Journal of the Geological Society, London, Vol. **144,** 1987, pp. 569–575, 4 figs, 2 tables. Printed in Northern Ireland

Age of Feiran basement rocks, Sinai: implications for late Precambrian crustal evolution in the northern Arabian–Nubian Shield

R. J. STERN & W. I. MANTON
Programs in Geosciences, The University of Texas at Dallas, Box 688, Richardson, Texas 75083–0688, USA

Abstract: Basement exposed on the perimeter of the Red Sea was created during the Pan-African event at the end of the Precambrian. Pre-Pan-African crust in the northern part of this region has not yet been identified. This paper reports the results of Rb–Sr whole-rock and U–Pb zircon dating of gneisses and related basement units from the Wadi Feiran area in the Sinai peninsula, where the existence of such older basement has previously been suggested. A post-tectonic extensional dyke gives a Rb–Sr age of 591 ± 9 Ma with an initial $^{87}Sr/^{86}Sr$ ratio of 0.7034 ± 0.0002. Rb–Sr whole-rock and thin slab dating of paragneisses gives ages of *c.* 610 Ma with an initial $^{87}Sr/^{86}Sr$ ratio of 0.7035. A U–Pb zircon age of 632 ± 3 Ma is interpreted as either the time of formation of these gneisses or the age of the crust sampled by protolith sediments. Granodiorite to the east gives a U–Pb zircon age of 782 ± 7 Ma and is interpreted as representing the westernmost extent of a 780 ± 50 Ma terrane that extends across Sinai into Jordan. Uplift and erosion of the 780 ± 50 Ma terrane supplied detritus to flanking terranes in N and SE Sinai. This region thus acted as a foreland to the younger accretionary and extensional units to the south and west that were active later in the Pan-African event. There is still no evidence for pre-Pan-African basement in the Precambrian units around the northern Red Sea east of the Nile.

Formation of the continental crust of NE Africa and Arabia occurred during the end of the Precambrian (450–950 Ma Pan-African Event of Kröner 1984), largely manifesting a telescoped series of island arcs, marginal basins, and accreted terranes (e.g. Gass 1977; Kröner 1985). One of the most remarkable aspects of this episode of crust formation is the apparent lack of involvement of pre-Pan-African continental crust, especially in the basement exposed east of the River Nile. There is no direct evidence that a well-defined stabilized foreland existed anywhere in this region, and there is a paucity of igneous rocks with isotopic signatures indicating the melting of such crust. An important exception to this generalization is the *c.* 1.6 Ga Afif Terrane in the S Arabian Shield (Stacey & Hedge 1984). Even this may not represent a pre-Pan-African foreland, because it is flanked east and west by wide expanses of Pan-African igneous rocks and so may be an 'exotic terrane' (Stoeser & Camp 1985). It is suggested that *c.* 0.9–1.2 Ga crust may exist in the southern part of the Eastern Desert of Egypt (El Shazly *et al.* 1973; El Manharawy 1977 *in* Cahen *et al.* 1984). These inferences are based on whole-rock Rb–Sr age determinations that either show a large degree of scatter around the regression line (El Shazly *et al.* 1973) or are unpublished and unavailable (El Manharawy 1977). We suspect that *c.* 1.1 Ga crust does exist in the Abu Swayel area (El Shazly *et al.* 1973) but these ages need confirmation by U–Pb zircon studies. The presence of 1.1–2.3 Ga cobbles of granodiorite, quartzite, and marble in the late Precambrian Atud Conglomerate of the Eastern Desert of Egypt indicates that a much older foreland must have been nearby (Dixon 1981), and probably lay just west of the present Nile (Schandelmeir *et al.* in press). With the exception of Archean basement at Uweinat, 1000 km west of the principal Pan-African exposures (Klerkx & Deutsch 1977), this foreland has not been found *in situ.*

Our ignorance regarding the relationship between Pan-African and older units severely limits our ability to reconstruct this region's tectonic evolution accurately. One place to look for pre-Pan-African basement is among the gneissic rocks of Afro-Arabia. It is important to know if these gneissic terranes are significantly older than the surrounding greenschist-facies ophiolitic and arc material and so may have represented a continental foreland during the Pan-African orogenies, or if these terranes are simply metamorphosed equivalents of the lower grade rocks. This problem was first addressed in studies of the Egyptian basement, where gneisses are principally exposed in three areas (Fig. 1): (1) Sinai, typified by the area drained by Wadi Feiran; (2) Central Eastern Desert typified by the rocks of Meatiq Dome; and (3) SE Desert, one representative of which is the gneissic terrane surrounding Gebels Migif and Hafafit.

Earlier investigators considered these gneisses to be palimpsests of much older continental crust (Hume 1934). Schürmann (1966) grouped the Egyptian gneisses into the 'Mitiq Series' which he interpreted to underlie (beneath one or more unconformities) the younger and less metamorphosed parts of the basement. Subsequent studies have not supported these interpretations. Sturchio *et al.* (1983*a*) interpreted the Meatiq Dome gneisses to have formed from sheared and recrystallized felsic igneous rocks at 580–625 Ma accompanying the development of a metamorphic core complex. Low initial $^{87}Sr/^{86}Sr$ (*c.* 0.7030) was interpreted to indicate a lack of remelted older continental crust (Sturchio *et al.* 1983*b*). Gneisses from the Migif–Hafafit area have a more complex history. Eleven deformational and five igneous events have been observed (Elbayoumi & Greiling 1984). A syntectonic tonalite intruded into the gneisses has a U–Pb zircon age of 682 Ma and an initial $^{87}Sr/^{86}Sr$ of 0.7024 (Stern & Hedge 1985). Metasediments and foliated granites from the region give values of ε_{Nd} (at 750 Ma ago) of +6.6 and +5.4 (Harris *et al.*

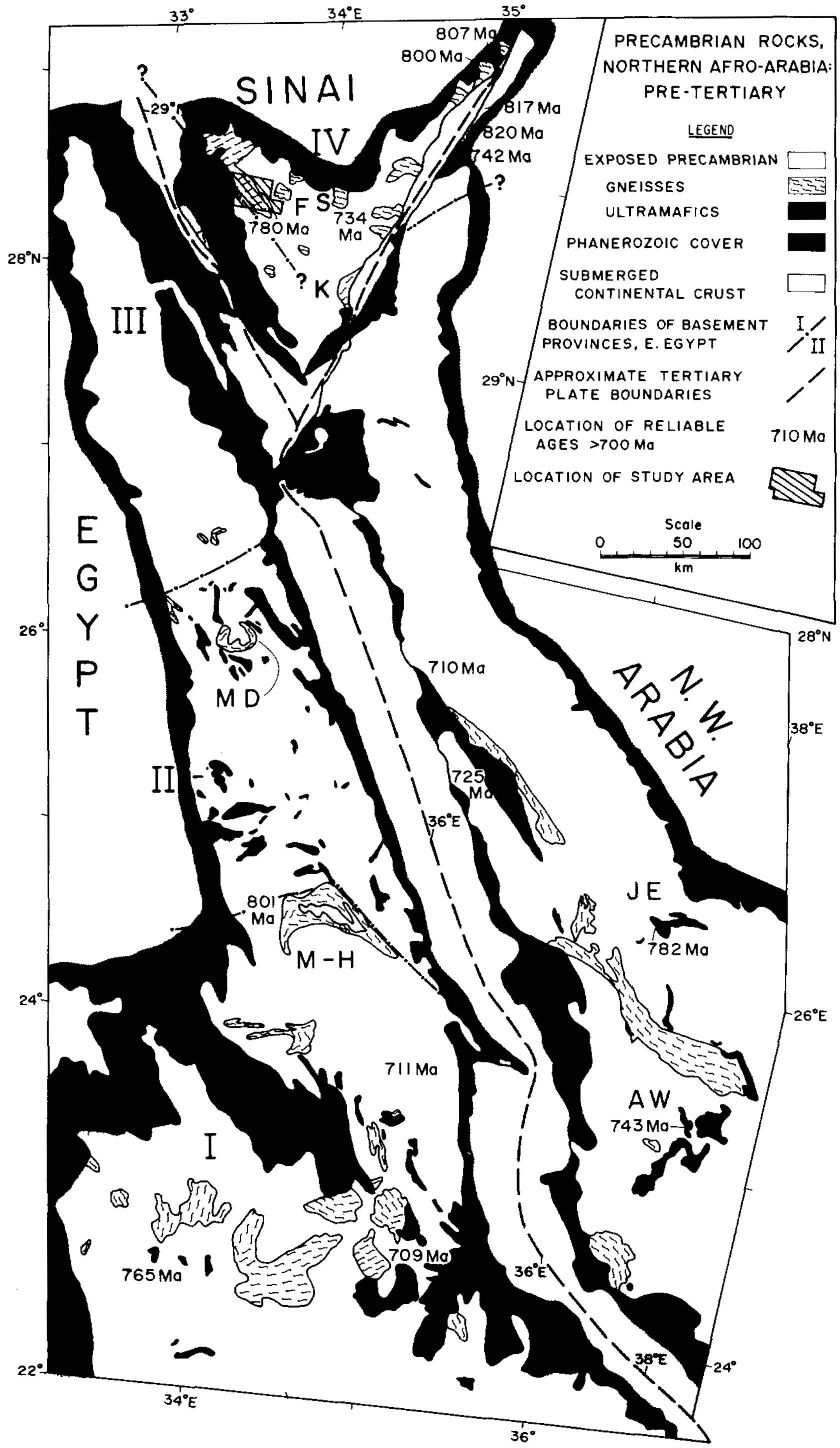

Fig. 1. Locality map of NE Africa, Sinai, and NW Arabia, modified to show the distribution of basement terranes prior to opening of the Red Sea and Gulf of Suez. Also shown are the location of gneisses and associated high-grade rocks and serpentinized ultramafics. Reconstruction includes 110 km translation along the Dead Sea–Gulf of Aqaba strike slip system and 30 km closure of the Gulf of Suez (Freund *et al.* 1970). The designation of basement province boundaries in the Eastern Desert of Egypt are those of Stern & Hedge (1985); designation of basement province IV, Sinai, is a new result. Sources for the geological compilation are as follows: Eastern Desert of Egypt, El-Ramly 1972; Sinai, Eyal *et al.* 1980; Jordan, Bender, 1974; NW Arabia, Johnson, 1983; Clark, 1985. Abbreviations for regions discussed in the text are as follows: M–H, Migif–Hafafit; MD, Meatiq Dome; F, Feiran; JE, Jebel Ess ophiolite; AW, Al Wask ophiolite; K, Wadi Kid; S, Sa'al. Also shown are the ages and locations of more reliable whole-rock Rb–Sr, U–Pb zircon, or Sm–Nd dates with ages greater than 700 Ma. Sources of geochronological data: Hashad *et al.* 1972; Halpern & Tristan 1981; Dixon 1981; Bielski 1982; Jarrar *et al.* 1983; Claesson *et al.* 1984; Hedge 1984; Stern & Hedge 1985; present study.

1984), a further indication that pre-late Precambrian continental crust was not involved in generating the Migif–Hafafit gneisses.

The gneisses of Wadi Feiran represent an excellent opportunity to investigate further the origin of Pan-African high grade terranes. This paper reports our efforts to determine the timing of the principal crust-forming events in the Feiran area and presents our interpretation of their significance for deciphering late Precambrian crustal evolution in N Afro-Arabia.

Geological setting and previous work

The Feiran gneisses are located near the NW corner of exposed basement in Sinai (Fig. 1). The gneisses and associated synorogenic granodiorites are bracketed to the north and south by post-orogenic granites (Fig. 2), dated elsewhere in Sinai and Egypt at 570–600 Ma (Bielski *et al.* 1979; Halpern 1980; Fullagar 1980; Halpern & Tristan 1981; Stern & Hedge 1985). El Gaby & Ahmed (1980) interpreted the Feiran gneisses to represent a thick (>5000 m)

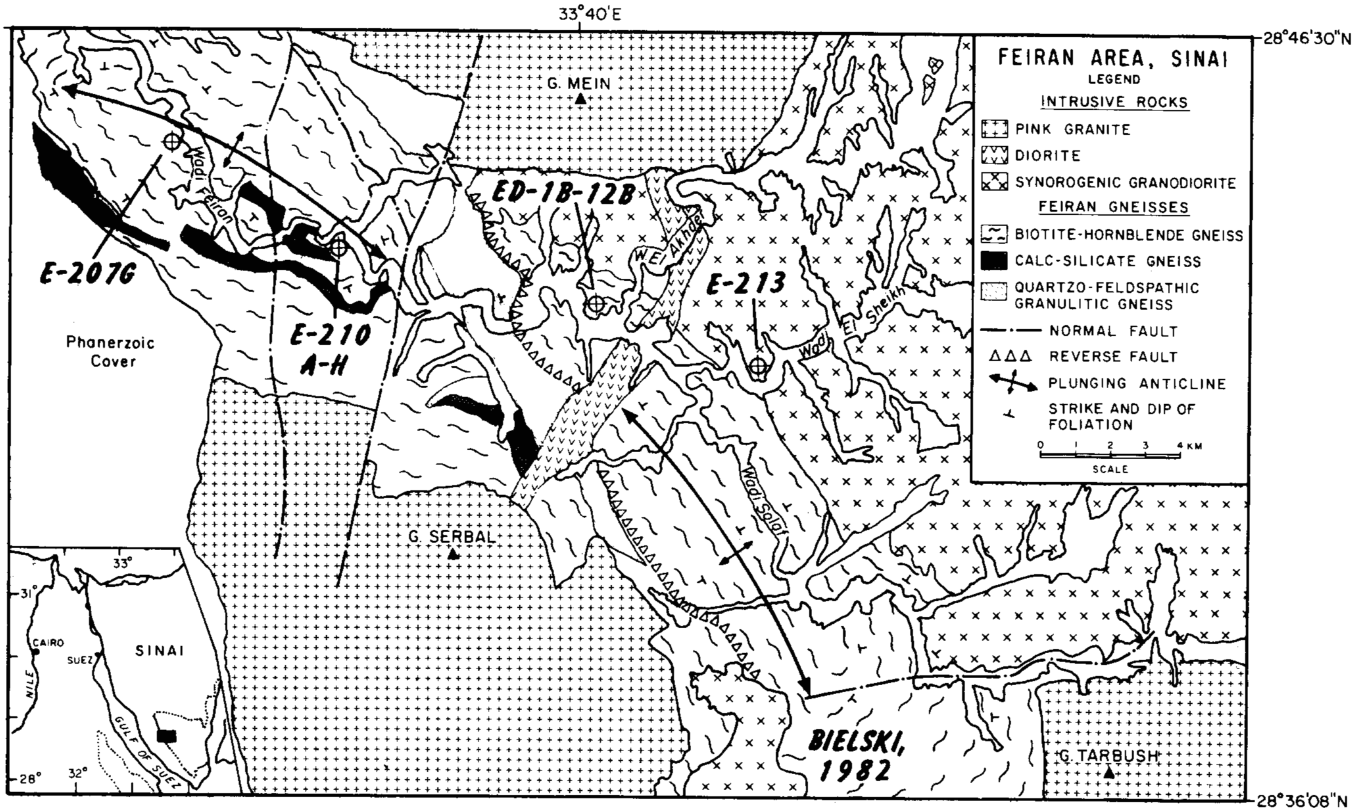

Fig. 2. Geological map of the area around Wadi Feiran, modified after El-Gaby & Ahmed (1980), showing the distribution of samples analysed here along with the approximate location of the suite analysed by Bielski (1982).

sedimentary succession, with minor mafic intercalations, that was folded into three NW–SE trending anticlines overturned to the west and separated by NW–SE trending thrust faults.

Previous investigators have considered the Feiran gneisses to be of pre-Pan-African age (Schürmann 1966; Shimron 1980). Shimron (1980) argued that the gneisses formed prior to 1100 Ma largely on the basis of unpublished geochronological data. Bielski (1982) reported a Rb–Sr whole rock age of 643 ± 41 Ma and an initial $^{87}Sr/^{86}Sr$ ratio of 0.7032 ± 0.0007 for five samples of gneiss collected within the present study area. The gneisses are flanked to the east by an extensive body of foliated granodiorite and are dissected by abundant N30°E trending bimodal dykes (Voegeli 1985). The latter postdate all compressional deformation and are manifestations of late regional extension (Stern 1985).

Analytical techniques

Samples for Rb–Sr whole-rock geochronology were processed according to standard procedures, including cleaning, crushing, and column separation. Rb and Sr were analysed by isotope dilution on the UTD 6"- and 12"-radius solid source mass spectrometers. Total processing blanks are ≤ 3 ng Sr and ≤ 0.1 ng Rb. Age regressions were carried out using the York II treatment of data (York 1969) and all uncertainties are 2σ. Samples containing zircons were crushed, sized, and the nonmagnetic zircons isolated. These were cleaned by boiling in HNO_3 and HCl before dissolution in Krogh-type bombs. Total processing blanks for Pb are 0.5–1.0 ng. Analyses were carried out on the UTD 12"-radius instrument. Regression calculations to concordia are those of Ludwig (1980); uncertainties are at the 95% confidence level. A lower intercept of 15 ± 20 Ma was forced; the justification for this procedure for dating rocks in the Arabian Shield has been presented elsewhere (Stacey *et al.* 1984).

Results

The location of the four suites studied here along with that of the suite analysed by Bielski (1982) are shown in Fig. 2. analytical results are listed in Tables 1 and 2 and are presented in Figs 3 and 4. The Rb–Sr results will be discussed first.

The late dyke swarms are represented by 10 samples, ED–1B to 12 B, taken from a single composite basalt–rhyolite dyke (Voegeli 1985). These give an excellent spread in $^{87}Rb/^{86}Sr$, defining an age of 591 ± 9 Ma and an initial $^{87}Sr/^{86}Sr$ ratio of 0.7034 ± 0.0002 (Fig. 3a). The MSWD of 1.8 is less than the F-variate of 2.02, so this is an isochron age, taken to represent the time of emplacement of this and other NE-trending dykes in the region.

Eight samples of hornblende–biotite quartzo-feldspathic gneiss were collected from outcrops separated by a total distance of no more than 400 m. These rocks were mapped as metasediments by El Gaby & Ahmed (1980); we describe these as paragneisses in the field. The eight points define a regression line with an age of 614 ± 27 Ma and an initial $^{87}Sr/^{86}Sr$ ratio of 0.7034 ± 0.0003 (Fig. 3b). The MSWD of 4.7 is significantly higher than the F-variate of 2.18, indicating that this is not an isochron. Nevertheless, the low intercept and moderate degree of scatter indicate this age is

Table 1. *Rb, Sr, and $^{87}Sr/^{86}Sr$ data*

	Rb (ppm)	Sr (ppm)	^{87}Rb/^{86}Sr	^{87}Sr/^{86}Sr[1]
1. *Dyke*				
ED-1B	78.7	699	0.326	0.70597 ± 6
3B	81.4	639	0.368	0.70639 ± 7
4B	101	518	0.564	0.70819 ± 8
6B	164	35.0	14.05	0.82307 ± 13
7B	184	27.2	20.38	0.87689 ± 8
8B	182	16.2	34.19	0.99169 ± 10
9B	166	22.1	22.46	0.88791 ± 23
10B	102	573	0.515	0.70765 ± 7
11B	78.4	687	0.330	0.70620 ± 7
12B	76.3	672	0.328	0.70631 ± 7
2. *Gneiss*				
E-210A[2]	35.5	342	0.283	0.70606 ± 9
AD	63.1	326	0.560	0.70862 ± 9
AL	18.4	346	0.154	0.70506 ± 5
210B	80.8	309	0.757	0.71000 ± 7
210C	53.4	395	0.391	0.70677 ± 7
210D	36.2	393	0.267	0.70587 ± 7
210E	19.4	407	0.138	0.70445 ± 6
210F	16.8	313	0.155	0.70518 ± 10
210G	46.0	466	0.286	0.70595 ± 7
210H	82.1	482	0.492	0.70775 ± 11
3. *Single samples*				
E-210 Peg.	136	161	2.44	0.72430 ± 7
207G	36.5	358	0.295	0.70654 ± 7
213	74.3	299	0.718	0.71139 ± 6

[1] ^{87}Sr/^{86}Sr normalized to E + A $SrCO_3$ ^{87}Sr/^{86}Sr = 0.70800.

[2] E-210A, 210AD, and 210AL are whole rock, dark layers, and light layers, respectively, of sample E-210A.

probably close to the true age; whether variable initial ^{87}Sr/^{86}Sr or metamorphic disturbances are responsible for the scatter is not clear. Analyses of adjacent 5-cm-thick light and dark layers from E-210A (Fig. 3b inset) define an isochron age of 610 ± 44 Ma with an initial ^{87}Sr/^{86}Sr of 0.7037 ± 0.0002 K-feldspar-rich pegmatite from this locality (E-210 peg.) gives model ages of 606 Ma (I.R. = 0.7032) to 586 Ma (I.R. = 0.7039).

Zircons were separated from a gneiss sample (E-207G) that has a very similar field appearance and petrography to the samples dated by Rb–Sr whole-rock techniques. The zircons were few in number and of small size (<140 mesh). Microscopic examination disclosed that all were euhedral with no apparent overgrown or metamict cores. The single fraction analysed gave a nearly concordant model age of

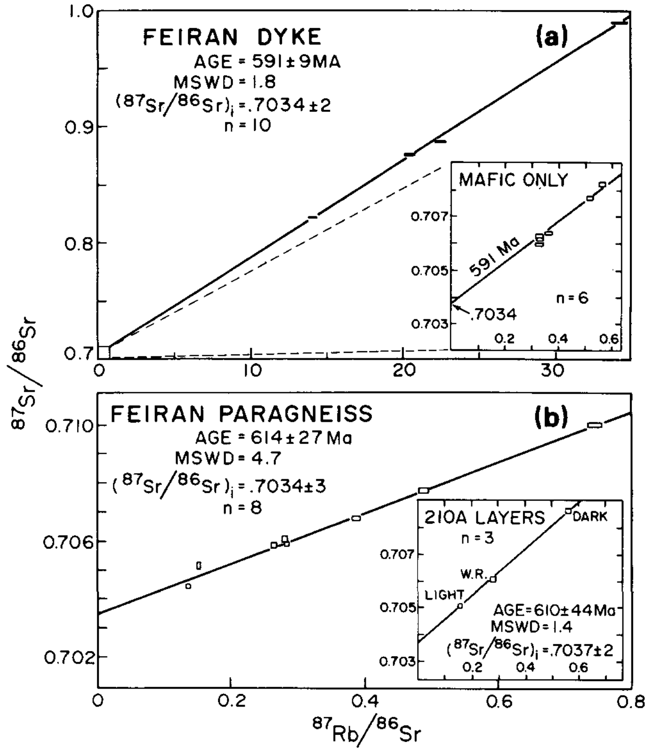

Fig. 3. Rb–Sr whole rock isochron diagrams for rocks from the Feiran area. (a) Late extensional dyke; (b) Paragneiss (inset shows results for thin-layer dating of one sample).

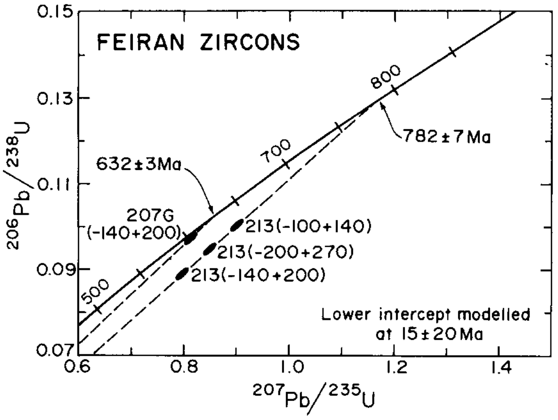

Fig. 4. U–Pb zircon concordia diagram for rocks from the Feiran area.

Table 2. *U, Pb, and Pb isotopic composition data*

Sample, mesh	Concentration U (ppm)	Concentration Pb (ppm)	Measured Pb ratios 206/204	Measured Pb ratios 207/206	Corrected ratios 207 Pb/235 U	Corrected ratios 206 Pb/238 U	Corrected ratios 207 Pb/206 Pb
207 G paragneiss[1]							
−140 + 200	434.4	40.59	7812	0.06255	0.8096	0.09662	0.06077
213 granodiorite[2]							
−100 + 140	475.8	51.06	2667	0.07054	0.9010	0.1002	0.06520
−140 + 200	552.9	53.20	1873	0.07279	0.7968	0.08883	0.06506
−200 + 270	629.0	63.32	6211	0.06736	0.8499	0.09476	0.06505

[1] Corrected for common lead at 620 Ma (Stacey & Kramers 1975).

[2] Corrected for common lead at 790 Ma (Stacey & Kramers 1975).

632 ± 3 Ma (Fig. 4). Using this age and the Rb–Sr data in Table 1, E-207G had an initial $^{87}Sr/^{86}Sr$ ratio of 0.7039.

Strongly foliated granodiorites comprise an extensive terrane east of the paragneiss. Three fractions of zircons separated from sample 213 were euhedral, with no obvious overgrown cores or metamict components. Analytical results were quite discordant but defined an age of 782 ± 7 Ma (Fig. 4). Interpreting this as the crystallization age, the Rb–Sr data in Table 1 indicate that this granodiorite had an initial $^{87}Sr/^{86}Sr$ ratio of 0.7034.

Discussion

There are two possible interpretations of the gneiss zircon age of 632 Ma. This age represents either the time of the metamorphism that formed the gneisses, or the time of igneous activity sampled by the sedimentary protolith. Selection of the correct interpretation is critically dependent on a better understanding of the nature of the gneiss protolith. The gneisses are quartzo-feldspathic and lack minerals that would be found in metamorphosed polycyclic sediments, especially aluminosilicates and garnet; there are no quartzites among the gneisses. The protolith for the Feiran paragneisses was probably very immature, volcanigenic wackés or perhaps lithic arenites interbedded with felsic tuffs. Such sediments predominate in the late Precambrian of this region and were deposited very rapidly following consolidation of their source rocks (Engel *et al.* 1980; Massey 1984). The zircons from the gneiss may have formed in granitic rocks that cooled and were eroded just prior to the deposition of these sediments and tuffs; in this case, the 632 Ma age corresponds to pre-metamorphic igneous activity and the time of gneiss formation is better defined by the Rb–Sr whole rock and thin-slab ages of 610–615 Ma. Alternatively the 632 Ma age dates the gneiss-forming event. This could be the result of either the loss of all lead from detrital zircons or growth of new zircons in metasediments that were originally deposited free of zircon. In this case, the 632 Ma refers to peak metamorphic conditions with 614, 610, and 606–586 Ma Rb–Sr ages corresponding to later times of cooling through whole-rock, thin layer, and K-feldspar blocking temperatures.

In either interpretation, there was very little time between the igneous activity and sedimentation resulting in protolith formation, metamorphism, and uplift. This is demonstrated by the relatively low initial $^{87}Sr/^{86}Sr$ ratios (0.7034–0.7039) of the gneisses and the overlap between the ages given by U–Pb zircon, whole rock Rb–Sr and thin slab Rb–Sr, and the model age of the late pegmatite. Regardless of which interpretation is preferred, the gneiss formed at or just after 632 Ma. The high-grade metamorphic event must have been complete by the time that the 591 ± 9 Ma dykes were emplaced. The tectono-magmatic episode responsible for gneiss formation may have been related to the 610-630 Ma compressional events important in the evolution of the Central and NE Deserts of Egypt (Stern & Hedge 1985). The major NW-SE trending structures clearly developed during this event.

The 780 Ma age from the Feiran granodiorite is especially significant for regional tectonic reconstructions but is more difficult to interpret within the local geological context. The contact between the paragneisses to the west and the granodiorite parallels the paragneiss structural trends and on the basis of the geochronological date must either be an unconformity or a fault. This contact has not been identified in the field as either, probably due to overprinting by the 610–630 Ma event. More detailed structural studies will be required to resolve this problem.

The granodiorite is the westernmost known representative of 780 ± 50 Ma crust which extends across Sinai into Jordan but has not been found in adjacent portions of Egypt or Arabia. In Jordan, sillimanite–garnet schists, 2-mica granite, and paragneiss give U–Pb zircon ages of 742–820 Ma (Jarrar *et al.* 1983). Pelitic schists from a high grade metamorphic terrane around Elat, Israel, give whole-rock Rb–Sr ages of 807 ± 35 Ma (I.R. = 0.7032 ± 0.0005; Halpern & Tristan 1981) or 800 ± 43 Ma (I.R. = 0.7030 ± 0.0006; Bielski 1982). Metavolcanic rocks of the Sa'al Group in Sinai yield Rb–Sr whole-rock ages of 734 ± 17 Ma (I.R. = 0.7029; Bielski 1982). There is thus clearly an extensive terrane of 780 ± 50 Ma basement underlying much of N Sinai and S Jordan. These are the oldest reliable ages from this region, and suggestions of ages greater than 1100 Ma (Shimron 1980) cannot be confirmed.

The 780 ± 50 Ma terrane in northernmost Afro-Arabia has not been recognized in adjacent terranes immediately to the south and west (Fig. 1). Rocks older than 700 Ma have not been found north of 25°N in Egypt in spite of efforts to date the oldest rocks there (Stern & Hedge 1985). Units older than 700 Ma are quite common in Egypt south of 25°N (Fig. 1) and in Sudan (Fitches *et al.* 1983; Klemenic 1985). In the Arabian Shield, units older than 700 Ma are not common north of 27°N (Hedge 1984). To the south, 780 ± 50 Ma ages are first encountered for the 782 ± 38 Ma J. Ess ophiolites at 26°20′N (Fig. 1; Claesson *et al.* 1984). Basement dominated by >700 Ma rocks is only found south of 23°30′N in Arabia (Cooper *et al.* 1979; Fleck *et al.* 1980; J. S. Stacey pers. comm. 1986).

Further support for the interpretation that 780 ± 50 Ma crust is common in Sinai and Jordan and absent from terranes immediately to the south and west comes from consideration of initial $^{87}Sr/^{86}Sr$ ratios. Paragneiss and the dyke analysed have initial $^{87}Sr/^{86}Sr$ ratios of 0.7034–0.7039, similar to the initial ratios for rocks of like age elsewhere in Sinai ($\bar{x} = 0.7040$; Mittlefehldt & Reymer 1986) but slightly higher than initial $^{87}Sr/^{86}Sr$ ratios for rocks of similar age from the Eastern Desert of Egypt (0.7025–0.7035; Stern & Hedge 1985) or from NW Arabia (0.7028–0.7035; Duyverman *et al.* 1982; Hedge 1984). The consistently higher initial $^{87}Sr/^{86}Sr$ ratios of Sinai basement rocks probably reflects the participation of 780 ± 50 Ma crust and/or lithosphere in the generation of the Sinai melts. This component was not present during melt evolution in adjacent regions to the south and west.

The juxtaposition of 630 Ma paragneiss against the 780 ± 50 Ma basement in the Feiran area is especially interesting. This may represent a major crustal boundary within the Precambrian basement, separating 540–670 Ma sialic crust of province III in the NE Desert from 780 ± 50 Ma basement of Sinai and Jordan that was nevertheless severely overprinted during the younger igneous pulses. The latter is identified as a new tectono-magmatic basement province (IV in Fig. 1) to differentiate it from other parts of the Egyptian basement, even though the definition of its southern limits will require further detailed studies in NW Arabia and Sinai.

Finally, these data permit us for the first time to identify an erosional unconformity that was developed early in the

Pan-African evolution of N Afro-Arabia. Priem *et al.* (1984) produced U–Pb zircon ages for granitic pebbles from metaconglomerates from the Sa'al and Kid areas (Fig. 1). The Sa'al metaconglomerate pebbles gave an upper intercept age of $757 \pm^{28}_{20}$ Ma while the Kid pebbles gave an upper intercept age $735 \pm^{28}_{22}$ Ma. These pebbles must have been eroded from a basement similar to that of the 780 ± 50 Ma terrane identified here. This is a further indication that much of the basement of Sinai and Jordan had been consolidated and was being subjected to uplift and erosion while processes of accretionary and extensional tectonics were continuing to the south and west.

Conclusion

The Precambrian basement exposed in Wadi Feiran records almost 200 Ma of crustal growth. The emplacement of granodiorite at 782 Ma is part of a major episode that resulted in the formation of large portions of the basement of Sinai, southern Israel, and Jordan at 780 ± 50 Ma. This terrane was uplifted and eroded, supplying course debris to flanking basins. A second major episode is reflected in the formation of paragneisses at or slightly after 632 Ma; this may represent an eastern extension of tectonic and magmatic activity of this age, commonly found in the Eastern Desert of Egypt. Finally, extensional tectonism and magmatic activity accompanied the emplacement of N30°E-trending dyke swarms at 591 Ma and related A-type granites. Although it is becoming increasingly clear that an older foreland (or accreted terrane?) existed in northernmost Afro-Arabia early in the Pan-African, there is still no evidence for either a pre-Pan-African foreland or reworked pre-Pan-African basement in northernmost Afro-Arabia east of the River Nile.

Field work in Sinai was made possible by the co-operative research programme of Professors E. M. El-Shazly and W. H. Kanes (NSF INT-7801469). Laboratory research was supported by NSF EAR-8205802. The assistance of David Voegeli in the field and in the laboratory was appreciated. David Mittlefehldt made it possible to review critical parts of Miriam Bielski's dissertation. The comments of P. D. Fullager, H. N. A. Priem, A. Reymer, J. S. Stacey and two anonymous reviewers are gratefully acknowledged. This is University of Texas at Dallas Programs in Geosciences contribution number 499.

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Received 16 June 1986; revised typescript accepted 29 September 1986.