

PETROGENESIS AND TECTONIC SETTING OF LATE PRECAMBRIAN ENSIMATIC VOLCANIC ROCKS, CENTRAL EASTERN DESERT OF EGYPT

ROBERT JAMES STERN

Department of Terrestrial Magnetism, 5241 Broad Branch Road N.W., Washington, DC 20015 (U.S.A.)

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ABSTRACT

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Early stages in the geologic evolution of the central eastern desert of Egypt (CED) reflect an intense episode of ensimatic volcanic activity similar to modern magmatism of the ocean floors and island arcs. This paper reports results from studies of the petrology and petrogenesis, and interprets the significance of these Late Precambrian volcanic rocks.

A three-fold stratigraphy is preserved in the basement of the CED. A basal section of oceanic crust includes ultramafics, gabbros and pillowed basalts. These older metavolcanics (OMV) are conformably succeeded by dominantly volcanogenic metasediments, which are in turn succeeded by a dominantly andesitic, calc-alkaline sequence of younger metavolcanics (YMV). The OMV and YMV are largely restricted to the CED in Egypt, but analogous terranes are found in northern Arabia. (40–400 ppm) and Ni (30–260 ppm). They are poor in K_2O (0.05–0.92%), Rb (0.3–5.0 ppm) and Ba (11–89 ppm). On Ti–Zr–Cr–V–Ni–P discriminant diagrams, the OMV plot in the field of modern abyssal tholeiites. High K/Rb (450–1800) and light REE depletions support this inference, although K/Ba (25–45) is lower than modern mid-ocean ridge basalts (MORB). The sum of OMV geochemical characteristics requires that these magmas were derived by the fractional fusion of the mantle. It is suggested that the OMV were generated by 20–25% fractional melting of previously depleted mantle at depths of less than 60 km. Relatively little fractionation accompanied ascent to the surface, where the OMV were erupted in a primitive crustal environment, either a small oceanic rift or a back-arc basin.

Metamorphism of the YMV resulted in little elemental redistribution. These andesites have sub-alkaline clinopyroxenes and major-element geochemical characteristics indistinguishable from modern calc-alkaline andesites. YMV andesites in the central and western CED have K/Rb = 400–600, K/Ba = 20–40 and are light REE-enriched and heavy REE depleted. High concentrations of Cr (50–150 ppm) and Ni (20–100 ppm) and low initial $^{87}Sr/^{86}Sr$ ratios (0.7028–0.7030) indicate that these magmas were generated by melting in the mantle. Modelling studies and consideration of experimental data indicate that these andesites were formed by 2–10% fractional fusion of hydrous, undepleted, garnet lherzolite at depths of 65 km or more in the mantle.

The data show that an intense episode of instability, convection, and widespread melting occurred in the mantle beneath Afro-Arabia at the end of the Precambrian.

INTRODUCTION

A growing body of evidence indicates that an episode of ensimatic crustal generation and related sialic crustal disturbance took place in parts of North Africa and Saudi Arabia at the end of the Precambrian. Such activity was largely confined to the northern extensions of cratonic terranes rejuvenated during the Pan-African orogeny (Fig. 1). Especially intense activity was concentrated around the modern Red Sea, in Saudi Arabia, Sudan and eastern Egypt. Igneous and tectonic processes in this region resulted in the development of primitive ensimatic crustal terranes which formed, collapsed, and were engulfed by granites during the final 400 Ma of the Precambrian (Engel et al., 1980).

The purpose of this paper is further to elucidate the nature and signifi-

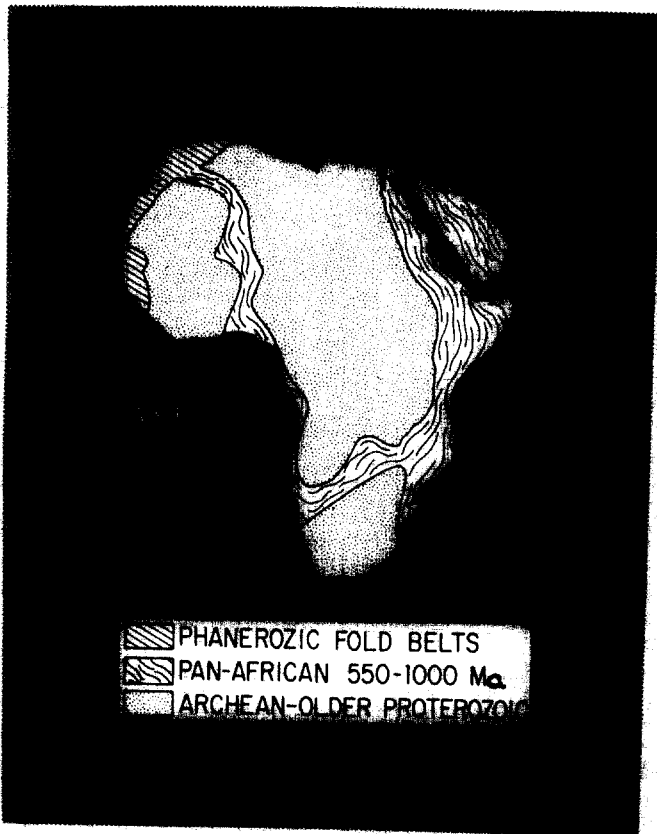


Fig. 1. Generalized tectonic map of Africa, showing the relationship of terranes affected or generated during the Pan-African event to older and younger crustal blocks. Pan-African crustal units studied in this paper are situated in NE Africa, west of the modern Red Sea.

cance of Late Precambrian igneous activity in North Africa. The volcanic rocks which constitute the earliest manifestation of such activity are considered to be especially important. Thus, this study is focused on the petrology and petrogenesis of the volcanic rocks of a portion of NE. Africa, the central eastern desert of Egypt. Such data are important because very little is known of Late Precambrian volcanism in NE Africa and Saudi Arabia. Similar sequences in Saudi Arabia have been investigated by Bakor et al. (1976); petrochemical studies of Egyptian volcanic sequences similar to those described here are limited to that of Awadallah and Shaalan (1979).

REGIONAL GEOLOGY AND STRATIGRAPHY

Most of the Precambrian outcrops in Egypt are restricted to the area between the Nile and the Red Sea. Within this region, the prebatholithic volcanic and sedimentary succession is concentrated in the central eastern desert (CED). This study concentrates on the volcanic succession of the CED, especially between $25^{\circ} 30'$ and $26^{\circ} 30'N$. The following areas were studied in detail: Wadi Ambagi, Wadi El Dabbah, Wadi Kareim, Wadi Um Seleimat and the region around Fawakhir; Wadi Arak, Wadi El Mahdaf, and the region just east and north of the junction of Wadis Arak and Zeidun (Arak/Zeidun). The location of these areas is shown on a geologic map (Fig. 2) of a portion of the CED. This map represents the integration of the author's geologic mapping in these areas with other regional investigations (Noweir, 1968; Akaad and Shazly, 1972; El Ghawaby, 1973).

Despite more than 150 years of study, the basement stratigraphy of this area is incompletely understood. The volcanic units discussed here would be included in the Shadli Middle Schists of Hume (1934). Akaad and El Ramly (1960) suggested that these metavolcanics and associated ultramafics and immature sediments reflected activity in a developing geosyncline. El Shazly (1964) divided this episode into 'early- and late-geosynclinal structural stages'. The latter stage was characterized by the intrusion of 'synorogenic plutonites' (tonalites and granodiorites) and the eruption of the Dokhan Volcanics. The volcanic units discussed herein would be included in the early geosynclinal structural stage. El Shazly further subdivided the volcanic rocks of this stage into: (1) a 'basic—ultrabasic association'; and (b) an 'intermediate association'. The first was known as the 'Baramia Association'; El Shazly noted that it "exhibits ophiolite characteristics". In contrast, the 'intermediate association' was dominated by andesites. The author's field-work substantiates the most important aspects of El Shazly's (1964) stratigraphy.

The results of the present stratigraphic investigation are schematically summarized in Fig. 3, condensed from the more extended discussion of Stern (1979a). The volcanics shown in Fig. 3 are informally subdivided on the basis of relative stratigraphic position into 'older metavolcanics' (OMV) and 'younger metavolcanics' (YMV).

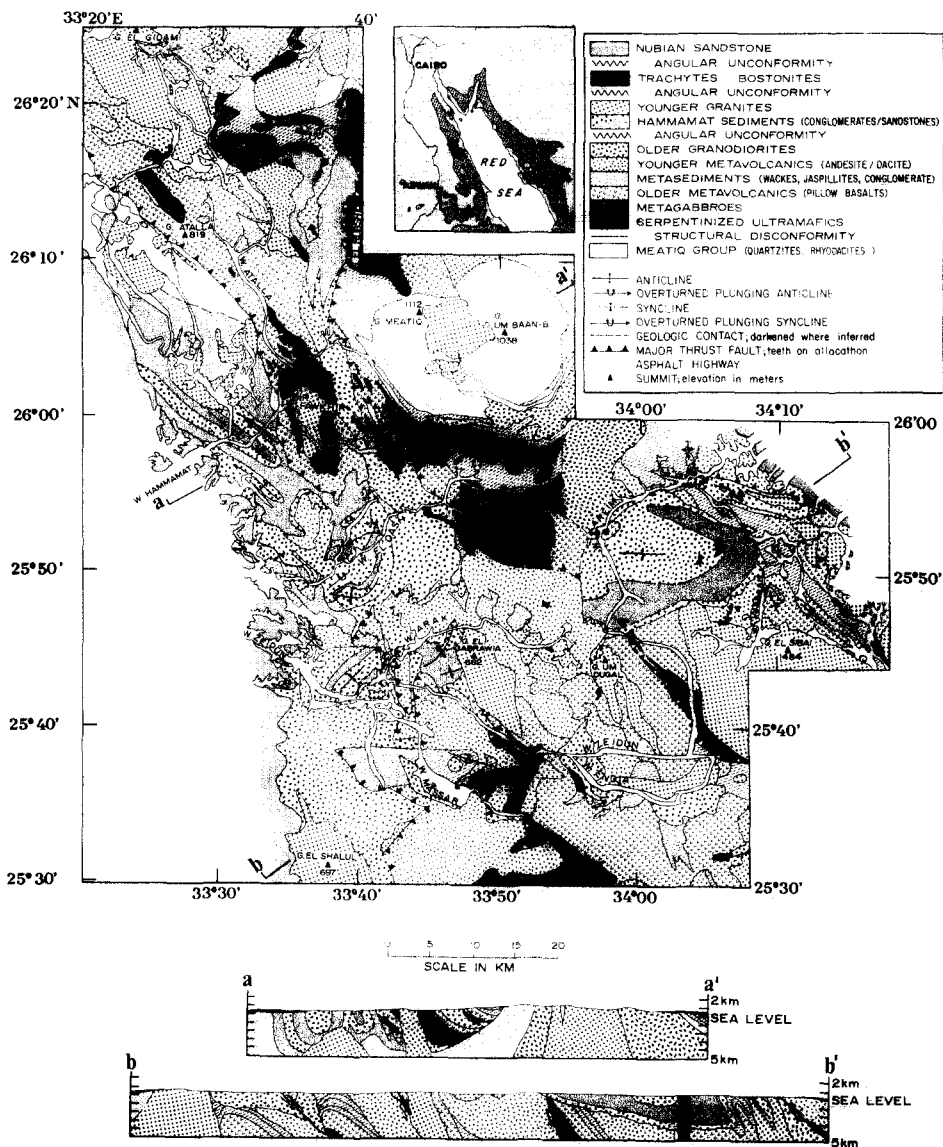


Fig. 2. Geologic map of a portion of the central eastern desert (CED) of Egypt.

Older metavolcanics

The OMV are characterized by a thick, monotonous succession of aphyric pillowed metabasalts. Sedimentary interbeds are rare. At Arak/Zeidun, the OMV are associated with 10–30 m thick metagabbroic sills. These metagabbros appear to be comagmatic, and have been included with the volcanic rocks for the purposes of petrologic study. As a result of the homogeneous nature of the OMV, it is difficult to ascertain the importance of tectonic

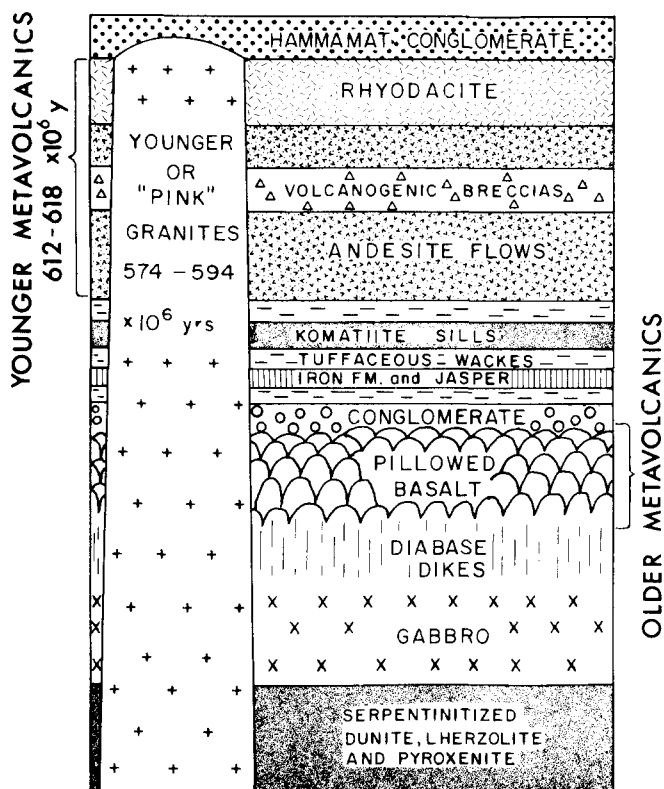


Fig. 3. Generalized stratigraphy for Late Precambrian basement units of the central eastern desert (CED), Egypt. Range of ages reported for the younger metavolcanics (YMV) are from the work of Stern (1979a); ages for the younger of 'pink' granites are from the work of Fullagar and Greenberg (1978) recalculated for $\lambda = 1.42 \times 10^{-11} \text{y}^{-1}$.

repetition or deletion within the succession; hence, thickness determinations are equivocal. At Arak/Zeidun, the OMV is folded, but appears to be ca. 1 km thick. East of Fawkhir, the OMV appears to extend ca. 1 km beneath the metasediments before it passes downward into diabase dikes (?) and coarse-grained gabbro.

Although widespread, the succession of pillowed basalt—gabbro—ultramafics is rarely preserved in its original sequence. Fawkhir-Um Seleimat is the only area where gabbros and ultramafics have been seen (by the author) to unequivocally underlie pillowed basalt. The top of this sequence consists of pillowed basalts with the local development of sedimentary pillow breccia. Down-section, the pillows are succeeded by aphyric, more massive metabasalt which may include diabase dikes. These, in turn, are underlain by metamorphosed anorthositic gabbros around Bir El Sid. Along the Qena—Quesir road, the gabbroic succession is ca. 500 m thick. This grades down-section into 500 m of pyroxenite. The pyroxenites overlie more altered

serpentinites, derived from dunites and harzburgites. The serpentinites form the base of an allochthon thrust to the west over the Hammamat Formation.

The Fawakhir-Um Seleimat succession shows characteristics similar to ophiolites. Ophiolites have been interpreted as fragments of oceanic (Moore and Vine, 1971; Church, 1972), marginal basin (Pearce, 1975; Bakor et al., 1976; Hawkins, 1977), or island-arc crust (Ewart and Bryan, 1972; Miyashiro, 1973). The volcanologic and stratigraphic characteristics of the OMV and its gabbroic—ultramafic substrate are consistent with such paleo-environmental reconstructions.

The OMV or similar ophiolitic units are of regional extent. Similar successions are widespread in northern Saudi Arabia (Bakor et al., 1976; Delfour, 1977; Shanti and Roobol, 1979).

The OMV are conformably succeeded by a series of dominantly volcanogenic sediments, especially tuffaceous sandstone and breccias. These immature metasediments include distinctive beds of hematite-rich Fe-formations, jaspers, rare marls, and distinctive, poorly-sorted conglomerates (Stern, 1979a). These conglomerates contain cobbles of granite, arkose and quartzite. Pb-isotope ages on zircons from granite cobbles in the range of 1.1–2.3 Ga (Dixon, 1981). These conglomerate beds attest to the proximity of a continental mass during the generation of the volcanic rocks; they may be glaciogenic.

Younger metavolcanics

The YMV are a dominantly andesitic volcanic sequence which overlies and interfingers with the immature metasediments. In the field, the YMV are distinguished from the OMV by the abundance of porphyritic and more felsic lithologies in the former succession. The abundance of volcanoclastic sediments, and the paucity of large pillows further distinguishes the YMV from the OMV. The absence of serpentinites in the YMV contrasts with their more general association with the OMV.

Interpretation of the thickness of the YMV is complicated by its deformation. West of Wadi El Dabbah, the section is at least 1.3 km thick. In Wadi Arak, the YMV is at least 5 km thick, and is overlain by 500–1000 m of rhyodacite. In Wadi El Mahdaf, the YMV is 1.2 km thick before it is terminated by a fault. Thus, the YMV is everywhere over 1 km thick.

GEOCHEMISTRY OF THE METAVOLCANIC ROCKS

Secondary remobilization

Three metamorphic facies are recognized in both the OMV and YMV: (1) prehnite—pumpellyite; (2) greenschist; and (3) epidote—amphibolite. Metamorphic conditions have been discussed in greater detail elsewhere

TABLE I

Analyses of older metavolcanic pillowed flows

	2A	26B	56B	60A	64O	64P	64R	85	91B
SiO ₂	51.9	57.3	48.9	50.4	54.8	54.6	56.8	52.0	55.2
TiO ₂	1.21	0.87	1.20	1.52	1.49	1.51	1.15	1.56	1.93
Al ₂ O ₃	16.5	16.4	17.0	17.1	15.6	15.9	15.3	13.6	14.7
Fe ₂ O ₃	2.70	2.69	5.06	4.07	3.62	3.64	2.60	1.38	2.71
FeO	6.26	3.15	5.29	2.23	3.42	3.79	5.31	8.55	3.75
MnO	0.13	0.11	0.17	0.11	0.13	0.14	0.14	0.17	0.13
MgO	5.96	3.92	3.96	2.28	4.50	4.28	6.70	6.48	4.19
CaO	12.00	8.36	11.54	20.12	10.80	11.45	6.11	9.41	7.77
Na ₂ O	3.0	6.3	3.1	2.5	4.6	4.3	4.5	4.0	5.5
K ₂ O	0.25	0.08	0.24	0.07	0.05	0.09	0.29	0.17	0.07
H ₂ O ⁺	1.07	1.45	1.31	1.00	1.68	0.90	2.03	0.49	1.65
H ₂ O ⁻	0.16	0.01	0.22	0.05	0.02	0.22	0.34	0.17	0.18
P ₂ O ₅	0.07	0.07	0.19	0.16	0.11	0.10	0.11	0.15	0.17
CO ₂	1.25	0.43	1.10	1.15	0.50	0.23	0.42	0.02	1.47
Total	102.46	101.14	99.28	102.76	101.32	101.15	101.80	98.15	99.42
Rb	2.7*	0.52*	3.9*	0.33*	0.91*	1.26*	5.0*	1.8*	0.71*
Sr	262	95	376	49	66	104	161	95	101
Ba	47.6*	23.6*	74.0*	19.54*	11.1*	23.4*	88.8*	35.3*	40.2*
Zr	80	45	91	105	89	99	77	108	123
V	269		413				286	337	373
Cr	237	394	41	285	237	230	278	272	214
Ni	103	182	34	117	258	226	197	98	122
Cu	136	47	121	49	54	54	52	66	61
Zn								92	118
FeO*/MgO	1.46	1.42	2.49	2.58	1.48	1.65	1.14	1.51	1.48
K/Rb	769	1277	511	1761	456	593	481	784	818
K/Ba	44	28	27	30	37.4	32	27.1	40	14.5

* Analysis by isotope dilution.

TABLE II
Analyses of older metavolcanics

	Dikes				Metagabbroes					
	27	90E	94L	28	64J	93B	111A	111B		
SiO ₂	52.0	49.9	52.8	48.2	48.7	48.7	49.1	48.2		
TiO ₂	1.72	2.90	2.25	1.09	0.89	0.83	0.13	0.13		
Al ₂ O ₃	15.0	15.0	16.1	15.7	17.2	16.7	18.1	16.9		
Fe ₂ O ₃	4.16	4.21	3.76	3.57	2.68	3.04	1.29	1.07		
FeO	5.96	8.54	5.63	6.19	7.12	5.46	4.80	4.73		
MnO	0.13	0.18	0.16	0.15	0.16	0.14	0.13	0.13		
MgO	6.89	4.25	3.03	5.35	7.49	7.02	9.44	9.42		
CaO	6.49	7.37	7.86	12.45	11.05	11.00	14.27	13.47		
Na ₂ O	3.5	3.0	4.3	2.3	2.8	2.8	0.5	1.5		
K ₂ O	0.85	0.50	0.67	0.57	0.16	0.09	0.09	0.30		
H ₂ O ⁺	2.89	1.98	1.74	2.30	1.37	2.17	1.53	1.22		
H ₂ O ⁻	0.07	0.11	0.18	0.15	0.07	0.10	0.15	0.10		
P ₂ O ₅	0.44	0.58	0.42	0.07	0.06	0.06	0.03	0.03		
CO ₂	1.50	0.01	1.18	2.56	0.15	0.73	0.23	0.65		
Total	101.60	98.53	100.08	100.65	99.90	98.84	99.79	97.85		
Rb	10.4*	4.8*	5.8*	11.0*	2.5*	1.6*	1.79*	6.69*		
Sr	398	528	505	215	202	83	244	186		
Ba	292.3*	229.5*	277.1*	47.9*	41.9*	38.8*	40.0*	92*		
Zr	161	274	251	71	42	42	22	20		
V	204	188	194	334	289	258	171	198		
Cr	216	14	4	305	237	339	249	54		
Ni	144	25	6	77	210	153	176	218		
Cu	36	28	7	77	71	75	25	97		
Zn		168	125			90	47	51		
FeO*/MgO	1.41	2.71	2.97	1.78	1.27	1.17	0.63	0.60		
K/Rb	678	865	959	430	531	467	417	372		
K/Ba	24.1	18.1	20.1	98.8	31.7	19.3	19	27		

* Analysis by isotope dilution.

TABLE III

Analyses of younger metavolcanic flows at Wadi El Dabbah

	49B	54F	55A	55B	55C	55E	55G
SiO ₂	55.9	52.1	51.2	50.6	49.8	52.0	52.9
TiO ₂	1.47	0.63	0.63	0.77	0.57	0.69	0.67
Al ₂ O ₃	12.3	18.5	16.8	20.2	17.9	15.8	22.2
Fe ₂ O ₃	0.56	3.27	2.47	2.98	3.26	2.43	2.94
FeO	7.77	3.33	5.55	4.97	5.59	5.74	3.20
MnO	0.20	0.12	0.14	0.14	0.16	0.15	0.09
MgO	5.55	3.91	6.27	5.55	5.29	6.70	3.03
CaO	8.54	16.36	12.73	13.48	12.27	10.25	10.69
Na ₂ O	5.4	0.3	3.0	2.0	2.9	3.2	4.6
K ₂ O	0.09	0.02	0.18	0.04	1.02	1.20	0.33
H ₂ O ⁺	0.83	0.73	0.89	1.37	1.01	1.20	1.15
H ₂ O ⁻	0.10	0.11	0.10	0.11	0.11	0.12	0.10
P ₂ O ₅	0.19	0.07	0.10	0.09	0.08	0.10	0.10
CO ₂	0.00	0.01	0.01	0.15	0.00	0.00	0.66
Total	98.90	99.46	100.07	102.45	99.96	99.58	102.66
Rb	2.39*	0.37*	7.77*	0.99*	56.2*	34.8*	7.4*
Sr	83	184.1*	494	230	355.2*	376.2*	392
Ba	33.7*	8.10*	43.1*	23.9*	281.8*	220.5*	66.9*
Zr		54			51	52	
V		292			293	316	
Cr	305	155	179	141	162	175	210
Ni		94			209	100	
Cu	27	38	11	30	18	8	37
FeO*/MgO	1.49	1.60	1.24	1.38	1.61	1.18	1.93
K/Rb	311	449	192	335	151	286.2	370
K/Ba	22.2	20.5	34.7	13.9	30.0	45.2	40.9

*Analysis by isotope dilution.

(Stern, 1979a), but in general it cannot be assumed that the present chemical composition of these rocks is the same as that prior to metamorphism. Recognition of primary geochemical characteristics of the metavolcanic succession is critical for the successful reconstruction of Late Precambrian igneous processes and tectonic environments in the CED. The determination of the extent to which a particular element has been remobilized requires intensive investigations at the outcrop scale (e.g. Smith, 1968; Hellman et al., 1977) or experimentally (e.g. Bischoff and Dickson, 1975). It is unlikely that the results of such studies are more generally applicable to systems with differing physico-chemical environmental parameters (e.g. T, P, X_{H_2O} , X_{CO_2} , etc.).

TABLE IV

Analyses of younger metavolcanic flows at Wadi Massar

	190D	190E	190F
SiO ₂	78.4	69.2	57.9
TiO ₂	0.10	0.46	0.74
Al ₂ O ₃	13.6	16.3	15.6
Fe ₂ O ₃	0.31	1.26	2.98
FeO	0.32	1.56	4.27
MnO	0.02	0.03	0.12
MgO	0.16	1.79	6.44
CaO	0.55	3.11	8.83
Na ₂ O	4.8	3.2	1.7
K ₂ O	1.83	2.24	0.35
H ₂ O ⁺	0.73	1.29	2.04
H ₂ O ⁻	0.16	0.12	0.25
P ₂ O ₅	0.03	0.10	0.11
CO ₂	0.32	0.90	1.56
Total	101.33	101.56	102.89
Rb	46.5 *	38.0*	5.4*
Sr	81.0 *	580.4*	246.6*
Ba	309.9 *	551*	71.7*
Zr		99	79
V		67	224
Cr	12	18	148
Ni		4	142
Cu		6	58
FeO*/MgO	3.74	1.51	1.10
K/Rb	327	489	538
K/Ba	49.0	33.7	40.5

* Analysis by isotope dilution.

The approach to the problem of secondary elemental remobilization in the Egyptian metavolcanics has been two-fold. First, specimens analyzed were those with the least traces of visible alteration. These traces would include the presence of colored domains (Smith, 1968), extensive carbonate veining, or obvious penetrative deformation. Second, reconstruction of the original chemical compositions is based on the 'immobile elements', especially Ti, P, Cr, V and Zr, the rare earths, and to a lesser extent, Ni and Ba. While any element may be remobilized under the appropriate geochemical conditions, experimental studies show that these are among the least affected during metamorphism. In contrast, Si, Mg, Ca and the alkalis are generally much

TABLE V

Analyses of younger metavolcanic flows at Wadi Arak

	13	14	15	16	17	25	41
SiO ₂	78.0	50.4	58.7	56.0	63.4	54.9	78.4
TiO ₂	0.13	1.58	0.96	1.20	1.05	1.53	0.03
Al ₂ O ₃	13.3	16.3	16.5	16.5	15.5	16.6	13.4
Fe ₂ O ₃	0.33	2.8	2.78	3.03	2.67	2.62	0.69
FeO	0.16	5.54	3.54	4.25	3.33	5.19	0.33
MnO	0.02	0.13	0.10	0.12	0.09	0.13	0.03
MgO	0.14	5.42	3.68	4.81	1.76	5.17	0.33
CaO	0.20	6.65	6.34	7.62	3.82	7.67	1.32
Na ₂ O	3.6	5.8	3.5	2.6	3.9	3.0	1.6
K ₂ O	2.01	0.17	0.87	1.15	2.01	0.75	3.30
H ₂ O ⁺	0.51	2.35	1.73	1.70	1.44	1.11	1.23
H ₂ O ⁻	0.04	0.20	0.13	0.05	0.10	0.16	0.32
P ₂ O ₅	0.04	0.31	0.19	0.25	0.25	0.27	0.02
CO ₂	0.01	1.60	0.60	0.01	0.01	0.23	1.26
Total	98.49	99.25	99.62	99.29	99.33	99.33	102.26
Rb	34.3*	5.1*	16.6*	22	28	8	130.3*
Sr	107.3*	298.8*	577.7*	303	424	364	66*
Ba	275*	78*	186*	326*	558*	312*	
Zr	122	179	140	148	301	160	
V	13	240	158	199	94	235	
Cr	12	128	52	107	12	128	12
Ni	2	102	49	78	13	97	6
Cu	2	33	27	30	20	32	1
Zn	48	117		101	105	84	
FeO*/MgO	3.26	1.49	1.59	1.45	3.26	1.46	2.88
K/Rb	486	277	435	434	596	1152	210
K/Ba	60.7	18.1	38.8	29.3	29.9	29.5	

* Analysis by isotope dilution.

more easily affected by such environments (for instance see Bischoff and Dickson, 1975; Pearce and Cann, 1971; Hart et al., 1974; Humphris and Thompson, 1978).

It is worth noting that the approach followed here is a conservative one, and that in many instances the CED volcanics may have experienced very little secondary elemental redistribution. This is especially true for the YMV, where good Rb—Sr isochrons with low initial ⁸⁷Sr/⁸⁶Sr indicate limited mobility of the labile elements (Stern, 1979a). For this reason, some use is made of the more labile elements, i.e. Fe, Mg, Ni, K, Rb, Ba and Sr, although conclusions regarding their significance are less compelling. These data are then used to reconstruct the crustal environment and tectonic

TABLE VI

Analyses of younger metavolcanic flows at Wadi El Mahdaf

	106D	106E	109J	109K	109L	109N	109O	109P	109R	109S	109W	109X
SiO ₂	58.6	66.0	64.2	62.8	77.8	63.1	61.6	67.2	77.5	60.4	79.7	60.3
TiO ₂	1.08	0.67	0.49	0.49	0.08	0.48	0.50	0.39	0.07	0.73	0.05	0.85
Al ₂ O ₃	15.7	15.9	16.5	15.9	12.2	15.9	15.7	16.5	12.8	14.8	12.0	15.4
Fe ₂ O ₃	2.28	1.10	1.64	1.77	0.51	1.68	1.31	1.08	0.57	1.69	0.53	1.77
FeO	3.29	1.96	2.02	2.18	0.44	2.35	2.43	1.59	0.63	3.56	0.36	3.51
MnO	0.06	0.06	0.06	0.07	0.02	0.07	0.06	0.04	0.02	0.10	0.02	0.09
MgO	4.38	1.21	3.19	3.65	0.05	3.66	3.99	2.15	0.09	4.19	0.11	3.75
CaO	4.76	2.50	3.61	3.92	1.99	4.66	5.22	2.57	0.94	6.45	0.98	4.74
Na ₂ O	5.0	5.6	5.0	4.1	1.2	4.6	4.2	5.3	2.6	3.2	2.2	4.2
K ₂ O	0.98	1.25	1.14	1.18	2.77	0.60	0.61	1.33	2.84	0.74	2.16	1.68
H ₂ O ⁺	1.88	1.12	1.14	1.90	0.91	1.30	1.95	1.49	0.64	1.89	0.58	1.90
H ₂ O ⁻	0.15	0.17	0.21	0.16	0.26	0.19	0.11	0.36	0.15	0.22	0.25	0.21
P ₂ O ₅	0.28	0.19	0.13	0.15	0.02	0.15	0.14	0.13	0.02	0.16	0.02	0.21
CO ₂	0.20	0.26	0.29	0.10	1.85	0.37	0.31	0.10	0.73	0.01	0.76	0.29
Total	98.64	97.99	99.62	97.67	100.1	99.11	98.13	100.23	99.60	98.14	99.72	98.90
Rb	14.6*	19.1*	18.8*	18.6*	72.5*	13.1*	10.2*	24.5*	78.2*	17.2*	79.6*	27.7*
Sr	1029	388.3*	510	510	110.1*	511	217	463	49.8*	372	44.7*	456.3*
Ba	471.0*	325.6*	331.9*	335.2*	108.0*	220.5*	236.0*	379.2*	95.9*	261.3*	59.0*	419.1*
Zr	198	183	121	130	168	121	117	136	212	122	149	177
V	149	95	88	105	22	99	95	73	< 10	128	14	141
Cr	111	56	113	140	10	135	152	19	6	121	6	108
Ni	75	23	61	71	3	72	77	22	4	67	1	58
Cu	37	31	16	27	1	50	36	34	4	36	2	7
Zn	77	66	74	74	65	67	72	62	113	65	49	74
FeO*/MgO	1.22	2.44	1.08	1.03	18.0	1.06	0.91	1.19	12.7	1.21	7.61	1.36
K/Rb	557	543	503	527	317	380	496	451	301	357	225	503
K/Ba	17.3	31.9	28.5	29.2	213	22.6	21.5	29.1	24.6	23.5	304	33.3

* Analysis by isotope dilution.

setting in which they were erupted, and to examine the processes by which such melts were generated.

Analytical techniques

Analytical techniques for major and minor elements were largely based on combined use of XRF and AA as outlined by Clague (1974). The oxidation state of iron was determined by titration (Peck, 1964). Water and carbon dioxide were determined after the methods of Shapiro and Brannock (1956). Trace elements, including the REE, were analyzed by several different methods. Cu and Rb were determined using atomic absorption spectrometry calibrated with various U.S.G.S. standards. Sr, Cr and Zn were analyzed using the AA by the method of additions (Chow and Thompson, 1955). V, Ni and Zr were analyzed by X-ray fluorescence on pressed powder pellets. Some Rb and Sr, and all Ba, were analyzed by isotope dilution (Chow et al., 1980). REE were analyzed both by isotope dilution at the U.S.G.S. laboratories in Denver, using methods described by Divis (1975), and by instrumental neutron activation analysis at Washington University using methods outlined by Jacobs et al. (1977). Analytical error is defined as the

TABLE VII

Rare earth elements (older metavolcanics)

	2A*	28*	56B**	64J*	64O**	64R**	85*	91B**
La			8.51		2.17	2.39		3.86
Ce	8.23	5.04	18.30	2.037	7.12	7.60	15.541	12.97
Nd	8.114	5.706		3.664			11.310	
Sm	3.048	2.447	3.07	1.806	3.47	2.52	3.432	4.53
Eu	1.170	0.994	1.17	0.763	1.21	0.88	1.202	1.58
Gd				3.092			4.310	
Tb			0.86		1.06	0.89		1.30
Dy	4.240	4.557		3.706			4.643	
Er	2.636			2.385			2.694	
Yb	2.423	2.804	2.42	2.312	3.79	2.47	2.364	3.89
Lu			0.35		0.56	0.37		0.62
Values normalized to chondrites								
La			25.79		6.58	7.24		11.70
Ce	9.35	5.74	20.80	2.31	8.09	8.64	17.66	14.74
Nd	13.52	9.51		6.11			18.85	
Sm	16.84	13.52	16.96	9.98	19.17	13.92	18.96	25.03
Eu	16.96	14.41	16.96	11.06	17.54	12.75	17.42	22.90
Gd				12.42			17.31	
Tb			18.30		22.55	18.94		27.66
Dy	13.99	15.04		12.23			15.32	
Er	13.18			11.93			13.47	
Yb	12.12	14.02	12.10	11.56	18.95	12.5	11.82	19.45
Lu			10.29		16.47	10.88		18.24
(Ce/Yb) _N	0.77	0.41	1.72	0.20	0.43	0.70	1.49	0.76
Eu/Eu*	1.03	1.02	0.99	1.00	0.87	0.82	0.96	0.88

* Analysis by isotope dilution.

** Analysis by neutron activation.

TABLE VIII
Rare earth elements (younger metavolcanics) — andesites and dacites

	15*	17*	25*	55E**	106D**	106E**	109J**	109N**	109X**
La				4.2	17.28	15.60	11.43	11.71	14.97
Ce	25.725	58.051	27.126	10.10	41.56	36.18	24.55	25.62	35.57
Nd	14.369	31.866	16.839						
Sm	3.508	7.337	4.278	1.93	4.82	3.83	2.28	2.34	3.85
Eu	1.189	1.852	1.399	0.61	1.22	0.95	0.65	0.71	1.04
Gd	3.435	7.586							
Tb				0.51	0.95	0.47	0.22	0.36	0.78
Dy	3.318	7.692	4.717						
Er	1.878	4.360	2.676						
Yb	1.669	4.190	2.432	1.51	1.72	1.24	0.61	0.71	1.55
Lu				0.19	0.25	0.17	0.08	0.10	0.23
Values normalized to chondrites									
La				12.72	52.36	47.27	34.64	35.48	45.36
Ce	29.23	65.97	30.83	11.48	47.22	41.11	27.90	29.11	40.42
Nd	23.95	53.11	28.07						
Sm	19.38	40.54	23.64	10.66	26.63	21.16	12.60	12.93	21.27
Eu	17.23	26.84	20.28	8.84	17.68	13.77	9.42	10.29	15.07
Gd	13.80	30.47							
Tb				10.85	20.21	10.00	4.68	7.66	16.60
Dy	10.95	25.39	15.57						
Er	9.39	21.8	13.38						
Yb	8.35	20.95	12.16	7.55	8.6	6.2	3.05	3.55	7.75
Lu				5.59	7.35	5.00	2.35	2.94	6.76
(Ce/Yb) _N	3.50	3.15	2.54	1.52	5.49	6.63	9.15	8.2	5.22
Eu/Eu*	1.04	0.76	0.94	0.82	0.72	0.57	0.95	0.92	0.84

* Analysis by isotope dilution.

** Analysis by neutron activation.

absolute deviation of triplicate standards from working curves or that of triplicate unknowns from mean values, whichever is greater, as analyzed at the 95% confidence level. For oxides and elements determined by XRF and AA, these are: SiO₂, 0.55%; TiO₂, 0.02%; Al₂O₃, 0.31%; FeO*, 0.19%; MnO, 0.01%; MgO, 0.26%; CaO, 0.15%; Na₂O, 0.20%; K₂O, 0.06%; P₂O₅, 0.01%; Rb, 2 ppm; Sr, 10 ppm; Ni, 3 ppm; Cr, 8 ppm; Zn, 10 ppm; V, 15 ppm; Zr, 15 ppm; Cu, 2 ppm; and FeO (by titration), 0.20%. Water and Rb, Sr and Ba determined by isotope dilution are considered in references cited for their determination.

Forty-six analyses of metavolcanic rocks from the CED are reported in Tables I–VI which list data for the major and minor oxides as well as for the trace elements Rb, Sr, Ba, Zr, V, Cr, Ni, Cu and Zn. REE data are presented for 21 representatives in Tables VII–IX.

TABLE IX

Rare earth elements (younger metavolcanics) — rhyodacites

	13*	109L**	109R**	109W**
La		22.72	22.16	6.25
Ce	65.282	65.59	61.58	22.38
Nd	29.740			
Sm	5.993	12.82	14.54	7.58
Eu	0.817	0.15	0.50	1.75
Gd	4.591			
Tb		3.04	3.10	2.00
Dy	2.950			
Er	2.029			
Yb	2.780	9.42	11.95	11.73
Lu		1.46	1.77	1.76
Values normalized to chondrites				
La		68.85	67.15	18.94
Ce	74.18	74.53	69.98	25.43
Nd	33.80			
Sm	33.11	70.83	80.33	41.90
Eu	11.84	2.17	7.25	25.36
Gd	18.44			
Tb		64.68	65.96	42.55
Dy	9.74			
Er	10.15			
Yb	13.9	47.10	53.75	58.65
Lu		42.94	52.06	51.76
(Ce/Yb) _N	5.34	1.58	1.30	0.43
Eu/Eu*	0.46	0.03	0.10	0.60

* Analysis by isotope dilution.

** Analysis by neutron activation.

DISCUSSION

Crustal environment and tectonic setting

The tectonic setting and crustal environment in which the volcanic rocks of the CED were erupted must be understood in order to address broader problems of Pan-African crustal evolution in NE Africa. Assuming that igneous activity 600 Ma in the CED manifested sea-floor spreading/plate tectonic activity similar to that now operating, the data presented here on volcanic rocks can be used to identify specific tectonic settings. These include island arcs, mid-ocean ridge spreading centers, or intra-plate 'hot-spots'. The latter group would include magma-types erupted on continental crust. These data can also be used to evaluate the crustal environment in which these volcanics formed, i.e. older continental crust or young ocean floor.

Miyashiro (1974, 1975) suggested that relationships between the ferromagnesian elements during fractionation are diagnostic of tholeiitic and calc-alkaline volcanic suites and may persist through low grades of metamorphism. In his scheme, the enrichment of FeO^* and TiO_2 with increasing fractionation, as indicated by increasing FeO^*/MgO in the tholeiitic series, contrasts with the depletion of these oxides with increasing FeO^*/MgO in calc-alkaline suite. Figure 4 shows the OMV and YMV successions plotted on such diagrams. The OMV show a strong tendency towards Fe- and Ti-enrichment, with slight increases in FeO^*/MgO . This trend mimics that of modern mid-ocean ridge basalts, or MORB (Engel et al., 1965). In contrast, the YMV show iron and titanium depletion with advancing fractionation. With one exception, analyses of the YMV plot in the field defined by modern calc-alkaline lavas.

The affinities of the OMV and YMV may be further considered by examination of the relative behaviors of titanium and phosphorus. Both are relatively immobile during low grades of metamorphism. Bass et al. (1973) considered that modern oceanic volcanic rocks could be distinguished on the basis of the relative and absolute abundance of these elements. Figure 5 shows the fields of modern MORB, ocean-island tholeiites, and alkali basalts. The OMV generally plot in the field of MORB. Notable exceptions are three analyses of dikes which demonstrate closer affinities to the field of alkali basalts. Some metagabbros are depleted in Ti and P and thus plot outside of the field for MORB. Nevertheless, the similarity of the OMV to modern MORB is apparent. The YMV have a greater P/Ti ratio than do the OMV, defining a trend which occupies part of the MORB field but is steeper than and distinctive from that of the OMV. The trend defined by the YMV andesites and dacites is similar to that of the Cascades calc-alkaline igneous series.

Pearce and Cann (1971, 1973) proposed that modern volcanic rocks can be discriminated on the basis of their Ti and Zr concentrations. On this

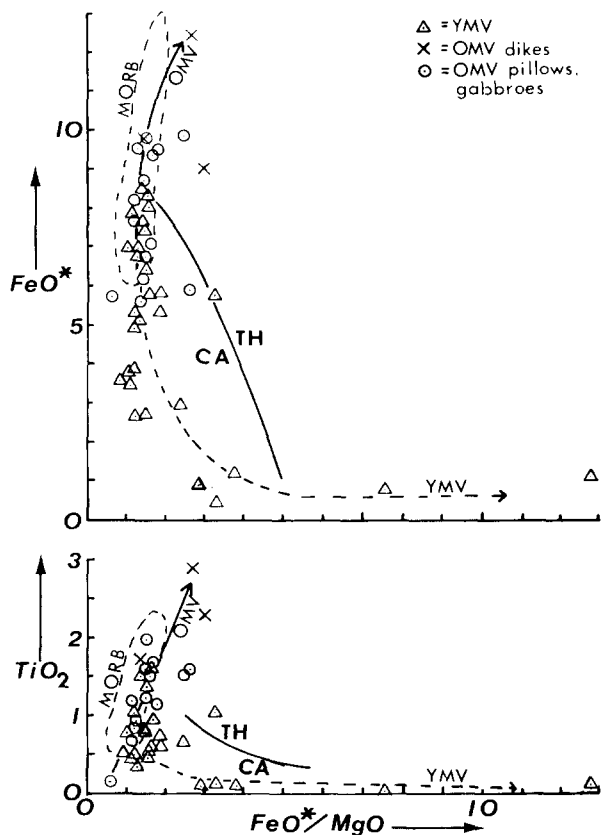


Fig. 4. Ferromagnesian major element discriminant diagrams, modified after Miyashiro (1974). Note that the OMV tend towards TiO_2 and FeO^* enrichment with advancing fractionation (increasing FeO^*/MgO), similar to that shown by modern MORB. The YMV show depletions of TiO_2 and FeO^* with advancing fractionation, characteristic of the calc-alkaline suite.

basis, analyses for OMV and YMV rocks generally plot in the fields of MORB and calc-alkaline rocks, respectively (Fig. 6). Three analyses of the OMV dikes are not shown, since these would plot in an alkalic field to the right of the diagram. It is noteworthy that while the OMV and YMV trends are generally distinctive, there is some overlap in the field of island-arc tholeiites (IATH).

Pearce (1975) noted that, while it is difficult to distinguish between MORB and IATH on the basis of most geochemical parameters, Cr is less abundant in arc tholeiites than in MORB at a given TiO_2 concentration. Figure 7 shows the Ti—Cr boundaries between MORB and island-arc lavas, with plotted analyses of the OMV and YMV. Again, the OMV are restricted to the field for modern MORB. The YMV generally plot in the field for modern arc lavas, but the relatively Cr-rich composition of the YMV is evident since

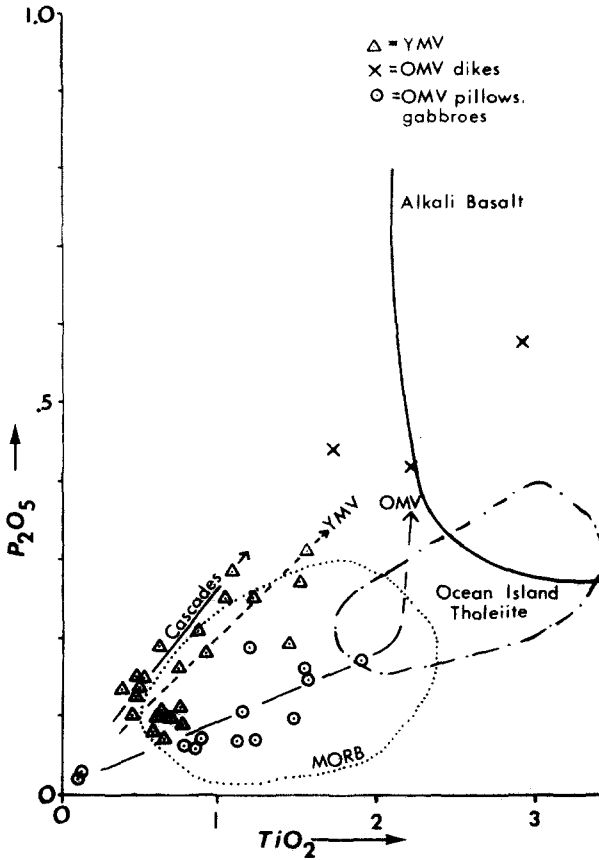


Fig. 5. TiO_2 and P_2O_5 discriminant diagram, with fields of modern MORB, ocean-island tholeiite, and alkali basalt shown for comparison with OMV and YMV analyses (Bass et al., 1973). Note that while most of the OMV plot in the field for MORB, OMV dikes are more alkaline. YMV analyses trend parallel to that of the Cascades (Smith and Carmichael, 1968).

most analyses are concentrated towards the upper limits of chromium concentrations in modern arc lavas; some even plot in the field of MORB. The Cr-rich nature of the YMV is especially apparent when we consider that while this diagram was designed for basalts, most of the YMV are andesitic.

A similar discrimination between metamorphosed igneous rocks of various affinities has been proposed on the basis of relationships between V and Cr. Figure 8 shows boundaries defined for modern igneous rocks by Miyashiro and Shido (1975). All but one analysis of the OMV plot within the field of MORB; this exception nevertheless manifests effects of tholeiitic fractionation. The YMV generally plot in a field transitional between tholeiitic and calc-alkaline rocks, a field also occupied by modern arc-related lavas of the western Pacific (Taylor and White, 1966).

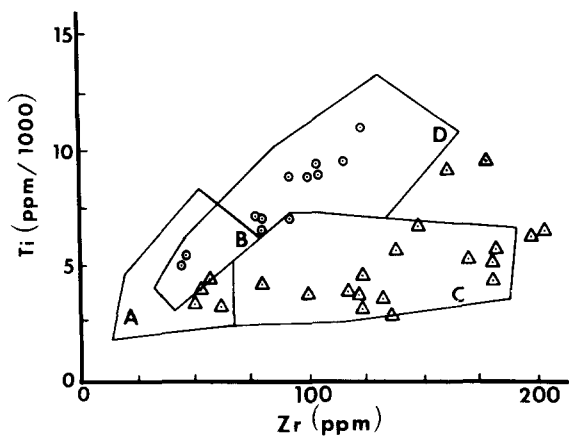


Fig. 6. Ti—Zr discriminant diagram, with analyses of OMV (circles) and YMV (triangles) plotted. MORB plot in fields B and D while arc tholeiites and calc-alkaline lavas plot in fields A, B and C (Pearce and Cann, 1973). Note that the OMV are indistinguishable from modern MORB while the YMV are most similar to modern arc lavas.

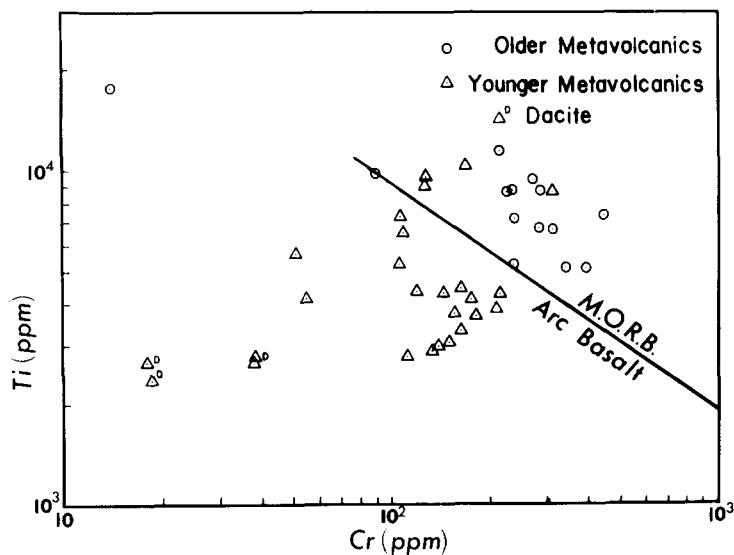


Fig. 7. log Ti—log Cr discriminant diagram, with OMV and YMV analyses plotted. Note that the OMV are indistinguishable from modern MORB while the YMV are similar to island-arc lavas (from Pearce, 1975).

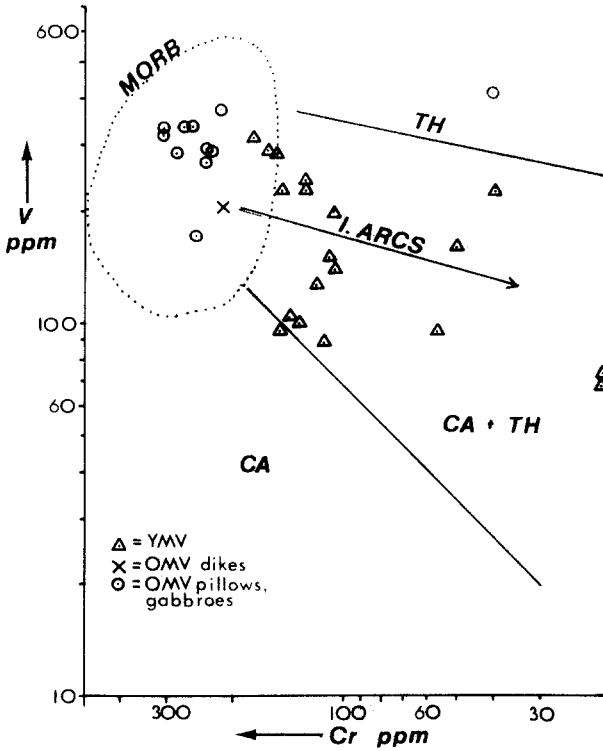


Fig. 8. log V—log Cr discriminant diagram, with OMV and YMV analyses plotted. Note that OMV analyses fall within the field of modern MORB while YMV fall in the field of calc-alkaline and tholeiitic are lavas (CA + TH). Trend of modern arcs of the Western Pacific is shown for comparison to YMV (Taylor and White, 1966) (modified after Miyashiro and Shido, 1975).

Finally, Ni abundances in the OMV are similar to oceanic tholeiites, while those of the YMV are more akin to island-arc rocks (Fig. 9).

Rare-earth element (REE) patterns for six metabasalts and two metagabbros from the OMV are shown in Fig. 10. With two exceptions, the patterns show strong depletions in the light REE with a chondrite-normalized Ce/Yb ratio ($(\text{Ce}/\text{Yb})_N$) of less than unity (0.20–0.77). There are no indications of the Ce anomalies expected from sea-floor weathering (Masuda and Nagasawa, 1975; Menzies et al., 1977). On this basis, the REE data are interpreted to be the same as concentrations in the original lavas. Many investigators have noted the distinctive REE patterns of MORB (Frey et al., 1968; Gast, 1968; Kay et al., 1970; Schilling, 1971). Ocean-floor tholeiites commonly manifest light-REE depletions and have nearly-flat heavy-REE patterns of 10–20 times chondritic abundances. More recent investigations have shown that light-REE-enriched MORBs also exist (Group II of Bryan and Thompson,

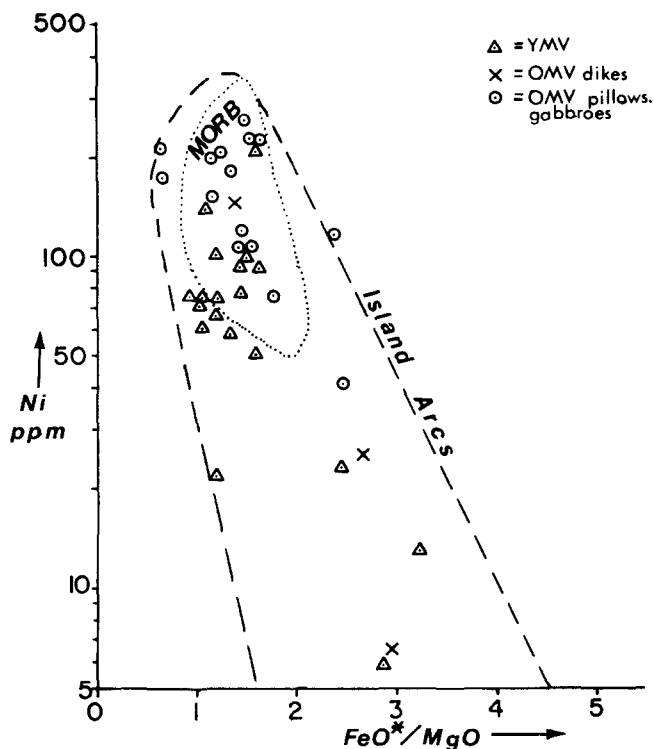


Fig. 9. Ni-FeO*/MgO diagram. Note that most of the OMV fall in the field of modern MORB while the YMV plot in the field of island arc lavas (after Miyashiro and Shido, 1975).

1976; White and Bryan, 1977). These lithologies are less common than the light-REE-depleted type of MORB. Figure 10 shows that nearly all the REE patterns of the Egyptian OMV fall within an envelope for typical light-REE-depleted MORB.

REE data for the nine meta-andesites and metadacites of the YMV are presented in Fig. 11. Compared to OMV, the YMV are light-REE-enriched and heavy-REE-depleted. In all cases, $(Ce/Yb)_N$ is greater than 1.5, and may even be greater than 10. Cerium anomalies are not evident in YMV REE patterns, indicating that these elements were probably not remobilized. Also shown in Fig. 11 is the field occupied by recent calc-alkaline lavas of the Cascades (Condie and Swenson, 1973). The similarity of the YMV to the Cascades emphasizes the strong calc-alkaline affinities of the YMV already apparent from considerations of their major and trace-element compositions.

Finally, REE data for four YMV metarhyodacites ($SiO_2 = 77.5-79.7\%$) are listed separately from those of the rest of the YMV, since their major,

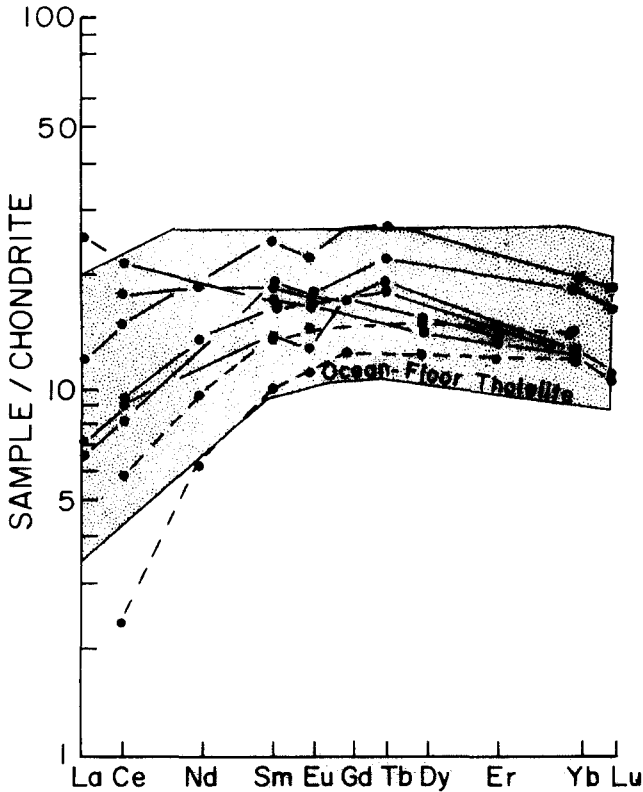


Fig. 10. REE patterns for OMV metabasalts (solid lines) and OMV metagabbros (dashed lines). Field of MORB (mid-ocean ridge basalt) shown for comparison (Schilling, 1971). Note that the OMV show similar light-REE depletions and plot within the MORB field.

trace and REE characteristics are so distinctive (see Table IX, Fig. 12). The four patterns show overall enrichments in REE relative to both the OMV and YMV intermediate rocks. In contrast to the OMV and YMV, the REE patterns for the metarhyodacites showed no consistent trends. They may be either light-REE-enriched (13), flat (109L, R) or light-REE-depleted (109W). All four samples show strong depletions in Eu.

The sum of geochemical characteristics for Egyptian metavolcanics indicates that the OMV is most similar to modern MORB while the YMV are indistinguishable from modern calc-alkaline, andesitic sequences. The case for the YMV is especially strong, and if plate-tectonic processes were responsible for the generation of YMV magmas, this must have occurred in an island-arc setting. The strong similarity of the OMV to modern MORB, on the other hand, does not preclude the possibility that other types of modern crustal environments are also possible analogues for that of the OMV. REE data from basalts of modern back-arc basins are variable, with both enriched

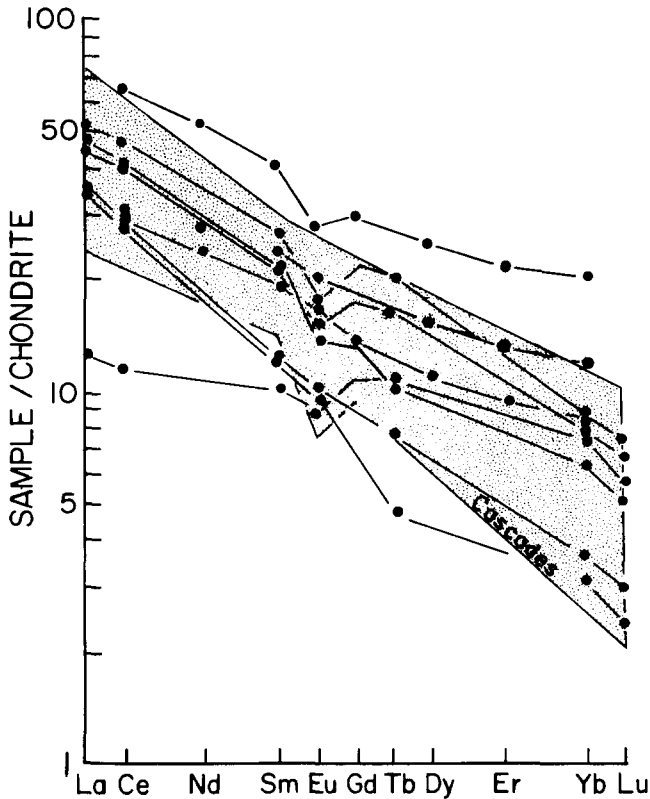


Fig. 11. REE patterns for YMV meta-andesites and metadacites. REE concentrations normalized to chondritic values of Frey et al. (1968) with the exception of Dy which is after Philpotts et al. (1971). Note all show light-REE enrichments and heavy-REE depletions, similar to that of modern calc-alkaline volcanic rocks of the Cascades (Condie and Swenson, 1973).

and depleted light-REE varieties (Hart et al., 1972; Hawkesworth et al., 1977; Hawkins, 1977). Thus, the REE data for the OMV are also consistent with the hypothesis that their origin was in a back-arc basin. Furthermore, some primitive island arcs erupt basalts strongly depleted in the light REE (Jakes and Gill, 1970; Yajima et al., 1972). It is unlikely that the other geochemical discriminants used here to distinguish between MORB, island-arc rocks, and intra-plate basalts are capable of unequivocally distinguishing between MORB-like rocks of oceanic-spreading centers, back-arc basins, and island-arc tholeiites, although both Ti—Cr and Ti—Zr diagrams argue against the OMV being arc tholeiites. In this problem, consideration of sediments associated with the OMV are useful. These sediments indicate that: (a) older continental crust was at all times nearby; (b) calc-alkaline volcanic terranes volumetrically dominated sedimentary sources; and (c) a restricted geometry

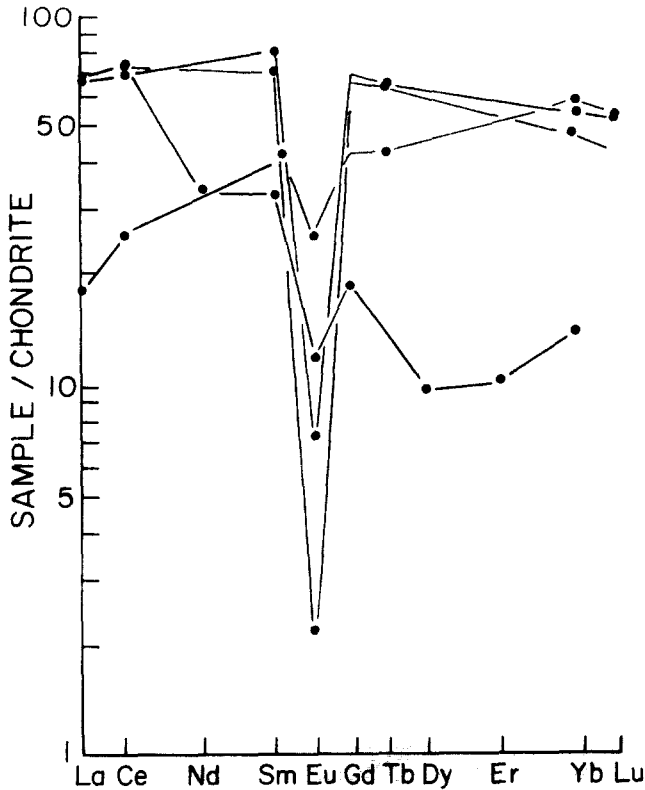


Fig. 12. REE patterns for YMV metarhyodacites. Note extreme, negative Eu anomalies. While total REE contents are high, light-REE depletions are common.

for the basin is indicated for the generation of the sedimentary Fe-ores (Stern, 1979a). Models calling for the OMV to be erupted in a great ocean basin such as the present Atlantic or Pacific must be rejected. Similarly, models calling for a setting in very primitive island arcs, analogous to those of the Western Pacific must be rejected as well. Models calling for the OMV to have been erupted in an incipient rifting environment such as the modern Red Sea or Gulf of California or a restricted back-arc basin near a continent are consistent with both the geochemical and sedimentological data.

Petrogenesis of the metavolcanic rocks

Discussion here focuses on the generation of OMV and YMV melts. The intention is to determine whether the OVM and YMV reflect anatexis of older continental crust or are primary melts of the mantle or perhaps subducted crust. The OMV basalts will be examined first, followed by the YMV andesites.

Older metavolcanics

Any discussion regarding the petrogenesis of the OMV must be consistent with its affinities to tholeiitic basalts of modern mid-ocean ridges, primitive arcs, and marginal basins. This discussion will thus focus on aspects of the generation of the OMV which have been discussed by investigators of MORB and MORB-like basalt.

In their pioneering work on the origin of MORB, Engel et al. (1965) noted these were the most depleted of modern lavas. They suggested that these were about 20% partial melts of a peridotitic upper mantle. Due to the similarities in especially Cr, Ni, Zr, REE, K, Rb and Ba, between the OMV and modern MORB, it is reasonable to infer that the OMV are manifestations of a similar degree of partial melting of the Late Precambrian upper mantle, which resembled the source of modern MORB.

Green and Ringwood (1967) suggested on the basis of extensive experimental investigations that high-Al MORB ($\text{Al}_2\text{O}_3 = 15\text{--}17\%$) are generated by ca. 25% fractional melting of lherzolite at a depth of about 30 km. Experiments by Kushiro (1973) indicate that such magmas are formed by anhydrous melting of spinel-lherzolite at depths of less than 50 km. These results have found further confirmation in subsequent experiments on the melting of natural lherzolites, and there is now a consensus that MORB-type tholeiites are generated by ca. 20% fractional fusion of peridotite at mantle depths of less than 60 km (Kushiro et al., 1972). Insofar as the relatively immobile element geochemical characteristics of the OMV are indistinguishable from those of modern MORB, the OMV are interpreted as the products of ca. 20% fractional fusion of anhydrous, depleted mantle peridotite at depths of less than 60 km.

Younger metavolcanics

The calc-alkaline affinities of the YMV are apparent from their chemical composition as well as from the composition of relict clinopyroxenes (Stern, in press). Calc-alkaline volcanic rocks are now erupted at convergent plate margins, above subduction zones. Several processes have been proposed to account for the genesis of modern andesites. These include:

- (1) Anatexis of older sialic crust;
- (2) Low-pressure fractional crystallization of basaltic magma;
- (3) Amphibole-controlled fractionation of basalt;
- (4) High-pressure melting of eclogite;
- (5) Fractional fusion of hydrous peridotite.

These possible modes of generation of YMV andesites are examined briefly in the following. For a more detailed discussion, see Stern (1979a).

Anatexis of sialic crustal materials. Considerations of the field evidence coupled with YMV chemical and isotopic compositions indicate that YMV meta-andesites were not generated by anatexis of underlying sialic crust.

The YMV were erupted on a substrate of immature metasediments deposited on OMV basalts; the affinities of this substrate to modern ocean-floor basalts makes it extremely unlikely that these were erupted on continental crust. Geochemical data for the YMV may also be used to evaluate this hypothesis. The experimental studies of Green and Ringwood (1968) indicate that a relatively high degree of partial melting (35–85%) of mafic rocks of a granulite-facies lower crust can produce andesitic liquids. Assuming that the source at the base of the Egyptian Late Precambrian continental crust corresponded to garnet granulite and had a trace element composition similar to that estimated for the lower crust (Leyreloup et al., 1977), a model calculation of partial melting for Ba, REE and some transition elements can be made. This shows that the expected concentrations of light REE are too high, and most transition elements are too low to account for the observed concentrations in the YMV andesites. Especially since the V, Cr and Ni enrichments of the YMV meta-andesites are characteristic, this argues against an origin for the YMV by the large-scale fractional fusion of the lower crust. This agrees with the conclusion of Chikhaoui et al. (1978) on the origin of similar Late Precambrian andesites from Algeria.

Finally, Rb–Sr systematics of the YMV may be used to evaluate the hypothesis that these andesites are anatectic melts of the lower crust. If it is assumed that, during the Late Precambrian, the lower continental crust in Egypt was similar in age to the granulites now exposed in SW Egypt, it would have a low initial $^{87}\text{Sr}/^{86}\text{Sr}$ (2.6×10^9 y; $i = 0.7033$; Klerkx and Deutsch, 1977). As upper and lower limits, let us assume that the lower crust in Egypt had an Rb/Sr ratio between that of the lower crust beneath France (Rb/Sr = 0.103; Leyreloup et al., 1977) and that beneath Western Australia (Rb/Sr = 0.265; Lambert and Heier, 1967). During the 2.6×10^9 y prior to the eruption of the YMV, such a lower crust would have evolved a $^{87}\text{Sr}/^{86}\text{Sr}$ between 0.7145 and 0.7521. The very low initial $^{87}\text{Sr}/^{86}\text{Sr}$ of the YMV (0.7028–0.7030; Stern, 1979a) cannot have been derived by the anatexis of such a lower crust. Rb/Sr in the upper crust should be as high as, or higher than, that of the lower crust, ruling out the anatexis of any crustal material to generate YMV melts.

Melting of graywackes such as those overlying the OMV could generate andesites with requisite chemical and isotopic characteristics, but this explanation is not favored for two reasons:

(1) There is no evidence that these sediments were exposed to thermal regimes required for melting. Except where these contact intrusive granites, the sediments are metamorphosed to epidote-amphibolite facies or less.

(2) These sediments are volcanogenic, derived from the eruption and erosion of an andesitic volcanic terrane. If melting of sediments generated YMV andesites, then an explanation is required for the generation of the low $^{87}\text{Sr}/^{86}\text{Sr}$ andesites which eroded to form the sediments. Ultimately, this 'chicken-and-egg' argument requires the generation of low $^{87}\text{Sr}/^{86}\text{Sr}$ andesitic melts without the participation of sediments.

Fractional crystallization of basaltic magmas at crustal levels. This hypothesis for the origin of the andesitic liquids is attractive in light of the low $^{87}\text{Sr}/^{86}\text{Sr}$ of the YMV as well as by the presence of occasional gabbroic xenoliths. However, simple calculations of the behavior of K, Rb, Ba, REE, Cr and Ni are incompatible with such a model. If we take the magmatic evolution of Agrigan in the Mariana island arc as an example, then the generation of the major element characteristics of andesites by such a process requires that more than 70% of the original basaltic liquids be precipitated as olivine, pyroxene, plagioclase, and magnetite (Stern, 1979b). Removal of olivine, magnetite and pyroxene should deplete the Ni, V and Cr contents of residual liquids. Moderate Sr and Eu depletion should result from the removal of plagioclase. In contrast, YMV andesites show high concentrations of Cr and Ni. Eu anomalies are uncommon and Sr is enriched ca. 400% over concentrations in OMV basalts. Furthermore, concentration of heavy REE in the YMV are less than for the OMV. These data preclude the generation of the YMV meta-andesites from the fractional crystallization of basaltic liquids similar to the OMV. In general, the amount of separation of mafic phases required to generate andesitic liquids from any basalt would deplete the Cr and Ni contents of the resultant liquids much more than is observed in the YMV. It is unlikely that the YMV meta-andesites could have been generated by shallow-level fractional crystallization of any primary mafic melt.

Amphibole-controlled fractionation of basalt. Fractional fusion of subducted, amphibolite-facies oceanic crust derived from MORB may produce liquids of andesitic composition (Allen et al., 1972; Holloway and Burnham, 1972). Crystallization of low-silica amphibole from basaltic magma could also yield andesites (Green and Ringwood, 1968; Nishimori, 1976). Models of REE-partitioning during melting of amphibolite indicate that this can cause strong depletions of the heavy REE similar to that of some YMV meta-andesites (i.e. 106 and 109 series; Table VIII). Amphibole-controlled fractionation of basalt thus can cause REE patterns similar to that produced by garnet-controlled fractionation (Arth and Hanson, 1975). The most likely candidate for a basaltic parent to the YMV would be OMV basalts metamorphosed to amphibolite facies. These would have the low $^{87}\text{Sr}/^{86}\text{Sr}$ required for the YMV parent and would be expected to be widespread beneath present YMV outcrops at the time of magmagenesis. More than 30% melting of this material would be required to generate andesitic liquids (Green and Ringwood, 1968; Holloway and Burnham, 1972). It is difficult to reconcile the heavy-REE depletion, which requires significant residual amphibole, with either the large extent of melting required or with the general enrichments of the incompatible elements (especially K) in the YMV. Potassium in an OMV-derived amphibolite would be strongly partitioned into the amphibole. At 30% melting of such a low-K parent with enough amphibole residual to cause the strong high-REE depletions, K_2O concentrations should be much lower than

observed in YMV andesites. Furthermore, andesites generated by such large degrees of melting of light-REE depleted OMV amphibolite should manifest as well the depletion of the light REE. This is not observed in the REE patterns of the YMV. These data support the conclusion that the YMV meta-andesites were not generated by the fractionation of amphibole from an OMV-like basaltic parent.

Eclogite-controlled fractionation of basalt. Green and Ringwood (1968) have suggested that andesites could be generated by a two-stage process of mantle distillation, first by the generation of MORB at oceanic-spreading centers followed by the generation of andesite by the fractional fusion of eclogite-facies MORB in subduction zones at 100–150 km depth. Gill (1974) tested this model using major- and trace-element studies. Assuming typical MORB as parent, the eclogite-fusion model satisfies the major element characteristics of Fiji andesites if a large fraction of liquid (30%) was at equilibrium with a residuum of garnet (26%) and clinopyroxene (44%). Trace element data are inconsistent with this model. Concentrations of K, Rb and Ba in the YMV are much greater than that predicted by this model. The alteration of basalts may be able to satisfy the K and Rb requirements, but Ba is always much lower in modeled liquids than in observed YMV meta-andesites. The problem is more severe for Cr and Ni. Using average concentrations of Cr and Ni in either modern MORB or the OMV and estimates of distribution coefficients for these elements in clinopyroxene and garnet (Davis and Condie, 1977), the 30% melt fraction would be expected to contain only 25–30 ppm Ni and less Cr. This is much less than that observed in the YMV andesites. Also, REE patterns expected from the fractional fusion of eclogite-facies MORB are inconsistent with those of the YMV andesites. While it would be possible for the garnetiferous residuum to cause the HREE-depletions of the YMV, there is no way to generate the smooth, LREE-enriched profiles of these rocks. Instead, the depleted nature of the parent in especially La and Ce should be reflected in derivative melts (Gill, 1974). Thus, the LIL, transition element, and REE data preclude the generation of the YMV by eclogite-controlled fractionation of MORB or OMV.

Fractional fusion of hydrous peridotite. Recent experiments indicate that melting hydrous peridotite can produce andesitic magmas (Kushiro, 1972; Kushiro et al., 1972; Mysen et al., 1974; Mysen and Boettcher, 1975). The proposal of American and Japanese investigators that andesitic melts may be direct fractional melts of hydrous upper mantle have since found support from the results of investigations by the Canberra group (Nicholls, 1974). Earlier conclusions that andesites could not be primary melts of the mantle were largely based on their low Ni and Cr concentrations. Taylor et al. (1969) noted that the low abundances of transition elements in andesites precluded their direct derivation from fractional fusion of the mantle. However, YMV meta-andesites contain concentrations of Cr (52–152 ppm) and

Ni (58–97 ppm) much greater than the 18 ppm Cr and 56 ppm Ni for the 'average andesite' of Taylor et al.; therefore, their objections to the generation of andesites directly from the mantle do not apply to the YMV.

Isotopic compositions of Sr in the YMV are consistent with a hypothesis of their derivation from the upper mantle. $^{87}\text{Sr}/^{86}\text{Sr}$ initials for this sequence average 0.7029 (Stern, 1979a). Assuming an initial $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.6998 some 4.6 Ga ago, these magmas were derived from a reservoir with an average Rb/Sr of about 0.023. This is as low or lower than estimates of the Rb/Sr of primitive mantle (Rb/Sr \approx 0.03; DePaolo and Wasserburg, 1976).

The geochemical and isotopic data are thus qualitatively consistent with an origin for the YMV meta-andesites by the fractional fusion of hydrous peridotite. An attempt to quantitatively model this hypothesis follows.

For the purposes of this model, it is assumed that the composition for the Late Precambrian upper mantle is similar to that of modern, undepleted upper mantle. This might contain ca. 1500 ppm Ni and 2500 ppm Cr (Taylor et al., 1969) and 3X chondritic abundances for Ba and REE (Kay and Gast, 1973). The model will be further dependent on the choice of peridotite mineralogy. The high Sr content and general lack of Eu anomalies in the meta-andesites suggest that the YMV were not derived from partial melting at relatively low pressures (plagioclase peridotite). Following the assumption of a source with 3X chondritic REE, the heavy-REE depletions generally observed in the YMV require a model involving garnet in the residuum. Following the model of Kay and Gast (1973), it is assumed that it consisted of 55% olivine, 25% orthopyroxene, and 10% each of garnet and clinopyroxene.

The third problem in this model of the YMV is the choice of appropriate distribution coefficients. This model will use REE data for garnet at 1275°C, 30 kbar (Shimizu and Kushiro, 1975) and pyroxenes and olivine at 950–1025°C and 20 kbar (Mysen, 1978a). Distribution coefficients for Ba are from experimental studies on olivine and orthopyroxenes (McKay and Weill, 1976) and clinopyroxene (Shimizu, 1974). The absence of experimental data for Ba in garnet requires the use of values determined for natural phenocryst-matrix pairs (Arth, 1976). Partition coefficients for Ni are for hydrous melting, $T = 1025^\circ\text{C}$, $P = 20$ kbar (Mysen, 1978b). Distribution coefficients for Cr are sensitive to ambient oxygen fugacities, but at typical values for andesitic liquids ($\log f_{\text{O}_2} = -9$ at 1000°C; Gill, 1978) can be estimated from the summary of experimental data by Irving (1978). These are taken from investigations into problems of lunar systems which typically have f_{O_2} much lower than those of terrestrial systems. Since distribution coefficients for Cr generally decrease with decreasing f_{O_2} , those used in this modeling are probably minimum values.

Finally, it is necessary to assume melting conditions and proportions of phases contributing to the melt. Kushiro et al. (1968) noted that hydrous melting stabilizes olivine on the liquidus over a wide range of pressures. The model thus assumes that olivine does not melt and that orthopyroxene,

clinopyroxene and garnet contribute in their modal proportions. Equilibrium melting is most likely under these conditions; for this reason equation 15 of Shaw (1970) is used.

The results of this model are graphically presented in Fig. 13. For comparison, the composition of a typical meta-andesite (25) is also plotted. Considering the assumptions and uncertainties in the model, these results permit the interpretation that the YMV meta-andesites reflect ca. 3–10% fractional melts of hydrous mantle. Several inconsistencies in the model, however, are worth discussing further. While most of the REE are satisfied by the model,

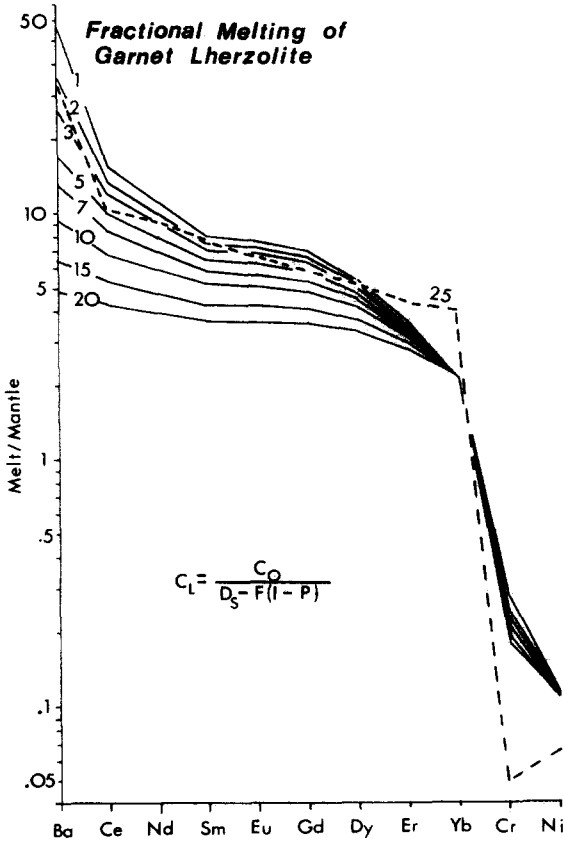


Fig. 13. Results of modeling for Ba, REE, Cr and Ni. Inferred parent peridotite consists of 55% olivine, 25% orthopyroxene, 10% garnet and 10% clinopyroxene and the composition: 1500 ppm Ni, 2500 ppm Cr (Taylor et al., 1969) and 3X chondritic abundances of Ba and REE (Kay and Gast, 1973). Expression for equilibrium non-modal melting is: $C_L = C_o / [D_s - F(1-P)]$ (Shaw, 1970). Parent is assumed to melt in the proportion 56% orthopyroxene, 22% clinopyroxene and 22% garnet. Distribution coefficients are after the results of high-pressure hydrous experimental data whenever possible (Shimizu, 1974; Shimizu and Kushiro, 1975; Arth, 1976; Schreiber and Haskin, 1976; McKay and Weill, 1976; Irving, 1978; Mysen, 1978a, b). Numbers on solid lines refer to % melting of postulated source.

Ba is relatively more enriched in the melt than predicted. Such a discrepancy might be the result of the fractionation and preferential transport of Ba, along with K and Rb, in a hydrous or CO₂-rich fluid medium (Best, 1975; Green, 1979). Er and Yb are significantly higher than that predicted from the model. This discrepancy can be reconciled if a slightly higher proportion of garnet to clinopyroxene enters the melt. Finally, the transition elements Ni and Cr are lower in the meta-andesites than predicted in the model. The discrepancy can be resolved if a small amount of refractory spinel ($D_{Cr} = 50-570$; $D_{Ni} = 5-10$; Irving, 1978) in the parent peridotite or the further precipitation of olivine and pyroxene or separation of an immiscible sulfide phase is allowed during ascent.

A minimum depth for the generation of the YMV meta-andesites can be estimated from the requirement that garnet be present in the residuum after melting. Garnet is not stable in the mantle at pressures of less than 20–30 kbar, corresponding to depths in the order of 65–100 km (Ito and Kennedy, 1967; Kushiro et al., 1972; Wyllie, 1971). The REE data thus indicate that the depth of mantle fusion was greater than the 65–100 km required to stabilize garnet peridotite.

Although relatively little of the research on the petrology of Pan-African volcanics has been reported, it is worthwhile comparing the conclusions of those investigations that have been completed. Late Proterozoic volcanic rocks from Algeria and Morocco are very similar to YMV andesites. In the Tassendjanet Area, Algeria, Chikhaoui et al. (1978, 1980) report andesites with similar major-element compositions to YMV andesites which are also enriched in Cr and Ni relative to modern andesites. These investigators argue that Tassendjanet andesites were generated by melting of hydrous peridotite. Modeling of andesites was not reported, but model results of calc-alkaline basalts indicated that these were generated by 10–20% equilibrium partial melting of garnet lherzolite. McCurry and Wright (1977) studied Late Precambrian andesites, dacites, and rhyolites from NW. Nigeria which have $(Ce/Yb)_N$ of 9–35. These investigators conclude that the andesites were generated at depths great enough to permit garnet fractionation. Thus all available data on Late Proterozoic calc-alkaline volcanic rocks support an origin from melting of garnet peridotite. This suggests that Pan-African calc-alkaline igneous activity was rooted at depths of greater than 60 km in the mantle.

CONCLUSIONS

The following are the most important conclusions of this study:

(1) The end of the Precambrian was a period of instability and melting in the mantle which lead to the formation of the entire crust of the central eastern desert of Egypt.

(2) The earliest stage was the generation of MORB-like tholeiitic basalts (the 'older metavolcanics') in a restricted back-arc basin or incipient Red Sea-like rift. The OMV represent about 20% melts of depleted upper mantle, at depths of 60 km or less.

(3) The next stage was generation of a calc-alkaline sequence of minor basalt, andesite and rhyodacite in an arc setting. The 'younger metavolcanics' represent lesser degrees (< 10%) of partial melting of a less-depleted mantle than the OMV, at depths greater than 65–100 km.

Thus, the volcanic rocks of the central eastern desert of Egypt record an episode of primitive crustal growth that is distinct from other terranes of similar age elsewhere in Africa which indicate that large-scale remobilization of older sial was common during the Late Precambrian.

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