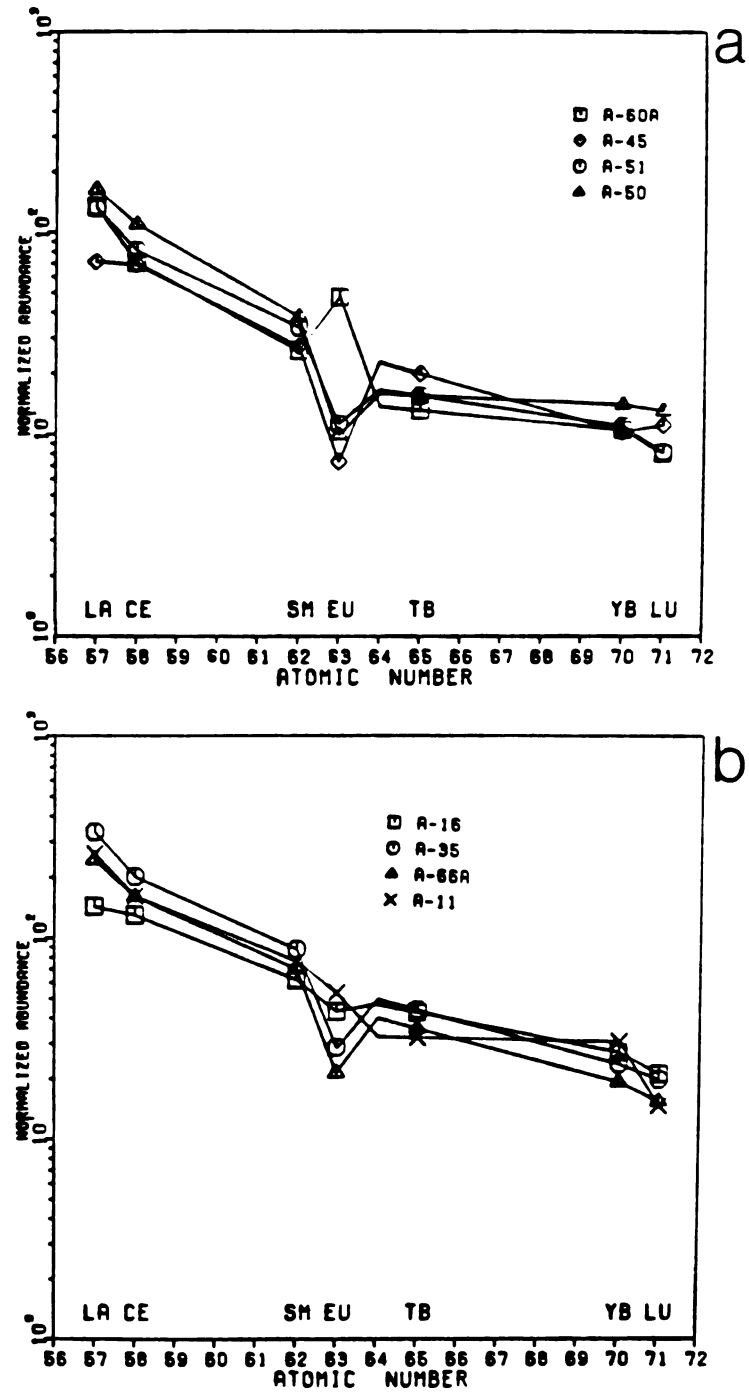


Figure 10.

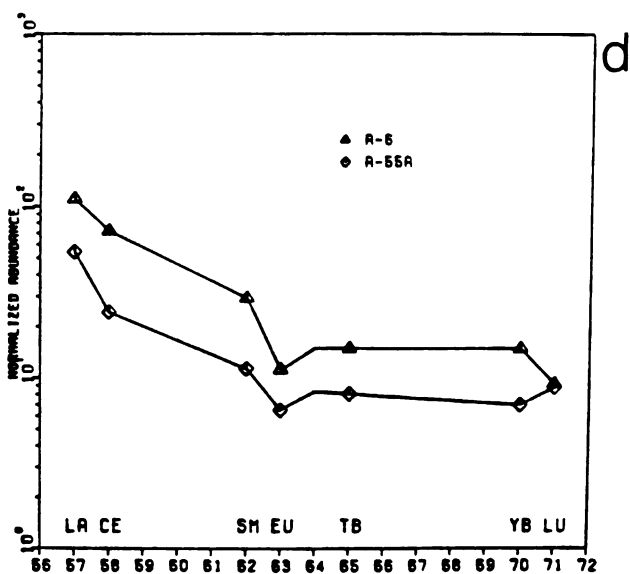
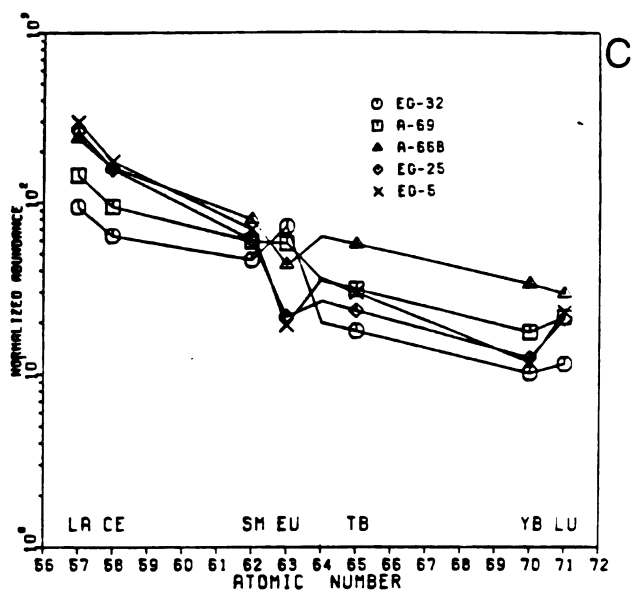


Figure 10 (cont'd)

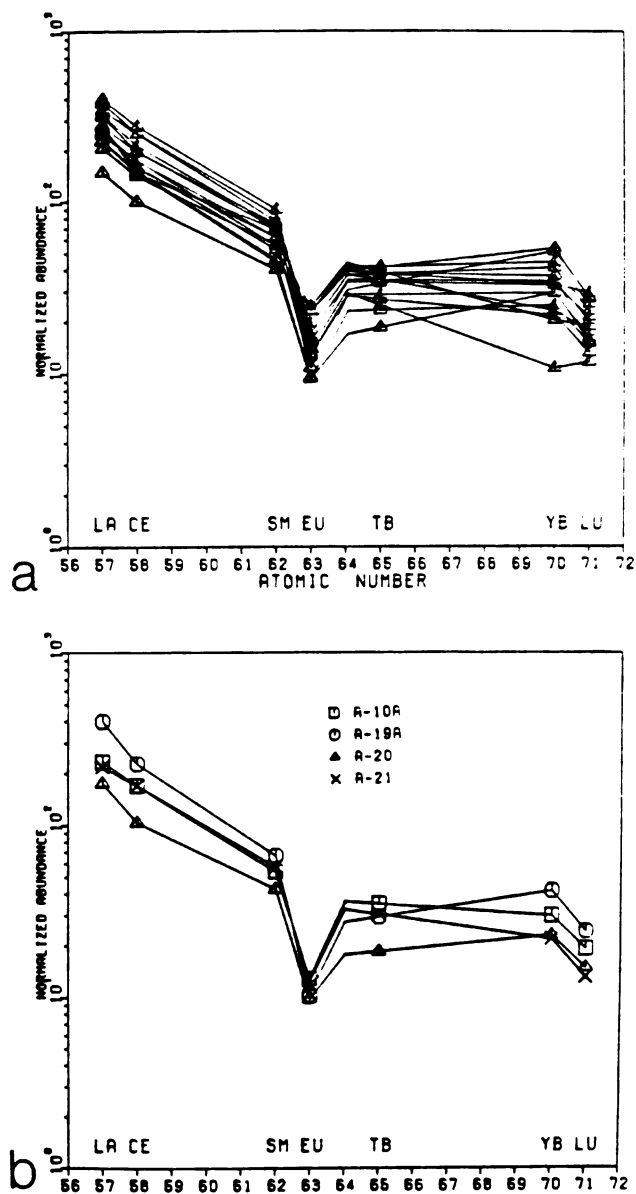


Figure 11.

Normalized REE patterns for the high-REE rock-types from Gabal Abu Khrug.

- a) Highly silica-undersaturated foyaitic syenites.
Note overall consistent pattern between individual samples.
- b) Extremely sodalite-rich foyaites.

Figure 11 (cont'd.)

- c) Quartz syenites. Note overall consistent pattern between individual samples and the higher concentrations relative to the foyaïtes.
- d) Quartz syenites which do not display the marked negative Eu anomalies common to most of the highly silica-oversaturated samples. These samples may represent more mafic syenites with large amounts of secondary quartz.
- e) Volcanic silica-oversaturated samples. Note similarity to the quartz syenites.

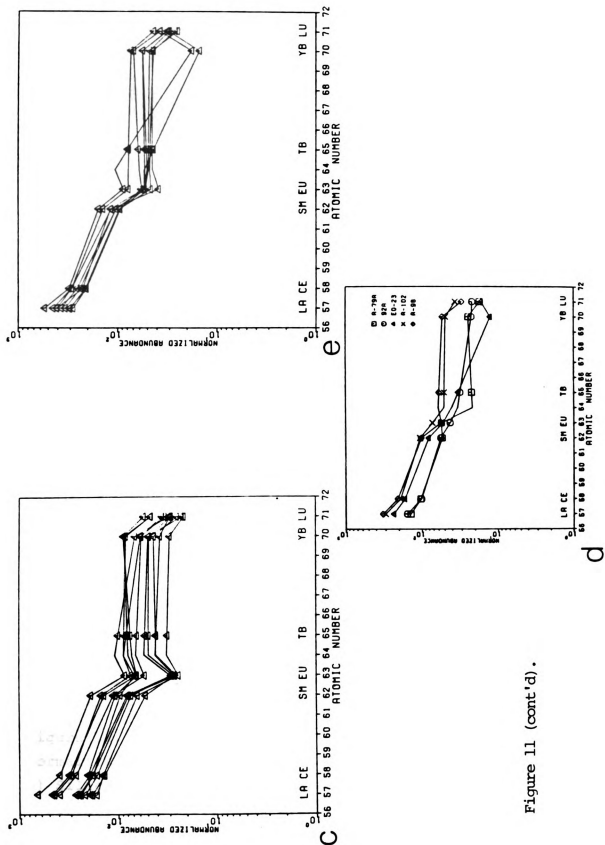


Figure 11 (cont'd).

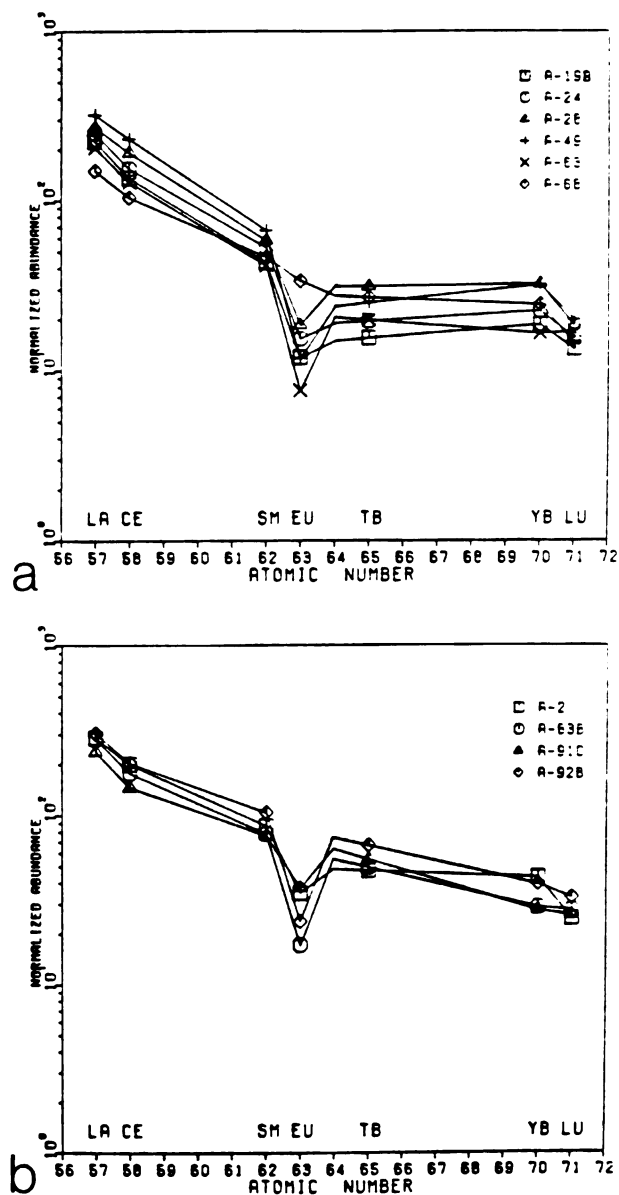


Figure 12.

Normalized REE patterns for the dike rocks.

- a) Silica-undersaturated dike samples. Note similarity to the foyaitic syenites.
- b) Silica-oversaturated dike samples. Note similarities to the quartz syenites.

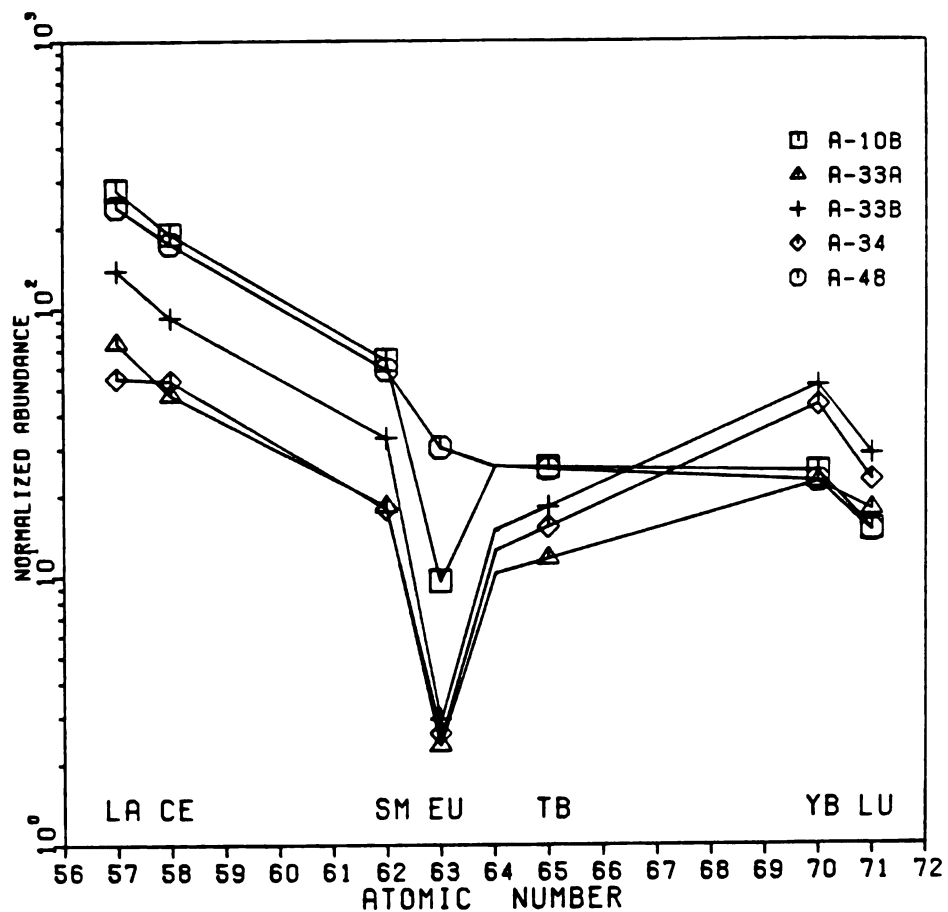


Figure 13.

Normalized REE patterns for xenoliths. Although some samples display very large negative Eu anomalies, they are otherwise similar to the silica-undersaturated dike rocks.

Several features should be noted in these normalized REE plots:

1. There is an overall sub-parallel relationship between the different varieties of rock-types.
2. The REE patterns are generally very similar for each rock-type, i.e. the REE distributions are not erratic.
3. The REE concentrations of the gabbros are less than those for the syenites and volcanics.
4. There is a wide range of Eu characteristics for this entire alkaline suite; from large positive Eu anomalies in the gabbros, to positive, flat, and negative Eu patterns in the pulaskites and alkaline syenites, to small negative Eu anomalies in the quartz syenites and volcanics, and finally, very large negative Eu anomalies in the foyaïtes.
5. The REE patterns of the pulaskites and alkaline syenites are very similar; both rock-types display a transition in Eu anomalies. The average absolute REE concentrations are slightly higher in the alkaline syenites than in the pulaskites.
6. The REE patterns of the quartz syenites and the volcanics are very similar.
7. The volcanics, which erupted early in the history of Abu Khruq, contain the highest average REE concentrations.
8. The foyaïtes, which were the last major stage of crystallization at Abu Khruq, contain average REE concentrations which are lower than the average REE concentrations of any of the other syenite and volcanic rock-types.

9. Careful inspection of the REE data reveals that some samples which appear to be highly differentiated, contain lower concentrations of REE than some samples which appear to be less differentiated (i.e. in thin section or with respect to other elements). This can be seen especially within the foyaites. Figure 10d illustrates an extreme example of this apparent REE-enrichment reversal for highly-evolved, pegmatoidal silica-undersaturated samples which according to field relationships, crystallized in the final stages of the development of Abu Khruq.

A number of dikes were analyzed for REE because of their abundance, their possible metasomatic affects on the host rock within the complex, and the possibility that some would provide examples of a parent magma (e.g. see Upton, 1974; Larsen and Steenfelt, 1974). Although there are many varieties of dikes (see El Ramly et al., 1969a), Figure 12 simply divides them into silica-undersaturated and oversaturated types. They are similar to the silica-undersaturated syenites, and quartz syenites/volcanics, respectively. They do not appear to have unusual REE concentrations to indicate associated metasomatic activity, nor do their compositions suggest a large mafic component.

Xenoliths have been shown to reveal important features about alkaline magmas (see DeLong et al., 1975; Ludden, 1977; Rock and Scoon, 1976). They occur in especially large quantities in silica-undersaturated suites, often coming from great depths in the mantle (Wilkinson, 1974).

At Abu Khruq, only the silica-undersaturated syenites contain significant amounts of xenoliths (see Appendix I). Petrographically, apart from some samples containing large amounts of Fe-rich biotite, they appear similar to the silica-undersaturated dikes, and therefore, consist of cognate material rather than mantle or deep-crustal xenoliths. Their REE patterns in Figure 13 are similar to the silica-undersaturated syenites, although they display larger negative Eu anomalies.

B. Other Trace Elements

Table 2 presents a limited number of analyses for compatible and other incompatible trace elements. The Rb and some of the Sr data is from Lutz (1979). The important features to be noted about these trace element distributions are:

1. The concentrations of the compatible trace elements with known ferromagnesian affinities (Ni, Cr, Co) are very low in the majority of the syenites and volcanics. The concentrations of the incompatible trace elements for these same rock-types are moderately high compared to the more mafic rock-types.
2. The Sr concentrations in the gabbros are very high; the Sr content is low in all other rock-types.
3. The Ba concentrations in a few syenites are very high. These samples appear more mafic in thin section, and contain higher concentrations of elements with mafic affinities. The Ba content is essentially zero in the foyaites.

Table 2. Other Trace Elements for the Egyptian Alkaline Complexes: G. Abu Khruq; G. El Kahfa; G. El Naga; G. Nigrub El Fogani.
Chlorine concentrations in ‰, all other elements are presented in ppm.

I. Gabal Abu Khruq				
Gabbros	Sr	Ba	Ni	Cr
A-83a	806	140	35	20.2
A-91a	967	140	26	94.4
<u>Silica-Undersaturated Syenites (F= foyaite; P= pulaskite)</u>				
A-10a	27.6	<1	7	10.0
A-16	65.7	440	11	<0.8
A-20	42.8	<1	7	6.0
A-21	29.3	<1	8	2.5
A-56a	286	<1	66	7.7
A-59a	317	470	33	3.9
A-60a	521	980	11	8.0
A-66a	128	310	9	9.3
<u>Silica-Oversaturated Syenites (Q.S.= quartz syenite; A.S.= alkaline syenite)</u>				
A-72	23.2	80	15	7.7
A-77	54.4	440	13	9.6
Eg-25	37.5	70	6	7.5
Eg-32	536	1300	9	11.9
<u>Volcanics</u>				
A-41a	31.3	570	9	<0.8
A-90	18.9	140	9	<0.8
Eg-14	22.5	170	7	9.3
Eg-16	28.6	150	9	11.5
<u>Dike (Silica-Oversaturated)</u>				
A-2	40.4	210	7	6.3
<u>Xenoliths (Silica-Undersaturated Host Rock)</u>				
A-10b	32.4	<1	8	6.7
A-48	356	490	53	7.3
<u>Wehrlite (Ultramafic Olivine-rich Cumulate)</u>				
A-94	96.5	<1	1040	989

Sr-Rb Data from Lutz(1979) for G. Abu Khruq				
Gabbros	Sr	Rb	Volcanics	
E-7	919.0	16.82	E-66	31.19
E-10	904.2	12.56	E-67	26.17
E-64	668.1	13.91	E-68	28.80
<u>Syenites: Undersaturated</u>				
E-2	12.57	104.2	E-1	16.13
E-3	49.56	103.7	E-11	177.5
E-3a	9.37	117.3	E-21	18.01
E-5	11.64	98.87	E-22	67.85
E-9	223.1	79.27	E-23	26.54
E-25	77.45	49.59	E-29	15.04
E-26	44.31	124.9	E-30	100.9
E-27	28.97	113.2	E-31	32.41
E-28	23.14	161.2	E-61	10.37
E-69	13.53	88.15	E-62	7.75
E-70	23.79	121.0	E-63	6.70
E-71	10.52	98.50	E-65	25.14
<u>Oversaturated</u>				
E-1	16.13			88.44
E-11	177.5			84.96
E-21	18.01			88.67
E-22	67.85			57.69
E-23	26.54			88.43
E-29	15.04			78.64
E-30	100.9			49.94
E-31	32.41			54.45
E-61	10.37			84.57
E-62	7.75			107.5
E-63	6.70			149.7
E-65	25.14			51.42

Chlorine Data from El Ramly et al. (1969a) for G. Abu Khruq. El Ramly et al.'s nomenclature is used.

<u>Silica-Undersaturated Syenites</u>		
11-Ditroite	0.14%	Average Nepheline Syenite
12-Foyaite	0.03%	Average Foyaite
13-Umptekite	0.20%	Average Ditroite

0.07%	0.09%	0.03%
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Table 2 (cont'd).

II. Gabal El Kahfa				IV. Gabal Nigrub El Fogani			
	Sr	Ba	Ni	Cr	Co		
<u>Gabbros</u>							
K-9	566	950	64	16.0	22		
K-39	895	380	24	14.8	30		
<u>Syenites</u>							
K-13	202	760	55	12.9	19		
K-16b	578	520	31	27.8	35		
K-24	16.8	<1	5	4.9	9		
K-28	163	1460	8	3.7	9		
K-33	22.7	70	7	7.3	8		
K-41	198	2010	9	8.9	17		
K-43c	510	780	25	10.0	24		
K-44	389	470	33	52.3	20		
K-52	365	310	36	50.5	20		
K-56a	541	690	14	14.9	21		
<u>Volcanic</u>							
K-19	587	530	31	25.6	32		
K-62	794	29.0	15	157	28		
III. Gabal El Naga							
<u>Syenites</u>							
N-7a	760	1750	7	6.6	6		
N-10	348	4270	9	8.9	3		
N-14	106	1720	6	6.0	3		
N-23	671	2730	7	6.2	9		
N-33	342	4900	9	7.2	11		
<u>Volcanic</u>							
N-1a	335	2080	8	7.0	11		

4. Sample A-94 represents an isolated outcrop of an ultramafic, serpentized rock-type which is considered to be part of the early development of Abu Khruq.

5. There are large differences between some samples in terms of certain trace elements, and very little differences in terms of other trace elements for those same samples.

This is considered to be due to the cumulate nature of some individual samples, and perhaps due to the unknown affect which mechanisms other than crystal fractionation may have exerted on the trace element distributions.

CHAPTER 4

THE PETROGENESIS OF SILICA-UNDERSATURATED/ OVERSATURATED ALKALINE COMPLEXES

Introduction

This chapter will discuss the interpretation of the chemical data on the Egyptian alkaline complexes in general, but with more emphasis on the four silica-undersaturated/oversaturated associations sampled by this study. In view of the general chemical similarities between these alkaline complexes, Gabal Abu Khruq is considered to serve as an example of the type of processes operative in the evolution of many of the Egyptian alkaline complexes. The obvious differences with respect to some of the other complexes will be discussed, and these will serve as examples of the types of variations possible.

The Egyptian alkaline suites are composed of predominantly salic, low-pressure alkaline rock-types which have many characteristics of a volatile-rich residual liquid. The presence of typical residual mineral phases, high chlorine concentrations, and widespread metasomatic/hydrothermal features give some indication of the low-temperature, halogen-rich nature of these alkaline magmas. This is supported by the projection of the compositions of the syenites into the thermal minimum of the Q-Ne-Ks residual system (Lutz, 1979, pp. 122-123). These types of melt systems have often been associated with many complex,

erratic, and poorly-defined mechanisms. Further, the majority of the rocks in many of the Egyptian alkaline complexes are coarse-grained, and unlike volcanic rocks, they often cannot be considered to represent the composition of the melt at the time of their crystallization. Rather, they may represent an unknown combination of crystals and trapped liquid (e.g. Irving, 1979). These features complicate a petrogenetic solution and cause scatter in the compositional trends of the chemical data for the Egyptian alkaline suites.

Section I

Metasomatism

The chemistry of many salic alkaline suites have been reported to be strongly affected by metasomatic fluids, especially in the final stages of crystallization (Mineyev, 1963; Ganzeyeva and Ganzeyeva, 1975; Alekseyev, 1970; Bowden and Whitley, 1974; Mitchell and Brunfelt, 1975; Eby, 1975; Borodin and Pavlenko, 1974; Barber, 1974; Rock, 1976; Martin et al., 1978). Before the chemistry of the Egyptian alkaline suites can be interpreted as primary petrogenetic indicators, the extent of metasomatism for each complex should be addressed.

Gabal Nigrub El Fogani contains signs of extensive alteration (see Appendix I). Much of it appears to be autometasomatism and may relate to the presence of CO₂-rich phases during the latter part of the crystallization of the

melt and the formation of carbonatite dikes. There is substantial evidence for alteration due to migrating fluids at Gabal El Kahfa. Rather explosive events are likely to account for the large cataclastic zones within the complex, as well as the fused and altered appearance of many of the rocks within the complex and the surrounding gabbros, volcanics, and amphibole schists. Gabal El Naga shows the least amount of alteration, both within and outside the complex (contrary to the observations of El Ramly et al., 1969b, see Appendix I). The country rock surrounding Gabal Abu Khruq contains little signs of widespread fenitization. Figure 14 shows the normalized REE distributions of three samples of country rock; one from the granite gneiss and two from the amphibole gneiss. Their REE concentrations (as well as Th concentrations, see Chapter 3) are low and not unlike the patterns of similar metamorphic rocks (refer to Shaw, D.M., et al., 1976; McClennen et al., 1979; Nance and Taylor, 1977). Petrographically, only A-64 shows signs of extensive alteration and its REE distribution is not significantly different than A-67 which is otherwise similar in thin section. A-57 is a particularly useful sample because it was collected from the granite gneiss-alkaline complex contact. It appears to contain its original quartz and have an overall fresh appearance in thin section. This would not support extensive metasomatism at Abu Khruq, nor a silica-transport mechanism from the surrounding country rock to the alkaline complex (see Lutz, 1979, pp. 233-242).

Figure 14.

Normalized REE patterns of three samples of country rock which surrounds Gabal Abu Khruq. In general, these REE patterns, similar to their petrographic appearances, do not indicate extensive fenitization or alteration due to the adjacent alkaline magmatic activity. These REE patterns are similar to those of equivalent metamorphic rocks in other areas (see text).

Figure 15.

SiO_2 versus Sm for the rock series at G. Abu Khruq. The positive relationship between SiO_2 and Sm is consistent with fractional crystallization rather than silica-leaching and associated REE-enrichment via fenitization (see Martin et al., 1978).

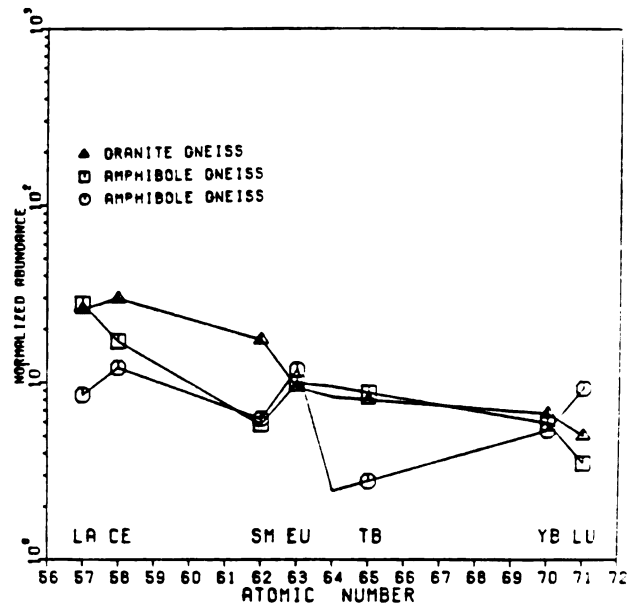


Figure 14.

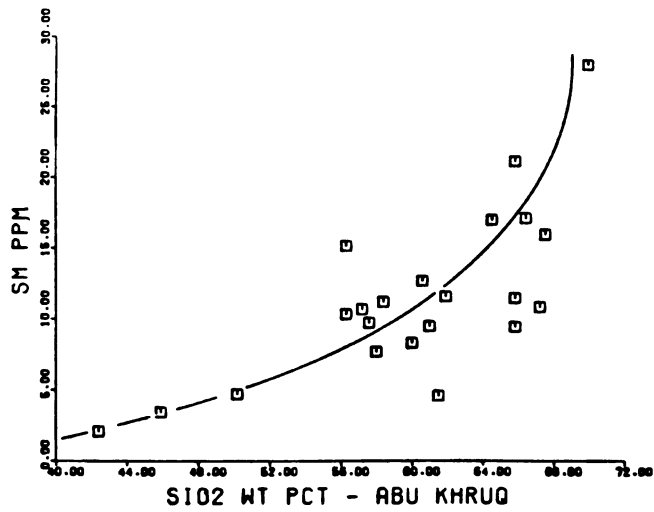


Figure 15.

Silicate rocks derived by fractional crystallization should show a sympathetic increase between SiO_2 and REE. A negative correspondence might indicate fenitization and the transport of silica (see Martin et al., 1978). Figure 15 demonstrates a positive relationship. This trend does not support extensive post-magmatic fenitization, nor the formation of the silica-undersaturated rocks via fenitization of originally silica-undersaturated rocks.

A general indication of the absence of large amounts of fenitization at G. Abu Khruq (as well as El Naga) is the overall sub-parallel normalized REE patterns for the entire rock series.

To summarize, there is evidence of strong alteration at G. El Kahfa, and late-stage metasomatic activity at G. Nigrub El Fogani. There is little evidence for widespread metasomatic fluids in disequilibrium with the alkaline melt systems at G. El Naga and G. Abu Khruq.

Section II

Fractional Crystallization

Introduction

The gabbro-syenite association has been noted in alkaline complexes throughout the world (e.g. Sørensen (ed.), 1974). Since Bowen's (1945) postulation of the "plagioclase effect" many workers have drawn upon the fractionation of large amounts of plagioclase (found in the gabbros) from an originally basaltic magma to explain the development of

peralkaline residual melts which crystallize and form the syenites and alkaline volcanics (see Upton, 1974; MacDonald, 1974; Edgar, 1974). In a preliminary way, the field relationships between the syenites and gabbros, and the general major element and trace element compositions of the rock series in the Egyptian alkaline complexes are qualitatively consistent with the involvement of crystal fractionation.

There are two general ways in which fractional crystallization may be shown to have occurred in the evolution of the Egyptian alkaline complexes. One approach entails a comparison of the chemistry of the Egyptian alkaline complexes with that of alkaline suites in other areas which have frequently been demonstrated to have evolved largely through the process of crystal fractionation. The other approach qualitatively assesses the chemical trends of the Egyptian alkaline complexes in light of the trends expected for crystal fractionation. The paucity of highly mafic end-members in the four alkaline complexes of this study does not allow a rigorous petrogenetic solution. This lack of mafic end-members is considered to be due to the efficient differentiation of mantle-derived, volatile-rich alkaline melts, resulting in the formation of ultramafic cumulate strata mostly below the exposed level of the alkaline complexes. Plutonic/volcanic alkaline complexes similar to those found in Egypt frequently contain evidence of associated lower units of gabbro and ultramafic stratiform complexes (e.g. Muñoz, 1969; Gastesi, 1969; Upton,

1974; Bridgewater and Harry, 1968; Fernanadez, 1980).

1) Comparison of the Chemistry of the Egyptian Alkaline Complexes to Other Alkaline Suites

The worldwide similarity of many primitive alkaline suites was discussed in Chapter 3. Many of the compositional trends of these suites have been frequently interpreted to be the result of crystal fractionation. Due to the lack of representation of mafic end-members in the Egyptian alkaline complexes, it is difficult to reconstruct the early differentiation stages during which crystal fractionation is likely to have been an important process. However, by comparing the compositional trends of the Egyptian alkaline complexes to these alkaline suites in other areas, it is possible to qualitatively note the conformity of the Egyptian data to the trends expected for crystal fractionation.

Figures 16-20 illustrate the similarity between the chemical trends of numerous alkaline suites--many of which have been interpreted to be the result of crystal fractionation. Similar chemical data was plotted for a large number of alkaline suites from different areas of the world. In addition, Dr. R. Schwarzer (Rice University) supplied this study with unpublished AFM and silica-alkali compilations of alkaline suites formed in different tectonic settings which supplemented the data presented by Schwarzer and Rogers (1974). These compilations further substantiate

Figure 16.

AFM ternary diagrams for several well-known oceanic alkaline suites which are similar to those of the Egyptian alkaline complexes, and similar to trends expected for fractional crystallization. Data compiled from Bishop and Wooley, 1973; Goldich et al., 1975; and Schmincke, 1973.

*Refer to Figures 4 and 5 in Chapter 3 for AFM ternary diagrams of the Egyptian alkaline complexes.

Figure 17.

Composite calcic-alkali ternary diagram of several well-known oceanic alkaline suites which are similar to those of the Egyptian alkaline complexes, and similar to trends expected for fractional crystallization. Data from Schmincke, 1973.

*Refer to Figure 6 in Chapter 3 for calcic-alkali ternary diagram of the Egyptian alkaline complexes.

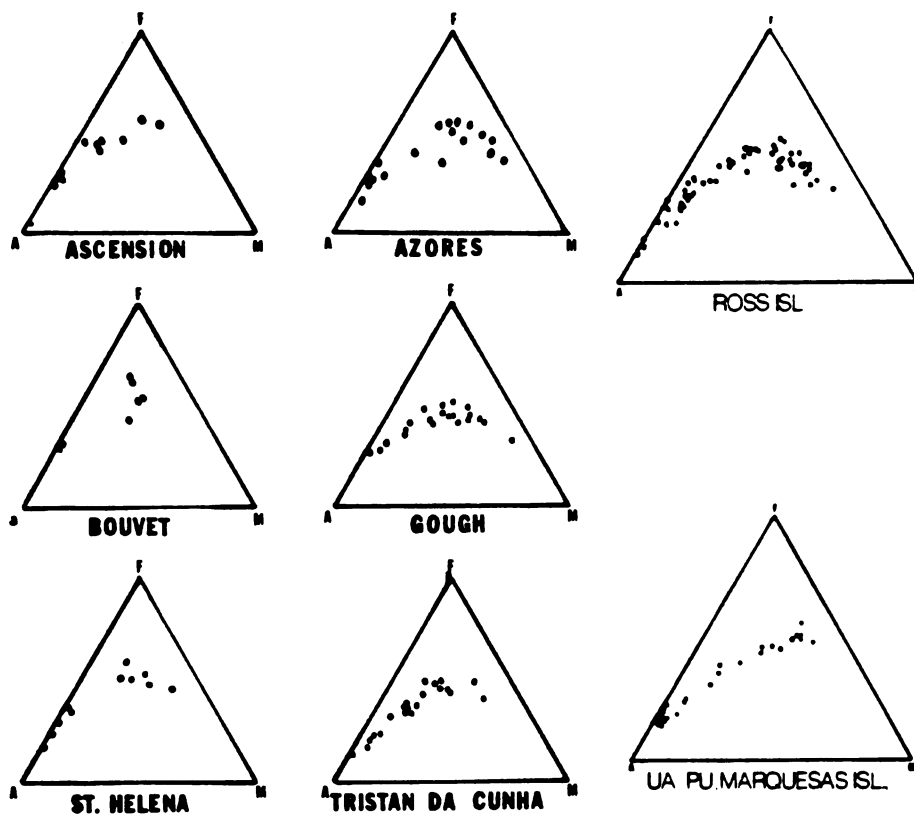


Figure 16.

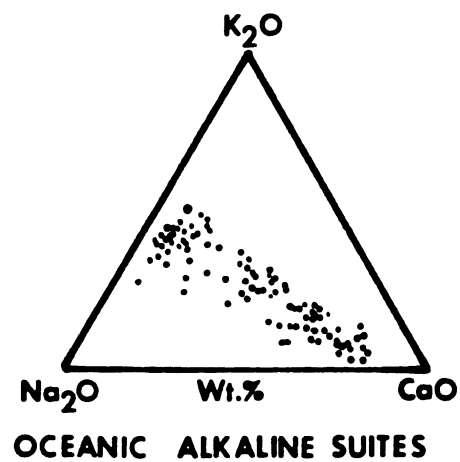


Figure 17.

Figure 18.

Silica-alkali diagrams for a selection of alkaline suites from continental, oceanic, extensional, compressional, and intraplate tectonic settings. These trends are similar, and there is no apparent systematic difference according to crustal setting. See text for further explanation. Data compiled from:

1. Tristan da Cunha, Atlantic Ocean,
2. Ross Island, Antarctic Ocean, Goldich et al., 1975.
3. Gough Island, Atlantic Ocean, Le Maitre, 1962.
4. Tahiti, Pacific Ocean, McBirney and Aoki, 1968.
5. Ua Pu, Marquesas Island, Pacific Ocean, Bishop and Woolley, 1973.
6. Reunion Island, Indian Ocean, Upton and Wadsworth, 1972.
7. Hanish-Zukur, Red Sea, Gass et al., 1973.
8. Jebel al Abyad, Saudi Arabia, Baker et al., 1973.
9. Kenya Rift, Africa, Saggerson, 1970; see also Lippard, 1973;
King, 1965; King and Chapman, 1972.
10. Monchique, Portugal, Rock, 1978.
11. Sintra, Portugal, Sparks and Wadge, 1975.
12. Cuttingsville, Canada, Laurent and Pierson, 1973.

*See also Figure 29 for another silica-alkali trend from the Canary Islands which is very similar to that of G. Abu Khruq in Egypt.

**Refer to Figure 3 for the silica-alkali diagrams of the Egyptian alkaline complexes.

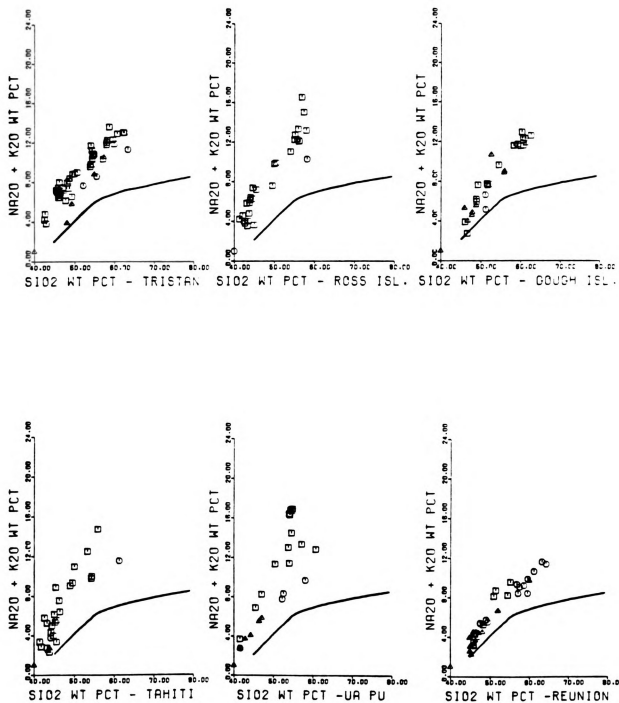


Figure 18.

- NE-NORMATIVE
- QTZ-NORMATIVE
- ▲ SiO_2 -SATURATED

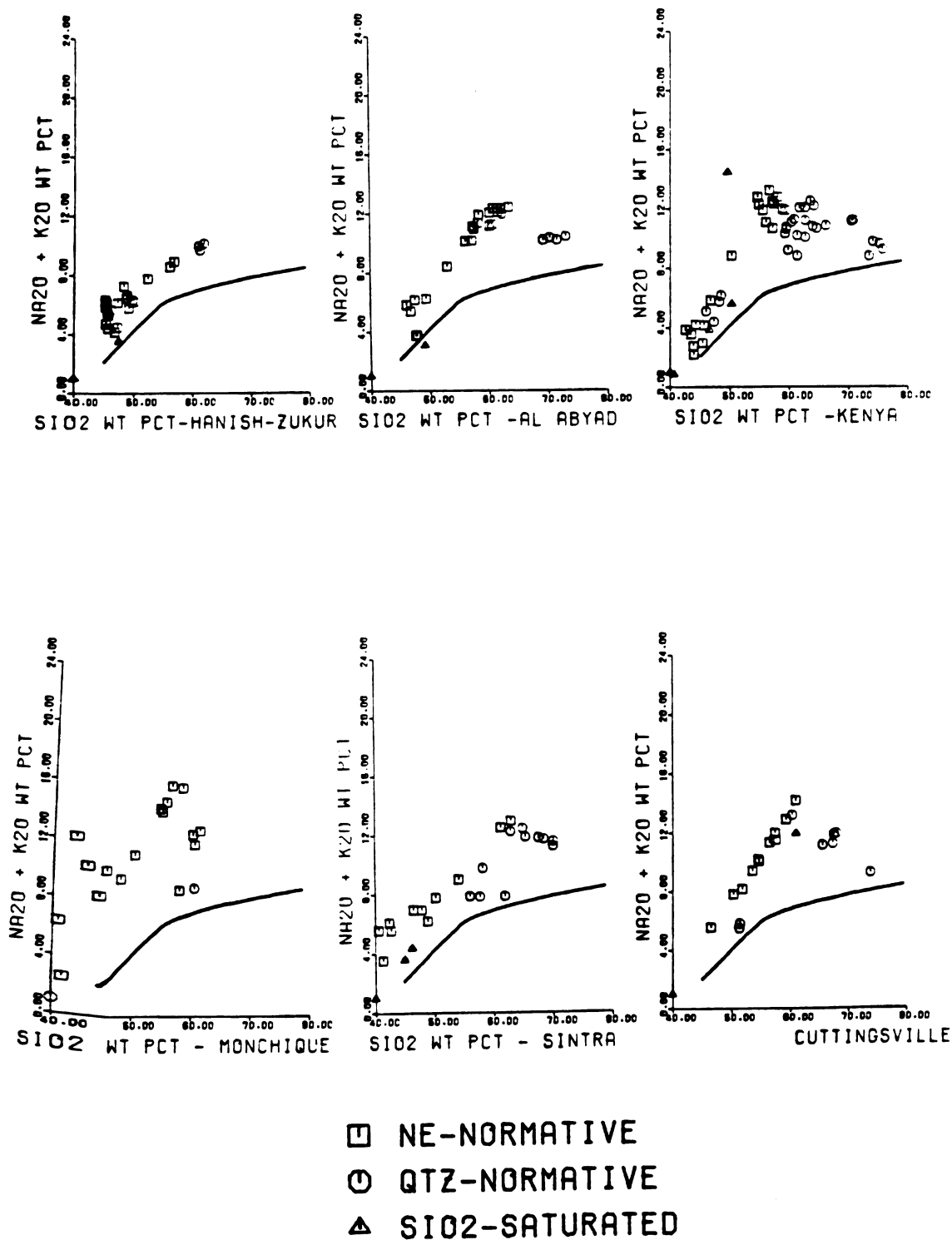


Figure 18 (cont'd).

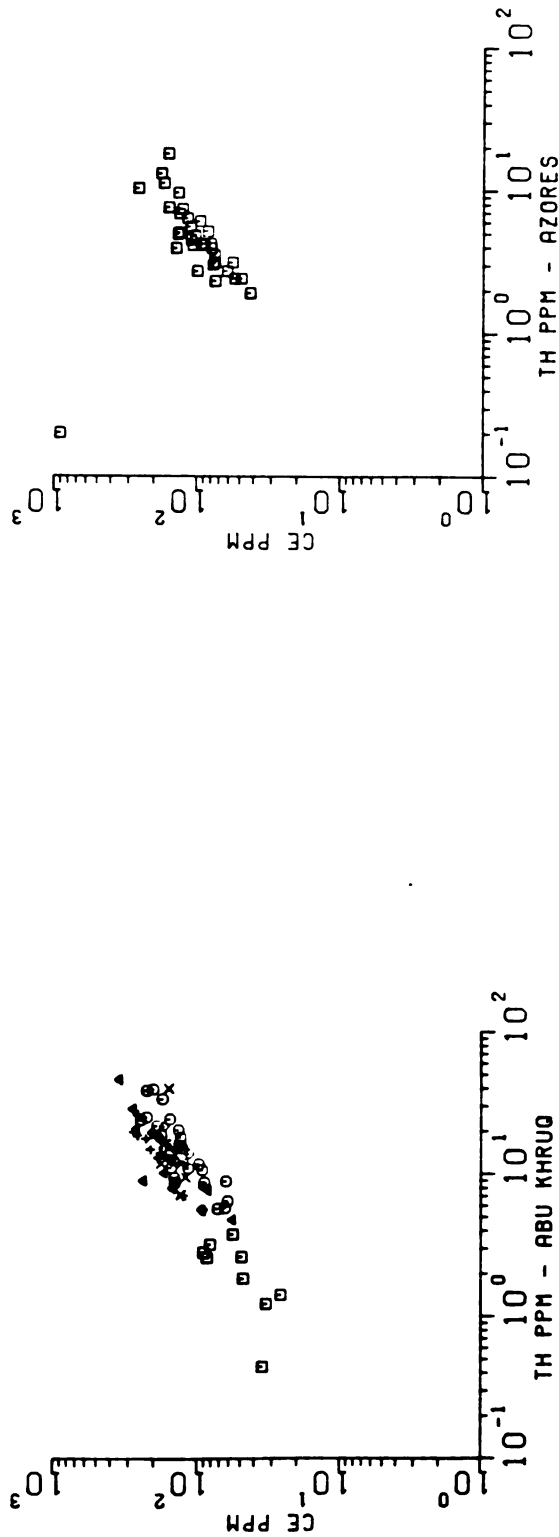


Figure 19.

Log thorium versus log cerium diagrams for an alkaline suite from the Azores and G. Abu Khruq, Egypt. Note similar concentration ranges and enrichment slopes in these two trace element trends from very different crustal settings. (REE data from Azores: White et al., 1979).

Figure 20.

Ce/Sm versus Ce diagrams comparing the REE enrichment and fractionation trends of well-known oceanic alkaline suites and that of G. Abu Khruq, Egypt. Though G. Abu Khruq demonstrates more scatter as a single suite, the concentration ranges and enrichment slopes are very similar. The larger amount of scatter is attributed to the plutonic nature of G. Abu Khruq. Data on the oceanic suites is compiled from:

1. Azores, White et al., 1979.
2. Reunion, Ludden, 1978; Zielinski, 1975.
3. Gough, Zielinski and Frey, 1970.
4. Grenada, Arculus, 1976.
5. Ross, Sun and Hanson, 1976.

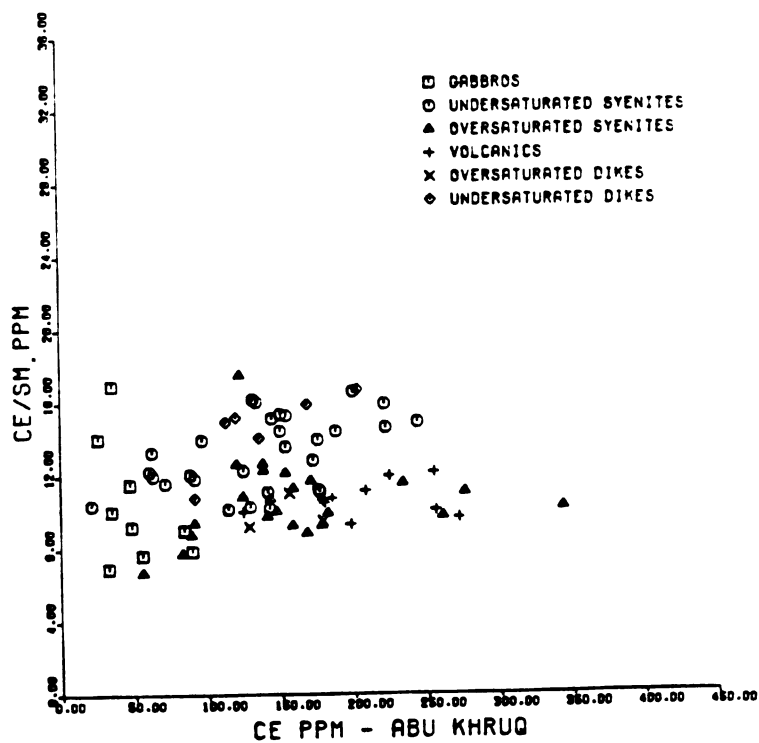
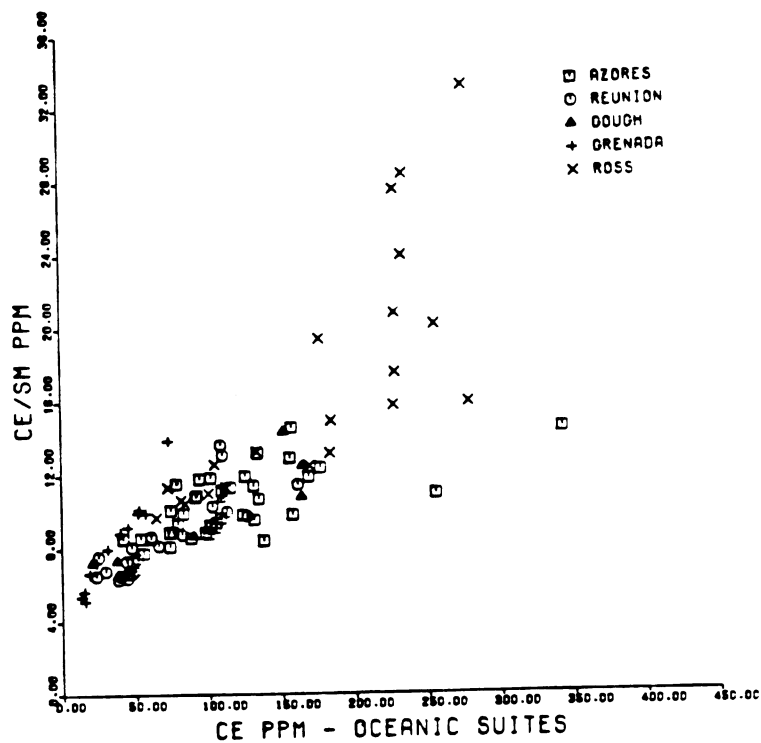


Figure 20.

their observation of the worldwide compositional regularities of alkaline suites. By referring to the diagrams presented here and in Chapter 2, one can note the similarity of these chemical trends to those of the Egyptian alkaline complexes. This similarity exists on both the major element and trace element levels, and appears to persist between alkaline occurrences found in different crustal settings. For the data on the Azores alkaline suite in particular, White et al. (1979) concluded that the dominant mechanism of differentiation was fractional crystallization. The strong similarity of the REE and Th data between the Azores and Abu Khruq in terms of both enrichment slope and concentration ranges, may indicate that fractional crystallization was an important mechanism in the melt at Abu Khruq also, Figure 19.

2) Fractional Crystallization in Relation to the Trace Element Distributions of the Egyptian Alkaline Complexes

The general depletion trends of the compatible trace elements and the general enrichment trends of the incompatible trace elements in the four Egyptian alkaline complexes of this study are similar to those expected for the differentiation of an alkaline magma in which crystal fractionation was an important mechanism.

A. Compatible Trace Elements -- Gabal Abu Khruq

Ni, Cr, and Co are strongly incorporated into the mafic mineral phases. Ni is most strongly partitioned into

the structure of olivine, Cr into pyroxene and Fe-oxides, and Co into olivine and pyroxene, but to a much lesser extent than Ni or Cr. The distribution of Co with respect to Th (used as the magma crystallinity index) for Abu Khruq is shown in Figure 21. The distributions for Ni and Cr are similar, but show an even sharper depletion (see Chapter 3). The lack of exposed mafic rock-types at Abu Khruq does not provide well-defined trends for these compatible trace elements with ferromagnesian affinities.

Sr and Ba are especially sensitive to feldspar fractionation. Sr serves as an indication of the fractionation of particularly plagioclase; Ba is strongly partitioned into the structure of alkali feldspar (e.g. Hanson, 1978; Baker et al., 1977). Variation diagrams are shown for Sr and Ba in Figure 22. The Sr trend demonstrates the efficient fractionation of plagioclase into the gabbros with the resulting severe Sr depletion in the majority of the exposed rock-types. Ba is interpreted to show a curve which is typical for a rock series in which alkali feldspar becomes the latter dominant fractionating phase and depletes the remaining melt in Ba. Note the inferred similarity to the Gregory Rift alkaline volcanic series shown in the inset of Figure 22b, where Zr is used as the index of crystallinity.

The five compatible trace elements presented are all interpreted to show the general trends expected for the crystal fractionation of an alkaline basaltic magma. The

Figure 21.

Cobalt versus thorium (index of crystallinity) for G. Abu Khruq. It is interpreted to illustrate the efficient fractionation of the ferromagnesian minerals early in the evolution of the melt at G. Abu Khruq.

Symbols are the same in this figure and Figure 22.

□ OL-CPX CUM.
 ○ DABBROS
 ▲ UNDERSATURATED SYENITES
 + OVERSATURATED SYENITES
 X VOLCANICS

Figure 22.

Trace elements with feldspar affinities versus thorium (index of crystallinity) for G. Abu Khruq.

- a) Strontium versus thorium; interpreted to illustrate the efficient fractionation of plagioclase. See Figure 26 for a similar presentation with a crystal fractionation path superimposed.
- b) Barium versus thorium; interpreted to illustrate efficient fractionation of alkali feldspar during the intermediate stages in the evolution of the melt at G. Abu Khruq. The inset diagram is from Baker et al. (1977) of data from the alkaline volcanics of the Gregory Rift. It serves to illustrate with less scatter than the intrusive Egyptian data, a similar type of Ba pattern (using Zr as the index of crystallinity). As is common in the literature on alkaline suites, this pattern is attributed to the fractionation of alkali feldspar during the intermediate stages of fractional crystallization.

Symbols are the same as in Figure 21.

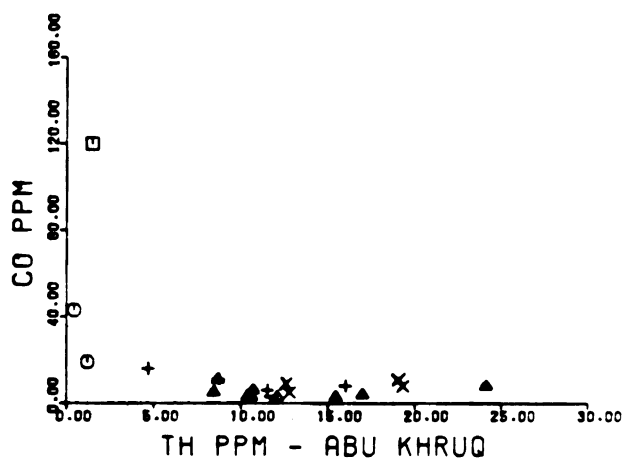


Figure 21.

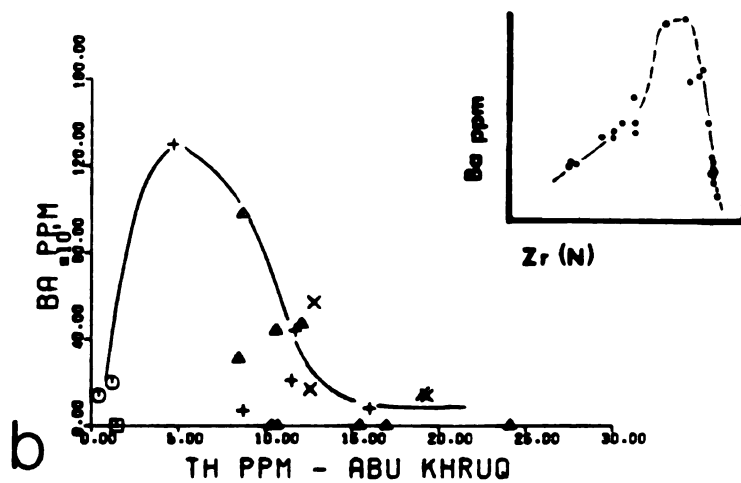
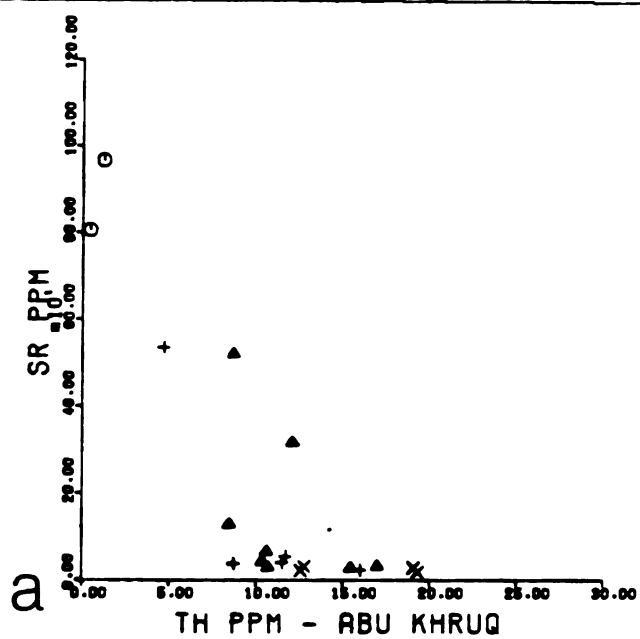


Figure 22.

lack of mafic rocks at Abu Khruq and the number of analyses does not allow a more rigorous treatment of these compatible trace element distributions. However, based on the order of crystallization of the major phases in a basaltic magma and the compatibility of these trace elements for these individual phases, these depletions are interpreted to be the result of efficient crystal fractionation of especially olivine, diopside, labradorite, and alkali feldspar--all observed in the mineralogy of the rocks at Abu Khruq.

B. Incompatible Trace Elements -- The Four Alkaline Complexes of this Study

Alkaline suites are characteristically bimodal in the distribution of their major elements (e.g. Petro, Vogel and Wilband, 1979). As is evident in many of the major element diagrams, the Egyptian alkaline complexes appear to be well-represented by the salic and some mafic compositions, with few intermediate compositions. Figure 23 is a frequency distribution of Th, a strongly incompatible trace element, for 188 samples from the four alkaline complexes sampled by this study. Although there is a sampling bias due to the method of collection, there is no hint of a bimodal pattern which coincides with the mafic-salic bimodal distribution of the major elements. The lack of such a compositional gap is consistent with fractional crystallization. It may be inferred that the bimodal distribution of the major elements is simply a function of efficient crystal

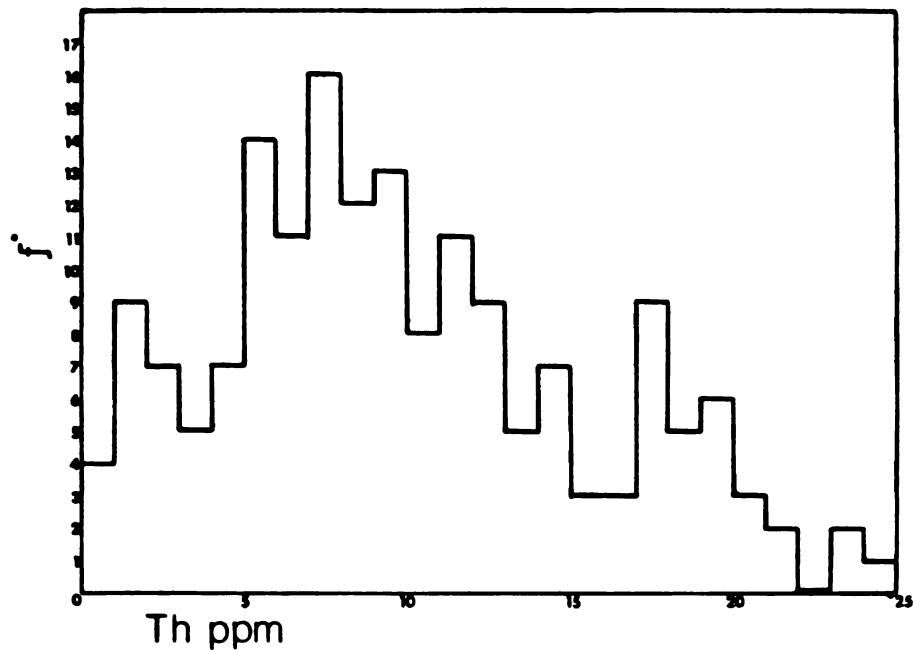


Figure 23.

Frequency diagram of thorium concentrations from four Egyptian alkaline complexes (188 analyses from G. El Naga, G. Nigrub El Fogani, G. El Kahfa, and G. Abu Khruq). There does not appear to be a bimodal thorium distribution (incompatible trace element) which coincides with a mafic versus salic bimodal distribution as noted in the major element diagrams for these and many other alkaline suites in other areas. Frequency plots of individual alkaline complexes are similar to this composite diagram.

fractionation and the formation of distinct intervals of crystallization of the major phases (see Clague, 1978; Wood, 1978).

A major problem in solving the petrogenesis of many alkaline complexes is the origin of the gabbros which are typically associated with the syenites and alkaline volcanics. Upton (1974) suggested that the REE data might be used to test whether these gabbroic bodies can be derived by extensive plagioclase crystal fractionation from a basaltic magma. Although this study's sampling does not include any plausible parental basaltic representatives, the overall sub-parallel REE-enrichment and opposing Eu anomalies of the gabbros and the syenites/volcanics are typical of plagioclase separation (Eu-rich) and a residual Eu-depleted liquid.

The sub-parallel nature of the normalized REE plots for the entire suite of major rock-types at Abu Khruq is illustrated in Figure 24. Assuming a mantle source region containing garnet, such a gradual sub-parallel enrichment of REE is unlike that expected for a partial melting relationship (which tends to produce large LREE/HREE changes), and similar to that expected for the fractional crystallization of the dominant minerals observed in the rock-types at Abu Khruq (clinopyroxene and feldspars).

Figure 25 shows Ce versus Ce/Sm for the four Egyptian alkaline complexes of this study. The REE enrichment (Ce) versus the REE fractionation (Ce/Sm) trends of Abu Khruq,

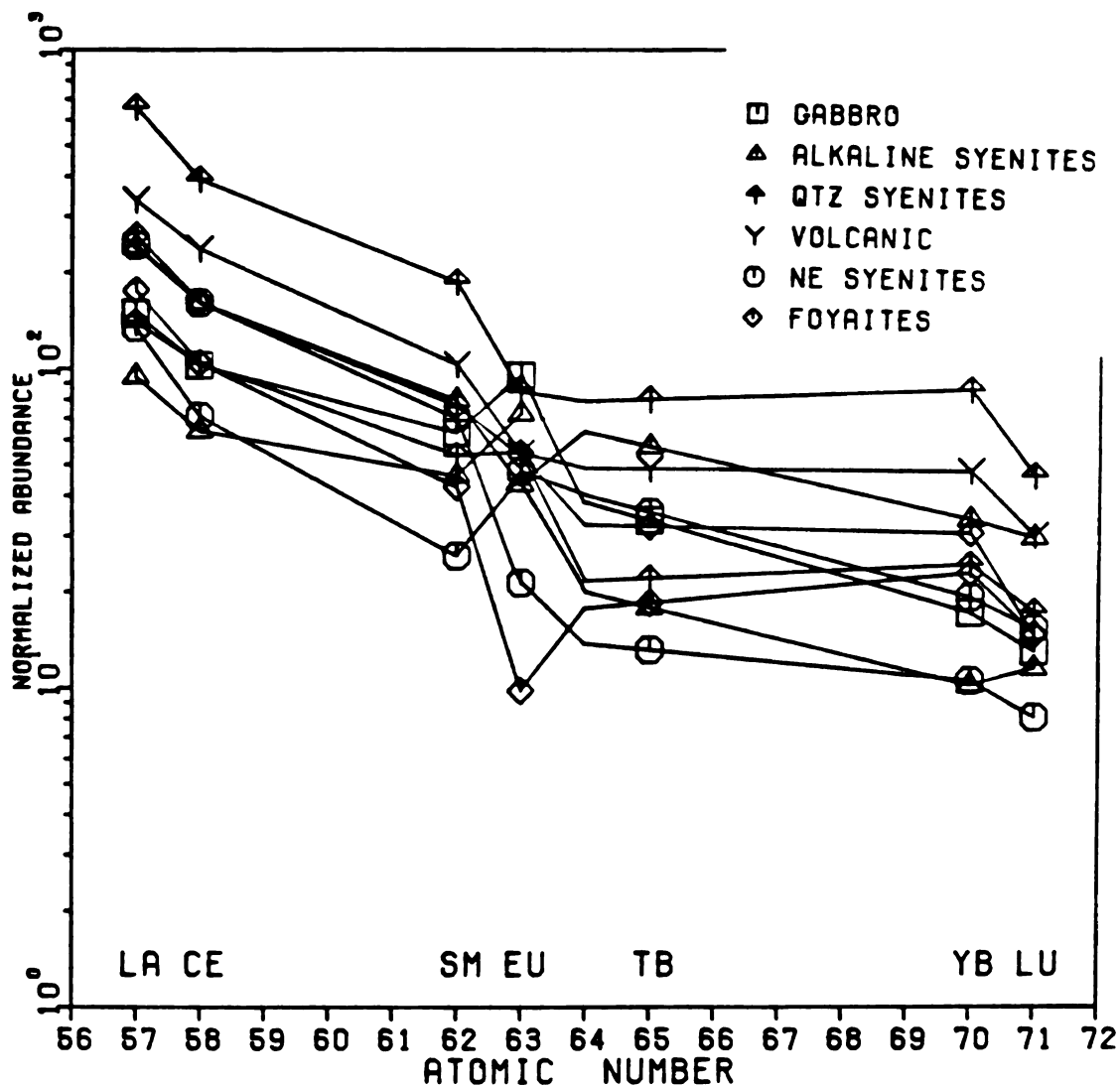


Figure 24.

Normalized REE plot of the major rock-types at G. Abu Khruq. Note the overall sub-parallel REE-enrichment pattern which supports a comagmatic relationship for this rock suite. In view of the mineralogy of these rocks (predominantly feldspars and clinopyroxenes), and the lack of minerals which cause large REE fractionations, this sub-parallel pattern is consistent with a fractional crystallization relationship rather than one involving partial melting.

Figure 25.

Ce/Sm versus Ce diagrams for some of the Egyptian alkaline complexes. These types of diagrams represent REE-fractionation versus REE-enrichment. Arrows are intended to show the general directions of enrichment/depletion via the fractionation of the dominant mineral phases.

- a) G. El Naga: this pattern is similar to that in which amphibole is an important crystal fractionation phase. The arrow is arbitrarily drawn to illustrate the effect of fractionating amphibole as a single phase.

Where $D_{Ce} = 1.44$ and $D_{Sm} = 3.21$ *

- b) G. El Kahfa: this pattern is typical for the fractionation of clinopyroxene and plagioclase. The arrow is arbitrarily drawn to illustrate the effect of fractionating 70% plagioclase and 30% clinopyroxene.

Where $D_{Ce} = 0.27$ and $D_{Sm} = 0.13$ for plagioclase*,

and $D_{Ce} = 0.50$ and $D_{Sm} = 1.67$ for clinopyroxene*,

to give the following bulk distribution coefficients:

$\bar{D}_{Ce} = 0.34$ and $\bar{D}_{Sm} = 0.59$ for 70% plagioclase
and 30% clinopyroxene.

- c) G. Abu Khruq: this pattern, similar to G. El Kahfa, is typical for the fractionation of clinopyroxene and plagioclase. The arrow is arbitrarily drawn to illustrate the effect of fractionating 70% plagioclase and 30% clinopyroxene. The distribution coefficients are the same as those used for G. El Kahfa in (b).

* Distribution coefficients are taken from Sun and Hanson (1976) and Arth and Hanson (1976).

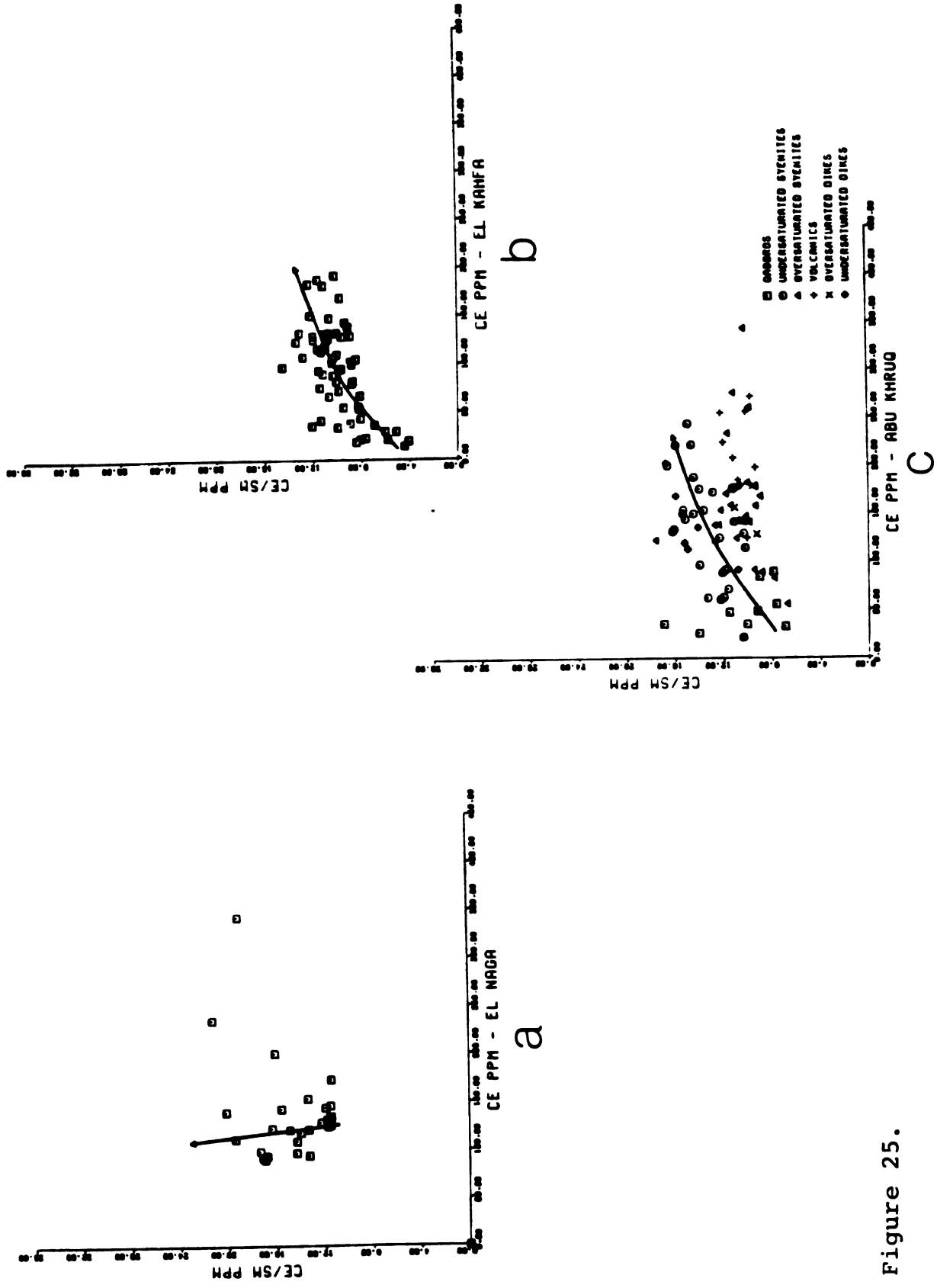


Figure 25.

El Kahfa, and El Naga are similar to those expected for the crystal fractionation of the observed dominant minerals; that of Nigrub El Fogani is difficult to evaluate due to the scattered nature of its predominantly altered rock-types. The Ce versus Ce/Sm patterns for Abu Khruq and El Kahfa are typical of clinopyroxene and plagioclase crystal fractionation; that of El Naga is characteristic of amphibole crystal fractionation (refer to Arth and Hanson, 1976; Tanaka and Nishizaga, 1975; Arth and Barker, 1976; Irving, 1978; Sun and Hanson, 1976).

C. The Relationship between Incompatible and Compatible Trace Elements -- Gabal Abu Khruq

Figure 26 shows Sr, a compatible trace element in feldspar-rich rocks, versus Rb, an incompatible trace element, for all the major rock-types from G. Abu Khruq (Rb-Sr data from Lutz, 1979). As can be seen by the superimposed fractional crystallization and partial melting curves, fractional crystallization can account for the sharp depletion of Sr, whereas the partial melting curve cannot. These curves have been drawn based on partition coefficients which appear to be reasonable for Abu Khruq, as well as being similar to average values often used in the literature (refer to Allégre et al., 1977; Frey et al., 1978; Baker et al., 1977). Reasonable changes in the relative bulk partition coefficients will not alter these relationships.

Figure 26.

Sr versus Rb for the rock series at G. Abu Khruq. The rapid depletion of Sr, a compatible trace element, cannot be explained by a partial melting relationship. As can be seen by the superimposed partial melting and fractional crystallization curves, the data is more closely approximated by the fractional crystallization relationship. Data from Lutz (1979).

- a) This diagram shows only the syenites and volcanics from G. Abu Khruq. The gabbros are excluded because they are considered to have a cumulate origin. Calculated partial melting and fractional crystallization curves are based on:

$\bar{D}_{\text{Sr}} = 3.0$ (a compatible trace element for alkali feldspar and plagioclase-rich rock suites).

$\bar{D}_{\text{Rb}} = 0.1$ (a known strongly incompatible trace element).

- b) This diagram includes the gabbros. If the gabbros did not have a cumulative origin, the Sr depletion is more pronounced, and the partial melting relationship deviates from the data trend even more than in diagram (a).

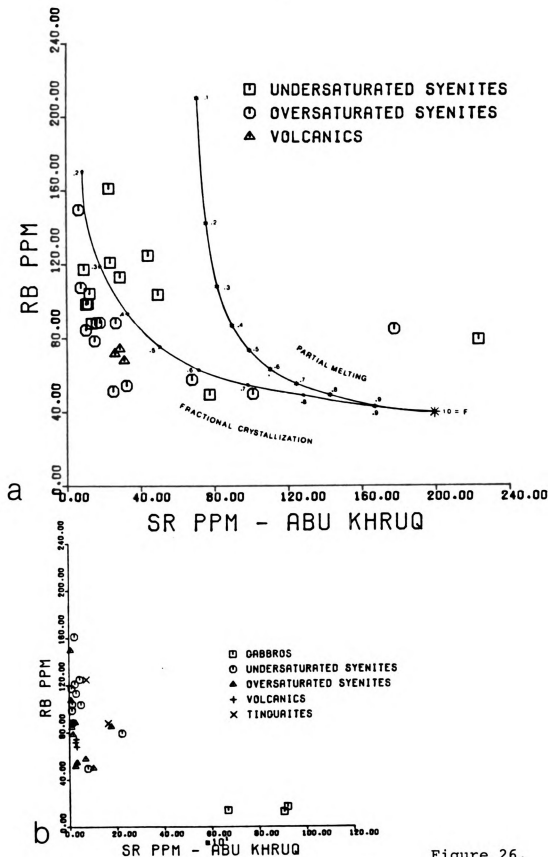


Figure 26.

D. Crystal Fractionation Modelling (See Appendix II for a more detailed model.)

Detailed trial-and-error crystal fractionation modelling in terms of the compatible and incompatible trace elements was attempted for the data from Abu Khruq. Minerals used in these fractionation models included olivine, clinopyroxene, plagioclase, Fe-oxides, alkali feldspar, apatite, and nepheline (nepheline appears to have partition coefficients similar to alkali feldspar, Eby, 1975). The calculations were performed on both calculated relative bulk partition coefficients (after the method outlined by Allegre et al., 1977) and published partition coefficients. The choice of the parent magma's composition was approximated by noting the chemical characteristics of the Egyptian suites and comparing them to many other similar alkaline suites which contain more mafic end-members. Although the use of 5 compatible trace elements (Ni, Cr, Co, Sr and Ba) and 4 incompatible trace elements (Ce, Sm, Lu and Th) served to constrain the models, the lack of an exposed mafic sequence of rocks at Abu Khruq did not allow a unique, rigorous solution. Consistently however, these models yielded very large amounts of crystal fractionation: about 2/3s of a moderately alkaline basaltic parent (similar to the average Hawaiian alkaline basalt) needed to be crystallized in order to produce the gabbros as cumulates, and greater than 90% of the original parent needed to crystallize in order to produce the bulk compositions of the syenites and volcanics.

Table 3. First Approximation of Fractional Crystallization
Based on Incompatible Trace Elements
(See Appendix I for a more detailed presentation).

Element	C_o	C_1	O_1^*	\bar{D}	F
Ce	44.00	184.97	188.27	0.46	0.07
Sm	5.53	16.38	16.50	0.57	0.08
Lu	0.23	0.91	0.94	0.51	0.06
Th	2.90	18.52	18.04	0.23	0.09

C_o = concentration of element in postulated alkaline basaltic parent magma.

C_1 = concentration of element in calculated remaining liquid after crystal fractionation.

O_1 = concentration of element in the observed rocks.

\bar{D} = bulk distribution coefficient for respective elements. Derived by Allegre et al.'s (1977) method which is explained in Chapter 4, Section III, 4. They compare closely to Allegre et al.'s (1977) coefficients in another alkaline suite.

F = fraction of original magma remaining after crystal fractionation and which contains the concentrations of the elements listed in C_1 .

* the observed concentrations are based on an average of 40 analyses.

Alkaline Basaltic Parent Magma

crystal  fractionation

Bulk Salic Composition (Average
of Foyaites, Quartz Syenites and
Volcanics).

The incompatible trace elements are not as affected by the early stages of crystallization as the compatible trace elements. Ignoring the early enrichment of the incompatible trace elements during the initial mafic stages of crystallization simply results in a more conservative fractionation model for only the incompatible trace elements. By comparing the compositional trends of the moderately alkaline, silica-undersaturated rock series at Abu Khruq to those of similar alkaline suites with more mafic members, a moderately alkaline basaltic parent melt similar to an average Hawaiian alkaline basalt was chosen (Schilling and Winchester, 1969; Kay and Gast, 1973; Frey et al., 1978). The bulk partition coefficients for the REEs were calculated via the method outlined by Allegre et al. (1977) (see pp.123-24). The bulk partition coefficient for Th was derived by taking an average of the values used by Baker et al. (1977) for the early mafic and late salic stages of differentiation in the Gregory Rift alkaline volcanics. The "Observed" concentrations of the derived liquid are a result of averaging 40 analyses of syenites and volcanics (an estimated bulk salic composition). The derived F values in this scheme all indicate that greater than 90% of the original alkaline basaltic parent must crystallize before the estimated bulk salic composition is attained.

Crystal fractionation models of other salic alkaline complexes have resulted in similar extreme amounts of crystal fractionation. Engell (1973), for example,

demonstrated that if fractional crystallization accounts for the evolved rock-types of the Ilimaussaq alkaline complex in Greenland, about 75-95% of the original augite syenitic magma must have crystallized before the last sodalite-rich foyaite stage was reached. This is the same rock-type which is thought to have ended the magmatic activity at Abu Khruq. If this amount must be fractionated from an augite syenitic magma to produce such rock-types, then even a greater proportion must be separated from an alkaline basaltic parent magma.

The relatively small volume of the Egyptian alkaline complexes does not create as great a problem in explaining reasonable amounts of the original basaltic magma undergoing fractional crystallization as in the more voluminous alkaline occurrences such as the East African Rift volcanics (see Baker et al., 1977; Baker and Henage, 1977). If the volume of the evolved rocks at Abu Khruq (syenites and volcanics) is approximated by a cone with the maximum diameter of Abu Khruq (8 km) and an estimated thickness of 1.5 km (based on a reconstruction by El Ramly Et al., 1969a), and if these evolved rocks represent the final 10% of the original alkaline basaltic parent melt, then about 754 cubic kilometers of alkaline basaltic magma is needed for the initial parent melt volume. This would mean that a simplified cylindrical magma chamber about twice the diameter of the present complex (16 km) would have to be about 3.75 km in height. The magnitude of these dimensions are

considered small enough to allow the possibility of in situ fractionation of a primitive alkaline basaltic magma at shallow crustal depths below the complex.

Section III

The Silica-Undersaturated/Oversaturated Problem

Introduction

One of the major problems in petrology involves rock associations which contain both silica-undersaturated and silica-oversaturated rock-types. The critical plane of silica-saturation in the basaltic tetrahedron (Yoder and Tilley, 1962) appears to effectively control the direction of fractionation so that the evolution of the liquid is strongly dependent on the initial composition of the parent (see Coombs, 1963). This prevents the transition of a basaltic magma from silica-undersaturated to silica-oversaturated compositions, and vice versa, via low-pressure crystal fractionation. This feature can be observed in the Q-Ne-Ks system of salic rock compositions in which the alkali feldspar join forms a thermal barrier at low pressures so that quartz-normative and nepheline-normative liquids are prevented from passing from one side to the other via crystal fractionation (Bowen, 1937). This problem is pertinent to the development of many of the alkaline complexes in the Eastern Desert of Egypt, as well as many other alkaline suites throughout the world.

1) The Common Occurrence of Silica-Undersaturated/Over-saturated Associations

There are many examples of silica-undersaturated/over-saturated alkaline suites of different ages and different crustal settings. Similar to the overall worldwide chemical regularity of many alkaline suites (e.g. Schwarzer and Rogers, 1974), silica-undersaturated/oversaturated associations tend to form regular trends in terms of both major elements and trace elements. Figure 27 shows a summary of the silica-alkali trends based on this study, as well as work by Schwarzer and Rogers (1974), Schwarzer (personal communication, 1979), Upton (1974), Miyashiro (1978) and others. The accompanying diagram from Upton (1974) shows a similar regular silica-undersaturated/oversaturated trend in the Gardar alkaline province of Greenland. The course of these trends appears to be strongly dependent on the initial composition of the parent magma (see Coombs, 1963). There are three fundamental trends; the Egyptian alkaline complexes sampled by this study fall into Trend 2 types.

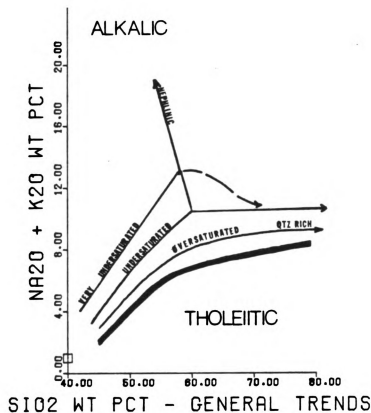
Trend 1: The first trend shows the differentiation pattern in silica-oversaturated alkaline suites in which a silica-oversaturated or hypersthene-normative mafic parent gives rise to slightly alkaline rock-types, ending with highly-evolved quartz-rich differentiates.

Trend 2: The second trend consists of a slightly silica-undersaturated parent giving rise to more alkaline rock-types than Trend 1, and in many suites of this type,

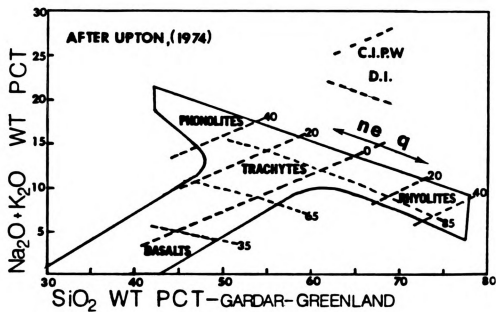
Figure 27.

Summary of silica-alkali trends.

- a) Summary of the silica-alkali trends noted for the Egyptian alkaline complexes and alkaline suites in other areas. See text for explanation.
- b) A compilation of the silica-alkali trends noted in the Gardar alkaline province of Greenland (after Upton, 1974). This diagram demonstrates a similar regular occurrence of silica-under-saturated/oversaturated trends.



a



b

Figure 27.

the differentiates are both silica-undersaturated and silica-oversaturated (see Miyashiro's (1978) "straddle type").

Trend 3: The third trend begins with a highly silica-undersaturated parent and leads to the development of very silica-undersaturated, highly alkaline differentiates, and in some cases, some silica-oversaturated rock-types.

This study noted that there is widespread occurrence of alkaline suites with an initially silica-undersaturated parent giving rise to both silica-undersaturated and oversaturated differentiates. However, similar unambiguous examples of an initially silica-oversaturated mafic parent giving rise to both varieties was not noted in this study. In some cases of silica-undersaturated/oversaturated associations, the most mafic rock-type was silica-oversaturated, but these were fairly well-differentiated rocks and not basaltic representatives (e.g. Pankhurst et al., 1976). It is possible that these salic complexes are underlain by silica-undersaturated, mafic rock-types.

Figure 28 shows a summary of these silica-undersaturated/oversaturated trends noted in the Egyptian and other alkaline suites in terms of Ce versus Ce/Sm. The consistency of these silica-undersaturated and oversaturated trends in terms of the major elements, and even on the trace element level, is considered to reflect a regular mechanism whereby the silica-saturation transition is accomplished.

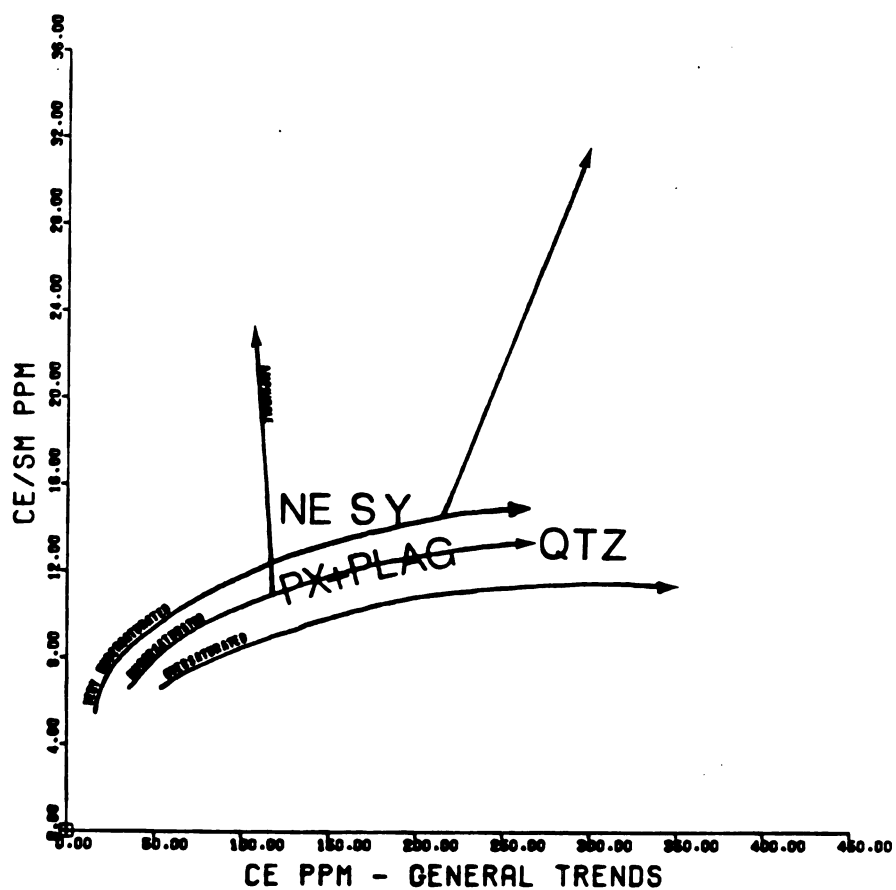


Figure 28.

Summary of the types of Ce/Sm versus Ce trends which this study noted to regularly occur in many alkaline suites throughout the world, including the Egyptian alkaline complexes.

The similarity between silica-undersaturated/oversaturated associations is particularly striking upon comparison of plutonic alkaline complexes, such as the Egyptian complexes of this study. There is a remarkable similarity between the Pajara alkaline ring complex on the island of Fuerteventura, Canary Islands, and Abu Khruq in Egypt. This similarity can be seen in the major and trace element chemistry, structure, mineralogy, and even the "secondary" alteration features (Muñoz, 1969; De Paepe et al., 1971; see also Fuster et al., 1968; Gastesi, 1969). The Pajara alkaline ring complex contains the same type of radial gradient in varying silica content, from nepheline syenites in the center, to quartz-rich syenites on the outer perimeters. Figure 29 shows the silica-alkali diagram for the syenitic and cumulate rock-types (gabbros and ultramafics) of the Pajara ring complex. Figure 30 demonstrates the similarity between the same rock-types from Abu Khruq and the alkaline rock-types of Fuerteventura, in terms of normalized REE plots.

The recently discovered alkaline complexes of the Velasco province in Brazil are very similar to the Egyptian alkaline complexes. They contain the same types of silica-undersaturated and oversaturated rocks, with a similar radial arrangement (see Darbyshire and Fletcher, 1979).

The alkaline province in southern Greenland is similar to the Egyptian province in general structural setting, as well as in the petrography and chemistry of its alkaline

Figure 29.

Silica-alkali diagram for the Pajara alkaline ring complex in the Canary Islands (Fuerteventura). This trend, as well as many features of this Canary Island complex, is very similar to the silica-alkali diagrams of the Egyptian alkaline complexes--G. Abu Khruq in particular (Figure 3, Chapter 3). Pajara data from Muñoz (1969) and Gastesi (1969).

Figure 30.

Normalized REE plot comparing similar rock-types from Fuerteventura, Canary Islands and G. Abu Khruq, Egypt. Note the close resemblance, even on an incompatible trace element level. Fuerteventura data from De Paepe et al. (1971).

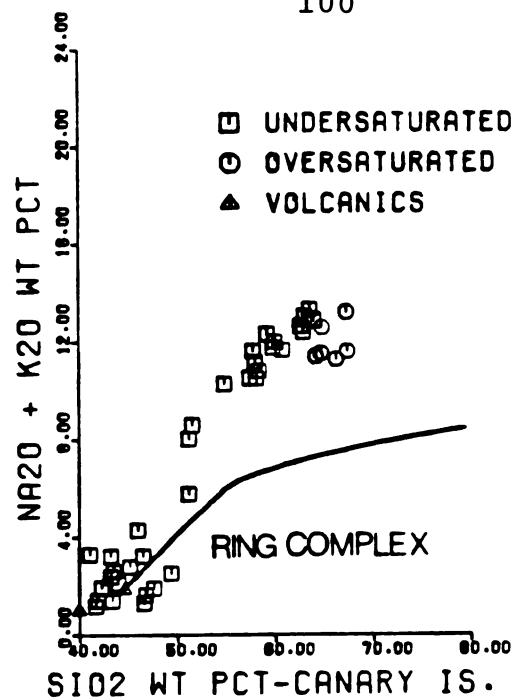


Figure 29.

ABU KHRUG & CANARY ISL. REE COMPARISON

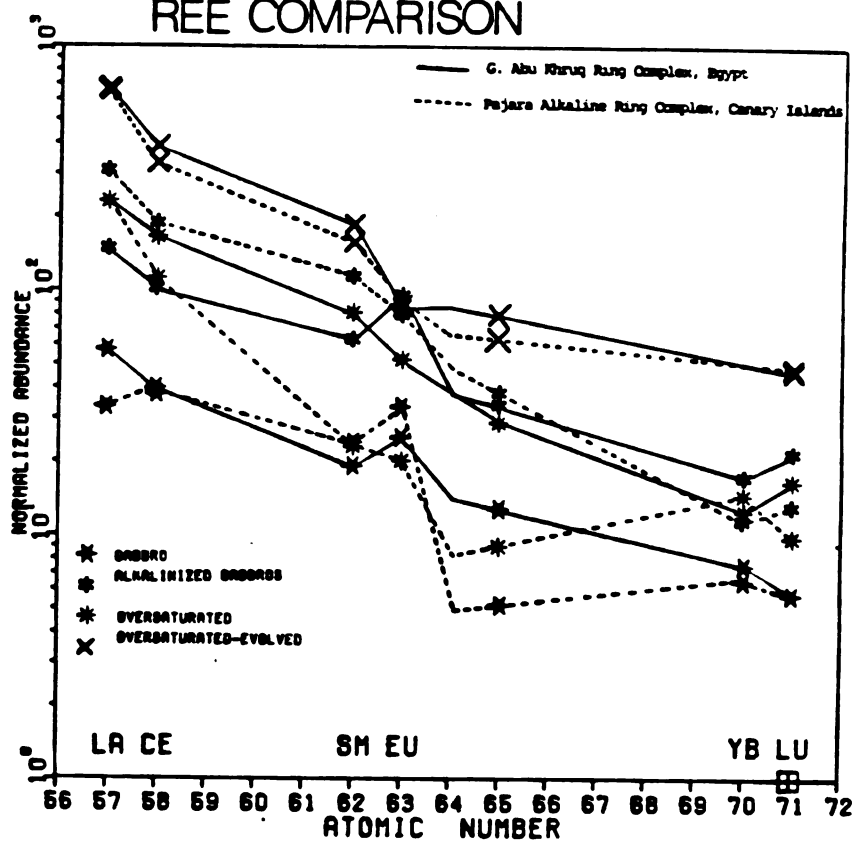


Figure 30.

complexes (e.g. Upton, 1974). The Kangerdlugssuaq intrusion in this alkaline province may prove to be very important for understanding the petrogenesis of these silica-undersaturated/oversaturated alkaline complexes. It consists of the same rock-types as Abu Khruq, but rather than having a conical faulted ring structure, it is an intact, largely unfaulted intrusion containing only gradational petrologic boundaries, and a continual gradient of SiO_2 , Na_2O and K_2O in a radial fashion (Pankhurst et al., 1976). Thus, without any sharp discontinuities, the inner silica-undersaturated foyaites grade into slightly silica-undersaturated syenites with up to 5% feldspathoids; these grade into slightly silica-oversaturated syenites with less than 2% modal quartz, and finally, these syenites give way to quartz syenites with up to 10-15% modal quartz. A similar radial gradient appears to exist at Abu Khruq and many other plutonic alkaline complexes, but in a disrupted, faulted pattern (see Lutz, 1979).

It is also important to point out that there are plutonic alkaline complexes with similar rock-types, but different apparent orders of crystallization. Tilley (1958) noted that alkaline complexes often are composed of silica-undersaturated centers surrounded by silica-oversaturated perimeters, or alternatively, silica-oversaturated centers surrounded by silica-undersaturated perimeters. The relationship between the syenites and gabbros also appear to vary in apparent order of intrusion (e.g. Eby, 1979). These

different field relationships between very similar rock-types are puzzling in terms of a regular petrogenetic model. This study considers that any proposed model on the differentiation of one type of intrusion must also allow for these other variations in apparent order of emplacement and crystallization.

2) The Silica-Undersaturated/Oversaturated Problem in the Egyptian Alkaline Complexes

As was shown in the previous section, there is a considerable amount of data which may be interpreted to indicate that crystal fractionation was an important process in the evolution of the Egyptian alkaline complexes. However, the relationships between the salic rock-types present four main observations which cannot be explained by simple crystal fractionation:

1. the field relationships of the silica-undersaturated and oversaturated rock-types,
2. the REE-depleted and REE-enriched stages of development in the silica-undersaturated and oversaturated rock-types, respectively,
3. several chemical trends which are consistent with the simultaneous low-pressure development of the silica-undersaturated and oversaturated rock series,
4. the incompatible trace element relationships which indicate that the REE in the silica-oversaturated series behaved as though they were more mobile than the REE in

the silica-undersaturated series, and further, that the HREE in both series have been partitioned as though they were unusually mobile relative to the other REE.

Each of these observations deviate from that expected for crystal fractionation, and is considered to be related to the comagmatic development of the silica-undersaturated and oversaturated rock-types. The mechanism of this development must be compatible with the data supporting crystal fractionation. The regular occurrence of these types of alkaline associations in the Eastern Desert of Egypt, as well as many other parts of the world, is considered to be indicative of a fairly consistent process leading to the evolution of silica-undersaturated/oversaturated alkaline associations. This fairly consistent process is considered to involve a thermogravitational diffusion process and its attendant fluid effects. The following pages will review the nature of the silica-undersaturated/oversaturated problem in the Egyptian alkaline complexes, and the mechanisms which have been proposed in the literature to explain silica-undersaturated/oversaturated associations.

A. Major Element Trends in Relation to the Silica-Undersaturated/Oversaturated Problem

Figure 31 displays the silica-undersaturated/oversaturated problem in terms of normative nepheline and quartz with respect to MgO for the rock series at Abu Khruq. This shows that both the silica-undersaturated and oversaturated

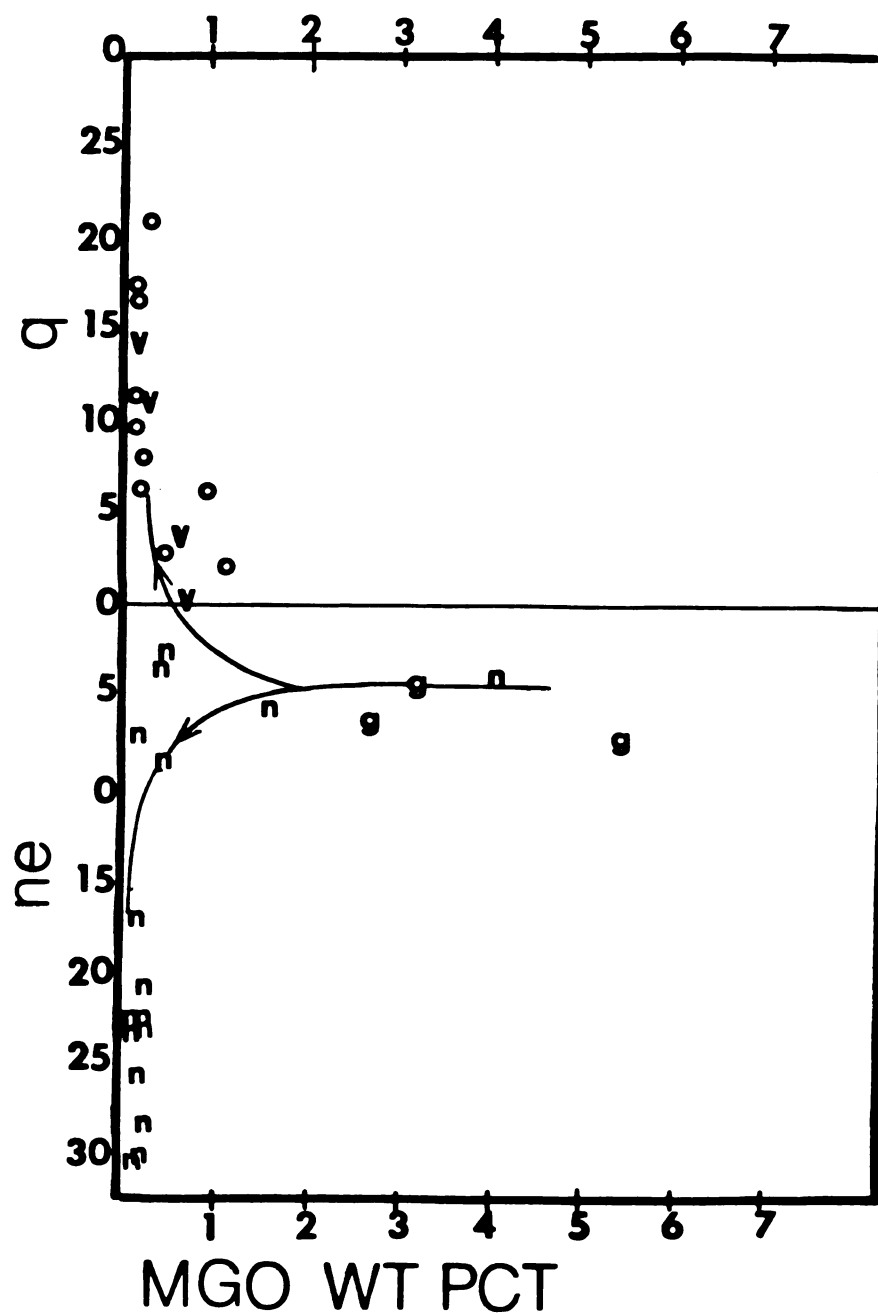


Figure 31.

Normative ne and qtz versus MgO (index of crystallinity) for the rock series at G. Abu Khruq. Note the progressively diverging salic trends of silica-undersaturation and oversaturation from the silica-undersaturated mafic rock-types. g=gabbro; n=silica-undersaturated syenites; o=silica-oversaturated syenites; v=volcanics.

trends converge toward the more mafic silica-undersaturated compositions. The silica-undersaturated rock-types cover the entire range of MgO concentrations; the silica-oversaturated rock-types cover only the less mafic compositions. The most mafic syenites of both silica-saturation levels were collected from apparently continuous rock bodies.

The major element variation diagrams in Chapter 3 demonstrated the highly-evolved, salic nature of the majority of the rocks at Abu Khruq. Although more samples of mafic syenites need to be analyzed for major elements, some general conclusions may be derived from these major element trends.

The CaO, TiO_2 , and P_2O_5 variations are consistent with fractional crystallization of a mafic parent melt. These element oxides are included in the early crystal phases: calcic plagioclase, calcic pyroxene, titanium oxides, and apatite. Due to the predominance of salic compositions in the exposed rock-types at Abu Khruq, these element oxides are not expected to show much variation. The SiO_2 , Al_2O_3 , Na_2O , K_2O and $\text{FeO}_{\text{total}}$ variations cannot be interpreted as a result of simple crystal fractionation of a single magma. These element oxides all have bifurcating trends corresponding to the silica-undersaturated and oversaturated series, similar to the ne/qtz versus MgO diagram of Figure 31.

B. Incompatible Trace Element Trends in Relation to the Silica-Undersaturated/Oversaturated Problem

Figure 32 clearly illustrates the separation of the silica-undersaturated and oversaturated trends at Abu Khruq. It becomes apparent that the silica-undersaturated and oversaturated trends begin from similar concentration ratios and progressively diverge along separate directions. This divergence becomes more noticeable for the HREE. An indication of the initial similarity between the less-evolved silica-undersaturated and oversaturated rock-types may be seen in the normalized REE plots of Figure 33. The more evolved rock-types, however, gain increasingly greater difference in terms of silica-saturation levels and REE concentrations. For a given thorium concentration, the silica-undersaturated rocks are depleted in REE, and especially HREE, relative to the silica-oversaturated rocks. The silica-undersaturated rocks, however, attain similar high Th concentrations, and much higher REE fractionation, i.e. higher Ce/Sm and Ce/Lu ratios. The relative REE depletion in the silica-undersaturated series and the development of two separate REE trends cannot be explained by simple crystal fractionation.

C. The Apparent Order of Crystallization in Relation to the Silica-Undersaturated/Oversaturated Problem

The apparent order of emplacement and crystallization of the different rock-types in the Egyptian alkaline

Figure 32.

REE versus thorium diagrams for G. Abu Khruq. Note the progressive separation between the silica-undersaturated and oversaturated trends with increasing Th concentrations (index of crystallinity). The most Th-enriched silica-undersaturated rocks (foyaïtes) are REE-depleted relative to the most Th-enriched silica-oversaturated rocks (quartz syenites and volcanics).

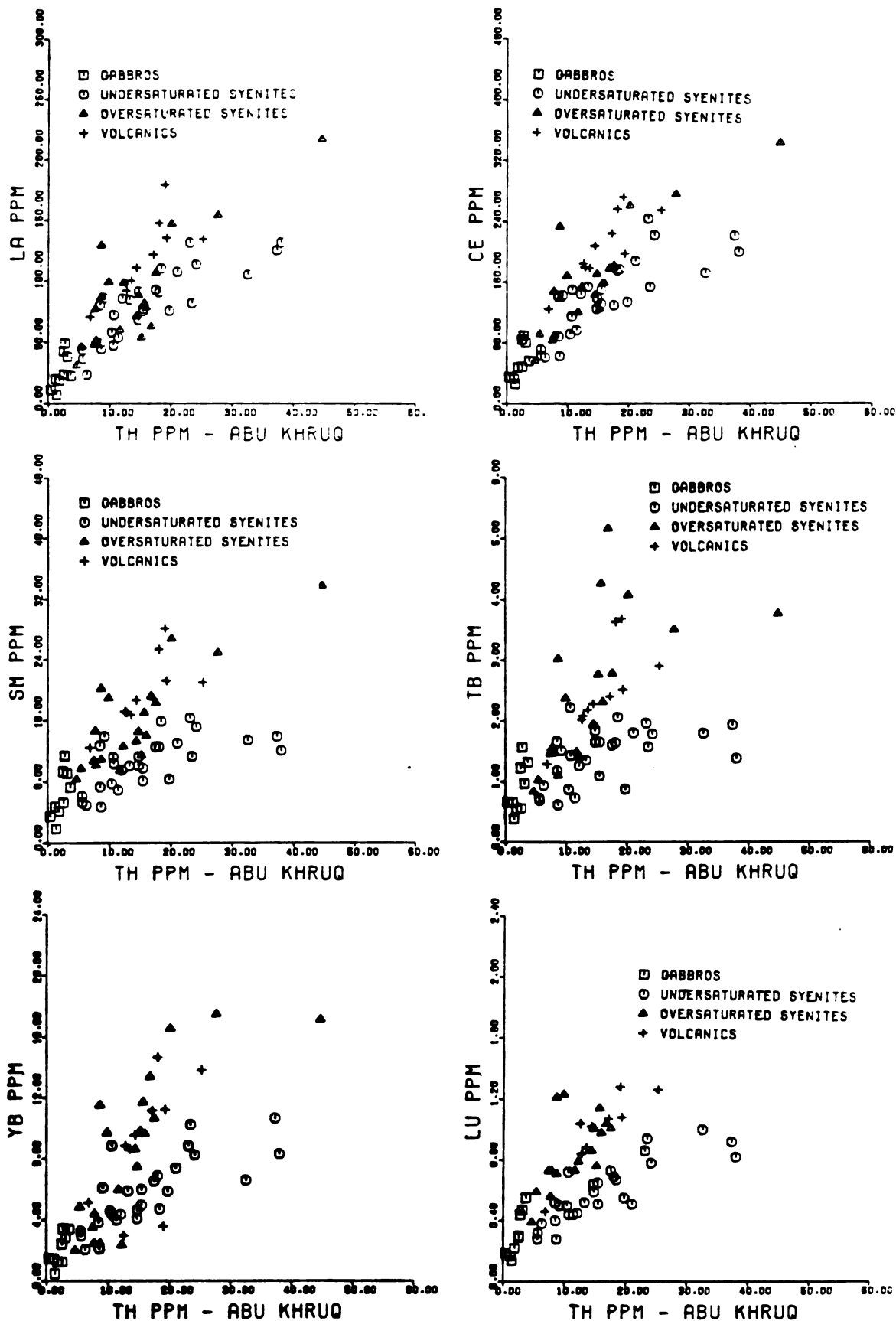


Figure 32.

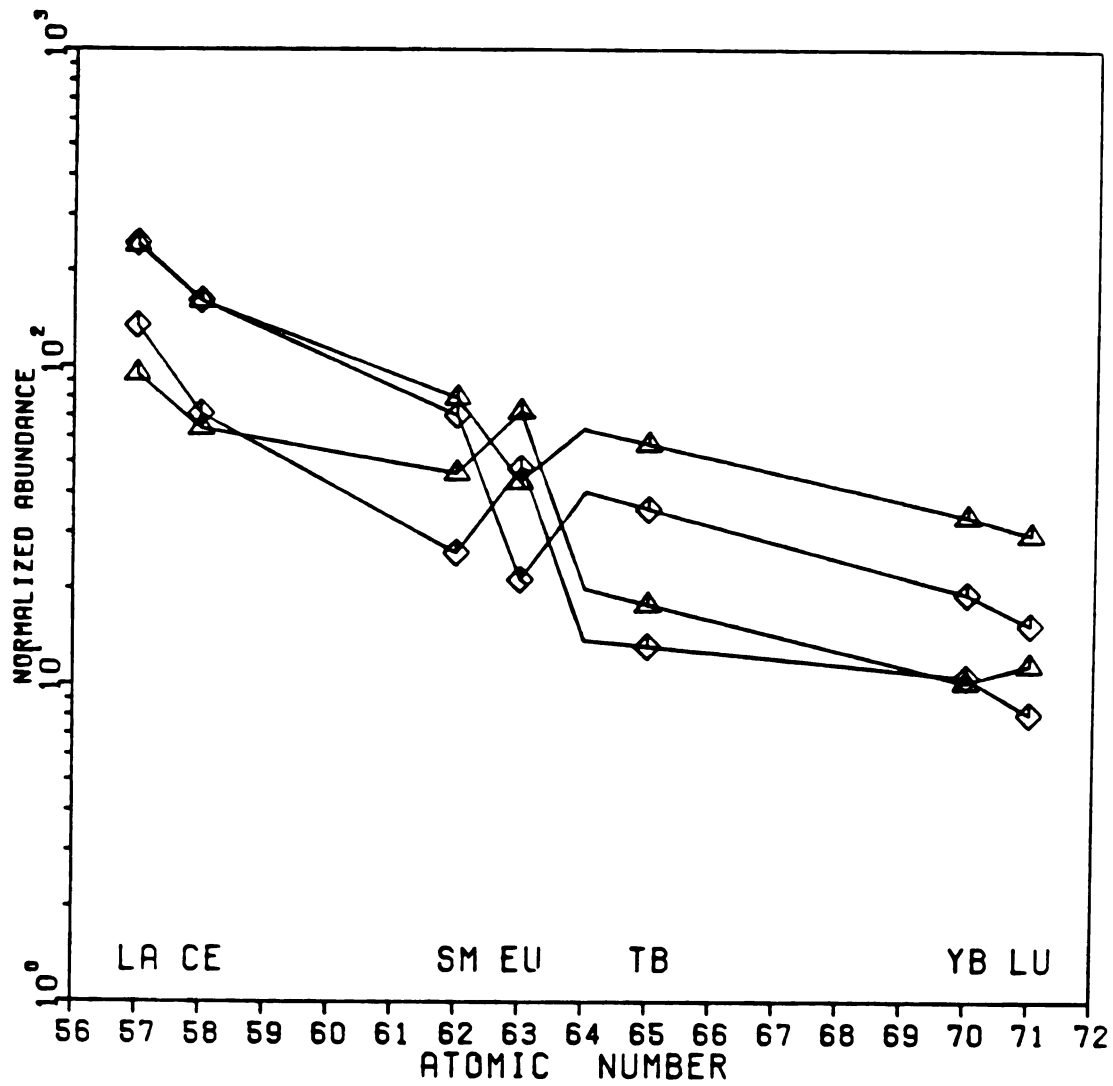


Figure 33.

Normalized REE plot showing the similarity between the more mafic, less incompatible-rich syenites of the silica-undersaturated and oversaturated rock series.

Diamonds = silica-undersaturated syenites

Triangles = silica-oversaturated syenites.

complexes is extremely difficult to incorporate into a petrogenetic model. For example, the order of intrusion/extrusion at G. Abu Khruq according to El Ramly et al. (1969a) was:

The volcanics (silica-oversaturated) initiated the magmatic activity, followed by the emplacement of the gabbros (silica-undersaturated), then the alkaline syenites (which included both silica-undersaturated and oversaturated varieties in El Ramly et al.'s rock classification), then a slightly later emplacement of the quartz syenites (silica-oversaturated), and finally ending with the intrusion of the foyaites (strongly silica-undersaturated). This apparent alternation of silica-saturation levels cannot be explained by a simple crystal fractionation model. Thus, either this interpretation of the field relations is incorrect, and/or a more complicated petrogenetic solution than simple crystal fractionation exists.

D. REE Depleted/Enriched Stages in the Development of Silica-Undersaturated/Oversaturated Complexes

In an igneous sequence derived by crystal fractionation, the initial stages of eruption are expected to be the least REE-enriched, and the final stages of crystallization are expected to be the most REE-enriched. Figure 34 shows the average Ce concentrations (representative of total REE) for the major rock-types at Abu Khruq in their apparent order of crystallization. As can be seen, the siliceous

Figure 34.

REE concentration levels (represented by Ce) in the major rock-types of G. Abu Khruq arranged according to their apparent orders of intrusion/extrusion. Similar to the pattern noted by Balashov (1972) in Figure 33, the last rocks to crystallize at G. Abu Khruq (foyaïtes) are relatively depleted in the REE with respect to the previous stages of crystallization. Similar to the enriched nature of peralkaline flows early in the magmatic history of an area, the REE-enriched volcanic cone formed early in the development of G. Abu Khruq. These depletion and enrichment levels with respect to the apparent order of intrusion/extrusion are not consistent with crystal fractionation.

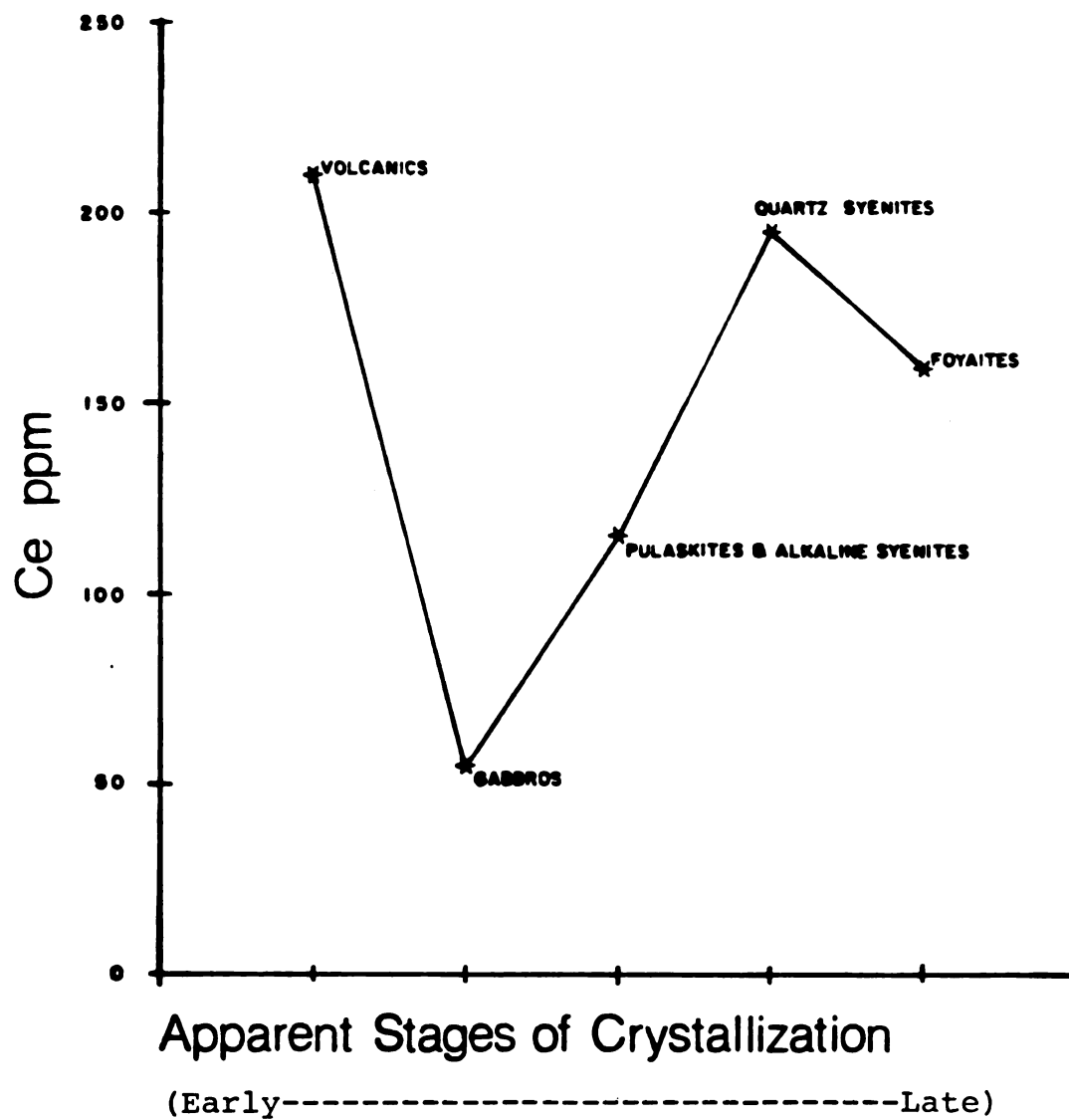


Figure 34.

volcanics represent the most REE-enriched rock-type at Abu Khruq in spite of their early formation in the development of the complex. The presence of such evolved, REE-enriched effusives before the crystallization of the more mafic rock-types lower in REE, cannot be explained by fractional crystallization. The development of such siliceous REE-enriched volcanic or ignimbrite flows early in the magmatic history of an area is not unique to Abu Khruq. This pattern has been frequently noted in peralkaline volcanic areas (Hildreth, 1981; see Bulletin Volcanologique, Special Issue - Peralkaline Rocks, v. 38, #3).

The foyaïtes at Abu Khruq present a similar problem in terms of the expected REE distribution for a melt evolving via crystal fractionation. Similar to their relationship in many alkaline complexes, these highly silica-undersaturated rocks form the last stage of crystallization. Despite their highly-evolved major element chemistry and their extreme depletion in compatible trace elements, they contain lower abundances of REE than the previously formed silica-oversaturated volcanics and quartz syenites, Figure 34. Balashov (1972) demonstrated that intrusive alkaline complexes typically end their magmatic activity with the formation of rock-types relatively depleted in REE with respect to the previous stages of crystallization, Figure 35. The regular occurrence of this depleted REE pattern in the final stages of crystallization of alkaline intrusions demands attention. According to the observations of this

Figure 35.

REE depletion in the last phase of intrusive activity which Balashov (1972) found to typically occur in many alkaline complexes.

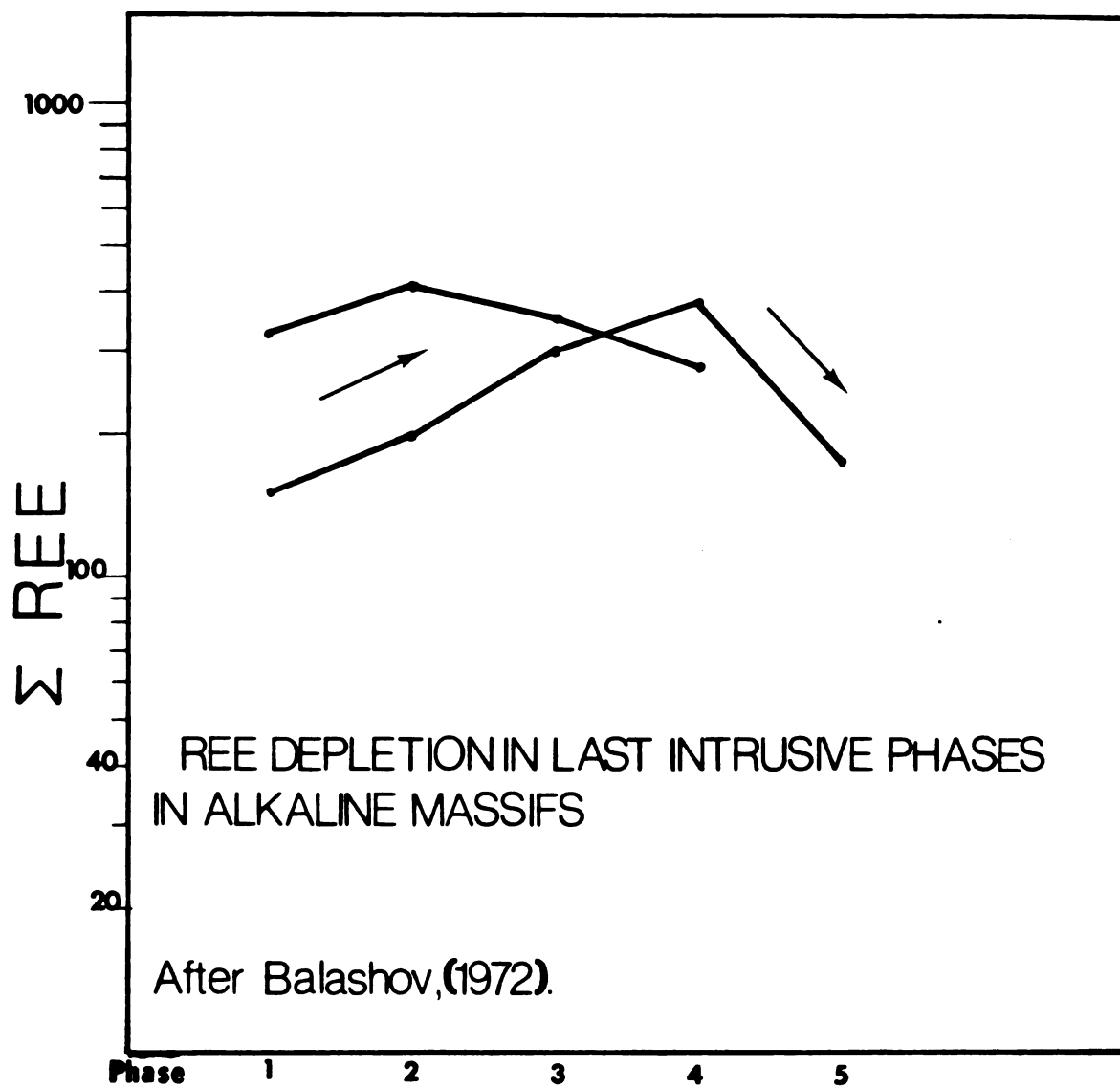


Figure 35.

study on silica-undersaturated/oversaturated suites, there appears to be a consistent enriched versus depleted REE relationship between the silica-oversaturated and undersaturated sequences, respectively. These relationships must be taken into account in any petrogenetic model on the evolution of these types of alkaline melts.

3) Silica-Undersaturated/Oversaturated Models

There are several models which have been proposed to account for the development of silica-undersaturated and oversaturated rock-types in close proximity. Four of these models are rejected for the Egyptian alkaline complexes based on the primitive isotopic data (Lutz, 1979), the radiometric age-dating of the rock-types within individual complexes (Serencsits et al., 1981; Lutz, 1979), the nature of the chemical trends, and the magnitude of the change required in the silica-saturation levels for the Egyptian alkaline complexes.

1. Crustal contamination models.
2. Resorption mechanisms (see Luth, 1976; MacDonald, 1974; Edgar, 1974; Gittins, 1979).
3. Fractionation of silica-deficient phases (see Oftedahl, 1948; Muir and Smith, 1956; MacDonald, 1974)
4. Liquid immiscibility (see discussion in Chapter 2).

There are three remaining possibilities for the origin of the types of silica-undersaturated/oversaturated complexes

observed in Egypt. Two of these involve high pressure fractionation, and the other invokes thermogravitational diffusion and fluid effects. High pressure fractionation is discounted; thermogravitational diffusion and attendant fluid effects are considered to be responsible for the development of a silica-oversaturated rock series from a silica-undersaturated magma.

5. High pressure fractionation

Experimental work has demonstrated that at high pressures, the thermal barrier between silica-deficient and silica-excess systems is lowered (Schairer and Bowen, 1935; Tuttle and Bowen, 1958). This allows for two suggestions on the origin of silica-undersaturated/oversaturated associations which involve high pressure fractionation.

a. The first possibility involves the coincidence of a silica-undersaturated magma and a silica-oversaturated magma in the same localized area, either commingling or in discrete episodes. This involves the emplacement of separate batches of melt from a common magma at very great depths, or separate batches of partial melt from a source region at very great depths. The regular occurrence of similar silica-undersaturated/oversaturated alkaline complexes in Egypt as well as other areas, the general lack of sharp contacts between the major rock-types, and the nature of some of the chemical trends, all serve to discount this type of simultaneous two-magma model.

b. The second possibility involving high pressure fractionation makes use of high vapor pressures within shallow magma chambers in order to explain the transition of silica-saturation levels in many alkaline associations. The thermal barrier between silica-undersaturation and oversaturation may break down at pressures greater than 10 kilobars (Morse, 1969, 1970). As a result, some petrogenetic models on silica-undersaturated/oversaturated associations have implemented models of very high vapor pressures in the magma chambers (e.g. Pankhurst et al., 1976). Alkaline magmas tend to have high $\text{CO}_2/\text{H}_2\text{O}$ ratios (e.g. Mysen, 1976), low H_2O solubilities (Kogarko, Burnham and Shettle, 1978), and large amounts of halogens (e.g. Bailey and MacDonald, 1975; Baker, et al., 1977). The effect of these volatiles on silicate magma systems is uncertain, and more experimental work is needed in order to fully evaluate their influences in alkaline magmatic systems. However, it is unlikely that such high vapor pressures may be attained in shallow crustal magma chambers. If vapor pressures build up enough to lower the thermal barrier at shallow depths, the vapor separation and circulation cannot lower the temperature of the magma very much or its effect on lowering the thermal barrier will be counteracted (see Pankhurst et al., 1976). More importantly, even if the thermal barrier can be lowered sufficiently, a chemical mechanism is still needed to cause the magma to change its composition toward the new silica-saturation levels.

Figure 36 supports plagioclase fractionation which can occur at pressures equivalent to depths above about 25-30 kilometers (plagioclase stability limit, Green and Ringwood, 1967) with a rapid decrease in oxygen fugacity (close to the QFM buffer, Weill and Drake, 1973). (See Thorpe et al., 1977; Baker et al., 1977; and Kay, 1978 for interpretation of these types of diagrams.) Thus, these trends are consistent with low-pressure fractionation for the rock series at Abu Khruq.

Experimental data indicates that the agpaitic order of crystallization (the dominant felsic minerals crystallize earlier than the dominant ferromagnesian minerals) which occurs in the Egyptian and other similar alkaline complexes, cannot occur at P_{H_2O} greater than 100-200 atmospheres (about 0.2-0.3 kilobars) (Kogarko, Burnham and Shettle, 1978).

The field relationships including gradational, unreacted petrologic boundaries, the miarolitic textures in all the rock-types, and the shallow nature of the collapsed sub-volcanic ring structures (e.g. Bahat, 1980), all support low-pressure differentiation.

The evidence against high pressure fractionation and the fact that low pressure crystal fractionation cannot account for the development of both silica-undersaturated and oversaturated trends, leads to the conclusion that another process other than fractional crystallization must

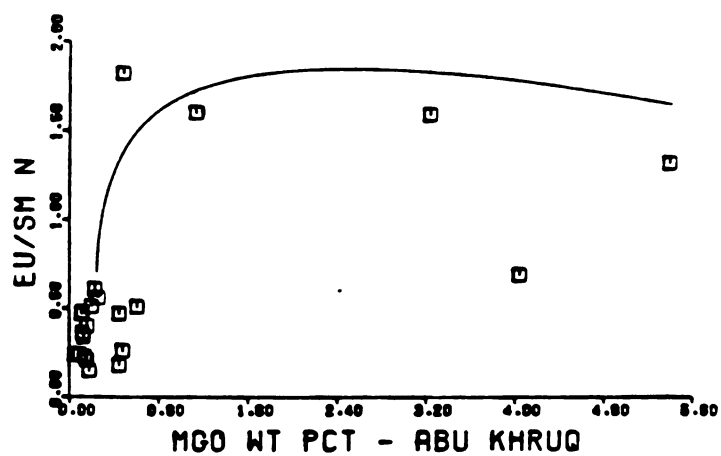
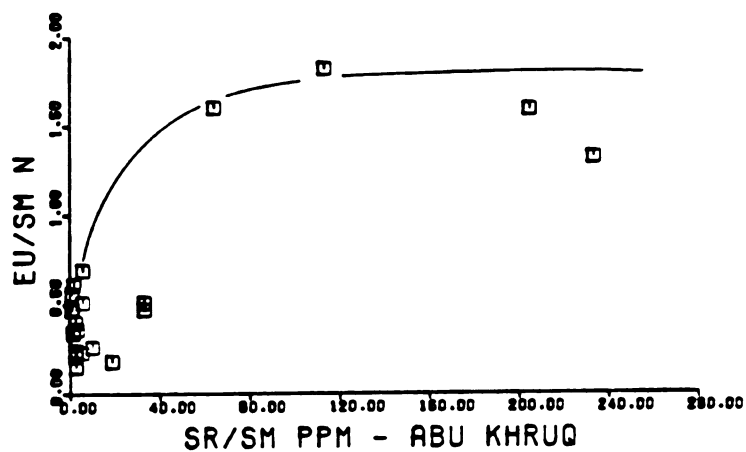


Figure 36.

Trace element and MgO trends which are considered to indicate low-pressure crystal fractionation (see Thorpe et al., 1977; Baker et al., 1977a, Kay, 1978). The diagrams are based on the plagioclase stability limit and the affinity which these trace elements display for plagioclase.

be responsible for the origin of these silica-undersaturated/oversaturated complexes.

6. Thermogravitational diffusion and attendant fluid effects.

This is the final possibility which is considered for the origin of silica-undersaturated/oversaturated complexes such as those in Egypt. Several mechanisms have been suggested to be responsible for the development of liquid fractionation in alkaline magmas. Desilication of an oversaturated magma via volatile transfer was suggested by Currie (1970) in order to explain the silica-transition of some of the alkaline complexes in Canada. Highly differentiated, halogen-rich alkaline melt systems of long-term crustal residence appear to provide the ideal conditions for a fairly stable thermogravitational process similar to those suggested for peralkaline, siliceous effusive rock suites (Hildreth, 1977, 1979, 1981; Saggerson, 1970). There is evidence that excess silica and other components will tend to accumulate in the roof and outer portions of certain types of volatile-rich magma chambers (e.g. Hildreth, 1977, 1979, 1981). Although structural and magmatic events may disturb the radial chemical gradients which are set up by convective circulation, Soret diffusion, and the movement of discrete vapor phases through the magma, it is possible that near-equilibrium mass transfer of certain components in these alkaline systems may occur along similar compositional trends as those directed by the overall

differentiation of the magmatic system. More specifically, this may involve the gradual change from solid-liquid equilibria to a gathering influence of solid-liquid-vapor equilibria. This approach to the silica-undersaturation/oversaturation problem is considered to be the most consistent with the available data on the Egyptian alkaline complexes.

4) A Quantitative Approach to Assessing the Nature of the REE Distributions

Use of the REEs as petrogenetic indicators in evolved alkaline melts is of special concern because these systems contain large amounts of alkalis and volatiles which may have a strong effect on both the primary REE distribution in the crystallized phases, as well as the alteration of these distributions due to possible metasomatic effects from the increasing proportions of these components in the residual liquids (Flynn and Burnham, 1978; Mysen, 1976, 1977a, 1977b; Mysen et al., 1975; Kogarko, 1974, 1977; Kogarko et al., 1978; Balashov and Krigman, 1975; Wendlandt and Harrison, 1979; Wyllie, 1979; Hards, 1976).

A preliminary precaution on assessing the petrogenetic validity of the REE is to make note of any known REE-complexing minerals which formed early in the crystallization history of the rock suite. Two REE-rich minerals which are common in the Egyptian alkaline rocks are apatite and magnetite (see Schock, 1979). By comparing the P_2O_5 and

FeO_{total} concentrations to the REE distribution in these alkaline rock series, it was concluded that neither mineral exerted a significant control on the overall REE distributions. No other unusually REE-enriched minerals have been observed in the Egyptian alkaline complexes (see also Lutz, 1979, p. 84; Fryer and Edgar, 1977; Miller and Mittlefehldt, 1982).

As was presented in the previous pages, the REE distributions of the rocks at Abu Khruq appear to bear a relationship to the development of the silica-undersaturated and oversaturated series. If the REE reflect their primary magmatic distributions, it is possible that a detailed study of their patterns may yield evidence of the mechanism(s) which caused the development of the silica-undersaturated and oversaturated rock series. In the last few years, a quantitative approach has been developed in order to more critically evaluate the nature of the trace element distributions. This approach was initiated by Anderson and Greenland (1969), and then more fully developed by Treuil and Varet (1973), and then by Allégre et al. (1977), Minster et al. (1977), and Minster et al. (1978) in a three part series (see also Allégre and Minster, 1978; Hanson, 1978).

The result of this quantitative approach is to arrive at some calculated bulk partition coefficients for suites of trace elements in relation to a well-established "hygromagmatophile" or highly incompatible reference trace

element (see Treuil and Varet, 1973; Allégre et al., 1977; Wood et al., 1979). For an explanation of this method, the reader is referred to Allégre et al. (1977).

Thorium was chosen for the highly incompatible reference trace element. Several factors make Th an excellent reference trace element for the Egyptian alkaline complexes: it was analyzed with high precision (INAA), it very rarely has a preferred mineral phase, there is lack of large amounts of radioactive mineralization in the Egyptian alkaline complexes (Hussein and Hassan, 1973), and Th is one of the most immobile trace elements (Adam and Gasparini, 1970).

Figures 37, 38, 40 and 41 show the log Th-log REE plots involved in the method outlined by Allégre et al. (1977) for the four Egyptian alkaline complexes of this study. The relative bulk partition coefficients for the REE from these plots have been calculated only for Abu Khruq.

Gabal El Naga

The log Th-log REE plots for El Naga show an extremely well-defined trend, Figure 37. This complex contains the least amount of evidence for widespread autometasomatism or alteration of the four alkaline complexes sampled by this study. El Naga is the only complex of this study in which amphibole is the dominant ferromagnesian mineral. In spite of these differences, the general nature of its geochemistry is similar to the other Egyptian alkaline complexes. The differences in terms of alteration, ferromagnesian phases,

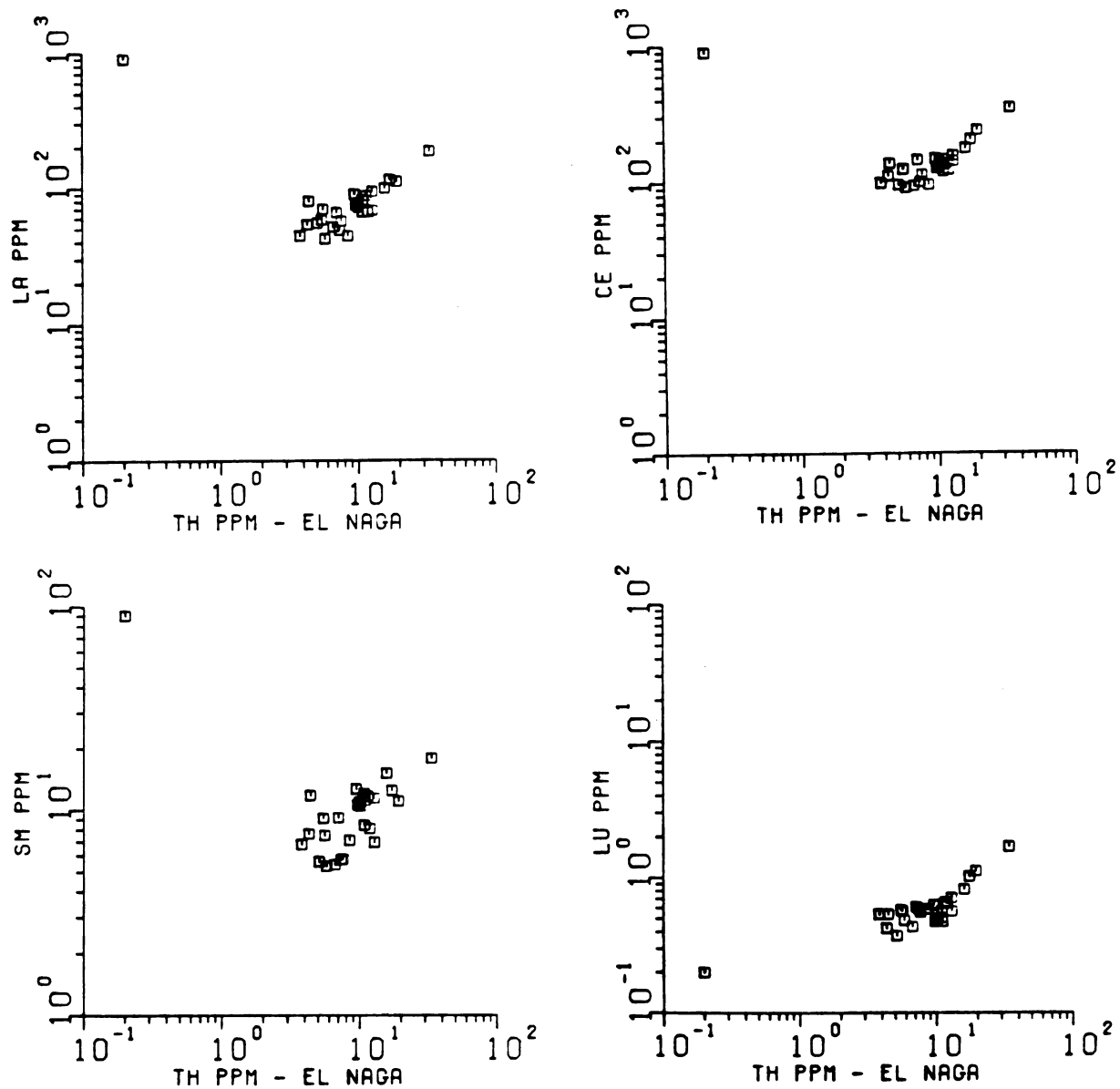


Figure 37.

Log REE versus Log thorium for G. El Naga.

and the well-defined REE trends may all be explained as the result of the retention of the volatile phases in the magma at El Naga. P_{vapor} is thought to be the dominant control on whether amphibole or pyroxene lies on the liquidus (Ernst, 1968; Ferguson, 1978a, 1978b; Stephenson, 1972; Larson, 1976). The retention of the volatile phases in the magma at El Naga would have caused the crystallization of amphibole rather than pyroxene, and would have avoided the alteration caused by otherwise migrating fluids.

Gabal Nigrub El Fogani

At the present level of exposure, Nigrub El Fogani contains predominantly high-evolved, altered rock-types, including carbonatite dikes. The presence of carbonatites at this complex attests to the importance of CO_2 -rich phases in the development of this alkaline melt. Experimental work indicates that CO_2 and carbonate complexing of the REE may have a significant effect on the final REE distributions of rocks crystallizing in CO_2 -rich systems (Wendlandt and Harrison, 1979; Balashov and Krigman, 1975). Large-scale, erratic effects on the REE distributions of other carbonatite-rich alkaline complexes have been attributed to this CO_2 -complexing (e.g. Eby, 1975). A similar effect may be reflected in the scattered REE trends of Nigrub El Fogani, Figure 38.

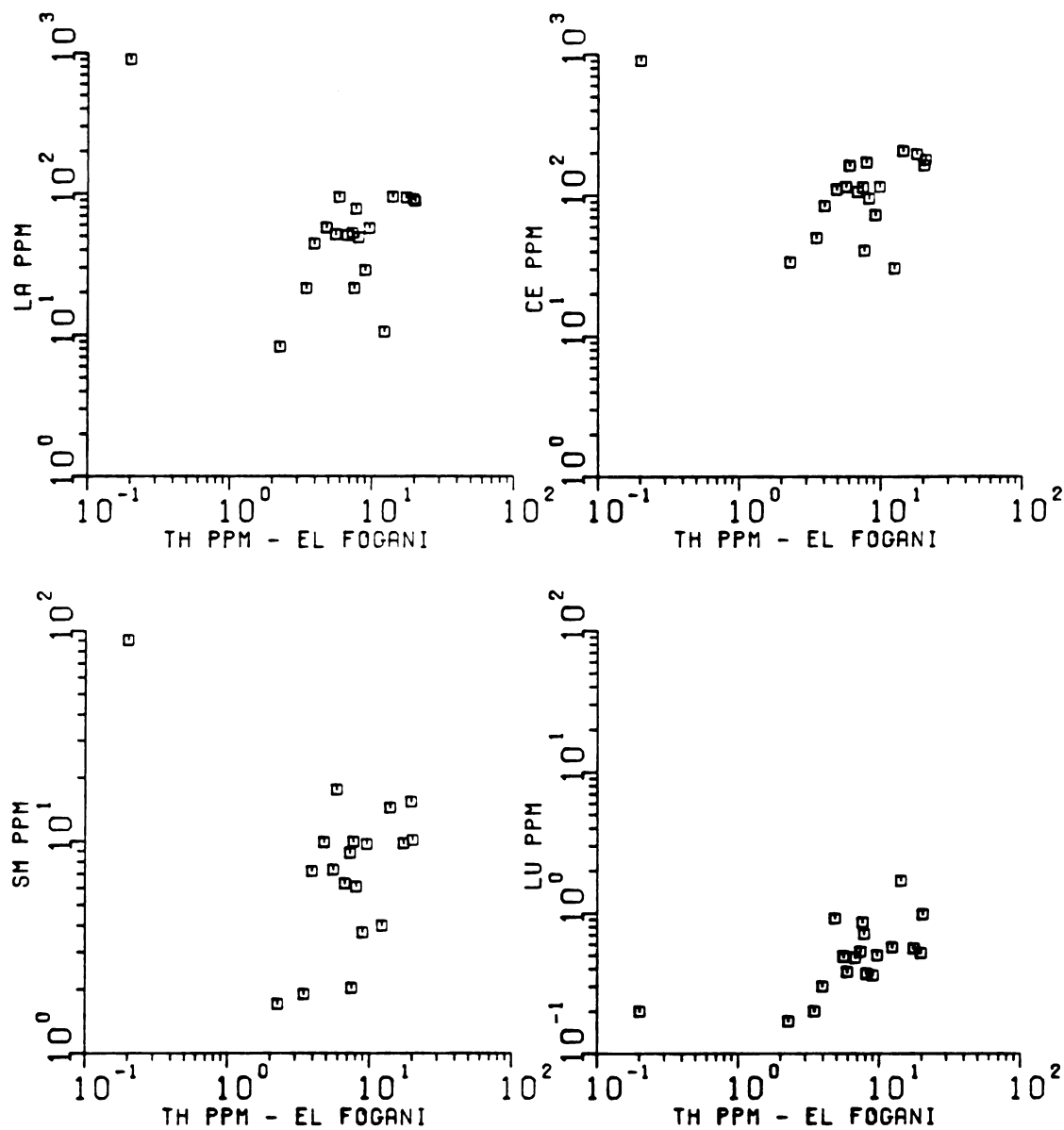


Figure 38.

Log REE versus Log thorium for G. Nigrub El Fogani.

Gabal El Kahfa

In the log Th-log REE plots for El Kahfa, the trends become increasingly less linear for the heavier REE, Figure 39. This "distortion" of the REE trends is considered to reflect the effective presence of volatile-rich phases which mobilized especially the HREE.

This complex is similar to Abu Khruq in age, mineralogy, and LREE distribution. Figure 40 shows the similarity between the two complexes in terms of Ce/Sm versus Ce. Both trends--as discussed earlier (p.81)--are similar to that expected for the fractionation of predominantly pyroxene and feldspar phases. However, El Kahfa lacks the large amounts of sodalite (chlorine-rich) and nepheline-rich syenites which occur at Abu Khruq. El Kahfa contains large cataclastic zones, plugs of granulated rock containing inclusions of olivine and diopsidic pyroxene, along with partially fused feldspar fragments. A large proportion of the exposed rocks at El Kahfa have a similar fused and altered appearance in thin section. The widespread development of these features was not noted in the other complexes, and suggest that El Kahfa's evolution involved large quantities of fluid phases which preceded the emplacement of the late stage dikes.

Based on the similarities between El Kahfa and Abu Khruq, it is assumed that the alkaline melt at El Kahfa contained high concentrations of chlorine. Depending on the silica activity of a melt, the alternative to the

Figure 39.

Log REE versus Log thorium for G. El Kahfa. Note the progressively more anomalous REE concentrations in the heavier REE.

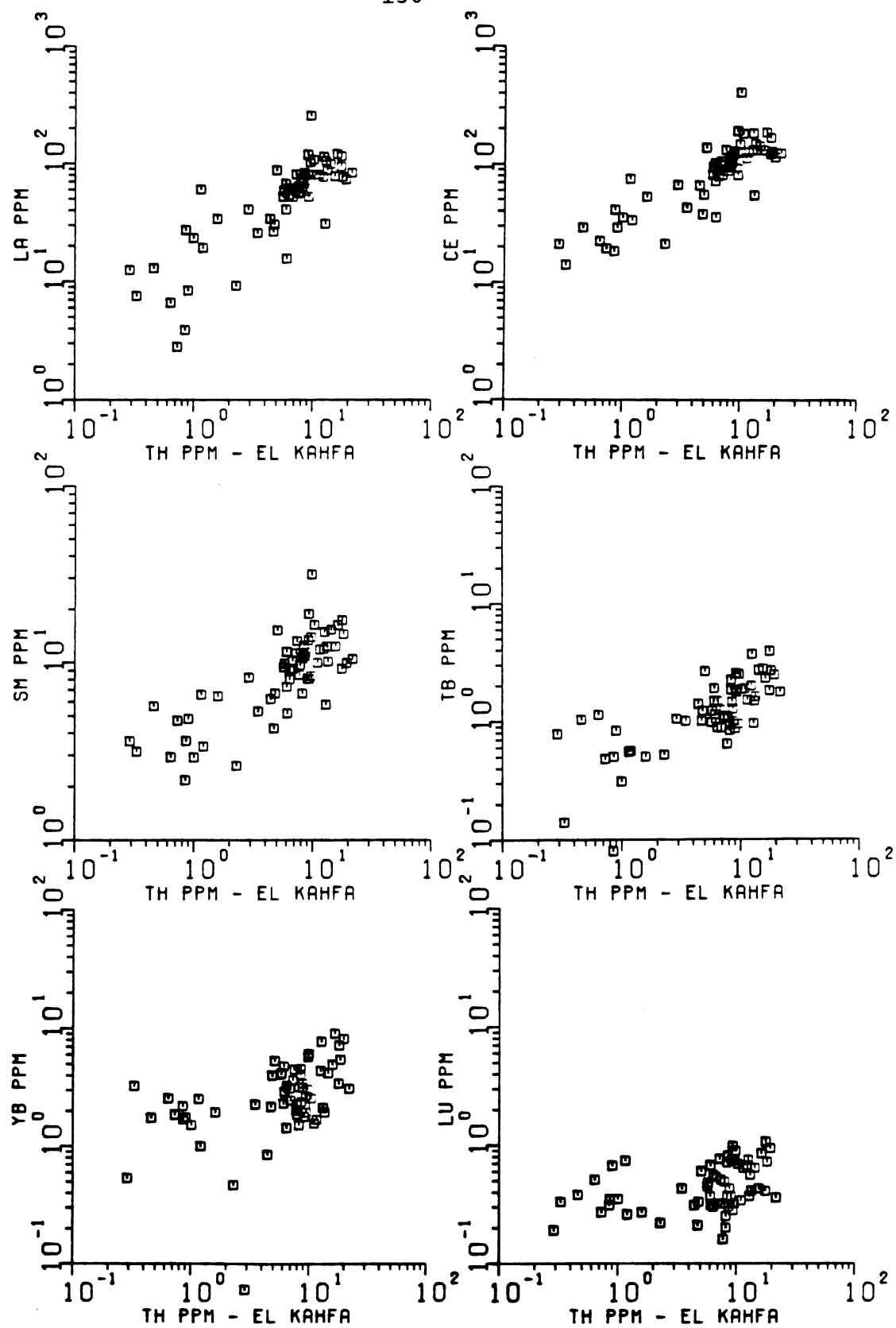


Figure 39.

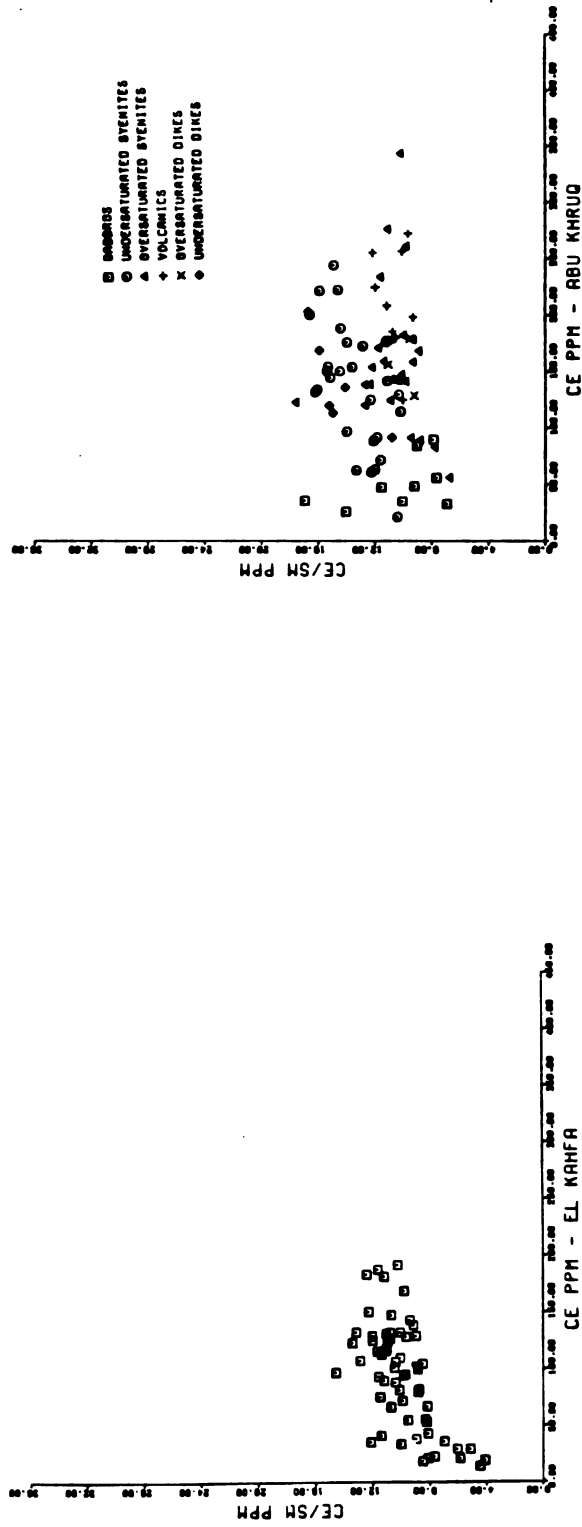


Figure 40 .

Ce/Sm versus Ce for G. El Kahfa and G. Abu Khruq, showing similarity between these two Egyptian alkaline complexes.

crystallization of Cl-rich minerals such as sodalite, is the exsolution of NaCl-rich fluids (Koster van Groos and Wyllie, 1969; Roedder and Coombs, 1967). The lack of the crystallization of Cl-rich sodalite at El Kahfa is considered to have resulted in the release, perhaps explosively, of large quantities of Cl-rich fluid phases from the magma. The experimental work of Flynn and Burnham (1978) indicates that Cl is perhaps the most efficient REE-complexing agent. Due to lanthanide contraction, Cl complexes the REE with greater efficiency as a function of decreasing ionic radii. These effects are interpreted to be reflected in the log Th-log REE plots of El Kahfa. The extensive signs of alteration, the lack of sodalite-rich residual phases, the increased mobilization of the HREE are all attributed to the widespread activity of Cl-rich fluids in the development of El Kahfa. This complex may represent an extreme of a Cl-related process which may have occurred in other complexes, such as Abu Khruq, and may be indicative of the importance of the volatile components in these types of salic alkaline complexes.

Gabal Abu Khruq

Figure 41 demonstrates the log Th-log REE trends for the alkaline suite at Abu Khruq. The relative bulk partition coefficients for the REE of Abu Khruq have been calculated using linear regression analysis to obtain the best-fit line for the data, Table 4. These calculated

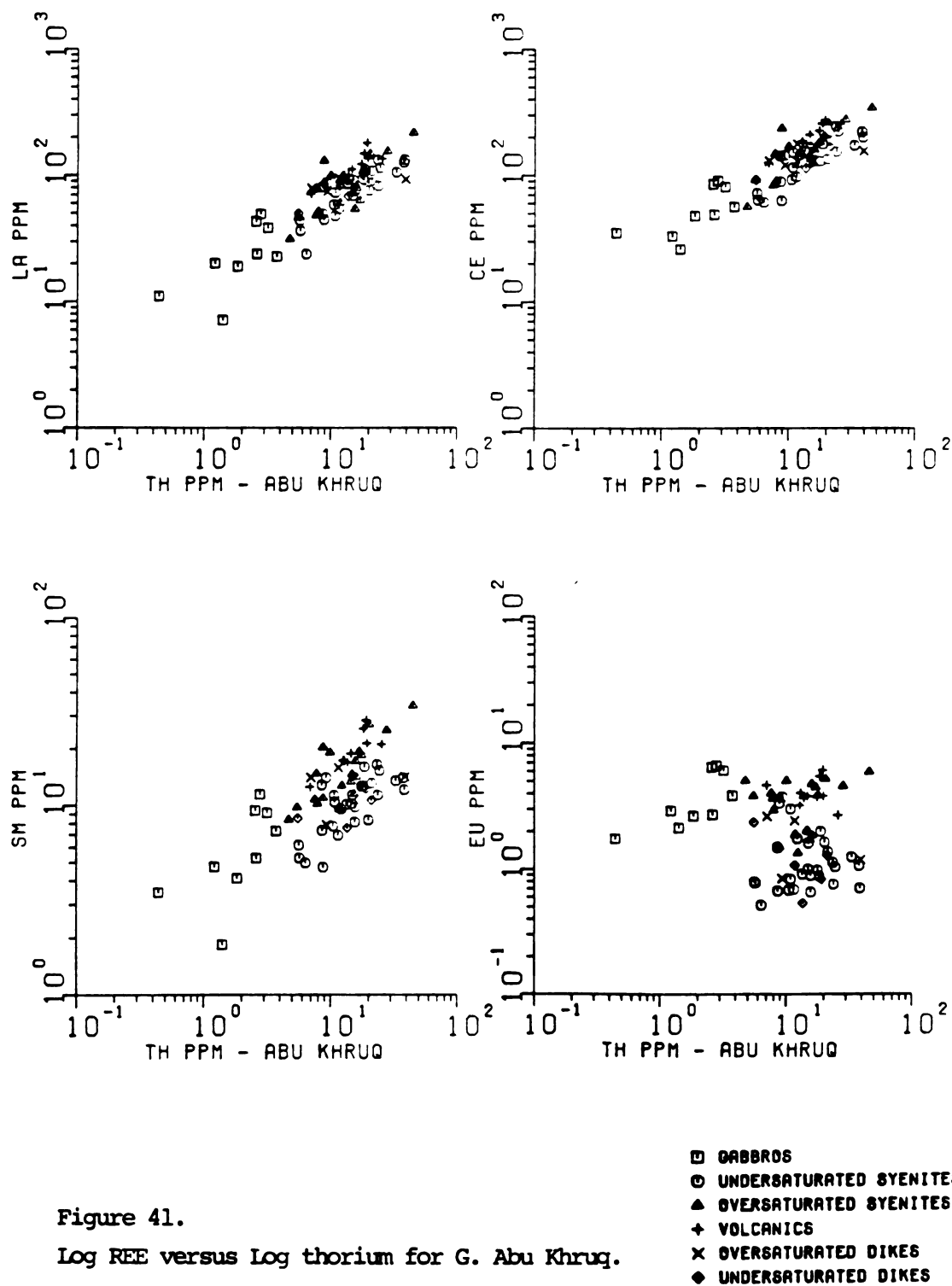


Figure 41.

Log REE versus Log thorium for G. Abu Khruq.

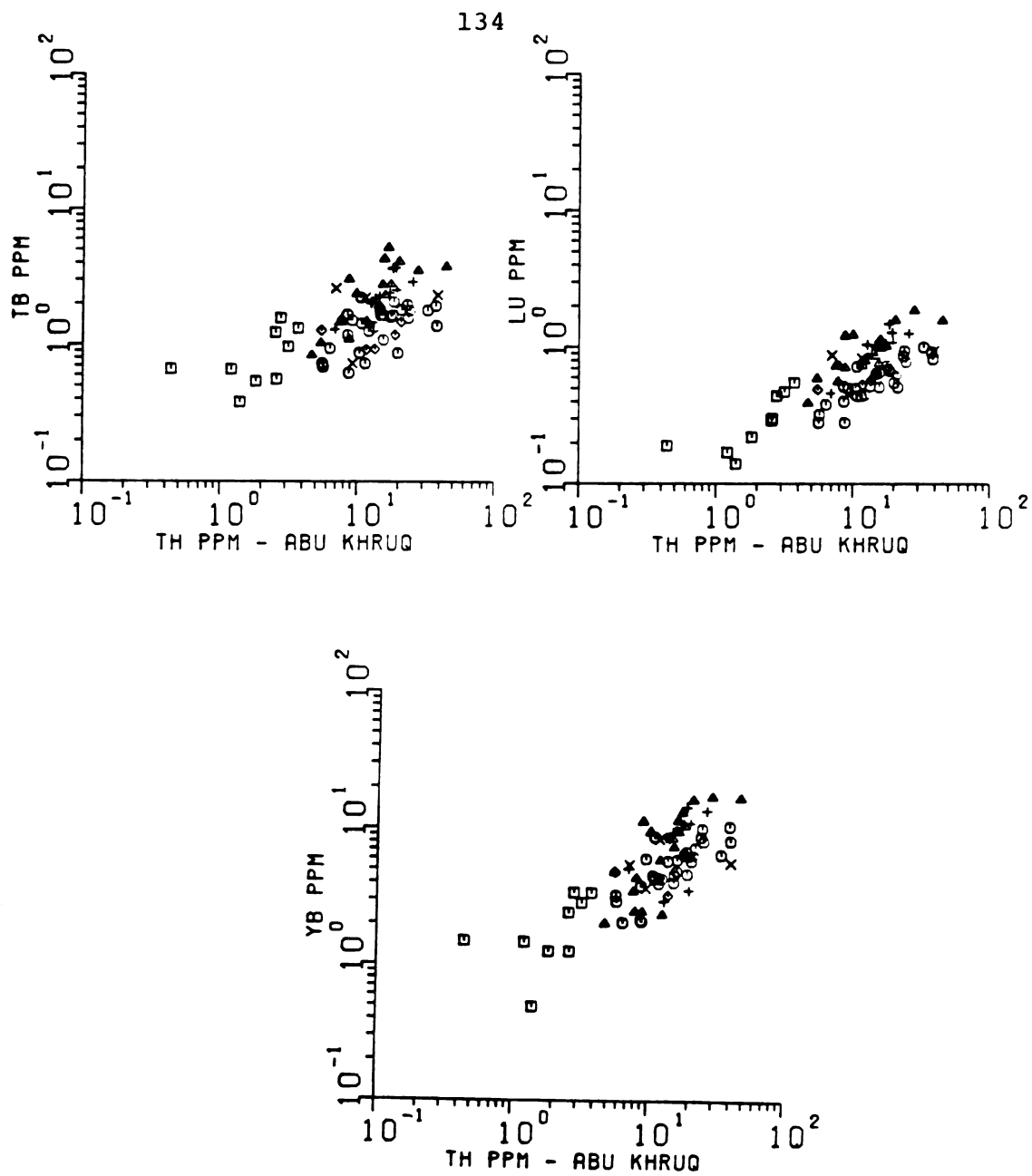


Figure 41 (cont'd).

Table 4. Calculated Relative Bulk Distribution Coefficients
for Gabal Abu Khruq

	\bar{D}	Slope	Correlation Coefficient	Standard Deviation	Intercept
La	.36	.64	.87	.13	+1.19
Ce	.46	.54	.86	.12	+1.55
Sm	.57	.43	.71	.16	+0.61
Tb	.59	.41	.64	.18	-0.24
Yb	.36	.64	.76	.20	+0.06
Lu	.51	.49	.78	.14	-0.70

Method of calculating these bulk distribution coefficients
is after Allegre et al. (1977).

\bar{D} = bulk distribution coefficient.

values indicate that the REE other than Eu, are behaving as incompatible trace elements. The relative bulk partition coefficients for the LREE (La, Ce, and Sm) demonstrate steadily increasing values, as expected for their normal behavior within a melt undergoing fractional crystallization. These values are very reasonable considering the mineralogy of the major rock-types, and compared to published partition coefficients (see Allegre et al., 1977 for comparison). The relative bulk partition coefficients for the HREE, however, are not consistent with their normal behavior in a melt undergoing fractional crystallization. These values indicate that the HREE were behaving anomalously incompatibly.* Although the calculated relative bulk partition coefficient for Lu is only slightly less than Sm, the importance of these numbers is that given the progression of bulk partition coefficients from the LREE to the MREE--values which are entirely reasonable for normal crystal fractionation--Lu would be expected to have a much greater value than the one observed. This apparent mobility of the HREE at Abu Khruq is similar to the extreme HREE mobilization at El Kahfa, although not as pronounced. It is possible that a Cl-rich fluid system similar to the one postulated for the development of

*The extremely low relative bulk partition coefficient for Yb may reflect an analytical problem, or alternatively, an extremely anomalous Yb behavior in alkaline melt systems.

El Kahfa, also existed at Abu Khruq, but due to the crystallization of Cl-rich sodalite in the foyaïtes, was much less pervasive.

Crystal fractionation may have been an important process of differentiation in the Egyptian alkaline melts, but as the volatile phases became progressively more abundant with crystallization of the early solid phases, crystal-liquid equilibria may have been gradually transformed to crystal-liquid-vapor equilibria. With the present data and modelling techniques, it is not possible to quantitatively assess such a multi-phase system. The chemical trends which were formed as a result of crystal fractionation would have begun to deviate from their expected patterns according to these vapor-fluid influences. In order to gain insight into these deviations from the expected compositional trends of crystal fractionation, calculations using the Rayleigh relationship for normal crystal fractionation were performed for the silica-undersaturated and oversaturated series, respectively.

The objective of these calculations is to investigate chemical relationships via the Rayleigh equation, and thereby, perhaps observe the nature of the deviations from normal crystal fractionation. Rather than using the standard "F" notation common to crystal fractionation calculations (amount of parent melt volume remaining), it is more appropriate to consider the analagous value in these calculations as assumed indirect measures of

enrichment, or "E". Average Ce, Sm, Lu and Th concentrations for individual rock-types were used in these calculations. The foyaite were calculated as derivatives of a pulaskite liquid (silica-undersaturated series); the quartz syenites were calculated as derivatives of an alkaline syenite liquid. The calculated relative bulk partition coefficients were used for the REE and an arbitrary bulk partition coefficient of 0.10 was used for Th (taken from Baker et al. (1977) for the Gregory Rift alkaline volcanics). Due to the objective of this scheme, the "E" values will be allowed to vary in order to yield the observed concentrations of the end-members. Table 5 summarizes the results of these calculations. It is clear from the widely varying E values that the process of differentiation between the mafic syenites and the salic end-members of each series did not consist of simple crystal fractionation.

First, it should be noted that the E values of the REE for the silica-undersaturated series are larger than those for the silica-oversaturated series. This reflects the greater REE enrichment in the silica-oversaturated rock series. In spite of the foyaite having much lower REE concentrations than the quartz syenites, their average Th concentration is slightly higher than that of the quartz syenites. This results in a slightly smaller E value for Th in the silica-undersaturated series. There is also a strong reciprocal relationship between the amount of

Table 5. Reciprocal Depletion/Enrichment Relationships Between the Silica-Undersaturated and Oversaturated Rock Series at G. Abu Khruq: REE and Th distributions using the Rayleigh relationship.

I. Silica-Undersaturated Series: Pulaskites Foyaite

	C _O ⁺	C _I	O _I ⁺⁺	D̄	E
(Pulaskites)					
(Foyaite)					
Ce	115.01	158.83	159.47	0.46	0.55
Sm	9.90	11.20	11.23	0.57	0.75
Lu	0.49	0.64	0.64	0.51	0.58
Th	10.37	19.71	19.78	0.10	0.49

II. Silica-Oversaturated Series: Alkaline Syenites Quartz Syenites

	C _O ⁺	C _I	O _I ⁺⁺	D̄	E
(Alkaline Syenites)					
(Quartz Syenites)					
Ce	115.99	195.59	195.29	0.46	0.38
Sm	11.62	18.97	18.61	0.57	0.32
Lu	0.74	1.12	1.12	0.51	0.43
Th	10.01	17.43	17.40	0.10	0.54

Sum of E Values: Silica-Undersaturated + Oversaturated

Ce	0.93	*Average of 7 analyses.
Sm	1.07	**Average of 18 analyses.
Lu	1.01	+Average of 6 analyses.
Th	1.03	++Average of 13 analyses.

C_O = concentration of element in parent liquid.
C_I = concentration of element in calculated remaining liquid.
O_I = concentration of element in the observed rocks.
D̄ = bulk distribution coefficients.
E = enrichment factor, analogous to usual F, see text.

Relative Bulk Distribution Coefficients Calculated for the Silica-Undersaturated and Oversaturated Rock Series (Calculated via the Allegre et al. (1977) method, separately for each rock series).

D̄ Slope Correlation Coefficient Intercept

I. Silica-Undersaturated Series

Ce	0.53	0.47	0.8041	1.5826
Sm	0.70	0.30	0.5498	0.6618
Lu	0.62	0.38	0.7478	-0.6806

II. Silica-Oversaturated Series

Ce	0.28	0.72	0.6550	1.4042
Sm	0.44	0.56	0.5506	0.5711
Lu	0.39	0.61	0.6481	-0.7015

*A method involving least-squares fit of data, see text.

enrichment necessary for the two series. The summation of the E values for each trace element results in a number near 1.00. In the standard crystal fractionation equations this represents 100% volume of the original parent melt; in these calculations it is used as the "original enrichment level" of the parent melt (see Table 5). Therefore, with respect to these calculations via the Rayleigh relationship, the unusual REE-depletion in the foyaites is roughly equal to the REE-enrichment in the quartz syenites. This may be taken as an indication that the REE were preferentially partitioned into the silica-oversaturated system, causing a corresponding depletion in the silica-undersaturated system. Based on the homogeneity of the REE distributions within each rock-type, it is likely that this occurred before crystallization of both the silica-oversaturated rock series and the silica-undersaturated rock series. A mechanism other than crystal fractionation must be called upon, one in which the REE components were transported in liquid state from the silica-undersaturated portions of the melt to the silica-oversaturated volumes. Unlike the E values of the REE, the E value of thorium in the silica-undersaturated series is smaller than that of the silica-oversaturated series. The relative difference of the E values for Th, however, is much less than those for the REE. This may reflect the ineffectiveness of the transport mechanism with respect to thorium, perhaps due to its extreme immobility.

These relationships will be discussed further in the next section.

The relative bulk partition coefficients were calculated separately for the silica-undersaturated series and the silica-oversaturated series. These values are given on the right-hand side of Table 5. The relative bulk partition coefficients for the silica-oversaturated series are smaller than those for the silica-undersaturated series. The apparent greater incompatibility of the REE in the silica-oversaturated system may be interpreted to indicate that the REE were allowed to become more concentrated in the final silica-oversaturated liquids. The lower incompatibility of the REE in the silica-undersaturated liquid and the inferred directions of REE-transport in this liquid zoned system did not allow as great an enrichment in the final silica-undersaturated liquids.

Summary of the Log Th-Log REE Plots

The log Th-log REE trends for the four Egyptian alkaline complexes are similar to each other in terms of their general degrees and directions of enrichment. The nature of especially the LREE are similar to those expected for an alkaline melt evolving by fractional crystallization. The HREE trends for El Kahfa and Abu Khruq clearly reflect a process(es) other than simple crystal fractionation. A large control on the more detailed distributions of the REE is considered to be exerted by the volatile and

halogen-rich phases which become volumetrically more important in the evolved, salic stages of the alkaline melts. Part of the nature and extent of the changes brought about due to these components depends on 1) their composition (i.e. CO_2 -rich, Cl-rich), and 2) the extent to which they are eventually released from the magmatic system. This helps to determine the crystallization of amphibole (El Naga) versus pyroxene (El Kahfa and Abu Khruq), the extent of alterations and autometasomatism, and the crystallization of residual minerals such as sodalite. Although many volatile species may have been important in the evolution of these alkaline melts, the obvious abundant presence of chlorine at Abu Khruq forms the basis for an initial qualitative assessment of the halogen and volatile affects. Widespread activity of Cl-rich fluid phases may explain the extensive signs of alteration and unusual HREE mobility (as predicted by experimental work). The importance of the volatile phases has always been emphasized in the petrogenesis of especially salic alkaline melts, but only recently has there been a growing body of quantitative work done on the specific effects of individual volatile components. Although this lack of understanding does not allow elaborate quantitative models, it is becoming increasingly evident that the halogens and CO_2 exert strong controls on the differentiation of magmas, and especially highly-evolved melt systems rich in these components (e.g. Hildreth, 1977, 1979, 1981; Konnerup-Madsen and Rose-Hansen, 1982).

Section IV

Thermogravitational Diffusion and Fluid Effects

Introduction

Magmatic processes other than fractional crystallization, partial melting, and magma mixing are difficult to unambiguously test using the geochemical data and methods available. There is considerable evidence to indicate that the quiescent shallow magma chambers of the Egyptian alkaline complexes allowed efficient crystal fractionation of an alkaline basaltic parent melt. However, there are regular mineralogic and chemical features found in many of these alkaline complexes which cannot be explained by crystal fractionation. Considering the large degree of crystal fractionation necessary to produce the bulk salic compositions observed in these alkaline complexes, it is likely that the progressive enrichment of the volatile and incompatible components had an important role in the differentiation of especially the salic end-members.

This study considers that a thermogravitational diffusion mechanism is responsible for major features in the Egyptian alkaline complexes--ones which cannot be explained by crystal fractionation. Recently, Hildreth (1977, 1979, 1981) has addressed the processes of thermogravitational diffusion and volatile complexing in silicic magmas with the intent to demonstrate quantitatively that compositional gradients within liquid zoned melts have extremely

important implications for the evolution of especially rhyolitic and granitic associations. This study considers that alkaline silica-undersaturated magmas contain chemical characteristics which may provide optimum conditions for the mechanisms which lead to magma zoning. Although these mechanisms may result in different chemical and mineralogic gradients in alkaline systems compared to rhyolitic systems due to unique chemical and thermal characteristics, alkaline silica-undersaturated/oversaturated systems such as the Egyptian ones may provide some of the most convincing evidence of liquid zoned magmas. The most important feature which may be explained by thermogravitational diffusion is the regular transition of volatile-rich alkaline magmas from silica-undersaturated compositions to oversaturated compositions. Such a mechanism may have relevance to many primitive alkaline associations in Egypt as well as other portions of the world.

1) Nature of the Mechanism

Little geochemical work has been done on quantitatively assessing the mechanism by which a magma becomes zoned. Peralkaline volcanic suites have often been explained in terms of a liquid zonation rather than normal crystal fractionation (Bryan, 1966; Saggerson, 1970; Bailey and MacDonald, 1975; see Bulletin Volcanologique, Special Issue - Peralkaline Rocks, v. 38, #3; Hildreth, 1981). Many early volcanic and ignimbrite flows which initiate

the magmatic activity of an area are more siliceous and incompatible-rich than the eruptions which follow. The composition of the different eruptions on a single volcanic edifice sometimes appear to vary according to the level of the eruption with respect to the volcanic cone. These unusual compositional patterns have often been attributed to the liquid distillation of a single magma in which the lighter and more volatile components are thought to accumulate near the top of the magma chamber, thereby establishing a stratified column of magma. The compositional gradients of numerous rhyolitic flows including extreme trace element depletions or enrichments, cannot be explained by crystal fractionation. Many of the chemical, isotopic, and field relationships, however, can be incorporated into a liquid zoned model (e.g. Hildreth, 1981).

Liquid zoned models have also been proposed for plutonic complexes (e.g. Currie, 1970; see Hildreth, 1981). The gradational zonations of the Kangerdlugssuaq silica-undersaturated/oversaturated alkaline complex in Greenland may represent an unfaulted, type-example of liquid zonation in plutonic alkaline complexes (the complex is described by Pankhurst et al., 1976).

The presence of $\text{Sr}^{87/86}$ and δO^{18} isotopic gradients in igneous suites may often be indicative of large-scale mass transport in magmas (see Hildreth, 1979, 1981). Such transport may be more feasible in alkaline magmas in comparison to other types of melts. This is due to the

polymerized state of alkaline melts because of the low concentrations of silica and high proportions of volatile and halogen components (e.g. Baker et al., 1977; Mustart, 1972). Long-term, low-pressure crustal residence is ideal for the development of efficient crystal fractionation and a progressive enrichment in the volatile, halogen, alkali, and incompatible components. This further decreases the polymerization of the melts. With a slow quiescent evolution, temperature and density gradients are likely to develop. Thus, a combination of convective circulation, Soret diffusion, and volatile complexing/diffusion all aide in forming a thermogravitational diffusion process which leads to the development of a liquid zoned magma. Such zoning may be capable of maintaining long-term liquid interfaces (Hildreth, 1977, 1979, 1981).

2) Evidence in Support of Simultaneous/Overlapping Development of the Major Rock-Types

There are numerous factors which may be interpreted to indicate a simultaneous or overlapping development of the major rock-types at Abu Khruq. Such a development is considered to support a liquid zoned magma.

1. The field relationships of the rock-types at Abu Khruq and many other similar alkaline complexes, are often interpreted to indicate that there was an alternating crystallization sequence of silica-undersaturated and oversaturated melts (e.g. El Ramly et al., 1969a). Explanations of such

alternations of silica-saturation levels may involve separate batches of melt of contrasting silica-saturations. This is unlikely, however, based on the gradational mineralogic and chemical relationships at Abu Khruq, and the frequent occurrence of similar alkaline complexes throughout the Eastern Desert as well as in many other areas of the world. An alternative model to such oscillating sequences is the simultaneous development of both silica-undersaturated and oversaturated compositions in a liquid zoned magma.

2. There are gradational boundaries between the pulaskites and foyaïtes (silica-undersaturated varieties), between alkaline syenites and quartz syenites (silica-oversaturated varieties), and between pulaskites and alkaline syenites (slightly silica-undersaturated and oversaturated varieties). Such gradational boundaries are even more evident in less faulted alkaline complexes (e.g. Kangerdlugssuaq, Greenland, Pankhurst et al., 1976). A similar gradational relationship may also explain the apparent lack of clear gabbro-syenite contacts.

3. The compositions of the different varieties of syenites may be used to support their simultaneous or overlapping development. The trace element characteristics of the pulaskites and alkaline syenites (slightly silica-undersaturated and oversaturated) are very similar. The gradations toward the foyaïtes and quartz syenites (strongly silica-undersaturated and oversaturated) demonstrate progressively diverging trace element characteristics. This

gradual chemical divergence is considered to reflect the general radial chemical gradient from the inner portions of the complex to the outer portions.

4. The composition of the volcanic cone is very siliceous and incompatible-rich. Due to the early formation of the cone, this is considered to indicate that a portion of the melt at Abu Khruq was highly differentiated and silica-oversaturated even in relatively early stages of the alkaline magmatic activity.

5. The major element and trace element concentrations of the quartz syenites are essentially identical to the volcanics. This is considered to indicate that a portion of the melt at Abu Khruq maintained a similar silica-oversaturated and highly differentiated composition--some of which was erupted early in the development of Abu Khruq, and the remainder crystallizing later as plutonic counterparts. Due to the detailed chemical similarity of the quartz syenites to the volcanics (e.g. the REE), it is likely that their process of differentiation was similar.

6. The general lack of evidence of large-scale reactions between the different varieties of rock-types at Abu Khruq is considered to indicate that the development of these different compositions occurred under quasi-equilibrium conditions within a single magmatic system. The absence of reactions between the syenites of contrasting silica-saturation levels may be used to support a near-equilibrium transfer of silica throughout the melt. The presence of

secondary quartz in the quartz syenites, and the lack of secondary quartz in the immediately underlying gabbros, for example, is difficult to explain unless this alkaline magmatic system contained specific directions of silica transport.

7. The radial chemical gradient across the complex of Abu Khruq is an extremely important observation. Lutz (1979) discusses the presence of a radial isotopic gradient at Abu Khruq in terms of $\text{Sr}^{87/86}$ and δO^{18} . These gradients are unlikely to represent crustal interactions because the isotopic ratios are within mantle-derived values (Lutz, 1979). An alternative explanation is that such isotopic gradients reflect the mass transport of material within the magma (Hildreth, 1979, 1981). Lutz (1979) also noted a strong radial gradient in terms of SiO_2 , Na/K , and Sr . Due to the somewhat asymmetric and faulted structure of Abu Khruq, these apparent radial gradients need to be studied more closely by detailed radial sampling and analysis. Even so, the general radial arrangement of the major rock-types and the radial distribution of silica are clear. Due to this radial geometry and the presence of gradational boundaries, it is likely that the chemical characteristics noted for each rock-type also reflect this radial relationship. Accordingly, this would mean a radial enrichment of the REE from the inner portions of the complex to the outer portions. This is similar to the REE gradient noted by Hildreth (1977, 1979) for the liquid zoning model of the Bishop Tuff.

3) A Preliminary Thermogravitational Diffusion Model for Gabal Abu Khruq

Although there are numerous parameters which are unknown in this model, there are many features which are qualitatively consistent with a thermogravitational diffusion mechanism for the differentiation of the salic rock-types at Abu Khruq.

The thick, stable crust of the Nubian shield provided a quiescent tectonic setting for the emplacement and differentiation of the Egyptian alkaline melts. This allowed the regular development of efficient crystal fractionation, stable thermal and density gradients, and a progressive enrichment of the alkali, volatile, and halogen components of the melt. This type of magmatic system provides the optimum conditions necessary for the development of a thermogravitational diffusion mechanism due to the low-polymerized nature of such an enriched melt, and the enhanced diffusivities of cations and complexed species (see Hildreth, 1979). It is proposed that thermodiffusion and volatile complexing caused mass transfer of Si, K in preference to Na, Sr, Ba, the REE, and many other components. These constituents were fluxed from the inner portions of the magma to the roof and outer portions of the chamber. This is depicted in the schematic diagram of Figure 42.

It is rather uncertain as to the details of how and when a thermogravitational diffusion process might have been established within the sodic alkaline melt at Abu

Figure 42.

A schematic illustration of a thermogravitational diffusion mechanism in shallow alkaline magma chambers. Note the radial volatile-flux pattern and the development of a liquid-zoned magma chamber. Such zoning may not have contributed to the formation of the gabbros early in the history of the development of the melt at G. Abu Khruq. It is likely that crystal fractionation was especially effective in the early development of these alkaline magmas, and it may be solely responsible for the formation of the gabbros. Crystal fractionation may have continued to operate in conjunction with the thermogravitational diffusion mechanism. Continued fluxing toward the outer crystallized margins of the complex resulted in the deposition of secondary quartz and possibly other components. This type of mechanism is considered to be responsible for the radial chemical and isotopic gradients in these complexes, and the overall development of the silica-undersaturated/oversaturated associations.

It should be noted that development of ring fractures and pulses of magma from different volumes in these types of liquid-zoned magma chambers may result in a variety of configurations in these alkaline complexes.

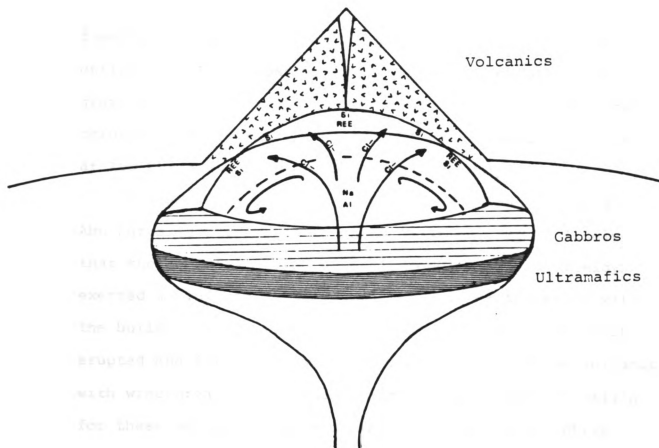


Figure 42.

Khruq. It is possible that its effectiveness was not operative until after the crystal fractionation of the gabbroic phases, resulting in the compositional drive toward peralkalinity via the "plagioclase effect", as well as an enrichment in the volatile and incompatible components. It is also uncertain as to the importance of fractional crystallization in the final stages of differentiation. It is possible that a combination of thermogravitational diffusion and crystal fractionation occurred throughout most of the evolution of these salic rock-types at Abu Khruq and other similar alkaline complexes.

The early formation of the evolved volcanic cone at Abu Khruq is attributed to magma zonation. It is likely that thermodiffusion and volatile complexing had already exerted a profound influence on the melt at Abu Khruq with the build-up of a silica-rich, volatile-rich cupola which erupted and formed the siliceous, REE-rich alkaline volcanics with widespread signs of autometasomatism. Such an origin for these volcanics may explain why almost their entire volume consists of silica-oversaturated compositions in spite of resting on, and being intruded by silica-under-saturated plutonic rocks. If such a liquid zoned mechanism accounts for the nature of the volcanics, it is not unreasonable that the same process affected other portions of the melt. This is, in part, supported by the very similar compositions of the quartz syenites to the volcanics and their position in the outer portions of the complex.

Once the outer margins of the complex began to solidify, silica and probably other components would continue to be deposited. This would explain the presence of secondary quartz in the syenites which form the outer perimeters of the complex. Because solidification is likely to have preceded gradually inward, the outermost quartz syenites would have received the largest amounts of secondary quartz. This is the relationship observed in the field and in thin sections.

The alkaline syenites (slightly silica-oversaturated) and the pulaskites (slightly silica-undersaturated) form the intermediate compositions of the complex. They may represent the initial stages of the thermogravitational diffusion process. Although much of their volume may have been partially crystalline, they may have played a part as the medium for thermodiffusion and volatile-fluxing.

The central foyaïtes crystallized last and represent the final residual-hydrothermal system, characterized by the metasomatic textural features common to many plutonic alkaline complexes (e.g. Kogarko et al., 1978). These rocks represent the de-silicated, transport-depleted portion of an originally silica-undersaturated melt. They may have formed a hot fractionating column within the central portions of the melt, yielding the silica and other components necessary for the formation of the silica-oversaturated perimeters of the magma. It is likely that this inner portion of the magma was involved in a complex

interaction of many processes including crystallization, re-melting, mixing, and vapor-fluxing. Similar to the suggestions by Luth and Tuttle (1953) and the agpaitic alkaline magmatic model proposed by Kogarko, Burnham and Shettle (1978), the discrete vapor phases are not lost outside the magma system, but rather are slowly re-dissolved back into the residual melt as temperatures decrease. The final stages of salic magmatism would then involve a continuous transformation from a magmatic to a hydrothermal system. The gradual development of these fluid effects and their genetic relationship to the magmatic system, may allow these alterations to occur under quasi-equilibrium conditions (e.g. albitization, sericitization, cancrinitization, feldspar-clouding, and secondary quartz). This may help to explain the lack of severe scatter in the chemical data in many alkaline complexes in spite of the extent of the alterations. The final stages of salic alkaline melts may be characterized by a regular association of residual-hydrothermal phases and low-temperature alterations. Such an approach possibly offers some insight into the abundance of the "secondary alterations" which have been noted to occur in an almost identical pattern in many alkaline complexes throughout the world. Given similar initial compositions and an uninterrupted differentiation sequence, these fluid effects may be nearly as regular as the crystal-liquid equilibria of crystal fractionation.

Roofward migration of certain chemical species does not require a discrete vapor phase (Kennedy, 1955; Shaw, 1974). Although the major elements may not have been significantly affected, volatile, isotopic, minor and trace element compositional gradients may have developed early in the history of the alkaline melt at Abu Khruq. As the melt cooled, volatile complexing and the strong thermal and density gradients may have begun to exert an increasing influence on the differentiation of the magma. Although it is uncertain as to the timing and extent of vapor phase separation, it is possible that halogen-rich and CO_2 -rich aqueous fluids began to flux through the magma and greatly enhance the development of a liquid zoned magma.

The polymerization of a melt is perhaps one of the most important factors in thermodiffusion and volatile transfer. Thus, an important component in liquid zoned systems is thought to be the halogens because they are highly soluble in peralkaline liquids, cause de-polymerization of the melt, and allow greater mobility of the cations (Mustart, 1972; Baker et al., 1977; Hildreth, 1979, 1981). It is clear that chlorine was an important component in the magma at Abu Khruq (crystallization of sodalite; Cl analyses by El Ramly et al., 1969a). Based on experimental and natural studies, it is likely that Cl plays an important role in the thermogravitational diffusion mechanism (Hildreth, 1979, 1981; Mustart, 1972). As temperatures declined in the sodic alkaline melt at Abu Khruq, there

would have been a tendency to exsolve a NaCl-rich fluid and/or eventually crystallize sodalite (see Koster van Groos and Wyllie, 1969; Roedder and Coombs, 1967). Experimental data indicates that Si, K, Na, the REE, and other components may become mobile in the presence of chlorine and other commonly associated volatile species such as the hydroxides (MacKenzie, 1960; Orville, 1963, 1972; Anderson and Burnham, 1967; Flynn and Burnham, 1978). Orville (1972) demonstrated that silica and potassium may be more mobile than sodium in the presence of a Cl-rich phase. Chlorine is also perhaps one of the most efficient REE-complexing agents (Flynn and Burnham, 1978). Similar to other REE-complexing agents, Cl complexes the REE with increasing efficiency in the heavier elements of the series due to the lanthanide contraction (Flynn and Burnham, 1978). Due to the high concentration of Cl in the volatile flux, and the complexing characteristics of Cl as predicted by experimental work, the REE, silica, potassium in preference to sodium, and other components, were preferentially carried toward the roof and outer portions of the magma. This resulted in the REE-rich silica-oversaturated syenites and volcanics in the outer and erupted portions of the magma. Due to chlorine's preference for the heavier REE, the HREE were more abundant in the melt than normally expected in silicate systems. This is reflected in the unusually small relative bulk partition coefficient of Lu (refer to Table 4).

As crystallization preceded inward, only a residual melt and the fluxing fluid phases remained as the final stages of the magma. Similar to the alkaline magmatic model proposed by Kogarko, Burnham and Shettle (1978), it is suggested that these REE-depleted fluid phases gradually re-dissolved back into this residual melt. This final melt-hydrothermal system is represented by the foyaïtes, which are strongly depleted in Si, K, the REE, Sr, and Ba (around zero ppm), and relatively enriched in other components such as Al, Na, and Th. The Rayleigh fractionation scheme performed in the previous section for the silica-undersaturated and oversaturated series, respectively, may provide some additional insight into the reciprocal enriched versus depleted pattern between the outer silica-oversaturated rocks, and the inner silica-undersaturated rocks. Assuming an exponential relationship in the distribution of the REE and Th, it appears that the degree of depletion in these final foyaïtic rocks is roughly equal to the degree of enrichment in the outer quartz syenites. This is interpreted to support the transport of REE from the central portions of the melt, toward the outer portions of the magma. Unlike the REE, thorium is interpreted to have been less affected by the transport mechanism--perhaps due to its greater incompatibility. Even so, its slight relative enrichment in the final crystallizing stage of the foyaïtes, is reciprocal to the slight relative depletion in the quartz syenites.

In spite of the overall REE depletion in the foyaïtes, the incorporation of Cl into the crystal structure of sodalite caused the remaining REE to be strongly incorporated into this final crystallizing stage because of its unusual REE-complexing characteristics. This is reflected in the high relative bulk partition coefficients for the REE as calculated for only the silica-undersaturated series (refer to Table 5). In addition to the relative depletion of the REE in the foyaïtes, the presence of large amounts of chlorine caused an increase in the relative abundance of the HREE compared to the LREE due to Cl's greater preference of the heavier elements of this series. This resulted in unusually low Ce/Lu ratios in the foyaïtes, and their "flattened" normalized REE patterns (see Figure 11). Thus, the inferred importance of Cl in these sodic alkaline melts is interpreted to be the cause of the decrease in the Ce/Lu ratios between the pulaskites and the foyaïtes, see Figure 43, (rather than an increase as would be expected in a case of crystal fractionation).

Although the roofward enrichment of silica and the LREE at Abu Khruq is similar to the pattern noted by Hildreth (1977, 1979, 1981) in his study of the Bishop Tuff, many of the other elements display opposite enrichment directions--including Eu, the HREE, Th, and Na/K. Further study of these types of sodic alkaline systems--as opposed to the silica-rich potassic rhyolite systems such as the Bishop Tuff--may reveal the reasons for such differences.

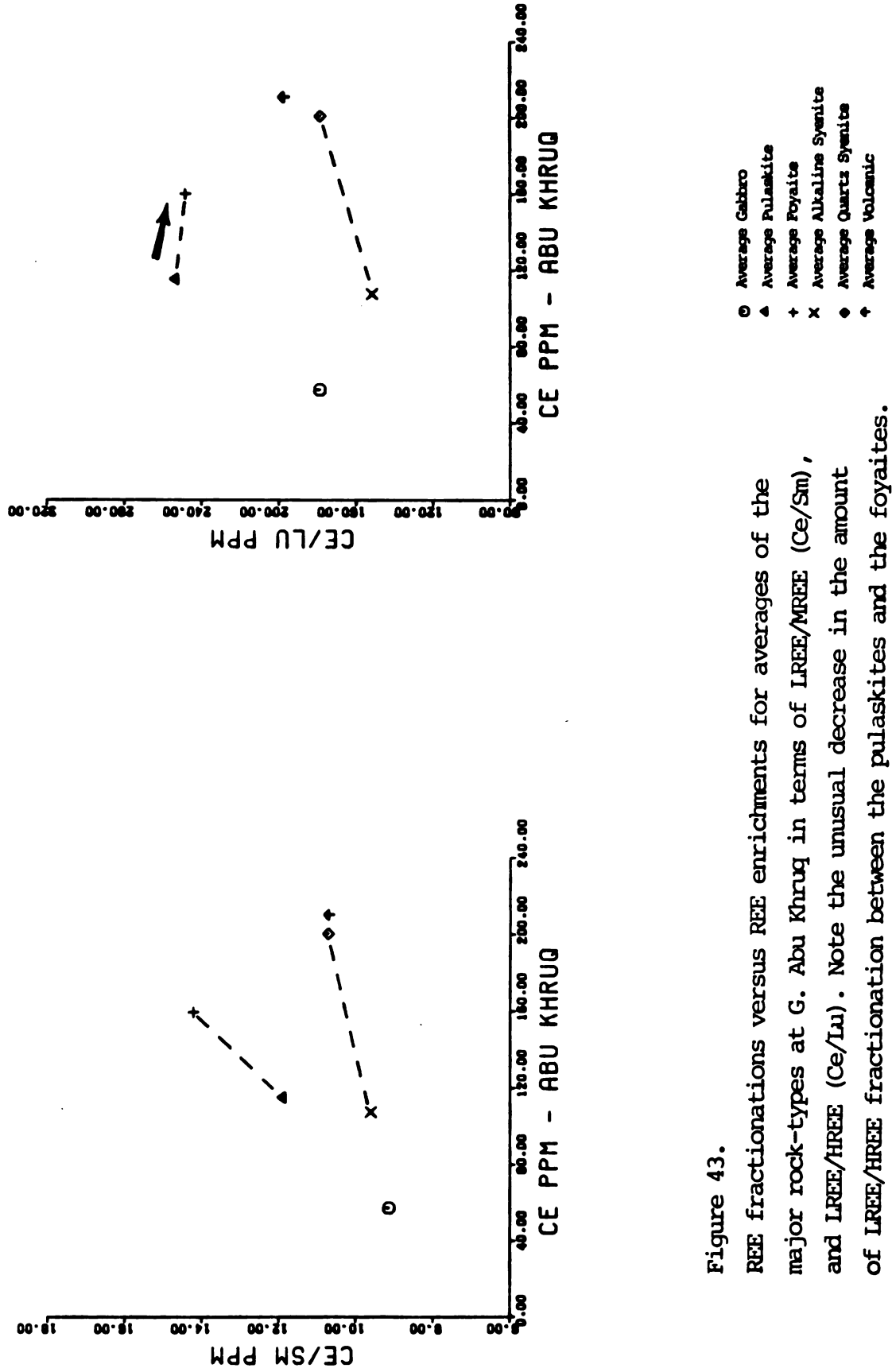


Figure 43.

REE fractionations versus REE enrichments for averages of the major rock-types at G. Abu Khrug in terms of LREE/HREE (Ce/Sm), and LREE/HREE (Ce/Lu). Note the unusual decrease in the amount of LREE/HREE fractionation between the pulaskites and the foyaites.

Differences in the thermal and density structures of melts, differences in the degree and nature of polymerization, the available volatile complexing, Soret coefficients, and many other factors may cause distinct gradients within different chemical systems. Models involving liquid zoning must take into account not only these differences in liquid structure and composition, but also the time factor involved in plutonic versus volcanic associations. The re-dissolution of the volatile-flux phases in the plutonic case, for example, may result in more complicated differentiation sequences and chemical patterns. It is also important to point out that many events may occur in the history of any specific alkaline magma so that unique geometries and compositions may be produced. This may involve structural changes in the surrounding crust, additional magmatic activity, or individual pulses of magma from different volumes in the zoned magma. Alkaline complexes with similar rock-types, but different apparent orders of intrusion/crystallization may be explained by any combination of pulses of magma from different volumes or levels of the liquid zoned magma.

The evolution of alkaline melts via fractional crystallization, thermogravitational diffusion, and the gradual transformation of the residual volatile-rich melt toward a hydrothermal system, may represent a regular sequence of processes involving crystal-liquid equilibria as well as crystal-liquid-vapor equilibria, all set within the thermal

and density structure of the magmatic system. Thus, given similar initial alkaline-volatile "packages" from the mantle, and a stable magma chamber, these alkaline melts may follow a fairly consistent course of differentiation so that similar alkaline complexes are frequently generated.

PART II

THE TECTONIC SETTING OF THE EGYPTIAN ALKALINE COMPLEXES

Introduction

The second objective of this study is to investigate the tectonic setting of the alkaline magmatism in Egypt. In spite of the similar chemical patterns of many primitive alkaline suites throughout the world, there is no well defined, unified model on the tectonic processes associated with alkaline magmatism. Thus, in order to approach this problem, the tectonic settings of other alkaline provinces are used as comparisons, the general tectonic development of Afro-Arabia is discussed, and common features are compared to the Egyptian alkaline province. A tectonic process is proposed for alkaline magmatism, and evaluated in terms of the common features noted for alkaline magmatism in general, and more specifically, for the timing and relationship of events in the Egyptian alkaline province.

CHAPTER 5

ALKALINE MAGMATISM AND PLATE TECTONICS

Introduction

The tectonic setting of alkaline magmatism has long been a major problem in petrology (Harker, 1896, 1909). Alkaline rock-types have been noted to occur in areas of relatively little crustal activity and low heat flow (Bowen, 1937; Tilley, 1958). This type of setting has been classically referred to as "anorogenic". This term implies no fundamental crustal process, and describes the state of most of earth's surface throughout much of its history. In spite of this, the term "anorogenic" often is used to describe the tectonic setting of alkaline magmatism (e.g. MacIntyre, 1977).

With the development of the theory of plate tectonics, a substantial amount of work has been devoted to characterizing magmas according to their tectonic setting (e.g. Barth, 1962; Martin and Piwinski, 1972; Sugisaki, 1976; Pearce et al., 1977; Petro, Vogel and Wilband, 1979). This type of approach led to generalized conclusions in which calc-alkaline magmas are most frequently produced at compressional plate boundaries, and alkaline magmas are most frequently produced at extensional plate boundaries. This may appear to be a better description for the tectonic setting of alkaline magmatism. Many models on the origin of alkaline magmatism have been based on the rift/dome setting, and

this has developed to such an extent that some alkaline provinces with no apparent rift association are sometimes presumed to be simply "pre-rift" or "failed rift" areas (e.g. Dostal et al., 1979; Currie, 1970; Upton, 1974; Upton and Blundell, 1978; Kisvarsanyi, 1980; Darbyshire and Fletcher, 1979; Ressetar et al., 1981; Strong and Minatidis, 1975; Sykes, 1978; Kumparapeli, 1970; Falvey, 1974). However, such models on alkaline magmatism ignore a large number of alkaline assemblages formed in tectonic settings which cannot be described as rifts.

This strong correlation often found in the literature between alkaline magmatism and rifting would seem especially applicable to the Egyptian alkaline province in view of its proximity to the Red Sea, Figure 44. However, age-dating of the Egyptian alkaline complexes demonstrates that this similar type of alkaline magmatism occurred sporadically over a period of more than 450 m.y., from 554 to 89 m.y. B.P. (Serencsits et al., 1981; Lutz, 1979; Hashad and El Reedy, 1979), see Figure 44. If one ascribes to the Red Sea rift/dome model for the generation of the younger complexes (see Ressetar et al., 1981), it is not unreasonable to expect similar tectonic events for the origin of the older complexes. However, there is presently little evidence for rifting in the immediate Red Sea area before the doming and strike-slip deformation of the Late Cretaceous, and the well-developed extensional activity of the Eocene (Garson and Krs, 1976; Youssef, 1968; Swartz and

Figure 44.

Map of the major alkaline complexes in the Eastern Desert of Egypt, and the general tectonic style of Afro-Arabia.

1. Wadi Dib, 554 m.y. B.P.
2. Zargat Naam, 404 m.y. B.P.
3. Tarbtie (North), 351 m.y. B.P.
4. El Gezira, 229 m.y. B.P.
5. El Naga, 145 m.y. B.P.
6. El Mishbeh, 142 m.y. B.P.
7. Nigrub El Tahtani, 140 m.y. B.P.
8. Nigrub El Fogani, 139 m.y. B.P.
9. Mansouri, 132 m.y. B.P.
10. El Kahfa, 91 m.y. B.P.
11. Abu Khruq, 89 m.y. B.P.

Dates from Serencsits et al. (1981) and Hashad and El Reedy (1979). Map and tectonic lineaments after Garson and Krs (1976).

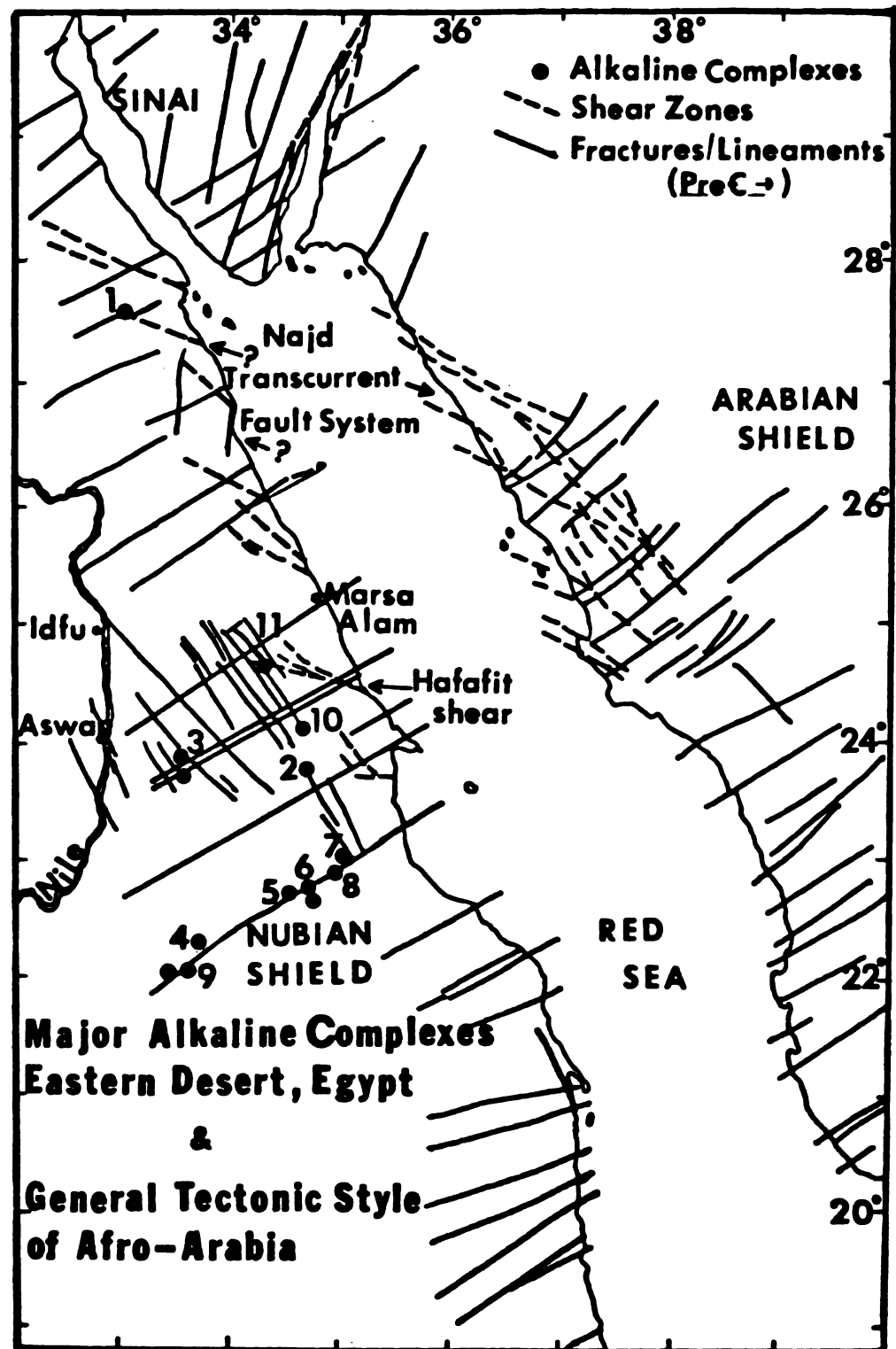


Figure 44.

Arden, 1960; Burek, 1969; Abdel-Gawad, 1969; Hancock and Kadhi, 1978; Brown, 1970; Coleman, 1974; Pilger and Rosler, 1976; Le Pichon and Francheteau, 1978). It is unlikely that each of these episodes of alkaline magmatism in Egypt occurred with rifts similar to the early Red Sea activity throughout most of the Phanerozoic eon.

It is proposed here that these multiple episodes of similar alkaline magmatism coincide with periods of tectonic reactivation during major changes in plate motion and inferred periods of thermal disequilibrium between the lithosphere and the mesosphere. These changes in plate motion have involved not only rifting/extension, but compression, and large-scale strike-slip motion.

1) Characteristics of Alkaline Magmatism

The Egyptian and many other alkaline provinces are similar in chemistry, general structural setting, and long-term reoccurrence of alkaline magmatism (see Part I). Thus, support for the proposed tectonic setting of the Egyptian alkaline province consists of comparisons to other alkaline occurrences as well as several approaches which corroborate the coincidence of the Egyptian episodes of alkaline magmatism with periods of regional structural re-adjustment and major shifts in plate motion.

Based on a review of the Egyptian and numerous other alkaline provinces, several features are characteristic of many occurrences of alkaline magmatism:

1. Similar alkaline magmatism occurs in oceanic and continental crust.
2. Similar alkaline magmatism occurs in regions dominated by either extension or compression (Schwarzer and Rogers, 1974; Frey et al., 1978; McBirney and Aoki, 1968; Miyashiro, 1978; Baker et al., 1977).
3. Alkaline magmas tend to occur away from the most thermally and mechanically active areas (Bowen, 1937; Tilley, 1958; Barth, 1962; Sørensen, 1974; Miyashiro, 1978).
4. This alkaline magmatism tends to occur along major crustal fractures, often with regional strike-slip deformation (Sørensen, 1974a; Sykes, 1978; Garson and Krs, 1976; De Long et al., 1975; Hawkins and Natland, 1975).
5. Episodes of alkaline magmatism can often be shown to coincide with major shifts in plate motion (Boyd, 1973; Marsh, 1973; Opdyke and MacDonald, 1973; van Breemen and Bowden, 1973; Foland and Faul, 1977; Herz, 1977; Sykes, 1978), or they may occur along zones of sustained strike-slip motion (see Hawkins and Natland, 1975; De Long et al., 1975; Arculus et al., 1977).
6. Alkaline magmatism often reoccurs over very long periods of time along the same set of major fractures in areas which are not disturbed by high thermal regimes (Zartman, 1977; Foland and Faul, 1977; Foland and Loiselle, 1981; Sykes, 1978; van Breemen et al., 1975; Currie, 1970; Doig and Barton, 1968; Kumarapeli and Saull, 1966;

Sørensen, 1974; Butakova, 1974; Blaxland et al., 1978; Lutz, 1979; Serencsits et al., 1981).

Crustal Thermal Regime and Structural Setting

Alkaline, and especially highly silica-undersaturated alkaline magmas, tend to occur away from the thermal high of active, tholeiitic-dominated mid-oceanic ridges and the normal convergent zones of island arcs (Miyashiro, 1978; DeLong et al., 1975). This same feature has been noted in analagous continental rift areas in which alkaline magmatism tends to occur on the flanks of the rift structure, away from the most thermally active areas (Wright, 1965, 1970; Le Bas, 1971; Baker et al., 1972) and in the foreland of convergent zones (van Breemen et al., 1979; Butakova, 1974). Schwarzer and Rogers (1974) stressed that the worldwide similarity of the major element chemistry of alkaline stuies in different crustal settings reflects a mantle process deep enough in the mantle to be unaffected by the chemical and thermal differences between mantle regions under the various tectonic settings.

Alkaline magmatism has been frequently noted to be associated with major crustal fractures and lineaments (e.g. Sørensen, 1974, p. 145; Sykes, 1978; DeLong et al., 1975; Hawkins and Natland, 1975; Garson and Krs, 1976). Although the exact nature of some of these lineaments is obscure, many prove to be major fault systems, frequently with significant strike-slip components (see Sykes, 1978;

DeLong et al., 1975). Some lineaments may represent crustal fractures, possibly annealed after several episodes of movement. Based on deep-sea dredgings and petrogenetic work (Thompson and Melson, 1972; Stebbins and Thompson, 1978), oceanic alkaline magmatism may be the result of leaky transform faults away from the active mid-oceanic thermal highs. These faults may provide access to a deeper source region in the mantle than the more voluminous tholeiitic variety of magmatism (Shibata, Thompson and Frey, 1979). There are numerous examples of alkaline assemblages associated with continental lineaments, and often these are on trend with transform faults in the adjacent oceanic plates (Foland and Faul, 1977; Ballard and Uchupi, 1972; Fletcher, Sbar and Sykes, 1974; Sykes, 1978; MacFarlane and Ridley, 1949; Marsh, 1973; Dingle and Scrutton, 1974; Rabinowitz, 1976; Cornelissen and Verwoerd, 1975; Fodor and Husler, 1977; Herz, 1977; Fitton and Hughes, 1977; Thompson and Melson, 1972; Stebbin and Thompson, 1978; Dunlop and Fitton, 1979; Booth et al., 1978; Shibata, Thompson and Frey, 1979). Even in areas which do not display obvious crustal offset, major episodes of alkaline magmatism have been explained in terms of fracture mechanics (e.g. Hawaiian Seamount chain; Jackson, Shaw and Bargar, 1975; Turcotte and Oxburgh, 1978; Shaw, 1980; Shaw et al., 1980; e.g. Easter-Sala y Gomez and Juan Fernandez island-seamount chains; Pilger and Handschumacher, 1981).

2) The Diverse Tectonic Settings of Alkaline Magmatism

In order to formulate a better concept of the possible tectonic conditions under which alkaline magmatism may occur, a review was made of many alkaline occurrences. Some examples of this review are briefly described and referred to in the following pages with the intent of isolating consistent tectonic aspects associated with the alkaline magmatism in these different settings.

The Atlantic alkaline provinces lend some good examples of alkaline magmatism before, during, and after major periods of continental rifting. The White Mountain Magma Series (Foland and Faul, 1977; Foland and Loisel, 1981), the Monteregian Hills (Kumparapeli, 1970; Foland and Faul, 1977), the Niger-Nigerian alkaline province (van Breeman and Bowden, 1973; van Breeman et al., 1975), the Angolan and South African alkaline provinces (Marsh, 1973; Neill, 1973; Sykes, 1978), the Brazilian alkaline occurrences (Amaral et al., 1967; Marsh, 1973; Neill, 1973; Opdyke and MacDonald, 1973; Herz, 1977; Sykes, 1978), and the Bolivian Velasco alkaline province (Darbyshire and Fletcher, 1979) are all clear examples of continental alkaline provinces which contain complexes emplaced during the major openings of the Atlantic Ocean. Many of the above workers have noted that these alkaline complexes tend to lie along major lineaments thought to have been activated during rifting. The major rifting activity and accompanying alkaline magmatism in the South Atlantic occurred around

140 m.y. B.P.; in the central Atlantic similar rifting and alkaline magmatism occurred earlier, around 180 m.y. B.P. (e.g. Sykes, 1978).

Initially it was often assumed that all the alkaline complexes in these Atlantic alkaline provinces evolved in the early rifting activity (see van Breemen and Bowden, 1973). The on-going work in the age-dating of alkaline provinces is revealing that isolated portions of especially the older continental areas are commonly the sites of long-term reoccurrences of alkaline magmatism. Similar to the long age span of the Egyptian alkaline province, the ages of the alkaline complexes in many provinces span several hundred million years (see Zartman, 1977; Blaxland et al., 1978; Butakova, 1974; Currie, 1970; Doig and Barton, 1968; Kumarapeli and Saull, 1966; Sykes, 1978; Herz, 1977; Marsh, 1973; Neill, 1973; Opdyke and MacDonald, 1973; Amaral et al., 1967; van Breemen and Bowden, 1973; van Breeman et al., 1975; Foland and Faul, 1977; Foland and Loiselle, 1981). This means that the alkaline magmatism in some of these provinces occurred in apparent episodes before, during, and after rifting (see especially Foland and Faul, 1977; van Breemen et al., 1975; Sykes, 1978). To date, the episodes of alkaline magmatism preceding continental rifting in these Atlantic alkaline provinces have not been adequately explained in terms of tectonic events. The two major episodes of alkaline magmatism which occurred after the continental rifting of the South and

Central Atlantic ocean basins, respectively, have been correlated with major shifts in the spreading of the Atlantic. Outbreak of alkaline magmatism in the South Atlantic alkaline provinces appears to coincide with a sudden shift in spreading of the South Atlantic around 80-90 m.y. B.P. during the Late Cretaceous (Marsh, 1973; Opdyke and MacDonald, 1973; Herz, 1977; Sykes, 1978). Similarly, an episode of alkaline magmatism in the provinces surrounding the Central Atlantic appear to coincide with a shift in the spreading of the Central Atlantic around 120 m.y. B.P. (e.g. Foland and Faul, 1977; Sykes, 1978).

Besides the separation in time from rifting, it is also important to note the emplacement of alkaline magmas along major lineaments far inland from the Atlantic continental margins during these periods of rifting and shifts in spreading. The recently discovered Velasco alkaline province in Bolivia intruded major lineaments in the Amazon craton at the same time the initial rifting of the South Atlantic was taking place 1600 kilometers away (circum 140 m.y. B.P.; Darbyshire and Fletcher, 1979). The Monteregian Hills alkaline complexes were emplaced during the proposed shift in the spreading of the Central Atlantic, well inland from the then mature rifted continental margins (Foland and Faul, 1977).

There are numerous examples of alkaline magmatism occurring within regional compressional tectonic regimes

(Schwarzer and Rogers, 1974; DeLong et al., 1975; Hawkins and Natland, 1975; Barth, 1962; Payne and Shaw, 1967; Appleyard, 1974; Barker, 1969, 1974; Sørensen, 1974, p. 146; Bowden, 1974; Marsh and Carmichael, 1974; Arculus, 1976; Arculus et al., 1977; Smith et al., 1977; Johnson et al., 1976; Bevier et al., 1979; Robin and Tournon, 1978; Miller, 1977; Miyashiro, 1978; Heming, 1979; Lutz, 1979; van Breemen et al., 1979).

The Samoan island chain represents one of the largest single occurrences of alkaline magmatism in the world. Silica-undersaturated alkaline volcanics erupted simultaneously along a 500 kilometer lineament which Hawkins and Natland (1975) interpret to be a fracture propagated by the subduction of the sharply flexed Pacific plate descending into the Tonga trench.

The island of Grenada, located at the southern end of the Lesser Antilles island arc, contains a highly silica-undersaturated alkaline suite (Arculus, 1976; Sigurdsson et al., 1973). There is seismic (Molnar and Sykes, 1969) and structural evidence (e.g. Alberding, 1957; Malfait and Dinkelman, 1972) indicating that this area is dominated by large-scale strike-slip faulting between the Caribbean and South American plates. DeLong et al. (1975) concluded that such strike-slip tectonics are genetically related to the presence of the alkaline magmatism rather than the subduction process itself. They made similar interpretations for eight other alkaline occurrences in oceanic

subduction zones. They concluded that these alkaline suites occurred either 1) "along or near the edge of a plate where hinge-faulting is suggested...or 2) when a fracture or other linear feature roughly perpendicular to the arc is being subducted" (DeLong et al., 1975, p. 731; see also Marsh and Carmichael, 1974).

Appleyard (1974) and Payne and Shaw (1967) have reviewed a "syn-tectonic" alkaline suite which appears to have been emplaced along strike-slip faults during the Grenville orogeny in Ontario, Canada. Paleomagnetic interpretation supports this style of strike-slip deformation with large-scale dextral shear between Gondwana and Laurentia (Morel and Irving, 1978).

There are some important alkaline suites which are associated with the Caledonian compressional tectonics. In the British Isles two major centers of alkaline magmatism are Loch Borrolan and Loch Ailsh. It has been concluded by several workers that they were emplaced in an episode of thrusting associated with the closing of the Iapetus Ocean (most recently by van Breemen et al., 1979). A similar tectonic setting may have existed for the Caledonian alkaline assemblages of northern Norway (Sturt and Ramsay, 1965; Appleyard, 1974), and Newfoundland (O'Driscoll and Strong, 1979). Each of these Caledonian alkaline magmatic centers occurred in regions that were dominated by oblique convergence and large-scale strike-slip deformation during the Caledonian compressional event

(Phillips et al., 1976; Mitchell and McKerrow, 1975).

Northern Norway, in particular, represents a transform boundary which, according to recent paleomagnetic evidence, was the site of extremely large amounts of strike-slip movement during both the opening and closing of the Iapetus Ocean (Kj de et al., 1978).

One of the largest and best-studied alkaline provinces in the world is the Gardar alkaline province in southern Greenland. This area was dominated by Proterozoic strike-slip tectonics (Bak et al., 1975; Grocott, 1979; Watterson, 1978; Upton, 1974; Stephenson, 1976). The dates of the alkaline complexes correlate with or slightly preceding the Grenville orogeny (Blaxland et al., 1978; van Breemen and Upton, 1972). Upton (1974) and others have suggested that the alkaline magmatism is actually associated with a Proterozoic rift system which was active during or slightly before the Grenville Orogeny (see Upton and Blundell, 1978). van Breemen and Upton (1972) simply state that the alkaline magmatism occurred in the foreland of the Grenville orogeny. The work of Bak et al. (1975), Watterson (1978) and Grocott (1979) on the shear deformation of Greenland appears to favor compressional tectonics during the same time as the alkaline magmatism.

Compressional forces may be more effective in fracturing the crust than extensional ones. Oblique compressional forces may be highly effective in forming or reactivating major zones of weakness in the crust (e.g. Benioff, 1962;

Fitch, 1972; Molnar and Tapponier, 1975; Mitchell and McKerrow, 1975; Holcombe, 1977; Tapponnier and Molnar, 1976, 1979; Watterson, 1978). Such fault systems may have great importance in the initial formation of intraplate dislocations, crustal response to latter deformation, (see Watterson, 1978) and localization of alkaline magmatism.

A major point in this study is the observation that the alkaline magmatic process is versatile with respect to many specific types of crustal activity. The reoccurring nature of alkaline magmatism in a given region does not support a model for the origin of these melts which is dependent on any single type of tectonic event such as rifting or subduction. Similar types of alkaline magmatism occur in a wide variety of crustal settings, including oceanic and continental crust undergoing extension or compression. The sum of the conditions which may be observed in the diverse crustal settings of alkaline magmatism, and the long-term reoccurrence of alkaline magmatism in old cratonic areas, may provide some constraints on a model for the origin of many alkaline occurrences. It is possible that alkaline magmatism represents a fairly regular geodynamic process which is not constrained to any single type of tectonic setting.

CHAPTER 6

MODELS FOR THE PRODUCTION OF ALKALINE MELTS

There are fundamentally four models on the initial production of mantle melts which may be applicable to the generation of primitive alkaline melts:

- 1) Localized mantle plumes,
- 2) Global mantle disturbances,
- 3) Lithospheric flexure,
- 4) Shear heating.

The first two approaches suggest that the mantle is the active system which creates melts and causes major structural changes in the overlying lithosphere via thermal or chemical disturbances (e.g. Morgan, 1971, 1979; Wilson, 1963; Anderson, 1975; Vogt, 1972, 1979, 1981; Burke and Dewey, 1973; Crough, 1978; Smith, 1979; Thiessen et al., 1979). The last two approaches suggest that the lithosphere plays a major active role in the creation of mantle melts rather than being the passive system (e.g. Schubert et al., 1976; Schubert and Yuen, 1978; Froidevaux and Souriau, 1978; Bailey, 1977; Sykes, 1978; de Gruyter and Vogel, 1981; Feigenson and Spera, 1980, 1981; Shaw, 1973, 1980; Shaw et al., 1980).

1) Mantle Plumes and Global Mantle Disturbances

The first two approaches are closely related and are the source of current debate (Gass et al., 1978; Vogt, 1981;

Pollack et al., 1981). One approach proposes that outbreaks of magmatism are the result of the localization of individual mantle plumes, plate thickness and plate velocity (Gass et al., 1978; Pollack et al., 1981). The second approach calls upon similar mantle plumes, but on a simultaneous, global scale (Vogt, 1972, 1979, 1981).

Although it is often difficult to prove or disprove the existence of mantle plumes in the geologic history of many areas, there are two features of the Egyptian and other alkaline provinces which make it difficult to impose any of the current plume models for the production of these alkaline melts. The ages of the Egyptian alkaline complexes--similar to the wide age distributions of many alkaline provinces--span a period of over 450 m.y. The complexes do not display a regular geometric pattern with respect to their ages, similar to those presented as evidence for the presence of mantle plumes in other areas. It is unlikely that mantle plumes may remain fixed to the lithosphere for the length of time over which alkaline magmatism reoccurred. It is also unlikely that mantle plumes may account for the irregular geometric patterns.

This type of long-term, non-systematic spatial and/or temporal distribution of alkaline magmatism has been used as evidence against a mantle plume origin in many areas: the White Mountain Magma Series, the Monteregian Hills (Foland and Faul, 1977), several African and Brazilian alkaline provinces (Marsh, 1973; Sykes, 1978), the Samoan

island chain (Hawkins and Natland, 1975), other alkaline occurrences in subduction zones (DeLong et al., 1975), and the Egyptian alkaline province by other authors (Serencsits et al., 1981; Lutz, 1979). The simultaneous eruption of alkaline volcanics along a 500 kilometer lineament in the Samoan island chain is a case in which any current plume model may be rejected (Hawkins and Natland, 1975). The synchronous emplacement of alkaline magmas in widely separated areas is difficult to account for by the localized mantle plume hypothesis. A good example of this is the synchronous emplacement of alkaline magmas in several areas along portions of the mature Atlantic margins, well after the opening of the ocean basins.

It is important to point out that even the best examples of the mantle plume model and linear age progression have also been explained in terms of fracture mechanics (e.g. the Hawaiian seamount chain: Jackson, Shaw and Bargar, 1975; Shaw and Jackson, 1973; Turcotte and Oxburgh, 1978; Shaw, 1980; Shaw et al., 1980).

Thus, if mantle plumes consistently account for alkaline magmatism, such models must become more complex in order to explain the spatial and temporal distribution of many alkaline igneous centers in rift areas, subduction zones, collisional fronts, continental margins of mature oceans, and intraplate areas. The global synchronism of mantle plume convection, as suggested by Vogt (1972, 1979, 1981), may provide such an alternative.

This study suggests an approach other than a planetary "hot spot" system.

2) Lithospheric Flexures

This approach also utilizes mantle upwellings, but the cause of the mantle disturbance lies in the motion of the lithosphere (e.g. Bailey, 1977). Bailey (1970, 1974) suggested that crustal arching of a rigid anisotropic plate by mechanical compression of surrounding plates, could cause partial melting via decompression in the underlying mantle. Volatile-fluxing would be an important result, thus explaining the enriched alkali-incompatible trace element composition of alkaline magmas being emplaced in these areas. This is a rather novel idea in that rifting would begin by compression. van Breemen, Aftalion and Johnson (1979) suggest that this approach may explain the presence of the Scottish compressional alkaline complexes (e.g. Loch Borrolan). Although this model cannot be discounted in certain areas, there are many alkaline suites which are not associated with crustal arching nor signs of immediately preceding compression.

3) Shear Heating

With the increasing number of advances in the understanding of creep properties of Earth materials and their application to mantle and lithosphere rheology, it is becoming evident that rather than the mantle always operating as the active system, deformation caused by the

movement of the lithosphere is of great importance in the production of magmas (Shaw, 1963, 1965, 1969, 1973; Shaw and Jackson, 1973; Jackson et al., 1972; Jackson and Shaw, 1975; Jackson et al., 1975; Shaw, 1980; Shaw et al., 1980; Feigenson and Spera, 1981; see Phil. Trans. Roy. Soc. Lon., v. 288-A, pp. 383-646, 1978; Phys. Earth Plan. Int. Special Issue, v. 17, no. 1, 1978). The current models on the mantle emphasize the slow, solid-state convection of the mesosphere (e.g. Weeterman, 1970; Weeterman and Weeterman, 1975; Elsasser et al., 1979), which is probably moving on the order of mm/yr (Minster and Jordan, 1978). The lithosphere, on the other hand, moves at relatively rapid rates, on the order of cm/yr., and often undergoes abrupt changes in acceleration. It is unlikely that the mantle, with the currently proposed rheology, could provide the forces necessary to consistently initiate these rapid changes. Many of the abrupt changes in the motion of the lithosphere may be caused by the geometry of the plates and the subduction-collision mechanism. If this is the case, it is the mantle which must adjust to these rapid changes in lithospheric motion. This is accomplished within the asthenosphere, a ductile layer (perhaps around 70 to 400 kilometers in depth, Weeterman, 1970; Stocker and Ashby, 1973) of relatively low viscosity which forms a zone of mechanical decoupling between the rigid lithosphere and the mesosphere.

It has been noted for some time that the flowage of material results in the dissipation of energy and the involvement of shear heating (Zener, 1948; Kennedy, 1963). There has been considerable theoretical work on shear heating, and recently, it has been applied to large-scale thermo-mechanical modelling of Earth's lithospheric and mantle systems (e.g. Schubert and Yuen, 1978; Melosh, 1976, Yuen and Schubert, 1977, 1979; Anderson and Perkins, 1974; Melosh and Ebel, 1979; Froidevaux and Souriau, 1978; Feigenson and Spera, 1980, 1981). It is possible that shear heating in the Earth as a result of changes in the motion of the lithospheric plates may provide a heat source necessary for the generation of mantle melts.

Indirect measurement of the rheology and shear stress at the base of the lithosphere shows that large lithospheric plates may have forces caused by basal shear stress that are comparable to the forces acting at the major plate boundaries (Melosh, 1977). This results in the concentration of large amounts of shear in the asthenosphere and the production of heat due to viscous dissipation. The regular translation of the lithosphere may thus result in the production of melts in the asthenosphere. However, based on indirect viscosity measurements (e.g. Drury, 1978) and the apparent balance between plate motion and the resistive forces (Jacoby, 1970), it is unlikely that this melt production due to the constant translation of the plates is large enough to be a regular source for crustal magmatism.

Studies on the thermal and mechanical considerations of the lithosphere-mantle systems using different assumptions and variables indicate that under a given rate of plate motion and other conditions, the thermal balance between the lithosphere and the mantle must move toward thermal steady-state conditions in order to avoid an eventual thermal runaway (Anderson and Perkins, 1974; Melosh, 1976; Schubert and Yuen 1978; Yuen and Schubert, 1979; Melosh and Ebel, 1979). However, plate accelerations cause an increase of shear stress on the base of the lithosphere and in the asthenosphere, and may result in increased heat dissipation and deep thermal anomalies (Froidevaux and Souriau, 1978; Melosh and Ebel, 1979). This causes a decrease in the viscosity of this decoupling zone which may lead to additional increases in shear stress and plate velocity so that heat dissipation may again increase and effectively lead to a thermally unstable state (thermal feedback: Grunfest, 1963; Shaw, 1969; Schubert and Yuen, 1978; Melosh, 1976; Yuen and Schubert, 1977, 1979; Froidevaux and Souriau, 1978; Melosh and Ebel, 1979). It is possible that under these conditions, significant amounts of asthenospheric melts are generated in order to aide in returning the lithosphere-mantle system to new thermal steady-state conditions. These partial melts may be produced in sufficient quantities to become a significant form of crustal magmatism (e.g. Shaw, 1969; Shaw and Jackson, 1973; Jackson and Shaw, 1975; Froidevaux and Souriau, 1978; Shaw, 1980; Shaw, et al.,

1980; Feigenson and Spera, 1981). Although it is possible that melts arise from the lithosphere/asthenosphere interface (see Shaw, 1973), the common lack of coincidence between the low-velocity/seismic attenuated zones and this interface (Froidevaux and Souriau, 1978) indicates that melt production probably occurs at different levels (Shankland, 1977) in a potentially unstable, thermally active "boundary zone" of the general lower lithosphere-asthenosphere regions (Koide and Bhattacharji, 1977). This boundary zone may be important in terms of providing a fairly continuous mantle source region for the production of magmas of different degrees of alkalinity and silica-depletion.

High pressure experimental work indicates that alkaline, silica-poor melts may be generated under large amounts of pressure, pressures similar to those expected for the asthenosphere (Green and Ringwood, 1967; Kushiro, 1968; Green, 1970). Trace element modeling of mantle compositions and alkaline rock-types indicates that alkaline melts are the result of small degrees of partial melting of mantle material (Gast, 1968; Kay and Gast, 1973; Frey et al., 1978). The presence of small amounts of melt in the asthenosphere has been suggested based on geophysical evidence of low-velocity and seismic attenuated zones (Nur, 1971). Electrical conductivity modeling supports the presence of small amounts of partial melt within the lower lithosphere and upper mantle (Shankland, 1977; Drury, 1978). Laboratory

(Unger, 1967) and field (Boudier and Nicolas, 1972; Padovani and Carter, 1977) evidence suggests that this partial melt exists as well-interconnected, low melt fractions.

The recent thermo-mechanical modelling which evaluates the theoretical aspects of shear heating in the asthenosphere and lower lithosphere, combined with the above geophysical and petrologic interpretations, appears to provide a reasonable basis for a model on the production of primitive alkaline magmas. Unlike the theoretical problems encountered in accounting for the large thermal anomalies necessary for the generation of silica-rich, non-alkaline melts via shear heating (e.g. Yuen and Schubert, 1979; Froidevaux and Souriau, 1978), alkaline melts require only relatively small thermal anomalies as effective heat sources. The shear heating model is consistent with all the common characteristics noted for primitive alkaline magmatism (p.168). Such a model would concur with the correlation between alkaline magmatism and major shifts in plate motion (e.g. Marsh, 1973; Opdyke and MacDonald, 1973, van Breemen and Bowden, 1972; Foland and Faul, 1977; Herz, 1977; Sykes, 1977; Hawkins and Natland, 1975; DeLong et al., 1975; Arculus et al., 1977; Shaw, 1980; Shaw et al., 1980; McHone, 1981). The common association between alkaline magmatism and major crustal fractures may be a result of the accompanying crustal failure under such plate motion, thus, providing a more feasible mechanism for the ascent

of the small degrees of asthenospheric melts. The viscous dissipation and the resulting generation of alkaline melts in the asthenosphere, as well as the initiation or reactivation of the overlying crustal fractures, may be responses to regional extension, compression, or transform plate movement. Such an approach provides a versatile process for the generation of alkaline magmas which requires only shifts in plate motion rather than specific types of tectonic settings. This may explain the common long-term reoccurrences of alkaline magmatism in many alkaline provinces. Intraplate fractures with dominant strike-slip motion maintain the relatively low thermal regime established by the underlying deep/low melt-fraction source region. The lower incidence of alkaline magmatism in areas containing high levels of mechanical and thermal activity--such as is common near subduction zones--may be a result of the dominant melt source region being too shallow, and generating degrees of partial melting which are too large to produce an alkaline character. In areas of gradually changing or fluctuating thermal/lithospheric activity, the corresponding changes in the depth and degree of melting of the shear zones in the melt source regions may produce the ranges of alkalinity and silica-depletion observed in alkaline provinces. These zones of shearing in the mantle may provide a suitable locus for the metasomatic fluids and enriched mantle material which appear necessary for the

generation of primitive alkaline melts (e.g. Frey et al., 1978; see Chapter 2).

Factors which are likely to have a bearing on the incidence of alkaline magmatism are the degree of thermal disequilibrium (magnitude of the shift in the motion of the plates and localization of possible concentrated shear dissipation), the ease of ascent of the alkaline melts (presence/activation of major fractures), and the involvement of shallow, larger degrees of non-alkaline melts which would effectively obscure the alkaline character of the deeper mantle melts (i.e. mixing). Ascent of the asthenospheric melts in the mantle is not addressed by this study, but is discussed by some workers (e.g. Turcotte and Ahern, 1978; Koide and Bhattacharji, 1977; Anderson, 1981).

Perhaps one of the most important aspects of alkaline magmatism which can be explained by shear heating model is the tendency for alkaline magmatism to occur simultaneously in widely separated areas of the world. Although more detailed work is needed on the age dating of alkaline complexes and their tectonic significance, there are a large number of alkaline occurrences which tend to cluster around similar dates (e.g. see Sykes, 1978; Darbyshire and Fletcher, 1979; Foland and Faul, 1977; Marsh, 1973). Commonly, these clusters of dates correspond to alkaline magmatism which occur in areas affected by the same episode of plate acceleration, but which may be separated by very large distances (see p. 174). The Egyptian and Atlantic

alkaline provinces are considered to be examples of the simultaneous incidence of alkaline magmatism in widely separated areas which were affected by the same shifts in plate motion.

CHAPTER 7

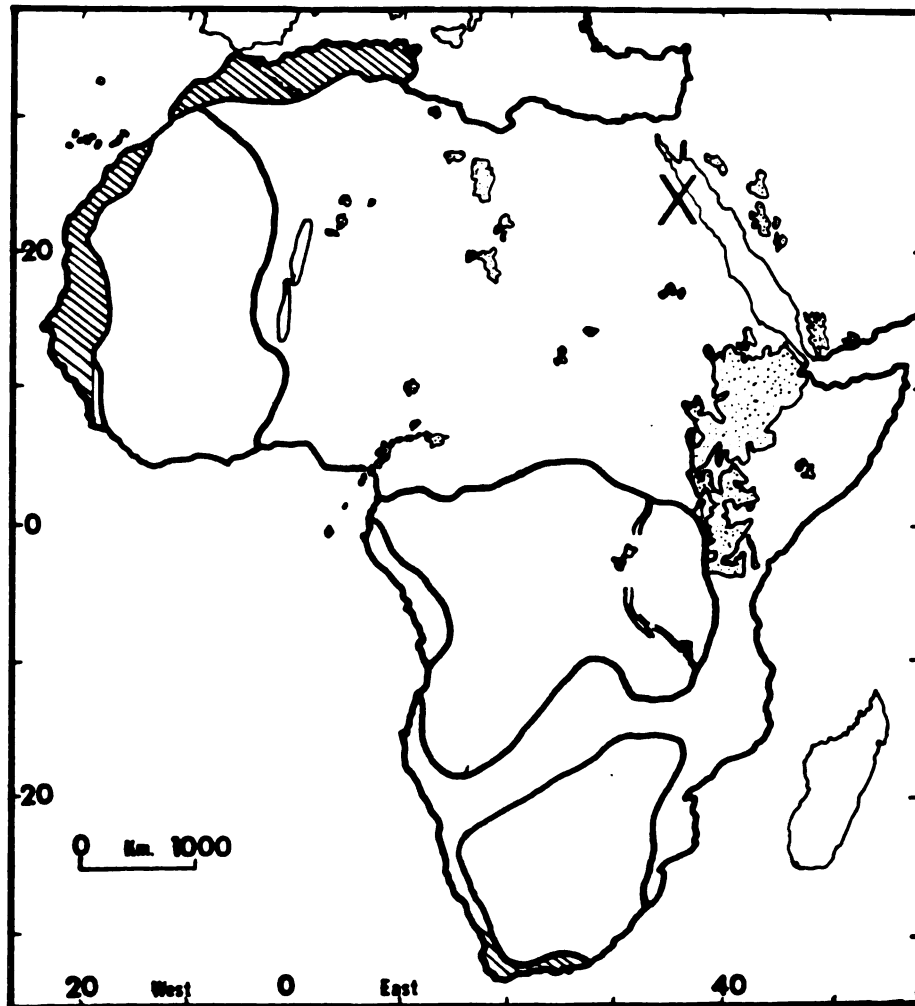
A TECTONIC MODEL FOR THE EGYPTIAN ALKALINE PROVINCE

1) The Tectonic Development of the Eastern Desert and Adjacent Areas

The African continent in general is composed of three major stable, old, and largely thermally inactive cratons surrounded by younger, more active crust which is thought to have been formed in the Pan-African tectono-thermal event, Figure 45 (see Kennedy, 1964; Clifford, 1970; Gass, 1970). The exact meaning of the "event" is rather obscure, occurring throughout portions of Africa from about 1100 to 500 m.y. B.P. (Gass, 1977). It was characterized by sequences of small amounts of folding, sedimentation, greenschist metamorphism, and a predominance of strike-slip faulting. Hurley (1972) perhaps made the initial suggestion that Africa may represent a landmass similar to one that might result if the Pacific Ocean were to close completely and leave a few old cratons surrounded by compressed island arc material and occasional oceanic crust. Detailed descriptions of the geologic development of Afro-Arabia can be reviewed in the following articles: Kennedy (1964), Clifford (1970), El Shazly (1977), Youssef (1968), Gass (1977), Engel et al., (1979, 1980), Greenwood et al. (1976), Fleck et al. (1976), Stern (1979), Dixon (1979, 1981), Greenberg (1978), Rogers et al. (1978); Marzouki and Fyfe (1977), Neary et al. (1976), Naseef and Gass (1977), Gass

Figure 45.

Map of the general tectonic configuration of Africa:
Three Precambrian cratons surrounded by younger,
largely Pan-African crust; stippled pattern represents Cenozoic magmatism diagonal lines indicate Hercynian deformation; "X" marks location of the Egyptian alkaline province within Pan-African crust (after Thorpe and Smith, 1974).



After Thorpe & Smith, (1974).

Figure 45.

et al. (1978), Bakor et al. (1976), Garson and Krs (1976), Garson and Shalaby (1975), Kröner (1977a, b, 1979), Kroner et al. (1979), Vail (1976), Swartz and Arden (1960), Said (1962), Brown and Jackson (1960), Brown (1970), Brown and Coleman (1972), Al Shanti and Mitchell (1976), Shackleton (1973), Thorpe and Smith (1974), Burek (1969), Abdel-Gawad (1969), Nur and Ben-Avraham (1978), Hancock and Kadhi (1978), Le Pichon and Francheteau (1978), Richardson and Harrison (1976), Pilger and Rösler (1976), Dewey et al. (1973), Moore (1979), Sengör (1979), Sengör et al. (1980, 1982), Bergougnan and Fourquin (1982), Laubscher and Bernoulli (1977), Al-Shanti (1980), Hadley and Schmidt, 1978).

The majority of the Afro-Arabian crust was formed during the Pan-African event by a cratonization process involving oblique compression and accretion of oceanic and island arc systems, shallow-water sedimentation, and intense plutonism (Engel et al., 1980; Gass, 1977). One of the most obvious indications of this cratonization is the presence of four or five ultramafic ophiolite zones which cross the Eastern Desert in a NW-SE direction, Figure 46 (see Neary et al., 1976; Bakor et al., 1976; Gass, 1977; Stern, 1979). This Pan-African geologic record in the Eastern Desert consists of a tholeiitic substratum with overlying immature volcanogenic metasediments and calc-alkaline volcanics, all of which were intruded by very deep-seated, primitive quartz diorites, quartz monzonites,

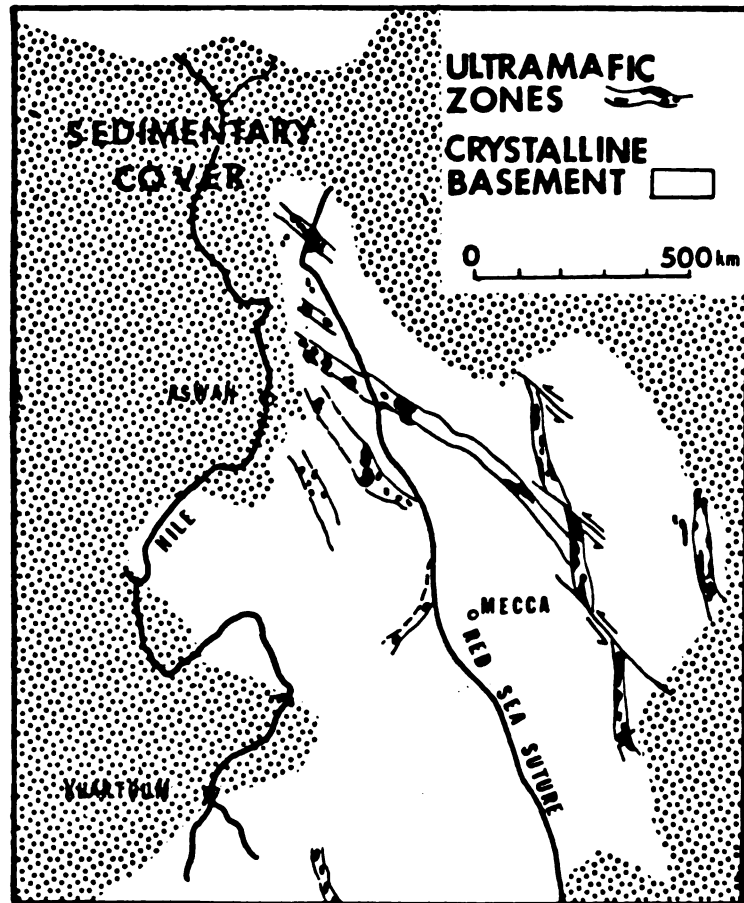


Figure 46.

Map showing palinspastic reconstruction of the Red Sea area and the distribution of the Pan-African ultramafic zones (ophiolites), (after Neary et al., 1976).

and LIL-enriched granites (Dixon, 1979; Greenberg, 1978; Boulos et al., 1979). This conversion from an oceanic setting to a thickened continental crust appears to have occurred from about 880 to 550 m.y. B.P. (Engel et al., 1980).

The structural deformation of this oblique compression and accretion was characteristic of the Pan-African tectonics throughout many parts of Africa, with the lack of large magnitudes of compressional deformation, and the predominance of strike-slip faulting. Oblique compressional tectonics have been shown to generate extremely effective strike-slip fault systems in both ancient and recent crustal settings (e.g. Benioff, 1962; Fitch, 1972; Mitchell and McKerrow, 1975; Phillips et al., 1976; Arthaud and Matte, 1977; Bak et al., 1975; Tapponnier and Molnar, 1976, 1979; Molnar and Tapponnier, 1975; Nur and Ben-Avraham, 1978; Watterson, 1978). Watterson (1978) emphasized the importance of these primary compressional shear zones which instill a strong anisotropic structural grain in the crust and provide preferential planes of weakness in subsequent tectonic activity. This style of deformation in the Pan-African culminated in major strike-slip dislocations in the Arabian shield during the late Proterozoic and the early Phanerozoic times (Brown and Jackson, 1960; Moore, 1979). Brown (1972) indicates that up to 240 kilometers of lateral displacement occurred along the Najd transcurrent fault system which Fleck et al.

(1976) estimate occurred over a period of about 50 m.y. at around 550 m.y. B.P. (see Figure 44). The magnitude of this fault is comparable to the Great Glen fault of Scotland and other major transcurrent fault systems (see Moore, 1979). Though more work is needed on the structural history of the Eastern Desert, a palinspastic reconstruction of the present Red Sea area makes it likely that the Najd fault system continues into the northern portion of the Eastern Desert (see Moore, 1979).

Strike-slip deformation did not only occur in the early history of Afro-Arabia. It was important throughout the Phanerozoic up until even recent times (Nur and Ben-Avraham, 1978; Brown, 1970; Swartz and Arden, 1960; Youssef, 1968; Abdel-Gawad, 1969; Burek, 1969; Garson and Krs, 1976; Moore, 1979; Hancock and Kadhi, 1978). Garson and Krs (1976), for example, emphasize the large-scale strike-slip faulting during the Late Cretaceous along the Precambrian lineaments (e.g. 100 kilometers of lateral displacement). Paleomagnetic evidence does not contradict such lateral displacement. On the contrary, such offsets may allow a better solution for the opening of the Red Sea (Le Pichon and Francheteau, 1978; Richardson and Harrison, 1976).

Recent work is beginning to decipher some of the complex history of the eastern Mediterranean and Afro-Arabian tectonics (e.g. Sengör, 1979; Hsü et al., 1977; Hsü, 1977; Laubscher and Bernoulli, 1977; Dewey and Sengör, 1979; Dewey et al., 1973; Nur and Ben-Avraham, 1978). The

pattern which is emerging from such work is the occurrence of large-scale strike-slip tectonics induced by oblique convergence and irregular continental margins (Dewey and Sengör, 1979; Hsü, 1977; Nur and Ben-Avraham, 1978). A consistent result of this style of plate interaction appears to be the "splintering" of the edges of the converging plates, back-arc spreading, and the formation of microplates. Thus, both compressional and extensional features are often recorded in close proximity in space and time (e.g. Hsü, 1977; Sengör, 1979; Laubscher and Bernoulli, 1977; Stocklin, 1968). A continuation of this intricate compression-transform-extension tectonics may be seen in the Zagros-Levantine-Red Sea activity of the Tertiary (e.g. Nur and Ben-Avraham, 1978) and that of the Aegean region in more recent times (Dewey and Sengör, 1979).

Although much of the detailed structural and tectonic history of Afro-Arabia is unknown, there is some evidence which may support this general style of tectonics since at least the end of the Paleozoic era (Sengör, 1979). Since this time, the African plate has been moving generally northward and rotating counter-clockwise (e.g. Laubscher and Bernoulli, 1977). This general movement continued throughout the spreading of the Atlantic and Indian Oceans, until the relatively recent Red Sea extension. This resulted in a dominant compressional stress in northeast Africa and Arabia during much of the Phanerozoic (Dewey et al., 1973). An important mechanism in this compression is considered to

be the oblique convergence of Africa and Eurasia with the closing of the Tethyan ocean basins and the corresponding opening of the Atlantic and Indian Oceans, Figure 47. This oblique convergence may have been accentuated by the irregular continental margin of the Arabian Promontory. The oblique compression and subduction of the Tethyan oceanic crust may have resulted in the reactivation and further propagation of the fractures and strike-slip faults within the Arabian Promontory and the adjoining Nubian Shield. The existence of the major Pan-African dislocations in Afro-Arabia coupled with these types of regional forces may help to explain the persistence of strike-slip tectonics throughout much of the history of Afro-Arabia.

A major point to note in relation to the small amount of data available on the tectonic development of the Eastern Desert is the importance of active intraplate zones of weakness. All the Egyptian alkaline complexes appear to lie along the major lineaments of the Nubian shield, Figure 44 (El Ramly et al., 1969a, 1969b, 1971; Garson and Krs, 1976; Lutz, 1979; Serencsits et al., 1981). The long-term reoccurrence of alkaline magmatism along these lineaments and the indications of multiple reactivations of these planes of weakness throughout the Phanerozoic, may be indicative of an intimate relationship between the alkaline magmatic process and this style of intraplate deformation.

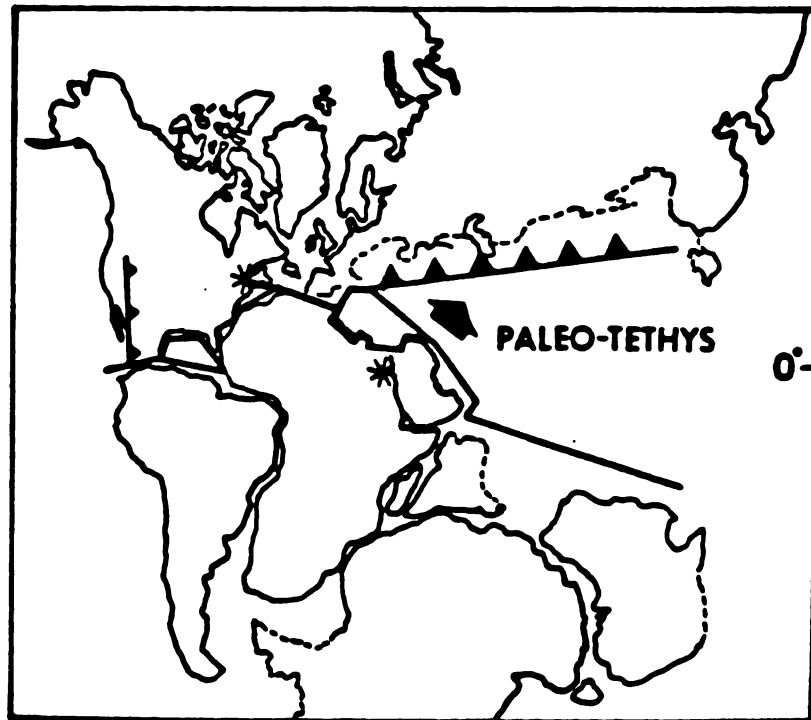


Figure 47.

Map showing paleopositions of the major continents during the Triassic, about 220±20 m.y. B.P. Note that the opening of the Atlantic will be accompanied by a corresponding closure of the Paleo-Tethys. Heavy black line represents the east-west Mesozoic break-up of Pangea; barbed lines represent Mesozoic subduction zones; "*" marks the alkaline provinces of Egypt and the White Mountain Magma Series.

(After Laubscher and Bernoulli, 1977).

2) The Episodes of Alkaline Magmatism in Egypt

Figure 48 illustrates a schematic interpretation of the development of Afro-Arabia and the emplacement of alkaline melts along the major fractures in the Nubian shield. These fractures are considered to have been initiated during the formation of this crust due to the oblique compressional tectonics of the Pan-African event. The complex interaction of the Tethyan, Eurasian and African plates caused the continued reactivation of these fractures. The relative rotation of the African plate and the irregular geometry of the leading continental margin of the Arabian Promontory both may have induced continued oblique compressional tectonics and a mechanism by which these reactivations occurred. The episodes of similar alkaline magmatism in the Eastern Desert are considered to have coincided with major shifts in plate motion. It is proposed that these alkaline melts were products of shear heating between adjacent volumes of the asthenosphere as the slow, solid-state convection of the mesosphere adjusted to the differential motion of the overlying plates, Figure 49. Their generation reflected the thermal disequilibrium between the lithosphere and the mesosphere during these plate accelerations. Thus, shifts in the motion of these plates caused the generation of small amounts of alkaline melt in the asthenosphere, and also aided in their accessibility to the surface by promoting the reactivation of the fracture system throughout the

Figure 48.

A schematic illustration of the development of Afro-Arabia:

- a) Oblique subduction-accretion, formation of Pan-African Afro-Arabian crust, the envelopment of oceanic material (ophiolites), the development of strike-slip fault systems (splaying etc; e.g. Najd transcurrent fault system), and a fracture pattern; emplacement of alkaline magma near the end of this Pan-African event along the Najd shear zone (Wadi Dib, 554 m.y. B.P.).
- b) Periods of reactivation and strike-slip faulting during major shifts in plate motion throughout most of the Phanerozoic eon (dominated by plate-wide African-Eurasian convergence), and the emplacement of alkaline magma along these zones of weakness--especially at their intersections.
- c) Late Cretaceous strike-slip deformation and the emplacement of alkaline magma during the intense convergence in northeastern Arabia, Iran, and Anatolia, followed by doming (not shown), mantle up-welling, and rifting along intensely weakened shear zones (Red Sea).

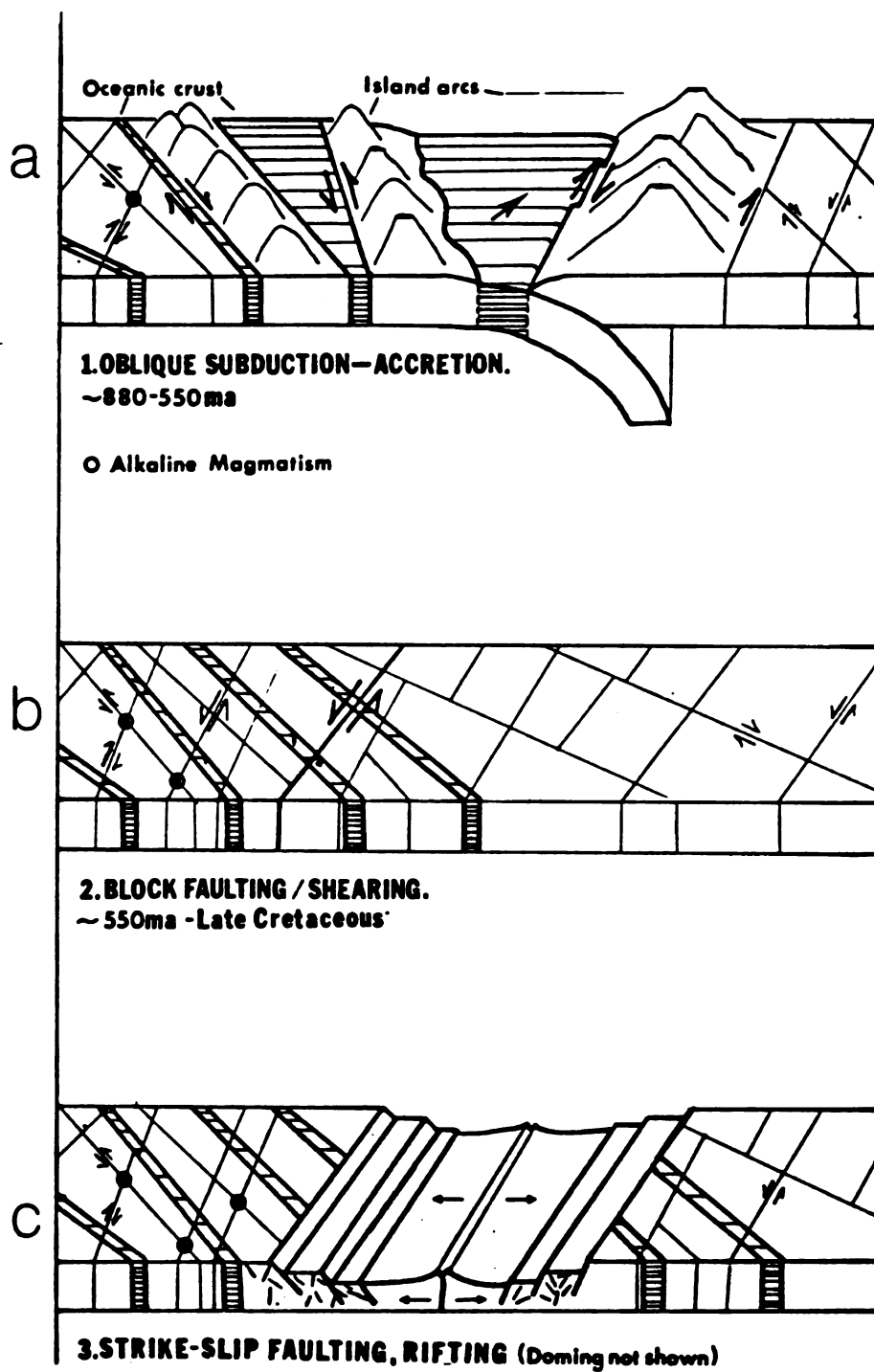


Figure 48.

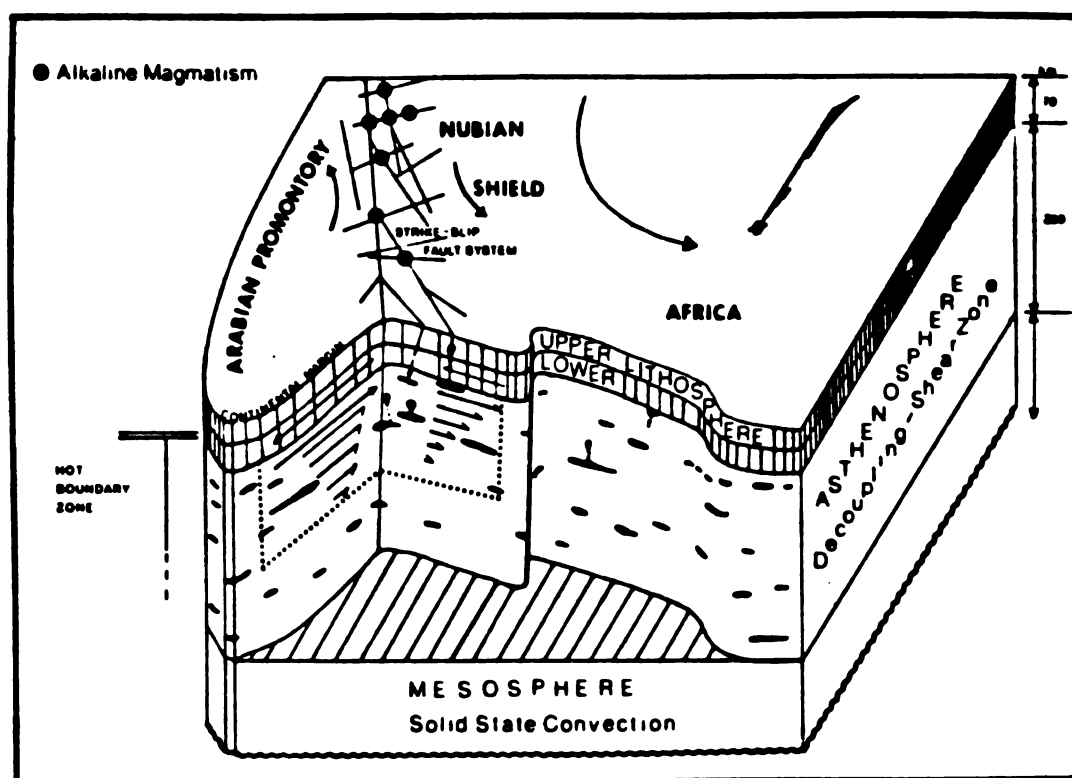


Figure 49.

Model of alkaline melt production via shear heating in the asthenosphere in response to major shifts in the motion of the African plate and their ascent along the Pan-African fractures. Reactivations of these fractures and the ease of ascent of the alkaline melts may have been especially favorable in Afro-Arabia due to shearing and strike-slip faulting between the Arabian Promontory and the Nubian shield. Such deformation may have been largely the result of the oblique convergence between the African and Eurasian plates (oblique subduction-transform motion-extension of the numerous Tethyan ocean basins), as well as the irregular continental margin of the Arabian Promontory. Dotted area outlines a general region of concentrated shear dissipation and melt release. No scale intended.

Phanerozoic eon. This reoccurring activation and intrusion of these fractures may have led to a significant weakening within the plate and the eventual formation of the recent Red Sea plate boundary at an oblique angle with respect to the Pan-African ophiolite suture zones, Figure 48c (and refer to Figure 46).

A. Thermal Setting

The early emplacement of alkaline magma at Wadi Dib (554 m.y. B.P.) in the last compressional stages of the Pan-African subduction-accretion process may reflect the relatively low thermal gradient characteristic of the Pan-African tectono-thermal event (Greenwood et al., 1976; Gass, 1977), and the lack of a continent-continent collision in the formation of Afro-Arabia (Kroner et al., 1979). Throughout the following numerous reoccurrences of alkaline magmatism in the Eastern Desert of Egypt, there appears to have been an overall change in the degree of alkalinity and silica-saturation, see Figure 3. The older complexes are consistently silica-oversaturated and relatively low in alkalis, whereas the younger complexes are richer in alkalis and contain silica-undersaturated mafic and salic rock-types as well as silica-oversaturated salic rock-types. This change in the compositions of the alkaline melts over most of the Phanerozoic eon is interpreted to be a function of the decreasing thermal gradient after the formation of the Afro-Arabian Pan-African crust. With the general

decline in the thermal gradient and a gradual thickening of the lithosphere, it is suggested that the localization of the effective zones of shear heating during subsequent periods of plate acceleration would migrate to progressively deeper levels and contain smaller degrees of melting.

According to high pressure experimental work (e.g. Green, 1970), this would lead to more silica-undersaturated alkaline melts. This long-term chemical trend throughout the evolution of the Egyptian alkaline province is interpreted to further substantiate that the mechanism responsible for these melts was not restricted to any single type of tectonic event. Rather, it is proposed that the Egyptian alkaline melts were generated as a result of a similar deep shear heating mechanism in response to any type of major shift in plate motion, within an overall tectonic regime of increasing stability and decreasing thermal gradient.

B. Tectonic Setting

There are four general approaches which support the coincidence of the episodes of alkaline magmatism with periods of regional structural readjustment and major shifts in plate motion:

1. limited information on the structural history of Afro-Arabia,
2. Africa's polar wandering path which may be interpreted as an indication of its motion since Pan-African times, Table 6,

Table 6 . Changes in plate motion associated with episodes of alkaline magmatism.

<u>Periods of Tectonic and Magmatic Activity</u>	<u>Periods of Rapid Polar Wandering (Africa)¹ (Myr)</u>	<u>Rapid Changes in Sea Level² (Myr)</u>	<u>Periods of Relative Motion Between Africa and Europe^{3-5,17} (Myr)</u>	<u>Periods of Alkaline Magmatism in Egypt⁶⁻⁸ (Myr)</u>
Pan-African event of Eastern Desert, Egypt: 1000-550 m.y.B.P. ⁹⁻¹¹ . Large scale transcurrent faulting (Wajd fault system, active 580-530 m.y.B.P. toward end of Pan-African event or "Hijaz Orogenic Cycle"), ^{12,13,14} .	680-660	480 410-390		554
Initial break-up of Pangea ¹⁵ leading to extension of W. America-Africa/ S. America. Overall compression and subduction of Paleo-Tethys north of Afro-Arabia; possible back-arc type spreading of "Cimmerian" land mass from NE portion of Gondwana (including Afro-Arabia) and formation of the Neo-Tethys ¹⁶ . Large-scale dextral shear (minimum of 3500 km) between Laurasia and Gondwana ¹⁷ .	420-220	350 330-320 280 220-185	250-200 ¹⁷	351 229,223,221
Initial rifting of the South Atlantic: Africa-S. America and Benue Trough extension ¹⁷ . Corresponding compression in Afro-Arabia and Tethyan domain, strike-slip faulting, closely followed by extension between Iran and Arabia ^{18,19} and possible separation of the Anatolian plate from Afro-Arabia ²⁰ .		140	165-80 (Mid-Cret. to Late Cret.)	145,142,140 139,132
Major shift in spreading of the South Atlantic and associated second major episode of alkaline magmatism 80-90m.y.B.P. ²¹⁻²⁴ . Large-scale strike-slip faulting in Afro-Arabia ²⁶⁻³² , intense compression in Iran and Turkey with associated ophiolites ^{18,19,33,34} , emplacement of Oman ophiolites in NE Arabia ³⁵	110-40	95 60	80-40 (Late Cret. to Early Eocene)	91,89
Red Sea coming and extension ^{36,37} .	40-0 (Early Eocene)	30		recent alkali magmatism associated with Red Sea opening ³⁸ .

1. Briden and Gass, 1974.
2. Vail et al., 1977.
3. Smith, 1971.
4. Heü, 1971.
5. Diets and Horden, 1970.
6. Lutz, 1979.
7. Serencsits et al., 1981.
8. Hashad and El Reedy, 1979.
9. Gass, 1977.
10. Engel et al., 1979.
11. Engel et al., 1980.
12. Moore, 1979.

13. Fleck et al., 1976.
14. Brown, 1972.
15. Gass et al., 1978.
16. Sençür, 1979.
17. Morel and Irving, 1980.
18. Dewey et al., 1973.
19. Alavi, 1980.
20. Laubecher and Bernoulli, 1977.
21. Marsh, 1973.
22. Sykes, 1978.
23. Opdyke and MacDonald, 1973.
24. van Breemen and Bowden, 1973.

25. El Ramly et al., 1971.
26. Yousoef, 1968.
27. Swartz and Arden, 1960.
28. Burak, 1969.
29. Abdel-Gawad, 1969.
30. Hancock and Kadhi, 1978.
31. Brown, 1970.
32. Garoon and Kra, 1976.
33. Heü, 1977.
34. Pinar-Erdem and İlhan, 1977.
35. Glennie et al., 1973.
36. Pilger and Möller, 1976.
37. Le Pichon and Francheteau, 1978.
38. Rasoatar et al., 1981.

3. motion of the adjacent oceanic plates which may be applied to the past 180 m.y. B.P., Table 6,

4. global sea level curves, Table 6. Unfortunately, only the second-order, smoothed curves have been published (Vail et al., 1977). The detailed global sea level curves are held as proprietary information by Exxon Production Research Company. With the exception of the earliest episode of alkaline magmatism in Egypt (554 m.y. B.P.) where the curve is supported by very little data, all the episodes of alkaline magmatism correspond to rapid changes in the curve. The episodes of alkaline magmatism at 140, 351 and 404 m.y. B.P. coincide exactly with major inflections in the global sea level curve. The episodes at 90 and 224 m.y. B.P. correspond to periods of numerous rapid changes leading toward a major peak in marine transgression. Obviously, more study is needed on the use and significance of these curves, however, one of the most important mechanisms for large changes in sea level is rapid tectonic movement. Assuming that these rapid changes in the curve correspond to large-scale tectonism, then these observations are consistent with a model in which alkaline magmatism tends to occur during major plate accelerations.

The first known alkaline magma in Egypt (Wadi Dib, 554 m.y. B.P.) intruded the northeastern portion of the Nubian shield during the final oblique compressional stages of the Pan-African event in the Eastern Desert. It is suggested that this alkaline magma was emplaced within

the large-scale strike-slip tectonics of the Egyptian portion of the Najd transcurrent fault system. This fault system was undergoing about 240 kilometers of lateral displacement during the same time as the emplacement of this early alkaline magma (see Moore, 1979) Figure 44.

Two alkaline complexes dated at 404 and 351 m.y. B.P. have uncertain tectonic events associated with their emplacement, but similar to the alkaline magmatic episodes that follow, they coincide with periods of rapid change in the inferred ocean levels around the world, Table 6. This correspondence is taken as an indication of abrupt changes in the tectonic pattern of the plates--changes which affected the ocean basin volumes and coincided with the outbreak of alkaline magmatism.

The 224 m.y. episode of alkaline magmatism occurs at the end of the second rapid shift in the motion of the African plate, Table 6. The general period of time marked the breakup of Pangea (Gass et al., 1978), with extremely large amounts of dextral shearing between Laurasia and Gondwana (a minimum of 3500 kilometers over a period of time from 250 to 200 m.y. B.P., Morel and Irving, 1981; see also Arthaud and Matte, 1977). Swanson (1982) considers large-scale dextral shearing as an early transform history for the initiation of the central Atlantic rift (see also Cherkis et al. 1973). It is in this period of time that the Paleo-Tethys began closing with the proposed back-arc spreading of the Neo-Tethys (forming the "Cimmerian"

landmass, Sengör, 1979; see also Stocklin, 1965; Tekeli, 1981). The Levantine transform boundary along the Arabian Promontory may have also been activated during this period of rapid plate movement (Dewey et al., 1973).

The 140 m.y. and 90 m.y. episodes of alkaline magmatism in Egypt coincide with the Jurassic and Cretaceous shifts in the relative motion between Africa and Europe, Table 6. These shifts have been inferred from the conjugate plate motion involved in the spreading of the Atlantic and the closing of the Tethyan domain (Dietz and Holden, 1970; Smith, 1971, Hsü, 1971). The late Jurassic episode occurred during the final closing and collision of the Paleo-Tethys (Sengör, 1979), with the corresponding opening of the South Atlantic (Pitman and Talwani, 1972). The Late Cretaceous episode occurred during the hairpin turn in the African polar wandering curve, Table 6, and a major shift in the spreading of the South Atlantic (Sykes, 1978). This was accomplished by corresponding compressional events in the Tethyan area. This involved vigorous strike-slip faulting in Afro-Arabia during the Late Cretaceous (Garson and Krs, 1976; Youssef, 1968; Swartz and Arden, 1960; Burek, 1969; Abdel-Gawad, 1969; Hancock and Kadhi, 1978; Brown, 1970), the obduction of the Oman ophiolites on the north-eastern margin of Arabia (Glennie et al., 1972) as well as the Late Cretaceous ophiolites in Anatolia (Pinar-Erdem and Ilhan, 1977), and the intense compressional deformation in Iran (Hsü, 1977; Alavi, 1980; Laubscher and Bernoulli,

1977; Dewey et al., 1973; Stocklin, 1968). Again, there is a correspondence between these episodes of alkaline magmatism and the sharp inflections in the proposed global sea level curves.

The coincidence of alkaline magmatism in widely separated areas is well illustrated in the case of the Egyptian and Atlantic alkaline provinces. The 224 m.y. episode of alkaline magmatism in Egypt coincides with the first known major episode of alkaline magmatism in the White Mountain Magma Series (Foland and Faul, 1977) along the crustal lineaments within the (still) adjoining North American continent. Figure 47 gives an indication of the relation of this North American tectonic lineament with Africa, possibly along the embryonic New England Seamount Fracture (Dewey et al., 1972; Foland and Faul, 1977), the South Atlas shear zone (Arthaud and Matte, 1977; Dewey et al., 1973), and the major plate boundaries between Africa and the Tethys (Dewey et al., 1973). These tectonic boundaries appear to have been the focal point of the tremendous strike-slip plate movement between Laurasia and Gondwana during the same period of time as this alkaline magmatism (Morel and Irving, 1981: a minimum of 3500 kilometers of dextral displacement) and may represent an early strike-slip stage of the Atlantic (Swanson, 1982).

The 140 m.y. and 90 m.y. episodes of alkaline magmatism in Egypt coincide with the two major episodes of alkaline magmatism occurring in the areas surrounding the

South Atlantic. Alkaline magmatism around 140 m.y. B.P. in Brazil, Uruguay, Bolivia, and western and southern Africa has been related to the initial rifting of Africa from South America (Marsh, 1973; Herz, 1977; Sykes, 1978; Neill, 1973; Darbyshire and Fletcher, 1979; Amaral, et al., 1967).

A second episode of alkaline magmatism in many of these areas around 80-90 m.y. B.P. has been correlated with a major shift in the spreading of the South Atlantic during the Late Cretaceous (Marsh, 1973; Opdyke and MacDonald, 1973; Herz, 1977; Sykes, 1978; Boyd, 1973).

Thus, periods of rapid convergence and associated strike-slip motion with the closing of the Tethyan basins coincide with conjugate stages of opening and transform motion in the development of the Atlantic. Considering this reciprocal plate motion, the synchronous emplacement of these alkaline melts is not likely to be the result of unrelated mantle disturbances. It is proposed that shear heating in the asthenosphere due to these major changes in plate motion is a more likely explanation. Depending on the presence and activation of major fractures, the localization of possible concentrated shear dissipation in the underlying asthenosphere, and the lack of interference or mixing from larger degrees of non-alkaline melts, alkaline magmatism may occur simultaneously in several portions of the world in response to the accelerative motions of the plates.

APPENDICES

APPENDIX I

APPENDIX I

DESCRIPTION OF FOUR EGYPTIAN ALKALINE COMPLEXES

- 1) Gabal El Naga: Latitude $22^{\circ}42'00''\text{N}$, Longitude $34^{\circ}28'25''\text{E}$; Elevation: 827 meters; Height above surrounding wadi: 300 meters; Maximum Diameter: 4 kilometers; Age: 142 m.y. B.P. (Lutz 1979).

Gabal El Naga is the most unique of the four complexes in this study. It alone contains amphibole as the dominant ferromagnesian mineral.* It also lacks the extremes of quartz-rich and highly silica-undersaturated rock-types seen in the other complexes. Even so, the more silica-undersaturated intrusive rock-types occur in the central portions of the complex. The difference in silica-saturation levels and the predicted chemical gradient would simply be less dramatic than in the other complexes. Unlike the other complexes, the majority of the rocks at El Naga have a less altered appearance. This is especially noticeable for the volcanics, which are fresh at El Naga and extremely altered in most of the flows at the other complexes.

G. El Naga is an extremely well-formed ring structure made up of an outer and inner ring, with the more stock-like

*El Ramly et al., 1969b, identify the dominant ferromagnesian mineral as aegerine-augite.

syenite bodies centered about the inner wadi. The country-rock is dominantly gneissic country rock, often with porphyroblastic textures. This study found that the outer ring consists of predominantly syenite rather than the gneissic country rock described by El Ramly et al. (1969b). The large zones of fenitization mapped by El Ramly et al. (1969b) were identified as syenites by this study. On the contrary, apart from narrow zones of alteration coinciding with fault zones and a rather altered condition in the syenites of the lower portions of the outer ring, El Naga contains much less evidence of widespread metasomatism than the other complexes visited.

The outer ring consists of silica-oversaturated syenites: hard, partially silicified, fine crystalline textures in the upper part of the ring; softer, coarse crystalline and yellow-stained syenites in the lower part of the ring. The outer ring is dissected by hydrothermally altered fault zones which El Ramly et al. (1969b) found to extend into sheared, granulated zones in the porphyroblastic gneissic country-rock. The inner ring contains three major intrusive rock-types capped by older, silica-oversaturated trachy-basalts. The oldest intrusive rock (cross-cutting relationships) is a rather enigmatic variety not seen at the other alkaline complexes, and called a "hybrid" rock by El Ramly et al. (1969b). This is a silica-oversaturated fine to medium-grained syenodiorite, which has a fused, recrystallized appearance in thin-section. It may be related to the

early magmatic phase accompanying the eruption of the trachybasalts. The remaining two types of intrusive rock in the inner ring are silica-undersaturated and are much more abundant than the small zones of "hybrid" or syenodiorite. The older variety (cross-cutting relationships) consists of extremely coarse-grained alkali feldspar laths, euhedral amphibole, large biotite books, and nepheline (called umptekite by El Ramly et al. 1969b). The later variety is more silica-undersaturated, contains more nepheline, and some interstitial analcime or sodalite.

The lack of widespread metasomatism, the crystallizations of amphibole rather than pyroxene, and lack of a strong silica-undersaturated to oversaturated gradient from the inner to outer portions of the complex may all be related to the retention of this magma's vapor phases and the lack of an effective transport medium for silica and other components.

- 2) Gabal Nigrub El Fogani: Latitude $25^{\circ}51'29''\text{N}$, Longitude $34^{\circ}56'49''\text{E}$; Elevation: 1078 meters; Height above surrounding wadi: 500 meters; Maximum Diameter: 6 kilometers; Age: 139 m.y. B.P. (Lutz 1979).

Gabal Nigrub El Fogani is the most intensely altered complex of the four complexes sampled. It alone contains carbonatite dikes and net vein complexes. It also contains the largest amount of volcanic rocks.

This is a highly eccentric ring structure consisting

of an incomplete outer ring which is fragmented in the southern portion, and a large central intrusive stock in the northeastern part of the complex, next to the eastern side of the outer ring. The surrounding country-rock is granitic and amphibole gneiss. Extremely quartz-rich syenites (alkaline granites) compose the outer and top portion of the outer ring. From the textures, much of the quartz appears to be secondary in origin. The inner side of the outer ring consists of silica-saturated or slightly undersaturated syenites which El Ramly et al. (1969b) call leucocratic essexites. This same rock-type abuts the central stock on the eastern side of the complex. The central stock is made of pink, intensely-altered nepheline syenites. This alteration has contains ferrugination and liebertinized nepheline (adds to the pink coloration) to the extent of reducing some outcrop slopes to a pink powder. From rather obscure crystal relationships in thin-section, much of this central stock may have contained a large amount of sodalite which has since been destroyed. These central syenites appear to have crystallized last, and since the surrounding earlier rock-types do not display similar intense alteration, it is inferred that most of this alteration of the nepheline syenites was caused by autometasomatism. This high abundance of fluid phases, carbonatite dikes, voluminous effusive rocks, and the inferred efficient silica-transport toward the outer margins of the complex all may be related to the effective

separation of this magma's vapor phases via discrete separation and volatile transfer.

- 3) Gabal El Kahfa: Latitude $25^{\circ}8'18''\text{N}$, Longitude $34^{\circ}38'55''\text{E}$; Elevation 1018 meters; Height above surrounding wadi: 120 meters; Maximum Diameter: 4 kilometers; Age: 91 m.y. F.P. (Lutz 1979).

Gabal El Kahfa contains evidence of explosive magmatic activity and widespread alteration which is different than that found at Nigrub El Fogani. El Kahfa is very similar to Gabal Abu Khruq in mineralogy, and representation of rock-types. However, it does not have large amounts of foyaites (sodalite-nepheline syenites).

This complex consists of a wide horseshoe-shaped outer ring which opens to the south. Between these southern terminations of the outer ring lies the central stock. The country-rock is amphibole schist which is often highly altered around the complex. Volcanics which are inferred to be comagmatic, and gabbros also occur around the ring structure. Both rock-types are intensely altered. (The highly altered gabbros were included as "epidiorites" or country rock by El Ramly et al., 1969b.) Once again, the outermost syenites are the most silica-oversaturated, the syenites on the inside of the outer ring (although very altered) are near silica-saturation, and the central stock contains highly silica-undersaturated syenites.

The outer ring consists of predominantly altered quartz syenites with fused textures, brown-stained alkali feldspar laths with vermicular crystal boundaries, corroded aegerine-augite subhedral crystals with associated opaque alteration products, and commonly blastic textures with the graphic quartz. The inner portion of the outer ring is a wide cataclastic zone made up of alkaline syenite with noticeably less quartz and altered essexite gabbros. These cataclastic zones consist of extremely broken-up rock--from large boulder-size blocks to fine granulated rock material--zones of shearing and hydrothermal alteration. This physical and chemical alteration occurred before the emplacement of the late-stage syenite dikes which cross-cut these cataclastic zones.

The inner stock consists of a small proportion of nepheline syenite. These nepheline syenites are surrounded by a predominance of essexite gabbros which are similar to the highly altered gabbros outside the ring structure, although the inner gabbros are less mafic. Considering this close association in the central stock between the gabbros and nepheline syenites, a special attempt was made to distinguish the contacts. Either these rock boundaries are obscure for some unidentified reason, or the contacts are gradational.

Within the central stock are several plugs which El Ramly et al. (1969b) refer to as "ore essexite diabases". These rock bodies have sheared and altered contacts with

the surrounding gabbros. They consist of fractured and granulated mafic material which are considered to represent possible diatremes of comagmatic ultramafic cumulates from the deep portions of the complex's substratum. The matrix is recrystallized or blastic alkali feldspar and plagioclase, scattered throughout with fractured euhedral pale-colored or colorless pyroxene (diopsidic), subhedral olivine, and biotite. Close by these possible xenolithic plugs are volcanic vent rocks (agglomerates and breccias) forcibly intruded (unusual "squirted" appearance) into the surrounding coarse-grained rock-types. There are cavities in these volcanics which contain free-surface quartz crystals. Some of the volcanic rocks include very mafic compositions. The exact structural relationship of these remnant flows to the surrounding intrusive rocks and xenolithic bodies is uncertain. However, field relationships allow the possibility that the olivine-diopside xenolith "plugs" and the cataclastic zones occurred in a single violent event, before the termination of the magmatic activity at El Kahfa. They may represent an explosive episode involving a build-up and release of magmatic vapor phases which aided in breaking up the syenites and gabbros, forcibly intruding the xenolithic cumulate material from an inferred ultramafic substratum, as well as causing widespread chemical alteration within and outside of the complex. This intense alteration is reflected in the chemical data on El Kahfa.

4) Gabal Abu Khruq

A. General Petrography of the Rock-Types at Abu Khruq Wehrlite

This study is not aware of any previous reports of this diopside-olivine rock from any of the Egyptian alkaline complexes. The sample was collected from a small outcrop on the outside of the northeastern portion of the outer ring, refer to Figure 2. The structure of this outcrop does not suggest that it is a dike. It may be a xenolithic mass brought up from the lower portions of the complex, or it may represent a partially exposed, in situ outcrop of an ultramafic substratum below Abu Khruq.

This rock-type displays a cumulate texture, it is highly serpentized, and is composed of predominately euhedral diopsidic-pyroxene crystals (about 60-70%, smaller amounts of fractured euhedral ferrosilite-olivine crystals (about 10-15%), small tabs of interstitial plagioclase (about 10%), some scattered nepheline crystals (less than 5%), and a widespread serpentized alteration/groundmass material. The presence of the nepheline crystals suggests the comagmatic relationship of this ultramafic rock-type, and the existence of an early silica-undersaturated alkaline magma (the REE contents of this rock-type are unexpectedly high). Drilling or geophysical techniques may be required to investigate the possibility of ultramafic cumulate layers below Abu Khruq and other Egyptian alkaline complexes.

Gabbros

The gabbros represent the most common basic rock-type at Abu Khruq. Chemical data indicates a comagmatic relationship as plagioclase-rich cumulates of an intermediate syenitic liquid.

The exposures of gabbro both outside and inside the ring-structure, and their apparent lack of conformity to the conical (ring) faults, makes them structurally different than all the syenites. El Ramly et al. (1969a) suggest that this relationship indicates the gabbros were emplaced before the syenites and the development of the ring-structure. They also indicate that there is evidence for the emplacement of the gabbros after the formation of the volcanic cone. This sequence would have great implications on the petrogenesis of the magma at Abu Khruq.

There are few areas where the gabbros show sharp contacts with the other rock-types, and these often can be shown to be dikes rather than the syenite stocks. The major rock boundaries often are obscured by colluvium, ring faults, and/or are gradational with respect to the syenites. (As at El Kahfa, considerable time was spent attempting to locate sharp contacts between the gabbros and syenites without success.)

The gabbros display coarse, ophitic, slightly miarolitic textures, and are composed of mostly labradorite laths (An_{66-52}), a second generation of smaller albitized rectangular plagioclase crystals (total plagioclase:

50-65%), and diopsidic augite (20-30%). The labradorite has well-developed albite twinning, occasional percline and albite twinning, often non-complex zoning, and frequent thin accumulus albite overgrowths. The pyroxene occurs interstitially with respect to the labradorite; the flesh-colored subhedral crystals often form large continuous "chains" of pyroxene around the plagioclase laths. Olivine is rare and always has a reaction rim of pyroxene \pm amphibole. There are very small amounts of orthoclase, occurring as small interstitial tabs in some of the gabbro specimens. A common mineral which can represent up to 8% of the rock is the Fe-rich biotite (stilpnomelane). This occurs as large irregular, red sheets, most often in association with alteration products of amphibole and opaques. An important minor mineral to note in the gabbros is nepheline. It occurs as small hexagonal and square euhedral crystals within and around the pyroxene. Thus, nepheline proceeds the crystallization of pyroxene. Apatite represents another important mineral phase as it can amount to almost 5% of the gabbros. The gabbro crystallization phase may represent a significant proportion of the P_2O_5 in the original melt. It occurs mostly as poikilitic euhedral crystals within the pyroxene, or immediately adjacent to the pyroxene grain boundaries. There are scattered accessory minerals, the most common of which are large subhedral grains of sphene, and small spinel euhedral crystals.

Although the gabbros, similar to all the major rock-types at Abu Khruq, contain certain amounts of alteration, the fundamental mineralogic and chemical character of these rocks has been preserved. The feldspars are commonly partially altered to cancrinite and sericite. Similar to the syenites, albitization is widespread, and in the apparent extreme cases, the grain boundaries of the plagioclase laths become wavy or vermicular. The pyroxenes often alter to amphibole, especially uralite and associated Fe-oxides. The Fe-rich biotite has been occasionally converted to chlorite or epidote, but the most common is its conversion to opaque Fe-oxides. Zeolites are present in the interstices of the more altered samples.

Volcanics

All the volcanics sampled from Abu Khruq have trachytic or rhyolitic compositions; there do not appear to be any basaltic flows. They are predominantly silica-oversaturated (one analyses from El Ramly et al., 1969a is slightly silica-undersaturated, 1.84 normative ne), and most samples contain quartz. The volcanic samples are typically altered, and based on the consistency and nature of this alteration, it is likely that autometasomatism was largely responsible. This cause for alteration is not unusual in many other peralkaline and alkaline volcanics.

Generally, the volcanics consist of 10-30% phenocrysts. These are dominated by large, partially resorbed alkali

feldspar phenocrysts with irregular, corroded boundaries which are commonly albitized and leached into the matrix material. These alkali feldspars display faint Carlsbad twinning, and are otherwise structureless. Often feldspar clouding and brown-staining has produced essentially opaque alkali feldspars; otherwise they have undulatory extinction. Occasionally, alkali feldspar also occurs as xenoliths of partially resorbed crystal aggregates. No other xenolithic components were noted in the volcanics. A much smaller generation of albitized alkali feldspar with very diffuse grain boundaries is scattered throughout the matrix.

The next dominant type of phenocryst is green pleochroic aegerine or blue pleochroic riebeckite. Both varieties of ferromagnesian minerals occur in the rhyolites and trachytes, but not together in a single sample. Some rhyolitic samples contain virtually no pyroxene or amphibole. Both aegerine and riebeckite tend to occur as small tabs with occasional larger fragmented grains which appear to have given rise to the smaller tabs before crystallization of the matrix. These ferromagnesian minerals commonly have reacted to form anhedral opaques and red-orange alteration products. Small amounts of the ubiquitous red Fe-rich biotite is commonly associated with this alteration.

There are small grains of anhedral quartz in the rhyolitic samples. Extinction is sharp in these quartz "blebs", but they often contain a large amount of fine, unidentified inclusions.

The matrix is usually cloudy, with indistinguishable mineralogies. Typically the matrix displays undulatory extinction, with a fused and altered appearance. Some samples show a glassy matrix with fine alkali feldspar, undulatory quartz, and a large number of zeolites.

It is difficult to make consistent distinctions between the trachytes and rhyolites. Often there appears to be examples of each type from apparently single flows. The trachytes frequently display well-developed flow textures; the rhyolites tend to be more altered and often contain brecciated textures.

There are five main areas in the complex which are capped by large amounts of effusive rocks. The peak area of Abu Khruq contains a larger proportion of rhyolites with vent, agglomerate and breccia facies. The large volcanic cap on the north side of the inner ring contains a larger proportion of trachytes. All contacts between the volcanics and syenites demonstrate an older age for the volcanics. These contacts form as altered, pegmatoidal boundaries for the syenites, and leached silicified boundaries for the volcanics.

Syenites

A strict petrographic division of the syenites is difficult because clear sharp contacts are uncommon and many syenite samples display a gradation of the characteristics noted for obviously distinct varieties. The extent

of this problem was not completely recognized in the field, and thus, caused a less effective sampling method. Future sampling should include systematic traverses across the syenitic bodies--especially in a radial fashion with respect to the ring structure. The majority of the contacts at Abu Khruq (as well as the other complexes visited) are obscured by the conical faults, the intervening wadis, colluvium, and modification of the rock due to sub-solidus and/or secondary affects. The presence of secondary quartz is a major problem in this respect. Though more study is needed, it is likely that many of the contacts between the different varieties of syenite are gradational. This is obvious for the boundaries between the different varieties of silica-oversaturated syenites, and undersaturated syenites respectively, but this gradational relationship also appears to exist between some of these undersaturated and oversaturated types. Such a physical relationship is fundamental to a petrogenetic interpretation.

This study uses the term alkaline syenite for the silica-oversaturated syenites with little or no modal quartz. The term quartz syenite is used for syenite with greater than about 5% modal quartz. Silica-undersaturated syenites without large amounts of nepheline (less than about 10%) and without sodalite or possible primary analcime are called pulaskites. The nepheline and sodalite-rich syenites are classified as foyaites.

Alkaline Syenites

The alkaline syenites are characterized by miarolitic, coarse-grained textures with a predominance of perthitic alkali feldspar laths (65-80%) and a variety of interstitial pyroxene (10-15%). The alkali feldspar is usually albitized with perthitic textures, more widespread replacement, and small amounts of adcumulus growth. Feldspar clouding and brown-staining, often in grid or linear patterns, are superimposed on the perthetic textures. Sericitization is common throughout the alkali feldspars. The pyroxene is dominantly interstitial aegerine-augite with the typical pleochroic green character (10-15%). It is typically miarolitic, with poikilitic or closely associated euhedral apatite (3-5%), euhedral opaques (Fe-Ti oxides), and Fe-rich biotite. The interior of some of the aegerine-augite grains zone toward progressively paler green, non-pleochroic cores which probably represent more augite-diopside compositions. There are small amounts of pleochroic brown-blue amphibole and needle-shaped, secondary uralite (less than 5%). The uralite forms with the ubiquitous amphibole-opaque-Fe-rich biotite alteration assemblage which is common through all the major rock-types at Abu Khruq. One feature which aides in distinguishing the more mafic alkaline syenites from the other silica-oversaturated syenites is the presence of small amounts of oligoclase (5-10%), in the form of small euhedral tabs commonly enclosed by the alkali feldspars. These frequently have been replaced by cancrinite.

The exact location and extent of the alkaline syenites is uncertain because of the gradational relationships with the surrounding syenites. In a general way, the alkaline syenites appear to predominate around the outer portions of the inner ring (surrounding the foyaïtes), and the inner portions of the outer ring (rimming the inside of the quartz syenites). However, the possibility of secondary quartz in originally quartz-absent alkaline syenites must be considered. More importantly, a feature which was ignored by El Ramly et al. (1969a, 1969b, 1971) is the close association of the silica-oversaturated syenites with the silica-undersaturated syenites. Both levels of silica-saturation were collected from apparently continuous rock bodies. Though this relationship needs to be studied further in the field, the gradational boundaries between silica-oversaturated syenites (alkaline syenites) and silica-undersaturated syenites (pulsaskites) has serious implications on a petrogenetic model, and needs to be closely documented. Such a gradational relationship between similar rock-types is clearly displayed in the Kangerdlugssuaq alkaline complex in East Greenland (Pankhurst et al. 1976).

Quartz Syenites

The mineralogy and alteration of the quartz syenites are very similar to the alkaline syenites other than the presence of large amounts of quartz (5-25%). The quartz occurs in two dominant habits; one is a polycrystalline,

interstitial type which does not appear to replace other mineral phases; the second is a graphic, myrmekitic type which forms huge optically continuous grains and can be observed to replace alkali feldspar laths and pyroxene (a minor variety of quartz can be found in the miarolitic cavities of the pyroxene subhedral crystals). The proportion of quartz in some specimens approaches 25%, leaving a "skeleton" of fused alkali feldspar laths in which the grain boundaries as well as the Carlsbad twinning become vermicular or wavy in outline. The blastic quartz-rich textures were verified to contain large amounts of secondary quartz in some samples using cathodeluminescence. The proportion of aegerine-augite and apatite is less in the quartz syenites compared to the alkaline syenites. However, the same type of ferromagnesian alteration products are present. Large zones of ferruginated quartz syenites are visible at Abu Khruq--especially in the outer ring. Some samples from these zones have been converted to an opaque, Fe-rich material, surrounded only by unaltered quartz.

Pulaskites

The pulaskites are silica-undersaturated syenites which contain about 70-80% perthitic alkali feldspar, around 15-20% green pleochroic aegerine-augite, about 10% or less feldspathoids (dominantly nepheline, with rare sodalite and some secondary analcime), and the remainder of the rock consisting of euhedral Fe-Ti oxides, red

Fe-rich biotite, and accessory zircon and apatite crystals. There is common sericitization and cancrinitization in the feldspars; nepheline is much less affected by alteration, but occasionally is converted to liebernite (pink coloration) and cancrinite. The texture of the pulaskites is similar to that of the alkaline syenites; the ferromagnesian alterations are also similar. Nepheline occurs as small poikilitic euhedral crystals and anhedral "blebs" within the alkali feldspar, as well as the euhedral crystals in the interstices. There is a small amount of oligoclase (5-10%) in the pulaskites. The order of crystallization appears to be oligoclase-alkali feldspar-nepheline-pyroxene, with the crystalline of nepheline overlapping with that of the alkali feldspar laths. The pyroxene, as is typical of the entire alkaline rock series, is strictly interstitial with respect to the silic mineral phases.

Foyaite

These highly silica-undersaturated, coarse-grained syenites found in the center of the ring structure, were the last major phase of crystallization at Abu Khruq. The dominant mineral is highly albitized, miarolitic alkali feldspar (50-70%) which is usually altered in the same fashion as the alkali feldspar in the other varieties of syenite. Nepheline accounts for about 20-30% of the rock in the form of hexagonal or cubic crystal habits. The majority of the nepheline is unaltered, unlike the alkali feldspar. Some nepheline, however, has been converted to

cancrinite, and more rarely, to analcime. The majority of the euhedral isotropic mineral in the foyaïtes is sodalite (based on preliminary microprobe data), and not analcime as indicated by El Ramly et al. (1969a). Analcime is observed to be a replacement mineral of sodalite, and rarely, nepheline. It also fills in fractures and cavities in the foyaïtes. Sodalite and analcime occupy about 5-15% of the rock. Pyroxene is again the typical green pleochroic aegerine-augite variety (5-10%), often slightly zoned, with occasional poikilitic inclusions of apatite and nepheline; it is again associated with the previously described ferromagnesian alterations. There is much less apatite and Fe-Ti euhedral oxides in the foyaïtes than in the pulaskites; apatite tends to occur as second generation (?) slender needles visible only at high magnification. No plagioclase was identified in these syenites. Accessory zircon (less than 1%) occurs as small scattered crystals.

El Ramly et al. (1969a) make a textural distinction between foyaïtes and ditroïtes in these sodalite-nepheline syenites. Though there is no significant chemical difference, the foyaïtes display a coarse trachytic texture, and the ditroïtes show a finer granular texture. The ditroïtes tend to be found structurally higher than the foyaïtes, and they contain a much higher proportion of xenoliths. The sodalite-nepheline syenites are the only rocks to contain significant amounts of xenoliths. These cognate xenoliths appear to be very similar petrographically and chemically

to some of the syenitic dikes. They can be generally described as fine-grained nepheline-alkali feldspar micro-syenites with higher proportions of aegerine-augite (needle crystal habit), and red Fe-rich biotite. The xenoliths which were analyzed from the ditroites were more mafic than the host rock; those from the foyaïtes were very similar to the host rock. Thus, the foyaïtes are considered to represent deeper portions of a magma which allowed slower cooling, more chemical equilibration with the cognate xenoliths, and the development of coarser, flow textures. The ditroites, on the other hand, probably represent the upper pulses of the same magma which accumulated the majority of the xenoliths from the wall-rock, ascended, and cooled more rapidly than the lower foyaïtic portions. Thus, less chemical equilibration was achieved between the magma and xenoliths, and fine granular textures resulted.

In all textures, the sodalite-nepheline syenites follow the same order of crystallization: alkali feldspar overlapping with nepheline, sodalite, and pyroxene.

The foyaïtes can be found only in the central, stock-like portion of the inner ring. In the outer portions of these syenite bodies, there are large zones of ferrugination and shearing. Two mine shafts tunneled well into the northeastern portion of the foyaïtes demonstrate how extensive this alteration is. Fe-oxidation, calcification, albitization, liebernitization of the nepheline, analcime replacement of nepheline, and corrosion of the pyroxenes

are all widespread. Fluorite and other secondary minerals are especially common throughout these altered syenites.

The structural relationships between the foyaïtes and other types of syenite are again obscure. This is commonly due to the ferruginated zones and intervening wadis. It is the observation of this study, however, that they display a gradational boundary with the pulaskites, and thus, indirectly grade into the silica-oversaturated alkaline syenites. Once again, such a relationship between the silica-undersaturated and oversaturated syenites has important implications on a petrogenetic interpretation.

Dikes

There are a wide variety of dikes at Abu Khruq which were sampled and studied in case they represented an important igneous phase of activity (e.g. early mafic phases), or had caused a strong affect on their host rocks (e.g. metasomatic). There is no evidence for them representing an extremely large volume of the magma, nor that they had a large affect on the host rock-types. Refer to El Ramly et al. (1969a) for a description of the different dike rocks.

B. Summary of the Mineralogy of Abu Khruq

Similar to the other Egyptian alkaline complexes, and many alkaline suites in general, the order of crystallization in the rocks at Abu Khruq begin with the crystallization of the salic minerals, and are followed by the

crystallization of the ferromagnesian minerals (agpaitic order of crystallization, e.g. Kogarko et al., 1978).

Only in the wehrlite (pyroxene-olivine cumulate) is there an indication of the ferromagnesian minerals crystallizing before the salic minerals. Even in this rock-type, however, nepheline forms in euhedral crystal habit. In the gabbros labradorite is the primary cumulus phase; in all the syenites alkali feldspar is the dominant primary cumulus phase; in the volcanics alkali feldspar is the major phenocryst.

Overall, the crystallization of the major minerals at Abu Khruq (as well as the other complexes) appears to have occurred in very discrete intervals of composition. Evidence of an ultramafic substratum similar to the wehrlite collected suggests that olivine and calcic-pyroxene precipitated into dense layers deep within the complex; the gabbros represent very plagioclase-rich accumulations in the lower part of the exposed complex. Because no rock-type was observed to have sub-equal proportions of plagioclase and alkali feldspar, it appears that there was a rapid transition to the crystallization of alkali feldspar which dominates throughout the syenites.

Olivine

Olivine is found in large amounts in only the pyroxene-olivine cumulate (wehrlite). In the only sample collected, it appears to be forsterite-rich. The only other

occurrences of olivine at Abu Khruq are small reacted grains in some of the gabbros. It appears that by the time most of the visible rock-types were crystallizing, olivine had moved off the liquidus.

Pyroxene

The pyroxene in the wehrlite sample appears to be diopside. The gabbros contain a flesh-colored, non-pleochroic diopsidic-augite (see El Ramly et al., 1969a). The most typical variety of pyroxene for the majority of the exposed rocks at Abu Khruq is aegerine-augite. This is a green pleochroic, often slightly zoned, miarolitic, subhedral mineral phase. It occurs interstitially throughout all the syenites. In much smaller proportions, there are also other varieties of pyroxene, such as a pleochroic blue-green type which is easily confused with the minor amounts of amphibole in the syenites.

Amphibole

Amphibole rarely occurs over 2-3% of any of the rocks at Abu Khruq. The most common form is uralite: a green pleochroic variety which forms as an alteration product from pyroxene as clusters of fine needle-shaped crystals. It is usually associated with the opaque Fe-oxides and red Fe-rich biotite which also commonly surround reacted pyroxene. Other minor varieties of amphibole include a brown, anhedral alteration product of pyroxene; there is also some kaersutite (blue-green variety) and some minor arfvedsonite.

Plagioclase

Large proportions of labradorite (An_{66-52}) occur in the gabbros as coarse-grained laths with well-developed albite twinning. Other than this, only very small amounts of oligoclase were identified in the other rock-types at Abu Khruq.

Alkali Feldspar

The initial composition of the alkali feldspars in the various rock-types at Abu Khruq is difficult to determine due to the widespread albitization and other sub-solidus or secondary reactions (sericitization, cancrinitization, feldspar clouding, dissolution). In many cases, it appears to be orthoclase laths in the syenites, with evident Carlsbad twinning, and albite exsolution textures. Some of the rhyolites contain sanidine phenocrysts, and the gabbros have very small amounts of orthoclase.

Feldspathoids

The dominant feldspathoid is nepheline. It appears in the ultramafic wehrlite sample, as well as in the gabbros, as small euhedral crystals. It is absent from all the volcanic samples (nor is there leucite in the volcanic rocks). It occurs as a cumulus phase in the silica-under-saturated syenites (pulaskites and foyaïtes), following only alkali feldspar in crystallization order.

The other common feldspathoid at Abu Khruq is sodalite. Based on preliminary microprobe data, the large crystals

of isotropic material in the foyaïtes are not alacime (El Ramly et al., 1969a), but sodalite. Analcime, however, was observed to replace some of the sodalite, more rarely, the nepheline, and also fill in fractures and cavities in the foyaïtes and pulaskites.

Quartz

Quartz occurs in two dominant habits: one is a polycrystalline, interstitial type which does not appear to replace other mineral phases; the second type is a graphic, myrmekitic type which forms very large, optically continuous grains, and can be observed to replace alkali feldspar laths and pyroxene crystals. This last type of quartz involves a secondary origin as verified by cathode-luminescence. Due to the presence of secondary quartz, it is difficult to determine if certain samples may not have originally contained quartz. In spite of this possibility, all intrusive samples with quartz are referred to as "quartz syenites". Although this classification may include some alkaline syenites and some slightly silica-undersaturated syenites, it is important to note that no quartz was observed in samples of gabbro or samples with large amounts of nepheline. Thus, either the second generation of quartz was deposited under near-equilibrium conditions in rocks of compatible mineralogies, or the original character of the rocks were radically altered along with the deposition of quartz. This second possibility is not considered likely.

The proportion of quartz in some quartz syenites approaches 25%. Such quartz-rich rocks are essentially "skeletons" of fused alkali feldspar laths and a few green pyroxene grains surrounded by blastic quartz textures. Small inclusions of silica are common in these syenites, as well as in some of the volcanic samples.

Fe-rich Biotite (Stilpnomelane)

The most ubiquitous minor mineral in all the rocks at Abu Khruq is a red, occasionally brown biotite variety which may be stilpnomelane. Based on the varying degree of coloration, it contains large differences in Fe content. In many cases, it becomes very dark red to opaque, and therefore, is difficult to separate from the commonly associated Fe-oxides. It most often occurs in typical pyroxene-Fe-oxide~~amphibole~~ alteration assemblages in all the intrusive rock-types. More rarely, it occurs interstitially with alkali feldspar as large subhedral flakes. It was observed in essentially all dike rocks, in the gabbros, and all syenites, but does not occur in the volcanics. The reason for this consistent appearance is uncertain, but it would appear to indicate a general characteristic for the entire magmatic series, one involving a hydrated, Fe-excess reaction.

Fe-Oxides

Fe-oxides occur in two textural varieties: one as euhedral cumulate varieties, the other as anhedral

alteration products from the corrosion of the pyroxenes and cumulate Fe-oxides. The euhedral crystals are Fe-Ti varieties (microprobe data and see El Ramly et al. (1969a), and are most abundant in the gabbros and more mafic syenites--usually in association with pyroxene (often poikilitically). The anhedral opaques are most often found in the pyroxene-Fe-Ti-oxide-stilpnomelane[†]amphibole reaction associations. No dense bands of cumulate Fe-oxides were observed at Abu Khruq, as sometimes occur in other alkaline complexes.

Apatite

Apatite is an important accessory mineral in alkaline suites--especially in studies using REE since it represents one of the most common REE-enriched minerals. Apatite, similar to the Fe-oxides, is sometimes found in dense cumulate bands in alkaline complexes. This was not observed at Abu Khruq. It occurs in two dominant forms: one is a large euhedral variety found within and beside the pyroxenes of especially the gabbros and more mafic syenites; the other type of apatite occurs as a second generation of fine needles seen only at high magnification throughout the majority of the rocks at Abu Khruq.

Other Minerals

Calcite, fluorite, and perhaps other low-temperature minerals occur in all the syenites--especially miarolitic cavities of the silica-oversaturated syenites. These types

of minerals were not identified in the gabbros. Zeolites, cancrinite, sericite, liebernite, chlorite, and epidote were observed in small amounts as alteration minerals of all the major rock-types.

Minor minerals originally crystallized in the major rock-types include sphene, spinel, and zircon.

APPENDIX II

APPENDIX II

A CRYSTAL FRACTIONATION MODEL FOR GABAL ABU KHURUQ BASED ON THE TRACE ELEMENTS

The objective of this crystal fractionation model is to present several stages of crystal fractionation in a manner which is reasonable for the observed rock-types at Abu Khuruq, and also taking into account the nature of similar alkaline complexes in other areas. Due to the lack of well-exposed mafic rock-types at Abu Khuruq, and the lack of a representative of a likely parent melt, this model is not well constrained. Review of the crystal fractionation models in the literature on similar alkaline suites aided in making certain decisions on the format of the model, particularly in the early mafic stages. The following fractionation scheme represents one possible sequence of crystal fractionations which is consistent with the observed trace element distributions of the major rock-types found at Abu Khuruq. An attempt was made in the model to use partition coefficients which 1) are reasonable for an alkaline suite, and 2) yield a more conservative fractionation model (i.e. smaller amounts of crystal fractionation). Although the detailed results of this scheme may be altered with the insertion of different partition coefficients, and

Table A II 1. A More Detailed Fractional Crystallization Model for Gabal Abu Khruq.

The compatible trace elements are used as indicators of crystal fractionation in the early fractionation history of the magma; the compatible trace elements are used for the latter stages of fractionation.

Compatible Trace Elements

First Fractionation Stage:

Alkaline basaltic parent magma

Ultramafic olivine-rich cumulate (Wehrlite, A-94)
+ Residual liquid (not represented in samples).

Results:

90% melt (C_1) + 10% olivine-rich cumulate (C_r), similar to sample A-94.

Second Fractionation Stage:

Melt from previous stage

Pyroxene-rich cumulates + Residual liquid (not represented in samples).

Results:

54% melt (C_1) + 36% pyroxene-rich cumulates (C_r) + olivine rich cumulate of previous stage.

Third Fractionation Stage:

Melt from previous stage

Plagioclase-rich cumulates (C_r) (gabbro, A-83a and A-91a) + Residual liquid (intermediate, mesocratic syenites, A-60a and Eg-32a).

Results:

32% melt (C_1) + 22% plagioclase-rich cumulates (C_r), similar to gabbros A-83a and A-91a + the previous pyroxene-rich cumulates + the previous olivine-rich cumulates.

Fourth Fractionation Stage:

Melt from previous stage (syenitic liquid)

Alkali feldspar-rich, feldspathoidal-rich residual liquid (C_1) in which there is on-going crystallization in close association with the in-situ cumulates of this stage. This is a crystal mush which is represented by a bulk average of the quartz syenites, the volcanics and the foyaite.

*Note that Ni, Cr, and Co are not treated further in this simplified model because they are in such low concentrations and are probably behaving incompatibly with respect to the dominant crystallizing phases.

Results:

7% residual liquid + 25% cumulate components consisting predominantly of alkali feldspar + feldspathoids, similar to the bulk composition of the quartz syenites, volcanics and foyaite, + 22% gabbros + 36% pyroxene-rich cumulates + 10% olivine-rich cumulates.

Table A II 1 (cont'd).

Incompatible Trace Elements

The fractionation scheme involving the incompatible trace elements begins with an alkaline basaltic parent magma (same as in the compatible trace element scheme), but the initial result begins with the production of the gabbroic cumulates which correspond to the Third Fractionation Stage of the compatible trace element scheme.

First to Third Fractionation Stage:

Alkaline basaltic parent magma



Plagioclase-rich cumulates (C_1), (gabbro, A-83a and A-91a) + Olivine and pyroxene-rich cumulates (unaccounted for due to serpentinized, REE-mobilized(?) nature of the sample) + Syenitic Residual liquid (intermediate, mesocratic syenites).

Results:

32% intermediate syenitic melt (C_1) (represented by the intermediate, mesocratic syenites, A-16, A-60a, Eg-32a, A-92a, A-79), + 64% gabbroic cumulates.

Fourth Fractionation Stage:

Melt from previous stage



Alkali feldspar-rich, feldspathoidal-rich residual liquid (C_1) in which there is on-going crystallization in close association with the in-situ cumulates of this stage. This is the same crystal mush stage as in the compatible trace element scheme, and is represented by a bulk average of the quartz syenites, volcanics,

and foyaïtes.

Results:

7% residual liquid + 25% cumulate components consisting of predominantly alkali feldspar + feldspathoids, similar to the bulk composition of the quartz syenites, volcanics and foyaïtes (+ the previous 22% gabbroic, 36% pyroxene-rich, and 10% olivine-rich cumulates).

*Note the consistency between the final fractionation stages the REE, Th, Sr and Ba.

Table A II 1 (cont'd).

Results and Parameters of the Crystal Fractionation Model

F	C _O	C _R	O _R	C _I	O _I	\bar{D}
<u>First Fractionation Stage:</u>						
Ni	.90	350	1254	1040	91	--
Cr	.90	350	932	989	271	--
Co	.90	60	165	120	46	--
Sr	.90	500	109	97	544	--
Ba	.90	400	44	1	440	--
<u>Second Fractionation Stage: (.54 of original parent melt)</u>						
Ni	.60	91	89	--	25	--
Cr	.60	271	268	--	78	--
Co	.60	46	51	--	39	--
Sr	.60	544	164	--	819	--
Ba	.60	440	63	--	697	--
<u>Third Fractionation Stage: (.32 of original parent melt)</u>						
Ni	.60	25	24	31	7	10
Cr	.60	78	77	57	22	10
Co	.60	39	43	31	33	14
Sr	.60	819	968	899*	594	529
Ba	.60	697	264	170	1017	1140
*This Sr average value includes A-83a, A-91a plus two additional sample values from data collected by Lutz (1979).						
<u>Fourth Fractionation Stage: (.07 of original parent melt)</u>						
Sr	.23	594	96	--	83	32 ⁺
Ba	.23	1017	23	--	5	1 ⁺⁺
+Average of 36 analyses from this study and Lutz (1979).						
++Average of 4 Ba analyses.						

F	C _O	C _R	O _R	C _I	O _I	\bar{D}
<u>First to Third Fractionation Stage: (Incompatible Trace Elements)</u>						
Ce	.32	44.00	37.44	38.33 [*]	81.40	82.60 ⁺
Sm	.32	5.53	5.15	4.03 [*]	9.03	8.82 ⁺
Lu	.32	0.23	0.21	0.20 [*]	0.40	0.55 ⁺
Th	.32	2.90	0.81	1.16 [*]	8.09	7.48 ⁺
* Average of gabbros, A-83a and A-91a.						
+ Average of intermediate syenites, A-16, 60a, 92a, 79a, Eg-32a.						
<u>Fourth Fractionation Stage: (.07 of original parent melt)</u>						
Ce	.23	81.40	82.80	--	180.00	188.27 [*]
Sm	.23	9.03	19.68	--	16.99	16.50 [*]
Lu	.23	0.40	0.42	--	0.82	0.94 [*]
Th	.23	8.09	7.36	--	21.03	18.04 [*]
* Average of 40 analyses of quartz syenites, volcanics and foyaites.						

C_O = concentration of element in the parent liquid.C_R = concentration of element in the calculated derived residual or cumulate after crystal fractionation.O_R = concentration of element in the observed rocks which represent the residual or cumulate phases.C_I = concentration of element in the calculated derived liquid after crystal fractionation.O_I = concentration of element in the observed rocks which represent the remaining liquid after crystal fractionation. \bar{D} = bulk distribution coefficient.

F = fraction of original magma remaining after crystal fractionation.

to a limited extent, with the use of another reasonable parent melt, the general conclusions are not greatly changed. Refer to Arth and Hanson (1976), Zielinski (1975), Sun and Hanson (1976), Allegré et al. (1977), Frey et al. (1978), and Baker et al. (1977) for the range of published partition coefficients used in this crystal fractionation model.

The major conclusions of the crystal fractionation calculations are:

1. Crystal fractionation was an important process of differentiation in the rocks at G. Abu Khruq.
2. Large amounts of efficient crystal fractionation occurred (about 2/3s removal of the parent melt for the stage after the separation of the gabbroic phases, and about 90% removal of the parent melt in order to achieve the bulk compositions of the most evolved rock-types).
3. It is likely that the parent melt of G. Abu Khruq was a moderately alkaline basaltic magma.
4. The distributions of the compatible and incompatible trace elements agree reasonably well so that a fairly unique solution for crystal fractionation is possible.

1) Choice of Parent Melt

The choice of a primitive parent melt for the alkaline rock suite at G. Abu Khruq was based on:

1. The relatively low concentrations of incompatible trace elements at Abu Khruq and the other Egyptian alkaline

complexes visited eliminated the possibility of a nephelinitic, basanitic, or other extremely enriched alkaline basaltic parent. The most likely parent is a moderately-enriched alkaline basalt.

2. The most mafic rock-types at Abu Khruq are silica-undersaturated, and therefore, the parent was probably silica-undersaturated.

3. The mafic rock-types of El Gezira, Egypt (El Ramly et al., 1971), Jebel Al Abyad, Saudi Arabia (Baker et al., 1973), Canary Islands (Munoz, 1969), Reunion Island (Upton and Wadsworth, 1972; Zielinski, 1973), eastern Australia (Frey et al., 1978), and Hawaii (Schilling and Winchester, 1969; Kay and Gast, 1973; Frey et al., 1978) all aided in selecting trace element compositions which are remarkably similar throughout the world for low-REE, fairly undifferentiated alkaline basalts.

It should be noted that the REE distributions for the theoretical parent of Abu Khruq is essentially that of an average Hawaiian alkali olivine basalt (Schilling and Winchester, 1969; Kay and Gast, 1973; Frey et al., 1978).

2) First Fractionation Stage

The first fractionation stage approximates the olivine-rich separation stage observed in sample A-94, a olivine (90%) clinopyroxene (10%) ultramafic cumulate (wehrlite).

$$\bar{D}_{Ni} = 15(.90) + 3(.10) = 13.8$$

$$\bar{D}_{Cr} = 2.15(.90) + 15(.10) = 3.44$$

$$\bar{D}_{Co} = 3.88(.90) + 1.21(.10) = 3.61$$

$\bar{D}_{Sr} = 0.2$... Sr is an incompatible trace element in this stage, and this bulk partition coefficient is reasonable compared to values used in the literature for similar alkaline mafic or ultramafic separations (e.g. Allégre et al., 1977).

$\bar{D}_{Ba} = 0.1$... Ba is an incompatible trace element in this stage, and this bulk partition coefficient is reasonable compared to values used in the literature for similar alkaline mafic or ultramafic separations. Compared to that used by Frey et al. (1978) it is larger and thus will result in a more conservative model.

The F value of .10 for this first fractionation stage was chosen based on the observed trace element distribution of sample A-94. About 10% olivine separation is not unreasonable for the earliest fractionations of an alkaline basaltic melt. About 15-25% total olivine fractionation is likely for alkaline basaltic suites which achieve trachytic compositions (e.g. Zielinski, 1975; Sun and Hanson, 1976; Baker et al., 1977). The remaining olivine separation will be accommodated in the following fractionation stage.

3) Second Fractionation Stage

The second fractionation stage approximates pyroxene-rich separations which were not observed at Abu Khruq. Although such cumulate rock-types were not observed, it is a geologically reasonable step between the preceding olivine-rich separations and next plagioclase-rich cumulate stage (gabbros). This stage does not seriously effect the general magnitudes of the results in this crystal

fractionation model. Pyroxenites are commonly observed in similar alkaline complexes with deeper exposure (e.g. see Gastesi, 1969; Muñoz, 1969). The presence of such rock-types may be indicated by the cognate xenoliths which were found in the diatreme-like features observed at G. El Kahfa (see Appendix I). In view of the fact that this study is the first to report the presence of the olivine-rich cumulate, it is not unreasonable to expect that further investigation of these alkaline complexes will reveal more ultramafic rock-types in their underlying structures.

$\bar{D}_{Ni} = 3.5$... relative bulk partition coefficient; this value is reasonable for the mineralogy of the predicted pyroxenites; it is similar to the relative bulk partition coefficient used by Allegre et al. (1977) for the mafic fractionations of an alkaline suite.

$\bar{D}_{Cr} = 3.44$... unchanged from the first fractionation stage; the relative bulk partition coefficient was uncertain due to the scatter in the Cr distributions. This scatter may be largely due to the unique amounts of Fe-Ti oxides in individual samples. Due to the extremely strong incorporation of Cr into the Fe-Ti oxides, and the pyroxene-rich nature of this fractionation stage, this rather high bulk partition coefficient was retained for this second fractionation stage.

$\bar{D}_{Co} = 1.3$... relative bulk partition coefficient; comments on the Ni value of this stage also apply to Co.

$\bar{D}_{Sr} = 0.2$... unchanged from first fractionation stage due to similar conditions.

$\bar{D}_{Ba} = 0.1$... unchanged from first fractionation stage due to similar conditions.

The F value of .60 for this stage is somewhat arbitrary. It considers the amount of ferromagnesian separations necessary before the following more salic rock-types may be derived (see the following fractionations based on Sr and Ba), and it also takes into account the amount of ferromagnesian separations which are typically removed in other studies for alkaline suites (e.g. Baker et al., 1977; Zielinski, 1975).

4) Third Fractionation Stage

The third fractionation stage approximates the plagioclase and pyroxene-rich separations similar to the observed gabbros (60% plagioclase, and 40% clinopyroxene).

$$\bar{D}_{\text{Ni}} = 3.5 \quad \dots \text{unchanged}$$

$$\bar{D}_{\text{Cr}} = 3.44 \quad \dots \text{unchanged}$$

$$\bar{D}_{\text{Co}} = 1.3 \quad \dots \text{unchanged}$$

$$\bar{D}_{\text{Sr}} = 2.59(.60) + 0.20(.40) = 1.63 \quad \dots \text{average values from published partition coefficients.}$$

$$\bar{D}_{\text{Ba}} = 0.36(.60) + 0.1(.40) = 0.26 \quad \dots \text{average values from published partition coefficients}$$

The F value for this third fractionation stage is a result of approximating the observed concentrations in the gabbros and the intermediate syenites. At this stage about 68% of the parent melt has crystallized.

5) Fourth Fractionation Stage

The compatible trace elements with strong ferromagnesian affinities are considered to start behaving incompatibly, and their low concentrations are below any feasible resolution for this crystal fractionation model.

$$\bar{D}_{\text{Sr}} = 3.87(.75) + 0.15(.15) = 2.98 \dots \text{where the syenites and volcanics are approximated by 75\% alkali feldspar and feldspathoids, and 15\% clinopyroxene (a less mafic aegerine-augite variety).}$$

$$\bar{D}_{\text{Ba}} = 6.12(.75) + 0.1(.15) = 4.61 \dots \text{same comment as for Sr.}$$

These two trace elements are the most difficult to model because their bulk partition coefficients change from highly incompatible ones in the early ferromagnesian crystallization stages to highly compatible ones in the intermediate crystallization stages. Thus, these trace elements become very depleted in the final stages of crystallization.

The fractionation scheme for the REE and Th is much more straight-forward because their bulk partition coefficients can be approximated by nearly constant values throughout the evolution of the entire melt. These values are relative bulk partition coefficients for the REE (derived by the method outlined by Allegre et al., 1977). The bulk partition coefficient of Th is arbitrarily chosen as 0.10 for the gabbroic fractionation stage. This value is in complete agreement with similar values used in the literature (see for example, Frey et al., 1978). This extremely small value is likely to increase slightly in

the final salic stages of differentiation. Thus, an arbitrary value of 0.35 was chosen with reference to the experimental data reviewed by Baker et al. (1977).

In this very late stage of fractionation, it is likely that there was not a complete separation between the residual liquid and the crystallizing phases. Thus, these final alkali feldspar and feldspathoidal-rich cumulates were probably distributed throughout the final crystallizing rock-types.

It should be noted that the bulk compositions of the final salic rock-types were used (averages). As was discussed in the text, crystal fractionation cannot account for the relationship between the various varieties of salic rock-types. In spite of the few rigid constraints in this model, there is good agreement between these trace elements in terms of very similar F values. Numerous variations of crystal fractionation models were performed, and consistently this same general type of result was concluded: in order to achieve the bulk salic compositions of the syenites and volcanics, about 90% of an alkaline basaltic parent melt must be removed.

APPENDIX III

APPENDIX III

NET-VEIN COMPLEX AT NIGRUB EL FOGANI: REE PATTERNS

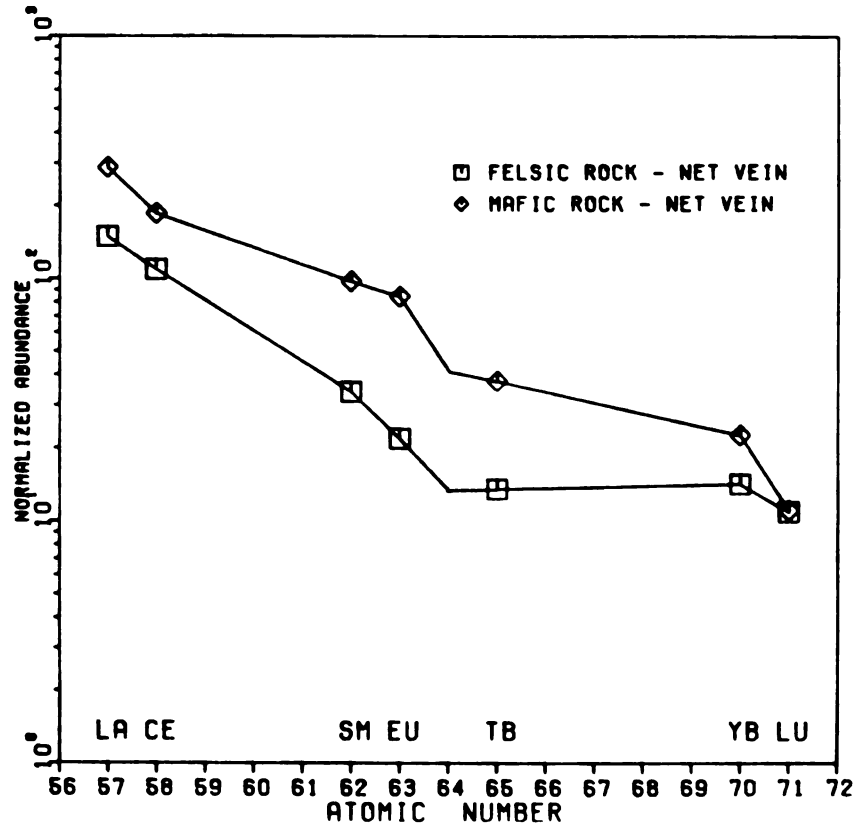


Figure A III 1. Normalized REE plot of the mafic and salic rock-types collected from a net-vein complex with apparent liquid/liquid contacts from G. Nigrub El Fogani. Similar to experimental data on mafic/salic liquid immiscible systems, these mafic and salic members show that the REE are richer in the mafic phase than in the salic phase. A crystal fractionation relationship would require the opposite. See Chapter 2, part B.

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