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A late Precambrian (~ 710 Ma) high volcanicity rift in the southern Eastern Desert of Egypt

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With 9 figures and 3 tables

Zusammenfassung

Die spätpräkambrischen Shadli-Metavulkanite in der südöstlichen Eastern Desert von Ägypten sind eine schwach metamorphe bimodale Basalt-Rhyodazit-Abfolge, die bisher als Ausdruck eines Inselbogen-Vulkanismus gedeutet wurde. Zwei Basalt-Typen können aufgrund ihrer unterschiedlichen chemischen Zusammensetzung unterschieden werden: der stratigraphisch untere Typ ist ein N-MORB ähnlicher Ferrobasalt, während der überlagernde Typ Charakteristika eines leicht fraktionierten E-MORB aufweist. Die beiden Basaltvarietäten werden von stark an inkompatiblen Elementen verarmten Schmelzen aus den oberen 60–75 km des Mantels abgeleitet. Die Entstehung der sauren Metavulkanite ist nicht eindeutig geklärt; sowohl Fraktionierung aus einer mafischen Schmelze als auch Aufschmelzung juveniler Kruste vom Inselbogen-Typ sind denkbar. Die mafischen und felsischen Gesteine definieren zusammen ein Rb-Sr-Isochronenalter von 712 ± 24 Ma, das wir als den Zeitraum der Eruption deuten. Die Spurenelement-Verteilung der Shadli-Metavulkanite weist keine der charakteristischen Merkmale von Subduktionsmagmatismus auf, und wir sehen daher keinen direkten Zusammenhang mit einer Inselbogen-Entwicklung. Wir interpretieren den Shadli-Vulkanismus als Resultat eines Riftprozesses in junger kontinentaler Kruste, ähnlich dem Rio Grande-Rift oder dem Afar-Dreieck, wo starke Lithosphärendehnung die Förderung großer Lavamengen ermöglichte. Diese Interpretation stellt das einfache Schema einer panafrikanischen Krustenbildung durch Inselbogen-Addition im arabisch-nubischen Schild in Frage und erfordert eine Neubewertung bisheriger Modellvorstellungen.

Abstract

The late Precambrian Shadli Metavolcanics of SE Egypt constitute a slightly metamorphosed bimodal sequence that has been previously interpreted as manifesting volcanic activity at an island arc. We report the first Rb-Sr geochronologic, trace element (including REE), and Nd isotopic data for these rocks. Two types of basalt are recognized, the stratigraphically lower suite having compositions like N-MORB ferrobasalt while the overlying basalt is similar to slightly fractionated E-MORB. The two basalt types were derived from melting of a strongly depleted source, most likely within the upper 60–75 km of the upper mantle. The origin of the felsic melts is problematic, and these could either have fractionated from a mafic melt or resulted from melting of juvenile crust. The mafic and felsic lavas yield a Rb-Sr isochron age of 712 ± 24 Ma that probably represents the time of volcanic eruption. The trace element characteristics of both mafic and felsic members of the Shadli Metavolcanics show few of the hallmarks of subduction-related melts, and we reject the hypothesis that these formed at an island arc. Instead, the field and geochemical data are most consistent with the hypothesis that these rocks originated in a magmatic rift, where the eruption of large volumes of lava accompanied large-scale lithos-

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pheric extension. This inference suggests that the tectonic setting of the important 700–715 Ma crust-forming event in NE Africa and Arabia needs to be critically reexamined.

Résumé

Les roches volcaniques faiblement métamorphisées du Précambrien tardif de Shadli dans le Sud-Est de l'Egypte se caractérisent par une séquence bimodale qui était jusqu'ici interprétée comme résultant d'un volcanisme d'arc insulaire. Nous présentons ici les premières données isotopiques (Rb-Sr, Nd) obtenues pour ces roches ainsi que des données d'éléments en trace (y compris les Terres Rares). Deux types principaux de basaltes peuvent être distingués: les basaltes qui se trouvent dans la partie inférieure de la colonne stratigraphique, ont une composition comparable aux ferrobasaltes de type N-MORB, tandis que les roches se trouvant dans la partie supérieure de la séquence s'apparentent plutôt aux E-MORB légèrement fractionnés. Les deux types de basalte proviennent de la fusion partielle d'un manteau fortement appauvri, et ce à une profondeur probable de 60 à 75 km dans le manteau supérieur. L'origine des laves acides pose, quant à elle, quelques problèmes: elles peuvent s'être formées par cristallisation fractionnée à partir d'un liquide basique, ou alors elles sont le produit de la fusion partielle d'une croûte juvénile. La combinaison des données Rb-Sr obtenues pour les roches acides et basiques permet l'obtention d'une isochrone définissant un âge de 712 ± 24 Ma. Cet âge est interprété comme datant les éruptions volcaniques. Les données d'éléments en traces obtenues pour les laves basiques et acides ne présentent aucune des caractéristiques associées au volcanisme d'arc insulaire. Par conséquent, nous rejetons l'hypothèse selon laquelle les roches volcaniques de Shadli se seraient formées dans un contexte d'arc insulaire. Les données géochimiques et de terrain seraient plutôt en accord avec une hypothèse selon laquelle les roches volcaniques métamorphiques de Shadli se seraient formées dans une zone de rift où de très volumineux épanchements de laves auraient succédé à une extension à grande échelle de la lithosphère. Cette interprétation des données nous conduit à suggérer un réexamen critique du contexte tectonique entourant la période de formation crustale se situant entre 700 a 715 Ma, dans le nord-est de l'Afrique ainsi qu'en Arabie.

Краткое содержание

Позднекембрийские метавулканиты Shadli, находящиеся на юго-востоке восточной пустыни Египта, являются слабо метаморфизированными свитами бимодального базальт – риодацита, которые до сих пор считали образованиями вулканизма островной дуги. По химическому составу различают здесь два типа базальтов : подстилающий ферробазальт, напоминающий N – MORB и перекрывающий его базальт E-MORB, проявляющий слабое фракционирование. Оба типа базальтов, у которых отмечается истощение взаимоисключающих элементов, происходят из верхних 60–75 км мантии. Происхождение кислых метавулканитов еще полностью не выяснено; возможным считают или фракционирование из мафического расплава, или возможное расплавление ювенильной коры типа островных дуг. Изохронный возраст мафических и фельзических пород, определенный с помощью метода Rb/Sr, составляет 712 ± 24 Ma. Предполагают, что и эрупция имела место в это время. Распределение малых элементов в метавулканитах Shadli не несет характера субдукционного магматизма; поэтому авторы не видят здесь прямой взаимосвязи с развитием островной дуги, но рассматривают метавулканиты Shadli, как следствие образования рифта в поздней материковой коре, подобно рифту Рио Гранде, или треугольнику Афар, где сильное растяжение литосферы способствовало поднятию больших количеств магмы. Такая интерпретация ставит под сомнение простую схему формирования панафриканской коры в результате присоединения островных дуг к арабо-нубийскому щиту; возникает необходимость пересмотреть существующие модели и представления.

Introduction

The formation of the Arabian-Nubian Shield during the Pan-African Event (ca. 950–450 Ma; KRÖNER, 1984) offers an excellent opportunity to study the formation of continental crust. Determining the age of volcanic sequences and the tectonic environment in which these were erupted is crucial for reliable reconstructions of crustal evolutionary pathways. Along with this, interpreting whether these melts interacted with older crust or were mantle-derived is important for addressing the controversy of whether the Arabian-Nubian Shield is juvenile crust or involved reworking of substantial amounts of older continental crust. To resolve this question it has been very useful to determine the age, initial isotopic composition, and geochemistry of volcanic successions in the Shield.

This approach has been useful in past studies of the Arabian-Nubian Shield, where it has demonstrated the geochemical similarities between the metavolcanic rocks of the Arabian-Nubian basement and those of modern arcs and back-arc basins (e.g., BAKOR et al., 1976; STERN, 1981; ROOBOL et al., 1983; HEIKAL & AHMED, 1984). We follow this general philosophy here to examine one of the largest exposures of basement metavolcanic sequences in Egypt, the Shadli Metavolcanics of the South Eastern Desert. In this paper, we report major and trace element geochemistry, Rb/Sr geochronology, and Nd isotopic data for these lavas at Um Samiuki, and use these to interpret their petrogenesis and tectonic setting.

Geological Setting

The study area is in the South Eastern Desert (SED), one of the three basements provinces defined by STERN & HEDGE (1985). The SED contains a diverse assemblage of lithologies, including abundant gneiss, granodiorite, metasediments and serpentinites in addition to the metavolcanic rocks being discussed here (Fig. 1). Reliable radiometric ages range from about 800 Ma to 580 Ma (HASHAD et al., 1972; DIXON, 1981; STERN & HEDGE, 1985); STERN & HEDGE (1985) concluded that the principal episodes of crustal growth in the SED occurred at 715–700 Ma and 665–685 Ma. The SED is bounded to the north by the Central Eastern Desert of Egypt (CED) which contrasts with the SED in being dominated by dismembered ophiolites, primitive arc assemblages, and associated immature sediments, all showing intense deformation related to arc accretion or Najd deformation (ELBAYOUMI & GREILING, 1984; SULTAN et al., 1988). The South Eastern Desert is also known as the Gerf Terrane (KRÖNER et al., 1987), and, with the Alaqi-Heiani and Onib-Sol Hamed sutures to its south, is unique among the Egyptian basement provinces in being bounded by well-defined suture zones.

The Shadli Metavolcanics are preserved in a 80 x 25 km belt, elongated WNW-ESE, and flanked north and south by granodiorites (Fig. 2). To a first approximation, this belt defines a broad synform and is cut by WNW-ESE-trending thrusts and strike-slip faults. The thickness of the Shadli Metavolcanics is not well known because of structural complications caused by faulting and folding; SHUKRI & MANSOUR (1980) state that it is more than 10 km thick. Previous studies on the geology, petrography, and major element geochemistry are summarized in SHUKRI & MANSOUR (1980), HAFEZ & SHALABY (1983) and KHUDEIR et al. (in press); the following summary is based on the well-studied area around the Um Samiuki Copper Prospect, from HAFEZ & SHALABY (1983) and SEARLE et al. (1978).

The Shadli Metavolcanics are a massive pile of cyclic, mafic to felsic, mainly submarine lavas, pyroclastics, and associated immature sediments. The sequence is subdivided into two groups, the older »Wadi Um Samiuki Volcanics« and the younger »Hamamid Group«. The Wadi Um Samiuki Volcanics are intruded by the granodiorites to the north, and are more than 1000 m thick. This group reflects repetitive eruption sequences of mafic and felsic volcanics and contains about 15% felsic material. Rocks of the Hamamid Group are the focus of this study. Because of massive sulphide deposits within the Hamamid Group, it is much better known than is the Wadi Um Samiuki Volcanics. The Hamamid Group attains a maximum thickness of 2000 m and consists of two distinct mafic-to-felsic cycles, both of which began with the eruption of mafic pillow lavas. Cycle I begins with the eruption of the »Lower Pillow Lavas«, consisting of aphyric basalt or olivine basalt. These are conformably overlain intermediate and felsic lavas and tuffs of Cycle I. The »Upper Pillow Lavas« of Cycle II conformably succeed the Cycle I felsic rocks, and are up to 100 m thick; these, in turn, are conformably overlain by Cycle II felsic rocks. The Upper Pillow Lavas consist of sparsely phyric plagioclase basalt. The entire sequence was intruded by dolerite sills following mineralization, folding, and thrusting and so these are of no interest here. In summary, the Shadli Metavolcanics is a very thick pile of largely submarine bimodal volcanics, with several mafic-to-felsic cycles preserved.

The samples studied here were collected by two of us (A. K. and A. A. R.) during field work in 1983, and include samples of the Cycle I or Lower Pillow Lavas, Cycle II or Upper Pillow Lavas, and Cycle I felsic rocks (Fig. 3). The samples are only slightly altered, leading SEARLE et al. (1978) to argue that the rocks of this area are not metamorphic rocks. We agree that these generally are remarkably fresh rocks, with beautiful preservation of original textures and fabrics, and with a general absence of veining or other indications of significant remobilization. However, the rocks are slightly altered, with the development of incipient greenschist-facies mineral assemblages. We emphasize that this is a reconnaissance study, but with results that are sufficiently surprising and important that further work is called for. In the following sections, we first present the geochemical data, followed by the Rb-Sr geochronologic results and Nd-isotopic

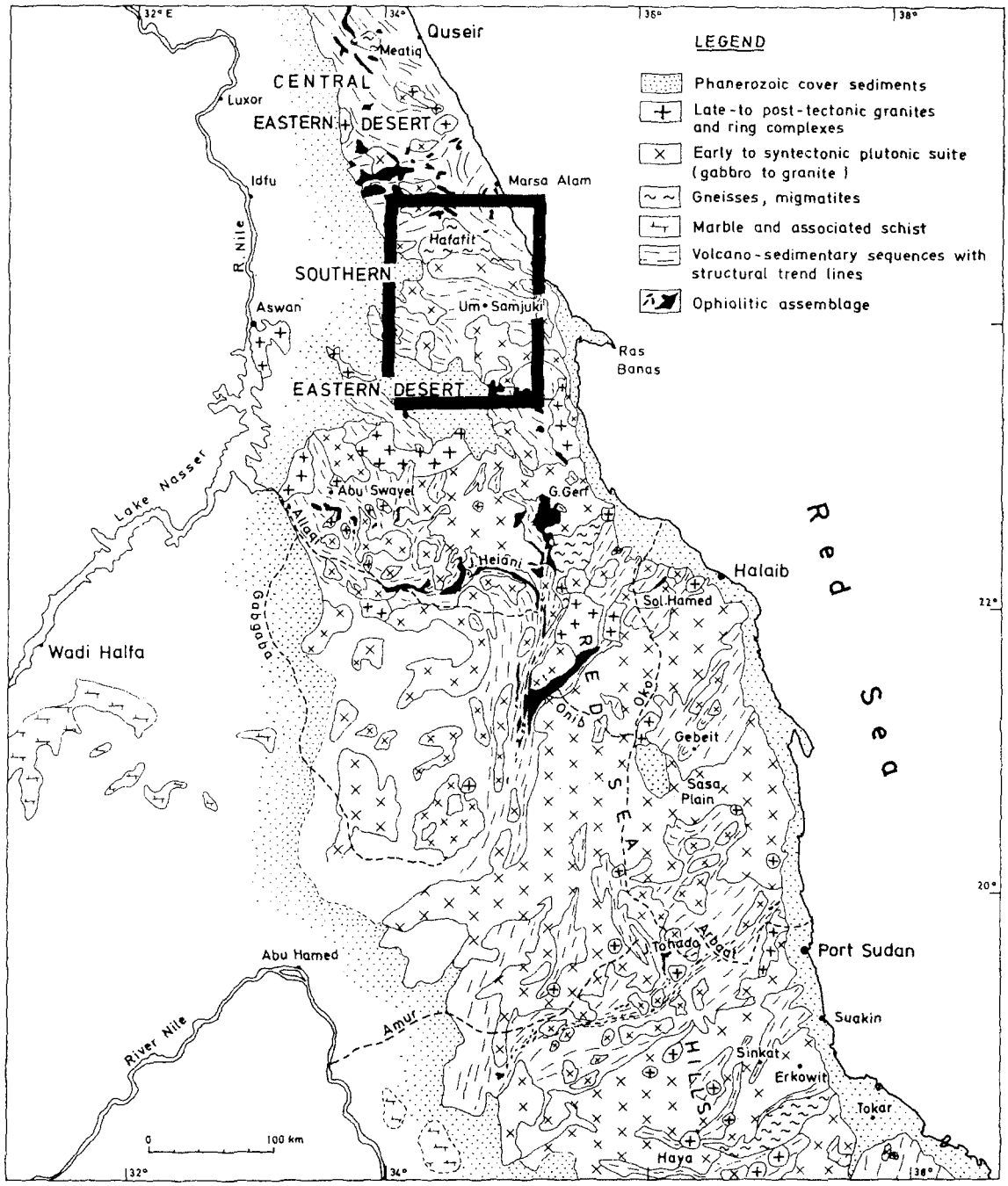

Fig. 1. Generalized geology of SE Egypt and NE Sudan, after KRÖNER et al. (1987): The rectangle shows the location of Fig. 2.

measurements. We then use this data to discuss the petrogenesis of the igneous rocks and infer their tectonic setting. Finally, we will discuss the significance of this work for our understanding of crustal evolution in the Arabian-Nubian Shield.

Analytical Techniques

Major element compositions were determined using conventional wet methods at Mainz University. Trace elements exclusive of Ba and the Rare Earth Elements (REE) were analyzed on the Siemens SRS 200 X-ray fluorescence spectrometer, using methods as detailed in WILLIS et al. (1972). Analyses of Ba and REE for six samples (EG-63, 64, 66, 68, 70, and 72) were done by isotope dilution techniques at UTD using the 12″-radius mass spectrometer, following methods outlined by LIN et al. (1989). Analyses of REE for two other samples (EG-65 and 71) were done by high pressure liquid

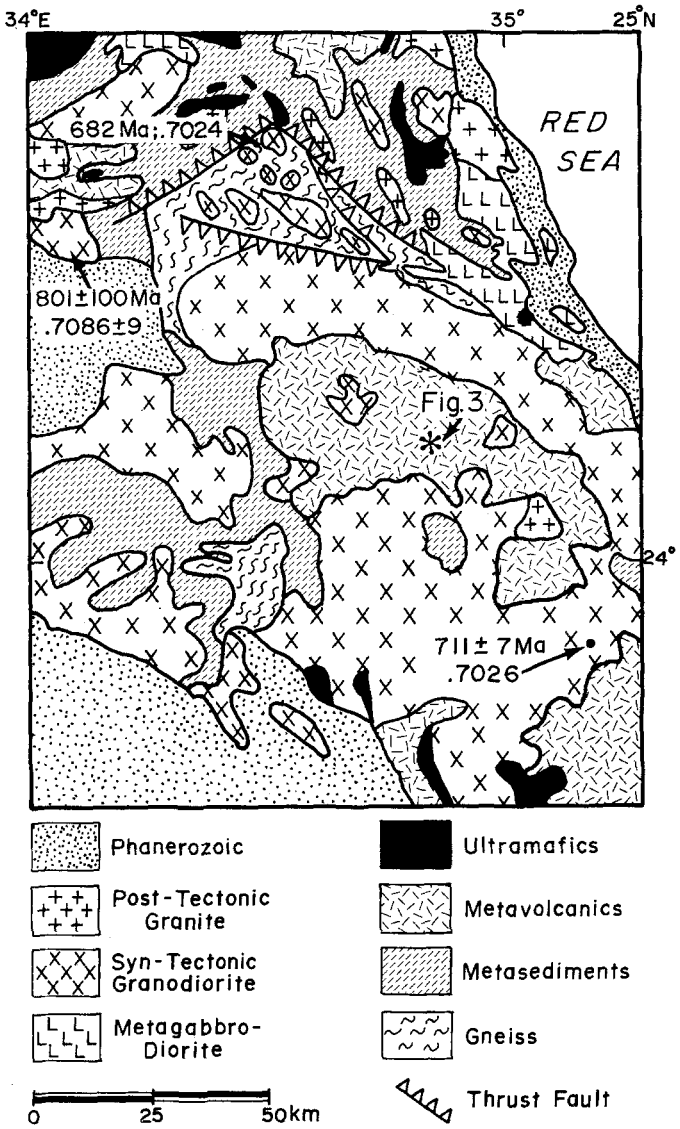

Fig. 2. Geology of the Shadli Metavolcanics belt, simplified from the map of EL RAMLY (1972). Thrust faults near the boundary of the CED and SED are from GREILING et al. (1984). Location of basement units with age determinations and initial $^{87}Sr/^{86}Sr$ are shown in boxes (Data sources, from N to S: STERN & HEDGE, 1985; HASHAD et al., 1972; DIXON, 1981). The location of the detailed study area is also shown.

chromatography at the Max-Planck-Institut für Chemie at Mainz, following the technique of CASSIDY (1988).

For the purposes of geochronology, Rb/Sr ratios were determined by XRF at Mainz following the peak-intensity method of PANKHURST & O'NIONS (1973) and as outlined by KRÖNER (1982); reproducibility is 1.5%. $^{87}Sr/^{86}Sr$ was determined for all samples at UTD using the 12″ radius solid source mass spectrometer. The total processing blank for Sr at UTD is about 3 ng. The isochron age was calculated using the York II model (YORK, 1969) and a ^{87}Rb decay constant of 1.42 x 10^{-11} y^{-1} (STEIGER & JÄGER, 1977). Uncertainties on the age and initial $^{87}Sr/^{86}Sr$ are quoted at the 2 sigma level. Nd isotopic determinations were performed on a MAT 261 mass spectrometer at UTD, following chemical procedures modified after those of RICHARD et al. (1976); total processing blanks for Nd are about 0.5 ng. $^{143}Nd/^{144}Nd$ are fractionation corrected to $^{146}Nd/^{144}Nd$ = 0.7219.

Geochemistry

The results of our chemical analyses for 11 samples are listed in Table 1. These include 6 basalts (SiO_2 = 48 to 53%), 1 andesite (60% SiO_2), and 4 rhyodacites (SiO_2 = 69–75%). The petrography and major element chemistry of the Shadli Metavolcanics have been reviewed by HAFEZ & SHALABY (1983) and KHUDEIR et al. (in press), and our results generally support their conclusions that the mafic rocks are tholeiites, while the felsic rocks have low contents of potassium. It is noteworthy that all of the mafic and intermediate samples analyzed by us contain less than 16.1% Al_2O_3, and we could not confirm the presence of high-alumina basalt types described by KHUDEIR et al. (in press). Our data are consistent with the interpretation of the above authors that the Shadli Metavolcanics constitute a bimodal suite, with modes at about 50% and 72% SiO_2 (Fig. 4).

We have analyzed two distinctly different types of basalt in the Hamamid Group. Samples from the Lower Pillow Lava (EG-71, 72) we have termed »Basalt I«. Basalt I contains high TiO_2 (2.28%), FeO* (14.4–14.8%), and P_2O_5 (0.36–0.46%); it has low SiO_2 (48–50%), Al_2O_3 (13.6–14.3%), MgO (3.8–4.4%), K_2O (0.25–0.29%), and Mg # (100 Mg/Mg + Fe; 32–35). »Basalt II« comes from drill core sampling the predominantly felsic upper part of Cycle I and the Upper Pillow Lavas (EG-63, 65, 66, and 69) and is characterized by low TiO_2 (0.96–1.12%), FeO* (8.0–8.7%), and P_2O_5 (0.16–0.18%), and moderate SiO_2 (51.2–52.4%), K_2O (0.72–1.17%), Al_2O_3 (15.5–16.1%), MgO (6.8–7.9%), and Mg # (59–64). Basalt I and II can also be identified in the major element data of earlier studies (HAFEZ & SHALABY, 1983; KHUDEIR et al., in press). For example, in the data of HAFEZ & SHALABY (1983), S5 and S7 are very similar to Basalt I, while S3 is similar to Basalt II; in the data of KHUDEIR et al. (in press), 92A and 11E are Basalt I and 22L and 85C are Basalt II.

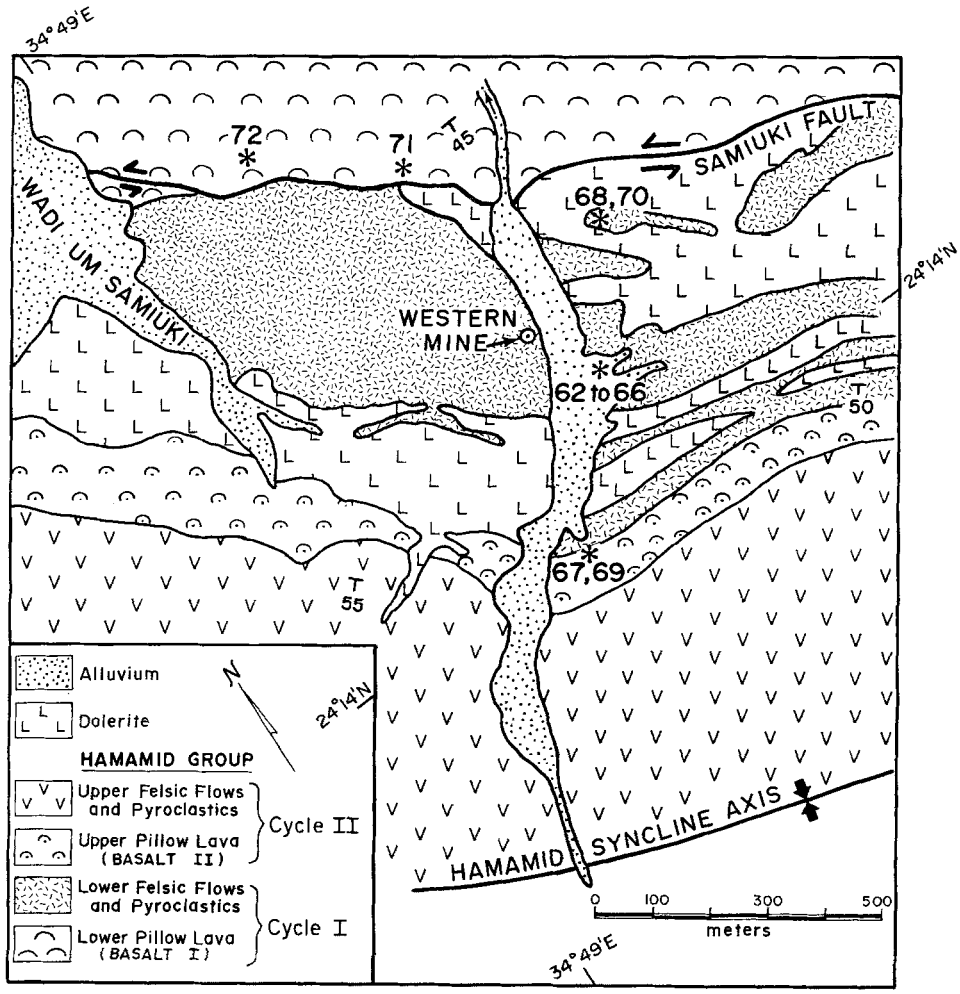

Fig. 3. Generalized geology of the region around the western mine, near the headwaters of Wadi Um Samiuki, modified from SEARLE et al. (1978). Location of samples studied for this report are shown. Samples 62 through 66 were collected from core drilled at the location shown.

Basalt I and Basalt II groups can also be seperated on the basis of trace element characteristics (Table 1). Compared to Basalt II, Basalt I has lower Rb, Ba, Nb, Cr, and Ni, and is strongly depleted in the light REE (Fig. 5, 6). Basalt II has much higher Rb, Nb, Cr, and Ni, and lower Y and V, and has a slight LREE-enrichment. Diagnostic trace element ratios such as Zr/Nb are lower for Basalt II (10–17) than for Basalt I (24–32), as is K/Rb (635–840 vs. 1660–3250). Sr/Nd in Basalt I (9–17) may be slightly lower than Basalt II (14–23). K/Ba is higher in Basalt I (89) relative to that in Basalt II (54–66), while Ba/La in Basalt I (6.8) is lower than that in Basalt II (13–19). The andesite sample (EG-67) shows chemical affinities to Basalt I, to which it might be related by fractional crystallization.

The felsic rocks have been identified as predominantly composed of low-K rhyolites and dacites (KHUDEIR et al., in press), an interpretation

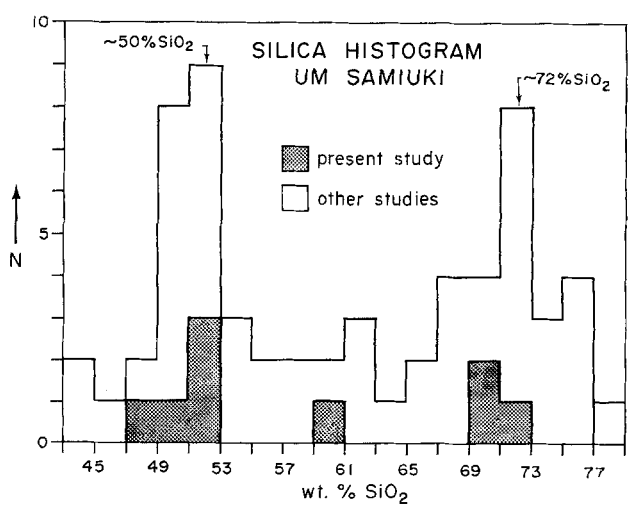

Fig. 4. Histogram of silica contents for rocks from the Shadli Metavolcanics. Note that the data define a bimodal assemblage, with peaks at about 50 % and 72 % SiO_2. Blackened area correspond to samples from the present study. Areas left blank correspond to data from SHUKRI & MANSOUR (1980), HAFEZ & SHALABY (1983), and KHUDEIR et al. (in press).

Sample	EG 62	EG 63	EG 64	EG 65	EG 66	EG 67	EG 68	EG 69	EG 70	EG 71	EG 72
Type	RD	BII	RD	BII	BII	A	RD	BII	RD	BI	BI
Major Elements(%)											
SiO_2	70.30	52.36	69.43	51.92	52.79	60.06	75.31	51.21	72.19	50.18	47.98
TiO_2	0.50	0.96	0.48	1.12	1.09	0.84	0.26	1.10	0.39	2.28	2.28
Al_2O_3	12.85	15.60	12.88	16.08	15.51	14.03	12.06	15.74	12.43	13.59	14.30
FeO	2.04	5.79	2.25	5.01	5.34	5.50	1.63	5.57	2.08	7.52	9.22
Fe_2O_3	3.50	2.47	3.30	3.93	3.19	3.56	1.03	3.44	2.27	7.69	6.16
MnO	0.17	0.18	0.17	0.17	0.17	0.20	0.05	0.13	0.14	0.28	0.25
MgO	1.01	7.85	1.07	6.85	6.83	2.76	0.54	6.96	0.64	3.80	4.41
CaO	1.94	7.13	2.00	8.10	7.33	7.51	0.93	7.92	1.89	7.47	9.25
Na_2O	5.20	3.76	5.22	3.44	3.67	1.91	4.95	2.73	5.17	3.90	2.77
K_2O	1.10	0.96	1.12	0.93	0.72	0.84	1.88	1.17	1.10	0.25	0.29
P_2O_5	0.12	0.16	0.13	0.16	0.18	0.13	0.04	0.16	0.08	0.46	0.36
LOI	1.12	3.05	1.26	2.66	3.11	2.78	0.71	3.05	1.15	1.76	1.86
CO_2	0.14	0.10	0.30	0.10	0.24	0.08	0.06	0.02	0.30	0.10	<0.01
Total	99.9	100.4	99.7	100.4	100.0	100.4	99.7	99.8	99.8	100.0	100.1
Trace Elements(ppm)											
Rb	9.4	12.3	9.5	12.2	8.7	8.7	24.6	13.8	9.3	2.1	2.1
Sr	118	189	119	244	197	187	54	208	102	140	198
Nb	6.8	7.6	7.2	6.4	13	2.8	4.3	8.4	10	4.2	4.1
Zr	44	99	424	106	131	73	135	110	486	133	96
Y	74	32	73	33	36	32	58	33	81	52	40
V	3	167	3	213	194	206	20.4	207	3	344	505
Co	2.4	43	2.0	41	37	21	5.3	41	3.0	36.4	41
Cr	2.6	242	2.6	178	190	23	7	190	2.7	2.5	3
Ni	6	140	3.5	108	109	12	6.0	109	2.4	10	7.6
Cu	---	---		45	34	39	2.3	41		37	30
Zn	---	---	129	85	97	107	24	56	112	171	135
Ba	---	120	153	---	111	---	277	---	138	---	26.9
La	---	6.17	12.2	5.37	8.43	---	6.59	---	13.2	5.10	3.93
Ce	---	16.3	32.7	14.2	20.5	---	16.6	---	35.1	16.2	12.1
Nd	---	11.4	24.9	10.5	14.0	---	12.9	---	26.4	15.5	11.6
Sm	---	3.02	7.91	3.01	4.04	---	4.38	---	8.23	5.17	3.98
Eu	---	1.05	2.91	0.92	1.32	---	0.67	---	2.92	1.94	1.85
Gd	---	4.14	9.07	3.67	4.55	---	6.06	---	10.0	6.94	5.34
Dy	---	4.64	10.2	4.46	5.33	---	8.33	---	11.8	8.00	6.11
Er	---	2.95	6.73	2.53	3.66	---	5.89	---	7.94	4.66	3.85
Yb	---	2.82	6.32	2.40	3.42	---	6.04	---	7.79	4.27	3.26
Trace Element Ratios											
K/Rb	970	650	980	630	690	800	630	700	980	1660	3250
Sr/Nd	---	16.6	4.8	23.2	14.1	---	4.2	---	3.9	9.0	17.1
K/Ba	---	66	61	---	54	---	56	---	66	---	89
Ba/La	---	19.4	12.6	---	13.2	---	42.0	---	10.5	---	6.8
$(La/Yb)_n$	---	1.46	1.29	1.50	1.65	---	0.73	---	1.13	0.80	0.81
$(Eu/Eu)^*$	---	0.91	1.05	0.85	0.95	---	0.40	---	0.99	1.00	1.23
La/Nb	---	0.81	1.69	0.84	0.65	---	1.53	---	1.32	1.21	0.96

BI = Basalt I; BII = Basalt II; A = Andesite; R-D = Rhyodacite

Table 1. Major and Trace Element Data

with which we agree. The felsic rocks in our collection are not extremely fractionated; they still contain appreciable TiO_2 (0.26–0.50%), CaO (0.93–2.00%), and MgO (0.54–1.07%). For rocks with 69–75% SiO_2, they contain remarkably little K_2O (1.10–1.88%) and Rb (9.3–24.6 ppm); only EG-68 contains more K and Rb than does Basalt II. Correspondingly, K/Rb for the felsic rocks (634–980) is generally higher than for Basalt II. The REE patterns for the rhyodacites are very nearly flat; two of these (EG-64, 70) show no Eu anomaly (Fig. 6).

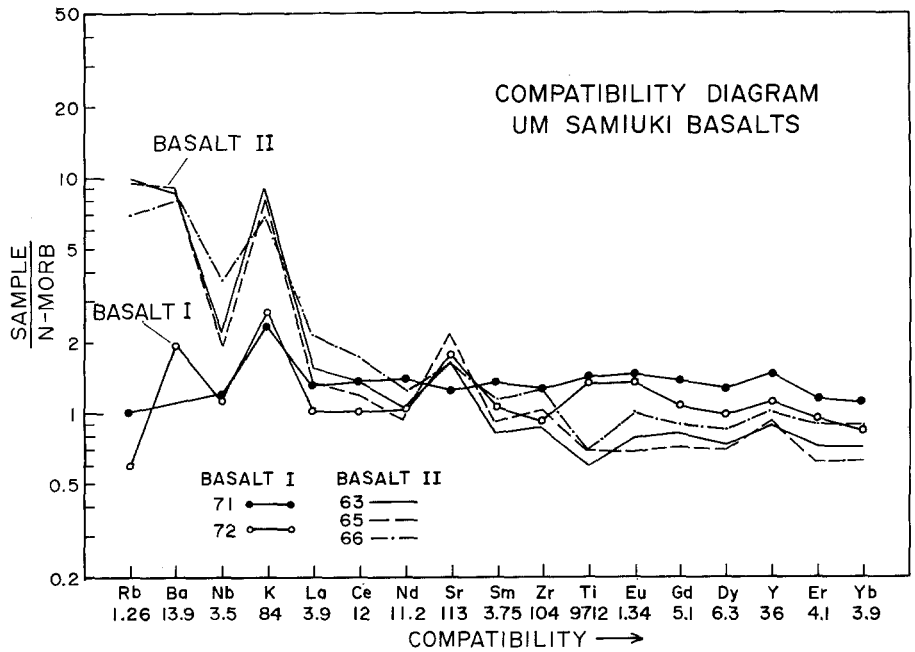

Fig. 5. Compatibility diagram, Um Samiuki basalts. The vertical scale corresponds to the abundance of an element compared to that in N-MORB, while the horizontal scale lists the elements in order of increasing compatibility in mantle phases to the right. Concentrations used for the N-MORB normalization are listed beneath each element (HOFMANN, 1988).

Geochronology

Six samples were selected for Rb-Sr dating. These samples encompass the full range of lithologies analyzed. Analytical results are listed in Table 2; a reasonably good spread in $^{87}Rb/^{86}Sr$ was obtained (0.02 to 1.35), although the age is largely controlled by one sample of rhyodacite (EG-68). The data define an excellent isochron (MSWD = 0.3; Fig. 7), with an age of 712 ± 24 Ma and an initial ratio of 0.7022 ± 1. This age is interpreted to mark the time of Shadli lava eruption, and, as previously discussed, corresponds to a major episode of crust formation in the SED. The initial ratio is also very

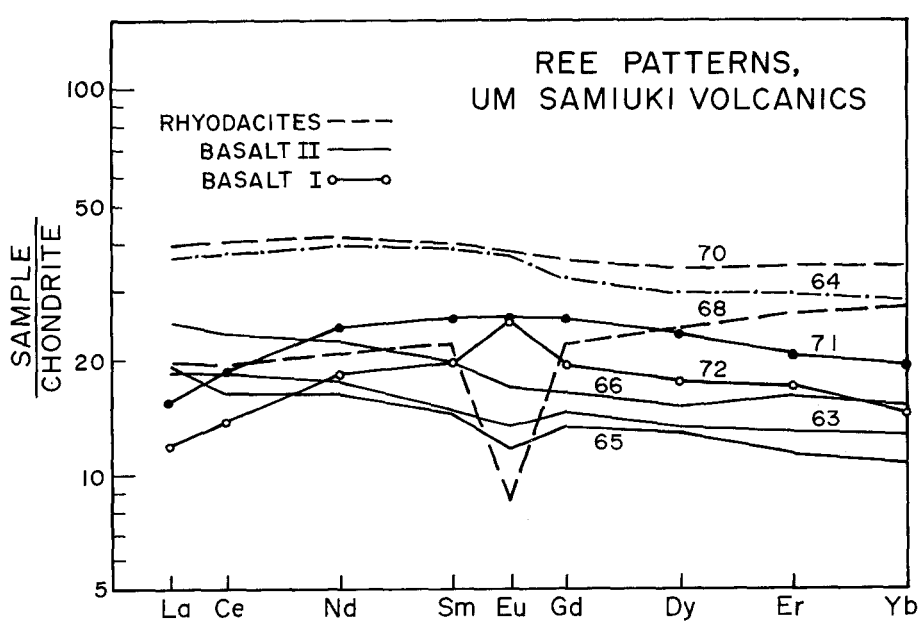

Fig. 6. REE patterns for Um Samiuki Volcanics, normalized to chondritic meteorites (REE from NAKAMURA, 1974; Ba from WOOD et al., 1979). Note that Basalt I is strongly depleted in LREE and has a positive or no europium anomaly, while Basalt II is slightly LREE-enriched and has a negative europium anomaly.

similar to that of 700–715 Ma plutonic rocks from the SED (0.7023–0.7026: DIXON, 1981; STERN & HEDGE, 1985). The low initial $^{87}Sr/^{86}Sr$ indicates that these melts were either derived from depleted mantle or from remelting of very young, low Rb/Sr crust that was derived from melting of depleted mantle. We tried to separate and date zircons from the acid samples of our suite, but the zircon yield was extremely poor. A few grains between 30 and 50 microns in size were evaporated using the technique of KOBER (1987) as detailed in KRÖNER & TODT (1988). No stable signal could be obtained after evaporation of common Pb, but the $^{207}Pb/^{206}Pb$ ratios obtained are consistent with the above Rb-Sr age.

Neodymium Isotopic Composition

Neodymium isotopic data for three samples of basalt are listed in Table 3. Epsilon-Nd at 710 Ma ranges from +6.3 to +7.8. These data are similar to ca. 740–780 Ma ophiolitic rocks from Saudi Arabia, which give Epsilon-Nd(T) of +6.6 to +7.6 (CLAESSON et al., 1984). The depleted mantle model of NELSON & DEPAOLO (1985) predicts that the depleted mantle had an Epsilon-Nd of +6.9 at 710 Ma, and the data for the Shadli basalts are very close to this value.

Petrogenesis

The chemical and isotopic data presented here allow us to make general observations regarding the orgin of the magmas for the Hamamid Group. We have already noted that the low initial $^{87}Sr/^{86}Sr$ (0.7022 ± 1) and Epsilon Nd (+6.3 to +7.8) fall in the range for depleted mantle about 710 Ma ago. This indicates that at least the mafic lavas were derived by melting of the mantle, with no discernible contribution from much older continental crust. We now turn to a more detailed examination of the basalt melt petrogenesis, followed by consideration of the felsic melt petrogenesis.

Because the variations in chemistry between Basalts I and II are systematic and involve immobile

	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
EG–63	0.184	0.70403±7
EG–64	0.222	0.70453±11
EG–66	0.124	0.70348±11
EG–68	1.35	0.71590±9
EG–70	0.288	0.70519±6
EG–72	0.0168	0.70240±8

$^{86}Sr/^{88}Sr$ normalized to 0.1194.
$^{87}Sr/^{86}Sr$ adjusted to E + A $SrCO_3$ = 0.70800

Table 2. Rb-Sr Geochronological Data

	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	$\epsilon Nd^{(710\ Ma)}$
EG–63	0.1602	0.51285±2	+7.8
EG–66	0.1745	0.51284±2	+6.3
EG–72	0.2074	0.51306±2	+7.6

$^{143}Nd/^{144}Nd$ for UCSD Nd = .511848 and for BCR–1 = 0.512611.

Table 3. Sm-Nd Isotopic Data

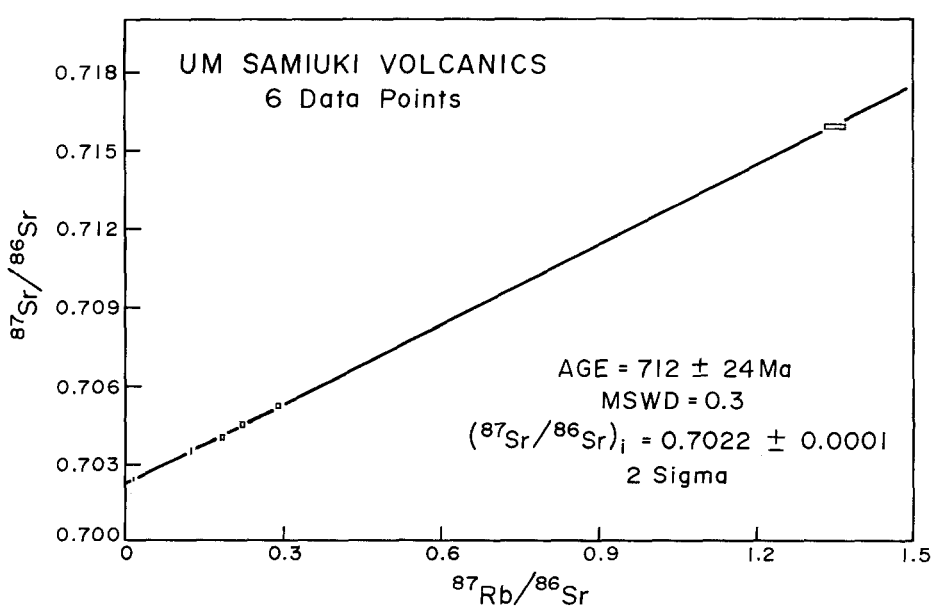

Fig. 7. Rb-Sr isochron diagram for Hamamid Group volcanics, Um Samiuki area.

(e.g., Ti, Zr, REE, Cr, P) as well as mobile (K, Rb) elements, we believe that the observed differences did not result from alteration but are original magmatic features. The relatively high MgO, Mg #, Cr, and Ni for Basalt II readily allows for these to have been generated by partial melting of the upper mantle, followed by limited fractionation of olivine and pyroxene (compare Mg # of 59–64 for Basalt II with Mg # of >70 expected for basalt in equilibrium with mantle peridotite; BVSP, 1981). All samples of Basalt II have negative europium anomalies, indicating that either plagioclase remained in the mantle after melting, or fractionated from the evolving melt; the petrographic observation that the melts were saturated in plagioclase leads us to prefer the latter interpretation. The elemental enrichment trends observed for Basalt II on the compatibility diagram (Fig. 5) are most comparable with slightly enriched abyssal tholeiites, or »E-MORB«. This inference is further supported by the moderately high K/Rb (635–843) and low Zr/Nb (10–16), Sr/Nd (14–23), and Ba/La (13–19) of Basalt II, all features characteristic of an E-MORB mantle source. For example, E-MORB has Ba/La = 11 to 20 (WILSON, 1989; GILL, 1981), Zr/Nb = 6–15 (WILSON, 1989; BSVP, 1981), and Sr/Nd of 10–30 (DEPAOLO & JOHNSON, 1979; WILSON, 1989). We note that Basalt II samples show a depletion in Nb, a characteristic of convergent margin magmas (GILL, 1981; WILSON, 1989).

The petrogenesis of Basalt I is more enigmatic. This has low MgO, Mg #, Cr, and Ni, indicating extensive fractionation of olivine and pyroxene. This fractionation was accompanied by strong enrichments in Fe, Ti, and V. The marked enrichment in ferromanganese elements indicates that fractionation was anhydrous. In spite of this, the primary melt from which Basalt I evolved must have been generated by melting of extremely depleted mantle, indistinguishable from that giving rise to modern N-MORB. The marked LREE-depletions $(La/Yb)_n$ = 0.8), very high K/Rb (1660–3250), and K/Ba (89) are all hallmarks of N-MORB. Basalt I could not have fractionated from a magma like Basalt II, because the contents of SiO_2 and incompatible elements K, Rb, and LREE are significantly lower than in Basalt II. Conversely, the difference in REE patterns and K/Rb leads us to conclude that fractionation of a Basalt I precurser could not have evolved to Basalt II, especially since the fractionation experienced by Basalt II has been relatively limited. We conclude that Basalt I and II reflect distinct magmatypes, both derived from melting of the mantle. The earlier phase, Basalt I, tapped a strongly depleted mantle (N-MORB type) and suffered significant anhydrous fractionation controlled by olivine and pyroxene. The later phase, Basalt II, tapped a distinctly less depleted mantle (E-MORB type) and suffered less fractionation of olivine and pyroxene, but substantial fractionation of plagioclase. The fact that the heavy REE are not depleted means that either melting occured at pressure below the stability field of garnet peridotite (i.e., shallower than about 60–75 km; DANCKWERTH & NEWTON, 1978), or that garnet was consumed.

The petrogenesis of the felsic rocks is also problematic. These may represent extreme fractionates of the mafic melts. If so, they could not have fractionated from a Basalt II parent, because much higher incompatible element contents (K, Rb, Nb) would be expected. Also, the REE patterns of the felsic lavas are slightly LREE-depleted, and two of the samples have no Eu anomaly (EG-64, 70). Basalt II is slightly LREE-enriched and characterized by negative Eu anomalies, and any deriviate would be expected to inherit and increase these features as a result of fractionation. Finally, the ratio K/Rb should decrease during fractionation (SHAW, 1968), but this is generally higher in the felsic rocks (630–980) than in Basalt II (630–700).

The felsic melts could be fractionates of Basalt I, as shown by the fact that the felsic rocks have low incompatible element concentrations, high K/Rb, and slight to moderate depletion in the LREE (Fig. 6). This is an attractive hypothesis because the Basalt I and the felsic rocks that we analyzed comprise a single magmatic cycle. The role of plagioclase fractionation in the evolution of the felsic rocks is shown by their low Sr/Nd (3.4–4.1), and by the strongly negative Eu anomaly in one sample (EG-68). A lack of zircon fractionation is indicated by high Zr (up to 486 ppm Zr) and Zr/Nb (up to 65); only the most evolved sample (EG-68) shows lower Zr (135 ppm) and Zr/Nb (31). This general lack of zircon fractionation in the felsic suite may have been responsible for our failure to isolate sufficient zircons for U/Pb dating. Alternatively, the felsic melts may represent anatectic melts of juvenile crust. At present, there are too few constraints to allow us to assess such a model.

Tectonic Setting

We are especially interested in resolving the tectonic setting in which the Shadli Metavolcanics were erupted. This is important not only for understanding

the geological history of the study area itself, but also for understanding the nature of the major 700–715 Ma crust-forming event in the SED. We appreciate that our data comes from a very small part of the Shadli Metavolcanics, but believe that, because the Shadli Metavolcanics is a repetitive pile of mafic-to-felsic volcanic cycles, inferences drawn from one part of the section may be applicable throughout. With this warning, we turn to discussing the tectonic setting of the Shadli Metavolcanics and use as a principal tool our new chemical data.

Previous workers have identified the tectonic setting for the Shadli Metavolcanics as that of an immature arc (HAFEZ & SHALABY, 1983; KHUDEIR et al., in press). This identification has been based on consideration of major element data alone, which show a general correspondence with modern convergent margin magmatic suites. The fact that the Shadli Metavolcanics comprise a bimodal suite causes difficulty for this interpretation, because such suites are more commonly interpreted as related to rifting (MARTIN & PIWINSKII, 1984). In fact, the recognition of the bimodal nature of the Kolet Umm Kharit metavolcanics, 50 km to the west of the present study area – and to which the Shadli Metavolcanics may be related – led EL RAMLY et al. (1982) to conclude that these were not erupted at a convergent margin.

Certain immobile minor and trace elements and their ratios can help to resolve tectonic environments (e.g., TiO_2, Cr, Zr, Y), especially for basalts. Discrimination diagrams for mafic rocks are shown in Fig. 8, along with fields for modern arc, MORB, and »within-plate« lavas. The data for Basalt II plot in the field of MORB on all three diagrams, while the data for Basalt I plot in the MORB field on two diagrams (Fig. 8a and c). The very low Cr content of the Basalt I samples causes them to plot in the arc field in Fig. 8b, demonstrating the weakness of these diagrams when very fractionated samples are involved. With this one exception, all of the mafic samples plot outside the field defined by modern convergent margin volcanic suites.

The tectonic setting of the felsic rocks can also be examined using discriminant diagrams (Fig. 9). These diagrams were designed for use with intrusive felsic rocks, but we see no reason why they cannot be applied to felsic volcanic rocks as well. All four felsic samples of the Shadli Metavolcanics plot in the field of Ocean Ridge Granites (ORG) on the Nb vs. Y diagram (Fig. 9a). On the Rb vs. Y+Nb diagram (Fig. 9b), three samples plot in the ORG field, while the fourth lies in the field for Within Plate Granites (WPG). It is especially noteworthy that all of the

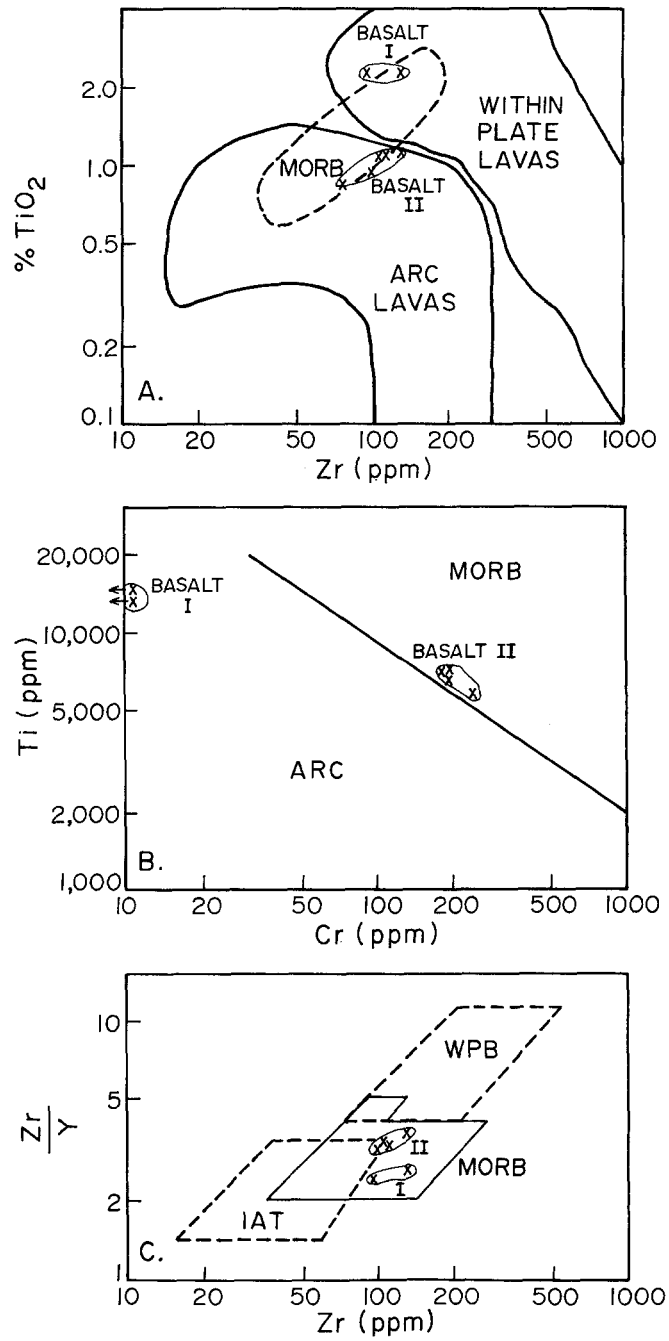

Fig. 8. Trace element discrimination diagrams for mafic rocks: A: TiO_2 vs. Zr (PEARCE, 1980), with data plotted for Basalts I and II. B: Ti vs. Cr (PEARCE, 1975), with data plotted for Basalts I and II. C: Zr/Y vs. Zr (PEARCE, 1980) showing fields for island arc tholeiite (IAT), MORB, and within-plate basalt (WPB), along with data for Basalt I (I) and Basalt II (II).

felsic samples plot outside of the field for volcanic arc granites (VAG).

The fact that both felsic and mafic portions of the Hamamid Group plot outside the field for arc rocks on various discrimination diagrams, and the fact that the suite is compositionally bimodal leads us to reject

the hypothesis that Hamamid volcanism occurred at a convergent plate boundary, either an intra-oceanic arc or Andean-type margin. Furthermore, and in spite of the fact that the various discriminant diagrams suggest an origin at a mid-ocean ridge spreading center, we reject this possibility as well, for the following reasons: (1) the Shadli Metavolcanics are associated with a much larger proportion of felsic volcanics and volcaniclastic sediments than ever found at modern spreading centers; (2) the Shadli Metavolcanics are preserved in a large WNW-ESE trending synclinorium that is autochthonous (i.e., not obducted); and (3) there is no evidence for gabbroic and ultramafic complements that might make the Shadli Metavolcanics part of an ophiolite section.

There remain two possible tectonic settings that are consistent with the field and geochemical data: (1) back-arc basin; and (2) magmatic rift. Both tectonic settings are extensional, but they differ because the first is closely associated with a subduction zone while the second is not. We are at present unable to unequivocally choose between these alternatives, but we find several aspects that do not favor the first hypothesis. First, ancient back-arc basins are generally identified as part of an ophiolite suite, although autochthonous back-arc basins are also known (e.g., Sarmiento Complex; TARNEY et al., 1976). Second, while felsic volcanics make up a small proportion of the back-arc basin volcanic succession (LONSDALE & HAWKINS, 1985), we know of no intra-oceanic back-arc basins composed of the 33% felsic volcanics reported for the Shadli (SHUKRI & MANSOUR, 1980). Third, back-arc basin basalts typically manifest a substantial arc signature, especially some depletion in HFSC (SAUNDERS & TARNEY, 1984). While some depletion Nb is observed in Basalt II, no depletion is seen for Zr. SAUNDERS et al. (1980) argue that, in many instances, La/Nb discriminates between MORB (La/Nb generally <1.4) and back-arc basin basalts (La/Nb generally >1); the Shadli basalts have La/Nb in the range 0.65 to 1.2 (Table 1). Third, the association of pillowed lavas and pyroclastic beds interpreted as subaerial eruptions (HAFEZ & SHALABY, 1983) indicates that volcanism occurred close to sea level and not at the 2–3,000 m water depth found in the modern back-arc basins of the western Pacific. Finally, if the Shadli Metavolcanics represent a back-arc basin, where is the arc associated with it? This has not been identified, and there is no obvious E-W trending arc-like magmatic axis that we know either to the north or to the south of the Shadli Metavolcanics. In summary, while the possibility that the Shadli Metavolcanics represent back-arc basin magmatism cannot be precluded, we do not favor this hypothesis.

We hypothesize that the Shadli Metavolcanics were erupted in a magmatic rift not associated with a convergent margin. Such a rift might find a appropriate modern analogue in the Rio Grande Rift or the Afar Triangle. At these locations, strong extension of continental crust is associated with voluminous bimodal volcanism involving the eruption of MORB-like basalts (some with strong Fe-enrichment) and low-K felsic lavas, all occurring

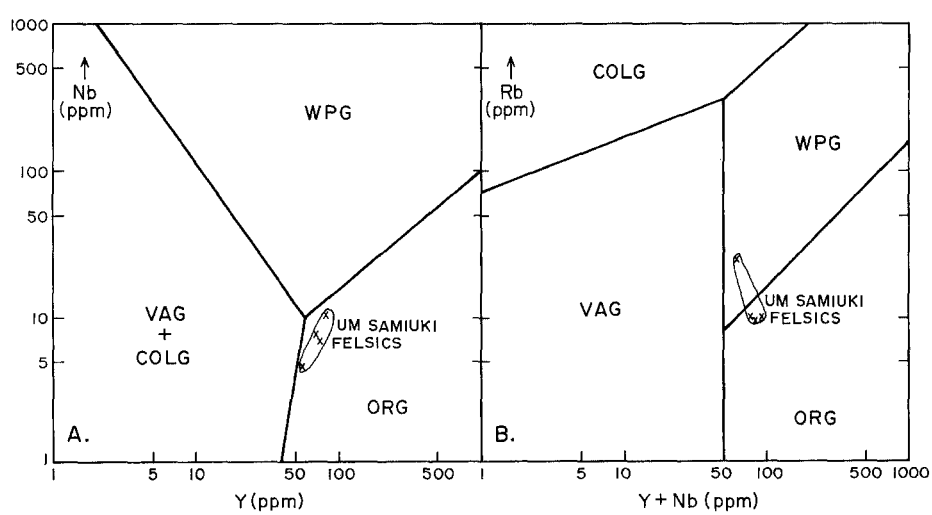

Fig. 9. Trace element discrimination diagrams for felsic rocks (PEARCE et al., 1984) and data for felsic samples ($SiO_2 > 69$ %) from the Shadli Metavolcanics: A: Nb vs. Y, showing fields for volcanic arc granite and collision-related granite (VAG + COLG), within-plate granite (WPG), and ocean-ridge granite (ORG). B: Rb vs. Y + Nb, with fields shown for VAG, COLG, ORG, and WPG. Note that the Um Samiuki felsic lavas plot consistently outside the field defined for convergent-margin suites (VAG).

near sea level. BARBERI et al. (1982) noted that intracontinental rifts could be subdivided into low and high volcanicity rifts. High-volcanity rifts are those rifts with high rates of eruption, such as Afar or the East African Rift, and have several diagnostic features. First, high volcanicity rifts erupt basalt with low K/K+Na and low LIL, such that a good inverse correlation exists between eruption rate and LIL element enrichment. Second, high volcanicity rifts are characteristically bimodal in composition, while low volcanicity rifts erupt a more unimodal distribution of lavas. Third, basalts from high volcanicity rifts tend to be saturated in silica (or nearly so), especially by comparison with the strongly undersaturated lavas erupted from low volcanicity rifts. Finally, high volcanicity rifts correspond to highly attenuated and extended continental crust. BARBERI et al. (1982) interpreted these features of high volcanicity rifts as resulting from the extensive penetration of asthenosphere into the lithosphere.

Following our inference that the Hamamid Group originated in a rift and using the inferences of BARBERI et al. (1982), the Shadli Metavolcanics can be argued to have formed in a high volcanicity rift, that is, in association with a very large amount of lithospheric extension. This is an attractive hypothesis, because it explains why volcanic activity continued very close to sea level in spite of the large amount of material erupted. We do not know what substrate the Shadli Metavolcanics were erupted on; more work on the relationship between the Shadli and the granitic rocks to the north and south and between the Shadli and the metasediments to the east will be required to resolve this question. Nevertheless, the observation that the Shadli Metavolcanics are an autochthonous sequence and are not associated with fragments of oceanic lithosphere indicates that this sequence was erupted on crust that was more continental than oceanic in thickness and composition. This could never have been submerged very deeply, and the fact that between 3 and 10 km of volcanics were erupted, always at or below sea level, requires that an equal amount of subsidence accompanied volcanism. Given the fact that the region must have had a very high thermal regime and so could not cool and subside during this activity, such subsidence could only be accomplished if volcanism was accompanied by large scale lithospheric extension.

Regional Implications

Recognition of the Shadli Metavolcanics as having formed in a high-volcanicity rift on juvenile continental crust requires a re-assessment of models for crustal evolution in NE Africa during the ca. 700–715 Ma crust-forming episode. This period has already been identified as a time when abundant juvenile continental crust was formed in the SED (STERN & HEDGE, 1985), largely as a result of the widespread intrusion of tonalites and granodiorites. It appears that this batholith extends in a general N-S direction, into the Midyan region of NW Arabia (PALLISTER et al., 1988) and well to the south into the Red Sea Hills of NE Sudan, where similar aged units are widespread (STERN et al., 1989) and comprise part of the Serakoit Batholith of ALMOND et al. (1984). This episode was thought to have resulted from subduction beneath the newly consolidated terranes of the Nubian Shield (STERN & HEDGE, 1985). If our age for the Shadli Metavolcanics is correct, this scenario is difficult to reconcile with the E-W trending high-volcanicity rift setting for the Shadli Metavolcanics.

Postulation of a N-S trending subduction zone at 700–715 Ma is difficult to support on a basis of regional considerations as well. Most of the suture zones of the Arabian-Nubian Shield trend NE-SW such that the Serakoit batholith cuts across these trends and so is unlikely to have resulted from subduction along the associated subduction zones. There are four structures that could mark the trace of the requisite N-S trending subduction zone in the Arabian-Nubian Shield: in Arabia, the Nabitah orogen and the Al Amar-Idsas suture and in Sudan, the Kabous melange zone and the Hamisana Shear Zone. The Nabitah Orogenic Belt of Arabia is now believed to be a zone of complex strike-slip deformation (QUICK & BOSCH, 1989). This leaves the Al Amar-Idsas region in the easternmost Shield as the only N-S trending suture in Arabia. With the Red Sea closed, this suture is 600 km distant, and subduction was to the east; furthermore the age of the ophiolite (694 $\pm$ 8 Ma) is younger than the batholith (STACEY et al., 1984). The Kabous melange zone of the NE Nuba Mountains in Central Sudan may be part of a major N-S trending suture that was active prior to the intrusion of the ca. 680 Ma Liri granodiorite (CURTIS & LENZ, 1985), but this subduction zone probably dipped to the west (ABDELSALAM & DAWOUD, 1991). For these reasons, we reject the idea that the subduction zone associated with the Kabous melange zone could have been responsible for the 700–715 Ma episode of crust-formation in SE Egypt, NE Sudan, and NW Arabia. Finally, the Hamisana shear zone of NE Sudan is known to manifest E-W shortening that is younger than 700 Ma, cutting obliquely across the Serakoit

batholith (STERN et al., 1989). If the Hamisana shear zone was situated on an older N-S trending suture, it would have to dip both to the east and the south to generate the batholith. This possibility cannot be precluded, but because of its complexity, we do not find it to be a very attractive hypothesis.

While it is now apparent that the tectonic setting associated with the 700–715 Ma magmatic episode in the Arabian-Nubian Shield needs to be critically re-examined, the rifting environment interpreted for the Shadli Metavolcanics has not been seen for other volcanic rocks of similar age from the region. The 723 ± 6 Ma Kadaweb Volcanic Group of NE Sudan (20°N, 36°20′E) is a unimodal volcanic sequence. Kadaweb basalts have the strong depletions in HFSC Ta and Nb diagnostic of arc sequences and are strongly enriched in the LREE, although they do plot in the OIB field on the Zr/Y vs. Zr diagram (KLEMENIC, 1985, 1987). The Hommagar Volcanic Group of NE Sudan (18°40′N, 36°50′E; 671 ± 8 Ma; KLEMENIC, 1985) still show the HFSC depletion characteristic of arc lavas, although this depletion is not as great as for the Kadaweb (KLEMENIC, 1987). Samples of the 712 ± 58 Ma Nafirdeib Volcanics of NE Sudan (ca. 22°N, 36°15′E) fall predominantly in the field of arc lavas on a Ti-Zr discriminant diagram and so are interpreted to have formed in an arc setting (FITCHES et al., 1983). In Arabia, the 688 ± 30 Ma Fatimah Group (ca. 21°20′N, 38°20′E; DARBYSHIRE et al., 1983) comprises a bimodal volcanic sequence associated with red fluviatile sandstones, marine carbonates, and conglomerates that nevertheless plot on discriminant diagrams in the fields for arc lavas (ROOBOL et al., 1983). Basalts from the 721 ± 55 Ma Surgah and Shwas volcanics of Arabia are strongly depleted in Nb and Y, showing strong affinities to modern arc magmatic sequences (BOKHARI & KRAMERS, 1981).

We believe that it is important to determine whether rifting of the sort observed for the Shadli Metavolcanics was unusual or was of regional extent during the evolution of the Arabian-Nubian Shield. At present, there is not enough high-quality geochemical data on well-dated volcanic successions in the Arabian-Nubian Shield to undertake a regional evaluation. It should be pointed out that ROOBOL et al. (1983) noted that the volcanic sequences of the Arabian Shield were commonly bimodal, an observation that caused them to argue that, although these were interpreted as arc sequences, they were nevertheless different than modern arc sequences which are characteristically unimodal (EWART, A., 1982). While it has recently been documented that ca. 575–600 Ma bimodal igneous activity in the NE Desert of Egypt was extensional and unrelated to subduction (e.g., STERN & GOTTFRIED, 1986, 1990), similar sorts of tectonic environments have not previously been advanced for older rocks. Obviously, the tectonic setting of all bimodal volcanic sequences in the Arabian-Nubian Shield will now have to be more carefully reconstructed.

Conclusions

The following conclusions are warranted from our study of the Hamamid Group and the Shadli Metavolcanics:

1. The Shadli Metavolcanics are composed of a 3 to 10 km thick sequence of bimodal volcanics, erupted near sea level. The preserved sequence, although deformed, is autochthonous.

2. The volcanic sequence was erupted at 712 ± 24 Ma ago, and is one manifestation of a major crust-building magmatic episode that encompasses SE Egypt, NE Sudan, and NW Arabia. Deformation of the Shadli Metavolcanics, including folding along E-W axes, N-or NE-directed thrusting, and NW-SE oriented strike slip faulting must have occurred after 712 ± 24 MA.

3. Initial Sr- and Nd- isotopic compositions indicate that the basalts were derived from melting of depleted mantle.

4. Two types of basalt are recognized in the Hamamid Group. Basalt I is as depleted in LREE as N-MORB but has suffered extensive fractionation. Basalt II is slightly enriched in LREE and has suffered much less fractionation than the older basalt. Neither basalt is a fractionate of the other. Both younger and older basalts formed by extensive melting of the upper mantle, most likely within the stability field of spinel peridotite.

5. The origin of the felsic rocks is unclear. These could have formed either by fractionation of a Basalt I-like mafic melt or could have formed by anatexis of juvenile crust.

6. The Hamamid Group did not form in an island arc. Neither the mafic nor the felsic lavas show geochemical affinities with modern arc suites.

7. The Hamamid Group most likely formed in a high-volcanicity rift. The data favor – but do not compel – the hypothesis that this rift was not formed in a back-arc setting, and that the rift has analogues in the modern Rio Grande Rift or the Afar Triangle. A very large amount of extension is required to have accompanied development of the rift.

8. In light of these data, the nature of the ca. 700–715 Ma crust forming event in NE Africa and

Arabia needs to be critically re-examined. Large-scale lithospheric extension may be more important in the early to middle Pan-African of the region than previously considered.

Acknowledgements

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