

ASPECTS OF METALLOGENESIS AND MICROORGANISMS
IN THE RED SEA REGION OF ~~SANDI~~ ARABIA

by

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Jabal Dhaylan lead zinc district in winter.



Wadi As Sirr rift valley at dawn, looking west.



Barite-jasper vein containing goethitic biomorphs near
Jabal Yahmum.

A B S T R A C T

ASPECTS OF METALLOGENESIS AND MICROORGANISMS IN THE RED SEA REGION OF SAUDI ARABIA

The mineral occurrences in the entire Red Sea region are described and related to the geological development of the Red Sea metallogenic province. The Cenozoic geology of the region is described and explained as a consequence of thermal plume generation in the mantle and subsequent plate separation. Microfossils associated with mineralization were found in the Red Sea deeps, the Arabian coastal plain and the Arabian shield.

Base-metal sulphide deposition from geothermal brines in the Red Sea deeps may be related to the closure of glacial periods and is not significantly active at present. Lead isotope and other geochemical data suggest that Pb, Zn and Cu in the sulphide facies of Atlantis II Deep were derived from sediments like those in the coastal region. Much of the widespread goethitic sediment is biomorphic. The morphology is very sensitive to diagenesis and the biomorphs have not previously been reported.

The coastal sub-provinces contain Pb-Zn-Ba mineralization of the Mississippi Valley type, Mn mineralization of various types, and traces of Cu mineralization of the Colorado red bed type. The writer discovered new minor occurrences of Cu, Mn and Ba. Field evidence indicates the Pb-Zn-Ba mineralization is epigenetic and lead isotope study suggests the lead was derived mostly from surrounding Cenozoic sediments. Phytoclast (trace plant material) alteration and trace element study suggest mineralization occurred at low temperature. The Jabal Dhaylan deposit is post-Miocene and was possibly penecontemporaneous with diapiric movement in the north and volcanism in the south.

In the Red Sea metallogenic province, as in other provinces containing Mississippi Valley type mineralization, the barite-dominant deposits form a separate group to the lead-zinc-dominant deposits. Much jasper occurs in many of the barite veins. This jasper contains remarkable goethitic biomorphs which indicate it is biogenic. This important discovery is believed to be the first record of distinctly epigenetic deposits in which microorganisms have actively participated in the process of mineralization. Microbial activity also appears to have participated in diagenetic and

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epigenetic manganese mineralization in the Arabian coastal plain.

Evidence of iron-precipitating microorganisms was found not only in the Red Sea deeps and in coastal regions but also in the adjacent Saudi Arabian shield. The Precambrian Wadi Sawawin iron ore deposit contains biomorphs and resembles other Lake Superior type iron ore deposits. Microbial remains were also found for the first time in stratabound sulphide deposits at Wadi Bidah and Wadi Wassat, and in some other areas in the Precambrian shield.

G L O S S A R Y A N D A B B R E V I A T I O N S

Names used in this work have been given their local spelling as far as possible but it must be noted that this is generally phonetic and varies somewhat, for example "Jabal" in Saudi Arabia is equivalent to "Gebel" in Egypt. For brevity some names have been shortened. "Wazra" for example refers to the area around Wadi al Wazra and "Azlam" refers to the general area of Qal at al Azlam.

Harrat	area dominated by a lava flow or by volcanic debris
Jabal	hill, chain of hills, or mountain
Marsa	sea anchorage
Qal at	old fortification or ruined castle
Ras	headland
Sabkha	saline silt, clay and sand flat
Sharm	small sea gulf

AAS	Atomic absorption spectroscopy
BRGM	Bureau de Recherches Géologiques et Minières
CP	Crossed polars (with the TLM)
DGMR	Directorate General of Mineral Resources, Saudi Arabia
DY	French diamond drill hole at Jabal Dhaylan
ED	Electron diffraction (with the TEM)
IC	Imperial College
IGS	Institute of Geological Sciences, London
μ	micron (0.001 mm)
PPL	Plane polarized light (with the TLM)
PS	Polished section
RLM	Reflected light microscope
SEM	Scanning electron microscope (Cambridge Stereoscan)
SM	Smear mount (dry and uncovered for XRD, mixed with glycerine for TLM)
TEM	Transmission electron microscope (specimens sedimented onto carbon coated grids; no replicas used)
TLM	Transmitted light microscope
TS	Thin section (used with the TLM)
USGS	United States Geological Survey
XRA	"X-ray analyser", solid state x-ray spectrometer (used in conjunction with the SEM)
XRD	X-ray diffraction
XRF	X-ray fluorescence spectrography

A C K N O W L E D G E M E N T S

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Many other people have been involved in the production of this work and space does not permit acknowledgement to all of them here however the reader will find further acknowledgements at appropriate places in the text.

Miss Fay O'Sullivan did the drafting, reference compilation and typing. She also provided moral and other support through the course of this study and comforted the writer during the dark periods of this work. The writers gratitude and love for this person is beyond verbal expression.

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INTRODUCTION

As the title implies, this study is concerned with a large region rather than a small area or district. It is not the study of a single ore deposit or even of a small group of deposits but rather the study of a metallogenic province. As such it deals as far as possible with all the mineral occurrences, whether of economic potential or not, in the Red Sea region, and the relation of these occurrences to each other and to the geological development of the region as a whole. Because of logistics and patronage special attention has been given to the northern Red Sea coastal plain of Saudi Arabia where field work was carried out over a period of seven months. The Red Sea region is of particular interest because it contains mineralization in young and unmetamorphosed environments which can be related to the modern concepts of plate tectonics.

The aim of this study was primarily to gain a deeper understanding of earth processes related to ore genesis and secondarily to gain information relevant to forecasting the types of deposits likely to be discovered in this region and to gain other information useful in the exploration for new deposits. As the study progressed it became evident that microorganisms have had a significant role in the formation of mineral deposits in all parts of the Red Sea region hence geomicrobiology became a necessary and integral part of the study of metallogenesis.

The Red Sea deeps, the coastal plains and the shield regions are tectonically related and iron precipitating microorganisms appear to have produced deposits, of different types, within them. The various aspects in this study are thus all ultimately related to processes of mineralization and their relationship to the geological development of this part of the Earth's crust.

In the presentation of this work observations have as far as possible been separated from interpretation. Geomicrobiological observations have been further separated from other geological observations, and all observational data are grouped according to geographical regions.

The literature on the Red Sea, its surroundings, and geomicrobiology, is far too extensive to review here however the interested reader will find ample references in the text. For this study an extensive analysis of the literature was made. This was a difficult task because of the enormous quantity of the literature, still growing like the mythological hydra, because of the wide range of subjects covered by it and because of internal inconsistencies within papers. Furthermore many papers conflict directly or indirectly with one another not only in theories and interpretations (which is understandable) but also in basic measureable observational facts. Few writers, for example, even agree the simple length of the Red Sea. It is 1000 miles long according to Degens and Ross (1970) and 1400 miles long according to Meigs (1966), (the Gulf of Suez being excluded). Even in the same book values range from 1680 km (Girdler 1969) to 1940 km (Said, 1969).

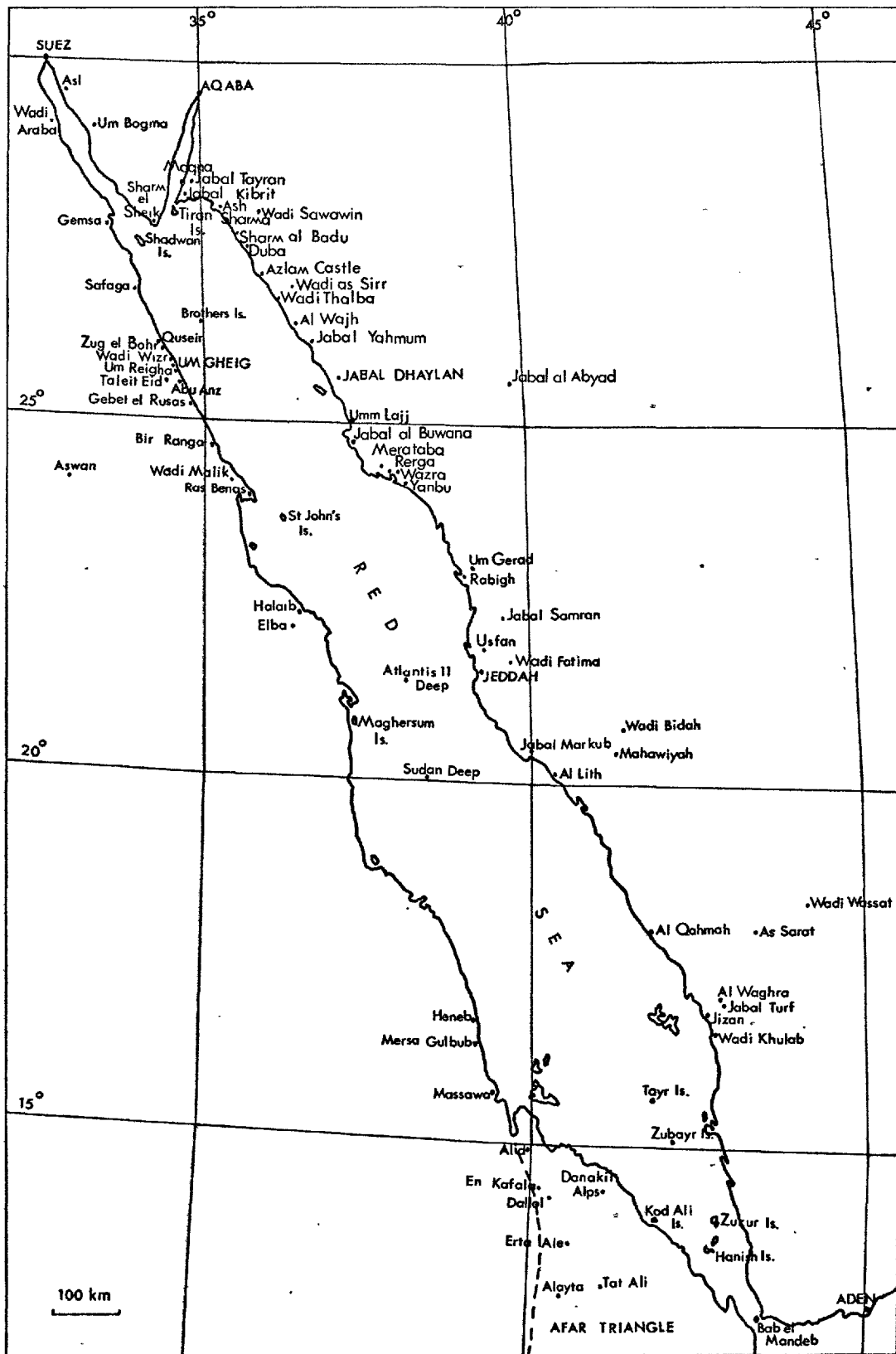


Fig. 1. Locality Map.

PART I

DESCRIPTIVE GEOLOGY

PART I DESCRIPTIVE GEOLOGY

Section 1 The Red Sea
Section 2 The Coastal Region
Section 3 The Cenozoic Igneous Realm
Section 4 The Precambrian Shield

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SECTION 1

THE RED SEA

1.1.1

TOPOGRAPHY OF THE RED SEA FLOOR

The Red Sea is deeper for its width than any other sea. Measured from Laughton's map (1970) it is 1740 km long. The mean sea depth is about 500 m and the width ranges from 180 km in the north to a maximum of 310 km between Massawa and Jizan (localities shown in fig. 1). The marginal shelves are narrow in the north but are wide in the south. The sides of the sea are stepped and have extensive great cliffs. The distance of the 500 fathom (915 m) isobath from the shore line ranges from 2 km near Sharm el Sheik, Sinai to 140 km near Massawa.

The main sea floor lies at about 1000 m depth. Within this is the axial trough which is 30 km wide at its maximum, and extends discontinuously from $16\frac{1}{2}^{\circ}\text{N}$ to $25\frac{1}{2}^{\circ}\text{N}$ (about 1000 km). The axial trough follows the centre of the main trough and forms a zig-zag pattern of NS and NW-SE alignments.

The hot brine deeps within the axial trough are enclosed by the 1900 m isobath. The largest is Atlantis II Deep at $21^{\circ}23'\text{N}$, $38^{\circ}04'\text{E}$. It is defined by the 2000 m isobath and is 14 km long by 5 km wide and is elongated NW-SE. Its flanks generally slope from 5° to 35° and its maximum depth is 2170 m. In places however the boundary walls are overhanging and appear to represent faults (Holmes and Tooms, 1972). Chain deep is south of Atlantis II Deep being separated from it by a sill at 2009 m depth. It is 3 km long and 0.7 km wide. Discovery Deep is west of Chain Deep and is separated from it by a sill at 1980 m. It is 4 km long and 2.5 km wide and is 2220 m deep at its maximum (Ross et al, 1969). Other unnamed small deeps are known near these three deeps. Beyond the Atlantis II group of deeps other deeps with somewhat similar sediments are ^{Suakin} ~~Sawakin~~ Deep at $20^{\circ}00'\text{N}$, $38^{\circ}30'\text{E}$, Nereus Deep at $23^{\circ}00'\text{N}$, $37^{\circ}10'\text{E}$ and Oceanographer Deep at $26^{\circ}20'\text{N}$, $35^{\circ}00'\text{E}$ (fig. 2).

The southern part of the Red Sea is separated from the Gulf of Aden by a sill 125 m deep and 24 km wide at the Straits of Mandeb. To the north it bifurcates into the Gulf of Aqaba, which is a chasm 1600 to 1800 m deep, and the Gulf of Suez which has an irregular flat floor similar to the main trough of the Red Sea but only 55 to 73 m deep.

The most comprehensive reference on the oceanography of the Red Sea is Morcos (1970).

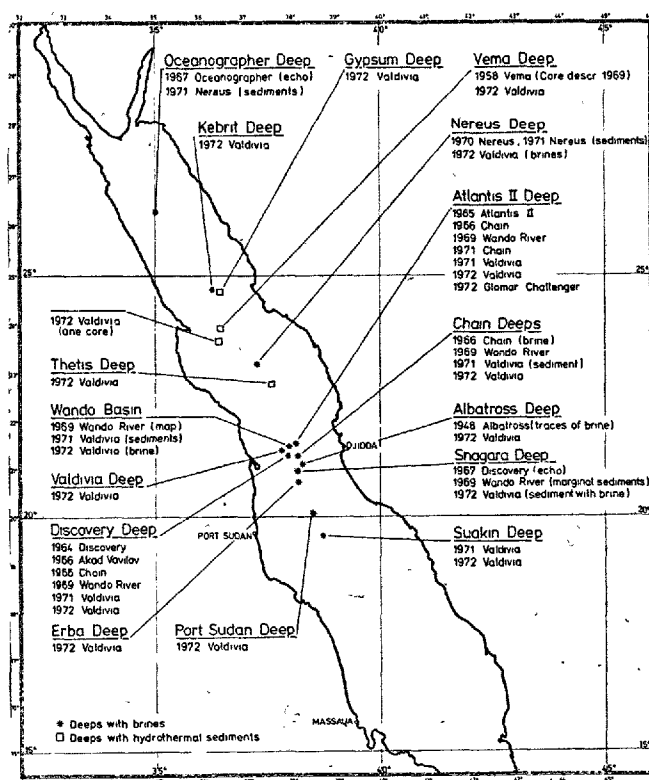


Fig. 2 Distribution of brine pools and brine derived sediments in the Red Sea. From Backer and Schoell (1972).

SECTION 1

THE RED SEA

Contd.

1.1.2

NATURE OF THE RED SEA WATERS

The Red Sea is the most saline body of water in world oceans (Morcos, 1970). Its only connection with world oceans is through the very restricted Straits of Mandeb. No permanent rivers flow into the Red Sea and evaporation greatly exceeds rainfall. Salinities range from 37 per mil in the south to 40 per mil in the north (Siedler, 1969). Water circulation is restricted to the upper 200 m. It enters in the upper part of the water column of the Straits of Mandeb and departs in the lower part richer in salt and organic matter but poorer in oxygen (Olausson, 1971).

The brines are stratified homogeneous layers of water with salinities up to seven times that of the overlying water. The largest and most concentrated are in Atlantis II Deep which is thought to be the source of the cooler brines in Discovery Deep and Chain Deep (e.g. Degens and Ross, 1970). The chemistry of the brines is shown in table 1 (from Emery, Hunt and Hays, 1969).

	Atlantis II Deep	Discovery Deep	Chain Deep	Dead Sea	Ocean Water
Na	92.60	93.05	24.0	32.30	10.76
K	1.87	2.14	0.78	6.17	0.39
Ca	5.15	5.12	1.18	13.98	0.41
Mg	0.76	0.81	1.42	34.45	1.29
Sr	0.04	0.04	0.01	0.20	0.008
Cl	156.03	155.3	41.9	178.20	19.35
Br	0.13	0.12	0.08	4.28	0.066
SO ₄	0.84	0.70	2.81	0.34	2.71
HCO ₃	0.14	0.03	-	0.18	0.72
Si	0.03	0.003	-	<0.01	0.004
Fe	0.08	0.003	-	<0.002	0.00002
Mn	0.08	0.05	0.005	0.004	0.00001
Zn	0.005	0.0008	-	<0.02	0.000005
Cu	0.0003	0.0001	-	<0.002	0.00001
Co	0.0002	0.0001	-	-	-
Pb	0.0006	0.0002	-	<0.002	0.000004
Ni	-	0.0003	-	<0.002	0.0000001
Salinity	257.76	257.37	72.19	270.11	35.71
Temperature	56.5°C	44.7°C	29.1°C	22.0°C	-
Density	1.178	1.183	-	1.233	1.03

Table 1 Major and minor ions in the Red Sea brines in g/kg (from Emery, Hunt and Hays, 1969).

The 56.5°C brine (table 1) in 1966 filled Atlantis II Deep to 2042 m b.s.l. above which was a 5 m thick interface. Above this interface was a 28 m thick brine layer with a temperature of 44.3°C and a salinity of 135 per mil (Ross, 1972). This was topped by a mixed zone 25 m thick across which the temperature ranges down to the normal deep Red Sea value of about 22°C. The Discovery Deep brine pool has a similar structure but the brines are thinner, cooler and less saline (Degens and Ross, 1970). The brines in the upper layers at both Atlantis II Deep and Discovery Deep have major element concentrations consistent with that expected for ocean water dilution (Tooms, 1970).

A dense accumulation of suspended colloidal matter and organic debris rests above each brine layer (Ryan et al, 1969). Low molecular weight hydrocarbons are enriched in the brines by a "factor of 10 to 1000" (Swinerton and Linnebo, 1969).

In the 52 months ending March 1971 the bottom brine in Atlantis II Deep increased from 56.5°C to 59.2°C (Ross, 1972). By April 1972 the temperature had reached 61.2°C (Ross et al 1972). By March 1971 the upper Atlantis II brine temperature had increased from 44.3°C to 49.7°C and the temperature structure in the uppermost zone had become more irregular. No salinity increase was noted but the level of the lower brine rose 7 m and the separation between it and the next layer decreased from 5 m to 2 m. The minimum temperature of the incoming brine for this is about 104°C and the corresponding average rate of increase of water is about 700 gallons/sec (Ross 1972). Although the brines are nearly isothermal the underlying sediments show wide temperature variations both laterally and vertically within the Atlantis II sediments. The maximum temperature (in 1966) was 62.3°C (Erickson and Simmons, 1969). There is also a positive thermal gradient in the sediment below Discovery Deep but the water at the sediment-water interface is 0.1°C cooler than the brine above (Erickson and Simmons, 1969).

Water stratification similar to that in Atlantis II Deep has been found in Suakin Deep and Sudan Deep but in other deeps small undifferentiated transition zones are typical (Backer and Schoell 1972). The brines of Atlantis II Deep and the neighbouring deeps have the highest chlorinities but depletion of Mg and SO₄ as in the Atlantis II Deep brines has not

been found with other brines in distant deeps. Hydrogen sulphide is abundant in Kibrit Deep and carbon dioxide occurs in Valdivia, Kibrit and Atlantis II Deep. In some areas with abundant geothermal sediments e.g. Thetis, Vema and Gypsum Deep only a slight increase of the chlorinities has been detected in the connate water (Backer and Schoell, 1972).

In a study of dispersion of metals from the Red Sea deeps Holmes and Tooms (1972) found that the effect of the Atlantis II Deep brines on the dissolved species and the particulate matter in the overlying near-bottom sea water could be detected for several miles but a much less marked effect was found with the waters around Nereus Deep. They also found that, except near the bottom in the median valley where metal concentrations are relatively high, there is a general decrease in the metal content of dissolved species and particulate matter in the sea water with increasing depth. Furthermore they found that the surface waters over Atlantis II Deep do not appear to have anomalous dissolved metal contents.

1.1.3

SEDIMENTS OF THE MAIN RED SEA TROUGH

GENERAL Beyond the coastal shelves, which are dominated by reefs and their associated sediments, the sea floor is covered mostly with pelagic calcilutite (or marl). The various sediment types are massive calcilutite, heterogeneous calcilutite, lithified aragonitic "hard layers", and geothermal brine sediments (Ku, Thurber and Mathieu, 1969).

The massive calcilutite contains planktonic forams, pteropods, coccoliths and clay as major components and detrital quartz, feldspar, mica etc. as minor components. The heterogeneous calcilutite has the same constituents as the massive lutite but it is composed of interbedded thin layers of this sediment having various textures. The lithified aragonitic hard layers consist of pteropods and forams in an abundant felted matrix of aragonite. They are uncommon and usually less than one meter thick (Ku, Thurber and Mathieu, 1969; Olausson, 1971; Friedman, 1972).

In addition to these types the writer found centimeter thick layers of detrital lutites. These are dominated by quartz, and contain lesser feldspar, chlorite, amphibole, mica and kaolinite in various proportions together with calcareous matter.

Goethite is a minor constituent of all the lutites and lutite is a common minor constituent of the goethite facies of the geothermal brine sediments. Fresh angular basaltic glass shards are also rare to abundant in some specimens of Holocene lutite. Upper Pleistocene tuff layers in the southern part of the Red Sea were noted by Ross et al (1972).

The proportion of biogenic carbonate to non-biogenic detrital minerals in the lutites decreases with distance from the shoreline (Shukri and Higazi, 1944; R. ^{Biggell} ~~Biggell~~, pers. comm. 1972). The lutite in the south is darker than that in the north, probably because of an organic sediment trap. Some organisms floating into the Red Sea from the Gulf of Aden would probably die from hypersalinity and gradually sink to the floor (Friedman, 1972). The lithified aragonitic layers are also more common in the south (Friedman, 1972) however this is probably due to salinity conditions when the Red Sea was isolated from the Gulf of Aden.

MINERALOGY The calcilutites contain both high- and low- magnesium calcites with a slight dominance of the latter; aragonite is generally absent from these sediments (Friedman, 1972). High-magnesium calcite, however, is associated with the aragonite in the hard layers (Olausson, 1971). Gypsum and anhydrite are almost entirely absent from the common Red Sea sediments outside the hot brine pools. Friedman (1972) considered this remarkable because the Red Sea was an evaporating barred basin several times in the Pleistocene. He explained the dominance of low-magnesium calcite over high- magnesium calcite, and the absence of sulphates and aragonite from the calcilutite as the result of widespread activity of sulphate-reducing bacteria. These bacteria use calcium sulphate to oxidize organic matter and so obtain energy. The products of this activity are low-magnesium calcite, hydrogen sulphide and carbon dioxide.

The aragonitic hard layers are thought to have formed during periods of high temperature and evaporation together with lowered sea level (Friedman 1972). Supporting evidence is the occurrence of heavy isotopes of oxygen and carbon. According to Ku, Thurber and Mathieu (1969) no hard layers have been formed in sediment deposited after the close of the Wisconsin-Würm glacial period 11,000 years ago. Friedman (1972) however claimed that "the youngest lithified aragonite layer having an age of 11,000-13,000 years B.P. coincides with times of maximum Quaternary warmth". Olausson (1972) found some dark bands of iron minerals in the hard layers together with traces of anhydrite, gypsum and sphalerite.

Shukri and Higazi (1944) found that most of the sediment greater than 0.3 mm diameter is biogenic carbonate and that most of the heavy minerals are less than 0.14 mm diameter. Typical heavy minerals, however, are not abundant in the Red Sea and are totally absent from some places. Where they do occur they are associated with pyrite and easily weathered common rock forming minerals (e.g. pyroxenes). They are relatively homogeneous over most of the sea floor and show no systematic decrease in volume with distance from the shoreline. This is characteristic of wind transported sediments.

Pyrite is more common in the north; in the Gulf of Suez it constitutes 43% of the heavy mineral fraction. It is associated with organic matter near Shadwan Island in the Red Sea, where it constitutes 25% of the heavy mineral fraction (Shukri and Higazi, 1944). It is authigenic and most commonly occurs as framboids filling hollows in microorganisms especially forams (personal observations, e.g. fig. 3, and Taylor, 1971).

PALAEONTOLOGY According to Herman and Rosenberg (1969), and others, (e.g. Firedman, 1972; Degens and Ross, 1969 etc) pteropods and planktonic forams are the commonest organisms in the Red Sea cores. The commonest organisms however are probably smaller than these. For example, using the electron microscope, the writer found coccoliths to be the dominant microorganisms in the calcilutites. This is not surprising when a gram of sediment may contain over a million coccoliths. A typical view is shown in fig. 4. Other fossils in the calcilutites are heteropods, gastropods, embryonic pelecypods, benthonic forams, sponge spicules, alcyonarian and echinoid spines and plates, ostracods, otoliths, radiolarians, diatoms, dinoflagellates and bryozoans (e.g. see Degens and Ross, 1969). Fossils of unknown forms like pteropods but much smaller (e.g. fig. 4) are also relatively common.

A variety of palaeontological and isotope-geochemical techniques, used by different workers on different cores, has produced confusing results, although there is general agreement on the date of the last glacial period i.e. 11,000 to 13,000 years ago. According to Berggren's summary (in Degens and Ross, 1969) of this work there were four distinct glacial-salinity cycles in the Red Sea during the last 70,000 years -

viz.	cycle I	80,000 to 50,000 years ago tropical to sub-tropical climate
	cycle II	50,000 to 25,000 years ago cooling and lowering of sea-level
	cycle III	23,000 to 13,000 years ago coolest time of late Pleistocene and high salinities
	cycle IV	11,000 years ago to the present general rise in temperature



Fig. 3. TLM photograph of framboids in foram in typical Red Sea detrital sediment (PS in PPL).

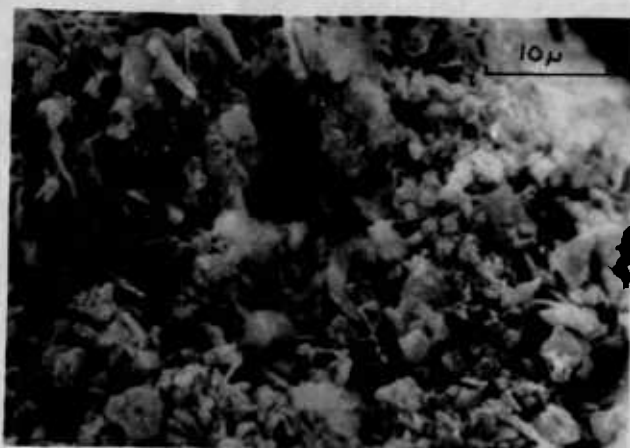


Fig. 4. SEM photograph of typical Red Sea detrital sediment showing coccoliths, mineral flakes and some tiny pteropod-shaped forms.

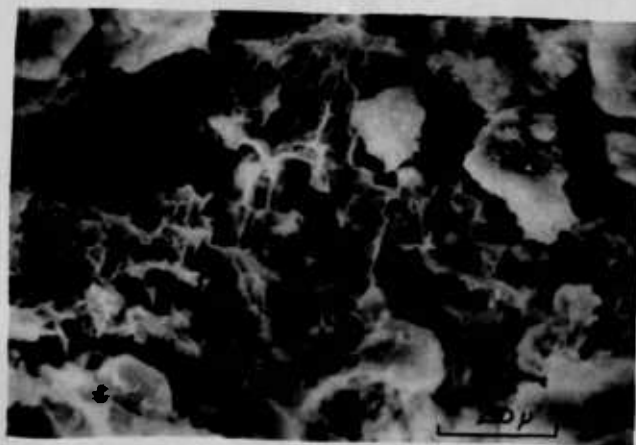


Fig. 5. SEM photograph of dessicated typical iron montmorillonite from Atlantis II Deep.

Another summary by Ku et al. of the same work and in the same book (Degens and Ross, 1969) gives the following periods as those of high temperatures and normal salinities

70,000 to 65,000 years ago

42,000 to 35,000 years ago

24,000 to 20,000 years ago

11,000 years ago to now.

The intervening periods correspond to low temperature, high salinity and isolation of the Red Sea from the Indian Ocean.

CHEMISTRY The chemistry of the sediments has been described by Tooms and Rugheim (1969), Bischoff (1969) and Holmes and Tooms (1972). An averaged analysis from Bischoff is given in table 3.

The writer made a semi-quantitative XRF analysis for Y Sr Ba Ni Cr Cu Pb Zn on a composite sample. Compared with coastal limestones the only anomaly is a high Sr content in the order of 1000 ppm. This is also much higher than the Sr content of the sediments in the geothermal brine deposits. The concentration is possibly due to coccoliths as it is well known that brown algae concentrate Sr (Zajic, 1969).

Holmes and Tooms (1972) found that the ore metal content of the calcareous sediment varies with depth, being greatest within the median valley and lowest near the coast. The calcareous sediment above the basalt along the central rift zone commonly has an increased Fe and Mn content similar to sediments found near the Mid-Atlantic Ridge (Backer and Schoell, 1972).

1.1.4 SEDIMENTS ASSOCIATED WITH THE GEOTHERMAL BRINES

GENERAL The sediments associated with the geothermal brines are of great interest because of their economic potential and because of their young age. Although sulphide minerals attract the most interest it should be noted that they generally form a minor part of these essentially goethitic sediments. The geothermal sediments occur at several localities in the Red Sea deeps (fig. 2) but the full range of mineralogical facies in Atlantis II Deep has been observed elsewhere only in Nereus Deep (Holmes and Tooms, 1972).

Atlantis II Deep lies within the axial zone of the Red Sea at a site of submarine volcanic activity associated with an anomalous magnetic gradient of 125 γ per km, and a Bouger anomaly of over 120 mgals (Hunt et al 1967). Both the rate of hot brine influx and the rate of sediment deposition are very high e.g. table 2.

Sediment type	rate of deposition
common sediments outside the axial trough	5 to 60 + cm/1000 years
carbonate sediments	8 cm/1000 years
detrital silicate sediments	2 cm/1000 years
geothermal sediments	40 + cm/1000 years

Table 2 Rates of sediment deposition in the Red Sea, from Ku, Thurber and Mathieu (1969).

The sediments are essentially brightly coloured finely layered gelatinous oozes with water contents up to 96%. They are well bedded on a scale ranging from a few mm to several m (Bischoff, 1969). Bischoff divided the sediment types into facies based on physical appearance and mineralogy. Although the facies are intermixed and finely interbedded, in places, they usually can be correlated from core to core. A generalized cross section of the geothermal sediments in Atlantis II Deep is shown in fig. 6.

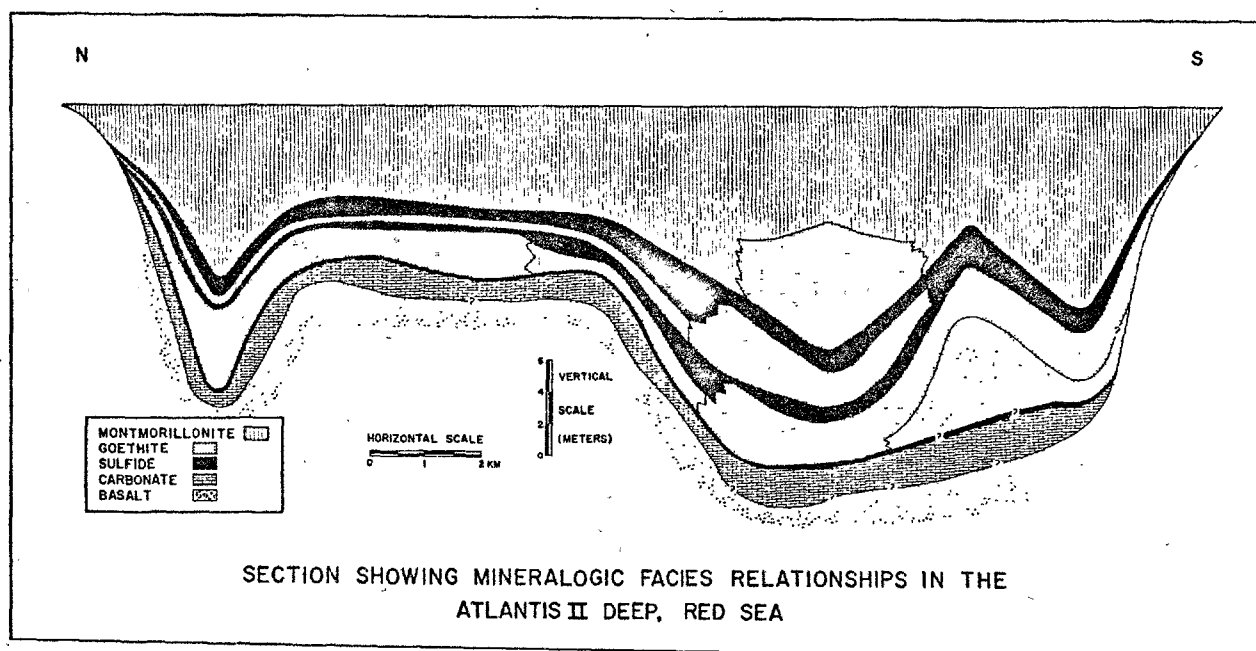


Fig. 6 Generalized cross section of the geothermal deposits in Atlantis II Deep. Section depicts only major events and correlative facies. (From Hackett and Bischoff 1973).

DETRITAL FACIES The detrital facies consists of the common sediments of the main Red Sea trough as described in section 1.1.3. This facies is typically pale brown and is found in varying proportions throughout the other facies.

GOETHITE FACIES The goethite facies is the dominant non-detrital facies of the sediments associated with the geothermal brines. Although Bischoff (1969) reported it as the second most abundant facies in Atlantis II Deep it is the main facies in all other deeps in which it has been discovered. It is widespread and is commonly found beyond the limits of the brine pools. It occurs as far south as latitude 17°N (Olausson, 1971) and as far north as 24°N (Herman and Rosenberg, 1969). Herman and Rosenberg found it in beds up to 0.5 m thick enclosed in sediments 100,000 years old. Although it is not now associated with hot brines in some areas, it is believed to have a similar genesis to the goethite found in Atlantis II Deep with which it is indistinguishable. In Atlantis II Deep it ranges in thickness from 0.01 m to 4 m with an average thickness of about 1 m.

(Bischoff 1969).

The goethitic sediment is yellow to reddish brown and orange, and consists essentially of uniformly fine grained well sorted goethite "spherules" 1 to 50 μ thick. The "spherules" are actually spiky balls of sub-microscopic crystals. It ranges from X-ray amorphous to well crystalline, and, in places, has been transformed to hematite by dehydration. Lepidocrocite and maghemite are very minor components of this facies. Bischoff found pyrite, pyrrhotite and greigite associated with the goethite, and Olausson (1971) found sphalerite associated with abundant goethite in an aragonitic "hard layer".

This facies is described more fully in section 2.2.2.

IRON MONTMORILLONITE FACIES Although this is recorded by Bischoff (1969) as the dominant and uppermost facies of Atlantis II Deep, the writer has observed it in only one of several cores from the upper part of these sediments. This facies is so far only known in Atlantis II Deep, Discovery Deep and Nereus Deep. Montmorillonite is rare in the common Red Sea sediments. According to Bischoff the iron montmorillonite is of intermediate composition between dioctahedral nontronite and a new ferrous-bearing trioctahedral montmorillonite. It would therefore seem possible to distinguish geothermal montmorillonite from detrital montmorillonite, however the montmorillonite studied by the writer has very little ferrous iron in it. The sample was collected in 1966 and allowed to dry in the air for years. It was therefore probably quite oxidized by the time it was obtained for examination. Its total iron content is about 40% (by XRF) and the FeO content is about 0.4% (kindly determined by A. J. Thompson using AAS). Bischoff's corresponding average figures are 37.1% (tot. Fe_2O_3) and 11.7% (FeO) with an extreme of 31% (tot. Fe_2O_3) and 1.5% (FeO).

This facies is a dark brown, finely bedded ooze with 90 to 96% interstitial brine. It is found in layers 4 to 6 m thick and consists essentially of montmorillonite and amorphous material with minor quantities of goethite, manganosiderite and sphalerite (Bischoff, 1969). Sulphide minerals were not found in the montmorillonite examined by the writer, probably because of oxidation. Although the scanning electron microscope revealed that this material has the typical flaky morphology of clays, it more

characteristically has a spongy structure with well developed boxworks in places (e.g. fig. 5). This is probably due to the growth of halite crystals as the material compacted and dehydrated. These delicate structures have survived the processes of leaching with deionized water, re-sedimentation and drying.

SULPHIDE FACIES The sulphide facies is represented by a homogeneous black bed "throughout the bottom of Atlantis II Deep" (Bischoff, 1969). It was the deepest horizon penetrated in 1966 and was found in only four cores. One of these passed through the sulphide layer showing it to be a bed 1 m thick resting on detrital material (Bischoff 1969). It consists essentially of pyrite, "amorphous iron monosulphide", sphalerite (marmatite), and amorphous silica with lesser amounts of associated chalcopryrite, calcite, barite and anhydrite (Bischoff, 1969). Stephens and Wittkopp (1969) found that marcasite is the dominant sulphide. They found marcasite and anhydrite as individual euhedral crystals up to 2 mm long. These minerals more commonly are intergrown, especially with smaller marcasite crystals situated along the axis of anhydrite crystals. Stephens and Wittkopp also found marcasite filling the chambers of several microorganisms. Chalcopryrite and marmatite occur as irregular masses attached to marcasite crystals. These minerals are intergrown such that each encloses the other and no regular order of crystallization could be found (Stephens and Wittkopp, 1969). Taylor and Volframs (Taylor, 1971) found veil-like films of pyrite, similar to an exsolution texture, in the marmatite.

A sample of the sulphide facies, collected in 1966 and left exposed to the air since, was examined by the writer but no sulphide minerals were found in it, presumably because of oxidation. Under the Scanning Electron Microscope, gypsum and anhydrite crystals were recognised but the ground-mass was irresolvable. Cores collected from the Red Sea by the Applied Geochemistry Group of Imperial College were also extremely poor in sulphides. Metal contents in these cores very rarely reached as much as half a percent Cu and Zn (Rugheim, 1970; R. Holmes, R. ^{Biggall} ~~Biggall~~ pers. comms., 1972). The writer failed to find sulphides in samples from these cores even after centrifuging dispersed sediment in heavy liquids. The gelatinous nature of the material of course was a hinderance. One sample of sulphide bearing sediment however was available from the Applied

Geochemistry Group. This is part of a lithified black laminated layer from the base of core 70/23 from Atlantis II Deep. It is part metallic and massive, and part dull and massive. The dull material is extremely fine grained and all of it except tiny pyrite-marcasite grains resisted polishing. X-ray diffraction showed it to consist of pyrite, marcasite, hematite, halite, iron sulphate(?) and anhydrite (or possibly hollandite, main XRD peak at $3.49 \text{ \AA}$). The metallic part is almost entirely marcasite. The only other sulphide observed was a tiny grain of chalcopyrite enclosed in marcasite. Some of the marcasite appears isotropic but it lacks the colour of pyrite. The marcasite occurs in wavy layers of radiating fibres and, in more massive areas, in wavy layers of spheres. Smaller spheres ranging from 0.01 to 0.10 mm diameter occur as concentrations of individual grains and as mosaic clusters forming larger spheres of 0.1 to 1.0 mm diameter.

MANGANITE FACIES The manganite facies locally occurs as black beds 15 to 50 cm thick in Atlantis II Deep. It consists of well-crystalline manganite, todorokite, groutite and woodruffite. Mn_3O_4 and Fe_2O_3 are the major components with the iron probably partly substituting for Mn and partly as amorphous "limonite" (Bischoff, 1969).

No manganite was observed by the writer but some dark goethite layers very rich in manganese were observed. These are more fully described in section 2.2 along with the goethite facies

MANGANOSIDERITE FACIES This "facies" consists of 1 to 20 mm thick, pale yellow-brown layers of manganosiderite which occur locally within the montmorillonite facies and the goethite facies of Atlantis II Deep (Bischoff, 1969).

ANHYDRITE FACIES This occurs as massive, white, pure anhydrite beds up to 20 cm thick in Atlantis II Deep (Bischoff, 1969).

MAGNETITE FACIES This has recently been found in association with goethite in the south western part of Atlantis II Deep (Hackett and Bischoff, 1973).

CHEMISTRY Bischoff's averaged chemical analyses for the facies are given in table 3 below.

	Detrital	Fe Mont- morillonite	Goethite- Amorphous	Sulphide	Manganite
SiO ₂	27.3	24.4	8.7	24.7	7.5
Al ₂ O ₃	8.4	1.7	1.1	1.5	0.7
Fe ₂ O ₃ (total)	6.5	37.1	64.2	24.3	30.5
FeO	1.4	11.7	2.7	13.4	0.4
Mn ₃ O ₄	0.6	2.1	1.1	1.1	35.5
CaO	23.6	4.8	3.4	2.5	2.9
ZnO	0.08	3.2	0.7	12.2	1.4
CuO	<.01	0.8	0.3	4.5	0.1
CO ₂	23.1	8.6	3.6	5.7	2.2
S	0.3	3.9	0.6	16.8	0.6

Table 3 Chemical analyses of Red Sea sediments averaged from 43 selected analyses (from Bischoff, 1969).

The silica content of the detrital facies is not surprisingly high considering the observations discussed in section 1.1.3. The silica contents of the sulphide and goethite facies however do seem high, especially when compared with the lime contents. This is possibly because calcite is unstable in the slightly acid brines. The silica is probably colloidal and in the form of fossils e.g. radiolaria. The alumina content of the iron montmorillonite seems very low and it is surprising that no one has previously drawn attention to this. A "pink" montmorillonite" quoted by Deer, Howie and Zussman (1966) has almost 20% Al₂O₃.

Burnett (1971) found that Cu is associated with Ba in areas of high organic productivity in the East Pacific Rise region. As the Red Sea is somewhat analogous with this region (discussed further in section 3), the writer checked on the association of Ba and Cu in the Red Sea sediment. Semi-quantitative XRF analysis revealed the following orders of concentration

<u>Facies</u>	<u>Order of Concentration</u>
Montmorillonite (oxidized)	4000 ppm
Sulphide (oxidized)	2000
Goethite	500
Detrital	100

Perhaps the sulphide facies should be higher than the montmorillonite facies as barite crystals have been found in it by Bischoff. These results appear encouraging so it has been suggested to the Applied Geochemistry Department (where the chemistry of the sediments is being investigated) that further correlations of Ba and Cu are worth looking for in the detrital and other facies.

1.1.5

IGNEOUS ROCKS IN THE RED SEA

PLUTONIC ROCKS

The plutonic rocks of the Red Sea have been considered to be part of the continental crust, and those close to the coast such as Tiran Island most probably are. This also appears to apply to Shadwan Island which occurs at the junction of the Red Sea and the Gulf of Suez (Shukri, 1954). St. John's Island ($23^{\circ}40'N$, $36^{\circ}10'E$) and the Brother's Islands ($26^{\circ}19'N$, $34^{\circ}30'E$) however are well out in the Red Sea, and are most probably unrelated to the continental crust. They are actually very steep submarine mountains that rise over 1000 m above the sea floor and are located near the median trough.

The basement of the Brother's Islands is essentially anorthositic gabbro intruded by an olivine gabbro, both of which are sheared and altered (Shukri, 1944). Other rock types include metabasalt, olivine dolerite, talc-carbonate rock, and an olivine-bearing limestone conglomerate.

St. John's Island is rather similar to the Brother's Islands and is geologically fascinating. It has been known since ancient times as a source of beautiful peridot gems up to 190 carats in weight.

Garnierite veins with traces of platinum have also been found in the Main Peridotite Hill (Moon, 1922). The Ni content ranges from 5% to 9%, and over 1000 tons had been mined by 1938 (El Shazly, 1959). Recently a new mineral, "magnesian cancrinite", has been found there by El Shazly and Saleeb in crystals up to 140 mm long. The island consists of a metamorphosed basic to ultrabasic complex partially overlain by Miocene to Recent sediments and Quaternary basalt comparable to that in the axial trough (El Shazly and Saleeb, 1972). Peridotites and pyroxenites are also common (Moon, 1922). The metabasalt forms thick apparently folded sheets overlying the metamorphosed dunite in the Main Peridotite Hill. A scapolite-cancrinite association was formed by "large scale alkali metasomatism of the metabasalt" (El Shazly and Saleeb, 1972).

Previous writers (e.g. Moon, 1922; Shukri, 1944; El Shazly and Saleeb, 1972) have considered St. John's Island and the Brother's Islands to be fragments of the Egyptian continental crust. Presumably no alternative

seemed possible. However it is probable that these islands are fragments of uplifted oceanic crust. Such a mode of origin is consistent with Carey's (1958) model of the Red Sea which preceeded the advent of "Plate Tectonics". Altered plutonic ultramafic fragments have recently been discovered at sites of ocean floor spreading e.g. within the Gulf of Aden (Gass, 1970a) and along the Mid-Atlantic and Mid-Indian Ridges (Miyashiro, 1972). Alternatively it is possible that the ultrabasic rocks were emplaced in leaking transform faults.

El Shazly and Saleeb (1972) claimed that the St. John's Island metavolcanics belong to the basic-ultrabasic Baramin Association on the Egyptian mainland but this "Association" is chemically quite different. The St. John's Island metabasalt has 1.12% TiO_2 , 6.55% total Fe oxide, 5.63% Na_2O , 0.07% P_2O_5 and 1.10% Cl (El Shazly and Saleeb 1972). These low iron, titanium and phosphorous contents are highly suggestive of oceanic basalt (Chase 1969), especially the titanium content because Cann (1970) found that titanium is relatively unaffected by alteration and metamorphism in oceanic basalts.

SUBAERIAL VOLCANIC ROCKS

All active subaerial volcanoes in the Red Sea are in the south. They are essentially in three groups viz. The Hanish - Zukur Group ($13^{\circ}43' \text{N}$, $42^{\circ}45' \text{E}$ to $14^{\circ}00' \text{N}$, $42^{\circ}06' \text{E}$), the Zubayr Group ($15^{\circ}00' \text{N}$ to $15^{\circ}12' \text{N}$, $42^{\circ}07' \text{E}$) and the Tayr Group ($15^{\circ}42' \text{N}$, $41^{\circ}45' \text{E}$). These volcanoes are in the fumerolic stage and are all essentially basaltic (Richard and Neumann van Padang, 1957; Griveau, 1966).

The Hanish - Zukur Group consist of undersaturated alkali olivine basalt, with trachybasalt, trachyandesite and andesite at Hanish Island (Gass, 1970a). These volcanoes are thought to lie on a leaky transform fault (Gass, 1970b). The Zubayr Group are intermediate between tholeiite and alkali basalt with associated picrite basalt and trachybasalt (Gass et al. 1973). The Tayr Group are of oversaturated olivine tholeiite of oceanic affinity similar to the Aden Series of the Ethiopian Rift (Gass, 1970b, Chase, 1969). Chase pointed out that the similarities with the Aden Series are notable because Jabal Tayr lies along the extension of the Wonji fault belt of the Ethiopian Rift along which fissure basalts have recently erupted. A detailed study of the volcanic islands has recently been published by Gass, Mallick and Cox (1973).

DEEP SEA VOLCANIC ROCKS

Basalt sporadically occurs over most of the floor of the axial trough of the Red Sea from its southern extremity to at least 23°N (Chase, 1969; Backer and Schoell, 1972; Schneider and Wachendorf, 1973).

Details of its field occurrence are generally unknown but it is generally thought to be in the form of submarine flows (e.g. Chase, 1969) with associated pyroclastics or hyaloclastics. Pillow forms were reported to have been photographed in situ by members of the Chain cruise 1971 (Professor G.R. Davis pers. comm.). Pillow lavas would be expected in this environment where the lava is basaltic, the depth of water (generally greater than 1,800 m) suppresses explosive activity, and the floor consists of soft muds. The occurrence of glassy lava fragments together with fine-grained crystalline fragments, as is usually found, further suggests pillow structures. Davies and Tramontini (1970) have suggested that solution of evaporites by present day volcanism could account for the numerous hot salty holes in the Red Sea floor, and for its unique bathymetric profile.

Fragments of glassy and fine-grained crystalline basalt from the axial trough ($20^{\circ}01'\text{N}$, $38^{\circ}27'\text{E}$; depth 2,440 m) examined by the writer were collected by the Applied Geochemistry Group. The largest of the glassy fragments has a ropy skin. The remaining angular fragments are mostly vesicular porphyritic black glass with vesicles widely spaced and ranging in diameter from 0.1 to 1.0 mm. The vesicles are spherical but the glassy rock also contains peculiar crack-like cavities giving a polygonal aspect to the fragment surfaces. These cavities are about the same width as the larger vesicles and have parallel sides which terminate downwards not as wedges but as hemi-cylinders. They are smooth sided and rarely intersect vesicles. They are possibly related to vesiculation, hyaloclastic brecciation and the depth of water. They were not observed in the phyrlic basalt which is much less vesicular and less porphyritic.

The glassy basalt is light brown in thin section and has an RI of 1.606 ± 0.003 (Chase, 1969, found 1.608 ± 0.002). It contains isolated euhedral phenocrysts of olivine (colourless even in handspecimen) and euhedral phenocrysts of plagioclase fringed by a dark colouration most probably caused by sub-microscopic pyroxene needles (figs 7 and 8).



Fig. 7. TLM photograph of plagioclase phenocrysts fringed by pyroxene(?) in brown basalt glass from Red Sea deeps (TS in PPL).

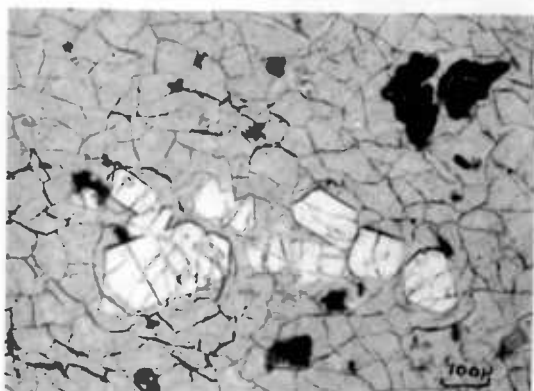


Fig. 8. TLM photograph of olivine phenocrysts in brown basalt glass from same specimen as in fig. 7 (TS in PPL).

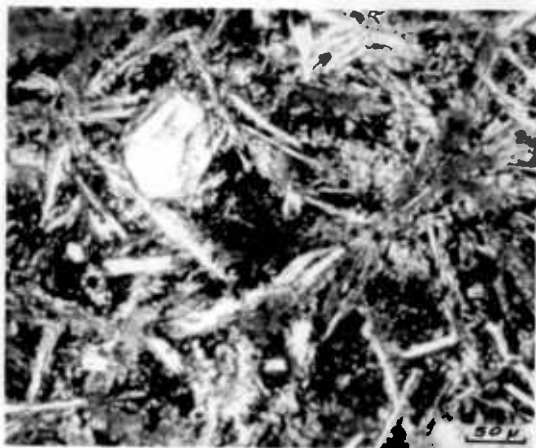


Fig. 9. TLM photograph of phyric-basalt from Red Sea deeps showing olivine (large white grain) plagioclase (white laths), pyroxene (? feathery material) and opaque minerals (TS in PPL.)

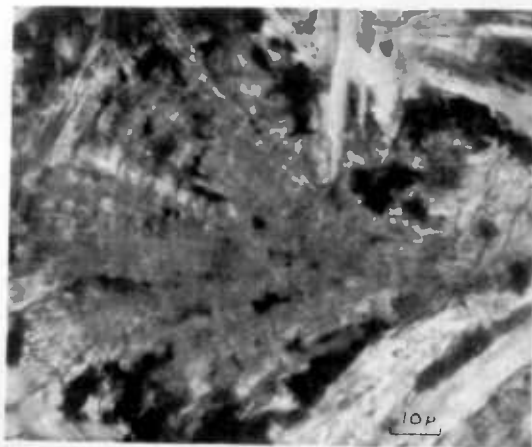


Fig. 10. Enlargement of centre part of fig. 9 showing probable pyroxene and euhedral opaque magnetite(?) (TS in PPL).

The phyric basalt is holocrystalline and very fine grained (fig 9). It consists of a felted mass of plagioclase laths (labradorite to bytownite), feathery clino-pyroxene(?) with a quenched skeletal texture (fig 10) ophitic to plagioclase, euhedral olivine phenocrysts, and very fine euhedral and skeletal opaque titan-magnetite(?). Schneider and Wachendorf (1973) found that the whole axial trough of the Red Sea appears to be floored with olivine tholeiite in which they found no recognisable modal pyroxene.

The chemistry of the basalt is discussed in section 1.3.6 where it is compared with the basalt plateaux of the shield area. It can be noted here however that the glassy basalt is olivine tholeiite (or sub-alkaline basalt) and the phyric basalt is alkali basalt (as are those of the shield and Jabal Zukur Island) as it is nepheline normative. The basalt fragments have similar composition apart from Fe, P and Ti to those studied by Chase (1969) and resemble oceanic basalts in their low potassium, titanium, phosphorus contents.

Trace element analysis gave surprising results (discussed in section 1.3.6). The phyric basalt was found to be abnormally rich in Cu and Zn but the glassy basalt has only average concentrations of these elements. No other trends are particularly evident in these results except that both kinds of basalt have consistently lower concentrations of Ba and Sr than the shield basalts. This is consistent with their more tholeiitic character and with the conclusion of Schilling (1969) who found that the rare earth contents of the Red Sea basalts are characteristic of submarine tholeiites from oceanic ridges and are different to continental tholeiites.

An effort was made to locate the Cu and Zn in the thin section of the phyric basalt but the grain size is too fine. Some tiny unidentifiable equant specks of a colour and reflectivity similar to that of pyrite were noted in association with the opaques but no definite sphalerite or chalcopyrite were found. Using special ultra thin doubly polished sections Bryan (1972) found that sulphides, especially pyrite, are common in the Red Sea lavas and that the sulphides are associated with magnetite which also is abundant.

Both the glassy and the phyric basalts are fresh however some yellow aphanitic material was adhering to them. This was removed and was found by XRD to contain quartz, chlorite and feldspar. A probable foram was also found in it hence it is probably ordinary Red Sea sediment and not altered basalt like the red material found by Chase (1969).

1.1.6

THE BASEMENT OF THE RED SEA

It is curious that the nature of the basement of the main Red Sea trough is controversial while that of the smaller, remote, axial trough is consensual.

Drill hole information together with seismic, magnetic, and thermal evidence indicates that the axial zone of the Red Sea overlies oceanic basalt, and is a zone of active seafloor spreading (e.g. Falcon et al 1970; Degens and Ross 1969). The only writers who disagree that the basement of the axial trough is oceanic basalt are El Shazly and Saleeb (1972) who stated, "The earth's crust under the Red Sea axial trough in particular (sic) is visualized as being constituted of a basement of metasomatic alkali-rich granitic rocks developed in the Cambrian at the expense of the pre-existing Precambrian igneous-metamorphic rocks and containing large xenoliths of the latter"!

Opinion differs on whether the deep basement of the main trough is oceanic or continental. This problem was discussed, but not fully resolved, at the recent Royal Society meeting on the structure and evolution of the Red Sea (Falcon et al. 1970). The continental crust around the margins of the Red Sea, has been stretched where horst and graben structures are now well developed. This however is probably confined mostly to the coastal region otherwise a new hypothesis would be necessary to explain the occurrence of matching disjuncts either side of the sea. Other evidence supporting the oceanic basement hypothesis is the large positive Bouger gravity anomaly over the Red Sea (Girdler, 1969) and the high velocities of seismic waves within the basement (Davies and Tramontini, 1970). The main argument against this hypothesis is the paucity of characteristic magnetic anomalies outside the axial trough. Such paucity however can be explained easily if the basalt is at a greater depth than expected, or if the basalt has been altered, or if sea floor spreading was very slow. The magnetic model itself is far from perfect. JOIDES drill hole 228, sited over a magnetic anomaly thought to be 2.2 million years old, bottomed in Miocene evaporites without intersecting any igneous rock (Ross et al, 1972).

The following generalized section (Table 4) derived from Davies and Tramontini (1970) is the most consistent and complete so far available.

Average thickness	Rock Type	Seismic velocity
0.2 km (0.33 km)*	unconsolidated sediments	2.0 km/sec
2 to 5 km	evaporites	4.3 km/sec
< 1 km	basic lavas	5.1 km/sec
(depth > 4.6 km)	oceanic basement	6.63 km/sec

* This figure from Phillips and Ross (1970) is probably more reliable.

Table 4 Generalized section through the main trough of the Red Sea (derived from Davies and Tramontini, 1970).

The formation of the Red Sea probably began in the early Tertiary (Cretaceous to Oligocene - e.g. Dubertret, 1970; Lowell and Genik, 1972) although it may have begun in the Triassic (Holwerda and Hutchinson, 1968). The formation has been discontinuous; the present stage is about 3 million years old (Phillips and Ross, 1970; Allen, 1970). The initial stages may have been somewhat similar to the development of the Danakil Depression in Ethiopia. This depression is between two continental blocks and appears to be underlain by new oceanic crust (Holwerda and Hutchinson, 1968). Filling the depression is a sequence about 1000 m deep of Quaternary shallow-water evaporites invaded by basalt.

Almost nothing is known about the basic lavas (postulated on seismic evidence) of Davies and Tramontini in table 4 however oil-well B1 situated well out in the southern part of the Red Sea terminated in Oligocene basalt (Lowell and Genik, 1972). Lowell and Genik also recorded Oligocene pillow lavas in the Dogali Series of Ethiopia but added that pillow lavas are not necessarily marine.

The evaporites are continuous over large areas of the sea beyond the axial trough and form a prominent seismic reflector (Phillips and Ross, 1970). The sequence is generally slightly folded and faulted but is tightly folded in some areas. JOIDES drill holes have confirmed that the sequence consists of anhydrite and halite dominantly with lesser pyritic clay and silt (Ross et al, 1972). The evaporite sequence is of late Miocene age and at least some of it is of shallow water sabkha origin (D. Shearman, pers. comm. 1972). It was probably deposited in an interval of 10 to 15 million years when the sea was open to the Mediterranean and closed to the Gulf of Aden (Davies and Tramontini, 1970).

The overlying, largely unconsolidated, sediments are more flat-lying than the evaporites from which they are separated by a probable disconformity and, in some places, an angular unconformity (Phillips and Ross, 1970). The thickness of this sequence of sediments is fairly uniform, averaging about 330 m, but it decreases toward the axial trough. The JOIDES holes showed these sediments to consist of pelagic carbonate ooze with admixed clay and silt (Ross et al, 1972). They are essentially the same as the common modern Red Sea sediments already described. A disconformity occurs between the late Pliocene and late Pleistocene in the sediments near Atlantis II Deep.

Pyrite was found in lower Pliocene argillites above Miocene evaporites in the three JOIDES holes that penetrated to the evaporites. The pyrite is associated with high Cu, Zn and V values, with Zn up to 5%. These core intervals are 25 m thick (Ross et al, 1972). Sulphide concentrations are probably widespread in the Red Sea sediments. Sulphides are found on the present floor and in older rocks near the coast. In the north pyritic carbonaceous argillite associated with evaporites was found in Barqan No. 1 well at 2,900 m (private communication), and in the south, slightly metamorphosed very pyritic argillite of early Oligocene age was encountered in Amber well at 3,550 m (Lowell and Genik, 1972).

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1.2.1

COASTAL PLAIN TOPOGRAPHY

Except in the Ethiopian sector, a narrow coastal plain, less than 1 to more than 30 km wide, fringes the Red Sea. The boundary on the continental side of the plain is marked by steep hills and mountains up to 3,700 m high in Yemen (Gillmann, 1968). This scarp also marks the general boundary between the Precambrian crystalline rocks and the post-Palaeozoic sedimentary rocks.

According to Nesteroff (1955) the Red Sea contains at least five different types of reefs and most of the coastline is bordered with fringing reefs. Nesteroff attributed the formation of beachrock in the Red Sea region to bacterial activity. The sand is veiled with an organic web packed with bacteria which produce an amorphous limestone cement.

An almost continuous band of emergent reef terraces 0.5 to 10 km wide and 3 m high lies at the waters' edge from Yemen to Al Wajh. These terraces are intermixed with raised beaches, dunes and gypsiferous gravels. Similar deposits occur along the Egyptian coast. Between these terraces and the boundary scarp is a pediment of alluvium and outwash sands and gravels. Between the raised terraces are supra-tidal sabkhas. Many drowned estuaries or "sharms" are also found along the coastal plain.

From the Gulf of Aqaba to Umm Lajj are terrace surfaces at heights of 6, 10, 20 and 30 m mostly on coralline rocks. From Umm Lajj to Jeddah are remnants of a terrace 20 m high, and south from Jeddah to Yemen the coastal plain is quite flat except for the 50 m high Jizan salt dome and two small volcanoes between Hali and Ash Shuqay (Skipwith, in press). Said (1969) recorded wave cut terraces in Egypt and the Sudan at 1 to 3 m, 7 m and 12 m above sea level. Terraces in the north, as high as 520 m asl. have been recorded by Brown (1970a).

The Ethiopian part of the Red Sea coastal region is dominated by the Afar or Danakil Depression which is essentially a subaerial evaporite basin partly below sea level. It is bounded by Precambrian basement to the south and west, and the Danakil and Aisha horsts in the east and north-east. The Depression is characterized by great flat salt pans,

extensive lavas and many fault scarps (Holwerda and Hutchinson, 1968). Although the scarp separating the Precambrian basement from the post-Palaeozoic sediments is related to major fault zones it is essentially an erosional feature in the areas in which it is best developed. This applies to both sides of the Red Sea, especially in the southern region where topographic relief is greater (Brown and Coleman, 1972; Whiteman 1971).

North of Jeddah the first major constriction in the coastal plain is in the Ras Jarbu - Wadi al Wazra area. The first extensive outcrops of Tertiary sediments are also situated here. Beyond this the next extensive part of the coastal plain is in a triangular region about 15 km wide and 150 km long essentially bounded by an easterly fault scarp near Jabal Yahmum and a NNW trending fault zone east of Jabal Dhaylan. This area also includes the great Wadi al Hamd and its extensive alluvial fan. North beyond this is the beautiful rift valley of Wadi as Sirr which is about 10 km wide and 100 km long. This virtually unknown rift valley trends NW and meets the coast near Ras Abu Masarib. It is probably part of the great Najd fault system. The only other extensive plain area beyond Ras Abu Masarib is the large triangular alluvial fan of Wadi Ifal in the Aqaba region.

1.2.2

GEOLOGY OF THE COASTAL PLAIN

INTRODUCTION

Until vital reports on the geology of the coastal plain (e.g. Croisile, 1967) are made available compilations must remain tentative. Nevertheless much information is available and some of this has already been collated by Skipwith (in press).

PALEOZOIC ERA

Very little is known of the Paleozoic geology of the Red Sea region due to a scarcity of outcrops. Brown (1970a) however found Cambro-Ordovician Wajid Sandstone in the outliers of Jabal Abu Hasan and Jabal Soda in the south about 75 km from the sea. He further noted that Nubian type sandstone probably covered much of the shield before the Permian.

At Tiran Island in the north of the Red Sea are rocks which can be correlated with the Devonian Tawil sandstone member of the Tabuk Formation (Bramkamp et al. 1963; and Said, 1969). In the Gulf of Suez are Carboniferous sandstones and shales of the Rod el Hamal and Um Bogma Formations, marine Permo-Carboniferous rocks of the Aheimer Formation, and Permo-Triassic red beds of the Qusaib Formation (Said, 1969).

MESOZOIC ERA

The general lack of outcrops of Mesozoic rocks appears to be the reason for Said's (1969) and Brown's (1970a) conclusion that uplift and non-deposition dominated this era.

TRIASSIC-JURASSIC The oldest Mesozoic rocks are in the south near Jizan. The Khums Formation of Gillmann (1968) is probably Triassic to Lower Jurassic. It consists essentially of argillaceous and siliceous sandstones, with a maximum thickness of 880 m near the Yemen border. Phosphatic inclusions were found in its middle feldspathic sandstone member. It rests with an angular unconformity on the Precambrian basement and has a general dip of 35°W. Overlying this formation with no apparent unconformity is the Upper Jurassic Hanifa Formation (Gillmann, 1968)

which consists mostly of limestones and which has a maximum thickness of 720 m. Intrusions are abundant in both the Khums and the Hanifa Formations (see section 1.3.4).

Jelenc (1966a) recorded the occurrence in Ethiopia of clastic rocks, limestone and gypsum deposited during a marine transgression in the Jurassic. This sequence is overlain by sandstone deposited during a marine regression. In an oil well at Maghersum Island off Sudan, Pre-Cretaceous basaltic breccia is overlain unconformably by the Mukawar Formation (Sestini, 1965; Said, 1969; Lowell and Genik, 1972). Middle and Upper Jurassic clastic sediments are also reported to occur along the Gulf of Suez (Said, 1969).

CRETACEOUS Overlying the Hanifa Formation in the Jizan area, but with a hidden contact, is the Lower Cretaceous Taouilla Formation (Gillmann, 1968) which consists of sandstone, limestone and shale; its maximum thickness at outcrop is 10 m.

The Mukawar Formation in Sudan consists of Upper Cretaceous to Lower Paleocene shales and sandstones with minor shallow water marine limestones and marls (Sestini, 1965; Said, 1969; Lowell and Genik, 1972).

In the Suez area are continental sandstones and shales of the Lower Cretaceous Malha Formation. Sestini (1965) however noted that a marine transgression from the north during the Lower Cretaceous reached as far south as Quseir. Assigned to the Upper Cretaceous is a thick sequence of marls and shales with lesser limestones, dolomites and red beds (Galala, Raha, Wata and Matulla Formations, Said, 1969). Cretaceous sandstone, marls and shales with phosphatic "occurrences" were observed further south along the Egyptian coast (Herbst et al 1965). Said (1969) also noted Cretaceous marine limestones near Quseir.

CENOZOIC ERA

PALEOCENE Uppermost Cretaceous to Lowermost Paleocene (Maestrichtian) marine marls, limestones (partly phosphatic) and lesser shales and sandstones were noted by Karpoff (1957) at Usfan, 55 km NE of Jeddah. No other data is available on the Paleocene from the eastern coast of the

Red Sea. Paleocene to Lower Oligocene sandstones, arkoses and basalt of the Hamamit Formation however were found in the oil well at Maghersum Island off Sudan (Sestini, 1965). This formation unconformably overlies the Upper Cretaceous to Lower Paleocene Mukawar Formation described above. Sestini (1965) considered that during the Paleocene, Eocene and Oligocene conformable clastic sediments were deposited in terrestrial, lacustrine and littoral marine environments with possibly more marine environments in the centre of the basin. Said (1969) records that thick sequences of Paleocene and Eocene rocks occur in the Gulf of Suez but no details are given.

EOCENE Eocene rocks are rare in the Red Sea coastal region. Skipwith (in press) noted that the Khurma Formation, 100 km SE of Jeddah consists of two small outcrops of Eocene(?) sandstone and red shale with numerous mollusc borings. The Usfan Formation near Jeddah consists of Marine and littoral Eocene sediments of marly limestone, sandstone and conglomerate. Brown (1970a) noted Eocene fossils in boulder beds 110 km inland from Al Wajh and at an elevation of 1,100 m.

OLIGOCENE Although the Oligocene was an epoch of great volcanic activity there are few sedimentary formations of this age in the coastal region. The best known formations are the Shumaysi Formation of Wadi Fatima between Jeddah and Mecca, and its equivalent, the Hamamit Formation found in the bore hole at Maghersum Island, Sudan.

The Shumaysi Formation, described by Al-Shanti (1966), consists mostly of marine and non-marine sandstones and shales with iron ore beds (described in section 1.2.4).

The Hamamit Formation conformably overlies the Mukawar Formation and consists of red quartzitic to arkosic sandstones and basalt (Sestini, 1965; Said, 1969). According to Heybroek (1965) there is an equivalent formation in the Suez region. Similar rocks of Oligocene age are also found at Jabal Dhaylan, 500 km north of Jeddah (see section 1.2.3).

According to Sestini (1965) the Oligocene was an epoch of uplift and terrestrial sedimentation.

MIOCENE The Miocene is well represented by a thick sequence of evaporites throughout the whole Red Sea region. Great facies variations characterize the Red Sea formations and detailed local sections are difficult to correlate, nevertheless broad correlations can be made from one end of the Red Sea to the other as shown in fig. 11, from Gillmann

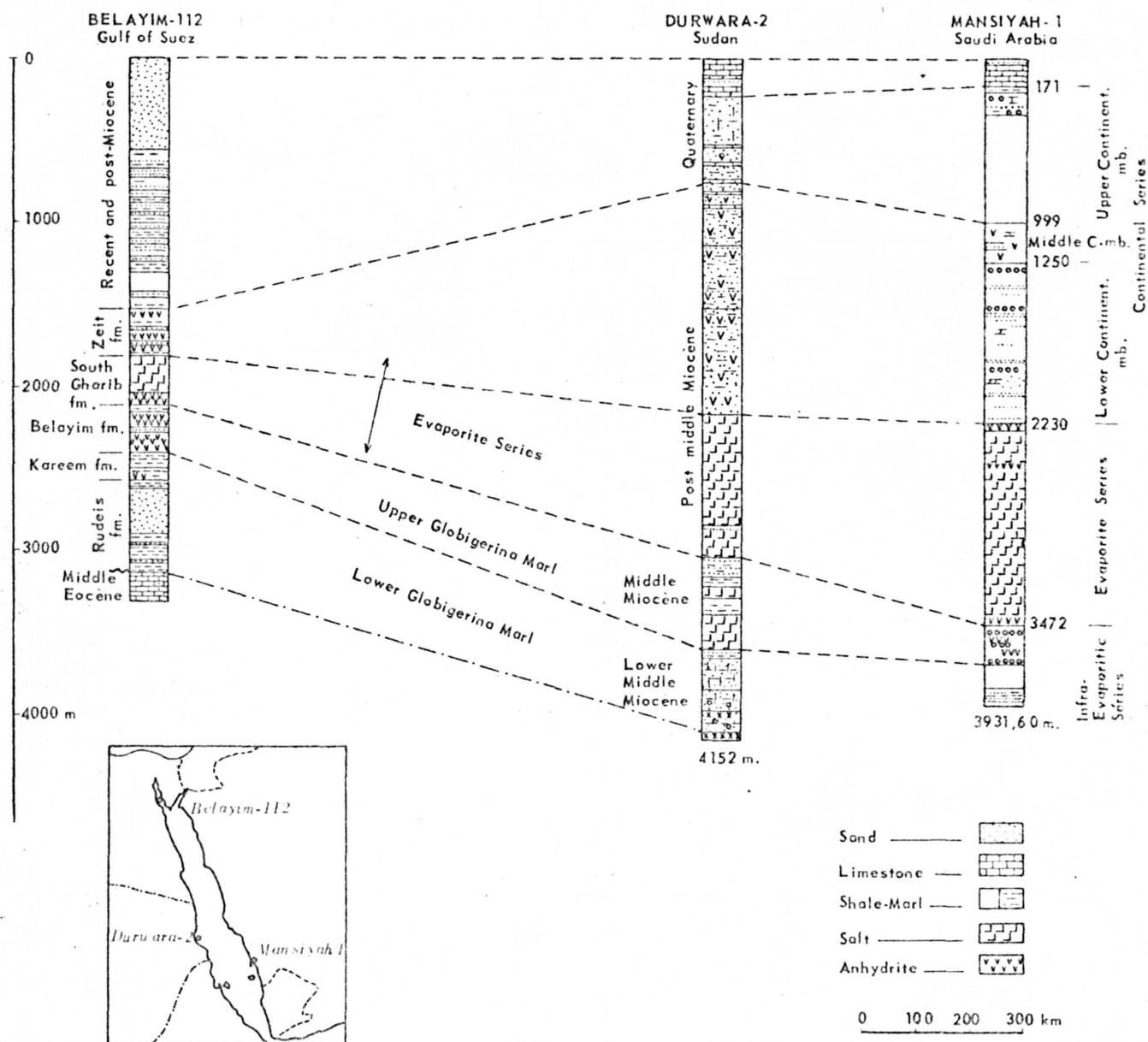


Fig. 11. Probable stratigraphic relationship in the subsurface Tertiary of the Red Sea gaben (from Gillmann, 1968).

(1968). A similar correlation has been made by Ahmed (1972). Ahmed also noted that seismic results showed that the evaporites and the unconformity below them extends throughout the Red Sea coastal region. As the evaporites from the coast and from near the centre of the Red Sea (Section 1.1.5) are of shallow water origin it appears that the entire Red Sea region was a lagoonal type of environment at this time. In the Lower Miocene it was apparently connected with the Mediterranean Sea (Ahmed 1972) but in the Upper Miocene it was isolated from both the Mediterranean Sea and the Gulf of Aden (Heybroek, 1965).

In the south-east part of the coastal region the Baid Formation rests unconformably on the Upper Jurassic Hanifa Formation near Jizan. It consists of shales, "silicite" with fresh water fossils, tuff, breccia and basic to acid lavas (Gillmann, 1968). It is densely intruded by andesite and quartz porphyry sills. The Baid Formation appears to be equivalent to the cap rock of the Middle to Upper Miocene Jizan salt plug hence it might in part be equivalent to the Lower Continental Series of the Mansiyah No. 1 bore hole 40 to 65 km away. The general stratigraphy of Mansiyah No. 1 borehole is illustrated in fig. 11. Evaporites of the Evaporite Series can be correlated with evaporites from the Farasan Islands (Saudi Arabia), the Dhalac Islands (Ethiopia), Durwara and Dungunab wells (Sudan), salt domes in Yemen, and oil wells in the Gulf of Suez.

Miocene outcrops are relatively rare between Jizan and Jeddah however evaporites have been penetrated at shallow depths in wells near Jeddah (Skipwith, in press) and Brown and Jackson (1958) mapped the Baid Formation from Jizan to approximately 20°N. Skipwith also reported that the Ghawas well near Al Lith penetrated 670 m of Plio-Pleistocene limestone, 2000 m of Continental Series sandstone and dolomite, and 1300 m of Miocene halite (cf. Mansiyah No. 1 well fig. 11). At 21 km east of Rabigh Richter-Bernberg and Schott (1954) found 107 m of Miocene tuff.

From Rabigh north to the Gulf of Aqaba generally poor outcrops of Miocene rocks are relatively common but in many places they are not readily distinguished from Pliocene rocks. Miocene outcrops however are more extensive between Al Wajh and Duba than is indicated on map I-204A by Brown et al. (1963). In this northern section of the coastline the

Miocene outcrops consist of limestone, marl, shale and sandstone. Gypsum is extensive in some areas but such areas are generally thought to be Quaternary. The rarity of halite outcrops could be partly due to their solubility. Beneath the surface of soft friable shale outcrops salt is commonly found. It also occurs as veinlets in some limestones. Tuffs also are probably more common than present literature suggests. Good exposures of basic tuffs were observed in Wadi Umm Lajj. Some of the "shales" elsewhere are probably reworked fine grained tuff.

The best Miocene outcrops from the eastern coast of the Red Sea are of the Raghama Formation in the Aqaba area (fig. 12). This formation is over 500 m thick and comprises three unconformable units. The lower unit consists of Oligocene to Lower Miocene conglomerates, sandstones and limestone resting unconformably on the Precambrian basement. The middle unit is of Middle to Upper Miocene marls, sandstones, shales and gypsum. The upper unit is of Upper Miocene to Lower Pliocene sandstones, sandy reef limestones and gypsum. The off-shore Barqan wells 1 and 2 in this area intersected a thick sequence of gypsum beds, about 200 m of Miocene halite and Globigerina marls similar to those in the Gulf of Suez (Skipwith, in press).

The Egyptian coastal plain has Middle Miocene basal lime grits, resting on the Precambrian basement, followed by Middle Miocene evaporites and the Samk Formation of marine marls (Said, 1969).

In the Sudanese coastal region the Middle Miocene Maghersum Formation unconformably overlies the Hamamit Formation. The Maghersum Formation consists mainly of marls and evaporites in one area and sandstones with minor calcarenites, reef limestones and basalt lavas in another area (Sestini 1965). It is overlain by the Khor Eit Formation of calcarenites with minor reefs and evaporites. It in turn is overlain by the Abu Imama Formation of limestones and the Upper Miocene Dungunab Formation of sandstones and evaporites (Sestini, 1965). Sestini noted that the Miocene clastics generally were not transported far. He further noted that the sedimentation pattern progressed from clastic to carbonate to evaporitic.

PLIOCENE Pliocene limestones, sandstones and marls are not easy to distinguish from Miocene lithological equivalents. The oyster Ostrea crassissima has been used as a stratigraphical marker but Brown (1970a)



Fig. 12. Outcrop of Raghama Formation
north of Al Bad, Maqna area.



Fig. 15. Outcrop in Wadi Al Hamd area
with small graben structures.



Fig. 14. Outcrop of Tertiary sediments
near Raas Duba showing position of salt
veinlets (behind the figure), fault and
tilted surface.

quoting Cox, pointed out that this fossil is found in both the Miocene and the Pliocene. Indeed, it seems that during the Pliocene the Red Sea was connected with the Indian Ocean where this oyster flourished, and was isolated from the Mediterranean. The writer has found midden-like heaps of these shells at various localities along the north-west coast of Saudi Arabia.

Pliocene rocks are not easily distinguished from Pleistocene rocks and some formations have been designated as Plio-Pleistocene e.g. at Jabal Dhaylan (Dadet et al. 1970) and in the Sudan (Sestini, 1965). Generally however it appears that the Pliocene was not an epoch of great deposition.

QUATERNARY The Pleistocene epoch began with mostly sandstone and shale deposition followed by limestone and conglomerate. Typical outcrops of younger Quaternary deposits consist of coral limestones, lag-gravels, aeolian sands, sabkha deposits, clay pans, gypsum beds, alluvial fans and wadi alluvia, and basalt. In some places algal mats have developed along the coast.

STRUCTURE

The Neogene formations of the coastal plain are generally flat lying with slight regional dips toward the sea. Horst and graben structures are widespread throughout the whole region on scales ranging from the full width of the Red Sea to that shown in fig. 13.

The main peripheral fault zone near the great escarpment of Saudi Arabia dips steeply to the west and is 1 to 2 km wide. It is pre-Miocene and has a displacement of 3 km (Brown, 1970a). To the south, on both sides of the Red Sea, it is associated with monoclinial flexures (Whiteman 1971; Brown 1970a). From the escarpment to the edge of the shelf is a series of step faults, horsts and grabens. Grabens have formed deep basins in which as much as 6 km of sediment have accumulated, (according to geophysical evidence, Skipwith, in press). Large faults are common in the northern part of the Arabian coastal plain. They generally strike north-west, and Brown (1970a) has suggested that they are related to the great Najd wrench fault system to the south-east. Large faults in the

central coastal plain commonly have a north-east strike, and those in the south have northerly and north-north-westerly strikes (Brown, 1970a; Brown and Hase, 1971, maps). North-north-west faults parallel to the trend of the sea are also common in the coastal plain. Magnetic lineaments due to dykes and altered fracture zones also have this trend (Brown and Hase, 1971). Some north-west faults are displaced by north-north-west faults. Many faults are covered by wadi alluvium.

Salt movement has produced intensely deformed zones in the southern part of the coastal plain. It is possible that salt movement has produced structural disturbances elsewhere also. Fig. 14 shows a tilted Quaternary surface on a faulted block of marly sandstone. A concentration of centimeter wide halite veinlets was found in the central part of this block (just above the Arab field assistant). In areas of gypsum outcrops bedding is chaotic, presumably due to the volume change associated with the transformation of anhydrite to gypsum.

SECTION 2

THE COASTAL REGION

1.2.3

JABAL DHAYLAN

GEOGRAPHY AND EXPLORATION HISTORY

STRATIGRAPHY

Basement
Oligocene
Miocene
Plio-Pleistocene
Quaternary

STRUCTURE

LEAD-ZINC MINERALIZATION IN LIMESTONE

Introduction
Prospect JJM2
Prospect JJM3
Prospect JJM4
Prospect JJM1
Prospect JJM10

LEAD-ZINC MINERALIZATION IN OLIGOCENE ARKOSE

COPPER MINERALIZATION

BARIUM MINERALIZATION

MANGANESE MINERALIZATION

PROSPECTING METHODS

GEOGRAPHY AND EXPLORATION HISTORY

Jabal Dhaylan is a chain of hills dominating the coastal plain between $25^{\circ}30'N$, $25^{\circ}36'N$ and $37^{\circ}02'E$, $37^{\circ}12'E$. The highest point on the jabal is 150 m above sea level. Jabal Dhaylan is approximately 500 km by plain from Jeddah and is 55 km NNW of Umm Lajj, and 15 km from the sea. It is a lonely place, short of water and almost unpopulated. Much of the landscape is covered with aeolian sand which freely moves in the windy season (figs. 15 and 16).

Jabal Dhaylan contains the largest known concentration of lead and zinc in the coastal plain. Bigot (1970) estimated the approximate tonnage of ore at Prospect JJM5 to be 300,000 tons assaying 10% combined lead and zinc. Most of this ore however is secondary.



Fig. 15. Aerial view of Jabal Dhaylan in winter, viewed from the south.



Fig. 16. Fault trace in Plio-Pleistocene limestone between JJM10 and Wadi Ank, viewed from the east.

Although perhaps a score or more of geologists have visited Jabal Dhaylan in the last decade, ancient workings (with the exception of a hole found by P. Chaumont) were unknown in this area. The writer found ancient workings at three localities and some remains of ancient dwellings at JJM3. At a very small excavation made at JJM3 the writer found some galena and artificial litharge, as well as a tiny piece of lead and a tiny piece of copper (possibly native copper) and some fragments of glass. The slag near the old workings contains traces of free lead. The ancient workings are primitive adits of irregular shape. Most probably they are much more than a millenium old as they lack the style and technique of the ancient mines that flourished in Saudi Arabia about 1000 years ago. In fact these mines at Jabal Dhaylan could be some of the oldest mines ever found. Even though the silver content of the ore is low it is possible they were worked for silver as were similar lead deposits at Laurion, Greece, which were worked at about 1000 B.C. It is impossible to estimate how much ore was removed from Jabal Dhaylan as the old workings have partially collapsed and have been partially or totally filled with aeolian sand. Less than 10 tons of slag is now visible in the sand at Jabal Dhaylan.

The rediscovery of mineralization at Jabal Dhaylan began in 1968 when Y. Berton of BRGM found traces of copper mineralization at Prospect JJM6 and near Prospect CJ1. In a follow-up study by J.J. Marchessau and J. Costans the lead-zinc mineralization was found. J. Bouladon (1968) claimed that until this time lead-zinc mineralization in Miocene rocks of the coastal region was "absolutely unknown".

In 1963, Brown et al assigned the sediments at Jabal Dhaylan to the Miocene Raghama Formation on their map I-204A at 1:500,000 scale (1963). In 1966 Al Shanti and Sultan mapped the area in relation to its petroleum potential. From 1968 to 1971 following the work of Y. Berton, J. Marchessau and J. Costans the area was mapped and prospected by geologists M. Bigot, P. Chaumont, C. Lafoy, P. Laloux and P. Pilet under the supervision of J. Bouldon, P. Dadet and J. Wolff. This work resulted in a general map at 1:50,000 scale, maps at 1:2,000 scale covering 1.4 km², and maps at 1:200 scale on individual prospects. The trenches cut have an aggregate length of 1,152 m, and 19 drill holes sunk at JJM3 have an aggregate length of 908 m (Chaumont, 1970a, Bigot, 1970). In addition to this Chaumont (1970b) collected 462 samples in a geochemical survey over 635 km². The general geology is illustrated in figs. 17 and 18.

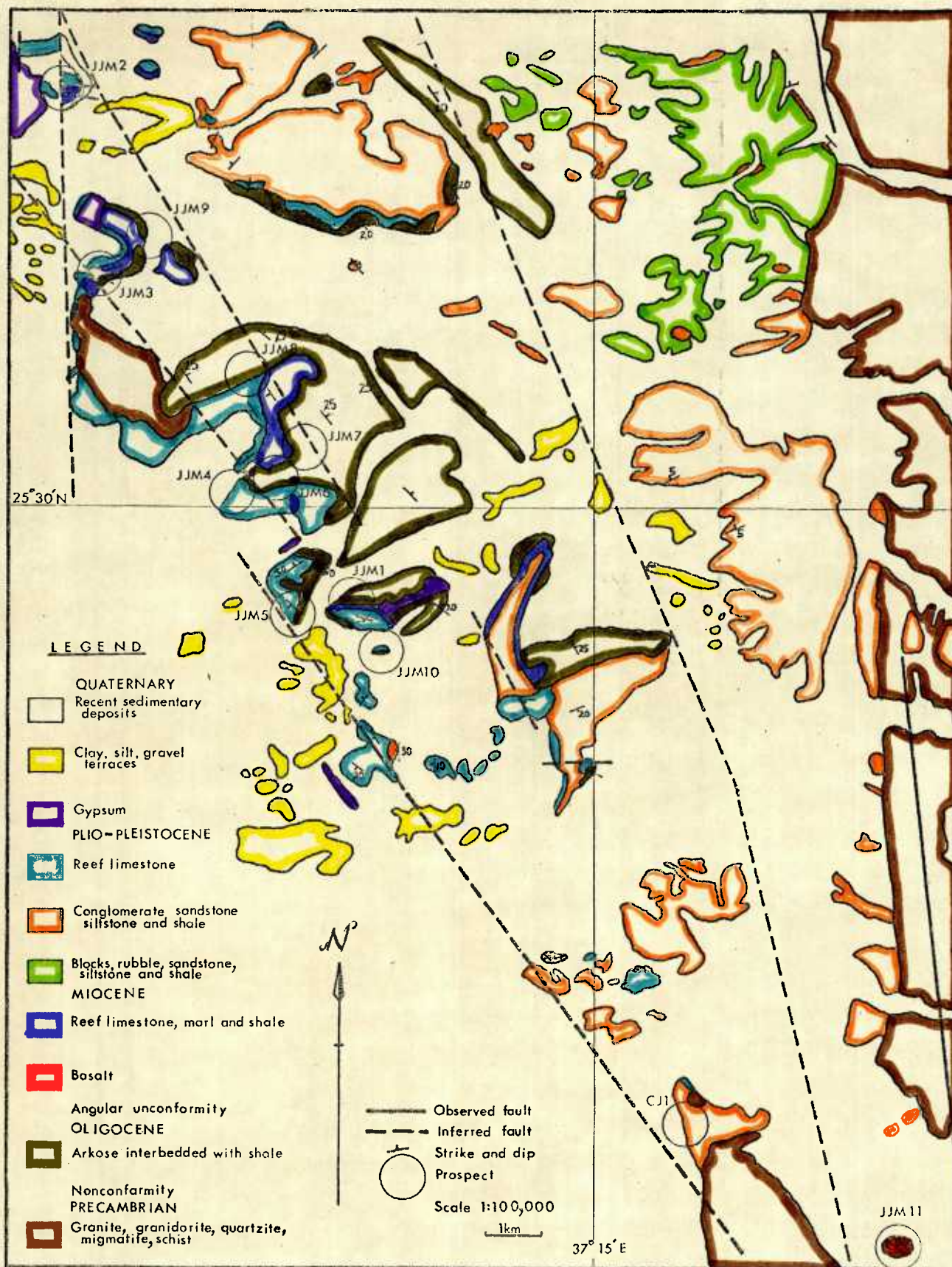


FIG. 17. GEOLOGICAL MAP OF JABAL DHAYLAN
from P. CHAUMONT (1970a)

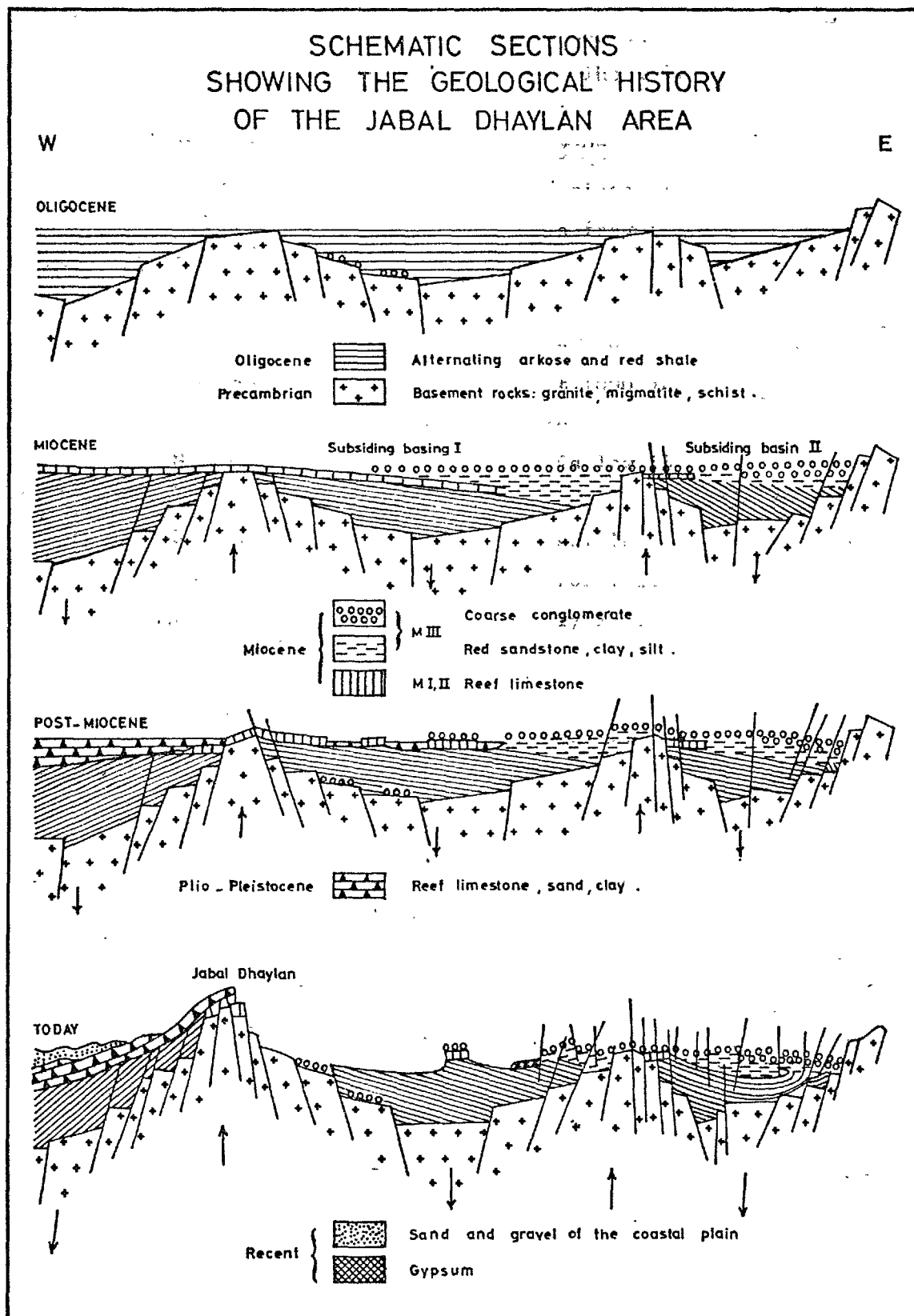


Fig. 18. Schematic sections showing the geologic history of the Jabal Dhaylan area. Directly from Dadet (1970).

STRATIGRAPHY

Basement

The basement of the succession at Jabal Dhaylan is a Precambrian complex of schists and granitic and dioritic rocks broken into a series of NNW trending fault blocks (Dadet, 1970).

Oligocene

Resting unconformably on the basement are pre-Miocene beds of friable brown arkose and red to yellow to green shale having a regional dip of 20° ENE. Although this formation has not been dated it has tentatively been assigned to the Oligocene by all those who have studied it (e.g. Dadet, et al. 1970, Chaumont, 1970a, Bigot, 1970). It is at least 36 m thick and is highly lenticular (Bigot, 1970). Locally it is conglomeratic at the base and at some higher levels.

Miocene

Approximately horizontal transgressive Miocene beds unconformably overlie the Oligocene beds and the Precambrian basement. Marchessau (1969) divided the Miocene strata into three parts. Miocene I ("Lower Miocene") consists of shale with gypsum veinlets, fossiliferous limestone and sporadic conglomerate, shale and sandstones. Miocene II consists of conglomerate and sandstone, and Miocene III ("Upper Miocene") consists of shale, sandstone and conglomerate. Chaumont (1970a) however considered the Miocene III division to be probably of Plio-Pleistocene age.

Bigot (1970) divided the Miocene into four main facies viz. -

1. Calcareous facies with corals which consist of reef and fore-reef deposits represented by reefoid, shelly and sandy limestones and breccia.
2. Argillaceous facies alternating with marl commonly spotted with manganese oxides. This indicates quiet deposition as in a lagoon.
3. Various detrital facies consisting of shale, and breccia of limestone and shale. These contain traces of gypsum and probably formed along the shoreline. Behind the shoreline a thick sequence of marl followed by shale, sandstone and conglomerate developed (Dadet et al 1970).

The reefoid limestones are up to 40 m thick, and the marls are 20 m thick (Dadet et al 1970).

4. Drilling at JJM3 revealed another previously unknown section of Miocene sediments. These are gypsum, anhydrite, argillite, and dolomite in an evaporitic facies 75 to 138 m thick. A synsedimentary fault abruptly divides this facies from the other three.

The limestones are abundantly fossiliferous but fossils are rare in the other rocks. The writer however found poorly preserved plant fragments and skeletons of bony fish at one locality near drill hole DY10. The shale in this area is yellow-green and contains sparse veinlets of halite up to 6 mm wide. Shrinkage cracks show it obviously contains swelling clay. It was thought that this could be a bentonite or possibly a reworked fine ash but no certain evidence of this was found under the microscope. A salt free specimen was analysed with a batch of basalts. The method of analysis (XRF) is discussed in section 1.3.6.

It is obvious from table 5 that the shale is much closer in chemical composition to ordinary shale than to basalt. Nevertheless its normative composition is that of trachyte which could be derived by differentiation from a basaltic magma.

	Qalib Basalt	DY10 Shale	"Average Shale"
SiO ₂	46.95	52.64	58.10
TiO ₂	1.36	0.78	0.65
Al ₂ O ₃	17.13	16.47	15.40
Fe ₂ O ₃	2.01	6.78	4.02
FeO	7.52*	0.68*	2.45
MnO	0.16	0.35	n.d.
MgO	8.98	3.21*	2.44
CaO	10.11	2.09	3.11
Na ₂ O	3.05*	1.98	1.30
K ₂ O	0.70	3.00	3.24
H ₂ O+	2.05	5.45	{ 5.00
H ₂ O-	0.57	1.77	
P ₂ O ₅	0.26*	0.10*	0.17
CO ₂	n.d.	n.d.	2.63
Total	100.85	95.30	100.00

Table 5 Chemical analyses of basalt and shale from Jabal Dhaylan and "average shale" (Mason, 1958). The following results in ppm were obtained by XRF for trace elements of the shale Pb 11, Zn 1600, Cu 124, Ba 218, Ni 71, Co 52*, Cr 113, Y 35, Sr 171 and Rb 53.

* kindly determined by A.J. Thompson using AAS.

Plio-Pleistocene

Most probably both Pliocene and Pleistocene rocks occur at Jabal Dhaylan but in most places they are indistinguishable in the field.

Along the western side of the Jabal Dhaylan horst a Plio-Pleistocene reef limestone, locally detrital or shelly, and a horizon of shale unconformably overlie Miocene rocks. Along the eastern side of the horst are Plio-Pleistocene conglomerate, sandstone, siltstone, and shale. Further to the north-east are blocks and rubble, sandstone, siltstone and shale resting on the Precambrian basement (Chaumont, 1970a, Marchessau, 1969, Dadet et al 1970). At outcrop the reef limestone is generally compact and tough. It is also black and manganiferous in places.

Quaternary

Conformably overlying the Plio-Pleistocene beds are gypsum, gypsiferous clay, terraces of argillaceous sandstone and gravel, alluvial fans and wadi fills of sand and gravel, and drifting aeolian sands (fig. 19). The gypsum is presumably Pleistocene as Dadet et al (1970) included it with the Plio-Pleistocene sediments, and Marchessau (1969) and Chaumont (1970) included it with the Quaternary sediments.

The basalt at Harrat Qalib to the south overlies Plio-Pleistocene sediment but is younger than the dykes at JJM11 (discussed in section 1.3.5).

STRUCTURE

Dominating the structure at Jabal Dhaylan is the Precambrian horst block which formed the island on which the Miocene and Plio-Pleistocene reefs grew. The horst is bounded by NNW faults which are expressed as magnetic lineaments (Dadet et al 1970). The horst is also broken by NS, NE and WNW faults.

The Precambrian basement to the east contains NS and ENE faults clearly visible on air photos. Small faults and joints of general NS and EW trends are also found at Jabal Dhaylan itself. The larger faults are mostly obscured by wadi alluvium and aeolian sand (figs. 16 and 19). As the wadi alluvium is extensive and the aeolian sand has little residence-time one is unlikely to detect solution channel ways by surface sampling in these zones. The mineralized veins are small features related to small faults and joints where they are visible in outcrop.



Fig. 19. Part of the main Jabal Dhaylan horst viewed from NNW showing general topography, western part of JJM5, JJM8, aeolian sand, and site of ancient dwellings.

The Oligocene rocks are slightly folded and have a regional dip of 20° to 25° ENE. The Miocene and Plio-Pleistocene rock along the Jabal Dhaylan horst dip down hill toward the sea (roughly ESE) at inclinations down to 30° . Dips are much less inclined between the horst and the eastern basement outcrops (Chaumont, 1970a).

LEAD-ZINC MINERALIZATION IN LIMESTONE

Introduction

The most abundant mineralization at Jabal Dhaylan is the lead-zinc mineralization in the Miocene and Plio-Pleistocene limestones. Prospects JJM1, JJM2, JJM3 and JJM10 in Miocene limestone and JJM4 and JJM5 in Pliocene limestone were found in 1968 by J.J. Marchessau (1969).

These prospects tend to occur on the seaward side of Jabal Dhaylan and are within a NNW trending zone 1.8 km wide by 12 km long (fig. 17). JJM3, JJM4 and JJM5 are colinear as are JJM1, JJM2 and JJM10 but this may be fortuitous. JJM3, JJM1 and JJM10 are almost colinear and are coincident with the extension of a fault trace.

The prospects are similar in structure and mineralogy and are extensively weathered and relatively small.

Prospect JJM2

The most northerly prospect is JJM2. This area contains a central NNW trending fault covered with aeolian sand and alluvium. The beds either side of this fault generally dip toward it at inclinations down to 30° (Chaumont, 1970a, map 2).

The most mineralized part of the area is a tiny basin the south-east where aeolian sand has buried some ancient workings. Using a Poclain hydraulic shovel at trench T20 Chaumont discovered an inclined cavity in the centre of this tiny basin. Above the cavity is a spotted white argillaceous limestone containing galena grains up to 60 mm long together with secondary lead and zinc minerals. A specimen from this locality assayed 13% Pb, 40% Zn and 5g/ton Ag (Chaumont, 1970a). Similar limestone elsewhere is generally not mineralized. Locally, replacement has been so intense that some specimens consist entirely of ore minerals. The best specimens from Jabal Dhaylan are from this locality.

A typical specimen from this south eastern part of JJM2 consists essentially of white hydrozincite, with fine irregular brown spots, and lesser galena, cerussite and hemimorphite. The grain size is mostly fine but has a wide range. Hydrozincite is generally micro-crystalline to sub-microscopic whereas galena is in masses of centimetric grain mosaics. The hemimorphite and cerussite are medium to fine grained. Hemimorphite occurs as sheaves of crystals but most of the other secondary minerals are of irregular shape commonly in the form of short tiny veinlets. Replaced fossils are discernable in places. Black manganiferous and brown ferriferous spots of various sizes and concentrations are typical in this ore.

Galena typically occurs as massive coarse granular aggregates invariably partially replaced. Cerussite occurs as three types. The first is found in contact with galena and is massive, vitreous and transparent. The second is translucent, white and has cleavages discernable under a binocular microscope. It is more common than the first type which it appears to replace. The third type is white and impalpable and has been identified by assays and by XRD. The first type was mistaken for anglesite which was expected from the oft quoted breakdown, " $\text{PbS} \longrightarrow \text{PbSO}_4 \longrightarrow \text{PbCO}_3$ ", however no anglesite was found at Jabal Dhaylan or anywhere else along the coastal section of Saudi Arabia. Ramdohr (1969) has commented on the general scarcity of anglesite.

Plattnerite is more common than expected. It occurs as thin brownish black powdery coats on cerussite, and as a brown colour stain when finely disseminated in hydrozincite. Laurionite was identified from an XRD trace of typical white hydrozincite ore. It is associated with cerussite, hydrocerussite, and pyromorphite(?). A very pale green mineral of amorphous appearance coats hydrozincite in places as an irregular film. It could not be positively identified by XRD but is probably a disordered form of pyromorphite. It is associated with a thin film of gypsum which appears to be the last mineral to form at this locality. A bright pale purplish pink powdery mineral occurs very rarely on hydrozincite in vugs. It could not be identified. It also very rarely occurs at JJM3 associated with secondary lead and zinc minerals.

Hydrozincite was identified from an XRD trace of a typical brown specimen of hydrozincite ore with lesser cerussite. It is evidently a white impalpable substance associated with cerussite. It commonly occurs as two types. Generally it is white and impalpable but in places it forms irregular veinlets in which it has a radiating fibrous sub-botryoidal habit. In veinlets with axial cavities a translucent type occurs at the margins and an opaque type at the interior. Calcite crystals separate these two types of hydrozincite in a few places and are rarely found at both sides of the opaque white hydrozincite. The opaque type has a chalky appearance in places but elsewhere, especially in cavities, it has a distinctive fluffy appearance and pearly lustre due to minute bladed crystals less than 0.5 mm long. Much of the hydrozincite probably pseudomorphs colloform sphalerite or wurtzite. While hydrozincite is abundant, smithsonite is apparently absent from this locality and all other localities of lead-zinc mineralization in the Arabian coastal plain, with one insignificant possible exception at JJM3. Dadet et al. (1970) recorded smithsonite from this locality and also noted that "efflorescent hydrozincite, doubtless due to winter humidity, marks a ring of rubble around a basin filled with aeolian sand". It is suggested that the "smithsonite" is actually non-efflorescent hydrozincite!

Hemimorphite (calamine) occurs in cavities within hydrozincite masses as radiating euhedral bladed transparent crystals up to a few millimeters long. Less commonly it occurs as tiny veinlets. In places hemimorphite crystals are black due to an intermixture with manganese oxide (probably cryptomelane). In another place it is pink due to an intermixture with iron oxides.

About 100 m NE of this locality the writer discovered a small, partially collapsed ancient mine (figs. 20, 21, 22, 23). The main excavation appears to follow a NNE fault and a branch off this follows an easterly fault. No slag heaps were found near this mine but it must be noted that as the workings found by Chaumont were hidden by sand much of the evidence of old workings is likely to remain hidden. The roof in the western branch appears to follow a limonitic contact between a dark spotted shale containing impalpable hydrozincite and cerussite and a spotted pale brown limestone with small geodes containing dolomite crystals.

The only occurrence of sphalerite in outcrop at Jabal Dhaylan is in Chaumont's trench 18 in the southern part of JJM2. Marchessau (1969) claimed to have found sphalerite crystals in vugs in trench 11 to the

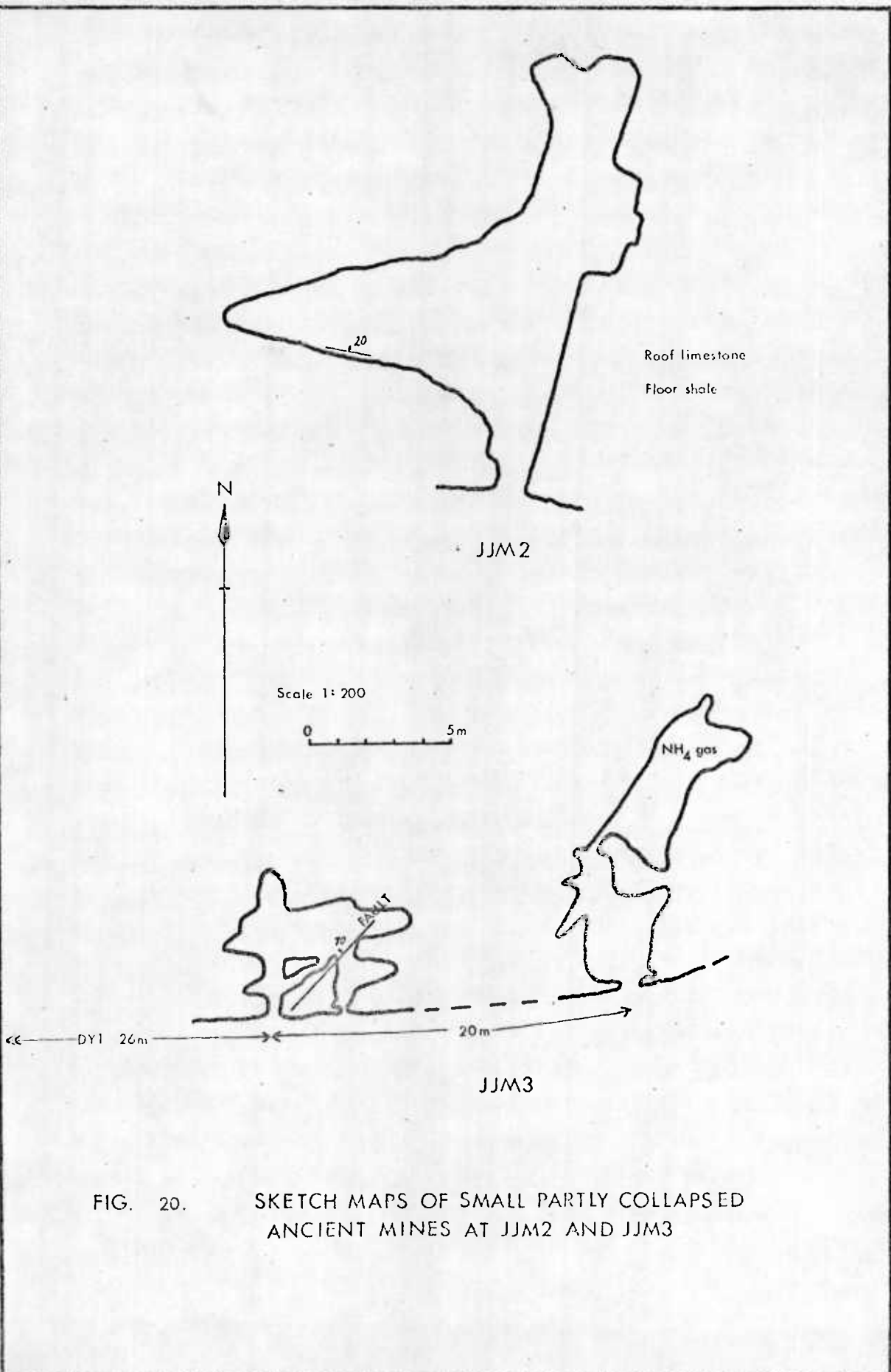




Fig. 21. Portal to ancient mine at JJM2 (circled with lateral extent of workings indicated) viewed from south. A small stone circle marks the summit of the Miocene limestone hill.



Fig. 22. Close view of portal shown in fig. 21.



Fig. 23. View inside ancient mine from western extremity. The small stalactites are composed of crystal laths of urea.

north but the writer diligently searched the locality and found none. The sphalerite in trench 18 is very difficult to identify (except by XRD) and is invariably in the form of very fine grained (< 1 mm dia.) disseminated specks. It is very similar to the sphalerite found at JJM3. The material found by Marchessau was probably manganiferous hemimorphite (Chaumont, pers. com. concurs with this). The sphalerite is associated with corroded galena and both are suspended in selenite.

Some near-by gypsum grains with abundant inclusions have cleavage planes extending over 80 mm. The gypsum has veined and replaced black spotted marl. At outcrops, spongy galena masses up to 35 mm long are partially replaced by both types of cerussite. The cerussite is partially replaced by plattnerite, and coating this is a very thin film of minium. Traces of pyromorphite also appear to be associated with the galena. The sphalerite is replaced by hydrozincite where it is not protected by gypsum. Corals within the marl at this locality have been corroded and replaced by hemimorphite and manganese oxides (fig. 24). Elsewhere at JJM2, particularly in the north, gypsum forms outcrops of white translucent massive material (opaque white at exposed surfaces) with sub-horizontal satin-spar veins very like those thought to have formed by hydraulic fracture (Shearman, et al. 1972). For details of other very minor occurrences of secondary lead and zinc minerals at JJM2 see Marchessau (1969) and Chaumont (1970a).

The only significant landmark at JJM2 is the well situated in Wadi Malah between JJM2 and JJM3.

Trench 18 is 475 m on a magnetic bearing of 2.5° from the well in Wadi Malah. Trench 18 to Chaumont's ancient workings is 378 m on a magnetic bearing of 350° ; Chaumont's ancient workings to the ancient mine in fig. 21 is 92 m on a magnetic bearing of 37° ; the ancient mine in fig. 21 to the barite locality (discussed later) is 200 m on a magnetic bearing of 360° .

Prospect JJM3

JJM3 was considered by BRGM geologists to be the most promising area at Jabal Dhaylan and 19 holes were drilled to basement in this area to "determine the nature and extent of mineralization, the plunge of the basement, and the variations in facies and thickness of the Tertiary rocks in so far as they can help explain the mineralization" (Bigot, 1970).



Fig. 24. Coral replaced by hemimorphite and lesser manganese oxide from trench 18, JJM2.

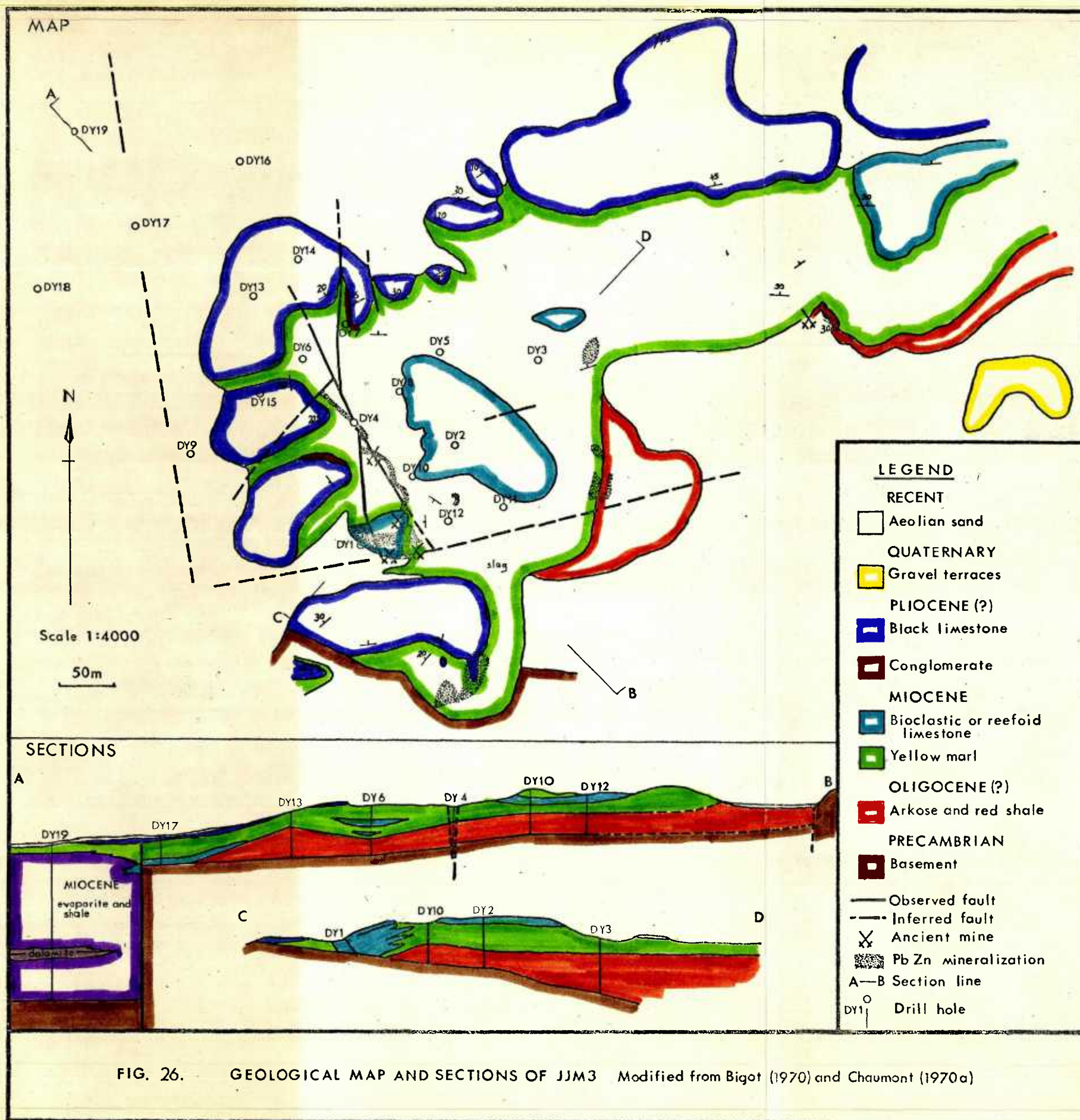


Fig. 25. Mineralized Miocene reef limestone at JJM5 viewed from south showing.
 1 & 2 Entrances to ancient mines
 3 & 4 BRGM drill hole collars
 5 & 6 BRGM pits and trenches

According to Bigot (1970) the mineralization at JJM3 is preferentially concentrated in Miocene reefoid limestone which was deposited along an ancient shoreline. The grade and the Pb/Zn ratio diminish away from the reef limestone into the associated marl. Galena, the only sulphide, occurs in limestone much more than in marl.

The Miocene has two domains separated by a normal synsedimentary fault (trend N20W, throw 65 m, W side down) which controlled submarine relief and the facies of sedimentation (Bigot, 1970). In the eastern domain the basement rises gently SSE and has a saddle trending N20W due to another fault with a throw of 10 m, E side down (fig. 26). Only two drill holes penetrated Miocene rocks of the western domain. The western domain is completely covered by superficial aeolian sand. By superimposing facies, mineralization and basement isobath maps Bigot found that the well-mineralized zones coincide with elevations of the basement. As elsewhere at Jabal Dhaylan, the mineralization at JJM3 appears to be related to faults, cavities, and porous zones, and is now mostly in the form of secondary lead and zinc minerals. Traces of these minerals are common throughout JJM3.

The most strongly mineralized part of JJM3 is the reef at drill hole DY1 (fig. 26). The 400 m² surface of this outcrop averaged 22% Pb + Zn (Chaumont, 1970a). Although mineralization in this area is patchy it is concentrated around a northerly trending fault. The writer discovered two ancient adit mines (fig. 20 and 25) and other signs of old workings here. This locality was probably more extensively mined than the sketches suggest. The workings have partially collapsed and are not easily recognised (fig. 25). The two adits were probably connected. As they are largely filled with aeolian sand few observations could be made however the limestone inside appears to be the same as that at outcrop and is irregularly mineralized with hemimorphite, hydrozincite, cerussite and relics of galena, together with small pockets of manganese oxides, goethite and plattnerite. In the eastern adit relics of a steep northerly trending galena vein 40 mm wide were found. This galena is surrounded by cerussite and hemimorphite, and can be broken away clean from the cavity it fills. Typical ore in this locality is white and fine grained and consists of hydrozincite, hemimorphite, cerussite, galena, calcite, barite (described later) and gypsum. Corals can be found replaced by cerussite, hydrozincite and galena. Galena crystals in places form porous masses up to 150 mm long of cubes averaging 3 mm wide. They are partially replaced



by cerussite and lesser plattnerite. Traces of minium were found on exposed plattnerite surfaces. Cubes of galena 1 mm wide, altered to cerussite and plattnerite were found on barite.

About 100 m north of the above locality is a steep NW trending shear zone mineralized over an area 120 m long and 2.5 to 5.75 m wide with an average grade of 21% Pb + Zn (Chaumont, 1970a). Although the country rock is mostly marl, the shear zone is occupied by a manganese oxide spotted marly limestone. The shear zone is silicified in places up to 3 m wide and the rock contains manganese oxides, secondary lead and zinc minerals and small patches of galena. Some Plio-Pleistocene limestone overlying part of the shear zone is silicified but not significantly mineralized. Near the centre, the shear zone is displaced 6 m by a NE trending dextral fault. At old workings near there the rock is very rich in manganese oxides which occur as spots and pockets replacing the marl. A specimen from there has a nodule of galena, surrounded by cerussite, resting on hemimorphite crystals lining a cavity. At first, it would seem that the galena was deposited on the hemimorphite but it is obvious from other examples nearby that the galena is primary and the deposition of the hemimorphite followed its corrosion. Associated secondary minerals are pyromorphite and a rare unknown pale purplish pink powdery mineral similar to that found at JJM2.

At the southern part of this shear zone are more ancient workings. The marl at this locality is veined and replaced by hydrozincite, pink hemimorphite (iron stained), galena and cerussite. Approximately 60 m further south in the saddle near DY11, near a dispersed slag heap, is a 10 mm wide vein of galena trending 120° . When the galena is removed the vein walls are clean and show no signs of the former ore.

In the north-eastern part of JJM3 are more relics of ancient workings. Nearby the marl at the Miocene marl - Oligocene arkose contact is spotted with manganese oxide and in places is completely replaced by sooty pyrolusite. This rock contains galena, calcite, hemimorphite, hydrozincite, cerussite (two types), plattnerite and traces of minium. Also present are irregular grains of green pyromorphite and pale brown pyromorphite or mimetite. A creamy white lump within the manganese-rich marl was found to contain calcite, hemimorphite, cerussite, hydrocerussite, pyromorphite and gypsum. Galena at this locality has curved cleavages probably due to growth defects.

In the southern part of JJM3 at "Calamine Hill" are hemimorphite crystals up to 20 mm long associated with calcite crystals up to 10 mm long in geodes within grey to black manganiferous Plio-Pleistocene limestone.

Information on mineralization in the western sector of JJM3 is available only from two drill holes. In DY18 the highest assay values are 1.65% Zn and 750 ppm Pb, and in DY19, 2.4% Zn and 650 ppm Pb (Bigot, 1970). In DY18 the highest zinc values are from two core intervals, one consisting of interbedded clay, gypsum and anhydrite, the other of porous white limestone.

In DY19 the zinc zone is in the upper 3.6 m of a 17.7 m core length of gypsum-veined dolomite. This is the only extensive dolomite found at Jabal Dhaylan. It is sandwiched between two clay layers within a generally evaporitic sequence. The dolomite is a massive, white, aphanitic rock except for places where it is interlayered with gypsum dipping at 70° . Under the binocular microscope myriads of tiny pores are visible in the dolomite. When it was crushed with a pestle and mortar some hydrogen sulphide gas was released. Vague pink patches in the dolomite are due to concentrations of disseminated tiny sphalerite grains. Under the microscope the sphalerite grains were found within the dolomite and more commonly in cavities and veinlets filled with later gypsum and lesser anhydrite. In a vein composed of a single gypsum crystal across the drill core, 1 mm diameter galena crystals and smaller pyrite - marcasite grains were found. The galena crystals are in cubes, octahedra and modified forms. Tiny irregular veinlets of galena 1 mm wide surrounded by patches of disseminated sphalerite about 5 mm wide, specks of pyrite-marcasite associated with sphalerite, and fine pyritic veinlets, are sparsely disseminated in the dolomite. Balls of pyrite-marcasite up to 1 mm diameter are widely spaced through the gypsum associated with the dolomite. Sphalerite occurs rarely in veinlets as concentrations of tiny grains and as small colloform masses. In one place a very pale part of the sphalerite mass could possibly be smithsonite.

In polished section the sphalerite has a colloform texture and is in the form of small spheres (fig. 27). The sphalerite was identified by XRD and it should be noted that the diffraction peaks are broad probably because of zoning (the XRD patterns of sphalerite and wurtzite are similar).



Fig. 27. PLM photograph of zoned spheroidal sphalerite in gypsum from DY19, JJM3, Jabal Dhaylan (PS in PPL).

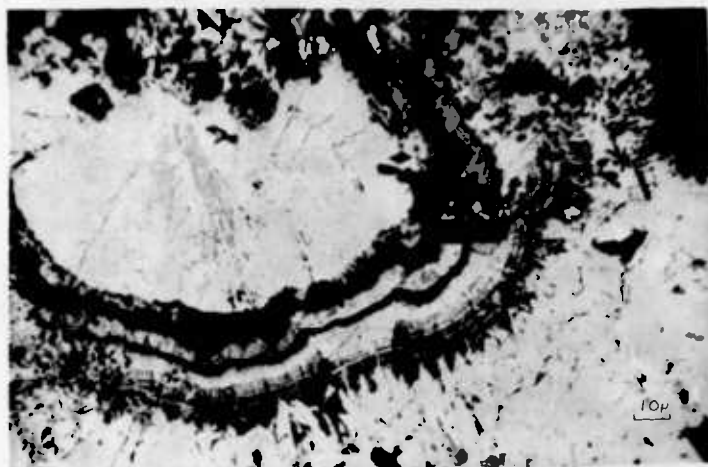


Fig. 28. RLM photograph of marcasite laths (lower right), pyrite (upper left) and intermediate material in dolomite from DY19, JJM3, Jabal Dhaylan (PS in PPL).



Fig. 29. PLM photograph of marcasite (large white mineral) and spheroidal zoned sphalerite in gypsum from DY19, JJM3, Jabal Dhaylan (PS in PPL).

In polished section it somewhat resembles goethite. Round pyrite grains are surrounded by radiating marcasite crystals (fig. 28). Between the two is a material of intermediate properties resembling Ramdohr's (1969) colloform melnikovite-pyrite. It is possibly a cryptocrystalline mixture of pyrite and marcasite. In general marcasite is much more common than pyrite and occurs mostly in bladed crystals. The marcasite appears to post-date (replace) sphalerite and pre-date gypsum (fig. 29).

In a thin section of the dolomite stained with alizarin red S, Dr. D. Shearman (pers. com.) found sparse relics of calcite as well as relics of anhydrite within gypsum. The dolomite apparently replaced calcite, and anhydrite replaced both calcite and dolomite. Gypsum then replaced the anhydrite probably under conditions of hydraulic fracture (Shearman, et al. 1972). Dr. Shearman considers that the rock most probably was a limestone that was replaced by anhydrite and dolomite almost simultaneously, and was then replaced by gypsum. Sphalerite is associated with dolomite and particularly with gypsum but not with calcite or anhydrite.

In general, the gypsum at Jabal Dhaylan other than that in veins has probably replaced nodular anhydrite - relics of which have been found within the gypsum. The gypsum occurs as large crystals, as satin spar and as an intermediate product. Dr. Shearman kindly examined specimens of the evaporites from Jabal Dhaylan and concluded that they are all quite typical of shallow water evaporites as found generally in sabkha environments.

Dr. R. J. Watling kindly analysed for mercury a mineralized gypsiferous dolomite (DY19 - 86.9) and an ordinary non-mineralized gypsum (DY19 - 23.0) and obtained the results listed below.

DY19 - 86.9	DY19 - 23.0
47.5 ppb Hg as HgS	13.5 ppb Hg as vapour
(10 ppb Hg as HgS trapped in crystals)	10 ppb Hg as HgCl ₂
15 ppb Hg lattice held	

In the presence of excess chloride and sulphide ions heat is necessary for mercury to form as the sulphide rather than the chloride. These results therefore imply that the mineralization was epigenetic and formed at a temperature possibly within the range of 150 to 250°C (R.J. Watling pers. com.). It also shows that sulphur activity was strictly limited to

certain areas. This further suggests epigenesis but these tentative conclusions have a low confidence rating because only two specimens were examined.

At Jabal Dhaylan the ore has a low precious metal content. Cu is $< 0.05\%$, Au < 0.5 g/t, Ag < 7 g/t with exceptions up to 124 g/t and Cd is about 0.07% (Bigot, 1970 and Chaumont, 1970a).

Prospect JJM4

Prospect JJM4, near the centre of Jabal Dhaylan, is the site of weakly mineralized Plio-Pleistocene marl, argillaceous limestone, calcarenite and reef limestone which unconformably overlie Oligocene arkose and shale (Chaumont, 1970a).

The limestones dip 15° to 30° NW and are intersected by NW and NE faults bearing secondary zinc, iron and lead minerals. Mineralization however is mostly in marl and is restricted mainly to an area about 10 m wide. Maximum assay values are 1.85% Pb, 33.75% Zn (Chaumont, 1970a, and Marchessau, 1969). Typically the mineralization occurs as fine stockworks of hydrozincite veinlets and as infillings and replacement features in the marl and limestone. The limestone is spotted with manganese oxides and contains tiny irregular veinlets of barite. Calcite and hydrozincite line geodes in pale brown fossiliferous vuggy limestones, the interior of which is chertified through replacement by silica. A soft black rock with irregular open white veinlets of hydrozincite from this locality consists mostly of plattnerite and hydrozincite with minor halite. Underlying this zone is a porous orange-brown "gossan" possibly a replaced limestone, with hemimorphite crystals and relics of barite.

North-west of the ore zone are NW trending veins of limonite, with traces of hydrozincite, up to 100 mm wide. These dip 75° SW and are parallel to a N40W joint set.

Relatively more secondary iron minerals accompany the mineralization at JJM4 than elsewhere at Jabal Dhaylan. Possibly sphalerite and pyrite-marcasite dominated the primary ores.

Prospect JJM5

Prospect JJM5 is south-east of JJM4 near the end of the main hill of Jabal Dhaylan (fig. 17). The mineralization occurs in Plio-Pleistocene reef limestone and marly limestone in a monocline dipping 20° SW over Oligocene arkose and shale. The limestones are commonly brecciated and contain numerous northerly trending joints and small faults (Chaumont, 1970a). Many features at JJM5 have a northerly trend e.g. contacts, scarps, alluvial terraces, and faults.

The main mineralization at JJM5 is a gossanous lens 56 m long trending N 20° E with an average grade of 3.3% Pb, 30.9% Zn (Chaumont, 1970a). This wedges out to the north, and in the south is represented by veinlets of calcite with manganese oxide. The mineralized zone is in brecciated marly limestone and dips steeply west. It probably consists of a series of weathered irregular veins. The ore zone is porous though the surrounding rock is not.

Adjacent to this mineralized zone the writer discovered some ancient mine adits now mostly collapsed and filled with aeolian sand (figs. 30, 31, 32 and 33). Dispersed slag heaps occur at the foot of the hill below the adits. Although the adits are difficult to enter, joints within them trending N 10° E and N 70° W were noted.

Within the adits and the mineralized zone the mineralization is essentially hemimorphite associated with cerussite, hydrozincite, irregular remnants of galena, and halite together with much goethite, hematite and pyrolusite in-filling and replacing the limestone in very porous zones. Hemimorphite is generally crystallized and colourless but rare pale purple and pale green varieties were observed in the centre of the area. Barite veins associated with manganese oxide also occur in this area but they are not directly associated with the lead and zinc mineralization.

Prospect JJM1

Prospect JJM1 is a very minor mineral occurrence in the black hill, east of JJM5, where Marchessau first found a few specks of galena associated with secondary lead and zinc minerals in Miocene limestone boulders within marl capped by Plio-Pleistocene limestone. The writer found only traces of hemimorphite and cerussite in manganese oxide spotted limestone, however Chaumont (1970a) observed rare galena specks at this locality.

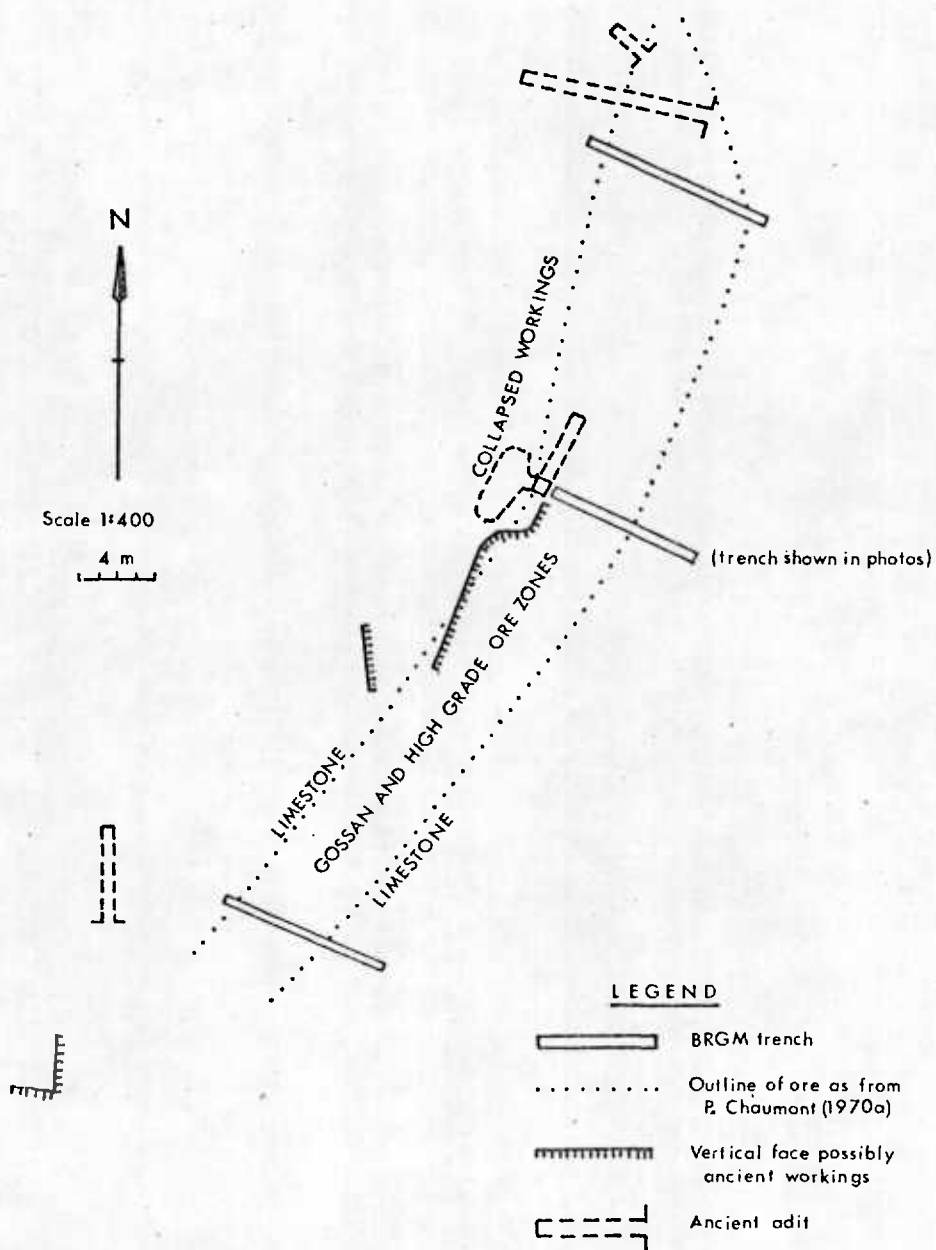


FIG 30.

ANCIENT WORKINGS AND NEW TRENCHES AT JJM5



Fig. 31. Prospect JJM5 viewed from south.



Fig. 32. Enlargement of fig. 31. Figure is beside ancient workings and BRGM trench



Fig. 33. Prospect JJM5 viewed from north. Slag heaps at base of hill. Could be sites of ancient furnaces. BRGM trench near figure.

Prospect JJM10

Prospect JJM10 is the southern most occurrence of lead and zinc at Jabal Dhaylan and is 1 km SSW of JJM1. This occurrence is in a small outcrop of Oligocene arkose, very thin altered subhorizontal white Miocene limestone, and thin brown Plio-Pleistocene limestone. The lead-zinc mineralization is associated with copper mineralization (discussed later) and occurs in patches around N 80° E trending joints. By trenching, Chaumont (1970a) found that the mineralized zone is about 20 m long and 5 m wide. The highest assay is 1.05% Cu, 4.80% Zn (Marchessau, 1969). An average mineralized specimen was found to have 1.55% Pb, 4.75% Zn and 0.05% Cu (Chaumont, 1970a).

A typical mineralized specimen of Miocene ore consists of white fine grained massive limestone with spots, patches and dendrites of manganese oxide, and with chrysocolla patches surrounded by cherty limestone and goethite in places. Chrysocolla commonly surrounds manganese oxide spots. Irregular galena grains are sparse, and are found in areas black with manganese oxide. The galena is partially replaced by secondary copper minerals and by cerussite of the two types described from JJM2. Plattnerite, rare hydrocerussite, and possibly minium are associated with the galena. Cupriferous galena (Clark and Sillitoe, 1971) was also observed under the reflecting microscope in the replacement zone at the edge of normal galena. Zinc occurs as crystallized hemimorphite and as powdery hydrozincite associated with the secondary lead and copper minerals.

In the massive pale brown Plio-Pleistocene limestone veinlets of calcite associated with manganese oxides are common. As with the Miocene, cerussite, hemimorphite, and rare galena are associated with manganiferous patches. Manganese oxide occurs as spots, dendritic masses and veinlets.

LEAD-ZINC MINERALIZATION IN OLIGOCENE ARKOSE

Lead and zinc mineralization in Oligocene rocks is meagre and occurs only as spots and hazy masses of cerussite, hydrozincite and plattnerite(?) cementing the arkose. The occurrences are vaguely oriented NNW and the main occurrences, JJM7 and JJM8 and JJM9, are colinear along a NNW line. All the occurrences are at low elevations peripheral to the hills.

The best example of this type of mineralization is at JJM7 where the maximum assay values are 12% Pb, 1.8% Zn (Chaumont, 1970a). The arkose at this locality is typically friable. In the mineralized patches it has fine white powder-like cement which under the microscope appears as fine fibrous crystals of cerussite around quartz and feldspar grains. The arkose ranges from pale brown to dark brown (due to organic matter), and in some mineralized patches has spots of manganese oxide. Mollusc(?) borings were found at JJM8.

Traces of secondary lead and zinc minerals also occur in Oligocene arkose at JJM3, JJM1 and PL 117 (found by P. Laloux of BRGM). Bigot (1970) found that the Oligocene at JJM3 everywhere has some secondary lead and zinc in it but this mineralization is not apparently influenced by variations in thickness of the arkose. Chaumont (1970a) found an outcrop area at JJM3 of 400m² of mineralized Oligocene arkose with assays up to 7% Pb and <2% Zn.

COPPER MINERALIZATION

Very weak copper mineralization occurs in rocks of Precambrian, Oligocene, Miocene and Plio-Pleistocene ages at Jabal Dhaylan.

JJM11, a Precambrian window SSW of Jabal Dhaylan (fig. 17) was thought by Marchessau (1969) to be typical of a possible source area for the copper in the younger rocks. The mineralization however is extremely sparse and is limited to specks of chalcopyrite and pyrite together with traces of chrysocolla in a few hand specimens of vein quartz from sheared quartzite and schist.

At JJM6 in the central part of Jabal Dhaylan chrysocolla occurs in Oligocene arkose. This is the most intense copper mineralization at Jabal Dhaylan. The maximum assay value is 6% (Marchessau, 1969) and the exposed dimensions of the "channel" or patch from which this came are 1 by 0.4 m (Chaumont, 1970a). Although the mineralization transgresses cross-bedding at one place it is nevertheless confined to arkose. Mollusc(?) borings were found in some of the arkose but are not common. Typically the cupriferous sandstone is bright green due to chrysocolla cement and rare chrysocolla veinlets about 1 mm thick. The least altered material is almost turquoise-blue but more exposed parts of the rock are apple

green. Atacamite was found in parts of weathered surfaces. Brown carbonaceous streaks occur in some of the mineralized patches but in other localities are not associated with mineralization.

At JJM10 an Oligocene arkose is locally cemented with chrysocolla and atacamite similar to that at JJM6 but far less extensive. Overlying this is the copper-lead-zinc mineralized limestone previously described. Chrysocolla in this limestone surrounds dark spots, some of which surround corroded galena. This galena is partially replaced, particularly along cleavage traces, by cupriferous galena, chalcocite, cerussite (both types) and covellite. In the outer part of the dark zone are traces of boleite, diaboite, atacamite and possibly malachite, azurite and tenorite. Very similar copper mineralization was found at Merataba (discussed in section 1.2.4). Dr. R. Sillitoe kindly examined the polished sections and identified the cupriferous galena. It is probably a meta-stable product and bears similarities with that found in Chile (Clark and Sillitoe, 1971).

At Prospect CJ1 in the SSW part of Jabal Dhaylan, J. Costans of BRGM found chrysocolla about 200 m from where Y. Berton previously found a similar occurrence. This is close to a major basement fault. In this area is a subhorizontal grey Plio-Pleistocene conglomerate weakly cemented with kaolinite and calcite. In places it has ENE trending joints and in some places away from the mineralization it contains rare geodes with manganese oxide and calcite crystals. The chrysocolla occurs in vague patchy areas as cement to the conglomerate and an associated coarse sandstone, as sparse infillings, and as veinlets in the cement and in pebbles. In places the chrysocolla is colloform but mostly it is massive and impalpable. It is pale green to greenish blue in hand specimen. Under the microscope it has a similar appearance to chalcedony and commonly has brush extinction occurring in bluish green spherulites and in layers of fibres with the fibres normal to the layering. It also occurs as a deeper green cryptocrystalline or amorphous variety deposited after the other variety. Atacamite is sparsely distributed on weathered surfaces.

An effort was made to find the source of the copper by breaking many of the pebbles in the conglomerate however none was found to contain possible primary copper minerals. Vague dark spots in the chrysocolla were thought

to be possibly chalcocite however XRD results (main peaks at 8.15, 3.13, 2.76 Å) suggest these are possibly amphibole! This puzzling result is possibly due to relics from replaced rock fragments, although partially replaced rock fragments are not obvious.

BARIUM MINERALIZATION

Barite occurs sparsely at Jabal Dhaylan in Plio-Pleistocene and Miocene limestone and shale at JJM2, JJM3 and JJM5. It is generally found as radiating lamellar aggregates and as layers of parallel lamellar aggregates with the crystals normal to the layers. The layers originate from limestone wall rock or from a plane about which a double layer has formed. Between double layers commonly are open spaces (fig. 34). This type of structure is very similar to a dolomitic structure which Dunsmore (1973) found at Pine Point, and probably has a similar origin. At Pine Point the dolomite grew from tiny joints and cracks in limestone; later acids attacked the rock dissolving out the calcite but not the dolomite.

The thickest aggregate of barite layers found at Jabal Dhaylan is 60 mm thick with individual layers up to 5 mm thick. Sparse calcite crystals rest on the barite. Four veins with a northerly trend were found in limestone up-slope and down-slope from the ancient workings at JJM5 but no barite was found with the main lead-zinc mineralization. In thin section the barite has a spherulitic texture. At the centres of the plumes are calcite mosaics and between the plumes are extremely fine grained mosaics of quartz (chert). Manganese oxide is common along the axes of the barite double-layers. It is also found in places on the outer side of the layers as well, and is probably younger than the barite.

Veins similar to those at JJM5 were found at JJM3. The specimen shown in fig. 34 is 40 mm thick and contains 1 mm wide cubes of galena, partially replaced by cerussite and plattnerite, resting on the barite.

At JJM2 individual layers of barite up to 20 mm thick were found near the ancient workings. This barite is botryoidal and translucent and is overlain by a chalky white opaque material which appears to have replaced translucent barite. From XRD this appears to consist of barite plus minor gypsum, hence it is probably a weathered form of barite. In the



Fig. 34. Specimen of layered barite from JIM3 showing double layers, cavities and galena.

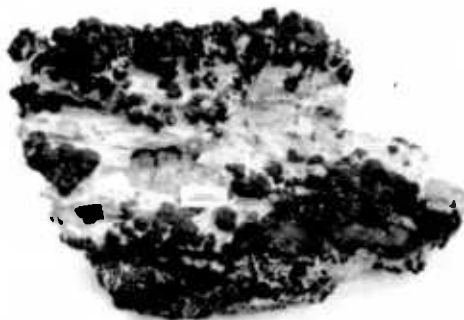


Fig. 35. Specimen of dendritic manganese oxide minerals, in soft marl, forming wall rock to small vein at JIM3.

northern part of this locality four irregular veinlets about 20 mm wide were found in Miocene limestone. They range in strike from WSW to NNW and in dip from 30° to 70° . In bulges in the veinlets axial cavities are filled with calcite crystals and manganese oxide powder.

Marchessau (1969) claimed that the barite at JJM2 is bedded however the writer found no evidence of this. Apart from that in veinlets the only other barite was found as fragments within soft shale creeping down slopes.

MANGANESE MINERALIZATION

Manganese mineralization is extensive at Jabal Dhaylan but is not sufficiently concentrated to form an ore deposit. It is found in rocks of all ages and is especially conspicuous in the black Plio-Pleistocene limestones (e.g. at JJM1, JJM2, JJM3 and JJM4) where weathering appears to have reconcentrated it on exposed surfaces (fig. 15).

Lead-zinc mineralization is almost invariably accompanied by manganese oxide and is accompanied by iron oxide to a much lesser extent except at JJM4. Manganese oxide spotted Miocene limestone is a guide to ore and the most concentrated occurrences of manganese are near lead-zinc occurrences.

The manganese minerals are generally difficult to identify and occur mostly as black powder (pyrolusite) and impalpable black masses with a range of hardness. It is likely that in many places two or more manganese minerals of similar appearance are intimately mixed together. XRD traces were made on 21 specimens, and several of these contained two or more phases. XRD alone is insufficient for the determination of many manganese minerals and 5 of the preparations were qualitatively analysed in the SEM using XRA. Pyrolusite appears to be fairly common; psilomelane "sensu stricto" was not found although minerals related to it such as cryptomelane and hollandite were.

The largest concentration of fairly pure manganese oxide is near the crest of the hill between JJM3 and JJM9. Here is a pocket, a meter or more square, essentially of dull aphanitic pyrolusite in porous coarsely crystalline brown Miocene limestone. The pyrolusite is finely veined by calcite and fibrous hollandite.



Fig. 36. Manganese mineralization near
DY11 collar, JJM3.

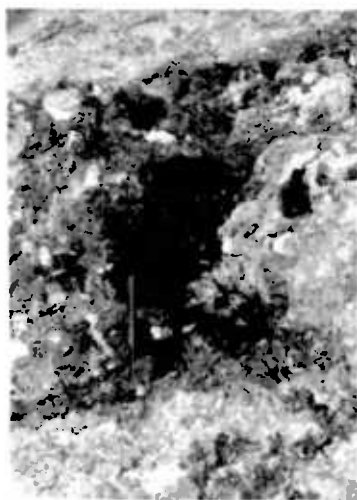


Fig. 37. Enlargement of part of fig. 36.



Fig. 38. Hydrohetaerolite specimen
with velvety lustre from area shown in
fig. 36.

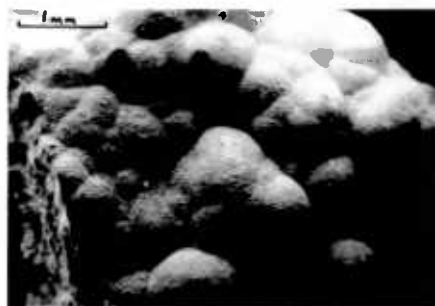


Fig. 39. SEM photograph of part of
specimen shown in fig. 38.



Fig. 40. Enlargement of part of
specimen shown in fig. 39.

The most interesting manganese occurrence at Jabal Dhaylan is in an exposure excavated for the positioning of drill hole DY11. This revealed irregular cavernous pockets of manganese minerals in Miocene limestone (fig. 36). The manganese minerals are crusty, nodular, botryoidal and stalactitic (fig. 37). Dull massive material within the crusts and nodules consists of coronardite with minor chalcophanite. The outer part of the nodular surface has a very unusual appearance ranging from drusy to velvety (fig. 38). This material and some thread-like stalactites consist of hydrohetaerolite. The cause of the unusual lustre was found with the SEM to be due to reflections from minute terminal crystal faces (figs. 39, 40). Rosettes of small plates of chalcophanite rest on the surfaces of some hydrohetaerolite nodules.

The manganese minerals associated with lead-zinc patches of replacement mineralization are coronardite, chalcophanite and pyrolusite.

In addition to pocket infillings and replacement patches the manganese minerals also occur in veins and at some localities along the unconformity between Oligocene arkose and Miocene limestones.

In the northern part of JJM3 is an irregular vein up to a decimeter wide in marl. It consists of interlocking blocky crystals of ramsdellite, plumes and radiating sheaves of coronardite and pyrolusite. Regular joint-controlled veins up to a decimeter wide are found elsewhere in marls, shales and arkose at JJM2 and JJM3. The veins are massive along the joints and are dendritic in the host rocks along the contacts. The dendritic material within the marl shown in fig. 35 is probably an intimate mixture of hydrohetaerolite, coronardite, cryptomelane and quartz. Its XRD pattern does not agree with any single manganese mineral; the d Å spacings and intensities are 6.90 (20%), 4.24 (40%), 3.49 (10%), 3.34 (30%), 3.10 (15%), 2.87 (20%), 2.46 (100%), 2.20 (20%), 1.56 (20%). XRA found the following elements, in order of concentration, Mn, Ca, Si, Zn, Pb, Fe, K.

Manganese oxide also occurs in tiny disseminated veinlets associated with calcite in limestones and as very fine grained dispersed material in limestones. As an acid residue this dispersed material corresponds to mixtures of chalcophanite and cryptomelane in five specimens, and pyrolusite

with minor hollandite in another specimen. Manganese mineralization has probably been a protracted process. Marchessau (1969) noted manganese rich layers in Quaternary sandstone west of Wadi Zarufi.

PROSPECTING METHODS

The rediscovery of mineralization at Jabal Dhaylan resulted from knapping mapping. As the secondary lead and zinc mineralization is generally difficult to observe with the naked eye BRGM developed field staining methods using reagents "Jerome" and "Gertrude". Marchessau and Costans subsequently found half of the mineral occurrences at Jabal Dhaylan with these reagents (Marchessau 1969).

"Jerome" is 10% acetic acid mixed with equal parts 5% potassium iodide. It gives a yellow stain with lead minerals. "Gertrude" is 3% potassium ferricyanide mixed (on the day of use) with equal parts 3% oxalic acid plus 0.5% diethylaniline. It gives an orange-red stain with zinc (Marchessau, 1969). According to Chaumont (1970a) "Jerome" is sensitive for lead contents of 0.5 to 1.0% minimum and "Gertrude" is sensitive for zinc contents of about 5% minimum. Stains are sometimes obtained however with lower metal concentrations. Apart from black manganese minerals the main secondary zinc minerals are hydrozincite and hemimorphite. Hydrozincite is readily soluble in acids but hemimorphite gelatinizes with common acids (Dana, 1932). It may be possible therefore to improve this technique by using a reagent capable of dissolving hemimorphite. Another way of possibly improving the technique would be to use a much more sensitive staining reagent. A.J. Thompson suggested zincon (Johnson, ed., 1964) for this (pers. com.).

A thin film of gypsum, almost invisible to the naked eye, coats some areas of outcrop at Jabal Dhaylan. Obviously, therefore, it is important when using staining reagents to apply them to freshly broken surfaces.

Manganese minerals appear to be a useful guide to mineralization especially as they commonly contain lead and zinc.

A program of geochemical sampling across faults was planned but in the field this appeared to be impractical because the faults are commonly obscured by scree blocks, migrating aeolian sand, or extensive wadi alluvium (e.g. figs. 16 and 19 in discussion on structure).

The results of Chaumont's (1970b) regional geochemical survey based on soil samples show no particular correlation between faults and lead-zinc anomalies. Copper anomalies, however, are independent of lead-zinc anomalies and show some relation to faults. Of the 462 samples collected only 17 contained copper values above 50 ppm; the highest being 300 ppm. Of these, 5 appear to be random and 12 were taken from possible N and NNW fault traces in arkose and conglomerate. Many barren samples however were also collected along these possible fault traces. Copper anomalies were not found near known occurrences at CJ1, JJM6 and JJM10 hence these copper prospects would not have been found if prospecting was based only on this geochemical survey. The nearest copper anomaly to CJ1 is 7 km away and appears to be random. The zinc anomalies however are useful and spread from the area of JJM2 and JJM3 for 9 km toward the sea in a SW direction. The accompanying lead anomalies extend less than 3 km from the deposits. There is a vague westerly trail of Zn anomalies from JJM4 and JJM5 but all the samples nearest JJM5, JJM1, and JJM10 are barren hence it is doubtful that they would have been found by this method alone.

Chaumont (1970a) noted that the Um Gheig lead-zinc deposit was discovered by geophysical methods and suggested their application at Jabal Dhaylan. In 1970 Guillon and Lafon made a ground geophysical survey at Jabal Dhaylan using magnetic, Turam, resistivity and self-potential methods for delineating the basement and following mineralization. The results were not very successful although Turam and resistivity appeared to be useful for delineating faults. An earlier airborne magnetometer survey (Sheets 306 to 308, 1:50,000 scale, DGMR, 1965-66) revealed the major NNW fault along the Precambrian horst at Jabal Dhaylan.

The writer attempted to use a new geothermometric method of prospecting based on phytoclast alteration (discussed in section 2.5.4) at Jabal Dhaylan however thermal alteration haloes were not found around the mineralized localities.

1.2.4

MINERALIZATION IN THE ARABIAN COASTAL PLAIN

MAQNA
 ASII SHARMA
 SHARM AL BADU
 AZLAM
 WADI THALBA
 JABAL YAHMUM
 JABAL DHAYLAN
 JABAL AL BUWANA
 MERATABA
 RERGA
 WAZRA
 UMM GERAD
 WADI FATIMA
 JABAL MARKUB
 AS SARAT
 AL QAHMAH
 AL WAGHRA
 JIZAN
 JABAL TURF
 WADI KHULAB
 YEMEN

MAQNA

The best exposures of Miocene sediments in the entire coastal region of Arabia are in the Maqna area. Here also are lead-bearing barite veins, and concentrations of sulphur, gypsum and oil.

The galena-bearing barite veins were found by BRGM geologists at $34^{\circ}54'23''$ E, $28^{\circ}25'23''$ N in 1969 (P. Dadet, pers. com.). One specimen contained 18% Pb, 1.5% Zn and 112 ppm Ag. The veins are a few hundred metres long and up to half a metre wide. They trend N to NNW and intersect subreefal Miocene limestone (Dadet, pers. com.). The vein minerals consist essentially of barite, calcite, quartz, galena, cerussite and hydrozincite. The veins lie along the extrapolated extension of a fault of Miocene age (fig. 41).

Sulphur forms a small deposit, 20 km due south of Maqna, in a broad breached anticline that strikes north-east. According to Bodenlos (1969) it occurs as crystals and crystal clusters in secondary gypsum veinlets within a bed of sandstone. The sulphur observed by the writer however

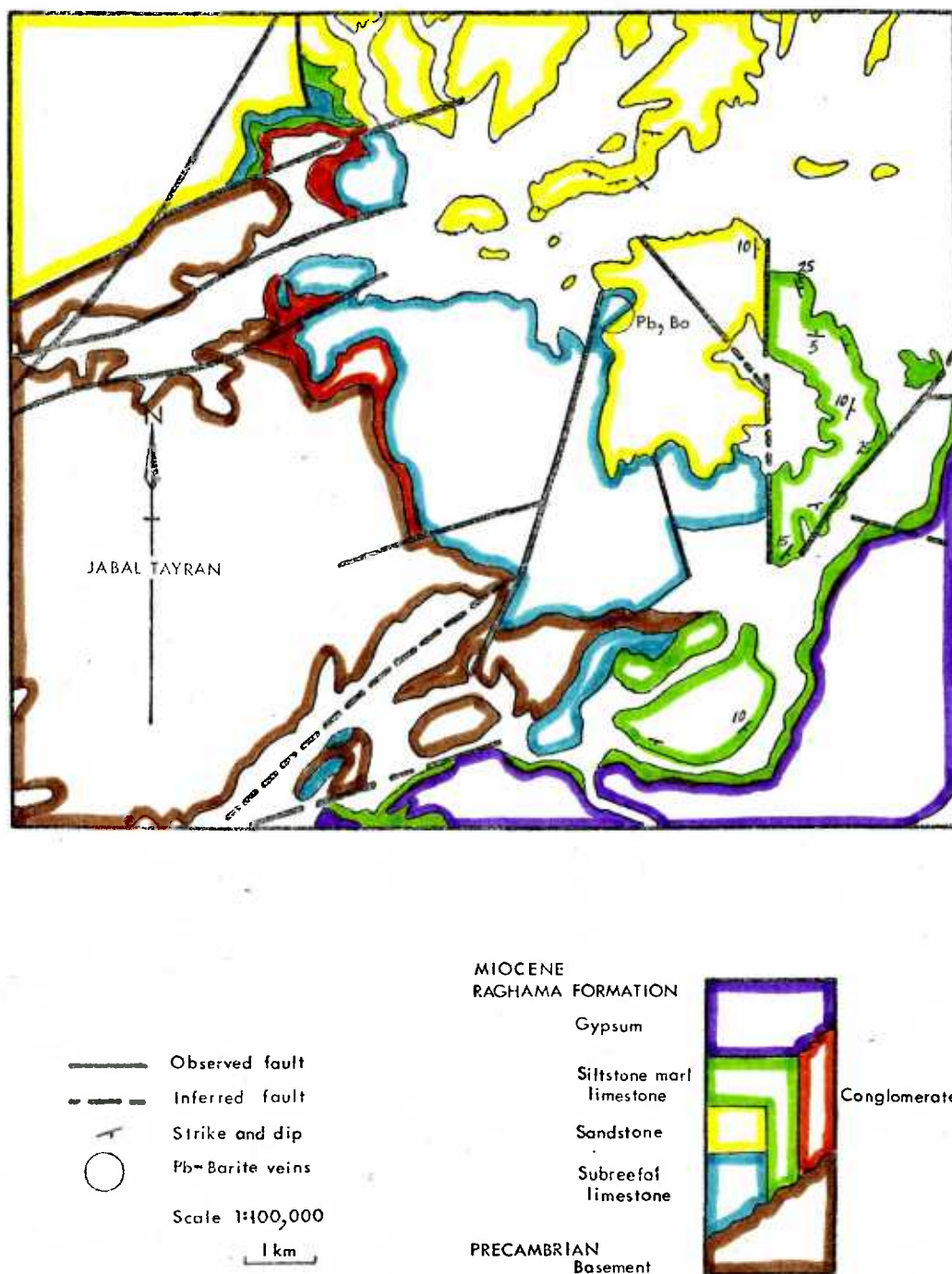


FIG 41. GEOLOGICAL MAP OF AREA CONTAINING GALENA-BARITE VEINS 15 km N80°E FROM MAQNA (modified from confidential source)

occurs as massive irregular veinlets within the bedded gypsum immediately underlying sandstone. The amount of visible sulphur at Maqna is discouragingly small however as the locality is known as Jabal Kibrit (sulphur mountain) it is probable that most of the accessible surface material has been excavated.

The gypsum in the Maqna area is finely layered and pure, and occurs in beds up to 30 m thick with an aggregate strike length of 74 km. Although faults are common in this area folds are mild with dips rarely greater than 25° . A dolomite associated with the gypsum contains widely disseminated pyrite crystals (Bodenlos, 1969).

A shallow salt dome is believed to occur "just north" of Al Lizan, $35^{\circ}02'E$, $28^{\circ}09'N$ (Skipwith, in press).

Oil seeps have long been known near Ash Shaykh Hamayd. The off-shore Barquan No. 2 well encountered oil and gas south of Al Lizan (Ahmed, 1972).

ASH SHARMA

In the Ash Sharma area at $35^{\circ}20'26''E$, $27^{\circ}57'13''N$, 45 km NNW of Duba, BRGM geologists found secondary zinc mineralization in geodes associated with fault fissures in grey coral limestone at the top of a 100 m high butte near the Precambrian basement (P. Dadet, pers. com.). The mineralization is associated with oxides of iron and manganese, and the best assay value is 11.25% Zn, 0.1% Pb. This occurrence was not examined by the writer however a pulverized specimen made available by BRGM was found to consist essentially of dolomite.

SHARM AL BADU

Near Sharm al Badu, 10 km south of Al Muwaylih and 32 km north of Duba is the sulphur prospect of Tuwayyil al Kibrit ($27^{\circ}36'N$, $35^{\circ}31'E$). Bodenlos (1969) did not find sulphur in this area and he reported that Bogue found only traces of sulphur. At the base of the hill of Tuwayyil at Kibrit the writer found a friable argillaceous sandstone with a greenish yellow matrix containing halite, sulphur and smectite. It seems likely that minor exploitation has removed the most obvious surface material. The outcrop forms a small flat-topped conical hill closely resembling the Tuwayyil al Kibrit south of Al Wajh. The base of the hill consists of green shale and siltstone with veinlets of gypsum probably derived from underlying evaporites. Gypsum beds up to several metres thick form out-

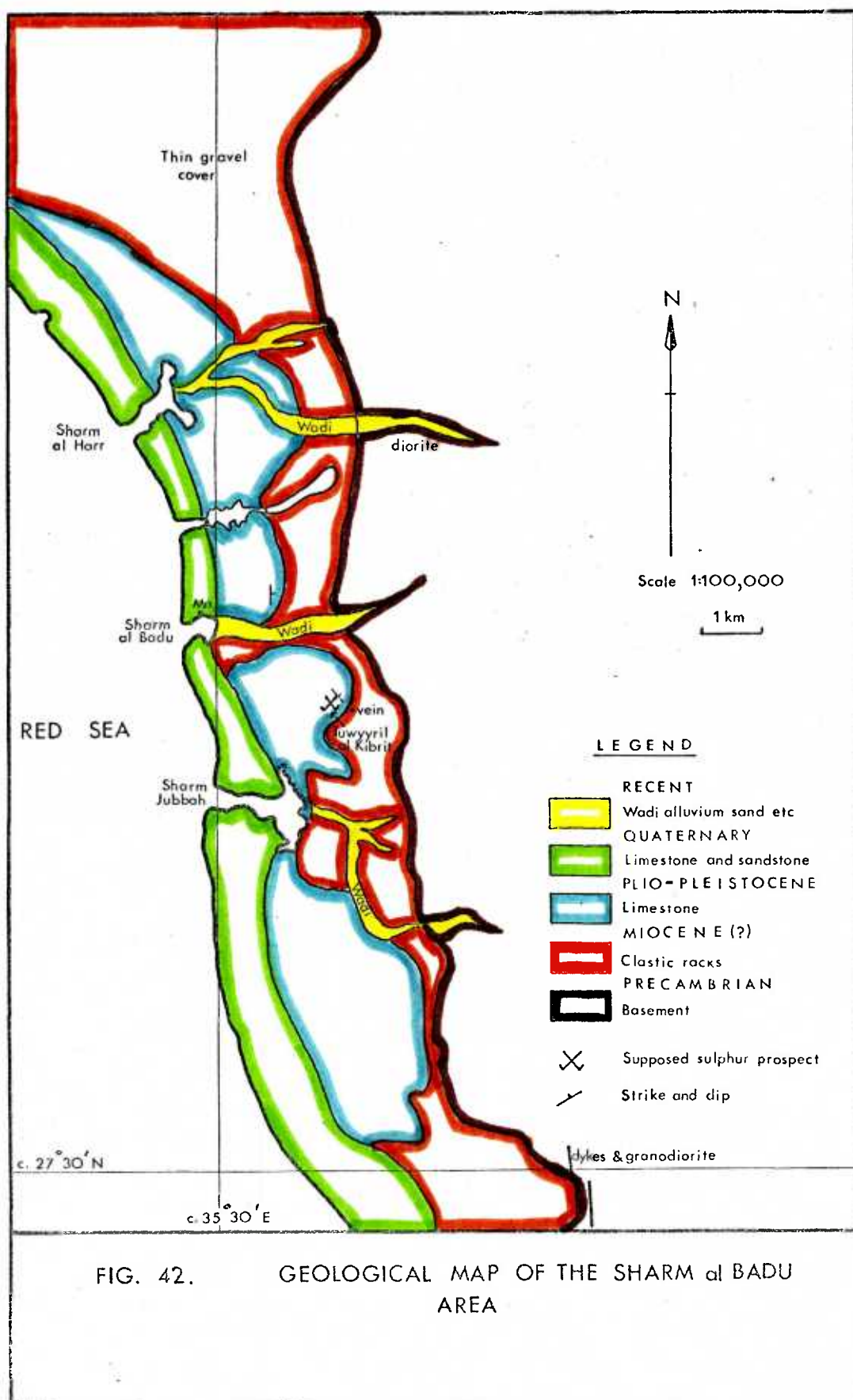


FIG. 42.

GEOLOGICAL MAP OF THE SHARM al BADU AREA



Fig. 43. Current bedded magnetite quartz sandstone near Sharm al Badu.



Fig. 44. Algal reef near shore line in Sharm al Badu area with partial replacement by manganese oxide.

crops north and south of the hill. Overlying the shale is about 10 m of conglomerate sandstone followed by 20 m of scree above which is about 20 m of limestone. Less than half a kilometer east of Tuwayyil al Kibrit (fig. 42) in Tertiary argillaceous sandstone is a poorly exposed NNW trending vein of iron and manganese oxides associated with common opal. Some of the opal contains extremely finely disseminated goethite as well as microveins of chalcedony and quartz.

In Sharm al Badu is a thin succession of marly limestone with algal reefs followed by conglomerate and magnetite sandstone and more marly reefoid limestone. The magnetite sandstone (fig. 43) is lenticular and has a maximum thickness in this area of about half a metre. Very similar magnetite sandstone was found at various localities almost as far south as Yanbu. Although it appears to be a massive magnetite bed in places, it is characteristically current-bedded in more siliceous parts. An algal reef (fig. 44) close to sea level has been partially replaced by todorokite and pyrolusite on its ^{landward} ~~seaward~~ side.

AZLAM

The main landmark in this area is Azlam Castle (Qal at al Azlam), 50 km SSE of Duba and 90 km NNW of Al Wajh. BRGM geologists found a zinc prospect about 8 km NW of the ruined castle at approximately $35^{\circ}56'28''$ E, $27^{\circ}04'43''$ N (P. Dadet, pers. com.). This was unknown to the writer at the time of his visit to the area however he found veinlets of barite and traces of copper in the same general locality. This was unexpected as the 1:500,000 geological map I-204A by Brown et al (1963) depicts this area as a Quaternary terrace deposit. Instead, it consists of a well exposed Precambrian inlier, probably a horst, of granite, hornfels, and slate overlain by Miocene(?) shale, sandstone and limestone. In places the limestone is "plastered" on the granite with an initial dip of 20° . A specimen collected by BRGM geologists from the limestone near a large WNW fault assayed 11.25% Zn and 0.1% Pb (P. Dadet, pers. com.). A pulverized portion of this received by the writer consists of dolomite, calcite and hemimorphite.

The barite occurs as small veins up to a centimeter wide in the limestone but is not apparently associated directly with other mineralization.

The copper traces occur in gravelly sand as cupriferous sandstone float pebbles which were traced to a notch between the highest hill (limestone-plastered granite) and its adjacent hill of hornfels on the seaward side of the jabal. Although the source was not found the cupriferous sandstone float appears to be a facies of the barren sandstone beyond the notch. XRD indicates that the copper mineral which occurs in the matrix of the sandstone is atacamite. BRGM chemists using AAS found typical fragments to contain 12.5% Cu, 0.1% Zn and 600 ppm Pb. A search was made for sulphides in a polished section of this rock but those found (pyrite, chalcopyrite?, bornite?) occur only as tiny specks in the quartz grains of the sandstone. The sparse magnetite in this sandstone incidently has textures similar to those in the Wadi Sawawin magnetite.

This locality resembles Jabal Dhaylan in several respects, the cupriferous sandstone, for example, superficially resembles that at JJM6. Also the limestone at this locality is manganiferous in places.

About 13 km south-east of this locality and 5 km east of Azlam Castle Own Sharif of the DGMR found minor sulphur mineralization. This sulphur cements a coarse iron stained sandstone in a small outcrop in which it overlies gypsiferous shale. The sandstone also contains traces of pyrite. This locality is near a large north-westerly fault bounding the westerly side of the Wadi as Sirr rift valley. About 2 km north-west of the sulphur occurrence is a basalt dyke. The copper-zinc-barite locality also is roughly in line with this fault zone.

WADI THALBA

The writer found manganese mineralization near the northern bank of Wadi Thalba at approximately $36^{\circ}10'E$, $26^{\circ}43'N$, 55 km NNW of Al Wajh (fig. 45). It appears to follow the subhorizontal bedding of the coarse grained sandstone which it infills and replaces. It is about a meter thick but only a decimeter of this is pure pyrolusite. The quartz grains in the less intensely mineralized zone are cracked and corroded and micro-veined by pyrolusite. The cement besides pyrolusite consists of goethite and calcite. A vein of pyrolusite about a decimeter wide and trending NNW was also found at this locality.

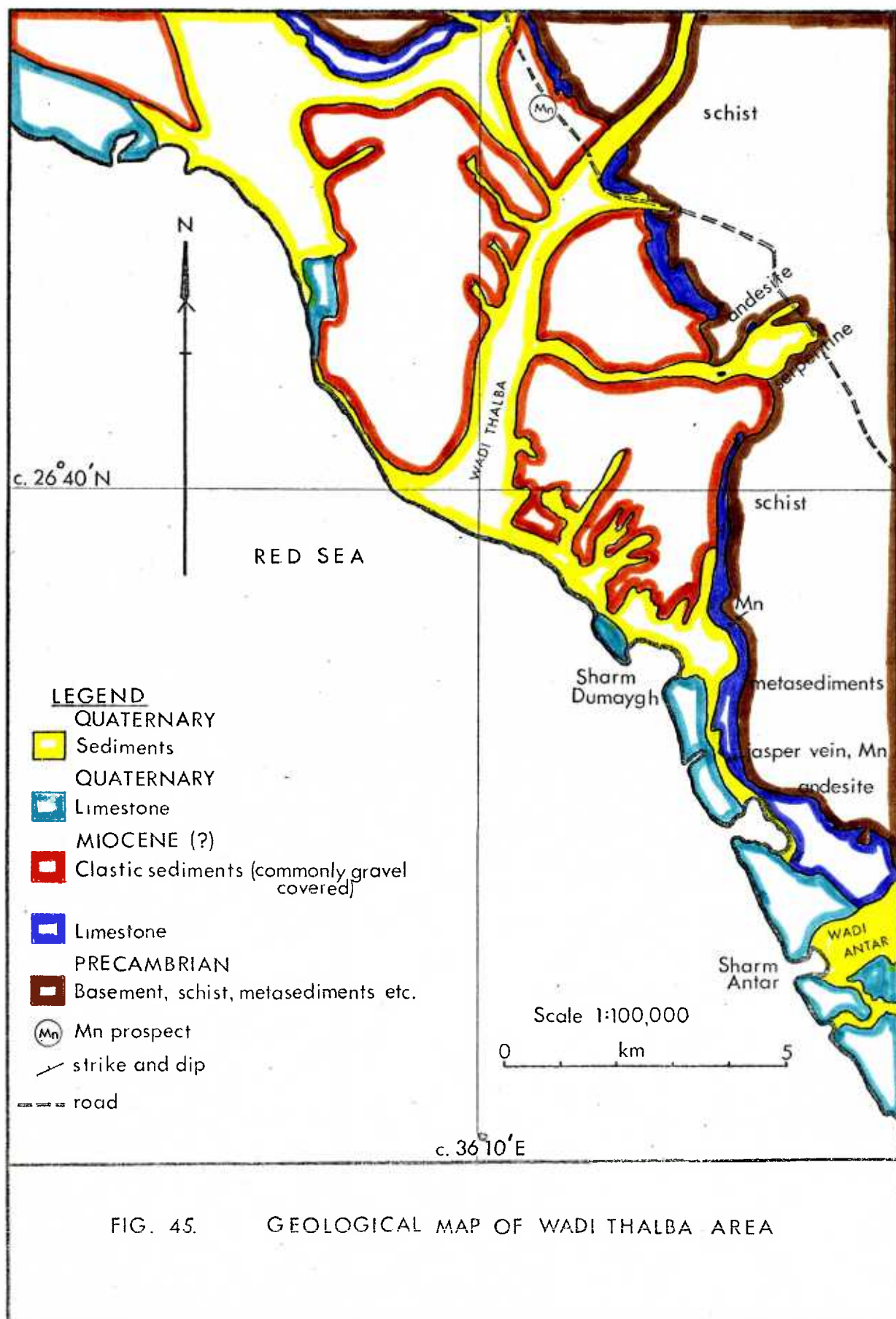


FIG. 45. GEOLOGICAL MAP OF WADI THALBA AREA

Between Wadi Thalba and Jabal Dhaylan limestone commonly has traces of manganese oxides either as disseminated spots or as tiny veinlets associated with calcite crystals and geodes.

About 10 km SSW of the above locality (fig. 45) manganese oxide is associated with calcite crystals in vugs in black crystalline limestone. A further 2 km SSW the limestone contains a brown jasper vein with vugs of goethite and manganese oxide. Nearby the associated subhorizontal Tertiary(?) conglomeratic sandstone contains pyrolusite cement and is very similar to that near Wadi Thalba.

About 11 km further south and about 35 km NNW of Al Wajh at the beach of a small wadi the writer found an extensive area of tarry sand. Similar material was not found in beaches to the north and south. The strata near this beach are rather faulted and contain subvertical veinlets of salt up to 15 mm thick. Ahmed (1972) recorded an oil seep less than 20 km south of this locality near Ras Marabit however the writer failed to notice any sign of oil near Ras Marabit during the 1970 field trip.

JABAL YAHMUM

Tuwayyil al Kibrit near Jabal Yahmum 30 km SSE of Al Wajh has been noted by Brown et al (1963) as a sulphur prospect. Bodenlos (1969) however, found only traces of light yellow powder in clusters of selenite crystals and in secondary "gypsite". He also recorded that Boque (unpublished report) found a few needles of sulphur south-west of Tuwayyil al Kibrit.

Tuwayyil al Kibrit is a prominent conical hill about 40 m high and is similar to the Tuwayyil al Kibrit at Sharm al Badu. The top of the hill consists of subhorizontal nonfossiliferous limestone that is porous and vuggy in some places and finely layered in other places. The hill slopes are scree covered. A large area around the hill consists of contorted gypsum (fig. 46) which is probably derived from anhydrite, the disrupted bedding being due to expansion. Massive white soft anhydrite occurs with the gypsum in places. Outcrops nearby show that the gypsum is overlain by subhorizontal sandstone. The only sulphur found by the writer occurs as disseminated powder in the clayey matrix to a gypsum veined marly sandstone in an outcrop a few hundred meters south-west of the hill. Own Sharif of the DGMR however found a new minor occurrence of sulphur as cement to a coarse grained iron-stained sandstone overlying gypsum and salty siltstone about 2 km north-west of Tuwayyil al Kibrit. Both sulphur

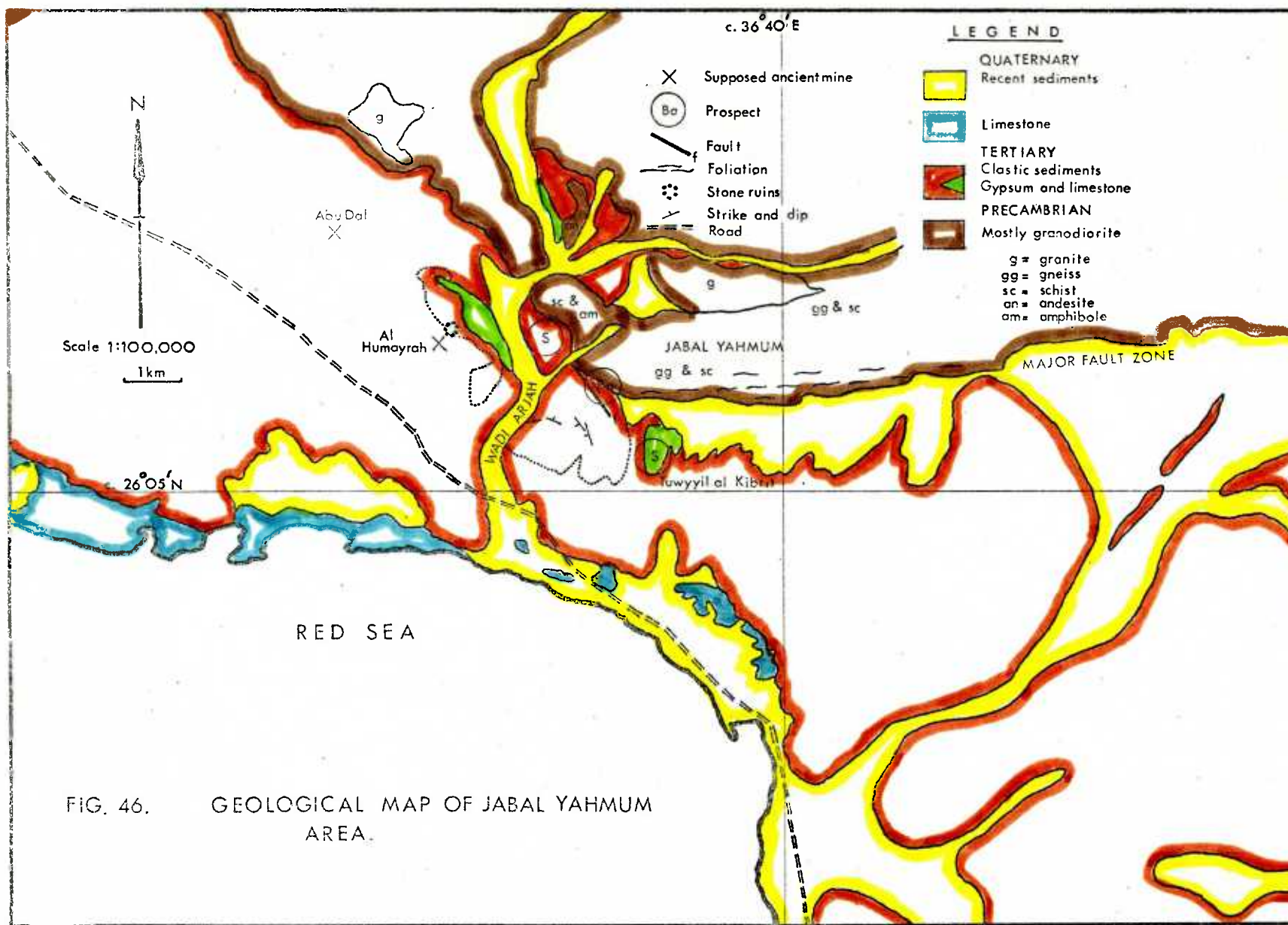


FIG. 46. GEOLOGICAL MAP OF JABAL YAHMUM AREA.



Fig. 47. Barite-jasper vein in siltstone near Jabal Yahmum, viewed from south.



Fig. 48. Specimen of barite and jasper from vein shown in fig. 47.

occurrences appear to lie along extrapolated traces of a fault zone (fig. 46).

Faint traces of probable hydrocarbons were observed in very thin vertical veinlets in Quaternary fossiliferous marly limestone about 3 km south of Tuwayyil al Kibrit.

About one and half kilometres north-west of Tuwayyil al Kibrit the writer found a swarm of beautiful small barite veins (figs. 46, 47, 48) in both the Precambrian basement granite gneiss and in the overlying Cenozoic soft brown sandstone. The veins have a maximum width of 0.15 m and are subvertical with a northerly trend parallel to associated joints. They are close to a large easterly trending fault zone and to a smaller north-westerly fault. The sandstone strikes N 15° W and dips 10° S, and rests directly on the gneiss except where a thin basal conglomerate has locally developed. It consists essentially of quartz, mica and lesser feldspar, 0.05 to 0.10 mm diameter in a weak matrix of calcite with lesser goethite and gypsum. The quartz grains appear to be corroded. The contact of the sandstone with the barite is abrupt even in thin section. Thin veinlets of anhydrite with lesser gypsum and halite occur along bedding planes in the sandstone near the barite veins. The large barite veins have axial zones of brown jasper. This consists of quartz and lesser chalcedony with zoning suggestive of colloform layering prior to crystallization. The goethite occurs as disseminated discrete bodies discussed further in 2.3.2. The jasper is also locally cavernous and contains concentrations of iron and manganese oxides (goethite and cryptomelane?). The veins appear to be thicker in the sedimentary superstructure than in basement. The best exposed vein (fig. 47) tapers upwards. These veins generally are very similar to those at Merataba and Umm Gerad.

Less than one kilometer north-east of the barite-veins on the western side of Wadi Arjah the sedimentary sequence is well developed and gypsum beds at outcrop are about 40 m thick. Also in this western part of the area are two prospects (or ancient mines), Abu Dal and Al Humayrah, noted by Brown et al (1963). Details of the prospects, such as what minerals were located, are lacking. The writer made a diligent search for these prospects but found nothing except a stone ruin, possibly a parapet, near the edge of a high cliff. Local people apparently do not know these prospects. They are in a barren gravel plain (terrace deposit), elevated

and slightly dissected. At the time of examination the area looked most unpromising however later experience at Wazra showed that mineralization can occur in such an area.

Between Jabal Yahmum and Jabal Dhaylan is Jabal Karkuma which Dadet et al. (1970) noted as a diaper structure.

JABAL DHAYLAN

Jabal Dhaylan is described separately in section 1.2.3.

JABAL AL BUWANA

Jabal al Buwana situated about 40 km south of Umm Lajj is somewhat similar to Jabal Dhaylan. It consists of a Precambrian basement horst of granite and amphibolite, within a NNW trending graben complex, partially overlain by Miocene(?) arkose, shale and limestone and Quaternary(?) and Miocene(?) gypsum. It was possibly an island before recent uplift (fig. 49).

Gypsum forms low outcrops over more than 15 sq km south of the Jabal and 2 km NE of Sharm Mahar. Despite the similarities with Jabal Dhaylan no lead, zinc or copper mineralization was found.

At the coast is a thick marly back reef Quaternary limestone overlying gypsum and green shale. This forms a fault block parallel to the coast, with a cliff on its landward side. Along the base of this cliff oxides of manganese have replaced the more porous parts of the limestone (fig. 50). This porous limestone is rather salty. The manganese occurs as black powder, as tiny black crystals and as fine soft dark purple needles replacing the tiny crystals. XRD indicated that it is cryptomelane possibly with some chalcophanite. XRA detected the following elements, in order of abundance, Mn, K, Zn, Ca, Fe, Si and Al. An acid residue of the limestone however has an XRD pattern corresponding to todorokite.

South of Jabal al Buwana at Marsa Tayyib Tusun is an area of sabkha overlying limestone. The black mud below this sabkha and above the limestone is fetid and contains traces of pyrite.

MERATABA

Merataba is an area of ancient mining activity approximately 40 km NW of Yanbu and about 100 km SE of Umm Lajj. Slag heaps were relocated by P. Dadet (pers. com.) and Marchessau (1969, unpublished notes) using



Fig. 49. Jabal al Buwana, a Precambrian horst surrounded by Cenozoic sediments.
viewed from north-east.

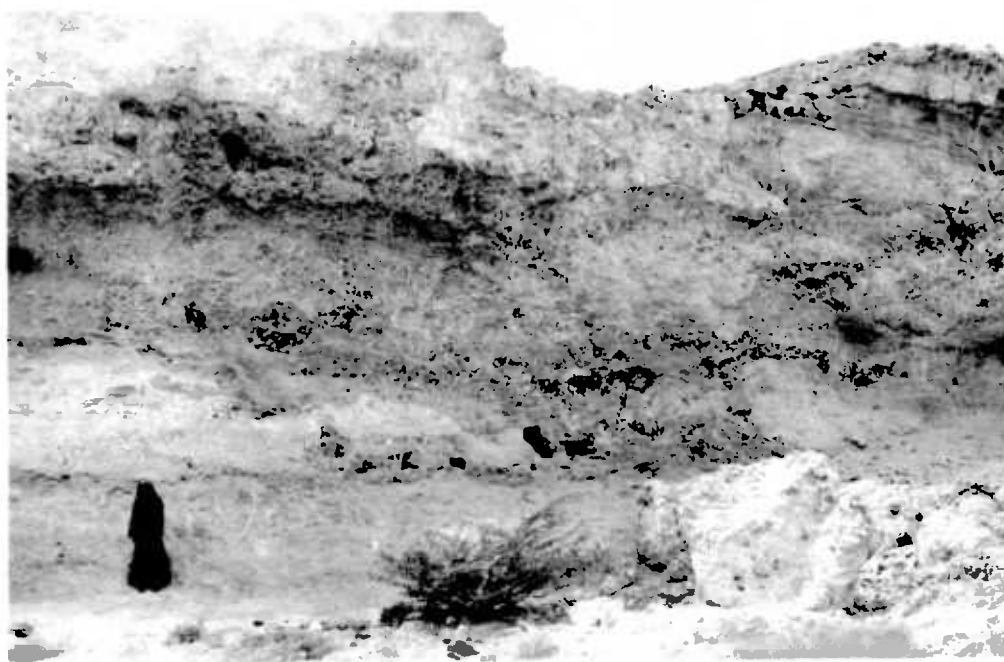


Fig. 50. Outcrop near Sharm Mahut, Jabal al Buwana area, showing manganese oxide
mineralization (black) in early limestone.

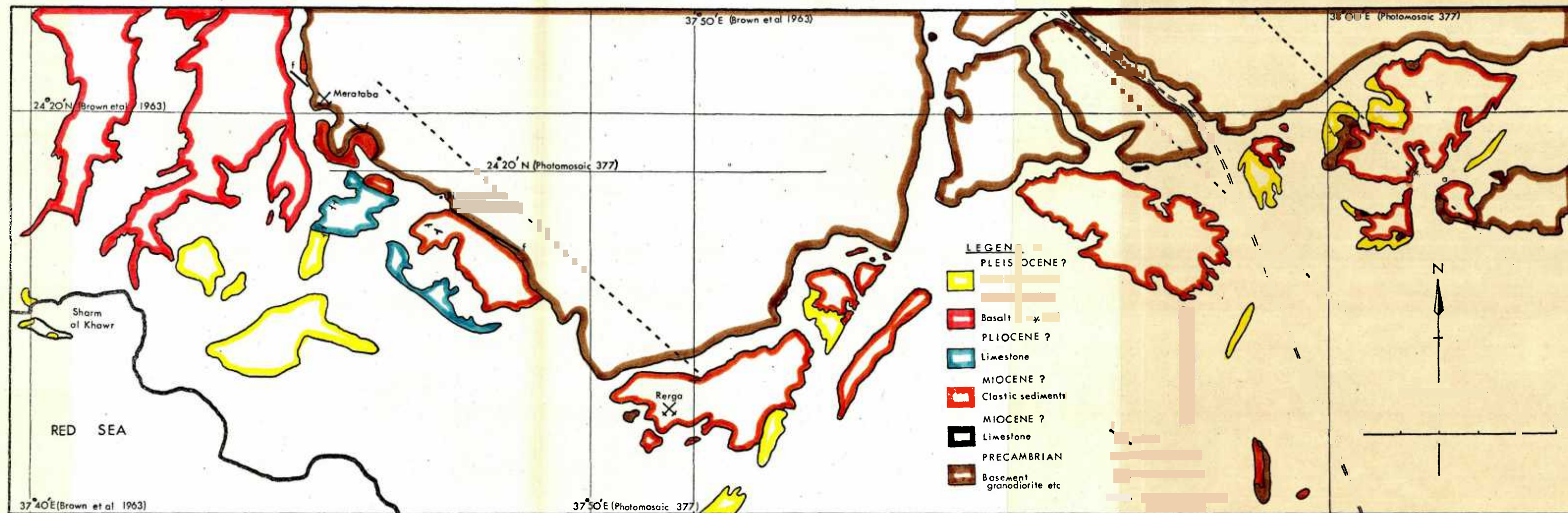


FIG. 51.

GEOLOGICAL MAP OF THE MERATABA-RERGA-WAZRA AREA
(Modified from incomplete maps by J. Marchessau, BRGM 1969)

information from an unpublished report by G. A. Shaw dated 7-1-1936. Dadet and Marchessau studied the region from Sharm al Khawr to Yanbu including the areas at Merataba, Rerga and Wazra. Fig. 51 is a map of this region based largely on Marchessau's mapping at 1:50,000 scale with additions and modifications by the writer.

The mineralization at Merataba is considered to be of Cenozoic age because of its similarity with that at Jabal Yahmum and because of new lead isotope data (section 3.2.4).

The country rock consists of migmatized amphibolite in the south; sheared aplite, quartzite, schist and gneiss in the central and northern part; and granite in the far north. The mineralization consists of veins of barite with or without brown jasper. The veins are from several millimetres to a metre thick, and up to several hundred metres long. They generally have a regular shape with parallel sides and are generally sub-vertical with an easterly trend. It is probable that these veins are mineralized tension fractures extending from a north-west fault zone which passes through Rerga. The Precambrian - Cenozoic contact more or less follows this line and two faults were found in this zone.

The barite in the smaller veins is apparently quite pure but that in the larger veins contains clusters of galena grains and very thin veinlets of galena (fig. 52). The galena is partially replaced by vitreous transparent cerussite, rare cupriferous galena, chalcocite, covellite, pyromorphite and mimetite(?). The vitreous cerussite is replaced by white translucent cerussite (as at Jabal Dhaylan) and rare pink hydrocerussite. The barite occurs as layers of radiating sheaths as at Jabal Dhaylan, Jabal Yahmum and Umm Gerad.

Filling weathering pores in the barite and parts of the jasper is a rare beautiful sky blue vitreous translucent mineral for which the XRD pattern corresponds to boleite (table 6).

The problem with this identification is that Dana (1932) described boleite as indigo blue. As it is a silver-bearing mineral ($9 \text{ Pb Cl}_2 \cdot 8 \text{ CuO} \cdot 3 \text{ AgCl} \cdot 9 \text{ H}_2\text{O}$) it was considered that the colour might deepen with exposure to light. Freshly broken material was exposed to sunlight for six months but no change became apparent. An attempt was then made to



Fig. 52. Slab of barite with galena and cerussite from Merataba.



Fig. 55. Specimen of barite vein (10 mm maximum thickness) in coarse arkose from Wazra area.



Fig. 53. Ancient mine entrance, filled with hay, at Merataba, viewed from the west.



Fig. 54. Ancient mining area near Wadi al Wazra.

Diffractometer Data		Camera data		Published data	
Merataba specimen (smear mount)		Merataba specimen (fibre mount)		standard boleite (Waldo, 1935)	
dÅ	I	dÅ	I	dÅ	I
5.10	40	5.06	50	5.10	40
-		-		4.86	20
4.40	100	4.44	100	4.40	100
4.26	45(quartz?)	-		-	
3.82	100	3.81	60	3.83	100
3.71	25	-		-	
3.49	35	3.52	20	3.50	80
3.34	55(quartz?)	3.33	50	3.34	20
3.12	40	3.11	30	3.13	100
3.05	25	-		-	
2.95	15	2.96	10	2.95	40
2.84	15	2.82	10	2.83	60
2.71	65	2.70	40	2.69	100
2.56	25	2.57	30	2.56	80
-		2.46	10	2.47	20
2.39	35	2.38	20	2.32	100
-		-		2.15	20
-		2.11	20	2.12	20
-		-		2.07	20
-		-		2.02	20
1.99	15	1.99	20	1.99	80
-		1.92	10	1.90	20
-		1.86	10	1.85	60
-		1.80	20	1.80	60
-		1.73	10	1.75	60
-		1.67	20	1.67	40
-		-		1.61	20
-		1.53	20	1.53	40

Table 6 XRD data on standard boleite and Merataba boleite.

check on the composition by using the dichromate microchemical test but the boleite would not dissolve in warm nitric acid. A specimen was then polished and subjected to electron probe microanalysis by G. Steed and N. Wilkinson of the Mining Geology Department. The results are :-

Pb 13.1%
Cu 10.4%
Ag 3.9%
Cl 10.4%

These low values, which were not increased with experimental repetition, are apparently due to surface damage of the specimen by the electron beam, to the inability of the instrument to detect elements of low atomic number e.g. oxygen, and possibly to loss of material by volatilization. Although the results are impaired for quantitative study they do nevertheless appear to show that the Merataba boleite when corrected on the assumption that the analytical results are proportional to the elemental concentrations has a higher copper content and a lower lead content than Dana's boleite. This could explain the less violet and more green pigmentation

	Pb	CuO	AgCl	Cl	H ₂ O	Total
Merataba boleite as analysed	13.1	12.5	5.2	9.2	n.d.	40.00
Merataba boleite "corrected"	31	30	12	22	5	100
Dana's boleite	49.73	16.98	11.47	17.02	4.8	100.00

of the Merataba boleite. Another possibility however is that the colour is due to a high water content. Insufficient material was available to measure the water content directly however it was possible to observe a few tiny grains on a heating stage under a microscope. Up to 250°C some grains became a deep green while others became only a greenish blue. At 250°C blackspots formed on the surface of the grains and at 265°C all grains became totally black due to the growth of the spots. At 275°C the grains began to melt. As the grains did not tend toward an indigo blue colour during heating it seems unlikely that the colour difference is due to a higher water content. It seems rather to be due to a compositional difference which suggests that this mineral is a new variety of boleite or a new mineral isostructural with boleite. Unfortunately this enquiry could not be continued due to a lack of material. The specimen in which the boleite was found came from the ancient workings shown in fig. 53. A green mineral associated with the boleite and which is possibly its weathering product appears to be atacamite but its XRD pattern is rather poor.

The jasper is generally massive and homogeneous though in places it contains small irregular cavities lined with chalcedony. A narrow vein of barite which thickens downward adjacent to a fault or joint is visible at the northern side of the portal to the ancient mine shown in fig. 53.

The jasper in contact with this barite appears to be younger than it and contains fragments of the barite vein. The jasper also has a faint foliation, reminiscent of bedding, dipping inward from both sides of the vein and horizontal in the central part. There is no rubble in the vein and no sign that it is a weathering phenomenon (it is discussed further in sections 2.3.2, 3.1.4, 3.2.1.). Smaller veins elsewhere suggest a close time relationship between the jasper and the barite.

Only one ancient mine (fig. 53) was found by the writer at Merataba however extensive slag heaps (perhaps covering 10,000 sq m) in the nearest flat country suggests that mining was extensive in this area. The mine photographed is now a hay store used by local inhabitants who unfortunately were not visible during the writer's visit. In their absence the writer was unable to remove much of the hay in order to gain access to the workings.

RERGA

Rerga, another ancient mining area, is situated about 11.5 km SW of Merataba and about 30 km WNW from Yanbu (fig. 51). At this locality are two slot-like ancient mines about 100 m apart in weakly cemented Tertiary sandstone and conglomerate. The mines obviously followed two parallel straight vertical veins for well over a hundred meters in a N 35° W direction. The lower parts of the mines have collapsed and the walls are bulging inwards. Another smaller mine nearby trends N 10° W. The walls of these mines are clean and, other than the excavations themselves and some tailings and slag, there is nothing to suggest that mineralized veins previously existed in this area. A sample from this locality assayed 4.3% Pb, 4.4% Zn and 1100 ppm Ag (P. Dadet, pers. com.). Marchessau and Dadet located the mines from data given by Johnson and Trent (1966).

Tailings from the mines consist of coarse sandstone, brown jasper and rare fragments of pyromorphite. Pyromorphite is associated with both sandstone and jasper and in some specimens with manganese oxide (coronardite?). Traces of galena were found with the pyromorphite, in the sandstone and in the jasper. In polished section it was observed to be partially replaced by rare covellite. Fragments of granite around the ancient mines suggest the basement may have been reached.

Using one of the staining reagents described in section 1.2.3., Marchessau (1969, unpublished notes) also found secondary zinc mineralization in Plio-Pleistocene(?) calcareous sandstone breccia 2.5 km ENE of the ancient mines. The writer was unable to re-locate this mineralization. One of Marchessau's specimens assayed 1.5% Zn (Dadet pers. com.).

The general stratigraphy of the Merataba-Rerga-Wazra region as elucidated by Marchessau (1969, unpublished notes) is :-

Quaternary	- Coral reefs, basalt lavas, alluvial fans, terraces, sabkhas and other sediments.
Plio-Pleistocene(?)	- Calcareous, shelly limestone and reef limestone.
Upper Miocene(?)	- Thick coarse conglomerate, sandstone and shale.
Lower Miocene(?)	- Limestone with coral and geodes, lesser shale and gypsum.
Precambrian	- Basement.

WAZRA

The ancient workings near Wadi al Wazra are approximately 20 km ENE of Rerga and 27 km N of Yanbu (fig. 51). They were located by Y. Berton and P. Dadet (Marchessau, 1969, unpublished notes) in 1967 when they found galena and vanadinite crystals with black sand as a concentrate in a prospecting dish.

The geology of the ancient mining area is perplexing because of poor exposure (fig. 54) and the irregularity of rock boundaries. The basement granite appears to have a very irregular surface (Marchessau, 1969 unpublished notes) and the overlying arkosic calcareous conglomerate appears to range from coarse calcareous arkose to fine grained massive limestone to localized boulder beds. The stratigraphy has been outlined in the preceding section on Rerga however the "Lower Miocene" limestone is very thin and locally is absent. The "Upper Miocene" conglomerate also is very thin near the ancient workings.

The quantity of slag on the flats near the conglomerate appears to be greater than can be accounted for by the apparently shallow excavations. The excavations observed by the writer are limited to six pits at various elevations in a line trending N 20°W parallel to vertical joints in the granite basement, and five pits in a northerly line to the west of the six pits.

The mineralization consists of galena, secondary lead and zinc minerals, manganese oxide, jasper and barite. The barite occurs as irregular veins up to 10 mm or more thick in arkose (fig. 55). The galena occurs as veinlets in granite and arkose, and as nodules in massive crystalline limestone. Within the jasper it occurs as grains in layers parallel to the layering of the jasper. The galena is partly replaced by cupriferous galena(?), chalcocite, covellite, transparent cerussite, translucent cerussite, plattnerite and bright green pyromorphite. Grey and orange barrel shaped crystals (some hopper shaped) were recovered from prospecting dish concentrates but the XRD patterns of these minerals are ambiguous. Probably both vanadinite and mimetite are present. Specks of pale greenish blue mineral in cavities in the jasper are probably diaboletite as found at Umm Gerad. Hemimorphite crystals occur in some cavities in limestone. Some limestone and some arkose is spotted with manganese oxide, and some limestone grades into massive manganese oxide. Soft patches of manganese oxide in the limestone consist of pyrolusite. Associated with this is a mineral having the appearance of dark red glass with a conchoidal fracture but its XRD pattern corresponds to pyrolusite. Dark brown material also with a conchoidal fracture has the XRD pattern of coronardite. XRA shows it to consist of Mn, Pb and much less Zn. Some massive hard aphanitic manganese oxide appears to be hydrohetaerolite though its XRD pattern is poor. XRA shows it to contain Mn, Zn, Si, Pb, K, Ba and Fe in that order of abundance.

The brown jasper here is very similar to that found at Jabal Yahmum, Merataba, Rerga and Umm Gerad. Some jasper however is red (quartz-hematite), some is black (quartz-pyrolusite?) and some grades into white "quartzite".

Gossan and secondary goethite are common in this and other areas but as they are superficial weathering products little attention has been devoted to them.

UMM GERAD

It was understood that the barite veins at Umm Gerad ($22^{\circ}55'N$, $39^{\circ}10'E$) 17 to 25 km north-east of Rabigh and about 150 km north of Jeddah, were the subject of an academic study by Ahmed Al Shanti of the DGMR hence the writer excluded this area from his field work. Al Shanti nevertheless, kindly discussed the geology of the area and gave the writer a collection of specimens.

The discovery of the barite mineralization is attributed to K.S. Twitchell working for the Saudi Arabian Mining Syndicate in 1936. The veins have a maximum width of 2 m and an average width of 0.5 m. The aggregate known length at outcrops is 1,865 m, and minimum reserves to 30 m depth are estimated to be 100,000 tons at over 90% BaSO_4 (Al Shanti, 1970).

The veins occur in a northerly trending horst of Precambrian rocks approximately 3 km wide and 20 km long. They have northerly, north-westerly and northeasterly trends, and the dips range from 60° to 90° . The veins occupy faults and fractures in granite, granodiorite, schists and metavolcanics. The general structure of the vein system is thought to be related to the Red Sea rift system (Al Shanti, 1970).

The mineralization is thought to be of Cenozoic age because of the general structure and because of lead isotope data (Delevaux et al 1967). New lead isotope data (section 3.2.4) and the close similarity of the Umm Gerad veins to the newly discovered Jabal Yahmum veins substantiates this dating.

Although the veins are generally regular they commonly "pinch and swell" along strike and dip, and in some places have angular deflections. Apart from mechanical control on the fracture system, the mineralization appears to be independent of its host rocks (Al Shanti, 1970).

The veins are chiefly composed of barite in white to pink crystalline masses however, as at Jabal Yahmum, the thicker parts of the veins contain lenses of brown jasper ("jasperoid") with associated manganese oxide. The barite ranges from transparent to translucent in thin section and contains scattered grains of galena and chalcopyrite which are commonly altered to malachite (Brobst, 1966). Some "black jasperoid" is also present. Some of the barite is pale green-blue and contains short vein-lets and patches of chrysocolla. The coloured barite grades into white barite and both have identical XRD patterns. The galena in the barite is partially replaced by cerussite, cupriferous galena, chalcocite and covellite (observed in polished section). Associated with the galena and in pores within the barite are rare specks of diableite and possibly atacamite. Some pale brown layered barite has cavities lined with manganese oxide on which are tiny quartz crystals.

The jasper closely resembles that from similar veins along the coastal plain. Brobst (1966) found that it contains anomalous proportions of Be, Pb, Y, B and Fe, and suggested that it has a hypogene origin. Brobst also noted that the "jasperoid" appears to have been deposited in openings in the earlier barite vein material.

Radiometric, gravimetric and magnetic surveys failed to find extensions to the vein system (Al Shanti, 1970) however one vein to the south of Wadi Rabigh was recently found through geological mapping by the DGMR (Abdul. R. Rashid pers. com.) and Jamel Kouther similarly found several barite veins north of those discovered by Al Shanti at Ria Harska.

WADI FATIMA

The bedded iron ore of Wadi Fatima, about 56 km ENE of Jeddah, has probable reserves of approximately 50 million tons of 44 to 50% Fe (Al Shanti, 1966). The discovery of the ore in 1954 is attributed to R. Richter-Bernburgh and W. Schott. Since then the areas have been investigated with 60 trenches, 8 diamond drill holes and 3 shafts.

The ore occurs as two parallel beds 3.3 m thick (upper bed) and 2.0 m thick (lower bed) about 7 m apart, and exposed over an outcrop length of 4.5 km. The ore beds are part of the Shumaysi Formation which consists of sandstone predominantly together with siltstone, shale, conglomerate, chert and tuff. This formation also contains some glauconitic(?) horizons and some shale beds rich in pyrite and carbonaceous material. The strike is northerly and the dip averages 25°E (Al Shanti, 1966).

The Shumaysi Formation is essentially a faulted homocline of Oligocene age resting unconformably on a Precambrian basement of granite, granodiorite, diorite, schist, metavolcanics and metasediments. It is thought to have formed in an essentially shallow marine environment that gradually evolved into a series of fresh water lagoons.

The source of the ore was possibly a ferruginous horizon in the nearby Precambrian Fatima Series (Al Shanti, 1966).

The ore is "oolitic" and consists of goethite and "brown hematite gel" with lesser hydrohematite in a matrix of quartz, goethite, "hematite gel" and lesser clay, apatite, feldspar, gypsum, anhydrite, calcite, siderite and zircon (Al Shanti 1966). The writer did not examine the area directly but Ahmed Al Shanti of the DGMR kindly provided a collection of representative core specimens for study. Typical ore in these specimens consists essentially of goethite "oolites" and "pisolites" in a matrix ranging from quartz sandstone to massive goethite to massive siderite to massive chamosite and to chamosite-quartz sandstone with disseminated siderite crystals. In some places the cement is calcite. Some "pisolites" consist of concentric layers of chamosite and goethite with micro-veins of pyrite. Finely divided carbonaceous matter is abundant in the matrix of some specimens.

JABAL MARKUB

Four hot spring localities lie along a north-westerly line about 70 km long and about 40 km from Al Lith. The localities are at Ad Darakah ($40^{\circ}02' \text{E}$, $20^{\circ}37' \text{N}$), Jabal Markub ($20^{\circ}29' \text{N}$, $40^{\circ}09' \text{E}$), Wadi al Lith ($20^{\circ}27' \text{N}$, $40^{\circ}31' \text{E}$) and Suq al Juman ($20^{\circ}18' \text{N}$, $40^{\circ}39' \text{E}$). At Jabal Markub are perennial springs, one issuing 50 l/min at 70°C , the other issuing 10 l/min at 55°C (Bhutta, 1965a). Both occur in closely jointed granodiorite. Yellow-red-brown iron oxides form crusts on the surface of the water at the outlets, and gypsum and calcite precipitate a "short distance" downstream. The precipitation of the goethite(?) is probably assisted by microorganisms as the iron content of the water is relatively low. The analysis given by Bhutta is :-

Cl	1675 ppm
Ca	475
SO ₄	427
Na	258
Al	82
HCO ₃	99
Mg	3.5
Fe	trace
CO ₃	trace
not determined	700
total solids	3730
pH	6.1

These springs resemble those at Khulab (Bhutta, 1965a).

AS SARAT

Laterite of probable Eocene age underlying the basalt of As Sarat, and overlying Precambrian basement was reported in bulletin number one of the DGM. It is exposed over a large area around Jama ($18^{\circ}00'N$, $43^{\circ}08'E$). No bauxite has been found but the laterite is locally rich in iron and is associated with deposits of kaolin.

AL QAHMAH

Of the coastal beach sand deposits the one so far considered to have the greatest economic potential is situated at $17^{\circ}53'N$, $41^{\circ}45'E$ near Al Qahmar village 40 km south of Al Birk. The beach is over 5 km long and is backed by many dunes. Short (1956) has recorded the reserves as 0.3 million tons of sand with 182,000 tons of strongly magnetic concentrate. The main heavy minerals are magnetite (28 to 32%) and ilmenite (4 to 10%). Some of the magnetite and ilmenite occurs as intergrowths.

Short (1956) classified the beaches of western Arabia into four types:-

1. Those containing quartz, feldspar, magnetite, ilmenite, titanite, garnet, epidote, zircon and rutile derived from granitic basement, mostly in the north.
2. Those containing magnetite, ilmenite, olivine, glass, pyroxene and hornblende derived from basaltic lavas e.g. at Al Qahmah.
3. Those containing garnet, amphibole, biotite, epidote, chlorite, tourmaline and apatite derived from metamorphic basement, mostly in the south.
4. Those containing coral fragments, oolites, and other forms of calcite, derived from coral, limestone and calcareous environments.

Beaches additional to Al Qahmah noted by Short (1956) are at the mouth of Wadi Tiryam; 10 km north of Wadi al Hamd; 2 km south of Umm Lajj, 18 km south of Al Lith, 10 km north of Wadi Hali, and 1 km south of Wadi Rin. The aggregate length of these beaches (excluding Qahmah) is 34 km averaging about 8% magnetite and 5% ilmenite.

The apparent paucity of heavy mineral concentration along the Red Sea coast is possibly due to a combination of low rainfall, a narrow "continental shelf", submarine cliffs, and quiet water conditions associated with carbonate and salt deposition.

AL WAGHRA

Two hot springs, 100 m apart, al Al Waghra ($17^{\circ}02'N$, $42^{\circ}59'E$), 55 km ENE of Jizan, have flow rates of 30 l/min and >180 l/min. and a temperature of approximately $55^{\circ}C$ (Bhutta, 1965b). They occur in NNW fractures in quartz-sericite and quartz-chlorite schists and lie along a NW line. The schists are very altered near the springs and white salts are deposited in the cooler regions around the springs. Quaternary olivine basalt lies within 10 m of the springs and Miocene intrusions (as at Jabal Turf) form outcrops 2 km WSW of the springs. These springs are very similar to those at Khulab.

JIZAN

The town of Jizan ($16^{\circ}54'N$, $42^{\circ}29'E$) is sited on a salt dome forming a hill 3 km long, 2 km wide and 50 m high. The cap rock is a highly deformed part of the Miocene Baid Formation. Skipwith (in press) states the reserves of halite to a depth of 3 m to be 1.5 million tons. Although sylvite, kainite and carnallite are apparently unknown at Jizan a thin bed of potassium salts has been located at the Farasan Islands 75 km to the west. Oil seeps and two salt domes are also known at the Farasan Islands (Ahmed, 1972).

JABAL TURF

Apart from the occurrence of possible magnetite accumulates and a possible pyrite-pyrrhotite-pentlandite association in Tertiary layered gabbro near Jabal Turf (Jebel at Tirf) nothing is known of base metal mineralization in this fascinating area currently under investigation by the USGS.

Jabal Turf ($16^{\circ}57'N$, $42^{\circ}57'E$) is a mountain 540 m high, 45 km east of Jizan and consists of a granite batholith 7 km wide and 25 km long trending NNW along a Tertiary flexure zone. The area consists of micrographic alkaline granite surrounded by a co-magmatic layered gabbro ring dyke which in turn is surrounded by rhyolite containing some dykes of lamprophyre (Gillmann, 1968).

The Miocene Baid Formation consists essentially of volcanic sediments intercalated with acid to intermediate flows. It has been extensively intruded by acid to basic magma some of which is associated with the Jabal Turf igneous complex (Gillmann, 1968).

From Gillmann's mapping it seems that the igneous complex is partly older and partly younger than the Baid Formation however he gave two K/Ar age determinations for it of 191 ± 28 m.y. and 120 ± 6 m.y. (Triassic to Cretaceous). Coleman and Brown (1971) and Coleman et al (1972) nevertheless claimed that the trough shaped layered gabbro was derived from tholeiitic magma about 22 m.y. ago (Miocene). They also noted that the gabbro is cut by post-Miocene basaltic dykes, and that in the late Pliocene renewed movements along the flexure zone tilted the complex toward the Red Sea. Tertiary basic hypabyssal intrusions following a Red Sea trend extend at least from the Yemen border north to Al Lith.

WADI KHULAB

A small island near the mouth of Wadi Khulab, about 35 km south of Jizan, contains disturbed gypsum, limestone and sandstone which probably cap a small salt plug (Heybroek, 1965). Agocs and Keller (1962) noted that pyrite crystals weather out of a vertical oolitic hematite bed on this island. They further noted that probable salt domes also occur at Al Madaya, 25 km SE of Jizan and at the mouth of Wadi Taashshar, 50 km SSE of Jizan.

Upstream in Wadi Khulab at $16^{\circ}45'N$, $43^{\circ}05'E$ (65 km ESE of Jizan) are several hot springs having a total output of at least 500 l/min. (Bhutta, 1965b). These springs issue from the NNW periphery of a closely jointed silicified Miocene andesite dyke. Calcite and gypsum precipitate from this spring water. At the hottest outlet the temperature exceeds $85^{\circ}C$, other outlets average $73^{\circ}C$.

Griveau (1966) dismissed the possibility of economically harnessing these and all other springs in Saudi Arabia for geothermal power. It should be noted however that subterranean geothermal anomalies may be more widespread than present surface features indicate. The temperature at the bottom (3930 m) of Mansiyah No. 1 well, 40 km NNW of Jizan was found to be $177^{\circ}C$ (Ahmed, 1972). This indicates a geothermal gradient of $40^{\circ}/km$ (Griveau quoted $30^{\circ}/km$ as normal).

YEMEN

Five salt domes occur along the Yemen coast between latitudes $15^{\circ}08'N$ and $15^{\circ}41'N$. The most noted of these are at Al Luhayyah, Jebel Guma and At Salif (Heybroek, 1965). Bulletin No. 1 of the DGMR recorded that barite veins cut Tertiary sediments in Yemen but no details were given. Sulphur is mined from Djebel El Esi volcano ($14^{\circ}31'N$, $44^{\circ}37'E$).

1.2.5 DEPOSITS OF THE AFRICAN COASTAL REGION OF THE RED SEA

ETHIOPIA

Volcanic Provinces
 Chercher Copper Occurrence
 Danakil Evaporites
 En Kafala Manganese Deposit
 Mersa Gulbub Barite Deposits
 Fuel Occurrences

SUDAN

Halaib and other Manganese Deposits
 Wadi Halfa and other iron deposits
 Coastal Evaporites

EGYPT

Ras Benas Copper Occurrence
 Wadi Malik, Elba and Wadi Araba Manganese
 Deposits
 Reigha, Ranga and Gemsa Sulphur Deposits
 Um Gerifat and other Iron Ore Deposits
 Anz, Ranga, Wizr, Talet Eid, and Rusas Lead-Zinc
 Occurrences.
 Um Gheig Lead-Zinc Mineralization
 Zug el Bohr Galena Deposit
 Quseir - Safaga Phosphate Deposits
 Hughada - Ras Gharib Oil Field

SINAI

Gulf of Suez Region - Um Bogma Manganese and Copper
 Gulf of Suez Region - Fuel and Gypsum
 Aqaba - Dead Sea Region

ETHIOPIA

Volcanic Provinces

The Cenozoic mineralization in Ethiopia is associated with volcanic and evaporitic rocks both of which are controlled by rifting. Although Jelenc (1966a) described the volcanic "sulphur province" as a continuous belt between Lake Chamo and Massawa channel it is better described as two separate provinces; one belonging to the south-westerly East African - Gulf of Aden rift systems, the other belonging to the north-westerly Afar - Red Sea rift system. Localities of active volcanoes are given in section 1.3.1. Kalinite occurs at Boina and Gabres volcanoes in the south-west province and sulphur is mined from Dofane volcano (Richard and Neumann van Padang, 1957).

The main sulphur bearing volcanoes in the north-west province are Kebrit Ale, Gada Ale and Dallol. Jelenc (1966a) noted that iron sulphides commonly accompany the sublimed sulphur in these volcanoes and that apatite and "salts" (kalinite?) also occur in the volcanic "sulphur province". A salt dome is adjacent to Gada Ale and the walls of some faults and fissures of the salt dome are covered with apatite crystals several centimetres across covered with octahedral magnetite crystals forming "an important deposit" (Barberi and Varet, 1970).

In the Alu horst fumeroles at 135° to 221°C emit boric, hydrofluoric and hydrochloric acids and deposit abundant sulphur and ammonium chloride. At Dala Filla volcano coquimbite and sassolite ($\text{B}(\text{OH})_3$) are abundant (Barberi and Varet, 1970).

DeKun (1965) noted that pozzuolanic sands occur at Mount Dala volcano (Jalua?) and elsewhere. He further noted that peridotites are found in xenoliths(?) in basalt at Edd ($13^{\circ}57' \text{N}$, $41^{\circ}38' \text{E}$) Kod Ali Island ($13^{\circ}57' \text{N}$, $41^{\circ}49' \text{E}$), Bahar Assoli ($13^{\circ}38' \text{N}$, $42^{\circ}08' \text{E}$), Sagan Valley (5°N , 37°E) and South Adaito Ela ($14^{\circ}28' \text{N}$, $40^{\circ}49' \text{E}$).

Lateritic iron ores related to Cenozoic volcanic rocks are widespread over Ethiopia. Localities given by Jelenc (1966a) include Ghimirh (7°N , 36°E), Nadda ($7\frac{1}{2}^{\circ} \text{N}$, 37°E), Entoto ($9\frac{1}{2}^{\circ} \text{N}$, $38\frac{1}{2}^{\circ} \text{E}$), Chercher (9°N , $42\frac{1}{2}^{\circ} \text{E}$), Aira (9°N , $35\frac{1}{2}^{\circ} \text{E}$) and Hamasien (15°N , 39°E).

Jelenc (1966b) noted that a small apatite-poor carbonatite occurs near the township of Harrar. Many hot springs also occur in the volcanic provinces so there is a geothermal power possibility.

Chercher Copper Occurrence

Jelenc (1966a) briefly noted that in the Jurassic Adigrat Series, which is dominated by limonitic quartz sandstone, conglomerate and shale; pyrite and chalcopryrite are associated with cobalt, nickel and traces of uranium in the Chercher area in Galetti and Kunni Valleys ($9^{\circ}00' \text{N}$, $41^{\circ}09' \text{E}$; $9^{\circ}00' \text{N}$, $40^{\circ}59' \text{E}$). A specimen from Galetti Valley assayed 18% Cu, 0.3% Ni and 0.1% Co. The average in this area however is 0.6% Cu Jelenc (1966b). The mineralization occurs as veinlets and lenses in sandstone associated with iron and manganese oxides.

Danakil Evaporites

DeKun (1965) noted that 200,000 tons of salt is mined annually from an area of 325 km² in the Danakil region (14°15'N, 40°18'E) and that the forecast annual production of potash from the same region was 300,000 tons, with initial reserves being greater than 6,000,000 tons. Gypsum is also widespread over this region and elsewhere in Ethiopia (Jelenc, 1966a).

The Danakil region contains the youngest (Quaternary) major marine evaporite sequence in the world. The potash ore being mined is a bed of pure sylvite in a sequence containing halite, carnallite, kainite, gypsum and anhydrite (Holwerda and Hutchinson, 1968). Selective leaching of carnallite before burial is considered to be the main source of the sylvite although there is good evidence that some of the carnallite has been converted to sylvite by water from sub-volcanic intrusions. Hot springs rich in MgCl, KCl, CaCl and NaCl are common in the Dallol area (the main sylvite locality) and some of the salt in this area contains inclusions of basalt (Holwerda and Hutchinson, 1968).

Seven salt domes occur in the islands off Massawa. Although Jelenc (1966a) stated that all the sulphur occurrences in Ethiopia are volcanic sublimation products DeKun (1965) recorded that at Zariga near Massawa gypsum contains 16 to 25% sulphur.

En Kafala Manganese Deposit

At En Kafala, north-west of Dallol Mountain, is a 200,000 year old bedded manganese deposit in a shallow water marine formation. From the basalt base upwards this consists of 1) basaltic tuff-conglomerate cemented with manganese oxide, 2) iron-rich manganese ore which laterally adjoins calcarenite and reef limestone, 3) massive manganese ore which laterally adjoins reef limestone and 4) another reef limestone at the top of the sequence (Bonatti, et al. 1972). The ore bed has a maximum thickness of over two metres, the thicker areas however have an iron-rich base. The iron-rich ore contains goethite and smectite. The massive manganese ore consists mostly of pyrolusite in nearly amorphous and in well crystallized states with much lesser birnessite and rhodocrosite. The massive ore also contains small clay lenses, geodes of calcite crystals and traces of strontian barite. The upper zone of the ore is soft and friable and consists of todorokite, birnessite and poorly crystalline pyrolusite.

The massive manganese ore contains several percent of barium but the iron rich ore contains less than 150 ppm.

The deposit is thought to have originated from hydrothermal solutions which were expelled onto a shallow open sea floor and then slowly oxidized with iron first precipitating out near the source (Bonatti, et al. 1972).

At Black Mountain adjacent to Dallol Volcano the salt is rich in manganese oxide. Hot springs at 108°C here carry ferrous chloride in solution which on exposure to air oxidizes to goethite (Holwerda and Hutchinson, 1968; Bonatti et al, 1972). The hot brines are similar to those in Atlantis II Deep and their components are thought to be derived by leaching from both basalt and evaporite (Bonatti et al 1972).

Bonatti et al. noted that several deposits similar to that above occur in the Danakil region. Jelenc (1966a) noted that manganese was being actively mined from one such deposit and that manganese also occurs in Eocene strata in the Ogaden region to the south (about 7°N, 44°E). DeKun (1965) noted that iron-manganese oxides occur in travertine near Mt. Ghedem (near Asmara).

Without giving further information Jelenc (1966b) noted that manganese oxide deposits occur in Paleozoic and Cenozoic rocks at Beliga (14°10'N, 40°09'E to 14°16'N, 40°07'E), Handeda (14°01'N, 40°08'E), Adi Berbere (14°27'N, 38°40'E), Adi Chigono (14°27'N, 38°47'E) and Mussley (14°17'N, 40°07'E).

Mersa Gulbub Barite Deposits

Post-Miocene barite veins, north-west and west of Mersa Gulbub at Heneb (16°20'N, 39°07'E), Meter, and Gharsa occur as "post-tectonic hydrothermal deposits" associated with Red Sea rifting (Jelenc 1966a, b). These veins occur in faults cutting sediments of the Pliocene-Miocene Dogali and Oligocene Dessert Series (mostly clastic rocks). The veins are up to 2 m or more wide and at Heneb contain galena nodules up to 0.15 m wide. Manganese and iron oxides are common in some occurrences particularly those at Ghedem. The deposits occur in an area about 30 km long and are within 20 km of the coast.

Fuel Occurrences

DeKun (1965) reported oil seeps in the Dahlac Islands east of Massawa. Ahmed (1972) further noted several on-shore coastal and off-shore oil seeps north and south of Massawa. Small lignitic coal seams in Miocene beds up to a metre thick are known at Adi Ugri ($14^{\circ}50'N$, $38^{\circ}50'E$) and in the Chelga basin near Gondar ($12^{\circ}3'N$, $37^{\circ}30'E$). Jelenc (1966a) also noted that Mesozoic coal occurs in the Wollega and Shoa districts.

SUDAN

Halaib and Other Manganese Deposits

Manganese oxide is mined from eight deposits at Halaib. The manganese occurs mostly as steep veins up to 2 metres wide and less commonly as concretions and cementations of pyrolusite and lesser manganite associated with calcite, quartz, and barite in Miocene conglomerate and sandstone. The geology of the deposits is very similar to that at Wadi Malik (Egypt). Kabesh and Afia (1961) postulated an epigenetic origin but Whiteman (1971) thought the deposits were syngenetic. The individual deposits are at Ankalidot ($22^{\circ}30'N$, $35^{\circ}56'E$), El Hobal ($22^{\circ}24'N$, $36^{\circ}04'$ and $36^{\circ}07'E$), El Shumhai ($22^{\circ}21'N$, $36^{\circ}19'E$), Takmanyai ($22^{\circ}16\frac{1}{2}'N$, $36^{\circ}30'E$), Amar ($22^{\circ}06\frac{1}{2}'N$, $36^{\circ}39'E$), Adargab ($21^{\circ}53'N$, $36^{\circ}43'E$), Metateib ($21^{\circ}51'N$, $36^{\circ}45'E$), and Tikaralet.

Fracture filled manganese oxide veins up to 4 metres wide of Miocene(?) Age in Precambrian host rocks also occur at Sinkinat ($19^{\circ}00'N$, $36^{\circ}30'E$; Kabesh (1964).

At Paloich and at Wabuit are Plio-Pleistocene(?) beds of iron and manganese oxides in the Umm Ruwaba Series which consist mostly of sand, clayey sand, gravel and clay (Kabesh and Afia, 1961).

Wadi Halfa and other Iron Deposits

Although Cenozoic iron ore deposits seem to be widespread in the Sudan, like the Egyptian deposits they are not easily related to the Red Sea rift system. Oolitic goethite-hematite occurs at Halfa within Cretaceous Nubian sandstone on the Nile River near the Sudan - U.A.R. border. This seems to be rather similar to the iron ores of Aswan. Tertiary bedded

oolitic iron ores are also found at several places in Sudan, the most notable being in Western Kordofan (12°N , 29°E), e.g. near En Nahud (Kabesh, 1964).

Coastal Evaporites

Miocene gypsum is mined from the coastal region between Dungunab and Marsa Salak, particularly at Mukawar Island. Salt is mined at Ras Roweiya (DeKun, 1965). Evaporitic deposits also are recorded at several other localities along the coast (Kabesh, 1964).

EGYPT

Ras Benas Copper Occurrences

Chalcocite associated with covellite, cuprite, atacamite, and malachite occurs near Ras Benas ($23^{\circ}58'\text{N}$, $34^{\circ}54'\text{E}$) as an irregular layer up to 15 mm. thick in a dense middle Miocene dolomitic shallow water algal limestone with minor clay in gypsum overlying arkosic sandstone and conglomerate (Anwar, 1964).

El Shazly (1959) noted that secondary copper mineralization similar to that at Um Bogma in Sinai occurs at Wadi Araba ($29^{\circ}11'\text{N}$, $32^{\circ}36'\text{E}$).

Sabet (1958) recorded that some dolerites of post-Eocene age in the Um Gheig area have veinlets of chalcedony up to 0.2 m wide in joint planes and one of these carries malachite and hematite (a selected specimen assayed 11% Cu).

Wadi Malik, Elba and Wadi Araba Manganese Deposits

The steep north-west trending manganese oxide veins in Miocene sandstone and conglomerate close to the Precambrian basement at Wadi Malik ($24^{\circ}10'\text{N}$, $35^{\circ}18'\text{E}$) and Elba ($22^{\circ}04'\text{N}$, $36^{\circ}23'\text{E}$) closely resemble those at Halaib in the Sudan (El Shazly, 1959; Whiteman, 1971; Basta and Saleeb, 1971). At Wadi Malik the vein minerals are pyrolusite, psilomelane and hematite; at Elba they are pyrolusite, ramsdellite, banded colloform psilomelane and cryptomelane, soft todorokite, black calcite and barite (which increases in quantity with depth, Basta and Saleeb, 1971). Similar veins in the same system at Wadi Malik also occur in the basement

amphibolite but are probably the same age as those in the younger rocks. More than 2000 tons of ore were extracted from Wadi Malik by 1951 (El Shazly, 1959). Similar veins are known at Aranqwab, Hubal, Mated Onqwab in the basal part of the Miocene section.

Several small lenses, pockets and bedding replacement deposits of pyrolusite, psilomelane, hematite, goethite, calcite and halite occur at Wadi Araba in Lower Cretaceous sandstone (Abdallah, 1961). Traces of barium are present and the deposits are associated with NW faults.

Reigha, Ranga and Gemsa Sulphur Deposits

At Um Reigha ($25^{\circ}42'N$, $34^{\circ}30\frac{1}{2}'E$) 60 km south of Quseir are lenses and pockets of native sulphur associated with calcite and selenite which is irregularly distributed over an area of about one square kilometre. The sulphur occurs in middle Miocene gypsum - anhydrite beds associated with oil-tainted limestone (El Shazly and Mansour, 1962).

Sulphur is similarly found at Bir Ranga ($24^{\circ}24'N$, $35^{\circ}12'E$) associated with pyrite, hematite and goethite in pockets and lenses interlayered with Miocene gypsum and shale (El Shazly 1968).

Sulphur is also found associated with bituminous matter and pyrite in Miocene dolomitic limestone and anhydrite at Gemsa ($27^{\circ}39'N$, $33^{\circ}35'E$; El Shazly, 1959). According to Schnellmann (1959) the mineralization in places is extensively contaminated with hydrocarbons. He described three types of ore

1. Anhydritic : finely disseminated sulphur in anhydrite.
2. Aragonitic : crystalline coarser sulphur in interspaces between rosettes of aragonite. The occurrence of aragonite incidently suggests the anhydrite has been reduced.
3. Dolomitic : coarse crystalline sulphur lines the walls of fissures and cavities in the dolomite but not penetrating far into the mass of the rock.

These horizons invariably occur below a bed of marl. Schnellmann (1959) eloquently argued for a biogenic origin for the sulphur.

Over 1200 tons of sulphur were mined from Reigha, Ranga and Gemsa by 1955. Sulphur also occurs at Ras Benas though details of the occurrence are not available (El Shazly, 1959).

Um Gerifat and other Iron Ore Deposits

At Um Gerifat ($25^{\circ}35'N$, $34^{\circ}32'E$) between Abu Anz and Um Reigha an "ochre" deposit occurs as lenses of hematite and goethite up to 2.5 m thick in dolomitic marly and arenaceous limestone of Miocene age. The deposit contains very minor tridymite and is thought to be a weathered hypogene deposit with an origin related to that of the lead-zinc mineralization at Um Gheigh (Kabesh et al 1970).

Other iron deposits are the bedded Eocene pisolitic and massive iron ores of the Baharia Oasis and the Upper Cretaceous Minette type iron ores of Aswan (El Shazly, 1959). In both areas the iron ores occur as goethite and hematite associated with sandstone, and possibly are of a similar origin to the Wadi Fatima iron ore. El Hinnawi (1965) considered the iron ore of the El Bahariyeh Oasis ($28^{\circ}00'N$, $28^{\circ}40'E$) to be related to Tertiary volcanism.

Anz, Ranga, Wizr, Talet Eid, and Rusas Lead-Zinc Occurrences

El Kholy et al (1970) stated that Miocene galena occurs at Ranga, Wadi Ithil, Abu Anz and Gebel el Rusas but no further details were given.

At Gebel el Rusas ($25^{\circ}11'N$, $34^{\circ}46'E$) the geology is very like that at Um Gheigh 50 km to the north; the mineralization occurs as replacement veins in a north-west trending anticline and as wadi boulders (Herbst et al, 1965). According to El Shazly (1968) the mineralization occurs near a north-north-west trending fault. Breadnell (1924) recorded that the lead-zinc ore is associated with iron and manganese oxides and occurs in gypsum and the underlying Middle Miocene lime-grit series. As at several other mineralized localities the associated overlying limestone is oil tainted. Brine springs occur about 2 km north-north-west of Gebel el Rusas (Breadnell, 1924). Ancient mines have been located at Jebel el Rusas and Bir Ranga.

The galena at Bir Ranga occurs as disseminated grains in Miocene gypsum (cf. Jabal Dhaylan) and as layers alternating with sulphur and iron oxides in anhydrite and limegrit host rocks. The Abu Anz deposit ($25^{\circ}32'N$, $34^{\circ}34'E$) also occurs in Miocene limegrits having a north-north-west strike (El Shazly, 1959).

Deposits at Wadi Wizr and Abu Ghorban are located within 10 km of Um Gheig (Herbst et al 1965). The Wadi Wizr deposit (north of Um Gheig) occurs in oil tainted Miocene calcareous sandstone (limegrit) and consists of galena, wulfenite, strontianite and traces of smithsonite in siliceous replacements along bedding planes (El Shazly, 1959; Herbst et al 1965). Pre-Middle Miocene dolerites occur to the south-west of the deposit. The calcareous sandstone host rock at Abu Ghorban is very thin (Herbst et al 1965).

Lead-zinc bearing veins occur at numerous localities in the Precambrian basement of the Egyptian coastal region but with few exceptions the age of the mineralization is unknown. Veins of barite-jasper (cf. Merataba) in Precambrian gabbro at Taleit Eid, 15 km south west of Um Gheig, contain J type galenas of probable Tertiary age together with lesser sphalerite, marcasite and chalcopyrite (El Kholy and Selim 1968; Sabet and Saleeb-Roufail, 1968).

El Shazly (1968) noted that Eocene limestones at Wadi Gassus near Safaga and at Asal (Asl) oil well in Sinai also contain lead-zinc mineralization but no details were given.

Um Gheig Lead-Zinc Mineralization

The Um Gheig lead-zinc deposit at $25^{\circ}43'N$, $34^{\circ}29'E$, 50 km south of Quseir occurs as a replacement body within a Middle Miocene calcareous sandstone (limegrit) confined between a conglomerate resting on a Precambrian basement and a sequence of sandstone, conglomerate, gypsum, marl and bituminous limestone (Anwar et al, 1972; Herbst et al 1965). This sequence generally dips $25^{\circ}ENE$ and is cut by steep NNW and less common NNE and NS faults (Anwar et al 1972). The nearest basalt is 10 km north-north-west of the deposit. The overall shape of the deposit is that of a stockwork dipping steeply SSW (Herbst et al 1965). The observed dimensions of the orebody are 150 m by 50 m by 20 - 40 m. The ore zone however might be 340 m long (El Shazly 1968). The orebody consists of several coalescing irregular lenses having lead-rich cores and zinc-rich haloes with the primary ore consisting of sphalerite, galena, marcasite and pyrite; with traces of jordanite in the galena. The gangue consists of dolomite, ankerite, calcite, quartz and gypsum. Much of the ore occurs as cerussite, calamine (smithsonite?), hemimorphite, pyromorphite and

hydrozincite. The carbonate and quartz form crusts and fissure fillings, limpid gypsum lines cavities in colloform sulphide ore. The sphalerite occurs as concretions and cavernous drusy masses with inclusions of galena (Herbst et al 1965). The outer margins of the deposit are more ferruginous and the ore in the outer zones is disseminated in the country rock.

Minimum reserves have been calculated as one million tons (Herbst et al. 1965) with the following composition

Zn	15.82 %	Sb	0.0018%
Pb	2.26	Cl	0.44
Bi	trace	F	< 0.005
Au	trace	CO ₂	10.9
Ag	12.7 (ppm)	CaO	8.97
Cu	0.013	MgO	4.69
Hg	< 0.001	BaO	0.14
Cd	0.025	Fe ₂ O ₃	11.34
Mn ₃ O ₄	1.25	SiO ₂	25.60
Ni	0.17	Al ₂ O ₃	6.29
As	< 0.01	H ₂ O	4.90
Co	0.02		

Zug el Bohr Galena Deposit

Four ancient mines have been found at Zug el Bohr 15 km south of Quseir and 35 km north of Um Gheig (El Shazly 1968). At this locality galena occurs in lenses up to 10 m thick parallel to bedding in replaced Middle Miocene limegrit, sandstone and shale. Mineralization however does not appear to have been influenced by the cross-bedding of the sandstone. Limestone overlying the deposit is oil tainted (El Shazly, 1959). The mineralization is associated with north-north-west faults and is cut by east-north-east faults (El Shazly, 1968).

The writer was able to examine specimens in the Mining Geology Department Museum collected by Mr. R. Blattner. The ore specimens consist of nodular galena and galena-cemented sandstone. The nodular specimens have maximum dimensions of 70 by 100 by 10 mm and probably represent simple nodules or weathered vein fragments. They are composed of massive galena in contacting grains up to 10 mm across. In places the contacts contain black material which appears to be a mixture of cerussite and plattnerite. The galena is pure except for rare relic quartz grains and pebbles.

Pockets in the galena contain cerussite (of two types as at Jabal Dhaylan) and limpid crystals of hemimorphite. The nodules are coated in aphanitic white material up to 2 mm thick. This consists of cerussite, paralaurionite, anhydrite and a trace of halite. A very thin vitreous net-like film of gypsum coats large areas of the opaque white coat. Pale green poorly crystalline pyromorphite also occurs with anhydrite in parts of the opaque white coat; and in some places there are beautiful yellow and orange wulfenite crystals up to 1 mm across.

The sandstone specimens range from ordinary massive quartz sandstone with an argillaceous matrix to massive galena with inclusions of quartz grains. The galena appears to have replaced some of the matrix and the quartz. Some of the sandstone has cerussite cement presumably through weathering and some surfaces of the sandstone are coated with cerussite, hydrocerussite(?) and plattnerite, pyromorphite, and anhydrite.

Quseir - Safaga Phosphate Deposits

Phosphate beds of Upper Cretaceous age associated with shallow marine limestone, argillite and marl are widely distributed between Quseir ($26^{\circ}06'N$, $34^{\circ}16'E$) and Safaga ($26^{\circ}45'N$, $33^{\circ}56'E$) along the Red Sea coast. The beds contain bones, teeth, shells, rock fragments, quartz grains, clay, pyrite and organic matter and have a maximum thickness greater than 2 m. The phosphate occurs as oolitic and colloidal collophane, and dahlite (El Shazly, 1959, El Tarabili, 1969). It obviously has a marine organic origin and was probably upgraded by leaching of calcite during diagenesis. By 1956, 12.5 million tons had been mined and reserves were calculated as 176 million tons. The phosphorous content correlates with the fluorine, vanadium and uranium contents. The U_3O_8 at Quseir is 0.004% and at Safaga is 0.008% (El Shazly, 1959).

Hughada - Ras Gharib Oil Field

Oil was found seeping into tunnels of the Gemsa sulphur mine in 1868 but the surrounding oil field was not discovered until 1909 (DeKun, 1965). The producing field of Hughada - Ras Gharib occurs along the Gulf of Suez from $27^{\circ}14'N$, $33^{\circ}50'E$ to $28^{\circ}21'N$, $33^{\circ}06'E$). Lignite also occurs along this area (El Shazly, 1959).

SINAI

Gulf of Suez Region - Um Bogma Manganese and Copper

The Um Bogma district, ($29^{\circ}00'N$, $33^{\circ}20'E$) includes more than 14 manganese orebodies and many copper occurrences in an area about 18 km long and 9 km wide.

All of Western Sinai is very intensely faulted with normal antithetic faults tilted away from the Gulf of Suez. These together with associated dyke swarms and hot springs are mostly of post Miocene age (Shata, 1956).

The manganese orebodies were thought by most early writers (e.g. Shata, 1956) to be related to these faults. If this were so they would be directly related to the rifting of the Red Sea, however later writers (e.g. El Shazly et al 1963, Mart and Sass, 1972) consider the ore to be syngenetic-diagenetic.

The ore occurs as massive lenses 1 to 200 m diameter and 0.2 to 6.0 m thick in subhorizontal shallow marine Lower Carboniferous dolomite near the base of an essentially clastic succession overlying Precambrian granite, diorite and gneiss (Mart and Sass, 1972).

Up to 1957 ore production was over 4 million metric tons of 22% Mn and 35% Fe ore. The larger lenses have manganese-rich cores and iron-rich margins. The manganese minerals are pyrolusite and psilomelane-cryptomelane with lesser polianite, ramsdellite, manganite, hausmannite and pyrochroite. Hematite and goethite are the iron minerals. Gangue minerals are montmorillonite, illite, quartz, barite, dolomite, calcite and halite. The ore generally has a colloform texture and is veined by pyrolusite-barite and pyrolusite-barite-quartz. The order of deposition appears to be barite, manganese minerals, iron minerals and lastly calcite. Dolomitic areas are richer in manganese and sandy areas are richer in iron (El Shazly, et al 1963, Mart and Sass 1972). The restrictions of the ore to one stratigraphic horizon and the relation of iron and manganese distribution to certain lithofacies suggest a syngenetic origin. Mart and Sass (1972) proposed that the iron and manganese in adsorbed and finely particulate forms were trapped in tidal pools of low energy. El Shazly et al (1963) proposed that the iron was deposited closer to the shoreline than the manganese. According to Mart

Sass (1972) diagenetic differentiation, which was concurrent with dolomitization of limestone, produced a gradient of Eh and pH that attracted the Mn^{++} to the cores of the lenses (low Eh, high pH) and Fe^{++} to the rims of the lenses.

Copper occurs in the sandstone about 50 m below the manganiferous dolomite at Um Bogma according to Mart and Sass (1972). Hilmy and Mohsen (1964) however noted that it occurs most commonly in reddish brown shales, above the dolomite, at the top of the Lower Carboniferous succession. They recorded 17 different copper minerals from 10 localities in the Um Bogma district. Atacamite, malachite and turquoise are the most significant minerals in this area. These minerals occur in veins generally parallel to the Suez rift, disseminated grains and concretions mostly in shales but also in sandstone, dolomite and rocks of the Precambrian basement.

The writer examined specimens from Um Bogma in the Mining Geology Department Museum. A typical manganese ore specimen consists of massive aphanitic material containing amorphous manganese oxide, goethite and a little hematite, drusy in places, in which are veins of crystalline pyrolusite. This pyrolusite occurs in beautiful botryoidal aggregates of radiating fibres up to 50 mm or more long. In places attached to the long fibrous material are randomly oriented blocky crystals of pyrolusite probably representing a faulted form of the long fibrous type. Spherulitic chalcedony is associated with some of the radiating pyrolusite. Another typical ore type consists of hard ($H > 5$) aphanitic masses of intermixed hematite and amorphous manganese oxide containing cavities and irregular patches of soft aphanitic material of the same composition. Barite associated with the ore occurs as limpid crystals with sides up to 20 by 30 by 10 mm long. This contains many fine inclusions, and some inclusions up to 0.5 mm diameter consisting of goethite which probably has replaced pyrite. The specimens of country rock associated with the ore are as described by Mart and Sass (1972) except for a dolerite dyke intersecting the ore zone. This is altered to a pale brown "white trap" consisting of fine grained smectite, kaolinite, feldspar, maghemite and magnetite. It contains irregular replacement veinlets of manganese oxide up to 1 mm wide in places associated with calcite, random mixed layered clay and hematite. The XRD pattern of the manganese oxide suggests that it is very poorly crystallized pyrolusite. The alteration of the dyke and the introduction of manganese can be attributed to endometamorphism (Hamilton, 1968).

At the southern tip of Sinai manganese oxide occurs as cement to a conglomerate (deKun, 1965) near Sharm el Sheik.

Gulf of Suez Region - Fuel and Gypsum

Oil has been found at Sudr ($29^{\circ}40'N$, $32^{\circ}42'E$) and Asl ($29^{\circ}33'N$, $32^{\circ}44'E$) in Eocene-Miocene strata, at Abu Rudeis, Sidri ($28^{\circ}48'N$, $33^{\circ}12'E$), Wadi Feiran ($28^{\circ}40'N$, $33^{\circ}12'E$) and Belayim ($28^{\circ}33'N$, $33^{\circ}19'E$) in Miocene strata, and at Abu Durba ($28^{\circ}29'N$, $33^{\circ}20'E$) in Nubian strata. Non-coking coal occurs at Oyun Mussa ($29^{\circ}53'N$, $32^{\circ}39'E$) and coking coal of mining potential occurs at Jebel Marag ($26^{\circ}33'N$, $31^{\circ}37'E$; deKun, 1965). Petrasheck and Nakhla (1961) noted that a coal seam 0.5 to 0.8 m thick lies in Lower Carboniferous strata about 100 m above the manganese-bearing dolomitic limestone at Um Bogma.

Gypsum forms extensive outcrops in south-western Sinai and has been mined north of Um Bogma (deKun, 1965).

Aqaba - Dead Sea Region

The Sinai Plate adjoins the Arabian Plate along the Dead Sea Rift which is a northern extension of the Red Sea Rift system. Quennell (1959) found that a pre-Neogene restoration of the rift zone brings ten features back into coincidence. For example with his restoration the mineralized areas containing cupriferous sandstone and manganiferous sandstone of Timna ($29^{\circ}45'N$, $35^{\circ}00'E$, 25 km north of Aqaba) and Fenan (130 km north of Aqaba) appear to be parts of the one large mineralized area. The copper at Timna occurs as veins of various copper minerals associated with barite, fluorite, calcite and quartz in the Precambrian granite and as more extensive "stratiform" ores consisting of chalcocite, chrysocolla, malachite and other secondary copper minerals occurring as cement, veinlets, thin layers and concretions in nearby Lower Palaeozoic sandstone (Wurzbürger, 1967). The manganese occurs in "zebra sandstone" stratigraphically higher than the copper and consists of pyrolusite with lesser polianite, psilomelane, hollandite and ramsdellite (Bentor, 1957b).

Secondary copper mineralization in Eocene limestone, chert and sandstone in Wadi Arava has been attributed to hot solutions rising up the Dead

Sea rift (Nissenbaum, 1970). The mineralization consists of malachite, atacamite, chrysocolla and chalcocite(?) in nodules, veinlets and as cement. The localities of this mineralization in Wadi Arava are at Nahal Ashosh ($30^{\circ}30'N$, $35^{\circ}10'E$), Nahal Shahaq ($30^{\circ}44'N$, $35^{\circ}11'E$) and En Gofra (near Galilee).

Hot springs now active in the northern Dead Sea rift valley have been found to contain very similar chemical constituents to each other and to the hot springs at Gebel Hamman Forun ($29^{\circ}12'N$, $32^{\circ}58'E$) in Western Sinai near the Gulf of Suez (Mazor 1968). As the waters are independent of rock type they were attributed to deeply infiltrated Sea water mixed with local ground water.

PART I DESCRIPTIVE GEOLOGY

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Section 2	The Coastal Region
Section 3	The Cenozoic Igneous Realm
Section 4	The Shield

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SECTION 3 THE CENOZOIC IGNEOUS REALM Contd.

1.3.1 ACTIVE VOLCANOES IN REGIONS AROUND THE RED SEA

Active volcanoes are sparsely scattered over Saudi Arabia (mostly in the north) and are rather common in Ethiopia and Yemen. Those in Saudi Arabia are relatively unknown and the present writer has not seen references by other writers to Neumann van Padang's catalogue (1963). Several of the volcanoes however are shown on Brown's map (1972). It seems most likely that there are more active volcanoes than those presently known, as the harrats are relatively uninhabited and unknown regions. The following data has been extracted from Neumann van Padang (1963) and Brown (1972).

SAUDI ARABIA

Name	Locality	Activity	Rock type
1. Western Harrat ar Raha	27°48' N 36°10' E	erupted c. 1300 BC	?
2. Hala al Ishqa	27°35' N 36°48' E	erupted 640 AD	ol. basalt
3. Hala al Bedr	27°15' N 37°12' E	" 640 AD	"
4. Hala al Asi	27°05' N 37°21' E	" before 600AD	"
5. Hala al Ethnein	26°23' N 39°58' E	sofataric	tuff
6. Harrat el Khaiber	25°34' N 39°52' E	erupted 700-800AD	basalt?
7. Harrat al Medina	24°20' N 39°44' E	" 1256AD	?
8. Harrat Lunaiyir	25°15' N 37°41' E	" 1000-1100AD	?
9. -	27°32' N 37°14' E	?	?
10. -	27°21' N 37°22' E	?	?
11. -	27°14' N 37°26' E	?	?
12. -	24°20' N 39°44' E	?	?
13. Djebel Yar	near Waghra hot springs	erupted 1820	ol. basalt

It is quite possible that 2 is the same as 9, 3 the same as 10 and 4 the same as 11; the records are very vague.

Volcanoes within the Red Sea have been discussed in section 1.1.5 and further details are given by Gass et al (1973). Richard and Neumann van Padang (1957) recorded that Jabal Tayr (15°42' N, 41°44½' E) erupted in 1833-1834 and possibly 1883 and that Saddle Island (15°07' N 42°06½' E) in the Zubayr Group) erupted in 1846.

YEMEN AND SOUTH YEMEN

Name	Locality	Activity	Rock Type
1. Kaulet Hattab	15°38'N 45°05'E	erupted 500-600AD	ol. basalt
2. Djebel Zebib	15°36'N 44°07'E	" after 200AD	ol. basalt
3. Djebel El Marha	15°17'N 44°13'E	?	basalt
4. Djebel Er Rabeke- et Tadawin volcanic chain	(15°20'N 44°15'E) (15°33'N 45°20'E)	erupted after 1200BC	ol. basalt
5. Harras of Dhamar	(14°25'N 44°25'E) (14°34'N 44°35'E)	erupted 1937	basalt
6. Djebel el Esi	14°31'N 44°37'E	solfataric	pantellerite and basalt
7. Djebel Isbil	14°39'N 44°37'E	solfataric	basalt
8. Djebel Hammam Demt-	14°03'N 44°45'E	solfataric	?
9. Djebel En Nar	13°20'N 43°44'E	solfataric	?
10. Es Sawad	(13°22'N 45°40'E) (13°49'N 46°34'E)	solfataric	ol. basalt
11. -	15°39'N 44°07'E	?	?
12. -	14°30'N 44°45'E	?	?

Note that 11 and 12 may coincide with some of the preceding areas.

ETHIOPIA

The active volcanoes of Ethiopia can be conveniently divided into two groups, one trending south-west which is related to the Ethiopian rift and one trending north-west which is related to the Red Sea rift. The following data have been extracted from Richard and Neumann van Padang (1957).

S o u t h - W e s t P r o v i n c e

Name	Locality	Activity	Rock Type
1. Tollo-dimetu	8°55'N 38°10'E	solfataric	?
2. Fantale	8°58½'N 39°56'E	"	sodic rhyolite and trachyte
3. Dofane	9°21'N 40°06'E	"	rhyolite
4. Damer Giog	10°00'N 42°51'E	"	?
5. Abida	10°04'N 40°50½'E	erupted 1928	basalt and andesite
6. Ayelu	10°05'N 40°42'E	" "	rhyolite
7. Gabillemma	10°47'N 41°21'E	solfataric	?
8. Boina	10°58'N 42°12'E	"	basalt
9. Gabres	11°07'N 42°30'E	"	?
10. Tiho	11°30'N 42°03½'E	"	?
11. Kurub	11°36'N 41°10'E	"	?

N o r t h - W e s t P r o v i n c e

Name	Locality	Activity	Rock Type
12. Afdera	$13^{\circ}24' \text{ N } 41^{\circ}05' \text{ E}$	erupted 1915	basalt
13. Ummuna	$13^{\circ}32\frac{1}{2}' \text{ N } 40^{\circ}37' \text{ E}$	solfatatic	?
14. Erta Ale	$13^{\circ}37' \text{ N } 40^{\circ}35\frac{1}{2}' \text{ E}$	erupted 1906	basalt
15. Dubbi	$13^{\circ}42' \text{ N } 41^{\circ}35' \text{ E}$	" 1863	basalt
16. Gabuli	$13^{\circ}47' \text{ N } 40^{\circ}26' \text{ E}$	solfatatic	?
17. Alu	$13^{\circ}49' \text{ N } 40^{\circ}21\frac{1}{2}' \text{ E}$	"	?
18. Kebrit Ale	$13^{\circ}55' \text{ N } 40^{\circ}15' \text{ E}$	"	basalt
19. Dallol	$14^{\circ}14' \text{ N } 40^{\circ}17' \text{ E}$	erupted 1926	halite (sic)
20. Alid	$14^{\circ}53' \text{ N } 39^{\circ}55' \text{ E}$	solfatatic	sodic rhyolite and andesite basalt
21. Jalua	$15^{\circ}02\frac{1}{2}' \text{ N } 39^{\circ}49' \text{ E}$	"	ol. basalt

Several of these places (1, 4, 10, 13, 14, 19) were noted as fumarolic fields and are not necessarily volcanoes. Dallol (19) for example is a 3 km long, $1\frac{1}{2}$ km wide dome of salt with a summit crater containing hot brine springs at 100°C (Holwerda and Hutchinson, 1968). Black Mountain merges with Dallol Mountain to the north-east (summits are about 4 km apart). A violent phreatic eruption in 1926 produced angular salt breccia around the rim of Black Mountain Summit in which now occurs a MgCl_2 saturated brine at 130°C . Gravity and magnetic anomalies together with iron salts suggest the activity is due to underlying mafic intrusions (Holwerda and Hutchinson, 1968).

Several more volcanoes have been found by Barberi et al (1970) and Barberi and Varet (1970). There are, for example, at least six active volcanoes in the Erta Ale range which is 95 km long. These volcanoes are mostly fumarolic and range from olivine basalt to sodic rhyolite in rock type. A lava lake occurs at Erta Ale itself from which fountains were observed in November 1969 (Barberi and Varet). In addition to Erta Ale range, dormant volcanoes also occur in the Alayta range ($12^{\circ}52' \text{ N } 40^{\circ}32' \text{ E}$). Fissure lavas are common in both ranges and pyroclastics are rare.

SECTION 3 THE CENOZOIC IGNEOUS REALM Contd.

1.3.2 CENOZOIC VOLCANISM IN YEMEN AND ETHIOPIA

DISCUSSION

Igneous activity is well developed in the southern part of the Red Sea region where it is also relatively well known. It therefore serves as a good background to studies in the central and northern parts of the region. It is also essential to the understanding of the general tectonic development of the Red Sea region.

Dominantly basic lavas have been produced in this region throughout the Cenozoic era. According to Gass (1970b) alkaline basalts of the Trap Series preceded and accompanied vertical uplift (geodoming) of the region. Following this, peralkaline volcanism of the Aden Series occurred along zones of crustal attenuation, and low potassium tholeiitic basalts of oceanic character developed in the central part of the Red Sea as well as in the northern Afar and the Gulf of Aden. Exceptions to this are the alkalic basalts of recent age in attenuated zones and on the flanks of geodomes.

YEMEN TRAP SERIES

The Trap Series covers $750,000 \text{ km}^2$ to a maximum thickness of 3,500 m in Ethiopia, and $30,000 \text{ km}^2$ to a maximum thickness of 2,000 m in south-west Arabia. The series ranges in age from Paleocene to Lower Miocene and is generally uniform and similar to alkali olivine flood basalts elsewhere in the world (Gass, 1970b).

The series is divided into two units. The lower unit is 1200 ± 200 m thick and consists of a few thick extensive and uniform flows of aphyric alkali basalt erupted from meridional fissures. The upper unit is 2600 m thick and consists of more widely varying, usually porphyritic, types erupted from central vent volcanoes which, although they contain abundant silicic and ultramafic varieties, are of alkaline affinity throughout (Gass, 1970b).

The Trap Series probably originated at a depth of about 60 km by partial fusion in the upper Mantle and the magma probably rose by stoping and wall rock reaction. With time the series became less alkaline giving rise to peralkaline siliceous eruptions (Cox et al 1970). Relative quiescence followed the Trap Series volcanism although uplift and attenuation continued (Gass, 1970b). There is nevertheless some overlap in time of the Trap volcanism and the Aden Series volcanism. This has led some writers (e.g. P. Mohr, 1972; Barberi et al, 1972) to abandon this terminology.

ADEN VOLCANIC SERIES

Although it is only a five hundredth of the volume of the Trap Series, the Aden Volcanic Series is nevertheless one of the largest peralkaline provinces of the world (Gass, 1970b). When volcanic activity recommenced after the deposition of the Trap Series at the end of the Miocene, it was of the Aden Volcanic Series central vent type and the Red Sea oceanic type. The Aden Series is considered by Gass to be transitional between the Trap Series type and the Red Sea oceanic type.

The Aden Volcanic Series covers most of the Afar triangle in Ethiopia apart from the northern Danakil, and even extends into the main Ethiopian rift zone but is less extensive around Aden in South Yemen. The series of Aden differs slightly from that in Ethiopia. Both contain silicic rocks but intermediate types are rare in Ethiopia (Gass, 1970b).

Cox, Gass and Mallick (1970) considered the most likely "parental" magma for the series at Aden to be a mildly alkaline olivine basalt which on differentiation would give a series ranging from trachybasalts to trachyandesite to trachytes to peralkaline rhyolites. These rocks are difficult to recognise in the field. Some rhyolites, for example, with more than 20% normative quartz have olivine, alkali feldspar and clinopyroxene as phenocrysts. Geomorphology and K/Ar dating indicate the series is about 5 m.y. old but the $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{87}\text{Rb}/^{86}\text{Sr}$ isochron plot suggests the series is related to a thermal event some 20 to 30 m.y. earlier than the age of eruption. This older age corresponds to the age of the Trap Series, the upper part of which is also chemically similar to the Aden Series. Cox, et al (1970) therefore postulated that after partial fusion in the upper Mantle had given rise to the Trap Series

some magma crystallized to give a layered plutonic sequence in high level magma chambers. This plutonic sequence was subsequently partially melted to produce the magma of the Aden Series which then reached the surface without significant mixing or fractionation. They also noted however that although this applies to the Aden Series at Aden it does not necessarily apply to the Aden Series of Ethiopia.

VOLCANISM IN THE AFAR TRIANGLE Barberi et al (1970) noted that the Aden Series in Afar bears some similarities to the volcanics of the Erta Ale and Alayta ranges but has a more variegated petrography and has a wide range of $^{87}\text{Sr}/^{86}\text{Sr}$ isotope ratios. This suggests to them that the series was affected by crustal silicic contamination. Brown and Coleman (1972), McKenzie et al (1970) and Schilling (1973) however consider that the Afar triangle is underlain mostly by oceanic crust.

The Danakil rift began to form in the lower Miocene and since then volcanism has been almost continuous (Barberi et al, 1972). Mohr (1972) rejected the terms "Trap Series" and "Aden Series" however his "Afar Series" appears to be equivalent to the Aden Series in Afar. He further noted that in central and southern Afar Lower Miocene - Upper Pliocene flood basalts of the Afar flood basalts are overlain by silicic volcanics, restricted Quaternary flood basalts and extensive Quaternary sediments. Peralkaline granites of Oligocene - Quaternary age were also noted to have pierced this succession.

Barberi, Tazieff and Varet (1972) rejected all other terminologies and introduced the terms "oceanic volcanism", "continental volcanism" and "stratoid volcanism". They considered that the main types of volcanism reflect the tectonic evolution of the Afar. The stratoid volcanics are an extensive thick series of fissure basalts and ignimbrites with some silicic piles. They are mostly of peralkaline composition but lack intermediate rocks. The high soda content makes their origin problematical. The continental volcanics are mainly silicic lavas and ignimbrites with minor basalts and almost no intermediate rocks. They occur along rift margins and are thought to have resulted from contamination or direct crustal melting. The oceanic volcanics occur at Erta Ale Range and its en echelon continuation in Alayta and Tat'Ali ranges. They consist mainly of basalts with lesser peralkaline rhyolite, ferro-basalts and dark trachytes and are thought to be derived from a primary magma transitional between tholeiite and alkali basalt.

ERTA ALE AND ALAYTA RANGES

Lavas of the Erta Ale and Alayta ranges in northern Danakil are very similar to the neo-oceanic basalt of the Red Sea axial trough and the Gulf of Aden. Although they possess a mineralogy typical of alkali basalts the lavas are olivine tholeiites and tholeiites on a normative basis (Barberi et al, 1970).

These lavas issued from fissure, small shield, and central volcanoes, and consist of abundant basalts and scarce soda-rhyolites. The intermediate fraction products include andesine basalt, ferro-basalt (tot. FeO > 15%) and dark trachyte. The end-products of fractionation are oversaturated alkali trachyte, alkaline and peralkaline rhyolite (Barberi and Varet, 1970). There are very few signs of explosive activity even with the acid rocks probably due to a rapid rise of magma. The volcanic activity began in the Quaternary with the extrusion of submarine basalt and continues today with lava lake activity at Erta Ale (Barberi and Varet, 1970).

The $^{87}\text{Sr}/^{86}\text{Sr}$ isotope ratios are low and constant throughout the series. This suggests the magmas were not contaminated with crustal material (Barberi et al 1970). Gass (1970b) described the series as low potassium tholeiites of oceanic character.

SECTION 3 THE CENOZOIC IGNEOUS REALM Contd.

1.3.3 CENOZOIC VOLCANISM IN SUDAN, EGYPT AND SINAI

Very little is known of Cenozoic volcanic activity in this part of the Red Sea region. Almond (1968) noted that Cenozoic volcanic rocks, mostly alkali olivine basalts with lesser peralkaline trachyte, rhyolite and phonolite are widely distributed in Sudan and cover an area exceeding 50,000 km². Medani (1973) further noted that the Trap Series of Ethiopia extends into Sudan and that phonolitic plugs occur near Doka. He also noted the occurrence of a swarm of NNW trending dolerite dykes.

According to Andrew (1937) Tertiary igneous rocks are sparsely scattered over the whole of Egypt with a concentration at Sinai and in the northern part of the Gulf of Suez. Mostly they are alkaline olivine basalts; acid rocks are almost entirely lacking and andesites are uncommon. He also noted that pre-Miocene dykes in the Suez area are parallel to the Gulf of Suez.

Sabet (1958) noted that many of the Tertiary volcanic rocks in Egypt are of Oligocene age. In particular he referred to post-Eocene porphyritic olivine "dolerite" lavas about 30 km south of Quseir, and to probable Pre-Middle Miocene "dolerite" patches at Sharm el Bahari, Wadi Wizr and an area north-north-west of Wadi Um Gheig. A mineralized lava also in this area is discussed in section 1.2.5.

Le Bas (1971) described an area of 700 km diameter covering parts of north-east Egypt, south Sinai and north-west Saudi Arabia as a Tertiary peralkaline igneous province. Although this province probably does exist it is difficult to see what evidence this description is based on. Le Bas gave no references to the source of his data or inspiration. He also described another such province covering parts of Ethiopia and Yemen. One would expect this to be the Aden Volcanic Series but the area on his map extends well beyond the outcrops of this series.

Chemical analyses of rocks from Egypt show them to be alkaline basalts (discussed further on). This also applies for a post-Miocene(?) dyke at

Um Bogma in Sinai. This is a typical alkali olivine basalt very similar to those described further on in Saudi Arabia except that, unlike a nearby lava which has been analysed by the writer it is very altered.

In the north of Saudi Arabia and extending beyond into Syria and Israel are flood alkali basalts of Miocene-Pliocene and Pleistocene-Holocene ages (Krasnov, 1968 , Bentor, 1957a). The main trend of these basalts is parallel to the Red Sea coast so even though they are far distant from the sea their origin is probably related to the rifting of the Red Sea.

SECTION 3 THE CENOZOIC IGNEOUS REALM Contd.

1.3.4

CENOZOIC VOLCANISM IN SAUDI ARABIA

The Cenozoic plateau basalts of Saudi Arabia cover an area of about 75,000 km² (over 100,000 km² if the Syrian basalts are included) and so form one of the great flood basalt provinces of the world. It is remarkable therefore that so little is known of this area. The absence of any published petrographic descriptions or chemical analyses of these rocks spurred the writer to look at a few widely spaced areas simply as a starting point. Work is currently underway however at Leeds University and at the USGS in Saudi Arabia. Recently Baker et al (1973) of Leeds University presented a paper on the "Alkaline volcanic rocks of Jebel al Abyad, Saudi Arabia" to the Geological Society of London.

Baker et al (1973) noted that the al Abyad rocks have similar differentiation trends to the rocks of the Ethiopian plateau (Trap Series). Hawaiiites, mugearites, trachytes, phonolites and comenditic lavas are common in the al Abyad area. Except for the comendites, the entire suite is nepheline normative. The parental magma is thought to have been relatively high in total iron and alkalis and to have had a high K/Rb ratio. Low concentrations of the incompatible elements in the basic rocks suggest to Baker et al. that the melt moved rapidly from depths of perhaps 60 to 100 km, to high level chambers. A low degree of partial melting in the mantle was advocated to account for the high total alkali content. The high K/Rb ratio in both the volcanics and basement rocks was thought to reflect a mantle source particularly low in Rb. These results are quite compatible with those of the writer except that the writer studied only basic rocks.

The USGS maps (I-200A, I-204A, I-205A, I-210A, I-211A, I-216A and I-217A) show the volcanic plateau to be mostly olivine basalt with minor and local andesite, trachyte, felsite, aplite and rhyolite. A calcite-bearing feldspathoidal syenite plug apparently has intruded basalt at 18°06'N, 43°11'E, and large areas of obsidian occur at Jabal al Abyad (25°40'N, 39°50'E). If Barberi et al (1972) are correct in their extra-

polations then stratoid volcanism of flood basalt and ignimbrite probably has occurred in Saudi Arabia though no ignimbrites have yet been found.

The distribution of the lavas and of large volcanic craters shows a trend parallel to the Red Sea in Harrat Uwayrid and Al Harrah, and a meridional trend in Harrats Hutaym and Rahat. The Red Sea volcanic trend is most probably directly related to rifting especially as it is part of a depression elsewhere filled with Quaternary sand. The meridional trends are probably due to the effects of rifting on a pre-existing structure. A pre-rift reconstruction of the Red Sea region shows Harrat Rahat to be an extension of a volcanic belt in Sudan near the Ethiopian border (see the International Tectonic Map of Africa, 1:5,000,000 1968, UNESCO, for the position of the lava in Sudan). Gass (1973) considered that the meridional trend was due to an obstruction to the anticlockwise movement of the Arabian plate by its impingement with the Eurasian plate. This does not conflict with the above view of the writer.

According to Gillmann (1968) all of the eastern edge of the plain of Jizan was the seat of very intense volcanism during the formation of the Red Sea graben in the Upper Cretaceous and particularly during the deposition of the Baid Formation which consists largely of tuffs and flows of basalt, trachyte, tinguaitite, dacite, "variolite", obsidian etc. Coleman and Brown (1971) noted that the Oligocene to Lower Miocene was a time of widespread alkali basalt volcanism in the Sirat Plateau, following continental attenuation and the taphrogeny of the Red Sea. They further noted that these rocks may be coextensive with the Yemen Trap Series. Basalt at Wadi Fatima is also of Oligocene age and may be correlated with basalt at Khor Shin Al in Sudan (Coleman and Brown, 1971).

In the Lower Miocene (about 22 m.y. ago) tholeiitic gabbro basalt and andesite dykes invaded the eastern margin of the Red Sea. These are exposed between Al Lith and the Yemen border, especially at Jabal Turf (described in section 1.2.4) and can be traced as far north as the Gulf of Aqaba as reversed magnetic anomalies (Gillmann, 1968, Brown, 1972, Coleman and Brown, 1971). These dyke swarms in the Jizan area are 5 to 10 km wide and represent a crustal dilation of as much as 300 m (excluding Jabal Turf).

The largest area of Tertiary basalt on Brown's (1972) map is of Middle Miocene age and the next largest is of Pleistocene-Holocene age.

Coleman and Brown (1971) pointed out that the alkali olivine basalt extrusion in the Miocene was contemporaneous with the extensive deposition of evaporites in the Red Sea region. Brown (1972) stated that volcanic activity continued through the Tertiary to the present with peaks at about 25 21 18 12 10 7-5 and 4-0 m. years ago. The Miocene-Holocene flows of Western Arabia are alkali olivine basalt chemically similar to the Yemen Trap Series. Xenoliths of mantle material in lavas 0-5 m.y. old extruded from faults point to mantle sources and greater depths of magma generation compared with the oceanic subalkaline basalts of the axial trough (Brown, 1972).

SECTION 3 THE CENOZOIC IGNEOUS REALM Contd.

1.3.5 AGE DETERMINATIONS OF ARABIAN BASALTS

During the early stages of this study no isotopic age determinations were available for any of the basalts of the Arabian Plateau. To overcome this problem specimens were collected for paleomagnetic age determinations but some of these were later submitted to Dr. N.J. Snelling of the Institute of Geological Sciences, London, for K/Ar dating. Just before the preliminary results of this dating became available Brown (1970a) published some age determinations on the basalts. These together with geological estimates based on stratigraphy and geomorphology are shown on his map depicting time relations for the Arabian Plateau basalts. Fortunately the data do not overlap.

More recently Brown (1972) revised his data making changes that effect an area of perhaps 30,000 km². Unfortunately the localities of the new specimens taken for age determination are not known.

The localities for the samples submitted to Dr. Snelling are given further on in this section. The samples selected for dating come from near mineralized areas, and the main purpose of the dating was to see if any relation could be found between mineralization and magmatism. Such a relationship would be suggested if the basalt were coeval with mineralization, or possibly with the host rocks to the mineralization. An indirect relationship should also exist because the events of mineralization, rifting and volcanism should have been controlled by the same tectonic system.

Samples AZL and JJM are both from intrusions possibly related to mineral occurrences as either chemical or heat sources. JJM came from a dyke-like plug which several geologists visiting Jabal Dhaylan thought was probably coeval with Harrat Qalib. A magnetic anomaly connects this (Dadet et al 1970) plug to the mineralized area at Jabal Dhaylan. Samples QAL, MER, and RAB are from lava flows. The results obtained by I.G.S. are:-

AZL	16.04	$\pm$	0.8	m.y.	Miocene
JJM	15.9	$\pm$	0.8	"	Miocene
QAL	1.78	$\pm$	0.59	"	Pliocene
MER	1.06	$\pm$	0.39	"	Pleistocene
RAB	79.7	$\pm$	18	"	Cretaceous(?)

The mineralization near AZL and JJM occurs in Miocene rock but is probably of Plio-Pleistocene age. It seems therefore that QAL is more likely to be related to the mineralization at Jabal Dhaylan than JJM. There is also possibly a relation between the new phase of sea-floor spreading in the Red Sea which began 2 to 4 m.y. ago and the mineralization and volcanism of Harrat Lunnayyir. The intrusive activity may relate to an earlier phase of the rift formation.

The age of RAB is greater than expected from field relations and was questioned, but nevertheless it was confirmed with a second analysis by I.G.S. Brown (1972) however noted that some of the samples studied by the USGS had anomalously old ages due to argon inheritance from older rocks. RAB comes from Harrat Rahat which Brown (1970a) depicted as Pliocene to Recent in age but later changed to Miocene. Despite this relatively young age for the Harrat he showed (1970) two sample localities south of RAB with ages of 42.8 and 55.2 m.y.

SECTION 3 THE CENOZOIC IGNEOUS REALM Contd.

1.3.6

ARABIAN BASALT CHEMISTRY

INTRODUCTION

The reasons for studying the basalt chemistry are, firstly, apart from the new data by Baker et al (1973), no chemical analyses of any samples from the Saudi Arabian basalt plateau appear to exist in the literature and any information on the basalt chemistry of this large province is therefore an advance in knowledge. Secondly, direct relations between the basalts and the mineralization were sought through trace element study and through dating for which K determinations were necessary. Thirdly indirect relations of magmatism and ore deposition were sought through relating the chemistry to tectonism.

SAMPLES

Unfortunately it was not possible to adequately sample the basalt plateau but some samples were collected during the coastal mineralization reconnaissance. These samples were originally collected for paleomagnetic dating and hence were large and fresh. The samples from the Red Sea deeps received from the Applied Geochemistry Research Group however were so small as to be of questionable value and were therefore analysed in duplicate. The weights of these samples were :-

Red Sea deeps	Glass A	1.7 g	RSA
" " "	glass B	0.8 g	RSB
" " "	phyric basalt	3.7 g	{RSC RSD

LOCALITIES

RSA	Sudan Deep core 37, 2, 427 m deep.	20°06.6' N, 38°27.5' E
RSB	" " "	
RSC	" " "	
RSD	" " "	
RAB	North Rabigh, Harrar Harshah, part of Harrat Rahat	23°00' N, 39°00' E lava flow

MER	Merataba, Jabal Ras Jarbul	$24^{\circ}20'N$, $37^{\circ}40'E$	lava flow
QAL	East Harrat Qalib near Jabal Dhaylan	$25^{\circ}15'N$, $37^{\circ}10'E$	lava flow
JJM	JJM11 Jabal Dhaylan,	$25^{\circ}25'N$, $37^{\circ}15'E$	dyke-like plug
AZL	Previously unknown outcrop, Wadi As Sirr rift valley,	$26^{\circ}55'N$, $36^{\circ}10'E$	probable dyke
MED	Medina, Harrat Rahat	$24^{\circ}30'N$, $39^{\circ}35'E$	lava flow
JAK	Harrat Al Buqum, Wadi Bidah	$20^{\circ}40'N$, $41^{\circ}25'E$	lava flow
SIN	Jabal Abu Hamata, Sinai	$29^{\circ}00'N$, $33^{\circ}20'E$	lava flow

The following localities refer to samples from Egypt. The analyses of these rocks are unpublished and have been kindly made available to the writer by Fikry Khalef.

SER	Serabit el Khadim, Sinai	$29^{\circ}02'N$, $33^{\circ}27'E$	
BIR	Bir el Hamma, Sinai	$30^{\circ}38'N$, $33^{\circ}33'E$	Cretaceous intrusion
TOW	Tower, Cairo-Suez area		
HAM	Hamza, Cairo - Suez area	$30^{\circ}15'N$, $31^{\circ}38'E$	
GOW	Gebel Owenait, Western Desert		
MAN	Mandisha, Western Desert		
MAY	Maysran, Western Desert		
BAS	Basalt Hill, Western Desert		
HUF	Hufuf, West Desert		

PETROGRAPHY OF BASALTS

The coastal basalts (including additional specimens from near Umm Lajj, Jeddah, Rabigh, Medina, Yanbu an Nakhl, and Harrat Snuewa) have a uniform composition and, where medium grained, fairly uniform texture. The very fine grained basalts (AZL, JJM, MED) range from phyrlic to porphyritic and are more variable in mineral proportions. All the basalts contain olivine though in some it is partially altered. It is commonly euhedral and slightly porphyritic. In RAB it is glomeroporphyritic. The pyroxene is augite and its common mauve colour suggests it is titan-augite. This is further suggested by hour-glass and sector zoning which is also common. Twinning and zoning however appear to be absent from augites in some thin sections. The augites form granular (when fine grained) to ophitic textures and are euhedral to anhedral respectively. The abundant augite in JJM is glomeroporphyritic.

Plagioclase ranges from An 45 (JJM) to An 65 (AZL); the lavas however are generally about An 50. Grains are euhedral to subhedral and are commonly normally zoned. Porphyritic plagioclase is associated with a fluidal fabric in several specimens. The opaque minerals are equant (magnetite?), tabular (ilmenite?), and less commonly interstitial and skeletal.

Most specimens appear to be holocrystalline however glass is apparent in a few fine grained specimens. Although zeolites were not identified some clear isotropic interstitial material in several specimens is possibly a zeolite or more probably analcite. Interstitial calcite occurs in several specimens and opal was found in JAK.

ANALYTICAL TECHNIQUES

Before the Red Sea deeps samples became available some of the coastal samples (AZL, JJM, QAL, MER, RAB) were partially analysed using AAS for Al_2O_3 , MnO , Fe_2O_3 (as tot. Fe), CaO , MgO , K_2O , Zn and Cu; and colorimetry for P_2O_5 and TiO_2 . This work was carried out in the departmental geochemistry laboratory under the guidance of A.J. Thompson.

When the Red Sea deeps samples became available all the samples were analysed for major elements (except P_2O_5) and trace elements by XRF in Dr. G. Borley's laboratory under her supervision. Several of the Na_2O results were obviously wrong due to machine trouble so the analyses were repeated except for Na_2O and trace elements. Mr. A.J. Thompson then very kindly made analyses for Na_2O , MgO , K_2O and Zn (by AAS), FeO (by Titrimetry) and P_2O_5 (by Colorimetry).

The methods used for analysis were the standard procedures used in the above laboratories. Dr. Borley's notes ("methods of sample preparation-analysis of silicate rocks by automatic X-ray spectrometry") were closely followed and need not be reiterated here. It is necessary to note however that for major elements the sample powders were fused one part to seven with lithium tetraborate. With two of the Red Sea samples (RSA, RSB) insufficient material was available for this so 0.4g of each of these samples were diluted to one part in 14 lithium tetraborate. To give a

duplicate of the other Red Sea sample this dilution was also applied. That is RSA, RSB and RSD were diluted to one part in 14 while RSC and all other samples were diluted to the standard one part in 7. H_2O^- was determined after heating overnight at $110^{\circ}C$ and H_2O^+ was determined after heating at $850^{\circ}C$ for 30 minutes or longer when necessary. Because of sample shortage with the Red Sea material trace elements were determined on undiluted powders before major element analysis began. The raw results for major elements were processed by computer using a program (available from the XRF laboratory) which also accounts for various corrections including matrix effects. These corrections were not made for the trace element results as they are relatively small with heavy elements and are likely to be within the limits of error.

GENERAL RESULTS

The accuracy of the analyses is not as high as was hoped but it is adequate for the present purposes. The totals of the major element oxides range from 97.42% to 101.18% but this is slightly misleading because the totals of 5 check analyses by (XRF on standard rock powder discs) range from 97.99 to 102.19%. It can be seen by comparing the pairs of the duplicates RSA, RSB and RSC, RSD that the values for SiO_2 are out by as much as 2.16% (note also that the corresponding totals differ by 3.28%). It can also be seen that the trace element results are rather inaccurate (though some AAS repetitions by the writer showed a reasonable precision). The results Zn (a) are from XRF, Zn (b) from AAS by the writer, Zn (c) from AAS by A.J. Thompson and Cu (a) from XRF, Cu (b) from AAS by the writer.

The norms of all the analyses of major elements were calculated by computer using a program devised by R.W. Le Maitre and based on the method of norm calculation of Kelsey (1965). The results are given in table 7.

MAJOR ELEMENT RESULTS AND SIGNIFICANCE

Although no nepheline was observed in thin section (possibly because it is unidentifiable in these very fine grained rocks) the basalts of the Red Sea deeps and the Arabian Plateau are nepheline normative except for

Sample	RSA	RSB	RSC	RSD	RAB	MER	QAL	JJM	AZL	MED	JAK	SIN	SER	BIR	TOW	HAM	GOW	MAN	MAY	BAS	HUF
SiO ₂	47.93	48.04	49.32	47.15	44.50	44.09	46.95	41.58	47.45	45.51	43.33	45.39	49.70	45.69	48.19	48.53	46.30	48.57	47.53	48.94	49.18
TiO ₂	1.48	1.48	1.62	1.55	1.15	1.78	1.36	2.72	3.42	2.80	2.54	1.32	1.25	2.11	2.86	2.41	2.72	1.73	1.98	1.94	1.82
Al ₂ O ₃	14.96	16.41	13.96	13.56	16.10	15.33	17.13	14.84	14.23	15.94	14.95	17.65	17.45	10.93	11.61	13.87	14.82	13.33	13.44	12.71	12.65
Fe ₂ O ₃	1.23	1.23	1.00	0.74	4.73	3.62	2.01	5.22	6.05	2.43	4.06	2.74	1.70	5.44	6.96	6.82	5.49	3.39	2.79	3.35	4.52
FeO	11.16	(11.11)	(11.10)	(11.10)	5.25	7.29	7.52	5.85	7.71	10.27	7.91	6.91	7.35	10.02	8.00	5.98	7.09	7.05	7.46	7.33	6.12
MnO	0.20	0.21	0.20	0.20	0.15	0.17	0.16	0.19	0.21	0.20	0.20	0.16	-	0.23	0.19	0.14	0.13	0.13	0.13	0.13	0.13
MgO	7.95	8.03	7.55	7.58	8.43	8.33	8.98	9.08	5.46	6.87	8.27	7.22	6.85	10.72	6.10	5.31	6.35	8.64	7.97	8.56	8.06
CaO	11.13	11.26	11.38	11.18	11.36	10.40	10.11	10.36	9.85	9.07	9.07	10.91	9.00	9.30	9.76	9.45	9.58	8.34	8.96	8.17	8.18
Na ₂ O	2.42	(2.41)	3.34	(3.34)	2.98	2.55	3.05	3.97	2.89	3.84	3.72	2.98	2.95	2.00	1.84	2.76	3.59	3.73	4.15	3.78	3.58
K ₂ O	0.14	0.15	0.19	0.18	0.45	0.84	0.70	1.11	0.56	0.77	1.12	0.51	0.90	0.63	1.74	0.95	1.20	1.64	1.61	1.63	1.58
H ₂ O ⁺	0.31	0.69	0.80	0.80	3.51	2.74	2.05	2.29	1.24	0.13	1.90	2.92	2.40	1.59	0.55	1.07	1.57	1.57	1.90	1.39	2.23
H ₂ O ⁻	0.00	0.00	0.10	0.10	0.40	0.55	0.57	0.77	0.17	0.00	0.59	0.54	0.00	0.36	1.55	1.37	0.69	0.68	0.49	0.56	0.47
P ₂ O ₅	(0.14)	0.14	(0.14)	(0.14)	0.15	0.36	0.26	0.73	0.42	0.66	0.51	0.15	-	0.54	1.18	0.43	0.46	0.52	0.61	0.67	0.56
TOTAL	99.05	101.18	100.70	97.42	99.16	98.05	100.85	99.31	99.66	98.49	98.17	99.20	99.55	99.36	100.33	99.09	99.99	99.32	99.02	99.14	99.08
Quartz	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	3.39	0.00	0.00	0.00	0.00	0.00	6.91	5.77	0.00	0.00	0.00	0.00	0.00
Orthoclase	0.83	0.89	1.12	1.06	2.66	4.96	4.14	6.56	3.31	4.55	6.62	3.01	5.32	3.72	10.28	5.61	7.09	9.69	9.52	9.63	9.34
Albite	20.48	20.39	26.47	22.69	18.94	19.60	22.46	10.25	24.45	24.78	18.00	20.77	24.96	16.92	15.57	23.35	27.01	28.38	25.85	30.81	30.29
Anorthite	29.55	35.52	22.54	20.94	29.23	27.91	30.99	19.40	24.24	23.99	20.79	33.28	31.72	18.99	18.28	22.66	20.78	14.79	13.29	12.90	13.79
Nepheline	0.00	0.00	0.97	3.02	3.40	1.07	1.81	12.65	0.00	4.18	7.30	2.40	0.00	0.00	0.00	0.00	1.82	1.72	6.10	0.63	0.00
Diopside	20.40	17.65	27.23	27.75	20.74	17.15	14.11	21.41	17.51	13.68	16.76	16.06	10.52	20.07	17.93	16.88	18.94	18.60	21.90	18.67	18.41
Hypersthene	9.95	7.89	0.00	0.00	0.00	0.00	0.00	0.00	9.14	0.00	0.00	0.00	13.99	17.61	11.19	6.91	0.00	0.00	0.00	0.00	6.20
Olivine	12.62	15.21	16.62	16.75	10.89	14.61	18.62	11.57	0.00	16.82	14.32	13.58	5.80	7.41	0.00	0.00	7.89	14.48	12.75	14.47	7.01
Magnetite	1.78	1.81	1.45	1.07	6.86	5.25	2.91	7.57	8.77	3.52	5.89	3.97	2.46	7.89	10.09	9.89	7.96	4.91	4.04	4.83	6.55
Ilmenite	2.81	2.81	3.08	2.94	2.18	3.38	2.58	5.16	6.49	5.32	4.82	2.51	2.37	4.01	5.43	4.51	5.17	3.29	3.76	3.68	3.46
Hydroxy-fluor-apatite	0.33	0.33	0.33	0.33	0.35	0.85	0.61	1.72	0.99	1.56	1.20	0.35	0.00	0.80	2.78	1.01	1.09	1.23	1.44	1.58	1.32
Diff. Index	21.30	21.28	28.56	26.77	25.00	25.64	28.41	29.45	31.15	33.51	31.92	26.19	30.28	20.65	32.76	34.74	35.93	39.79	39.47	41.08	39.63
S.I. unsat.	4.57	5.52	6.81	8.58	7.27	6.58	8.66	15.49	0.00	9.72	11.76	7.22	2.18	2.86	0.00	0.00	4.67	7.11	10.07	6.17	2.82
TRACE ELEMENTS in ppm																					
Pb	9	7	7	-	9	9	8	8	8	7	8	8									
Zn (a)	90	100	370	-	85	80	80	170	100	85	80	80									
(b)	n.d.	n.d.	n.d.	-	85	80	85	n.d.	120	n.d.	n.d.	n.d.									
(c)	n.d.	n.d.	n.d.	-	90	85	75	110	140	95	95	70									
Cu (a)	80	110	370	-	135	115	120	120	90	90	95	87									
(b)	n.d.	n.d.	n.d.	-	155	100	130	115	105	n.d.	n.d.	n.d.									
Ni	105	105	123	-	240	155	150	60	60	95	160	85									
Cr	220	250	230	-	285	270	270	95	95	125	235	210									
Ba	150	150	150	-	200	275	220	420	290	220	285	235									
Sr	90	90	95	-	255	550	430	350	350	440	640	475									
Rb	15	15	15	-	20	25	20	20	15	15	20	15									
Y	30	30	30	-	25	30	25	30	30	30	30	30									

Figures in brackets were estimated from duplicates.

Table 7. Chemical analyses and C.I.P.W. norms of basalts from Red Sea, Saudi Arabia and Egypt.

RSA, RSB and AZL which are hypersthene normative. They also are all olivine normative except for AZL which is quartz normative but which contains modal olivine. This rock is very fine grained and the olivine is generally altered. Euhedral feldspars occur as inclusions in an olivine grain from this rock. Using Wilkinson's (1967) criteria AZL is an oversaturated tholeiite, the other Arabian Plateau basalts are alkali basalts.

The glassy Red Sea basalt is olivine tholeiite as are those of Schneider and Wachendorf (1973) and one of Chase's (1969) Red Sea basalts. The phyrlic basalt however is alkali basalt. Chase's other Red Sea basalt analysis corresponds to oversaturated tholeiite. He pointed out however that the values for Fe_2O_3 are maximum values computed from the gain in weight when the samples were heated from 500°C to 1000°C . The normative quartz in the oversaturated tholeiite is therefore suspect. Chase's samples were analysed by the optical emission spectrometer method and his SiO_2 content was determined by difference of the sum of the cations plus P_2O_5 and ignition loss at 1000°C from 100.00%.

The Red Sea deeps material analysed in this study compared with that analysed by Chase is higher in CaO and lower in Ti, Fe tot., $\text{Fe}_2\text{O}_3/\text{FeO}$, K_2O and P_2O_5 and therefore is more indicative of an oceanic origin. The writer's results are also close to those for Atlantis II Deep glass as analysed by Schneider and Wachendorf (1973). All of these analyses correspond to those of oceanic tholeiites given by Chase (1969) and Cann (1971) except for some variation in SiO_2 , Al_2O_3 and Na_2O which is probably due to experimental error. Cann's values for SiO_2 are 49.61 ± 0.72 , Al_2O_3 16.01 ± 0.85 and Na_2O 2.76 ± 0.25 .

Cann further noted that there is a continuous gradation from tholeiitic types to alkaline types in oceanic basalts and recommended that the term "oceanic tholeiite" be abandoned.

The basalts of Jabal Zukur compared with those of the Red Sea deeps, Jabal Teir, the Arabian Plateau, the Ethiopian Rift and the East African Rift (Chase, 1969) are low in Mg and Ca, and high in Ti, Na and K. These basalts therefore are somewhat anomalous but resemble AZL to some extent in Ti and Mg. The basalts of Jabal Teir are intermediate between the Red

Sea deeps and the Arabian Plateau in K and Ti but are lower than both in Mg in which it tends to resemble Jabal Zukur (Chase, 1969). Compared with the Ethiopian Rift they are similar in Ti, Mg, Ca but are lower in Fe tot. and K.

Compared with the basalts in the Red Sea deeps the Arabian Plateau basalts have higher TiO_2 , K_2O , P_2O_5 , Fe_2O_3 and lower SiO_2 , FeO and CaO. Potash contents do not increase with distance from the coast. Intrusions and lavas distant from the coast however contain more Ti, P_2O_5 and H_2O but it must be emphasised that viable generalizations cannot be made of such a small number of samples.

The basalts of the Aden Series in the Ethiopian Rift have higher K_2O and TiO_2 contents and lower MgO than the Red Sea deeps basalts (Chase, 1969). They are rather similar to the Arabian Plateau basalts except that they have higher Fe tot. and lower MgO and K_2O .

Compared with the East African Rift, and Red Sea deeps basalts are higher in SiO_2 and lower in TiO_2 and K_2O . The Arabian Plateau basalts are also lower in K_2O , MgO and Fe tot. but higher in Al_2O_3 .

The Egyptian basalts are not as uniform in composition as the Arabian Plateau basalts. The analyses received were screened using Manson's (1967) criteria and several were rejected. Generally the Egyptian basalts are higher in SiO_2 , K_2O , P_2O_5 and lower in Al_2O_3 and MnO.

The two Cairo-Suez analyses represent oversaturated tholeiites and another that was rejected had 14% normative quartz. The basalts of the western Desert like those of the Arabian Plateau are mostly alkali basalt. The three from Sinai are intermediate between those of Cairo-Suez and those of the Arabian Plateau or the Western Desert.

Insufficient samples have been examined for differentiation trend studies nevertheless the traditional AMF and KCN diagrams are given in figure 56 for comparison with the Erta Ale rocks of Ethiopia (Barberi and Varet, 1970, Barberi et al, 1970). A disturbing feature of these diagrams is the wide

spread of values for the Red Sea samples although the two samples of glass plot together.

In general the results show that the Arabian Plateau and the Egyptian Western Desert are alkali basalt provinces, the Sinai-Suez-Cairo area is tholeiitic and the Red Sea deeps basalts are oceanic tholeiites with a slight trend toward rift type basalt (especially for those analysed by, Chase 1969). The composition to some extent also reflects the tectonic setting.

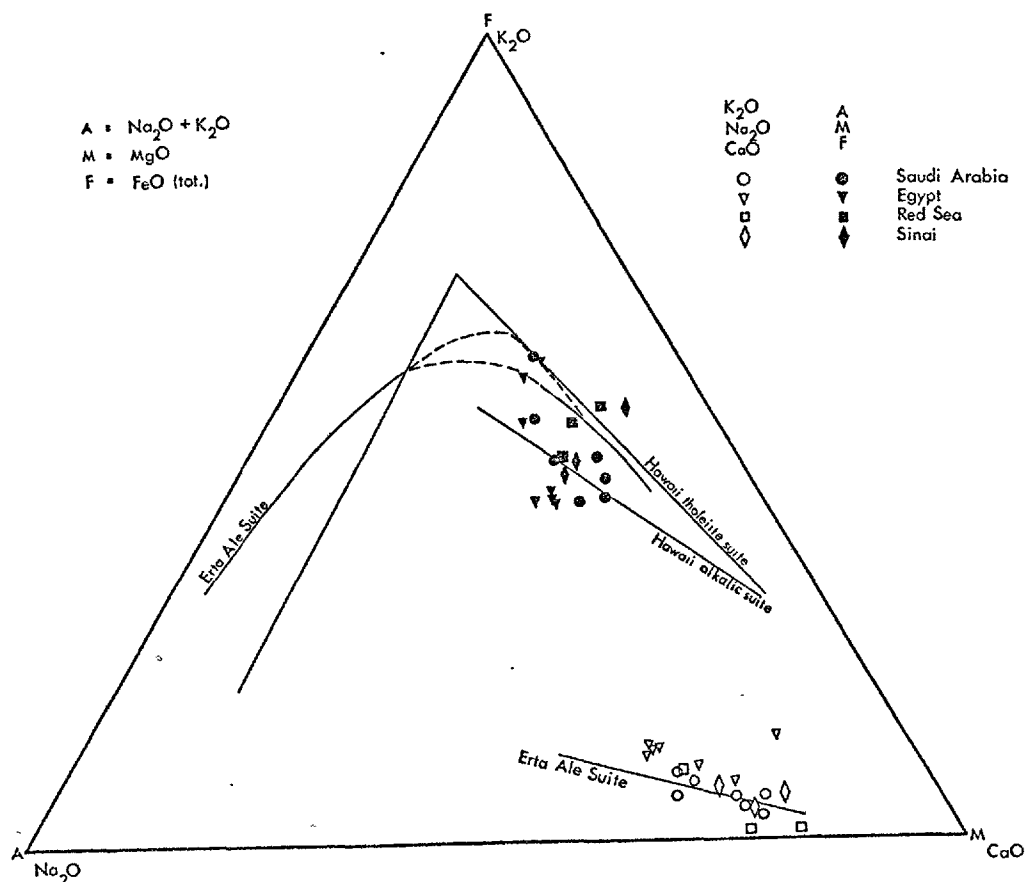


Figure 56 A-M-F and K_2O - CaO - Na_2O diagram of basalts from Saudi Arabia, Egypt, Red Sea and Sinai.

TRACE ELEMENT RESULTS AND SIGNIFICANCE

Apart from a tendency for the intrusive basalts to be slightly high in Zn and Ba, and low in Ni and Cr there is little indication of a direct relation between mineralization and basalt chemistry in the Plateau basalts. The phyric basalt of the Red Sea deeps however has an abnormally high Zn and Cu content but a normal lead content. As the glassy basalt has normal values for these elements it is likely that it is more representative of the magma as a whole. Possibly the magma was chilled to glass (with normal values) and behind the glassy skin volatiles with Zn and Cu collected as the magma crystallized to give the phyric basalt. The resulting Cu, Zn values would therefore be localized. Prinz (1967) noted that Cu tends to accumulate with differentiation of basalt magma. Unfortunately the copper values of the Arabian Plateau basalts tends to decrease slightly as the differentiation index increases. It is unlikely that the higher Zn and Cu values are spurious as they relate to the same rock and no other spurious results were found in the trace element data. Also the rock appears under the microscope to be quite fresh and the major element analysis collaborates with this.

The Cr, Ni and Y values appear to be normal for basalts (Prinz, 1967). Ba, Rb, Sr and K are higher in the plateau basalts than the deep sea basalt. JJM which is high in K and Ba however is not high in Rb and Sr. The Rb/Sr ratio is much higher in the deep sea basalt than the coastal plateau basalts. This indicates a much higher mantle source for the submarine basalt (Prinz, 1967). Haenel (1972), on the basis of heat flow measurements, considered that wet gabbro below the Red Sea axial trough would be molten at the relatively shallow depth of 12 km. The plateau basalts are more akin to the Yemen Trap Series and the Aden Volcanic Series than to the Red Sea deeps in many respects hence it is likely that the plateau basalts have a relatively deep source e.g. 60 to 100 km as postulated by Baker et al (1973) for the al Abyad volcanics. This is compatible with a depth of 60 km postulated by Cox, et al (1970) for the depth of magma generation for the Trap Series. Brown (1972) noted that mantle xenoliths in coastal lavas 5 m.y. old or younger point to magmatic sources at greater depth than those for the oceanic basalts of the axial trough.

PART I	DESCRIPTIVE GEOLOGY	Section 1	The Red Sea
		Section 2	The Coastal Region
		Section 3	The Cenozoic Igneous Realm
		Section 4	The Shield

SECTION 4

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SECTION 4

THE SHIELD

1.4.1

GENERAL GEOLOGY

Some aspects of metallogenesis and microorganisms in the Red Sea region of Saudi Arabia involve the Arabian Shield but this area, which exceeds 500,000 km², can only be treated here as a fringe subject. The tectonic development of the Red Sea involves the shield area particularly and evidence of the activity of iron precipitating microorganisms have been found not only in the Recent material of the Red Sea deeps and in the Tertiary formations of the coastal region but also in Precambrian basement rocks of the shield.

The shield is essentially a plateau of granitic and metamorphic rocks exposed by erosion coupled with broad arching centered in the region now occupied by the Red Sea. The Egyptian shield is roughly a symmetrical reflection of the Arabian shield about the Red Sea. The Arabian shield deviates from a semi-circular outcrop shape in the north-east. There is a large north-easterly trending fault zone in this area and the shield may have ruptured along this zone during arching. This fault zone has affected the Paleozoic sedimentary rocks to the north-east and may be related to a transform fault in the Red Sea because St. John's Island (section 1.1.5) lies along this direction (see the UNESCO International Tectonic Map of Africa 1:5,000,000 scale 1968). The easterly boundary of the shield is rather straight and has a direction intermediate between that of the Red Sea and the Najd fault system. Viewed at a scale of 1:10,000,000 the shield appears to have a NS (meridional) grain and a NW (Najd) grain caused essentially by the outcrop shapes of the granitic formations.

The oldest dated rock in the shield is granite 1000 $\pm$ 50 m.y. old, but an extensive "basement gneiss" is probably older than this and the greenstone intruded by the granite also is, provided the field relations have been correctly interpreted (Brown, Delfour and Coleman, 1972). An ophiolitic suite along the eastern margin is overlain by 10,000 m of chlorite-sericite schists. In the southern part of the shield migmatic gneisses and paraschists were metamorphosed during a period ending with granitization about 615 m.y. ago (Brown, Delfour and Coleman, 1972). Throughout the

northern part of the shield 8,000 m of volcanics associated with ophiolites are thought to have accumulated between 660 and 1000 m.y. ago. Psammitic clastic rocks 6,000 m thick, cut by 620 m.y. old granite disconformably overlie the volcanics and in turn were overlain by acid volcanics, 3700 m thick, during the Pan African orogeny which culminated at about 570 m.y. ago. The youngest suite of the shield disconformably overlies the others and consists of volcanic rocks, clastic rocks and limestone 520-550 m.y. old deposited during the Najd Orogeny (Brown, Delfour and Coleman, 1972).

Several orogenies affected the shield, however the main episodes of plate collision and magmatism were the Kibaran, the Hejaz and the Najd orogenies (Brown and Coleman, 1972). The Kibaran Orogeny is the oldest and occurred $1,000 \pm 50$ m.y. ago. It has a NE trend but otherwise is not yet well known. The Hejaz Orogeny culminated about 660 m.y. ago with acid volcanism and synorogenic granite intrusion dominantly in NS belts. It possibly correlates with the Mozambique Orogeny. The Najd Orogeny (using the broad sense of the term "orogeny") possibly correlates with a Pan-African tectonic thermal event. It partially overlapped the Hejaz Orogeny and culminated about 570 m.y. ago with the emplacement of granite batholiths. A later phase 520 m.y. was confined to emplacement of granite stocks, andesitic volcanism and left lateral displacement by as much as 240 km along a NW fault system which can be traced 1200 km (Brown and Coleman, 1972).

SECTION 4

THE SHIELD

1.4.2

MINERALIZATION

Although a large part of the shield is occupied by granitic rocks relatively few orebodies occur in them. The most economically significant rocks, according to Eijkelboom (1970) are weakly metamorphosed acid pyroclastic rocks, especially those of the Halaban Group. Such rocks contain the polymetallic sulphide deposits at Jabal Sayid, Al Amar, Musayna'ah, Nuqrah, As Safra, Jabal Smaran, Umm ash Shalahib, and Wadi Bidah. The pyrite deposit at Wadi Wassat also occurs in such rocks as do the gold bearing veins in the north-western Hejaz and at Mahd adh Dhahab (Johnson and Trent, 1966). Contrary to what might be expected, microorganisms appear to have flourished in Precambrian subaqueous environments, of pyroclastic deposition in Saudi Arabia. This is further discussed with respect to Wadi Wassat, Wadi Bidah, Mahawiya, Jabal Samran and Wadi Sawawin in section 2.4.1. The Wadi Sawawin iron ore deposit occurs in the Silasia Formation which Eijkelboom (1970) considered to be probably equivalent to the Halaban Group which would date it between 600 and 1000 m.y. old but the writer considers that as this deposit appears to be a Lake Superior type is therefore more likely to be nearer 2,000 m.y. old than 1,000 m.y.

PART II

GEOMICROBIOLOGY

PART II GEOMICROBIOLOGY

Section 1	Introduction
Section 2	Red Sea Deep
Section 3	Coastal Plain
Section 4	Shield
Section 5	Phytoclast Geothermometry

SECTION I

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SECTION I

I N T R O D U C T I O N

2.1.1

SIGNIFICANCE TO THIS WORK

One of the original aims in this study was to find new methods of using microorganic remains in the search for ore. This began with a study of the application of phytoclast geothermometry but went on to a broader study of the relations of microorganic activity to ore formation within the metallogenic provinces of the Red Sea, the coastal plain and parts of the shield.

It must be emphasised that the writer has not attempted a palaeontological study in the classic sense, concerned with taxonomy, dating and correlation, but has sought direct relationships between ore genesis and activity of microorganisms. This study is one of ore geomicrobiology or economic lithobiontics rather than palaeontology.

The Red Sea region was well suited to this study for the following reasons:

1. It contains sediments of widely differing ages with microbial remains.
2. Microorganic remains were found during this study in a wide range of environments, some of which are very unusual.
3. Evidence found suggests microorganisms have directly and indirectly participated in the deposition of ore minerals in this region.

2.1.2

PERSPECTIVE IN GEOMICROBIOLOGY

Views on the role of microorganisms in the formation of ore deposits commonly gravitate to one of three extremes viz. -

1. microorganisms play virtually no part in the formation of ore.
2. most syngenetic deposits are formed by microorganisms.
3. many syngenetic deposits probably are biogenic but it is almost impossible to establish a biogenic origin.

Much of the controversy on the origin of possible biosyngenetic ore deposits derives from epistemic-ontological confusion (the case as it exists confused with what can be known to be the case) and from imperspicuous use of the term "biogenic". The fact that the occurrence of an ore deposit can be explained without reference to microorganisms does not necessarily mean the deposit is abiogenic. Also, the fact that many deposits can be attributed to activity of microorganisms by itself does not disprove abiogenic theories. Sulphur deposits for example can be biogenic, volcanogenic or partially both. Biological processes can form major controls on ore deposition (e.g. as with phosphate deposits) or minor controls (e.g. coral reefs as favourable host rocks to certain epigenetic deposits) hence care is needed with terminology.

As the techniques of biologists (unlike those of palaeontologists) largely depend on active processes of living organisms they are naturally sceptical about fossils as evidence on the role of microorganisms in producing ore deposits in the past. This is reinforced by the fact that many microorganisms, e.g. iron bacteria, have relatively simple morphologies.

The great potential of microorganisms can be grasped from an understanding of their role in the world today. They form three quarters of all living matter by weight and have a great variety of metabolic processes (Butten, 1971). They can concentrate elements by modifying the physico-chemical environments, by acting as catalysts, and by accumulating elements in their soft bodies, skeletons or encrustations. The annual production of organic matter on earth is 50-100 billion tons (Kuznetsov et al, 1963). Mason (1958) pointed out that "If the mass of the biosphere has been

approximately constant over the last 500 million years and if the average life cycle were one year, then the total amount of matter that has passed through the biosphere in that time is comparable with the total mass of the earth". Microorganisms can also live in a wide variety of environments including fumarolic areas. Baas Becking and Moore (1961) found that the Eh-pH limits of biological systems and of the natural aqueous environments almost coincide. Various microorganisms can live at pressures ranging from 0.1 atm. to 1000 atm. (spores can survive 12,000 atm), temperatures ranging from 104°C to -30°C , pHs ranging from 1 to 12 (some can grow in hot normal sulphuric acid) and salinities ranging from 0 to saturated (many thrive in the 30% salt solution of the Dead Sea) and some can grow in solutions containing as little as 0.1 mg/l of organic matter (Kuznetsov et al, 1963, Zajic, 1969). H. Puchelt (Universitat Tübingen, W. Germany, pers. com.) found sulphate-reducing bacteria growing naturally in an oil reservoir at 120°C , and J. Angus (Imperial College, 1972, pers. com.) found that some sulphate-reducing bacteria apparently survived autoclaving at 130°C . Molten et al (1973) found bacteria (not sulphate reducers) that even survived 160°C .

It is a common misconception that the Precambrian was relatively barren. Although macro-fossils (other than stromatolites are rare) microfossils are relatively abundant. Microorganic remains have been found even in some of the oldest known sedimentary rocks (e.g. the Fig Tree Group, 3.1×10^9 years old, Schopf, et al 1971) and Barghoorn (1957) noted that Precambrian coals form impressive evidence of the abundant and widespread occurrence of organisms at that time. It should further be noted that much of the organic matter deposited in the Precambrian probably was consumed during subsequent metamorphism. Once life became established on the earth it would have spread like a plague at the fantastic speeds that organisms are capable of spreading at today in the absence of competition and predators. A single bacterium with a generation time of 20 minutes could produce a progeny of 2^{144} cells after only 48 hours (Stanier et al 1970). Although the average bacterium weighs only 10^{-12} g this population would weigh 2×10^{31} g or roughly 4,000 times the mass of the earth. After the initial colonisation, all the suitable environments for habitation would have quickly filled and the numbers of organisms would then depend largely on the available environments and adaptation to new environments.

The numbers of organisms increased with evolution but most probably not in the linear fashion commonly assumed. There could be a difference of the type shown diagrammatically in fig. 57.

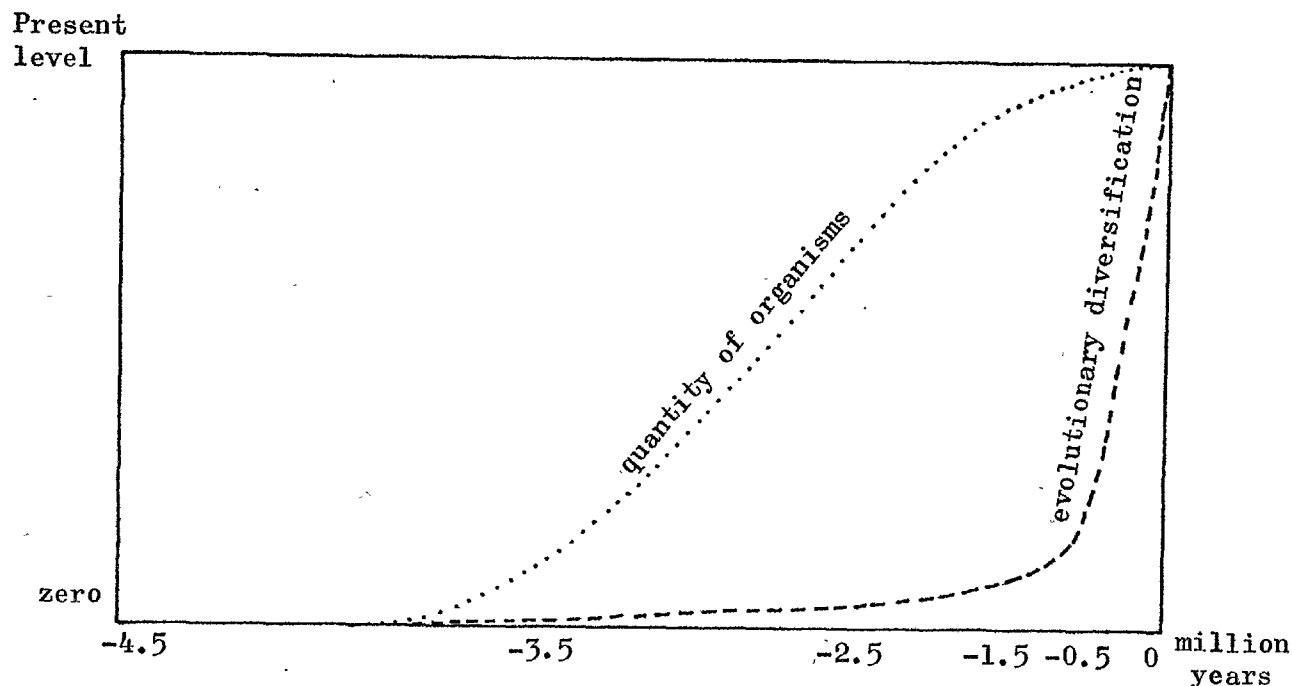


Fig. 57. Diagrammatic qualitative representation of quantity of organisms through time compared with evolutionary diversification.

As early as 1836 Ehrenberg found that organisms play an important part in the formation of bog iron ore. In 1919 Harder produced a thorough study of the geological possibilities of iron bacteria and found them to be quite widespread. In 1922 Gruner found fossil bacteria in the great iron ore deposits of the Mesabi range. Cloud (1965) refuted Gruner's discovery but claimed to have observed structures closely resembling iron bacteria in the Gunflint Iron Formation near the locality where a similar discovery was made by Tyler and Barghoorn in 1954.

It is clear that microorganisms have great possibilities as ore producers but with the evidence available one needs to maintain a careful perspective. It is fitting therefore to conclude with the following diverse statements of authoritative opinion.

1. "Plants and animals from the largest to the most microscopic play an intense and dramatic part in the concentration of certain elements Collectively these concentrations of heavy metals in sea animals are enormous" Lindgren (1933, "Mineral Deposits").

2. "There is a tendency now to ascribe a number of geological processes to microbial action without a thorough investigation of the geochemical and geobiological factors involved. Vague statements as to the virtues of microorganisms in geological happenings are of very little use to us. On the contrary, as they are discursive, and often not based on experiment, they give welcome fuel to those critics, who, subconsciously perhaps, want to maintain the 'sterile' 'status quo ante'". Baas Beeking and Moore, (1961, "Biogenic Sulphides").

3. "Many large and important ore deposits have been considered products of bacterial activities, but for most of these a hydrothermal origin cannot be disproved. If the metals were deposited by bacteria at the bottom of stagnant seas - similar to the present Black Sea - they should have been precipitated as simple minerals in an intimately mixed or rhythmic sequence; no such sulphide deposits are known". Park and MacDairmid, (1964 "Ore Deposits").

4. "Deposits whose localization and mineralogical zoning are clearly spatially related to sedimentary facies may well be sedimentary precipitates 'in toto' and localized at least partially by biological agencies". Stanton (1972, "Ore Petrology").

2.1.3

METHODS OF STUDY

The method of study depended on the type of material being examined and the nature of the examination. For much of the work it was necessary to isolate the phytoclasts by maceration. For normal silicates this meant routine maceration as practiced in Dr. M. Muir's laboratory and elsewhere (essentially dissolution of the carbonates and silicates in hydrochloric and hydrofluoric acids). For more difficult rocks such as sulphides, techniques such as those outlined by Gutjahr (1966) and Saxby (1972) were used.

Ordinary thin sections were used separately or together with maceration preparations and have the advantages of being more easily prepared and less structurally disturbed. Macerations however have the advantages of giving a concentration of organic matter which can be studied in a variety of ways. As few minerals can withstand maceration (especially Saxby's and Gutjahr's techniques) positive identification of the organic matter is enhanced. The disadvantages of maceration are that contamination is difficult to control, inorganic biogenic structures are lost, and some of the organic matter is broken down physically to an incoherent mess.

The Red Sea sediments were generally studied directly as smear mounts although some macerations and polished grain mounts were examined.

The preparations were studied by transmitted light microscope (TLM) and by SEM with XRA, mostly. Less commonly they were studied with a reflected light microscope, a TEM and an electron microprobe.

A series of etched specimens was also examined by SEM but unfortunately all the photographic results were accidentally lost by a technician. It was noted however that the technique is more successful with etched polished sections than etched rock fragments and that the degree of etching is critical.

Further details on methods are given by Muir, Hamilton, Grant and Spicer (1974).

2.1.4

EVIDENCE OF MICROBIAL ACTIVITY

One of the major problems in this study has been in obtaining evidence of microbial activity. This problem does not usually confront the biologist who deals with living organisms, nor the palaeontologist working with relatively complex forms. The difficulties become acute however when dealing with relatively simple structures that cannot be cultured in the laboratory. Often, especially in economic geomicrobiology, evidence of the reactions of microbes cannot be used as a criterion of genesis because one is concerned with showing that the reaction which deposited ore was caused by organisms whose presence and activity has been established (circular paralogism). Some deposits are easily established as biogenic e.g. coral reefs but others can be extremely difficult. The goethite facies in the Red Sea, for example, was studied by TLM and SEM without satisfactory results. This work was further continued by TEM for some time before a reasonable conclusion could be drawn. This fundamental problem is equally acute in the field of exobiology where much controversy has developed over the evidence of extraterrestrial life found in meteorites (Urey, 1966; Brooks and Muir, 1971).

As microorganisms in unstable environments often produce reactions that otherwise would inevitably have happened in their absence they are often overlooked. In these circumstances also the occurrence of such reactions cannot be used as evidence of the presence of microorganisms.

Evidence can be chemical, morphological and circumstantial. When organic matter is present chemical evidence can be substantial. Certain porphyrins, isoprenoids and n-alkanes found in rocks of all ages are considered to be almost exclusively biogenic. Carbon and sulphur isotope ratios are also commonly used criteria. Very often, due to autolysis, diagenesis and metamorphism, organic residues escape from localities of biogenic activity, and many fossil organisms e.g. corals and radiolaria, are characterized by carbonate and silica deposits rather than by organic

matter. With these one relies very much on morphological criteria. Schopf et al (1971) failed to find microscopic evidence of organisms in 30 thin sections of stromatolites. He attributed this to recrystallization. The apparent lack of microscopic evidence of biogenic activity has caused strong controversies in the past over the origin of bituminous coal, bog iron ore and certain manganese ore deposits that are now accepted as biogenic (Harder, 1919; Kuznetsov et al, 1963; Perfil'ev et al, 1965).

After a careful search of the literature the writer was surprised to find only two references on morphological criteria for biogenic processes. Cloud and Licari (1968, in a half page abstract) gave the following four criteria:-

1. Abundant indigenous microstructures that cluster narrowly in size.
2. Morphological complexity equivalent to living organisms and little resemblance to non-biogenic structures.
3. Microstructures implying evolutionary continuity with living forms.
4. Complex modifications consistent with reproductive or other specialized structures of living taxa.

The following eight criteria given by Hofmann (1972) are also useful but are applicable more to higher biogenic taxa:-

1. Evidence for the performance of vital functions.
2. Cellular differentiation combined with morphologic consistency in a large number of individuals.
3. Resemblance to living or known extinct forms, and consistency with an evolutionary sequence
4. Morphological diversity in an assemblage representing a population of several distinct groups.
5. Abundance and geographic extent, together with limited stratigraphic range.
6. Mode of preservation.
7. Chemical evidence.
8. Geological plausibility.

The criteria used by the writer are:-

1. **Composition:** preferably the microstructure should be organic or have traces of appropriate organic matter associated with it. (Appropriate isotopic ratios would enhance this).
2. **Morphology:** as noted above from Cloud and Licari (1968), preferably a shape which is typically biogenic and not abiogenic. Unfortunately however some shapes are ambiguous (e.g. microlites in lavas) but these are seldom a problem unless they are also organic as are the double-walled "prebiotic" but non-fossil cells produced by Fox (1965).
3. **Compatibility:** the microstructures should not only be compatible with a biogenic origin but their matrix should also be compatible with their having such an origin.
4. **Associations:** the microstructures should be associated with structures compatible with a biogenic origin. A single isolated microstructure is always suspect but a microstructure which is part of a relatively uniform population is unlikely to be accidental or a freak. Biogenicity is enhanced if the microstructure in question is closely associated with indubitable microorganic remains. Very often one deals not with individual organisms but with colonies, and even when there is no trace of the individuals left the colonial structures (e.g. corals) can be quite distinctive as biogenic objects. Associations should be examined on various scales. For example a possible tropical organism is unlikely to occur in a predominantly glacial sequence, or a photosynthetic organism in a dark environment.
5. **Arrangement:** the microstructures should be in an arrangement typical of a biocenosis however this does not exclude arrangements typical of allochthonous deposits.
6. **Quantification:** if a biochemical reaction is postulated it should at least be feasible in terms of reactor and reactant quantities (and other quantitative considerations) as well as being feasible from a biological point of view (e.g. no poisons).
7. **Minimum Ambiguities:** with the above criteria ambiguities will be minimized with indigenous microstructures. Precautions however need to be taken against exotic material such as contaminants and artifacts. During maceration procedures artifacts are easily produced through the interaction of fluoride with calcium, magnesium and other ions. The procedures are designed against such artifacts however they cannot be entirely eliminated, for example, because of the unusual compositions of some rocks. An example of a particularly troublesome artifact is

shown in fig. 58. The morphology of this artifact resembles some porifera and its composition (a fluoride) was particularly difficult to ascertain.

Contamination is a great problem especially with maceration procedures. Reagents, including hydrofluoric acid, and containers have to be particularly clean and precautions must be taken against airborne pollution. As far as possible, reactions were confined in sealed containers, and as a check against pollution blank samples, samples of dandruff and clothes-brushings, and samples of room dust were prepared and mounted in the same way as normal maceration preparations. Levels of contamination were kept low. Contaminants generally were readily identified by shape, composition, size and heterogeneity. Lycopodium spores were a common contaminant in Dr. Muir's laboratory and when more than a few were found in a preparation the preparation was rejected. Generally indigenous organic structures in rocks are altered by mild diagenesis or metamorphism to some extent. This causes them to gain a slightly darker colour than most contaminants. Thin sections are relatively free of contamination but precautions are still needed. Fig. 59 is an example of a mould in a basalt. After careful examination it was found that the mould was on the cover slip and during thin section preparation was accidentally pressed into the basalt. On enquiry it was found that the room in which the thin sections were prepared favoured such growths and the new coverslips were not always cleaned before use. It was then found preferable not to use permanent coverslips.

Another contaminant can be secondary mineralization. Crystallization in restricted environment, e.g. intergranular boundaries, can produce peculiar shapes (fig. 60). Even more of a problem is secondary biogenic mineral deposition.

Many of the above problems with artifacts and contaminants apply to electron microscope preparations also. In addition, electron beams can build up artifactual carbonaceous deposits on preparations and can form new structures through melting and vesiculation of parts of preparations. Uneven distribution of electrical charges can also give misleading effects. This is discussed further in the following section.

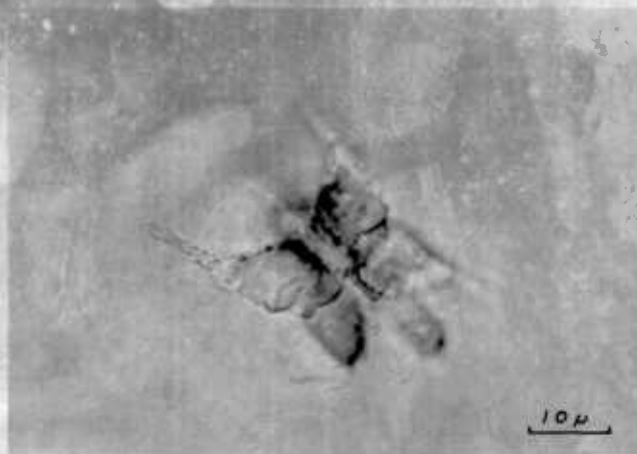


Fig. 58. TLM photograph of artifact in maceration preparation of rock from Jabal Dhaylan (SM in PPL).



Fig. 59. TLM photograph of mold contamination in thin section of basalt lava from Medina (TS in PPL).

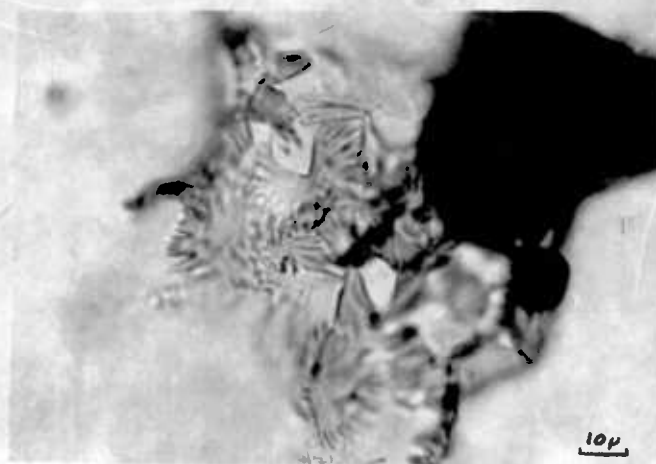


Fig. 60. TLM photograph of secondary goethite pseudo-fossils(?) in thin section of jaspilite from Bomvu (TS in PPL.)

PART II GEOMICROBIOLOGY

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Section 2	Red Sea Deep
Section 3	Coastal Plain
Section 4	Shield
Section 5	Phytoclast Geothermometry

SECTION 2

GEOMICROBIOLOGY
OF THE RED SEA DEEPS

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SECTION 2

G E O M I C R O B I O L O G Y
O F T H E R E D S E A D E E P S

2.2.1

OBSERVATIONS OF GOETHITIC SEDIMENTS

INTRODUCTION

At the time this study began no published data could be found which dealt with the detailed morphology of the goethitic sediment of the Red Sea deeps or with the possibility that the goethitic sediment might be biogenic. Recently however, in a paper concerning sedimentary iron deposits derived from hydrothermal solutions, Chukhrov (1973) noted that no definite relics of iron bacteria were found in sediments from the Red Sea. The writer sent photomicrographs of his observations to Professor Chukhrov in Moscow who replied that in some precipitates from Discovery Deep his team has found aggregates which resemble relics of Gallionella but they are not as typical as are samples from many other localities. He also agreed that the forms shown in some of the writer's photomicrographs do resemble bacteria.

The motivation for this study came from Dr. M. Muir who casually remarked, when observing one of the writer's smear mounts, that the material resembled the nonfilamentous marine iron bacteria Siderocapsa described by Zajic (1969). Closer examination failed to confirm this identification but thereafter stimulated the investigation of the goethite to try to determine whether or not it had a biogenic origin. This presented a major problem. Inorganically precipitated goethite can have complex morphology, and in the Red Sea deeps goethite would precipitate on the mixing of the geothermal brines with sea water even in the absence of microorganisms. Iron bacteria however often live in such environments (Harder, 1919). Dr. M. Muir and the writer now consider that at least some of the goethitic sediment resembles the filamentous iron bacteria Gallionella but as Professor Chukhrov noted it is not typical. It must further be noted however that most illustrations of Gallionella in the literature are of the type that grows in fresh water. Illustrations of salt water Gallionella are difficult to find. Furthermore the morphology of bacteria commonly changes according to the environment and as the environment in the Red Sea deeps is unusual it could be expected that unusual forms would develop there.

Iron bacteria such as Gallionella require little organic matter for growth hence chemical composition is not so useful as a criterion of biogenesis. Culturing and morphology are the main remaining avenues for investigation. Culturing of iron bacteria is difficult and requires specialized expertise. Morphology has been used mostly in this study but this suffers from the lack of information about deep sea iron bacteria, from the relatively simple forms of these bacteria, and from the very small size of the bacterial filaments. Judgement is therefore somewhat subjective but this was compensated somewhat by using an experimental preparation of artificial inorganic goethite as a non-biogenic standard.

Of the four main types of sediment in the Red Sea deeps only the iron-montmorillonite appears to be distinctly non-biogenic. The detrital sediment is largely biogenic. The sulphide sediment is problematical and is discussed later.

The goethitic sediment is commonly found in sediments above basalt in the central rift zone of the Red Sea (Backer and Schoell, 1972) and is not restricted to existing brine pool areas. It closely resembles goethitic sediments in the Mid-Atlantic Ridge region and the East Pacific Rise (Backer and Schoell, 1972; von der Borch et al, 1971 and personal observations on material kindly supplied by Dr. H.E. Cook of California and W.C. Burnett of Hawaii). Many samples collected by the Applied Geochemistry Group of Imperial College on cruises of the "Wando River" 1969 and the "Nereus" 1970 were examined by the writer and in them goethitic layers were found from localities $19^{\circ}42'N$ to $25^{\circ}23'N$. Such layers however are known from $17^{\circ}N$ to $26^{\circ}N$ (Backer and Shoell, 1972; Olausson, 1971). The specimens were examined as smear mounts with an ordinary transmitted light microscope (TLM) and of these, 8 were selected for detailed study by SEM and TEM.

OBSERVATIONAL TECHNIQUES

The use of three different methods of microscopic examination provided valuable data. The TLM obviously is a relatively rapid tool and was used for screening and selecting specimens for further study. It will be seen however that the grains are not the relatively massive objects they generally appear to be under the TLM. Considering the structure of the



Fig. 61. TEM photograph of Red Sea amorphous-goethitic material showing possible contamination (grey halo around filaments) and radiation damage (vesiculated material).

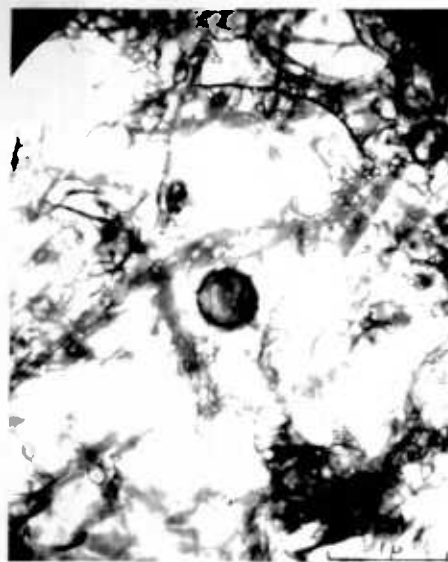


Fig. 62. TEM photograph of same sediment as in fig. 61 showing spherical body and radiation damage (holes in sheets).



Fig. 63. TEM photograph of typical goethitic sediment from Atlantis II Deep showing typical view with organic(?) filaments and "amorphous" material.

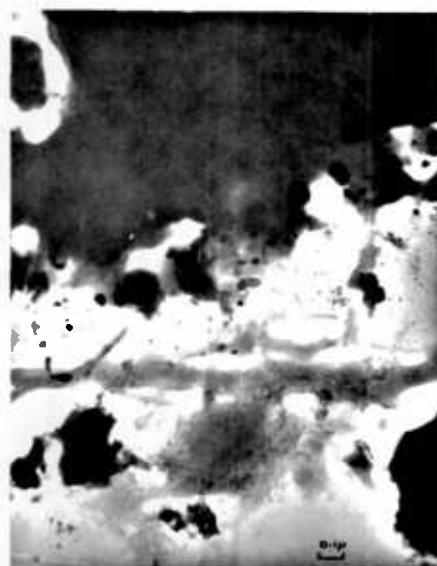


Fig. 64. TEM view of same area as in fig. 63, photographed seconds later, showing effects of radiation damage.

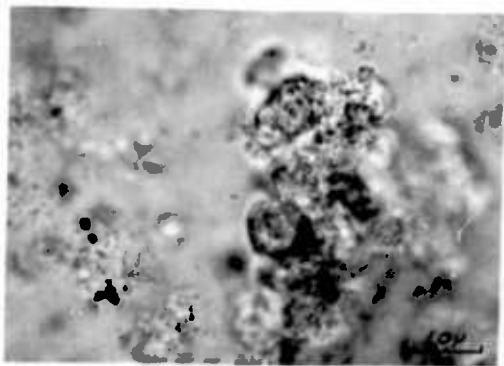


Fig. 65. TLM photograph of same goethitic sediment as shown in figs. 66 and 67. (SM in PPL).

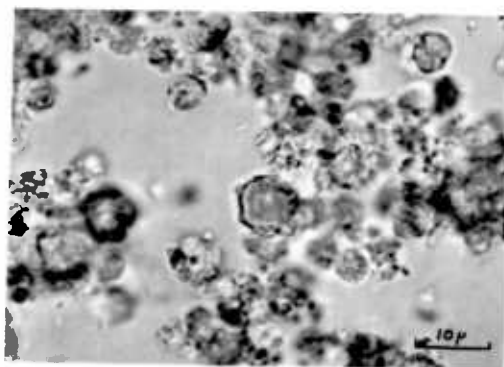


Fig. 68. TLM photograph of same goethitic sediment as shown in figs. 69 and 70. (SM in PPL).

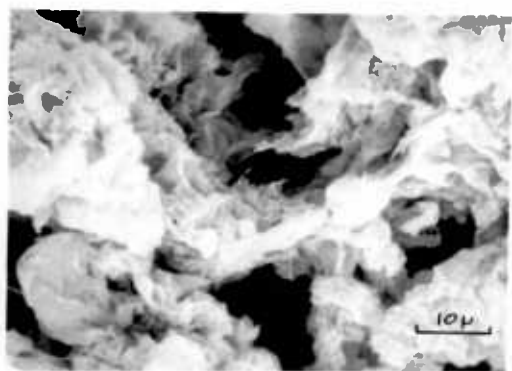


Fig. 66. SEM photograph of same goethitic sediment as shown in figs. 65 and 67.

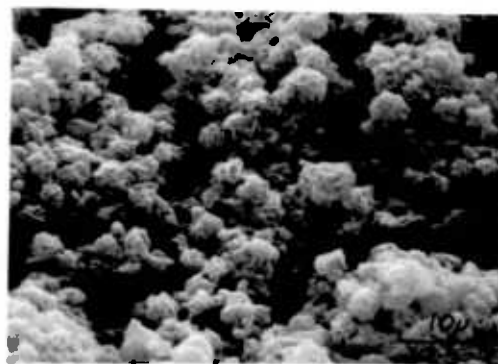


Fig. 69. SEM photograph of same goethitic sediment as shown in figs. 68 and 70. Enlargement shown in fig. 81.



Fig. 67. TEM photograph of same goethitic sediment as shown in figs. 65 and 66. Enlargement shown in fig. 97.

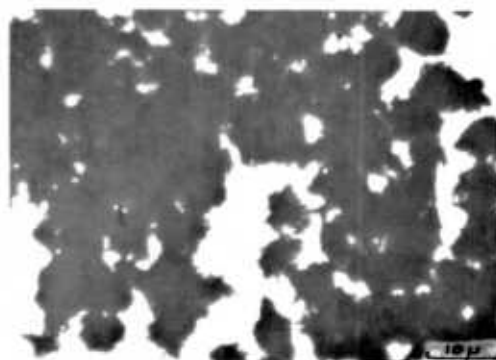


Fig. 70. TEM photograph of same goethitic sediment as shown in figs. 68 and 69. Enlargement shown in fig. 99.

The six photographs on this page are at the same magnification and illustrate three different methods of observation of two different specimens (X 1000).

grains, it is not surprising that Bischoff (1969) obtained refractive indices ranging from 1.55 to 2.0. It is contended that this is not evidence of the extreme range of hydration of the material, although its variable response to radiation damage does support this very reasonable hypothesis. The SEM was found to be particularly useful especially when used in association with an XRA. The specimens were coated in gold about 200 to 500 Å thick. If the coating was too thick detail was lost. If it was too thin then areas of the specimen, especially spiky parts, collected electrical charge and obscured the true nature of the morphology. The TEM was essential for some of the work especially for high resolution. The specimen material was carefully sedimented onto a carbon film. This allowed the use of electron diffraction for further study but as much of the material was relatively coarse, beam penetration was limited. The main problems were with contaminations and radiation damage. Fig. 61 shows grey haloes around filaments. This is possibly a result of contamination of the specimen whilst under examination under the microscope. Such contamination increases with time, beam intensity and imperfections in the vacuum system. The same figures also shows vesiculated material. This very probably is organic matter that has been overheated by the electron beam. Fig. 62 shows holes formed in very thin organic (?) sheets by radiation damage. Figures 63 and 64 show more effects of radiation damage. Generally, hydrated inorganic matter tends to shrink into blobs; organic matter tends to become plastic, to run and finally to vesiculate.

A comparison of the results of the three methods of examination can be made by comparing figures 65, 66 and 67 which are of one sample, and 68, 69 and 70 which are of another sample. All of these photographs are at approximately the same scale. Although there is generally close agreement between the results of the different methods (e.g. figs. 68, 69 and 70) it is clear that any method used alone can give somewhat misleading results, especially when the specimen contains very thin sheets and filaments.

GENERAL DESCRIPTION OF GOETHITIC SEDIMENT

The material from the "Wando River" cruise and typical goethite from the "Chain" cruise was received in a dry condition and is therefore somewhat altered. The goethite occurs as yellow-brown to mid-brown grains of 3 to 10 µ diameter. Some grains occur as aggregates, however most are of a

uniform size (i.e. well sorted). They are isotropic except for samples from deeper parts of the cores (e.g. 0.3 m) and those which are desiccated. The structure of the grains is too fine for good TLM resolution (as is most of the goethitic sediment from the Red Sea deeps) but it commonly appears to consist of spheroidal bundles of filaments and as irregular grains, more granular than flaky. Some grains appear to consist of concentric shells some of which have irregularities with the appearance of loops. Some appear to be ribbed and some have appendages. Under the SEM much of this material is rather nondescript. It occurs as irregular cellular masses, flaky grains, ribbed spheroidal bodies, and filamentous spheroidal bodies. Under the TEM little was observed because the grains are generally opaque. In the thinner areas are fine filaments, crystal laths, spheres $0.2\ \mu$ wide, and short rough filaments $0.1\ \mu$ wide. The goethite is relatively easily damaged by the electron beam. In places there are fine filaments of probable organic matter that also are easily damaged in a manner typical of organic matter (figs. 63 and 64). Some areas consist of tight clusters about $1\ \mu$ wide of laths about $0.03\ \mu$ wide. Other areas consist of very fine webs.

Morphological evidence requires illustration perhaps even more than verbal description and a series of photomicrographs arranged roughly according to magnification is given for this. Figures 71, 72 illustrate the size distribution and overall shape of the material. The chain structure in fig. 71 possibly represents a bacterial mat. Fig. 73 is a close-up of fig. 71 showing that the chains are relatively regular bodies. Figures 74 to 83 show typical grains with concentric shell, eccentric shell, platy, filamentous, webbed and crystalline structures. Fig. 79 illustrates a filamentous spheroidal body and figs. 80, 81 and 82 illustrate spheroidal bodies in which goethite crystallites appear to be present.

Figures 84 and 85 are of calcareous fossils. That in fig. 84 has been identified as probably Thoracosphaera deflandrei by Dr. S.A. Ali of the Geology Department, Imperial College. It is rare in this sediment but shows similarities with the goethitic grains. Figure 85 is of a coccolith more typical to this sediment. Coccoliths are the main calcareous fossils found during this examination of the goethitic sediment. In the SEM the coccoliths generally appear lighter than the goethitic material. Their



Fig. 71. TLM photograph of goethitic sediment with chain structure and dispersed grains. (SM in PPL.)

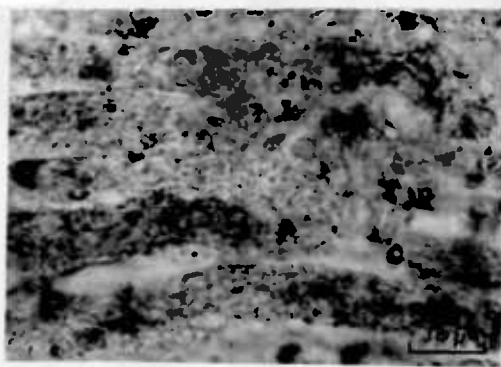


Fig. 72. TLM photograph of goethitic sediment with chain structure. Enlargement of area in fig. 71. (SM in PPL.)

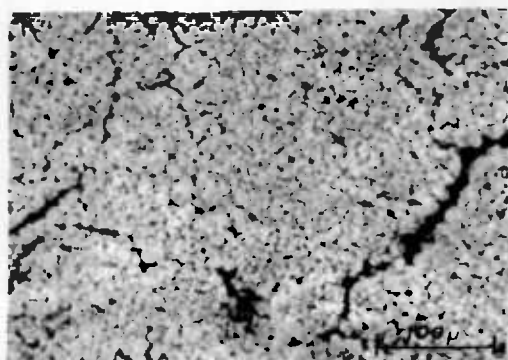


Fig. 73. SEM photograph showing well sorted goethite sediment.



Fig. 74. TLM photograph showing grain with concentric shell structure. Same sediment as in fig. 73. (SM in PPL.)



Fig. 75. TLM photograph showing grain with eccentric shell structure. Same sediment as in fig. 73. (SM in PPL.)

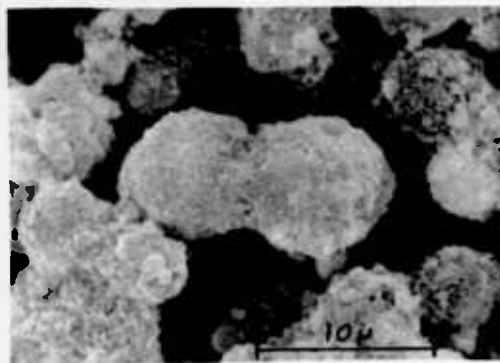


Fig. 76. SEM photograph showing peanut shaped double grain from same sediment as in fig. 73.

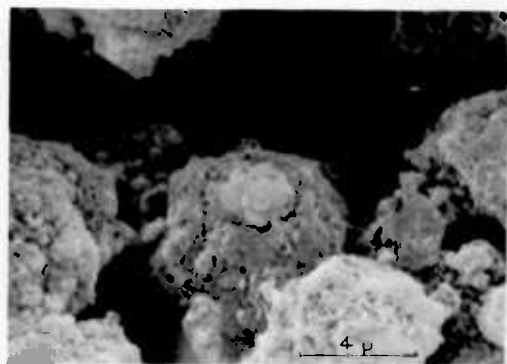


Fig. 77. SEM photograph showing plated spheroid from same sediment as in fig. 73.

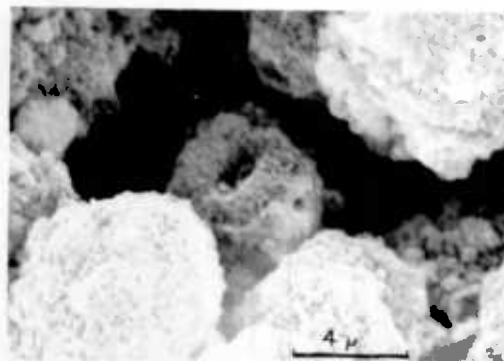


Fig. 78. SEM photograph showing spheroid with structured aperture from same sediment as in fig. 73.



Fig. 79. SEM photograph of filamentous spheroid.

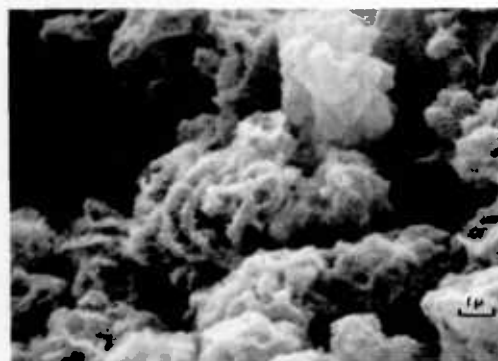


Fig. 80. SEM photograph of goethitic grain with unusual structure, from same sediment as in fig. 69.

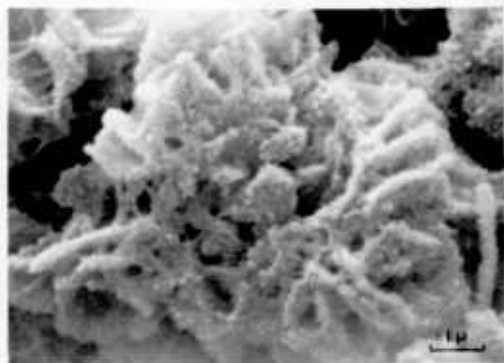


Fig. 81. Enlargement of central part of fig. 69 showing another goethitic grain with unusual structure.

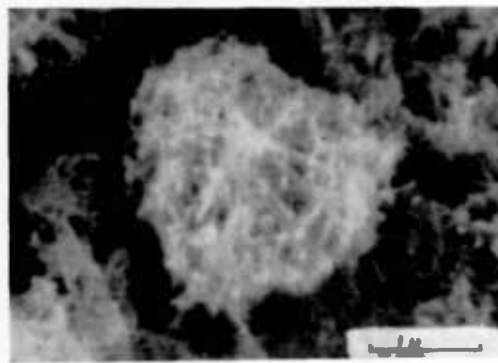


Fig. 82. SEM photograph of spheroidal grain composed of goethite crystallites.

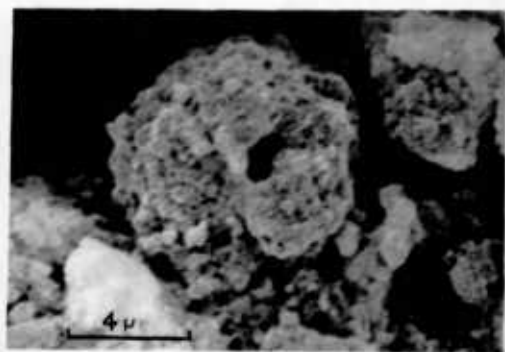


Fig. 83. SEM photograph showing goethitic spheroid with aperture somewhat resembling coccolith. From same sediment as in fig. 73.

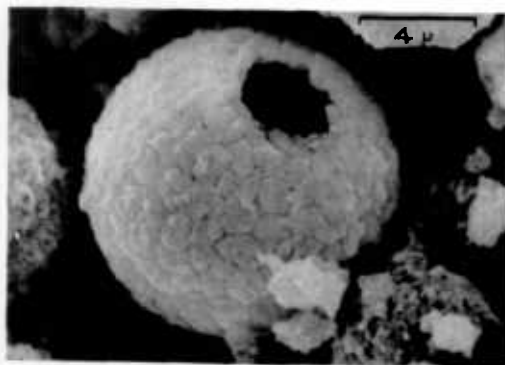


Fig. 84. SEM photograph showing calcareous *Thoracosphera deflandrei* in goethitic sediment. From same sediment as in fig. 73.

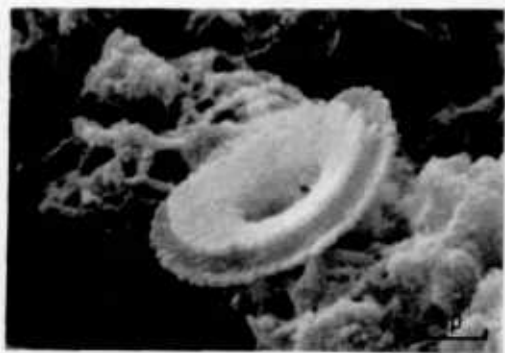


Fig. 85. SEM photograph showing typical coccolith, from same sediment as in fig. 69.

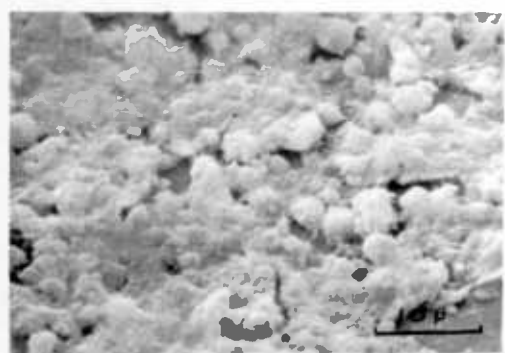


Fig. 86. SEM photograph showing zoogloal structures, from same sediment as in fig. 71.

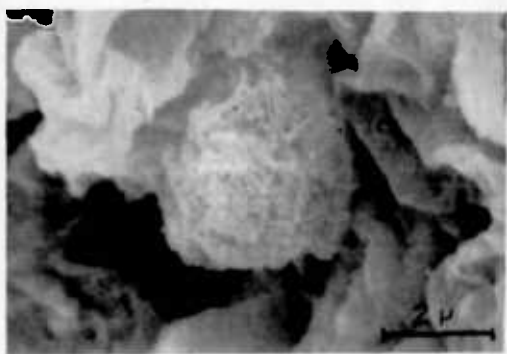


Fig. 87. SEM photograph showing structured spheroid attached to crumpled sheet from same sediment as in fig. 66.



Fig. 88. SEM photograph showing filaments, one having a thickened extremity, from same sediment as in fig. 71.

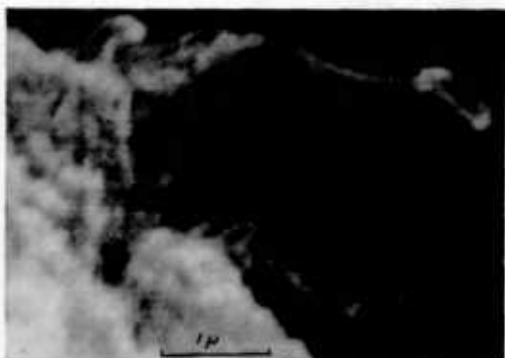


Fig. 89. SEM photograph of same specimen as in fig. 88, showing looped filaments.



Fig. 90. SEM photograph of same specimen as in fig. 88, showing curved filaments extending from webs.

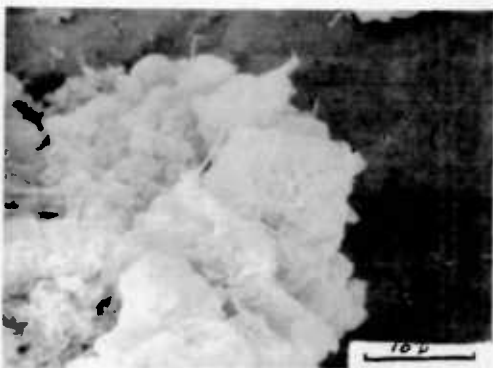


Fig. 91. SEM photograph of same specimen as in fig. 88 showing reticular webs or plates, and a bifurcated filament.

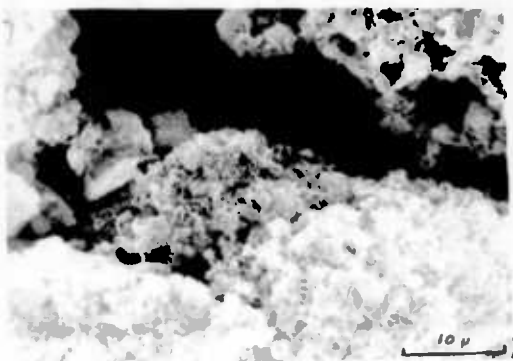


Fig. 92. SEM photograph of typical Red Sea goethite sediment showing boxwork-like reticular webs or plates and some spheroids.

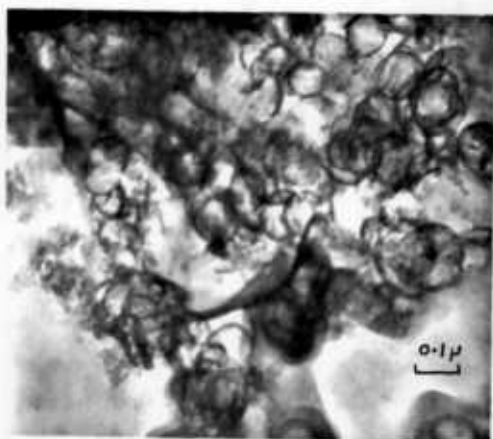


Fig. 93. TEM photograph showing cellular structure from same sediment as in fig. 71.

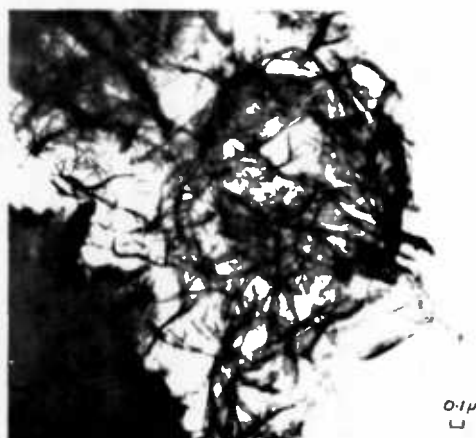


Fig. 94. TEM photograph showing irregular webbed filaments, from same sediment as in fig. 71.



Fig. 95. TEM photograph showing bifurcated rough filaments from same sediment as in fig. 71.



Fig. 96. TEM photograph showing rare tiny spore-like body and thin segmented filaments, from same sediment as in fig. 82.

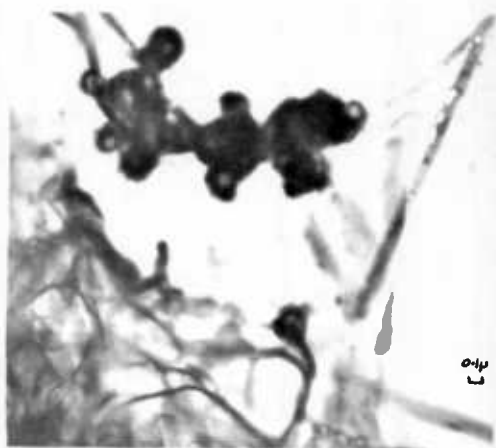


Fig. 97. Enlargement of lower right corner of fig. 67 showing segmented straight parallel and curved tapering filaments as well as mass of spheroids and spheroid attached to filament.

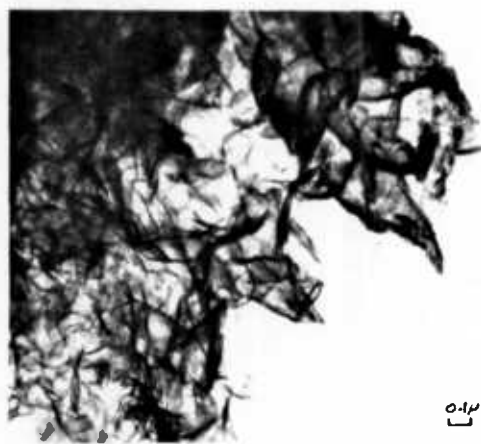


Fig. 98. TEM photograph showing webbs attached to filaments from same sediment as in fig. 92.



Fig. 99. Enlargement of lower central part of fig. 70 showing typical webbed grains.



Fig. 100. TEM photograph of same sediment as in fig. 70 showing structure.



Fig. 101. TEM photograph of same sediment as in fig. 82, showing goethite crystallites of two size ranges and irregular bifurcated filament.

calcareous nature also of course can be confirmed by XRA. Some of the goethitic grains, e.g. fig. 83 are easily mistaken for coccoliths. Figs. 86 and 87 show the relationships of the spherical grains to filaments and to the sheeted material. Figures 88, 89, 90 and 91 show filaments with curved, budded(?), looped and forked forms. Figures 91 and 92 show platy material in a boxwork-like structure. More spheres can also be seen (with some difficulty) in fig. 92 amongst the platy or webbed material and (with ease) in fig. 93. Figures 94 and 97 also illustrate filaments of various types (curved, tapering, cylindrical, rough, bifurcated and segmented). Fig. 97 also shows tiny spheres connected to each other and one connected to a filament. Fig. 98 shows thin crumpled sheet material. Fig. 100 shows a TEM photograph of a typical webbed grain and fig. 99 shows an unusual webbed grain. Fig. 101 is a good example of crystalline material.

MORPHOLOGICAL TYPES OF GOETHITIC SEDIMENT

Samples from the "Nereus" cruise were taken for study directly from the damp core and were therefore more reliable than other samples available. These form a group which the writer has divided into four types. These types grade from material with distinctly biogenic forms through intermediate types to material of distinctly inorganic form and it is later postulated that this essentially represents gradation in early diagenesis.

The first type occurs at and near the upper surface of the sediment. It consists of gelatinous aggregates of grains that commonly occur in chains and as dispersed grains. The chains are easily deformed and it is suspected that they were often broken and deformed during the preparation of the smear mounts. These delicate structures would not survive the treatment given to the cores from the "Wando River" cruise. Individual grains appear to be about 2μ diameter. The chains are about 6μ wide and were observed as parallel lines (figs. 71 and 72) and as branching lines. In one smear mount was a definite circular cluster of grains 200μ diameter. These grains are particularly difficult to observe by TLM (fig. 65) because of their ultra-fine structure. Some could be observed however as concentric shelled spheroids and some as stellate spheroids. These sediments range in colour from green (rare) to yellow to reddish brown. Under the SEM this material appears to consist of crumpled sheets, webs (or reticular plates, fig. 91), spheroidal bodies, filaments and transitional varieties (figs. 66, 86, 88, 89 and 91). The spheroidal

bodies range from 3μ to 6μ diameter and appear to consist of filaments, webs and tiny spheres but the structure is difficult to observe. The general structure commonly appears to be zoogloal. Filaments are relatively common and have a variety of forms ranging from simple straight filaments to forked, curly, tangled and connecting structures. Some form loops and some appear to have thickened extremities (bud-forming?) e.g. figs. 66, 86, 88, 89 and 91. Contamination and radiation damage complicate observations by TEM (figs. 61 and 62) however this also points to an organic origin. The thermoplastic material most probably is organic matter associated with the goethite although some may be non-crystalline hydrated iron oxide or hydroxide. Crystalline goethite (this term is used loosely here and is discussed further on) shrinks into more rounded masses but has not been observed by the writer to form bubbles. Much of the material appears as tangles of filaments and irregular sheets. Sheeted structures are much less apparent under the TEM than the SEM probably because they are very thin (cf. figs. 66 and 67). The filaments range in size from about 0.01μ to 0.1μ diameter and range in shape from straight parallel sided rods (or tubes) to curved tapering bodies associated with sheets (e.g. figs. 67, 93, 94, 95 and 97). Some rods or tubes appear to have segments (fig. 97) and fine bubbles were found around some of these suggesting they had organic sheaths. Others have a rough surface (fig. 95) perhaps due to crystal nucleation. These commonly bifurcate but the nature of the branching is uncertain. Regular spheres are not uncommon (e.g. fig. 93) but it is difficult to be sure that some of these structures are not artifacts of organic matter. Definite spheroidal bodies 0.5 to 1.0μ diameter isolated or connected to filaments or to other spheres were also readily observed (e.g. figs. 67 and 97). Lath-shaped goethite(?) crystals are rare.

The second type of goethite in the "Nereus" 1970 samples is the most common. It consists of very well sorted (figs. 73, 68, 69) yellow to dark brown isolated spheroidal bodies, having mottled surfaces and commonly a concentric shell structure (e.g. fig. 74). It occurs as dispersed grains 5 to 10μ diameter and as irregular branching chains. Some grain clusters 50 to 100μ diameter are possible globular clusters but are more likely artifacts caused by physical adherence of grains. Some grains have irregular shapes, some have eccentric shells (fig. 75) and a few are dumbbell-shaped with spherical shells overlain by dumbbell - or peanut-shaped

shells (e.g. fig. 76). They also appear to consist of flaky granules and boxwork-like webbed rough filaments (figs. 77, 78, 80, 81 and 83). Commonly the grains contain apertures (perhaps 5 to 10% of the population) and occasionally resemble degraded coccoliths (c.f. figs. 83 and 84). The grain structures, range from concentric to reticulate (cf. figs. 77 and 81). Under the TEM the reticulate types appear as balls of webbed filaments and as meshworks of filaments or crystals (e.g. fig. 100). Occasional grains have lobate structures (e.g. fig. 99 and possibly fig. 80) which essentially consist of fine radiating crystals outlining possible pre-existing biogenic structures.

The third type of "Nereus" 1970 goethite resembles the second but shows more surface structure under the TLM. It occurs as dispersed grains 2 to 10μ diameter and as short chains and clusters. The grains are generally spheroidal and under the TLM appear to be webbed or filamentous. Under the SEM and TEM the structure can be seen to be due to crystal lath clusters (e.g. figs. 82, 101). It can be seen in fig. 101 that the size distribution is bimodal and that associated with these crystal laths is an irregular bifurcating filament (nature of the branching is unknown). Thin straight filaments (0.03μ) wide with possible segments and irregular filaments are relatively uncommon in this material. Spore-like bodies were also observed under high power, however they are rare and therefore must be suspected as some kind of contamination. In one area vesiculating material moved along thin filaments while they were being observed.

The fourth type of "Nereus" 1970 goethite was found in one sample only (locality $25^{\circ}23'N$, $36^{\circ}12'E$). It consists of yellow granules too fine to be resolved with the TLM. Under the SEM and TEM however it was observed to consist of spherical crystal clusters averaging approximately 1μ diameter. This shows no obvious biogenic features, although it contains a much higher proportion of coccoliths, pteropods and forams than the other types. It is likely that the four types of "Nereus" 1970 goethite represent stages in the diagenesis of the sediment, with type 1 being the pristine form. Type 1 occurs much closer to the surface than type 3 in a core containing the two. Most probably the iron oxide material would become less hydrated and more crystalline with time and diagenesis. Indeed Bischoff (1969) found that some material south of Atlantis II deep

which is more crystalline is also much older (by radiometric dating) than the material in Atlantis II deep. XRD data also supports this. Type 1 is amorphous, type 2 is very poorly crystalline and type 3 is sufficiently well crystallized to give an unmistakable goethite XRD pattern. Organic matter is also very sensitive to diagenesis and its disappearance through autolysis would probably assist with the crystallization of the goethite. During this process the zoogloal structure would trend toward the crystal mesh structure probably beginning with the loss of thin sheeted material followed by the loss of filamentous material.

SECTION 2

G E O M I C R O B I O L O G Y
O F T H E R E D S E A D E E P S

Contd.

2.2.2

MACERATION RESIDUES OF GEOTHERMAL SEDIMENTS

Specimens from the goethite-amorphous, sulphide and iron-montmorillonite facies were macerated and examined. Generally it was found that the goethite-amorphous facies contained much more insoluble organic matter than the iron-montmorillonite facies and a little more than the sulphide facies. The organic matter appears to consist of disintegrated very pale brown thin sheets of mucilage commonly spotted (with cocci?). Tiny spore-like bodies and larger grass(?) spore bodies were also noted, the latter most probably being wind-blown from the land. Saxby (1972) studied the chemistry of macerated Red Sea cores and found the insoluble organic matter contents, on a dry salt free basis, to range up to 1%. He found an increase with depth in the C/H ratio in a core from Atlantis II Deep and attributed this to high temperatures and a higher rate of diagenesis in the sediment below this deep compared with other deeps. Comparing Atlantis II Deep with Discovery Deep he also found that in Atlantis II Deep the kerogen to pyrite ratio is low and the kerogen to carbonate ratio is high. This is supposed to support the hypothesis of Truper (1969), and Watson and Waterbury (1969) that the Atlantis II Deep brine area is sterile. It must be pointed out however that Saxby considered only the insoluble organic matter and only sulphate-reducing bacteria as a means of removing kerogen. He further assumed that there is a constant kerogen to carbonate ratio before microbial attack (this ignores sedimentation rates and transported land material). Organic matter precipitated or trapped with the goethitic sediment (c.g. Watson and Waterbury, 1969) would influence the kerogen to carbonate ratio, especially as the ratio of precipitation of goethitic sediment is much higher than that of detrital sediment.

2.2.3 COMPARISON OF NATURAL AND ARTIFICIAL GOETHITE

One of the main problems with using morphology to establish whether or not the goethite-amorphous facies is biogenic is that non-biogenic goethite may have a complex morphology. To alleviate this, artificial goethite was prepared inorganically under circumstances comparable to those in the Red Sea deeps. Ideally, hot sterilized brine and sterilized sea water would have been used. These were not available however, so Dr. J. Angus (Mining Geology Division, Imperial College, 1972) kindly suggested the following method and supervised the experiment. An aqueous solution of ferrous sulphate containing 600 ppm Fe was heated at 50°C in contact with air for two days in a reflux condenser. In addition, a separate solution of ferric chloride, having 600 ppm Fe, was adjusted to a pH of 4 by addition of sodium hydroxide and was heated at 50°C for two days in a reflux condenser. The ferrous sulphate gave a pale yellow-brown precipitate and the ferric chloride gave an orange-brown precipitate.

Under the TLM both precipitates appeared to consist of fine granules. Under the SEM both appear to consist of well sorted spheres about 1μ diameter with textured surfaces (fig. 102). The TEM showed (fig. 103) that the spheres consist of bundles of crystal laths about 0.03μ wide rather similar to those of the natural material in fig. 101. It also showed that the grains tend to occur in irregular chains. In detail the crystal laths closely resemble the obvious laths in the natural goethitic sediment. At high power they appear to consist of parallel fibres and are similar to the porous β - FeOOH (akaganeite) prepared by Gallagher (1970). This was prepared by slow hydrolysis of a hot aqueous solution of ferric chloride but even so the material prepared by Dr. Angus and the writer is goethite. Both precipitates have XRD patterns corresponding to goethite and akaganeite peaks are missing from the XRD patterns. The XRD patterns of the two precipitates are very close to each other but that derived from FeSO_4 is slightly weaker. The electron diffraction patterns of both however are virtually identical. The only obvious difference between the two precipitates is colour. It is well known that red hydroxide is derived

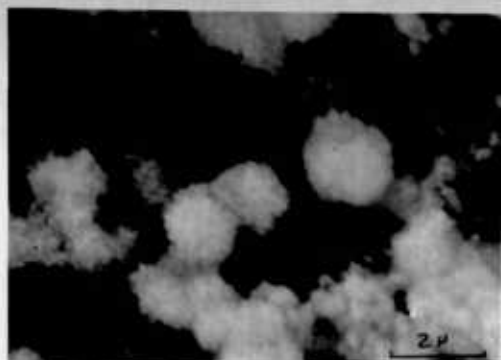


Fig. 102. SEM photograph of goethite produced artificially by oxidation of Fe Cl_3 .



Fig. 103. TEM photograph of same artificial goethite as in fig. 102.

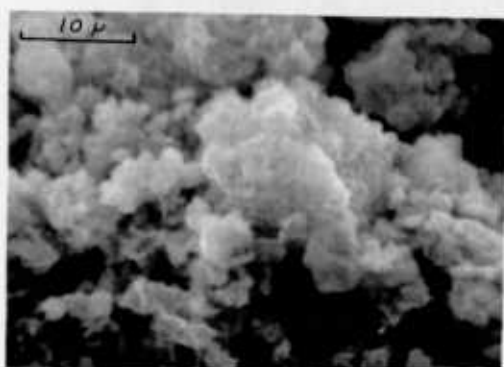


Fig. 104. SEM photograph of poorly crystallized mackinawite produced artificially by culturing sulphate reducing bacterial in same sediment as shown in figs. 71, 86 and 93.

from ferric salt and yellow from ferrous however the present study shows that the difference in colour is not easily attributed to sub-microscopic physical differences as previous writers have advocated (see Harder, 1919 for full discussion).

In appearance and in XRD and ED results, the artificial goethite closely resembles goethite of "Nereus" 1970 type 3. One difference is that a vague peak which appears to correspond to organic matter occurs on the ED pattern of the natural goethite. This organic matter may or may not be indicative of a biogenic origin of the natural goethite. It could simply have been trapped by the goethite as it precipitated. In appearance the artificial goethite also closely resembles "Nereus" 1970 type 4 goethite but is unlike types 1 and 2. XRD patterns of the latter are very weak and type 1 appears to be amorphous. Type 2 contains sufficient weak peaks to identify it as goethite (the $5\text{\AA}$ line tends to exclude hydrogoethite).

In general all the goethite is very poorly crystalline and XRA in the SEM was used to check the composition of the material. Generally the natural samples are homogeneous and are more or less alike. The Mn/Fe ratio however has a wide range from about 1% to 50%. The most manganiferous sample assayed about Fe 30%, Mn 15%, Zn 1.2%, Cu 400 ppm (Applied Geochemistry Research Group of Imperial College general results). No separate phases were found for the Fe and Mn under the SEM hence it is thought that both elements belong to the same phase. The "goethite" therefore is probably intermediate between goethite and groutite. The homogeneity of the "goethite" extends even to filaments however it is possible that SiO_2 occurs as a separate phase and was overlooked at the time of study. Ca occurs in coccoliths and K probably occurs in micaceous minerals occasionally observed.

2.2.4

BACTERIAL CULTURE

Dr. J. Tooms collected a core from Atlantis II Deep and found it to be quite fetid on site at the time of collection. Contamination was minimized by the manner of collection and by sealing the sample in the collection tube but precautions as rigorous as those used by Watson and Waterbury (1969) were not taken as the collection of biological material had not been anticipated.

In the hope of obtaining indubitable evidence of the occurrence of iron bacteria in the Red Sea deeps the writer tried to find means of culturing iron bacteria. Extensive enquiries were made in and around London but it was found that the culturing of such bacteria is difficult and requires very specialized knowledge. The main utilitarian significance of iron bacteria is that they are responsible for the build up of iron deposits in water pipes and it is not really surprising that they are relatively unknown. The only person found with any notion on how to culture iron bacteria was Mr. T.G. Tomlinson of the Water Pollution Research Laboratory, Stevenage. Mr. Tomlinson very kindly prepared Wolfe's medium and Leathen, McIntyre and Bradley's medium, and inoculated these media with tiny portions of the fetid Red Sea ooze. No microbial growth developed (letter dated 1 October, 1971, reference WPRS 44/29/1). These negative results are not particularly significant in view of the difficulties of culturing iron bacteria and the state of the material. By the time it had reached Mr. Tomlinson the material had been some weeks out of its normal environment of 56°C and 400 atmospheres pressure. Leathen (1970) pointed out that the relatively common iron bacteria of the Crenotrichaceae Family have never been grown in a pure culture or artificial medium.

Before meeting Mr. Tomlinson the writer visited Mr. D. Wakely and Dr. N. LeRoux at Warren Springs Research Laboratories, Stevenage, who though unable to assist with the culturing of iron bacteria suggested the writer should attempt culturing sulphate-reducing bacteria. These bacteria are hardy and easily cultured. Mr. Wakely and Dr. LeRoux kindly explained

the procedure and gave the writer 20cc culture bottles and written instructions. These instructions were followed exactly and every care was taken to avoid contamination. The culture medium is essentially Barr's medium (described by Truper 1969). An attempt was also made at using the method of Rickard (1969) but Dr. J. Angus and the writer were unable to adjust the Eh of the medium to the required value so Rickard's method was abandoned. In addition to the medium of Wakely and LeRoux a medium of the same composition was prepared and to this was added extra sea-salt so as to be comparable to the Red Sea brines. Six specimens of 0.1g were taken from the top of the sediment core and the remaining specimens of 1.0g, 0.1g and 0.01g weights were taken from a depth of a few centimetres in the core. Six 0.1g specimens in addition were taken and sterilized to serve as controls. The specimens, 30 altogether, were inoculated into the media and incubated at 30°C, 44°C and 60°C (10 specimens at each temperature). One specimen gave a positive reaction. This was of 1.0g from a few centimetres depth inoculated into the ordinary medium of Wakely and LeRoux and incubated at 44°C. The goethite in the bottle was converted into a black flaky substance which was shown by XRD to contain very poorly crystalline mackinawite (fig. 104).

Sulphate reducing bacteria were previously known to operate in Discovery Deep sediments and other sediments but not in those of Atlantis II Deep (Truper, 1969, Watson and Waterbury, 1969). Sulphur isotope evidence also suggests that the sulphides in the sampled sediments of Atlantis II Deep were dominantly precipitated by inorganic means (Kaplan et al 1969). This suggests the writer's results are anomalous. Laboratory contamination is a most remote possibility however there is a real possibility that the sediment has been contaminated with sulphate reducing bacteria known to occur in the brine-sea water transition zone (Watson and Waterbury, 1969). As the inoculated material came from well within a core sample it is unlikely that contamination occurred during sampling. It seems more likely that it is due to natural contamination that could have occurred when bacteria were carried down into the hot brine along with goethitic sediment. Truper (1969) and Watson and Waterbury (1969) pointed out that the environmental parameters of Atlantis II Deep independantly are not particularly hostile to sulphate-reducing bacteria but the overall effect of them taken as an integral group may inhibit bacterial growth. Obviously the bacteria were not prolific in the samples studied by the writer as it is

dominantly amorphous goethitic material. Nevertheless the XRA did detect some sulphur in the sample (perhaps around 1% - the analysis was qualitative).

2.2.5 BIOGENIC STATUS OF RED SEA GEOTHERMAL SEDIMENT

Despite prolonged morphological study the biogenic status of the Red Sea goethitic sediment is still slightly doubtful. Some material has a distinctly biogenic character but other material has distinctly inorganic morphology. It is possible that the sediment was precipitated by both types of processes. It is also probable that biogenic characteristics of goethitic sediment were quickly lost during the early stages of diagenesis.

The detrital sediment is largely biogenic and iron-montmorillonite is unlikely to be anything but inorganic. Sulphide sediment however is still enigmatic. Some sulphate-reducing bacteria were recovered from the deeps though it is not absolutely certain that they lived for long in the goethitic sediment in which they were found. The occurrence or absence of sulphate reducing bacteria in the present brine pools does not bear directly on the precipitation of the sulphide sediment as conditions were probably different when that sediment was deposited. The sulphur isotope data so far available do not offer a unique solution to the problem (which is discussed further in section 3.1.1).

PART II GEOMICROBIOLOGY

Section 1	Introduction
Section 2	Red Sea Deep
Section 3	Coastal Plain
Section 4	Shield
Section 5	Phytoclast Geothermometry

SECTION 3

GEOMICROBIOLOGY
IN THE COASTAL PLAIN

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SECTION 3

G E O M I C R O B I O L O G Y
I N T H E C O A S T A L P L A I N

2.3.1

WADI FATIMA IRON ORE

INTRODUCTION

The ore at Wadi Fatima is typical Minette type iron ore occurring dominantly as goethitic "oolites" and "pisolites" as described in section 1.2.4. Carbonaceous matter is locally abundant in the matrix and some silicified wood has been found in the ore. Pyritic carbonaceous shale occurs in the ore formation and bituminous matter veins limestone and basalt overlying this formation (Al Shanti, 1966).

The ore has been described as oolitic however the writer considers it to be onkolitic or microonkolitic. The American Geological Institute (AGI, 1960) defined oolites, inter-alia, as spherical to ellipsoidal bodies which have radial or concentric structures or both. A strictly descriptive definition of onkolites that enables clear distinction from oolites was not found in the literature and it appears that many oolitic rocks are probably onkolitic (e.g. Jenkyns, 1972). Onkolites are discoidal to ellipsoidal to spherical concentrically layered bodies which are a variety of stromatolite (Weiss 1969, Gatrall, Jenkyns and Parsons 1972, Jenkyns 1972). The layers in onkolites, unlike oolites, commonly have unconformities and irregularities in curvature, Al Shanti (1966) illustrated a beautiful example with at least three unconformities. Oolites are commonly thought to have a biogenic origin (e.g. Shearman et al 1970) but onkolites necessarily have a biogenic origin in common with other types of stromatolites.

OBSERVATIONS

Due to the opacity of the ore and its soft friable nature it is somewhat difficult to obtain good thin sections for studying the "oolites". The concentric layering however invariably is coarse, even if ill-preserved, and is commonly irregular. The onkolites are commonly deformed (or squashed) by one another when in contact. This could be expected with onkolites as they contain organic matter giving them a somewhat rubbery

consistency. Zeidler (1970) reported that average grade Wadi Fatima ore contains 44.3% Fe and 17% "volatile matter". Irregular coarse concentric layering and a deformed onkolite together with a ferruginous sandstone matrix are illustrated in fig. 105. Fig. 106 illustrates a spiral type of structure associated with a penecontemporaneous fracture surrounded by non-fractured but irregular concentric layering. Minor unconformities occur in the above onkolites but they are not easily observed in the photographs. Nuclei to the onkolites commonly appear to be indistinct probably because they are of similar material to the remaining parts of the onkolite. This impression however may be erroneous because few growth-centres lie in the planes of the thin sections. Some nuclei are quartz grains others are shells (e.g. gastropods) or organic matter. The example illustrated in fig. 107 appears to be bone or possibly woody material. Algal matter is rarely well preserved in the matrix to the onkolites. An exception is the well preserved material shown in fig. 108.

Abundant organic matter is obtained when the iron ore is macerated. This was identified by Dr. K. Alvin and Dr. M. Muir as mostly degraded algal mucilage and leaf cuticle (and possibly tissue) from monocotyledonous and dicotyledonous flora. Traces of probable animal matter were also found. Fig. 109 illustrate non-descript algal matter together with a piece of cellular tissue. Fig. 110 shows monocotyledonous leaf tissue with distinct stoma and containing a fungal hypha (magnified in fig. 111). Fig. 112 shows viral or bacterial erosion of leaf tissue.

To ascertain that organic matter forms an integral part of the onkolites an electron microprobe study was made. Most electron microprobes cannot detect carbon and other elements of such low atomic number however this is possible with the Cambridge Microscan Mark 5 at the Royal Army Research and Development Establishment, Fort Halstead. Through the kindness of Messrs. H. Priestly, J. Winn and P. Suddaby a polished section of the iron ore was studied with this instrument operated by J. Winn. Care was taken to avoid carbon contamination from diamond dust, silicon carbide, and oil. The writer painstakingly ground and polished the specimens using only alumina powders and deionized water. The surfaces were finally cleaned ultrasonically and the material was examined uncoated. The results clearly showed that the carbon distribution closely follows the concentric layering. This presumably is organic carbon as its distribution pattern is both

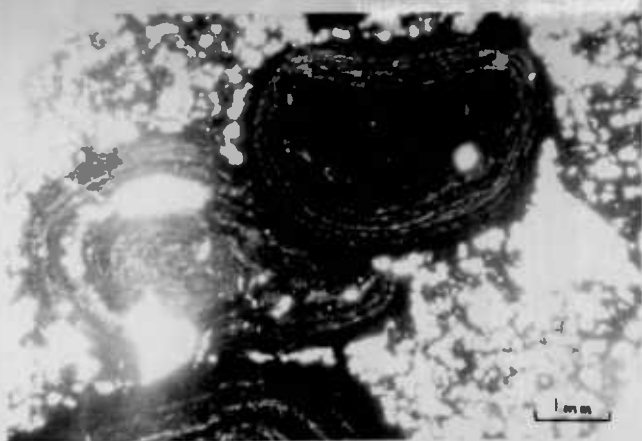


Fig. 105. TLM photograph of Wadi Fatima iron ore showing layering in onkolites, deformation at contact and matrix (TS in PPL).

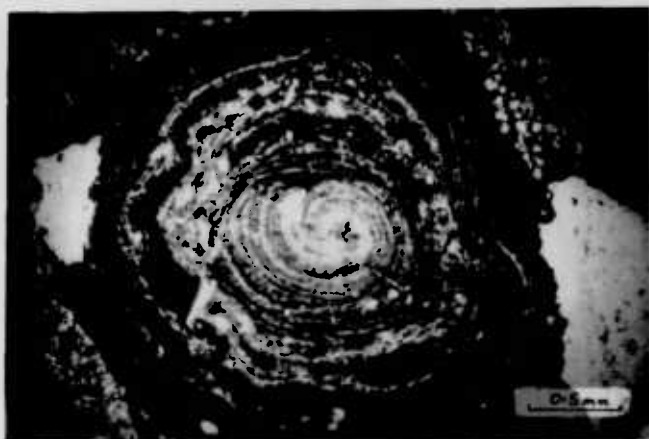


Fig. 106. TLM photograph of Wadi Fatima iron ore showing irregular and spiral type of layering associated with central fracture (TS in PPL).

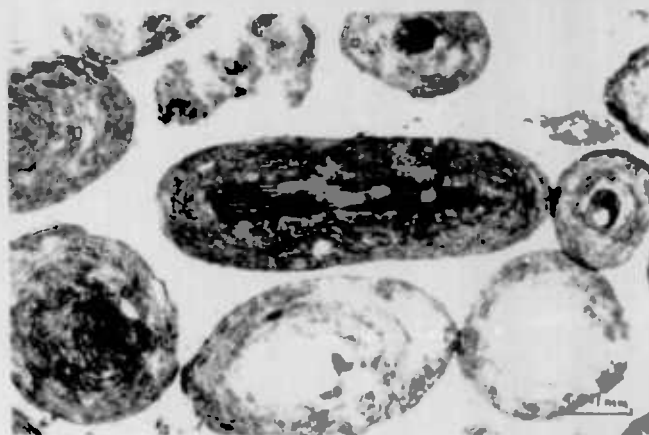


Fig. 107. TLM photograph of Wadi Fatima iron ore showing elongated onkolite with bone or wood nucleus (TS in PPL).



Fig. 108. TLM photograph of Wadi Fatima iron ore showing probable algal matter in matrix (TS in PPL).



Fig. 109. TLM photograph of maceration residue of Wadi Fatima iron ore showing non-descript algal matter together with cellular tissue or cuticle (SM in PPL).



Fig. 110. TLM photograph of maceration residue of Wadi Fatima iron ore showing typical well preserved cuticle (SM in PPL).

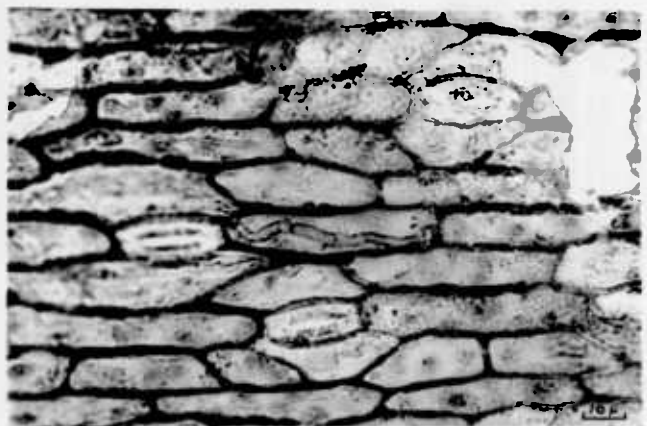


Fig. 111. Enlargement of part of fig. 110 showing details of cuticle structure including stoma and a fungal hypha.

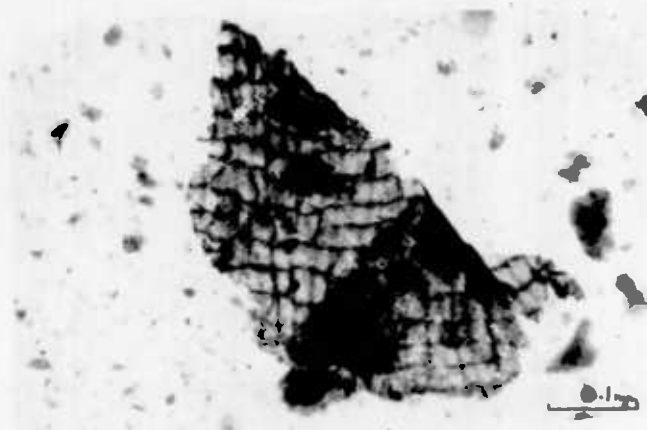


Fig. 112. TLM photograph of maceration residue of Wadi Fatima iron ore showing leaf cuticle with bacterial, viral or possibly fungal erosion (SM in PPL).

stronger and different to that of calcium and of background. The illustrated results have been presented by Muir, Hamilton, Grant and Spicer (1974).

Although the iron ore may be termed oolitic in a descriptive sense, its structural details and intimate association with biogenic matter shows that it is more than likely onkolitic.

SECTION 3 G E O M I C R O B I O L O G Y
 I N T H E C O A S T A L P L A I N

Contd.

2.3.2

IRON OXIDE BIOMORPHS IN JASPER VEINS

INTRODUCTION

The grouping of the barite-jasper veins in the coastal region is discussed in section 3.1.4 but it must be noted here that these fracture-filled transgressive veins form a distinct group even though there is a wide range in the proportions of mineral constituents. Some veins consist almost entirely of barite, others consist mostly of jasper. The minor constituents of manganese oxides, calcite, and galena also vary. The barite contains no recognisably biomorphic structures and its deposition generally preceded that of the jasper.

The term "jasper" is used here in its usual sense (e.g. Dana, 1932) and is preferred to the adjective "jasperoid" of Lovering (1972). Although Brobst (1966) described the jasper at Umm Gerad as jasperoid it should be noted that this conflicts with Lovering's definition which requires it to have a replacement origin. The coastal vein jaspers consist dominantly of quartz with lesser chalcedony and opal, and are coloured yellow, red, brown and black by disseminated iron and manganese oxides. These oxides are commonly biomorphic. They pre-date the crystallization of the quartz but may have formed contemporaneously with or after the introduction of the silica.

OBSERVATIONS

Umm Gerad

The Umm Gerad jasper is typical of the coastal jaspers except that it is slightly finer grained and has a preponderance of spherical over filamentous biomorphs. The general texture is shown in figs. 113, 114 and 115. Quartz boundaries can be seen by comparing 114 and 115. The grain size is about 30 μ and the fabric is granular to interlocking. Brobst (1966) noted that tiny particles of iron oxide and chalcedonic silica, that probably precipitated from a colloidal gel, are interstitial to isolated grains and groups of interlocking quartz grains. He further noted that T.G. Lovering "regards these features unusual in jasperoid because the

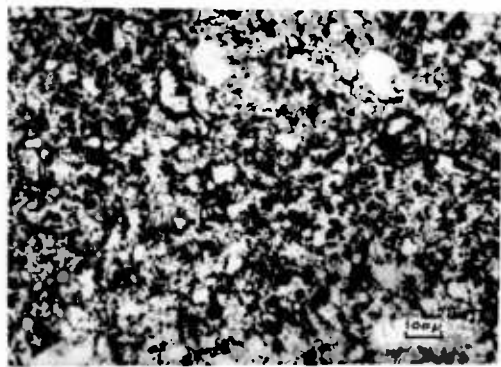


Fig. 113. TLM photograph of Umm Gerad jasper showing general structure. (TS in PPL).

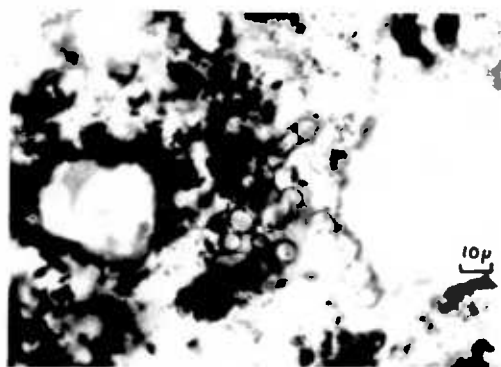


Fig. 116. TLM photograph of Umm Gerad jasper showing colonial mass of spheres. (TS in PPL).

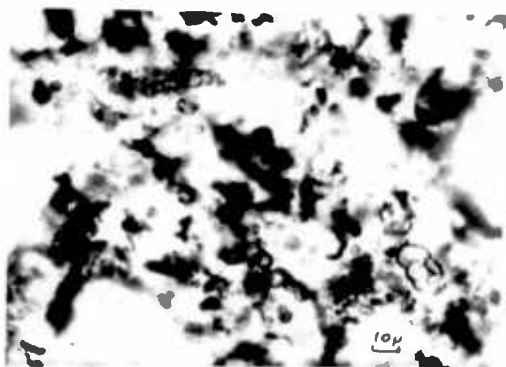


Fig. 114. TLM photograph of Umm Gerad jasper showing typical interpenetrating spheres. (TS in PPL.)

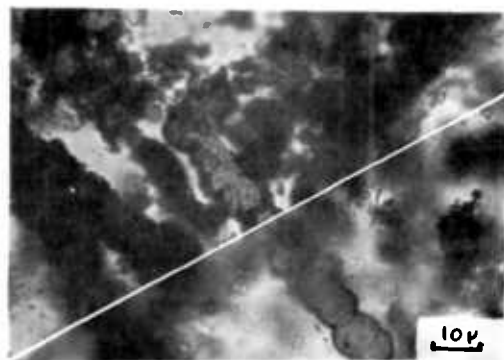


Fig. 117. TLM photograph of Umm Gerad jasper showing segmented tube. Composite picture. (TS in PPL).

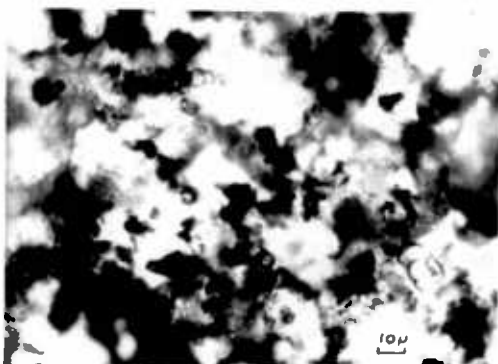


Fig. 115. TLM photograph of area shown in fig. 114 showing quartz grain boundaries. (TS in CP).

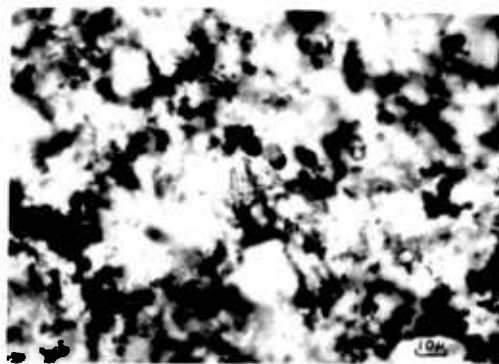


Fig. 118. TLM photomicrograph of Umm Gerad jasper showing straight filaments associated with tiny spheres. (TS in PPL.).

grains of quartz do not appear to have been formed from a gel, but rather by direct crystallization from solution, yet they contain disseminated inclusions of hydrous iron oxide that seem to have been precipitated in colloidal form prior to the crystallization of the quartz".

Typically the goethitic biomorphic forms in this jasper consist of spheres 2μ to 10μ diameter occurring either alone, in chains or in colonial masses (e.g. figs, 114 and 116). Over twenty broken spheres were found in one group and this probably represents only a fraction of the total number in the group because they were observed in thin section. When the spheres (or rather shells, as they are wall structures) occur in chains they interpenetrate to form segmented tubes. The example in fig. 117 is 10μ wide and 60μ or more long. Filaments (e.g. fig. 118) are uncommon, about one micron wide, and resemble those found in other jaspers.

Insoluble organic matter is very sparse in all the jaspers and macerations prepared in the usual way (described in section 2.1.3) are overcrowded with insoluble matter - probably degraded barite and gypsum. The sporopollenin obtained from the Umm Gerad jasper is typically yellow and in fragments 20μ wide and 20 to 150μ long. It has an indeterminate structure but is compatible with the cellular structure of the iron oxide e.g. fig. 119.

Wadi al Wazra

Pale brown jasper at Wazra closely resembles that at Yahmun macroscopically and microscopically. A red-brown jasper at Wazra however differs in colour and grain size from most of the other coastal jaspers. It has the appearance of reticulated quartz crystals in a matrix of granular quartz having an even grain size, about 200μ wide, and relatively straight boundaries. The biomorphs in this jasper are composed of manganiferous iron oxide (or oxyhydroxide) and predate the quartz grain boundaries (fig. 120). Typical views of this jasper are shown in figs. 121 and 123.

The goethitic matter occurs as spheres, commonly with concentric shells, as irregular masses, as filaments and as trichospheres (spherical masses of radiating tapered filaments).

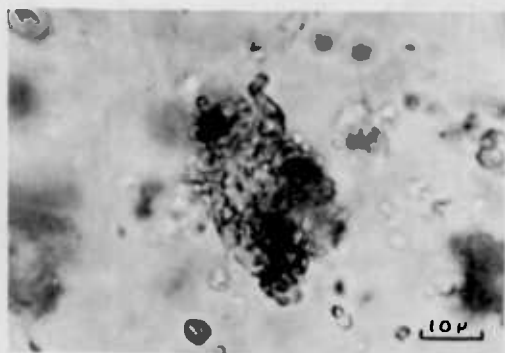


Fig. 119. TLM photograph of typical insoluble organic matter in Umm Gerad jasper. (SM in PPL).

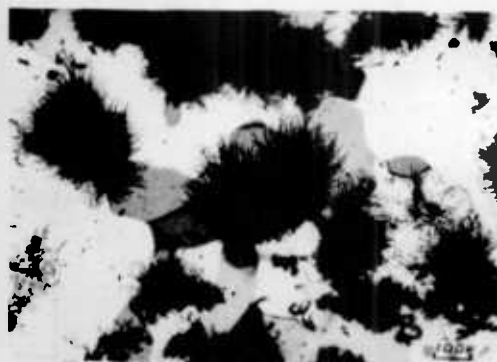


Fig. 120. TLM photograph of Wazra red-brown jasper showing quartz boundaries and goethitic trichospheres. (TS with CP).

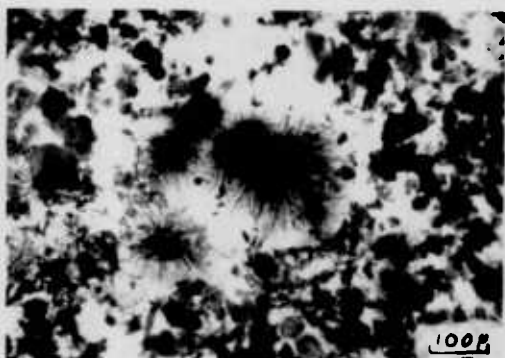


Fig. 121. TLM photograph of Wazra red-brown jasper showing typical view with spheres, filaments and trichospheres. (TS in PPL)



Fig. 122. Enlargement from fig. 121 showing tapering filaments of trichospheres.

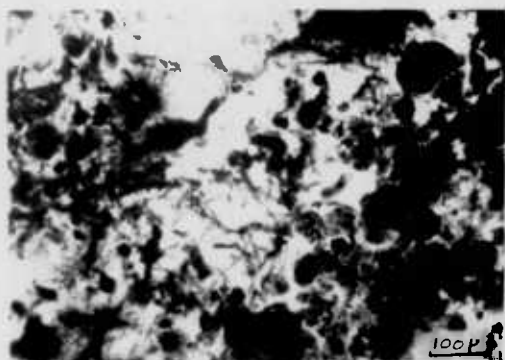


Fig. 123. TLM photograph of Wazra red-brown jasper showing typical filaments, concentric sphere, trichosphere and indeterminate matter. (TS in PPL)

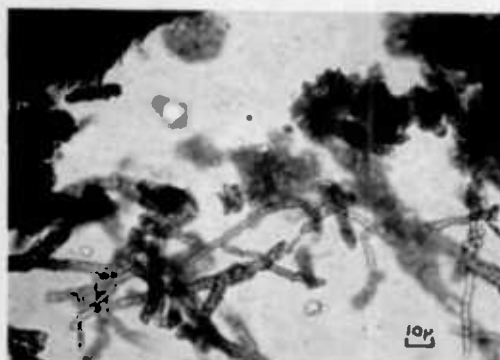


Fig. 124. Enlargement from fig. 123 showing detail of filaments.

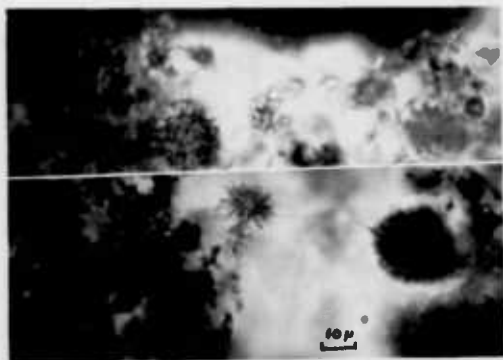


Fig. 125. TLM photograph of Wazra sulphide-bearing jasper showing trichospheres. Composite picture. (TS in PPL).

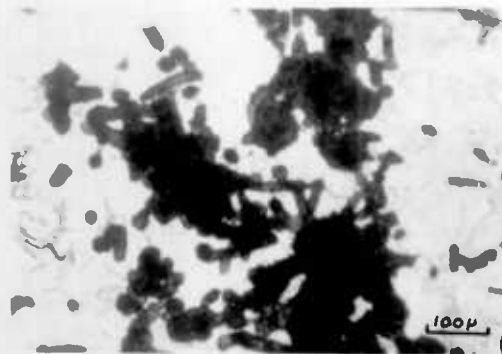


Fig. 126. TLM photograph of stromatolitic limestone from Wazra showing typical goethitic filaments. (TS in PPL).



Fig. 127. Enlargement from fig. 126, upper left, showing segmentation and termination of filament.



Fig. 128. TLM photograph of stromatolitic limestone from Wazra showing more typical goethitic matter. (TS in PPL)

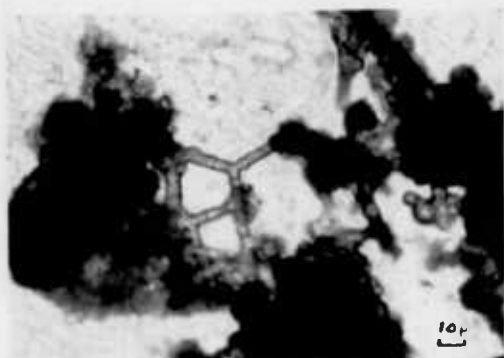


Fig. 129. Enlargement from centre of fig. 128 showing branching filaments.

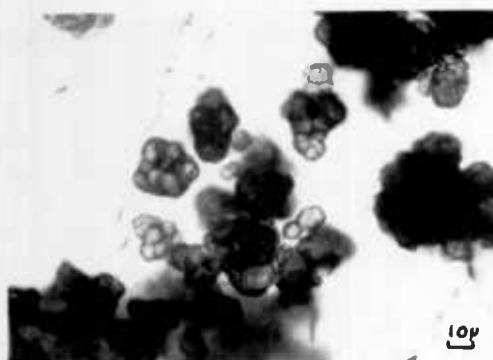


Fig. 130. TLM photograph of stromatolitic limestone from Wazra showing typical goethitic nodular masses. (TS in PPL).

The concentric spheres do not appear to be biomorphic. A large one (in fig. 123) was found to have at least 12 shells of very even thickness and regular shape with dark layers 2μ thick and light layers 1μ thick. Whether or not this pattern extends to the centre of the sphere is not known.

The filaments are 2 to 5μ in diameter and closely resemble those in the Yahmum jasper. They are slightly curved and commonly fiburcated (figs. 123, 124). The nature of the bifurcation is difficult to determine, but it appears to be true branching.

In another light brownish grey jasper containing galena at Wazra are biomorphs similar to those above, but generally the goethite in this rock is difficult to observe due to its fine size. The quartz however is very similar to that described above. Unfortunately clouding by alteration products obscures the regions around the galena. Trichospherical structures in this jasper differ slightly from those in the red-brown jasper (e.g. fig. 125).

A brown stromatolitic limestone associated with the veins at Wazra contains coarser goethitic biomorphs than the vein jasper. Typical views are shown in figs. 126, 128. The biomorphs consist of true-branched filaments (hyphae ?) about 3μ wide (fig. 129) and segmented filaments (hyphae ?) about 10 to 20μ wide (fig. 127) and masses of interpenetrating spheres or nodular masses (fig. 130). Also there are amorphous masses and uncommon concentric spheres. The segmented filaments closely resemble the limonitic-filled algal borings in ferruginous calcareous onkolites ("snuff boxes") described by Gatrall, Jenkyns and Parsons (1972). Dr. Jenkyns kindly examined fig. 126 and agreed that the material definitely appears to be biogenic and is either algal or possibly fungal. The filaments also closely resemble the calcified filaments of an endolithic alga described by Schroeder (1972) except that the calcified filaments are 30 to 40μ wide.

Rerga

The Rerga jasper consists essentially of a meshwork of fine goethitic filaments transecting quartz grains, about 100μ wide in a xenomorphic



Fig. 131. TLM photograph of Barga jasper showing zoning, patchy extinction and xenomorphic fabric of the quartz (TS in PPL).

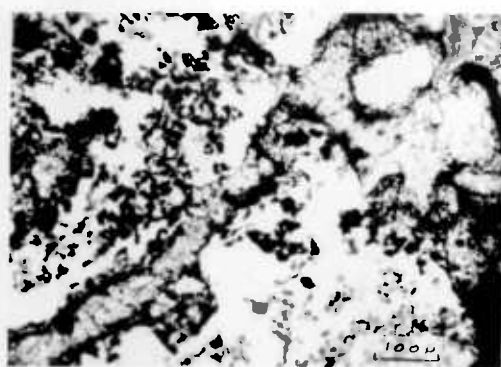


Fig. 132. TLM photograph of Barga jasper showing typical view with web-like masses of filaments in confined zones (TS in PPL).

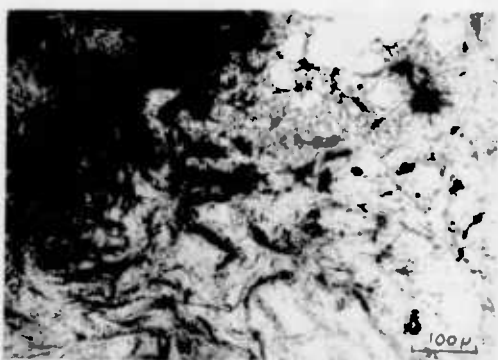


Fig. 133. TLM photograph of Barga jasper showing typical view with web-like nebulous masses of filaments. (TS in PPL)

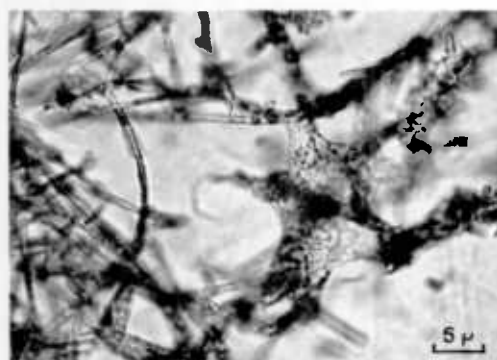


Fig. 134. TLM photograph of Barga jasper showing curved filament and tiny annulus associated with stellate body. (TS in PPL).

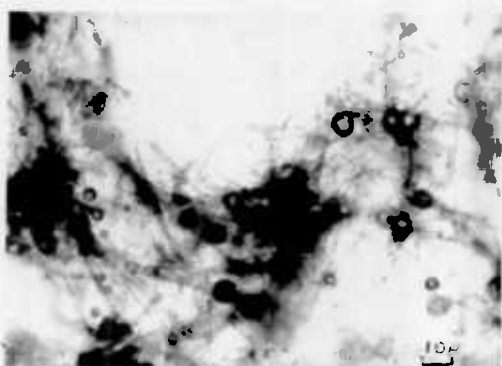


Fig. 135. TLM photograph of Barga jasper showing filaments and spheres. At upper right tiny annuli are just visible (TS in PPL).

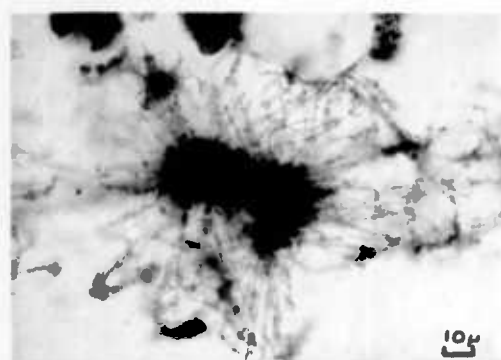


Fig. 136. TLM photograph of Barga jasper showing trichospherical filaments and tiny annulus (just visible in upper right). (TS in PPL).

fabric (using Lovering's terminology, 1972). In places the quartz is zoned in a manner characteristic of free crystal growth (fig. 131). This quartz commonly has undulose or patchy extinction between crossed polars.

Typically the goethitic filaments occur in dense web-like masses, confined to zones in some places (fig. 132), and in nebulous masses in other places (fig. 133). The filaments are very thin and even; some are 0.5μ diameter and 75μ or more long. They are straight to slightly curved and commonly branch. In places they form sheaves like the material from Wheal Jane mine described later. Tiny ring-like bodies 1 to 3μ diameter are associated with the filaments (in figs. 134, 135 and 136 but difficult to observed in these photographs). There are also spheres 2 to 8μ across associated with the filaments (fig. 135) and in irregular masses (fig. 137). Trichospherical developments of the filaments are also present (e.g. fig. 133, 136). Stellate bodies (fig. 134) are uncommon in this jasper and were not noted in the jaspers with coarser biomorphs.

Maceration of the jasper yielded little sporopollenin. Dr. M. Muir kindly identified some grass pollen and conifer pollen fragments of Tertiary to Recent age. The remaining material is unidentifiable degraded tissue. The pollen could have been incorporated in the vein at any time from the porous Miocene wall rock.

Merataba

This jasper resembles the jaspers at Yahmum and Umm Gerad. The quartz is granular (Lovering's terminology, 1972) with a grain size of approximately 70μ . A typical view is given in fig. 138. Much of the biomorphic matter in this rock occurs as cellular masses very like that at Umm Gerad. Interpenetrating spheres in chains and nodular masses are relatively common e.g. fig. 139. Filaments like those at Yahmum and Wazra are less common but nevertheless are abundant e.g. figs. 140 and 138. In places there are filaments (hyphae ?) up to 10μ or more in diameter (fig. 141). These coarse filaments are closely related to the spheres. In places are some concentric spheres associated with Yahmum type filaments (fig. 140).

When macerated, this jasper yielded little sporopollenin and this is unidentifiable (e.g. fig. 142).

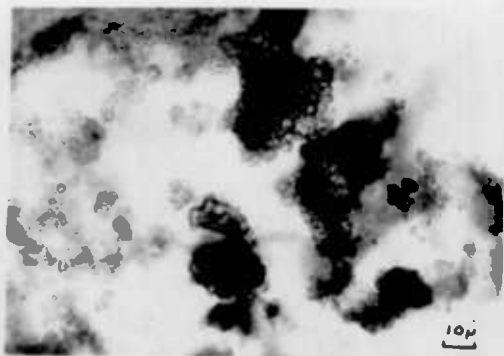


Fig. 137. TLM photograph of Rerga jasper showing irregular masses of goethitic spheres. (TS in PPL).

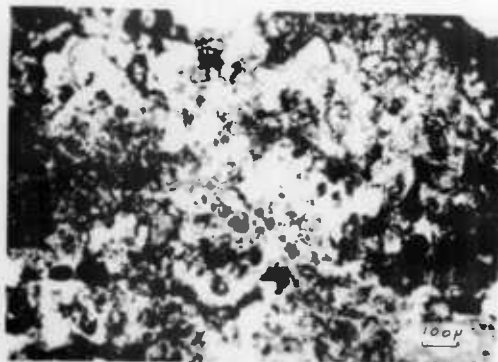


Fig. 138. TLM photograph of Merataba jasper showing typical structure. (TS in PPL).

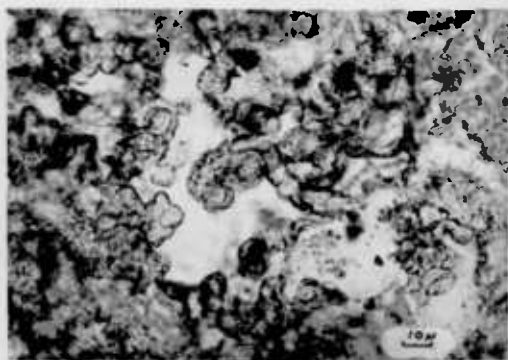


Fig. 139. TLM photograph of Merataba jasper showing interpenetrating spheres in chains and masses. (TS in PPL).

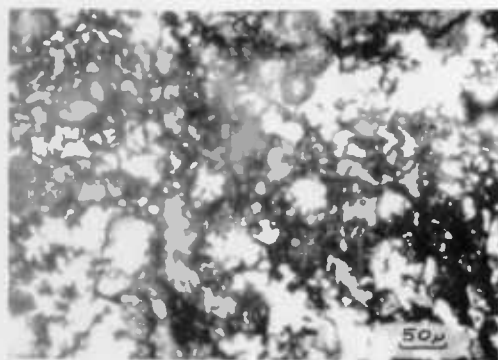


Fig. 140. TLM photograph of Merataba jasper showing typical filaments. (TS in PPL).

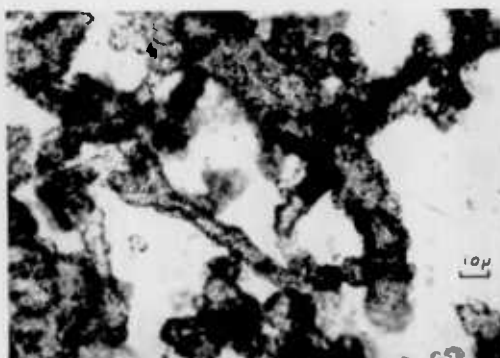


Fig. 141. TLM photograph of Merataba jasper showing detail of filaments. (TS in PPL).

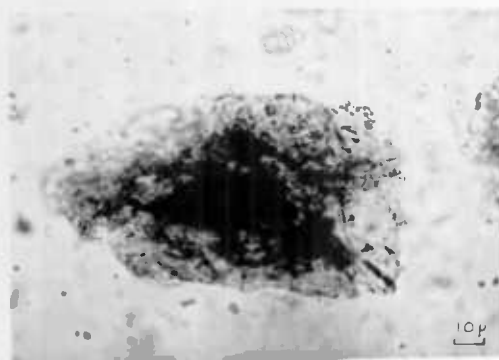


Fig. 142. TLM photograph of typical indeterminate insoluble organic matter from Merataba jasper. (SM in PPL).

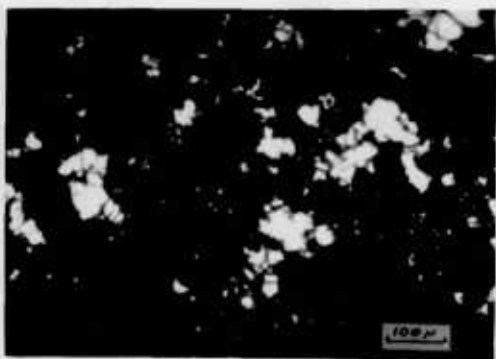


Fig. 143. TLM photograph of Yalmum jasper showing typical view. Enlargement shown in figs. 147 and 148 (TS in PPL).



Fig. 146. TLM photograph of Yalmum jasper showing filaments with rough surfaces and nodular developments (TS in PPL).

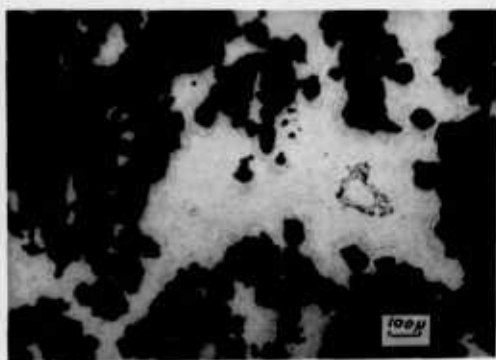


Fig. 144. TLM photograph of Yalmum jasper showing growth-line zoning in quartz (TS in PPL, substage lowered for increased relief).

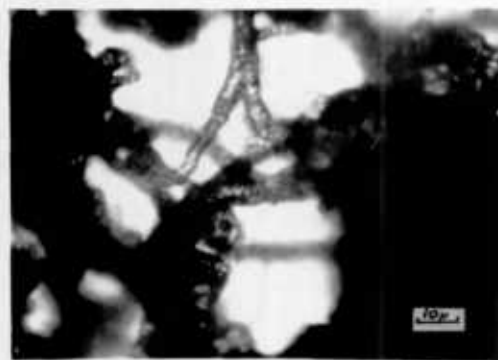


Fig. 147. Enlargement of area in lower centre of fig. 143 showing true branching. Same area as fig. 148 but at different focal level.



Fig. 145. Same area as shown in fig. 144 but with CP to show patchy extinction and relation of zoning to grain boundaries.



Fig. 148. Enlargement of area in lower centre of fig. 143 showing false branching. Same area as fig. 147 but at different focal level.

Yahmum

The Yahmum jasper typically consists of goethitic filaments and amorphous masses in quartz of grain size 50 to 100 μ e.g. fig. 143. As with the other jaspers the goethitic filaments and masses are so fine and tangled that they are commonly invisible due to the restricted depth of field or to the opacity of the overlapping elements. The quartz generally has a granular texture but in places is xenomorphic and of the jigsaw puzzle type (Lovering's terminology, 1972) and has patchy to undulose extinction between crossed polars. Colloform growth lines in chalcedonic quartz are clearly visible in some areas of the jasper (when the microscope substage is lowered). The growth-line zoning conforms to nucleation sites and to the outlines of iron and manganese oxide masses (figs. 144, 145).

The goethitic filaments are generally about 4 μ wide and are of an even width though the surfaces are rough and nodular in places (e.g. fig. 146). They are relatively straight and bifurcate with apparently true branching in some places (fig. 147) and false branching in others (figs. 148, 149). They terminate with hemispherical or nodular endings. The internal structure of the filaments is difficult to ascertain; some appear to be segmented, others do not, and some almost appear to consist of interpenetrating spheres (fig. 149). Nodular masses of interpenetrating spheres also occur isolated from the filaments (fig. 150). Some double spheres were also observed. Two are shown in fig. 151 (one is partially obscured). No concentric spheres of the type found at Wazra were observed in the Yahmum jasper but some non-biomorphic flat layered tabular fragments of goethitic material in the Yahmum jasper may be related to the concentric spheres. Manganese oxides in the jasper are discussed in section 2.3.5.

Maceration of this jasper yielded a little indeterminate fine mucilaginous matter and much insoluble mineral matter.

Other Jasper Veins

The best outcrops of jasper veins found are those from which the specimens described above were collected. Other veins to the north in the coastal region are poorly exposed and somewhat weathered. Doubtful biomorphs occur in several of these. The best of these is from a vein in Cenozoic limestone between Sharm Dumaygh and Sharm Antar (locality given in fig. 45). Stromatolitic and coral limestones occur near the vein. In thin



Fig. 149. TLM photograph of Yahmum jasper showing filament with false branching, nodular termination and cellular structure (TS in PPL).

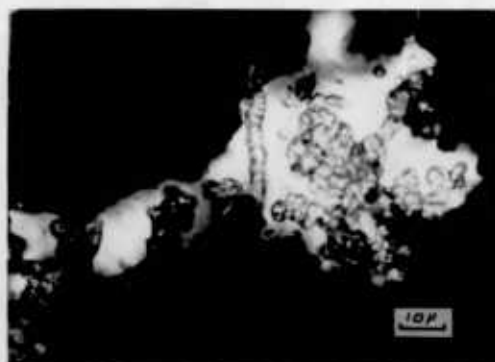


Fig. 150. TLM photograph of Yahmum jasper showing nodular masses of interpenetrating spheres (TS in PPL).



Fig. 151. TLM photograph of Yahmum jasper showing double spheres (TS in PPL).



Fig. 152. TLM photograph of jasper from between Sharm Dumaygh and Sharm Antar showing large inorganic(?) micro-structure (TS in PPL).



Fig. 153. TLM photograph of jasper from between Sharm Dumaygh and Sharm Antar showing coarse filaments (TS in PPL).

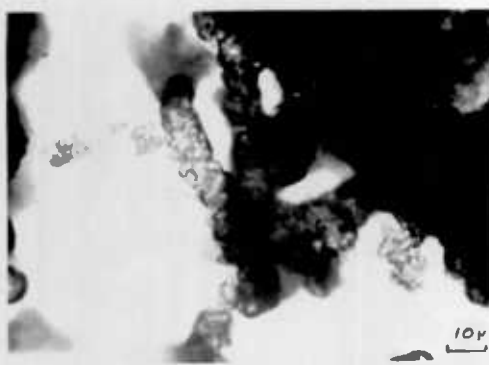


Fig. 154. TLM photograph of jasper from between Sharm Dumaygh and Sharm Antar showing chains and amorphous masses of spheres (TS in PPL).

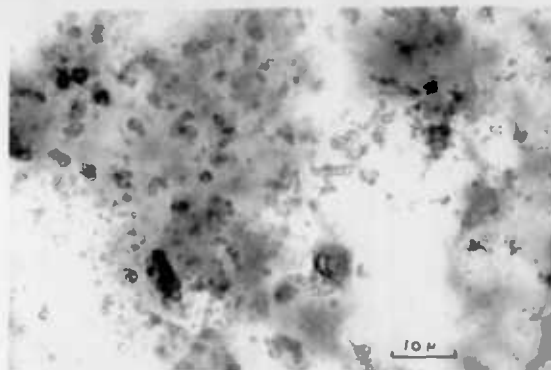


Fig. 155. TLM photograph of ferruginous opal from Sharm al Badu area showing indeterminate goethitic grains (TS in PPL).

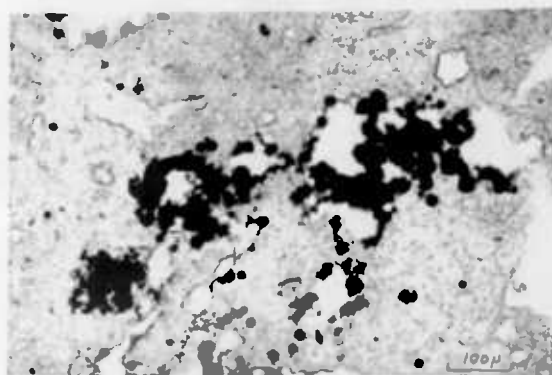


Fig. 156. TLM photograph of ferruginous opal from Sharm al Badu area showing goethitic spheres and masses of spheres near quartz-filled cavity (TS in PPL).



Fig. 157. TLM photograph of Sphaerotilus filaments and other sediment from Silwood Lake (SM in PPL). Specimen collected by R. Spicer of Imperial College.

section at low magnification are some unusual structures which are either probably inorganic forms or possibly fungal colonies (fig. 152). The quartz in this jasper resembles that at Yahmum. The goethitic biomorphs consist of coarse filaments (hyphae ?) 10μ wide which are possibly segmented, and masses and chains of interpenetrating spheres (figs. 153, 154). The biomorphs transect quartz grain boundaries in places.

The most northerly vein found by the writer is near Tuwyil al Kibrit in the Sharm al Badu area north of Duba (locality given in fig. 42). This occurs in Cenozoic sandstone and consists essentially of ferruginous material and common opal with micro-cavities and veins filled with chalcedony and quartz. Much of the opal contains goethitic material disseminated as clouds of indeterminant submicroscopic grains (fig. 155). It also contains some manganiferous trichospheres of the type found at Yahmum described in section 2.3.5 and goethitic spheres 10 to 20μ diameter and masses of spheres occur in relatively clear opal and more commonly at the margins of quartz-filled cavities (fig. 156).

2.3.3

BIOGENIC STATUS OF THE IRON BIOMORPHS

At least some and possibly all of the biomorphic structures described are most probably biogenic. It would be quite difficult to ascribe an inorganic origin to the interpenetrating spheres in the Umm Gerad and Merataba jaspers, the rings and stellate body in the Rerga jasper and the filaments in the Yahmum jasper. These structures and their assemblages are commonly produced by microorganisms. The ratio of biomorphic to non-biomorphic matter in these jaspers is higher than is found in many environments of biogenic iron deposition (e.g. see Harder 1919).

What organisms these biomorphs represent is difficult to determine. To distinguish between bacteria, fungi and algae is not always easy and can be almost impossible under conditions such as we have here. The biomorphs in the stromatolitic limestone at Wazra seem to represent boring algae. Muir et al (1974) have noted the similarity between the filamentous iron oxide biomorphs of Sphaerotilus (fig. 157) collected by R. Spicer at Silwood Lake and the filamentous biomorphs in the jaspers at Yahmum, Wazra and Rerga. It might be argued that the macerations provided evidence of insufficient organic matter for the support of Sphaerotilus but R. Spicer (Imperial College, pers. com.) found that the Sphaerotilus sheaths and much of the material associated with them at Silwood Lake is soluble in hydrochloric acid. He also found that the iron oxide in Silwood Lake is associated with amorphous silica. Leptothrix is another possibility. These bacteria live in media containing very little organic matter and form tubular threads which become coated with copious quantities of ferric hydroxide with increased growth. Zajic (1969) noted that some species exhibit true branching. The coarseness and common true branching of some of the filaments together with hemispherical terminations however suggests that the filamentous biomorphs in the jasper more commonly represent fungi than bacteria and it is probable that both lived together in the vein environments. The nodular irregularities in the filaments in the Yahmum jasper are very similar to features of the iron fungus Phycomyces Frodinhamii found by Ellis (1915). The branching shown in fig. 146

resembles Ellis's "whorled branching". The organisms in the veins would most probably not be algae (or other photosynthetic organisms) as the environment of growth would have been dark.

2.3.4

THEORETICAL CONSIDERATIONS ON BIOGENIC VEINS

Evidence has been presented to show that biogenic material occurs in the jasper veins associated with barite and galena in the Red Sea coastal region. Indeed most of the iron in the jasper appears to have been biogenically fixed by microorganisms which lived in situ as the veins formed. This is the first record of a biogenic vein and initially the proposition seems preposterous and revolutionary. It was previously inconceivable to some geologists. For example one of the early arguments against the biogenic origin of framboids was that some which occur in veins cannot be biogenic simply because they formed in veins (e.g. Glasson in Flinter et al 1969, p. 62).

The reasons for the general assumption that all veins are non-biogenic appears to be firstly that they are usually formed under conditions hostile to life (e.g. high temperatures and pressures) and secondly that there were none on record. It will be shown however that the environments in which certain veins form e.g. those found in this study, are not at all hostile to some organisms.

As the veins are fracture-fillings the pressure during their formation would have probably been normal hydrostatic pressure depending largely on the depth below the ground surface. This is unknown but as the Umm Gerad jasper was collected from a depth of 15 m and the vein has been eroded at the surface, it can be assumed that the jasper formed at a greater depth. The host rocks to the veins at Yahmum and Rerga are weakly lithified hence it is unlikely that the veins formed at depths greatly exceeding a kilometre. Microorganisms are well known to live at great depths. They live at about 2 km at the bottom of the Red Sea at about 400 atm (Truper 1969, and this study), and Zobell (quoted by Kuznetsov et al 1963) recorded that certain sulphate-reducing bacteria (Vibrio desulfuricans) which occur extensively in stratal waters can live and reproduce at a pressure of 1800 atm.

The temperature at which the veins formed is unknown though trace element data (section 3.2.3) and the lack of alteration of trace organic matter in the veins and of the host rocks suggests the temperature was relatively low. There is no evidence that the veins formed at a temperature much above that of the host rocks.

The chemical environment in which the jasper formed was probably conducive to the growth of iron bacteria. Formation of the jasper followed that of the barite and no biomorphs were found in the barite itself. The deposition of iron oxides together with silica and little else is a relatively common phenomenon in surficial aqueous environments, and deposition in subterranean fractures is probably similar. Some of the surficial iron oxide and silica deposits are biogenic but the origin of the silica in these deposits is seldom obvious. For the present purpose this is not important because the quartz in the jasper veins does not have a biogenic aspect, and non-biogenic silica is not harmful to many organisms. Zajic (1969) noted that manufacturers of colloidal solutions of silica gel encounter microbial growth so frequently that they have to use preservatives. He further noted that some water pipes in India were found to contain "iron-siliceous" encrustations formed by Leptothrix ochracea. Harder (1919) noted that some spring waters with iron bacteria growing in them contained 18 ppm Si amongst Ca, Mg salts. Presumably silica would precipitate before the Ca, Mg salts as the water evaporated. He also found a relatively high silica content (7%) in a bacterial iron scum collected inside a mine but considered that some of this could be due to silicate contamination.

Minette type iron ores and especially Lake Superior type iron ores contain silica and at least some biogenic iron.

The iron ore deposited from hydrothermal solutions associated with basic volcanism in the Red Sea deeps, the Mid-Atlantic Rise, the East Pacific Rise and at Santorini Caldera is associated with some silica (Puchelt, 1973; Cook, 1972 and Burnett, 1971). Biomorphic structures of goethite occur in the Red Sea deeps (section 2.2.1), in the East Pacific Rise (writer's observations and von der Borch, 1971) and at Santorini Caldera (writer's observations). These biomorphic structures are probably biogenic.

Allen (1934) and Walter, Bauld and Brock (1972) found growing stromatolites composed mostly of amorphous silica, around hot springs. The silica however is precipitated non-biogenically due to cooling and evaporation of water but nevertheless forms an integral part of the stromatolites. Amorphous silica is also deposited with iron bacteria in Silwood Lake (R. Spicer pers. com.).

The absence of light would help rather than hinder iron bacteria. The situation can be likened to water pipes and water drains in mines. Iron bacteria are well known for clogging water pipes (e.g. Zajic 1969) and iron bacteria commonly form deposits in mine waterways. Harder (1919) found them to be quite common in mines at depths of at least 100 m below surface. Professor G.R. Davis has also noticed this phenomenon and kindly presented the writer with some dried-out material from Wheal Jane mine. When crushed and prepared as a smear mount this material has the appearance of agglutinated filaments although much of the material is generally amorphous (fig. 158). A thin section cut normal to the weak foliation of epoxy-resin-impregnated material has a different appearance (figs. 159 and 160). Although it contains filaments very like those in the Yahmum jasper (fig. 160) most of the Wheal Jane material is nodular or amorphous. This material is amorphous to XRD although in appearance and composition it resembles goethite. Dr. M. Muir (pers. com.) notes that the filaments closely resemble those of the iron bacterium Gallionella.

Harder (1919) seems to have foreseen the possibility of biogenic vein formation. He stated (p. 25), "I have seen cultures of iron bacteria in places in mines where it seems quite improbable that they could have been brought by any other agency than the water itself, and it does not seem at all impossible that they may have been carried downward by surface waters through the larger openings in the soil or rock for many hundreds of feet. This opens the interesting possibility that iron bacteria may be active in fissures or other large underground openings entirely unconnected with mines and may thus be instrumental in the formation of many of the fissure deposits and replacement deposits of iron ore which are so common. Their presence in such localities is of course difficult to prove unless they should be found in underground openings which are penetrated during mining operations. Such a phenomenon to my knowledge has not been observed, but it might easily escape notice unless particular attention were called to it". The preservation of the material in silica enhanced the discovery of the Arabian material.



Fig. 158. TLM photograph of crushed dried iron bacterial colony from Wheal Jane mine waterway, Cornwall (SM in PPL). Specimen collected by Prof. G.R. Davis.

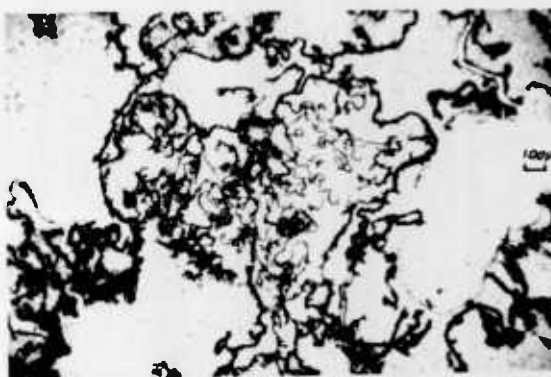


Fig. 159. TLM photograph of iron bacterial colony from Wheal Jane waterway as shown in fig. 158 (TS in PPL of resin impregnated dried material).

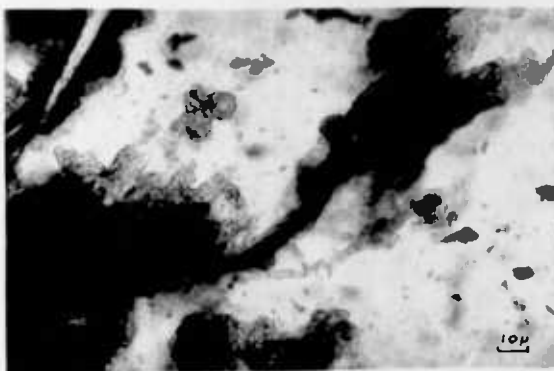


Fig. 160. TLM photograph of TS shown in fig. 159 showing typical view at high magnification with Yalumus type filament and nodular and amorphous matter.

The solutions from which the iron was deposited probably contained little organic matter because very little was found in the macerations and some iron bacteria such as Spirophyllum and Leptothrix thrive best in water containing little or no organic carbon (Harder, 1919; Zajic, 1969). The solution may have resembled those forming the hot springs at Jabal Markub and other localities along the southern Red Sea coast (section 1.2.4). The apparent lack of silica in these waters could be due to its precipitation at depth. It will be noted also that copious iron oxide deposits form around the outlets of these springs even though the water only contains a trace of iron.

Despite an extensive search of the literature no reference to biogenic jasper veins was found. There is scarcely any mention of biogenic activity in vein type environments. Perring (1973) recorded evidence of biogenic activity in a vein in North Derbyshire, England, but this activity appears to have been initiated by weathering. An important exception however concerns the source of sulphide ions in Mississippi Valley type environments. Jensen and Dessau (1967) proposed that the sulphate in connate water could have undergone bacterial reduction to hydrogen sulphide as the solutions passed through porous zones, such as reefs, that contained enough residual organic matter to provide an energy source for the sulphate-reducing bacteria. Further study of the sulphur isotopes by Sasaki and Krouse (1969) however has since thrown the origin of the sulphur into some doubt. The bacterial sulphate reduction postulated by Jensen and Dessau (1967) and others was thought of as a relatively near surface phenomenon but it is nevertheless a subterranean process and could occur at considerable depths. Subterranean bacterial sulphate reduction is a relatively well accepted hypothesis for some reactions in certain oil fields (e.g. Kuznetsov et al 1963). Dr. H. Puchelt (Universitat Tubingen, pers. com.) has found them to be active in a German oil well at 120°C. Kuznetsov et al (1963) recorded that bacteria occur down to 4 km depth, and Moore (1969) recorded that bacteria have been found in water from depths exceeding 5 km. Subterranean epigenetic sulphur deposition is also recorded by Kuznetsov et al (1963). A typical deposit at Shor-Su in Russia apparently was formed by biogenic oxidation of hydrogen sulphide derived from a distant biogenic sulphate-reducing source. Their evidence is compelling and it is interesting to note that the sulphur formed as tiny crystals. Hydrogen sulphide is also sometimes directly

oxidized to sulphate by subterranean bacteria e.g. in the oil fields of Kuybyshev Oblast (Kuznetsov et al 1963). Any non-photosynthetic microbial activity, especially if thermophilic, might occur at shallow to moderate depths, so many of the processes postulated for biosynthetic ore deposition can possibly be bioepigenetic and some theories may therefore need revision. Microbes can remove rock matter (including silicates) as well as deposit it so they can presumably form replacement deposits as well as fracture-fill deposits. Manganese oxide deposited in a subterranean environment has also been found by the writer as outlined in the following section.

SECTION 3 G E O M I C R O B I O L O G Y
 I N T H E C O A S T A L P L A I N

Contd.

2.3.5

MANGANESE OXIDE BIOMORPHS

LITERATURE ON MANGANESE DEPOSITING MICROORGANISMS

Far less is known about manganese depositing microorganisms than iron depositing microorganisms and references to the subject are difficult to find. It is known however that after many attempts Molisch in 1910 (quoted by Harder 1919) finally managed to grow Leptothrix and Cladothrix on a culture containing manganese salts.

In 1925 Thiel managed to culture fungi (Oidium manganifera) that precipitated manganese oxide but he also noted that no such fungi were previously known. The microorganisms (including bacteria) he studied were isolated from peat, soil, iron-spring slimes and a marine mud. He concluded that "since organisms that precipitate manganese are almost universal in their distribution, we are justified in assuming that in many cases they were instrumental in precipitating some of the manganese in the sediments that are protores of commercial deposits". Another result of Thiel's work is that he found that sulphate reducing organisms grown under anaerobic conditions can precipitate manganese oxide around the colonies.

Some iron bacteria are also capable of precipitating manganese and some, perhaps all, manganese bacteria are capable of precipitating iron. Bacterially deposited iron oxide can be relatively pure but bacterially precipitated manganese oxide usually is associated with some iron oxide. Siderococcus, Gallionella and Ochrobium oxidize iron exclusively and Metallogenium, Caulococcus, Kusnezovia oxidize mainly manganese (Perfilev et al 1965). Certain Crenothrix, Leptothrix, Cladothrix and Clanothrix collected from mine waters precipitate manganese as readily as iron (Thiel, 1925).

In 1952 Cornelius and Ploehinger postulated that some sedimentary manganese ore in Austria was precipitated biogenically and in 1962 Obut (quoted by Zajic 1969) claimed that manganese oxidizing bacteria participated in the formation of the Karu-Alma deposit in Russia. Zajic (1969) and Stanton (1972) considered that this probably also applied to the Nikopol and Chiatura deposits. Perfilev et al (1965) produced an outstanding treatise on manganese bacteria giving details of growth, morphology, new species and new techniques for geomicrobiological study.

Oceanic manganese nodules are considered to be at least partly biogenic by many writers (see Jedwab 1971, Andrushenko and Skornyakova 1970 and Zajic 1969 for references).

Recently Crear et al (1972) stated that manganese fixing bacteria are essential to the incorporation of manganese into lacustrine sediments. They also noted that the biota affects manganese behaviour by metabiolizing it, by catalysing manganese reactions and by influencing Eh, pH and organic ligand concentrations.

It should be noted that microorganisms can reduce certain manganese salts and can otherwise bring manganese into solution. Also sulphate reducing bacteria grown under anerobic conditions can precipitate manganese oxide around the colonies (Thiel, 1925).

Manganese oxide can be carried as a colloid stabilized by organic matter. When this is degraded by microorganisms the manganese oxide is released.

OBSERVATIONS

Yahnum

Manganese oxide biomorphs were first noticed by the writer in the thin sections of vein jasper from Jabal Yahnum in which the goethitic biomorphs were found.

The manganese biomorphs in this jasper generally occur as trichospheres i.e. spheres of radiating tapered filaments. The trichospheres range in diameter from 20μ to 100μ and the filaments are about 1μ wide. The filaments commonly are slightly curved and some bifurcate. Some trichospheres are connected by branching filaments (fig. 161). In this respect

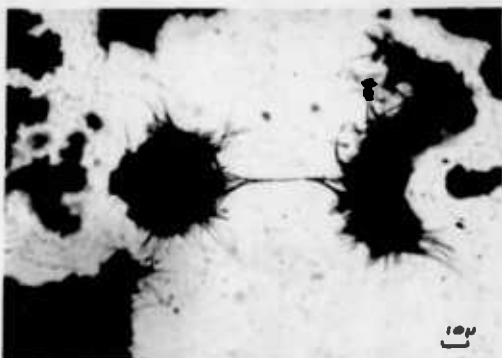


Fig. 161. TLM photograph of Yahnum jasper showing manganiferous trichospheres connected by branching filaments (TS in PPL).

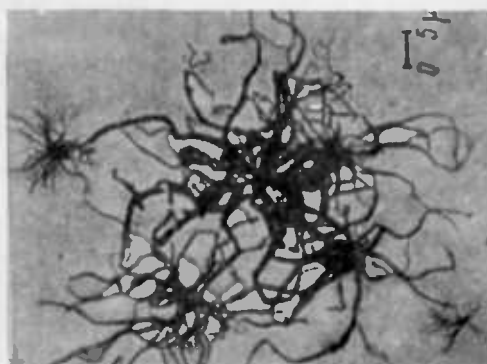


Fig. 162. TLM photograph of trichospherical colony of Metallogenium in capillary mount. Copied from fig. 7/17 of Perfilov et al (1965).

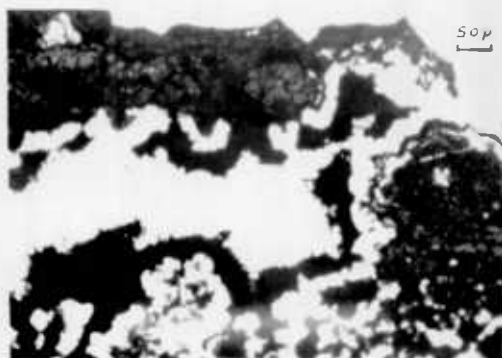


Fig. 163. TLM photograph of Yahnum jasper showing closely connected manganiferous trichospheres lining a quartz-filled cavity (TS in PPL).

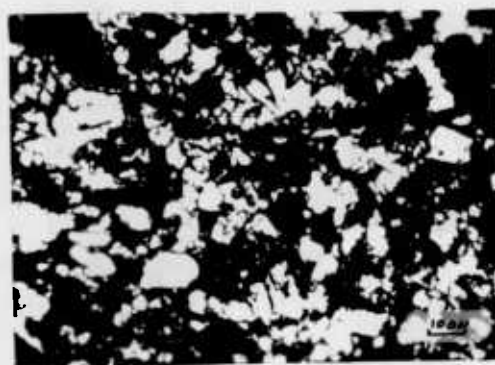


Fig. 164. TLM photograph of Yahnum jasper showing interconnecting manganiferous filaments (TS in PPL).

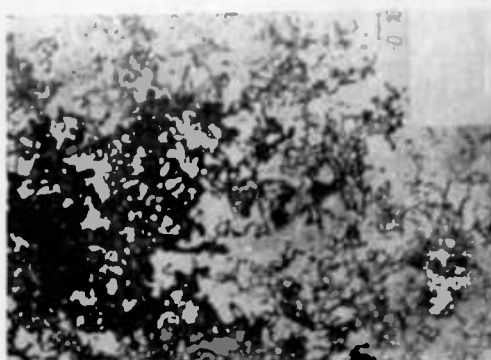


Fig. 165. TLM photograph of zoogloal-reticulate growths of Metallogenium in capillary mount. Copied from fig. 10/5 of Perfilov et al (1965).

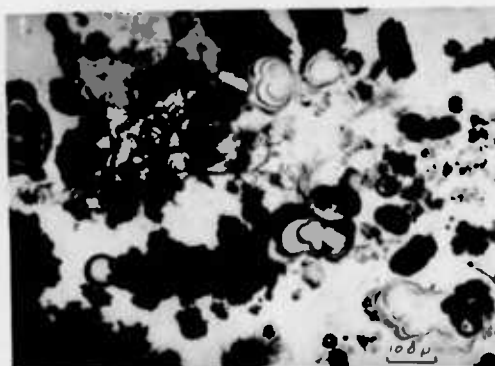


Fig. 166. TLM photograph of Wazra black jasper showing general view with trichospheres, filaments and colloform masses (TS in PPL).

they resemble Metallogenium in its stage of trichospherical colony development (e.g. fig. 7/17 of Perfilov et al 1965, reproduced here as fig. 162). They differ from Metallogenium however in having a greater number of filaments in each trichospherical colony and the filaments also appear to be straighter.

Manganese deposition is unevenly distributed over some filaments and in a few places is very thin. It is quite possible that many filaments are now invisible because of their thin transparent nature or because of autolysis. These trichospheres in the Yahmum jasper very closely resemble those found in jasper at Wazra (e.g. figs. 120, 121, 122 and 125). Like them the emplacement of the Yahmum trichospheres preceded the crystallization of the quartz in the jasper. The relation of the iron biomorphs to the manganese biomorphs however is less clear although many of the Wazra trichospheres appear to contain both manganese and iron. The manganese biomorphs at Yahmum are enclosed by masses of goethite in some places but more commonly they occur in large interstitial spaces between goethite masses (fig. 161). In some places they are separated from the goethite by a layer of quartz, e.g. fig. 163 where the trichospheres are closely interconnected and form a lining to a quartz filled cavity.

The manganese oxide and the goethite generally appear to be distinctly separate in the Yahmum jasper and the manganese oxide is a minor component.

Apart from the trichospheres, manganese biomorphs also occur in the Yahmum jasper as interconnecting filaments (fig. 164). These vaguely resemble zoogloea-reticulate growths of Metallogenium illustrated in fig. 10/5 of Perfilov et al (1965) reproduced here as fig. 165.

Wazra

A black jasper at Wazra contains a variety of manganese oxide structures including colorform masses (fig. 166) trichospheres, filaments and clavate masses (fig. 167) which resemble the clavate colonies of Metallogenium described by Perfilov et al (1965, fig. 168 is their fig. 11/4). The clavate colonies of Metallogenium studied by Perfilov et al. grow vertically. By analogy the structures in fig. 167 would be oriented with the right side up, left side down. In this position most of the filaments in the rock would be subhorizontal as they are perpendicular to the clavate structures.

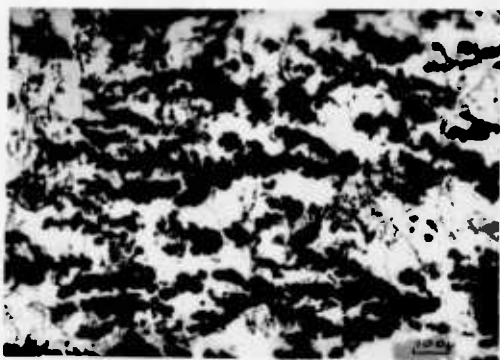


Fig. 167. TLM photograph of Wazra black jasper showing general view with filaments and clavate masses (TS in PPL).



Fig. 168. TLM photograph of clavate colonies of *Metallogenium* in capillary mount. Copied from fig. 11/4 of Perfilov et al (1965).



Fig. 169. TLM photograph of Wazra black jasper showing inorganic crystals (TS in PPL).

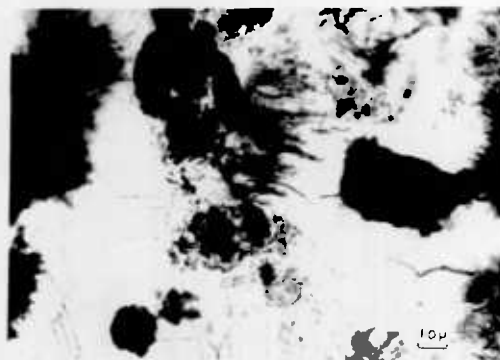


Fig. 170. TLM photograph of Wazra black jasper showing inorganic trichospherical growth (upper centre) and spherical structure of short curved filaments (lower centre; TS in PPL).

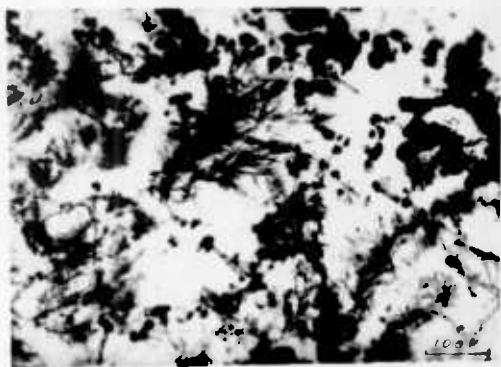


Fig. 171. TLM photograph of Wazra black jasper showing inorganic trichospheres and biomorphic filaments with true branching (TS in PPL).

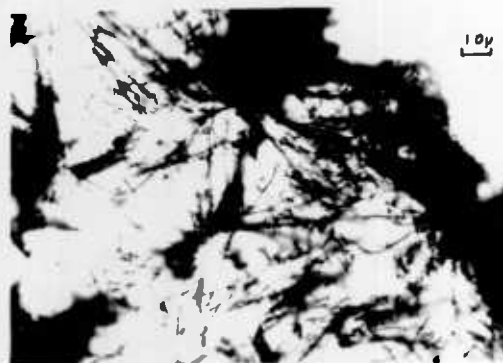


Fig. 172. TLM photograph of Wazra black jasper showing inorganic crystal filaments (TS in PPL).

The trichospherical structures in this rock are problematical. Some are actually distinct inorganic crystal growths e.g. fig. 169 and others which lack a crystal form are probably also inorganic (figs. 170, 171). These have bladed and wispy terminations which are very difficult to observe and to some extent they resemble the artificial trichospherical goethite with blade shaped crystals prepared inorganically by the writer for comparison with goethite from the Red Sea deeps (figs. 102, 103). Definite tubular filaments were not observed in the trichospheres in this jasper hence they are considered to be probably non-biogenic.

Biomorphic filaments in this jasper can also be confused with nematoid inorganic manganese oxide. That in figs. 169 and 172 appears to be inorganic but biomorphic filaments are also common and can be seen in figs. 173 and 171. These filaments are about 5μ diameter and have uneven coats of manganese oxide. They are relatively straight, have hemispherical or nodular terminations and commonly bifurcate with true branching. Septae are visible in some. Cellular masses are also visible in places and some uncommon filaments appear to consist of interpenetrating spheres. Spherical structures composed of short curved filaments (fig. 170) appear to be relatively rare but many may be obscured in thick manganese oxide coats. These structures resemble some structures observed in Lake Superior type iron ore deposits (e.g. type A structures of LaBerge, 1967).

A limestone in the mineralized area at Wazra contains massive manganiferous layers and disseminated manganese oxide which appears to be contemporaneous with the limestone. The massive pyrolusite is blocky in places with the cleavages or partings filled by calcite. The brown limestone is a medium to coarse grained sparite and contains numerous filaments, spheres and trichospheres. These are mostly composed of dark brown manganese oxide but some are coated with yellow to orange goethite and some very pale structures are probably composed largely of organic matter.

The trichospheres resemble the goethitic trichospheres found in the red-brown jasper at Wazra (fig. 122) and the manganiferous trichospheres in the black jasper at Wazra. In places they are so concentrated that the rock is almost opaque in thin section. Some appear to have hollow centres. They generally appear to be biomorphic but the nature of the

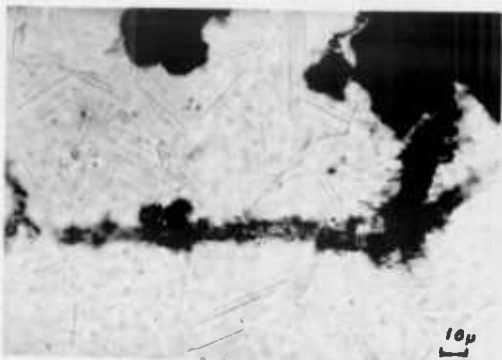


Fig. 173. TLM photograph of Wazra black jasper showing biomorphic filament with uneven manganese coating (TS in PPL).

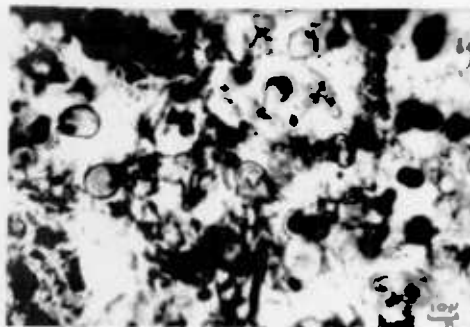


Fig. 174. TLM photograph of Wazra black limestone showing filaments and spherules (TS in PPL).

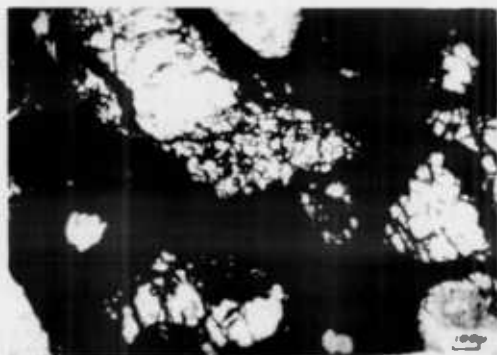


Fig. 175. TLM photograph of Wadi Thalba manganiferous sandstone showing structure with ribbon or sheet elements (TS in PPL).

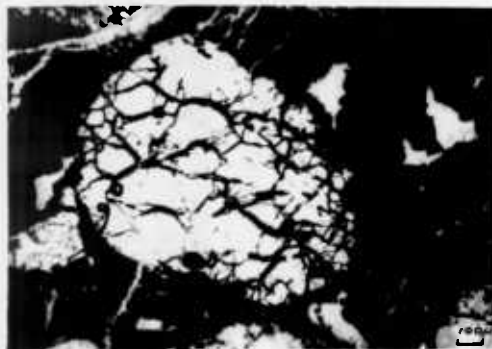


Fig. 176. TLM photograph of Wadi Thalba manganiferous sandstone showing structure with filaments (TS in PPL).

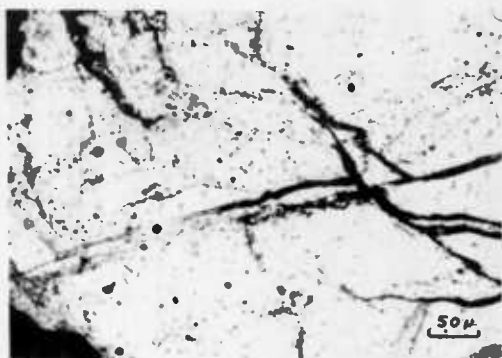


Fig. 177. TLM photograph of Wadi Thalba manganiferous sandstone showing uneven manganese oxide coating on filaments (TS in PPL).

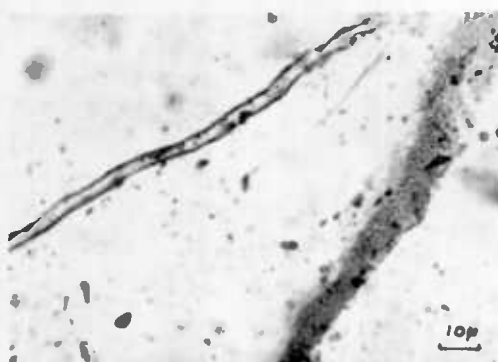


Fig. 178. TLM photograph of Wadi Thalba manganiferous sandstone showing difference between filament (left) and ribbon or sheet (right; TS in PPL).

trichomes is uncertain; some seem to be filamentous but others are ribbon-like. They are evenly sized and range in diameter from 10μ to 40μ .

The filaments are about 2μ wide and some appear to be segmented. The spheres are evenly sized and are about 5μ to 20μ diameter. They have well defined walls and some interpenetrate each other (fig. 174). It is possible that some of the trichospheres have developed from these smooth spheres. The manganese oxide coating on the spheres and filaments ranges from a stain to obscuring masses.

Wadi Thalba

The manganese ore at Wadi Thalba is friable and was impregnated with epoxy resin for thin sectioning. Unfortunately this resin is birefringent and hinders identification of the minerals. In the fringes of mineralization the manganese and iron oxides occur as relatively large interstitial masses, as cement, veins and microveins in a quartz sandstone. The quartz appears to be strained (undulose extinction) and appears to have been extensively cracked and broken after deposition and before emplacement or replacement by the manganese and iron oxides. Some of the opaque material consists of goethite but mostly it is pyrolusite. Calcite has replaced minor iron and manganese oxide in places.

The manganese oxide forms thin films (probably abiogenic) in quartz fractures but also forms reticulate meshes of straight filaments 1 to 3μ wide in strained quartz. Fig. 175 shows many ribbon structures, 176 shows many filaments. Some of the thicker filaments closely resemble channel colonies of microorganisms filled with iron and manganese hydroxide in oceanic nodules described by Andrushenko and Skornyakova (1970). There are also biomorphic structures at Wadi Thalba resembling their patch texture of manganese hydroxide accumulations on a colony of microorganisms. They recorded that "colony development results in the fusion of the separate fibrous like channels. In certain cases, minor filling of the channels results in concentration mainly of iron oxides, whereas mainly manganese oxides accumulate in the case of thinner moss-like forms. It is possible that there are various species of microorganisms, it should be mentioned that they resemble small unicellular organisms hypothetically related to algae known under the name of coating or boring algae".

Also in this rock are vug-like areas filled with chalcedony in which are fine filaments of goethite resembling the trichospheres at Wazra however these form linings to opaque areas rather than typical trichospherical colonies.

Details of the structures in this rock are difficult to determine because of the opacity of the manganese oxide. Black ribbons for example seldom can be distinguished from tubes. The manganese coating on the filaments is variable however and in places it is thin enough to make the filaments visible e.g. fig. 177. The difference between filaments and ribbons (or sheet fragments) is demonstrated in fig. 178. Some filaments appear to be segmented but others do not.

Azlam

In the area of copper-zinc-barite mineralization at Azlam is a black limestone rich in manganese oxide biomorphs. The manganese in this limestone (fig. 179) resembles both the well known pseudo-fossil dendritic form and the clavate colonies of Metallogenium (Perfilev et al 1965). It seems quite likely to the writer that some dendritic forms of manganese oxides are probably bacterial colonies even though these "pseudo-fossils" are classically thought to be non-biogenic. In this limestone the manganese occurs as large amorphous masses, clavate masses and as filaments (fig. 180). Filaments transect calcite grain boundaries so probably predate calcite crystallization. Some filaments (about 5μ wide) have spiky surfaces possibly due to encrustation of radiating pyrolusite crystals (fig. 181). Many filaments are unevenly coated with manganese oxide and commonly occur as projections out of manganese masses very much like the habit of the iron bacterium, Crenothrix (e.g. Harder 1919, fig. 2). These are irregular in shape, about 1μ diameter, tapered at the ends and segmented (fig. 182). They also appear to radiate from central points like Metallogenium but connections are obscured by manganese oxide deposition. In some places are cellular aggregates (cells about 2 to 3μ dia.) which form masses and bifurcating tubular colonies about 5μ dia. (fig. 183).

Other Areas

Manganese biomorphs probably occur in several other localities in the coastal plain not studied fully by the writer. Their discovery in the

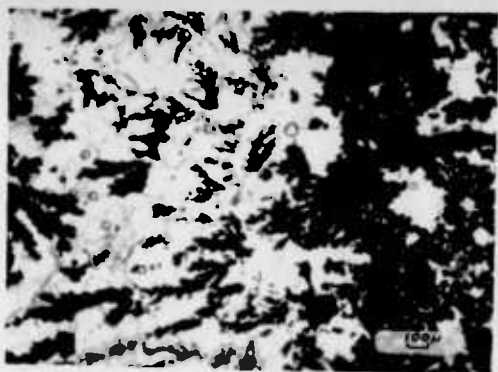


Fig. 179. TLM photograph of Azlam black limestone showing general view with clavate structure (TS in PPL).

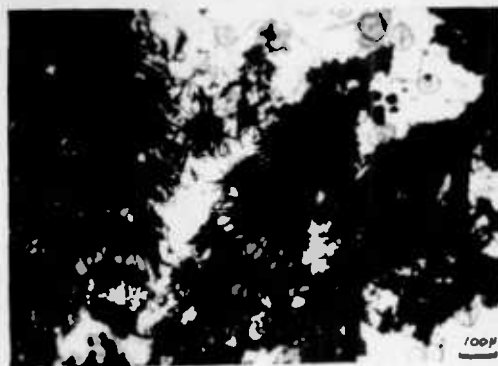


Fig. 180. TLM photograph of Azlam black limestone showing general view (TS in PPL).



Fig. 181. TLM photograph of Azlam black limestone showing filaments with spiky surfaces (TS in PPL).



Fig. 182. TLM photograph of Azlam black limestone showing segmented, curved filaments projecting through mass of manganese oxide (TS in PPL).



Fig. 183. TLM photograph of Azlam black limestone showing cellular filaments (TS in PPL).

rocks described above was made relatively late in this study. Also they are easily missed because of their fine size and their obscurity in opaque areas and in deeply stained areas. They may also be obliterated by some diagenetic processes.

No manganese biomorphs were found in the limestones at Jabal al Buwana or at Sharm al Badu. In these limestones the manganese oxide is loosely attached and is difficult to preserve during the preparation of the thin sections. Where it is observed in thin sections it occurs as pore-filling masses.

The limestone associated with the mineralization at JJM10, Jabal Dhaylan, is black in places with manganese oxide. This limestone is very fine grained and contains biomorphic goethite as found at Yahmum. It also contains irregular fuzzy clots of iron and manganese oxides which, at the edges, consist of tiny filaments a micron or less wide.

SECTION 3 G E O M I C R O B I O L O G Y
 I N T H E C O A S T A L P L A I N

Contd.

2.3.6 BIOGENIC STATUS OF THE MANGANESE BIOMORPHS

The trichospheres in the Yahmum jasper and some in other jaspers are probably biogenic but this suggestion is difficult to substantiate. Those with curved bifurcated filaments have a biogenic aspect but those with bladed and crystal forms definitely do not. Although the habit closely resembles that of Metallogenium the structure is different and no species of this exact shape has been found illustrated in the literature.

The filamentous and spherical biomorphs closely resemble some of the iron biomorphs and may even be of the same species hence they are probably biogenic and possibly represent both bacteria and fungi. The cellular filaments in the black limestone at Azlam ^{are} ~~is~~ almost certainly biogenic.

PART II GEOMICROBIOLOGY

Section 1	Introduction
Section 2	Red Sea Deeps
Section 3	Coastal Plain
Section 4	Shield
Section 5	Phytoclast Geothermometry

SECTION 4

GEOMICROBIOLOGY
IN THE SHIELD

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G E O M I C R O B I O L O G Y
I N T H E S H I E L D

2.4.1

SELECTED AREAS

INTRODUCTION

Although Brown and Jackson in 1960 reported the occurrence of limestones with stromatolitic structures in the younger Precambrian Fatima and Aba formations no publications yet exist on the discovery of microfossils in the Precambrian rocks of the Arabian shield. The writer discovered such microfossils in rocks from various parts of the shield in maceration residues but later found that Dr. V. Skiba and Dr. M. Muir of Imperial College together had discovered algal remains of Girvanella in a thin section of jasper from Wadi Sitaram area a few weeks before the writer's independent first discovery in 1970.

The rocks examined by the writer came from Wadi Wassat, Wadi Bidah, Mahawiyah, Jabal Samran and Wadi Sawawin. The samples from Wadi Sawawin were very kindly supplied by W. Liddicoat and Ahmed Al Shanti, the others were supplied by Barry Jackaman and Simon Rexworthy. The results reported by Jackaman (1971) and Rexworthy (1972) were based on the writer's work under the guidance of Dr. M. Muir and all the preparations for palynological study on these rocks, other than thin and polished sections, were made by the writer alone.

A marble from Jabal Samran ($22^{\circ}20'N$, $39^{\circ}33'E$) was treated in the same way as limestones from Jabal Dhaylan in an effort to establish maximum temperature to which the rocks had been subjected (described in section 2.5.1 on phytoclast geothermometry). The insoluble organic matter found in the Jabal Samran marble consists of non-descript mucilaginous tissue, short branched filamentous colonies very like those found in some Wadi Bidah rocks (e.g. fig. 196) and some small spheroidal bodies. One of these spheroidal bodies consists of two cells each of which contains a spot which possibly represents a nucleus. M. Muir (pers. com.) noted that this biomorph resembles Carvosphaeroides, a eukaryotic organism described by Schopf (1968). A jasper from Jasper Hill ($22^{\circ}18'N$, $39^{\circ}22'E$) also in the Jabal Samran district was also found to contain filaments and spheroidal

bodies, some with possible nuclei. A jasper from the Mahawiyah area (about $20^{\circ}18'N$, $41^{\circ}19'E$) studied by Abdul Razak Rashid (1971) also contains filamentous colonies and spheroids. Biomorphs from Wadi Wassat, Wadi Bidah and Wadi Sawawin were examined more closely and are described separately.

WADI WASSAT - WADI QATAN

The Wadi Wassat district ($18^{\circ}05'N$ to $18^{\circ}28'N$, $44^{\circ}05'$ to $44^{\circ}16'E$) described by Jackaman (1971) consists essentially of folded and metamorphosed basic andesitic and rhyodacitic lavas and volcanoclastic rocks more than 570 m.y. old, with massive pyrite lenses near the base of the Wassat Group. Similar rocks occur in the Wadi Qatan district in the southern part of the area.

Carbonaceous matter including algal and bacterial remains occurs in Jackaman's "Facies B" which consists of subaqueous calcareous tuffs and greywackes. It also occurs in "Facies D" which is calcareous volcanoclastic rocks with up to 2% carbonaceous matter, and in "Facies G" where it occurs as thin laminae in rhyodacitic tuffs. Similar carbonaceous matter occurs in Jackaman's "Facies B" of the Qatan Group where it comprises as much as 25% of andesitic tuff and is associated with pyrite.

Filamentous bacteria or algae or both, and spheroids are abundant in the andesitic pyroclastic rocks of the Wassat area. A filamentous meshwork found in thin section is shown in fig. 184. The filaments in this are poorly preserved and are somewhat altered and damaged however it must be noted that the host rock is of the lower greenschist facies of metamorphism. Dr. Muir (pers. com.) noted that these filaments closely resemble those found in parts of the Onverwacht Group. The meshwork in fig. 184 resembles others observed by the writer in Lake Superior type iron formations, and in general shape, but not in size, vaguely resembles the water-net green alga Hydrodictyon reticulatum. Poor preservation, optically contrasting mineralogy and the fine size of the minerals and biomorphs generally makes observation of the Wadi Wassat and Wadi Bidah microorganisms difficult or impossible in thin section.

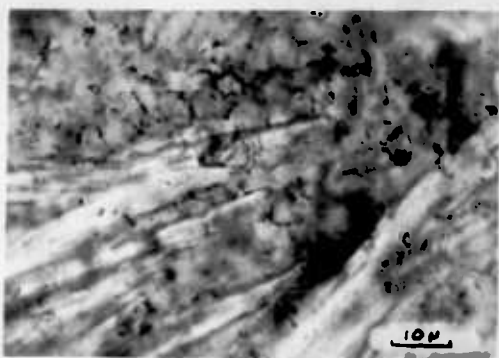


Fig. 184. TLM photograph of poorly preserved reticulated filamentous bacteria in a subaqueous pyroclastic rock from Wadi Wassat (TS in PPL).

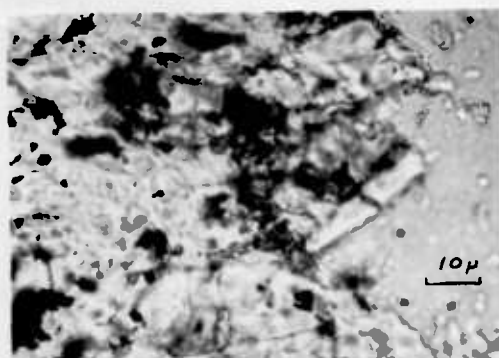


Fig. 185. TLM photograph of irregular mass of filaments in pyroclastic rock from Wadi Wassat (TS in PPL).

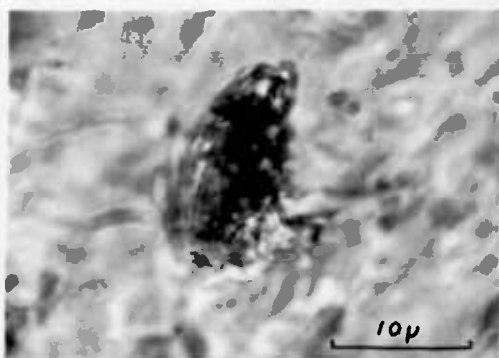


Fig. 186. TLM photograph of spheroidal filamentous colony in pyroclastic rock from Wadi Wassat (TS in PPL).

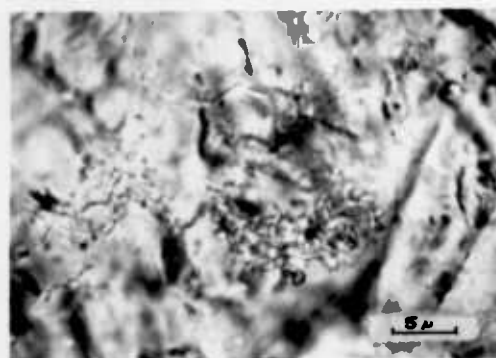


Fig. 187. TLM photograph of filamentous colony in pyroclastic rock from Wadi Wassat (TS in PPL).

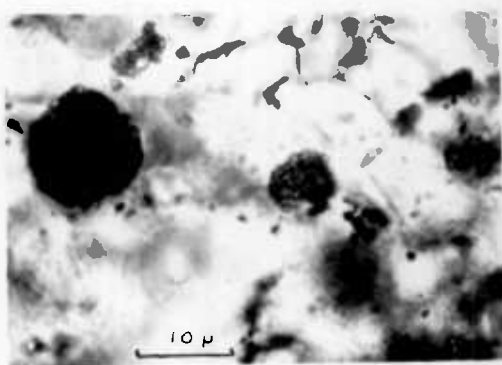


Fig. 188. TLM photograph of spheroidal filamentous colony in pyroclastic rock from Wadi Wassat (TS in PPL).



Fig. 189. TLM photograph of spore or colonial body in pyroclastic rock from Wadi Wassat (TS in PPL).

In addition to reticulated meshes, filaments also occur in connected irregular masses, e.g. fig. 185 and in tight sheaves similar to the iron bacteria colonies in Wheal Jane mine waterway shown in fig. 158. In general they resemble Gallionella (M. Muir pers. com.). The filaments also form ellipsoidal masses e.g. fig. 186 and fig. 187. In general, the filaments are thin, curved, short and bifurcated. Details could rarely be observed however some filaments appear to be fine tubes without segments, and others appear to be segmented. Some appear to consist of chains of ellipsoidal cells, and some filaments are probably helical e.g. fig. 188. Spheroidal bodies, possibly blue-green algal colonies, were also observed in thin sections e.g. figs. 188, 189. During thin section examination it was noted that the biomorphs generally appear to be independent of grain boundaries and commonly transect them.

When ore containing 80% pyrite was macerated abundant sporopollenin and some insoluble pyrite formed the residue. The residual pyrite is commonly spheroidal and probably occurs as framboids. Jackaman (1971) found framboids to be common in polished sections of the ore. Films of organic matter may have prevented acid attack on the pyrite. The residual sporopollenin occurs as fine degraded matter, non-descript mucilage containing cocci (fig. 190) and spheroids. This includes Botryococcus (fig. 191, M. Muir, pers. com), and other spore-like bodies which are possibly blue-green algal colonies (e.g. figs. 192, 193).

Some polished sections of pyrite ore selected by B. Jackaman contain biomorphic structures. These are rather vague and difficult to observe. They generally are outlined by tiny dots presumably representing inclusions of organic matter, or pores, and occur in the shape of nets, filaments, rings and spheres. Some ring structures resemble in size and shape "organized elements, type 2" found by Schidlowski (1965) in polished sections of detrital pyrite grains in Witwatersrand gold ore, about 2150 m.y. old (cf. figs. 194, 195). They also resemble biomorphs described by LaBerge (1967).

WADI BIDAHA

The Wadi Bidah district ($20^{\circ}20'$ to $20^{\circ}50'$ N, $40^{\circ}20'$ to $41^{\circ}27'$ E), described Jackaman (1971), consists of a sequence of folded metavolcaniclastic

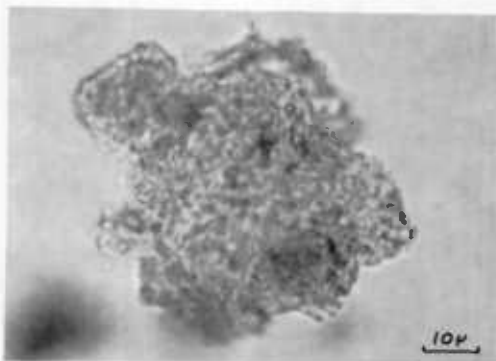


Fig. 190. TLM photograph of coccoid mucilage matter in maceration residue from pyritic ore from Wadi Wassat (SM in PPL).

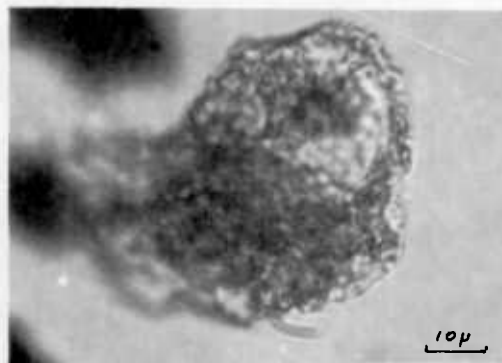


Fig. 191. TLM photograph of *Botryococcus* in maceration residue from pyritic ore from Wadi Wassat (SM in PPL).



Fig. 192. TLM photograph of spore or algal colony in maceration residue from Wadi Wassat pyritic ore (SM in PPL).

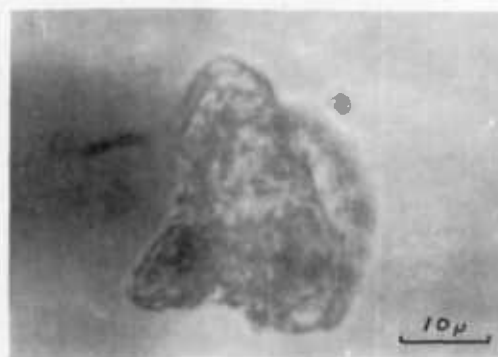


Fig. 193. TLM photograph of spore or algal colony in maceration residue from Wadi Wassat pyritic ore (SM in PPL).

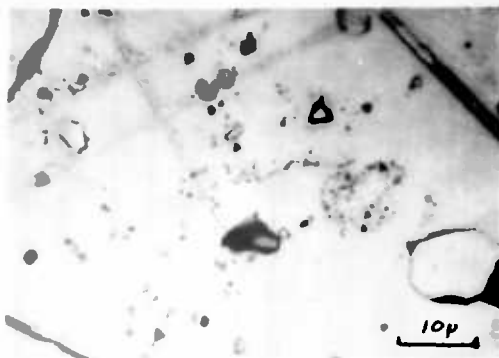


Fig. 194. RLM photograph of polished section of Wadi Wassat pyrite with filament and ring-type structures (central area, scratches in upper area; PS in PPL).

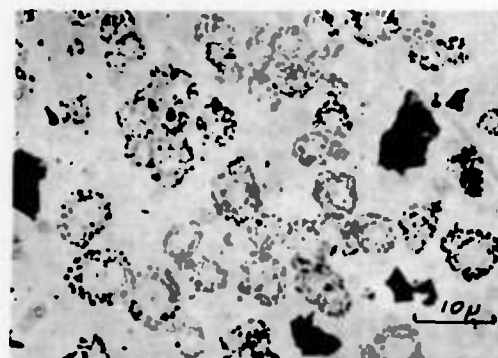


Fig. 195. RLM photograph of polished section of Witwatersrand (pyrite showing "organized elements, type 2". Copied from Schidlowski (1965; PS in PPL).

rocks and lavas of dacitic and andesitic composition having an age greater than 912 m.y. Stratabound massive sulphide lenses containing pyrite, chalcopyrite and sphalerite occur at Rabathan in a dolomitic, phosphatic, tuffaceous arenite and at Shaib el Tare and Gehab at the contact of dacitic volcanoclastic rocks and basic lavas and pyroclastic rocks.

According to Jackaman (1971) the carbonaceous calcareous volcanoclastic rocks enclosing the Rabathan mineralization contain up to 15% organic matter in places and most of the lithofacies in the district contain a minor proportion of carbonaceous matter. Hematite-cherts or jaspers, with 1 to 80% hematite, form the gangue to the Rabathan mineralization and occur elsewhere as marker horizons. Carbonaceous matter occurs throughout the jaspers and some areas of hematite are replaced by pyrite.

Dolomitic marble occurs in the footwall rocks in the Rabathan area. When macerated by the writer, this rock was found to contain abundant insoluble organic matter. Very typical of this rock and others in the area is the filamentous bacterial colony shown in fig. 196. This contains short irregular branching filaments 1 to 2 μ wide which in places appear to be septate. Also present in this rock are spore-like bodies resembling Huroniospora (fig. 197) and the brown alga Parka decipiens (fig. 198, M. Muir pers. com.).

The jasper (and pyritic chert) contains filaments and coccoid mucilaginous matter that can be seen in maceration residue smear mounts. Filaments, commonly subparallel to bedding, and some spheroidal bodies also were observed in thin sections by the writer. Commonly filaments occur at quartz boundaries and have chevron shapes (e.g. fig. 199). The writer has observed filaments with such shapes in other jaspers in which the filaments were unrelated to quartz grain boundaries hence it is feasible that the filaments governed the quartz growth rather than vice-versa. Elsewhere in the Wadi Bidah jaspers biomorphs occur within quartz grains or transecting grain boundaries. The filaments along the grain boundaries were not confused with irregular sheets of matter expelled during quartz growth (the distinction is sometimes difficult). Some of the jasper contains sheaves of fine filaments (fig. 200) which resemble the Wheal Jane filamentous iron bacteria. Spore-like bodies are associated with some of these filaments. Phantom structures of colourless material were also

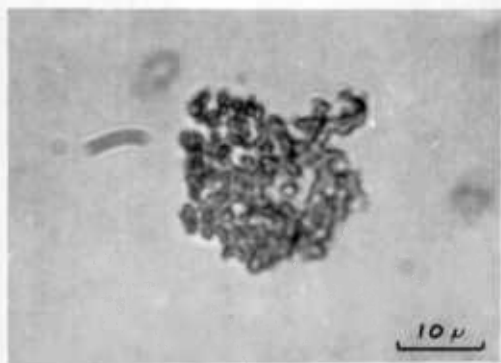


Fig. 196. TLM photograph of filamentous colony in maceration residue of Wadi Bidah limestone (SM in PPL).

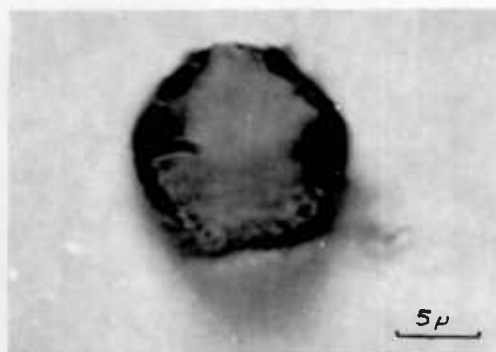


Fig. 197. TLM photograph of *Huronisporea* in maceration residue of Wadi Bidah limestone (BS in PPL).

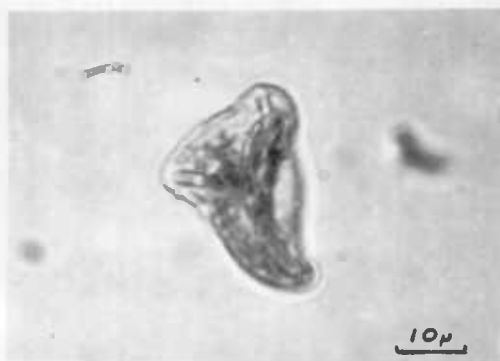


Fig. 198. TLM photograph of algal spore c.f. *Parka decipiens* in maceration residue of Wadi Bidah limestone (SM in PPL).

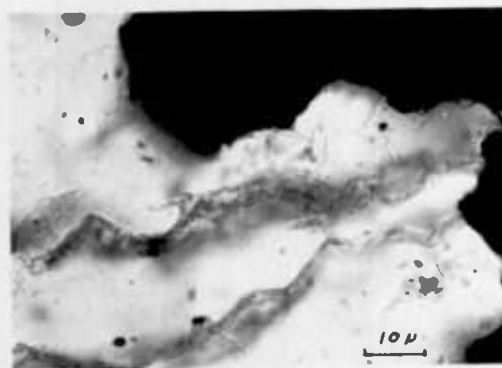


Fig. 199. TLM photograph of chevron filaments at quartz grain boundaries in Wadi Bidah jasper (TS in PPL).

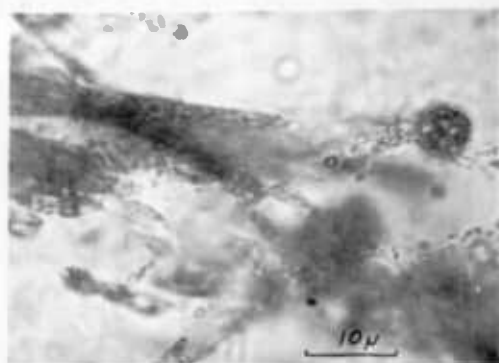


Fig. 200. TLM photograph of filaments and spore-like body in Wadi Bidah jasper (TS in PPL).

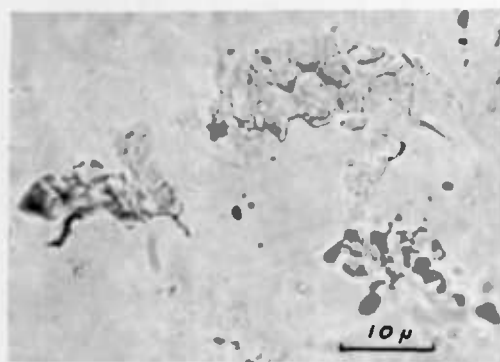


Fig. 201. TLM photograph of phantom structure within quartz grain in Wadi Bidah pyritic chert (TS in PPL).



Fig. 202. TLM photograph of algal colony in maceration residue of Wadi Bidah sulphide ore (SM in PPL).

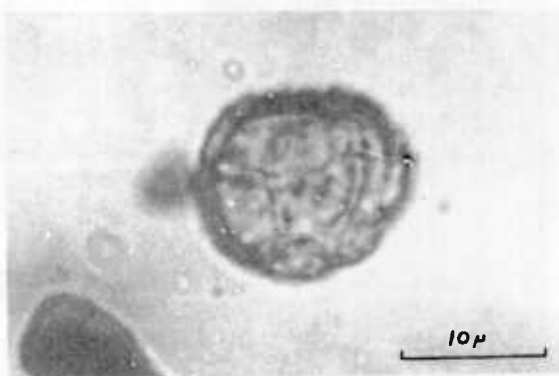


Fig. 203. TLM photograph of algal colony in maceration residue of Wadi Bidah sulphide ore (SM in PPL).



Fig. 204. TLM photograph of short thick filament in maceration residue of Wadi Bidah sulphide ore (SM in PPL).

found in jaspers from Wadi Bidah (fig. 201). These puzzling features are discussed further in section 2.5.2.

In thin sections and maceration residues of the Wadi Bidah volcanoclastic rocks, filaments, non-descript mucilage and spheroidal bodies were observed. The filaments are about 1μ diameter and are short, twisted and branched. True branching appears to occur in places and some of the filaments appear to have septae. Observation is difficult because of the fine size of the filaments and the various mineral grains.

Specimens from Rabathan were macerated and found to contain abundant filaments, spheroids and coccoid mucilage e.g. figs. 202, 203, 204. The rather complex body in fig. 202 is most probably a blue-green algal colony (M. Muir pers. com.). Filaments and mucilage were also found in maceration residues of the Shaib el Tare and Gehab ores.

WADI SAWAWIN

General Geology

The Wadi Sawawin area, at around about $27^{\circ}56'N$, $35^{\circ}53'E$, has been studied by the Japanese Geological Mission in Saudi Arabia as a potential source of iron ore. The following précis is derived from DGM reports "Mineral Resources Research, 1967-68" and "Mineral Resources Activities 1969-70".

The iron ore deposit is essentially a Lake Superior type Precambrian fine grained banded jaspilite composed of alternating layers of hematite and jasper. It occurs in the upper part of the Silasia Formation and is similar to the deposit at Wadi Karim in Egypt. The ore is folded and parts of the area are thermally and regionally metamorphosed. High grade ore consists of hematite 38%, martite 34%, quartz 25% and calcite 1%. Other minerals associated with the ore are magnetite, chlorite, sericite, garnet, epidote and pyrite. Reserves are 391 million tons of $>42\%$ Fe ore with a P_2O_5 content of 0.51 to 0.76%. The fine grain size of the ore has caused metallurgical problems.

Much of the area has been metamorphosed to greenschist facies and the Silasia Formation has been extensively intruded by subconcordant sheets of meta-diorite and by a complex ranging in composition from diorite to granite. The "Central granite" in the area is 550 to 660 m.y. old (from K-Ar and Rb-Sr methods respectively). The rocks in the

Silasia Formation other than jaspilite are sandstone, slate, tuff; tuffaceous sandstone, slate and breccia; meta-andesite and lesser conglomerate and limestone.

Purpose of Study

Microorganic remains have been found in Lake Superior type iron ores and associated rocks in North America and South Africa (Gruner 1922, Barghoorn and Tyler 1965, Schopf and Barghoorn, 1967, Cloud and Licari 1968) and many writers have suggested that this type of iron ore is biogenic. Dr. M. Muir showed the writer biomorphs in Australian Lake Superior type iron ores, and later the writer found such biomorphs in similar iron ore from the Scully Mine near Lake Superior which was studied by Jeff O'Leary (Imperial College). On this basis it seemed reasonable to question whether or not all Lake Superior type iron ores contained biomorphs and had a common biogenic origin. The late Dr. F.G. Percival very kindly made his extensive collection of iron ore specimens and thin sections from Africa, India and Australia available to the writer to pursue this matter. Biomorphs were found in thin sections of most of the relatively non-metamorphosed jaspers. This study however was made on a selection of random samples of typical ore and it was thought that more systematic sampling would be useful. The Wadi Sawawin deposit was well suited to this study as it extends the other work on the geomicrobiology of the Arabian shield as well as extending the geomicrobiological knowledge of world iron ores.

For this study the DGM provided 16 drill core specimens from depth 30 to 115 m in drill hole SW2, and 12 specimens from depth 2 to 66 m in drill hole SW3. These holes are situated in the western orebody in an area relatively unaltered by folding and metamorphism. The holes are 400 m apart and intersect the same "jaspilite" which occurs at 54.9 to 88.1 m depth in SW2 and 13.7 to 61.6 m depth in SW3.

Observations

The drill core samples essentially consist of 3 rock types namely tuff, jasper and opaque iron ore. Green tuff is the most common rock type and occurs in beds ranging in thickness from a millimetre to more than a decimetre. Graded bedding is visible in places. The tuff ranges in

composition from vitric tuff with shards completely replaced mostly by chlorite and hematite, to crystal tuff commonly with broken grains of feldspar, magnetite and lesser quartz, to lithic tuff. The mineral composition is feldspar, quartz, chlorite, calcite, siderite(?), epidote, clinozoisite(?), sericite, apatite, magnetite and hematite. The overall composition is probably andesitic. Three specimens logged as greywacke appear to be tuff although the distinction is difficult to make with this generally fine grained slightly metamorphosed sediment. One specimen contains vague shard outlines; however another specimen is medium grained and has a high proportion of feldspar fragments. The rock logged as slate also appears to be tuff but this tuff contains a relatively higher content of chlorite (which probably facilitated foliation development) and calcite (the introduction of which was probably facilitated by a weak foliation).

Magnetite is generally as common as or more common than hematite in the tuffs and some beds appear to contain as much iron as some of the jaspilite, and ought to be assayed together with the jaspilite (if this has not yet been done).

The massive ore is mostly hematite. Contacts with tuff and jasper are sharp or transitional and the main impurities in the ore are constituents of these two rocks. Some apparently massive iron-rich ore actually consists of hematitic tuff with fine hematite grains disseminated so closely that the rock appears opaque in all but the thinnest of sections. The presence of chert in similar circumstances is revealed by the overall hardness of the rock.

The iron in the jasper occurs essentially as hematite and only rarely as magnetite. The hematite occurs in granules ranging in size from less than 1μ to more than 10μ in diameter and more rarely as flakes or plates. Calcite and possibly siderite commonly occur as micro-veins and massive areas; chlorite and apatite are uncommon. At certain concentrations in thin sections hematite plates overlap and give an overall appearance of a cellular mass or a sea-weed like mass. Careful examination at high power dispels this illusion. Biomorphic structures nevertheless are abundant in the jasper and consist mostly of spheroidal bodies and less commonly of filaments.

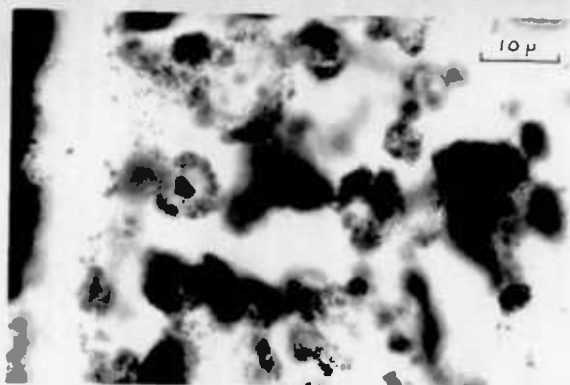


Fig. 205. TLM photograph of biomorphic shells, like type 2 of Schidlowski (1965) and type B of LaBerge (1967), in Wadi Sawawin jasper (TS in PPL).

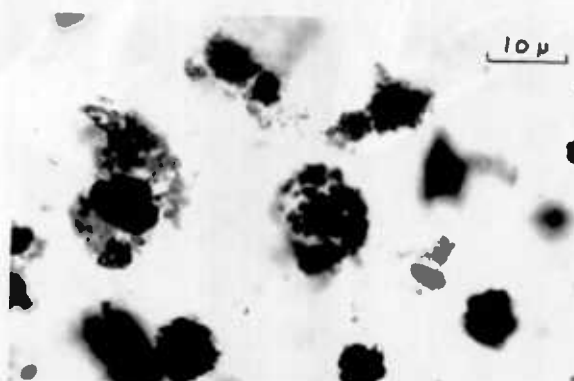


Fig. 206. TLM photograph of Wadi Sawawin jasper showing spheroidal biomorphs. One at the left has fine filaments like those in type A of LaBerge (1967; TS in PPL).



Fig. 207. TLM photograph of Wadi Sawawin jasper showing typical textures (TS in PPL).

The biomorphs shown in fig. 205 closely resemble the "organized elements type 2" of Schidlowski (1965) (see fig. 195). They also resemble "type B" of LaBerge (1967) although "type B" is twice as large. Other less distinct examples (fig. 209), however are the same size as "type B". "Type B" of LaBerge consists of probable and possible Eosphaera tyleri. This consists of two concentric shells separated by small satellite spheroids or ellipsoids. Depending on what parts of the structure are preserved, biomorphs consisting of a spheroid with a concentric halo, a broad halo or an annular series of ellipsoids, or a single large spheroid, may result. The biomorphs may consist of carbonaceous matter or, more commonly, extremely fine grained to medium grained hematite. Iron bacteria precipitate amorphous or goethitic iron oxide which is either assimilated within the cells or more usually is collected in and on the mucilaginous slime coating of the organisms; the actual mechanism is not fully understood (Harder 1919, Zajic 1969). The iron oxide presumably dehydrates to hematite during diagenesis or low grade metamorphism.

In fig. 206, left of centre, is a hematite grain surrounded by what appears to be a fragment of a sphere. In this outer part are fine filaments like those occurring in LaBerge's "type A" spheroids but his "type A" spheroids do not have solid hematite cores. This body and others like it resemble Schidlowski's organized elements type 1 (1965). It appears that some process has mobilized or removed some of the iron from these bodies. This is even more evident in fig. 207 which is perpendicular to the bedding. The cause of the iron migration is unknown but appears to be post-depositional and therefore due either to diagenesis, low grade metamorphism or weathering. As the specimen came from 66 metres depth weathering is somewhat unlikely. Furthermore Spencer and Percival (1952) found that acid leaching of similar rock removed the cores or nucleii of the spheroidal structures but left the haloes intact. In fig. 207 is an opaque area of concentrated hematite in the lower left; above this is an area of disseminated hematite grains (enlargement shown in fig. 208) and in the lower right is an area of nucleated spheres which shall be referred to as "Percival texture" (enlargement shown in fig. 209). These three areas have sharp boundaries in this thin section but in others the areas merge into each other. The Percival texture, which appears to consist essentially of LaBerge type structures, is almost ubiquitous in jaspilites throughout the world (Spencer and Percival 1952,

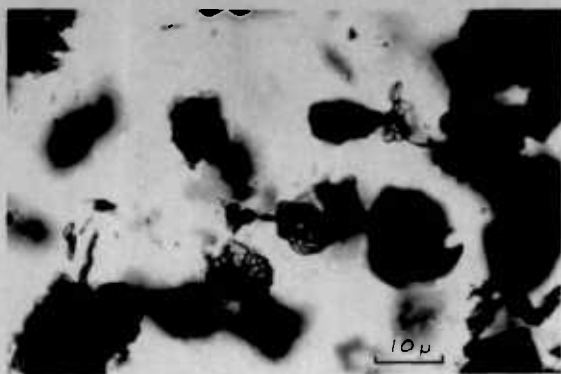


Fig. 208. Enlargement of area in upper left of fig. 207 showing hematite grains.

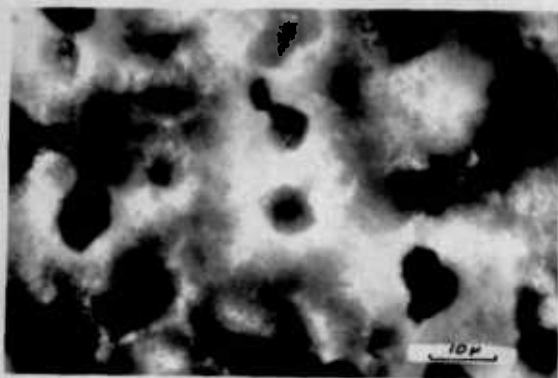


Fig. 209. Enlargement of area in lower right of fig. 207 showing shell structures.

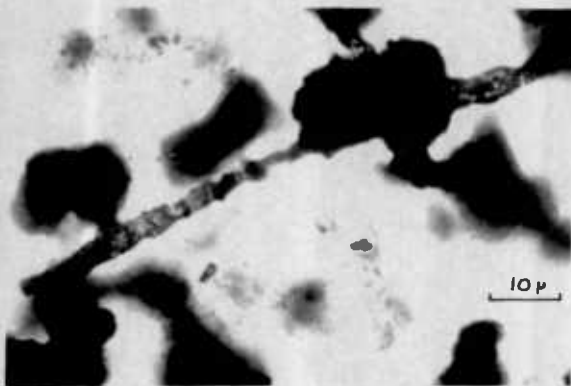


Fig. 210. TLM photograph of Wadi Sawawin jasper showing typical filaments (TS in PPL).

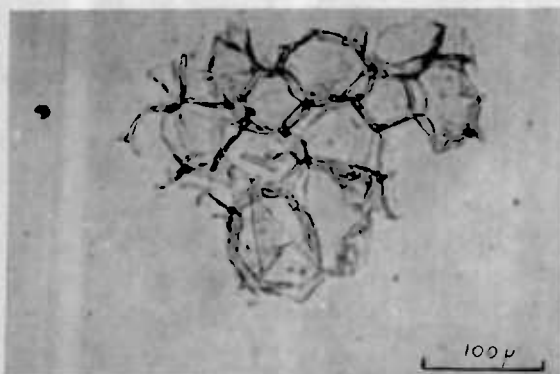


Fig. 211. TLM photograph of maceration residue of a Wadi Sawawin tuff, showing cellular body (SM in PPL).

Percival 1967, LaBerge 1967, and the writer's observations). Spencer and Percival (1952) found the spheroidal bodies consist of spherulitic quartz and are associated with polygonal syneresis(?) cracks in some non-metamorphosed jaspilites and concluded that the structures are colloidal in origin. LaBerge (1967) referred to their work but showed that several other writers considered the structure to be biogenic. No syneresis cracks were observed in the Wadi Sawawin jaspilite and the quartz is granular (in the terminology of Lovering, 1972). It is possible that the biomorphs could have been associated with colloidal silica and subsequent colloidal structures without themselves being caused directly by colloidal phenomena. Further evidence on the biogenic structures will soon be published in "Economic Geology" by LaBerge (pers. com.). Other spheroidal bodies having peripherally distributed hematite similar to LaBerge's types C and D are also very common in the Wadi Sawawin jasper. The filaments are of two sizes in the Wadi Sawawin jaspilite. Those in "type A" LaBerge structures are small ($0.5\ \mu$ wide and thinner, which is also the size given by LaBerge). Other less common filaments are 2 to $5\ \mu$ wide and rather straight. These appear to be septate and terminate as open tubes (e.g. fig. 210).

When macerated some specimens yielded very little organic residue and others yielded more typical quantities (perhaps 1000 ppm of the material macerated). The most common material is typical non-descript coccoid mucilaginous matter but segmented filaments, spheroids, and excellent cellular structures were obtained from the jasper and from the tuffaceous rocks. Fig. 211 is cellular matter from a hematite tuff and is very similar to the cellular matter found in jasper specimens. Other organic biomorphs in the tuffs are less similar to those in the jaspers.

2.4.2

SIGNIFICANCE OF BIOMORPHS IN THE SHIELD

Many of the biomorphs described in this section so closely resemble well known fossil microorganisms that their biogenic status can hardly be doubted. This study establishes that fossil microorganisms are widespread in the sedimentary formations of the Arabian shield even though they were previously unknown (except for Dr. Skibá's independent discovery), and it substantiates the claim that contrary to older established opinion, microorganisms were widespread and abundant in the Precambrian world.

Another important discovery is that fossil microorganisms occur in rocks metamorphosed to lower greenschist facies. It might be expected that microorganisms would be obliterated or decomposed by such metamorphism and there are very few references to discoveries of fossil microorganisms in metamorphosed Precambrian rocks. The knowledge that a rock has been metamorphosed actually deters some geologists from looking for microfossils in it (e.g. Barghoorn 1971). It should be noted however that biomorphs have been found in weakly metamorphosed rocks that have been previously described as non-metamorphosed (e.g. the fossiliferous Onverwacht Chert, commonly described as a sedimentary rock e.g. Nagy 1971, is actually metamorphosed, Dr. M. Muir, pers. com.). The writer (Hamilton 1967; Hamilton, Ramsden and Stephens 1970) has found well-preserved plant fossils metamorphosed to graphite in rocks of the hornblende-hornfels facies of metamorphism. Metamorphism however generally reduces the quantity of organic matter and alters its composition. This latter fact can be used for studying geothermometry. From the nature of the preservation of some of the organic matter in the shield it appears that some of the areas designated as lower greenschist facies are actually of the zeolite facies of metamorphism.

A further significant discovery is that some of the pyroclastic rocks of the shield contain relatively abundant organic matter. Although many people consider volcanic environments to be hostile to life, subaqueous volcanic environments provide warmth and nutrients that support prolific life (the writer has observed abundant subaqueous plant life around

volcanoes in Italy, Greece, New Zealand, New Guinea and Indonesia).

There is wide diversity in the fossil microorganisms of the shield rocks and both procaryotic and eucaryotic forms have been found. This suggests that biomorphs may be useful for correlation purposes and possibly age determination.

Jackaman (1971) concluded that the mineralization at Wadi Wassat and Wadi Bidah formed from the same geological processes, and that the main control on ore composition was magmatic rather than environmental. The great differences in ore compositions however do not appear to be reflected in the associated volcanic rocks, and the possibility that the differences are due to environmental factors is worth further consideration. Iron bacteria can precipitate iron oxides (or oxyhydroxides) from even very dilute solutions. Relatively pure siliceous iron oxide can result from such activity and facies of iron sulphide may occur in the same formation. Stanton (1972) noted that such iron sulphide facies differ from the oxide facies in that the iron sulphide is associated with carbonaceous matter rather than chert. Metals such as copper and zinc can be concentrated by microorganisms by means of biogenic alteration of Eh, pH conditions, or by precipitation with biogenic sulphide ions, or by incorporation into the microorganisms, or by adsorption onto organic matter (usually dead). Wadi Bidah mineralization, in addition to having copper and zinc concentrations far beyond those at Wadi Wassat, appears to have more carbonaceous and phosphatic matter associated with it. Also the microorganisms at Wadi Bidah appear to be more varied and some appear to be of a higher order (e.g. the eucaryotic Parka decipiens) than the microorganisms at Wadi Wassat.

Although the Wadi Sawawin deposit contains some well-preserved biogenic remains it does not necessarily follow that the whole deposit is biogenic. Nevertheless the close similarities of the less spectacular biomorphs, inter alia, with those in other Lake Superior type iron deposits suggests the deposit probably had a similar origin. Furthermore these less spectacular biomorphs (LaBerge types) are very prolific where observable and appear to account for much of the iron in the specimens studied. The deposition of the ore therefore was probably biogenic but the source of the ore may have been volcanic e.g. from distant volcanic rocks which may have had the iron leached from them.

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	Section 2	Red Sea Deep
PART II GEOMICROBIOLOGY	Section 3	Coastal Plain
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	Section 5	Phytoclast Geothermometry

SECTION 5

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SECTION 5

PHYTOCLAST
GEOTHERMOMETRY

2.5.1

INTRODUCTION

The irreversible effects of heat on carbonaceous matter are well known and have been used by coal geologists in rank valuation studies and by petroleum geologists in establishing limits for potential petroliferous areas. The effects have also been used in other coal studies such as problems of natural coke occurrences and less commonly in establishing temperatures of igneous intrusions (Hamilton, 1968). There are few references to other applications of this type of geothermometry and none that the writer could find on its application to ore deposits. Phytoclast geothermometry has great potential in ore deposit studies. The lack of its application so far is attributed to the unfamiliarity of ore geologists with specific methods of the coal and petroleum geologists and to the lack of realization that it can be used on very low concentrations of organic matter.

Phytoclasts are coalified fragments of plant matter, including spores and pollen (Bostick, 1970). The term "phytoclast" is used in preference to "kerogen" which is restricted to material in oil shales (A.G.I. 1960). It is also more concise than "coaly matter" and does not imply a restriction to the study of coals.

Phytoclasts are widely distributed in sedimentary and meta-sedimentary rocks. When present even in very small quantities, e.g. less than 1,000 ppm, they can be concentrated and isolated. Techniques for this are given in several of the references, the most detailed being those by Saxby (1970) and Gutjahr (1966). The information obtainable from dispersed phytoclasts is comparable with that obtainable from massive coal.

Phytoclasts and massive coals are very sensitive to heat which produces irreversible and continuous changes in them. These effects together with their widespread occurrence and their chemical independence of silicate and carbonate systems make phytoclasts very suitable for geothermometry. Phytoclast geothermometry also is applicable over a wide range of temperatures. For conditions under which heating was rapid, a close

estimate of temperature may be obtained but for conditions of extremely prolonged heating temperature estimates appear to be much less reliable.

Using measurements of the reflectance of vitrinite, Taylor (in Hamilton, Helby and Taylor, 1970) was able to show that phytoclasts in a volcanic breccia were never heated to a temperature much more than 100°C. This together with palynological dating of the phytoclasts was sufficient to show that the breccia is indeed a volcanic deposit and not an intrusive diatreme as was previously believed (Wilshire, 1961).

Phytoclast geothermometry should be applicable to epigenetic deposits in sedimentary rocks. If hot fluids accompanied the mineralization then a halo of thermal alteration should surround the deposit.

2.5.2

EFFECTS OF HEAT ON PHYTOCLASTS

With metamorphism, plant material goes through a series of changes known as rank progression, and terminates in fully ordered crystalline graphite. The main chemical effects of heating are loss of volatile matter and increase of carbon content with progressive molecular reorganization as illustrated diagrammatically in fig. 212.

Effects produced by relatively rapid heating can be duplicated in the laboratory and for these conditions close estimates of temperatures can be obtained. Such studies, on temperatures of intrusions into coal seams have been made by McFarlane (1929), Dapples (1939), Clegg (1955), Taylor (1961), Brown and Taylor (1961), Kisch and Taylor (1966), Kisch (1966) and Gosh (1967).

The effects of rapid heating depend on the composition of the phytoclastic matter especially its rank. In the absence of oxygen, plant matter and coals below the rank of bituminous form an amorphous isotropic char when heated rapidly. With slower heating rates, effects similar to rank progression are found. Bituminous coal forms coke beyond 300°C with rapid heating. Waxes and resins melt and disappear by 200°C and tarry substances melt at 200°C and become fluid beyond 300°C. Vitrinite undergoes great changes in structure and optical properties (Taylor 1961, Brooks and Taylor 1966). Anthracite and higher rank coals undergo changes similar to rank progression.

The effects of rapid heating under pressure results in coals having anomalous compositions and optical properties (van Krevelin, 1963; Brown and Taylor, 1961; Chandra, 1965). Chandra (1965) found that the measurement of anisotropy of vitrinite can be used for evaluating the pressure effect.

Coals heated at geological rates, giving normal rank progression, are only slightly affected by static pressure (M. and R. Teichmuller, 1968). The effects of shear stress are not well known but are likely to be far

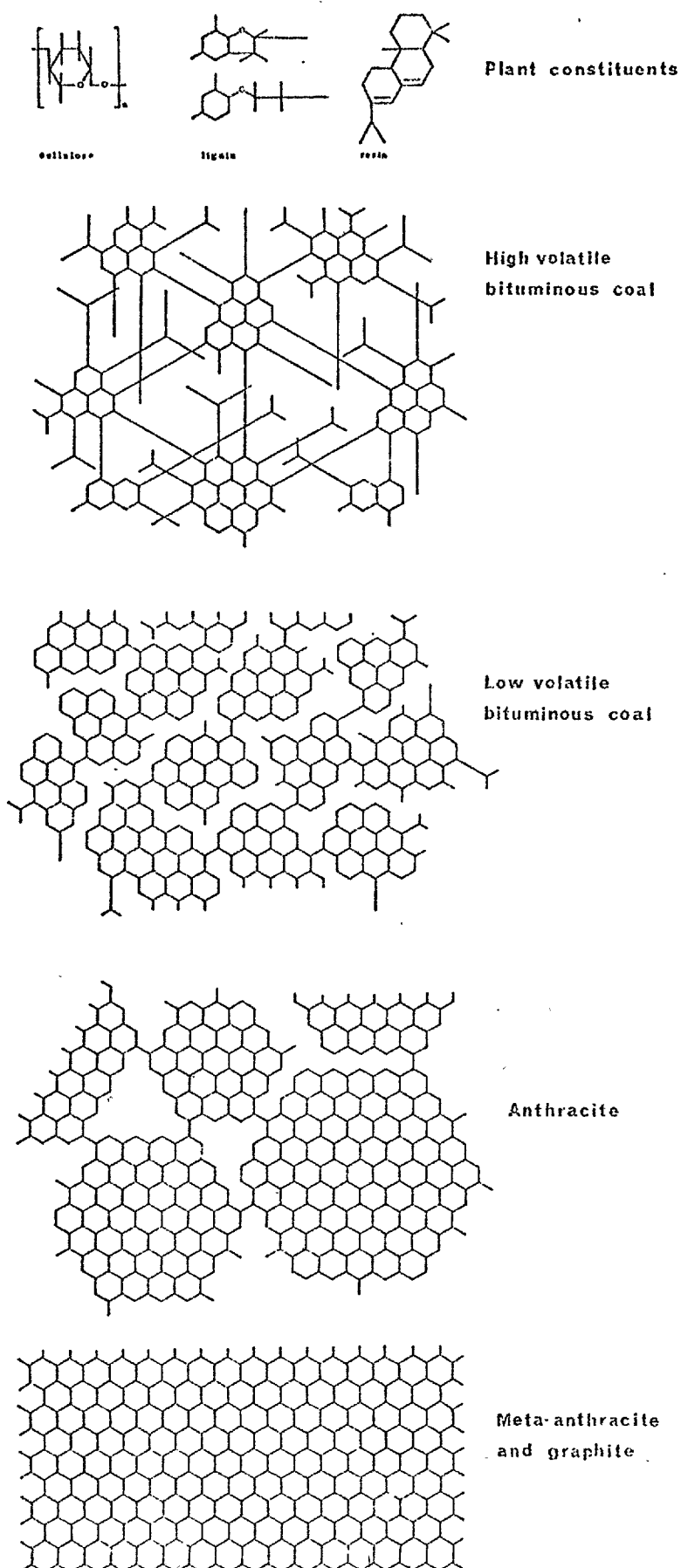


Fig. 212. Molecular framework diagrams of coaly matter ranging from plant material to graphite (using data from van Krevelen 1963 and M & R Teichmüller 1968).

greater than those of static pressure (Taylor, 1971). Coals heated at these geological rates are much less suitable for geothermometry than rapidly heated coals because the effect of geological time is not so well known.

Karweil (1965) has used chemical kinetic theory to show that time can have a profound effect on coalification (fig. 213).

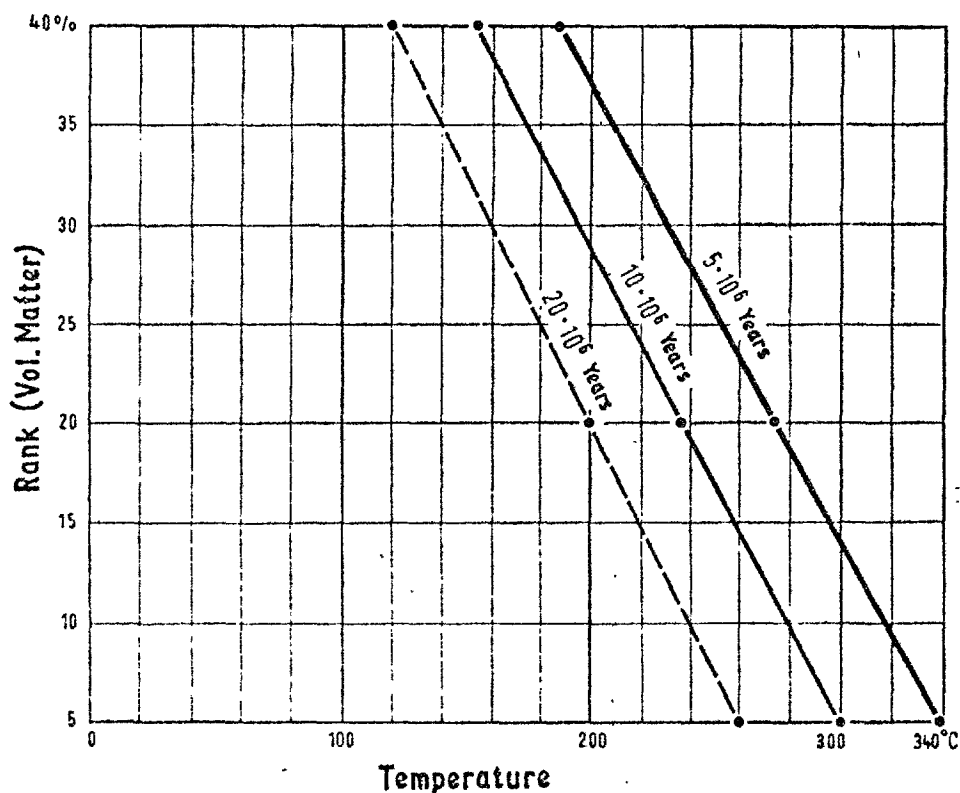


Fig. 213. Relation of time, rank and temperature according to Karweil, (Teichmüller, M. & R., 1968).

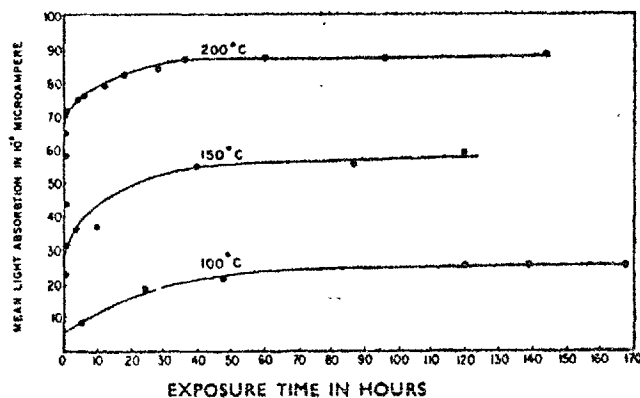


Fig. 214. Relation of time, rank and temperature from Gutjahr (1966).

Gutjahr (1966) however found that pollen grains reached almost full alteration after heating for as little as 100 hours (fig. 214). His results indicate that the rate of rank increase tends to zero with increasing time. Changes beyond a million years therefore ought to be negligible. Temperatures in the Crust would most probably not remain constant over millions of years. It is only the maximum temperatures that contribute to rank progression and these must vary with epeirogenesis, exhumation and geothermal gradient fluctuation.

Further evidence on the slight effect of time comes from the occurrence at several localities of only slightly altered phytoclasts from Precambrian rocks (Saxby, 1970; Barghoorn and Tyler, 1965; and observations by the writer).

Temperatures related to coal ranks can be estimated from laboratory studies on coal, and from comparison of coal ranks with rock metamorphism (see Kisch, 1968; M. & R. Teichmuller, 1968; Landis, 1971; French, 1964; and Castano and Sparks, 1970). Table 8 is a compilation of published work for obtaining rough estimates of temperatures from the properties of phytoclasts using metamorphic mineral reactions as a basis. This table must be used with care because it tends to gloss over the effects of composition and pressure. Also coal reactions can never fully be equated with mineral reactions. By comparing Table 8 with fig. 213 it can be seen that the figures obtained by Karweil (1956) on theoretical grounds of coal chemistry are probably in harmony with those obtained on the basis of associated mineral reactions for heating periods of less than 5 million years. Nevertheless it must be noted that M & R Teichmuller (1968) claimed that given sufficient time, anthracite rank theoretically can be reached at temperatures as low as 200°C. This is only half the temperature usually given for the formation of pyrophyllite from kaolinite plus quartz. Castano and Sparks (1970) on the other hand found coal of a rank bordering sub-bituminous and high volatile bituminous at a depth of 10,500 ft and a temperature of 230°C.

Estimates of maximum temperatures can be obtained from heating experiments because the effects of heat are irreversible. Heating for 200 hours, for example, will not affect the phytoclasts if the temperature is below a past maximum. Once the phytoclasts are heated to a temperature above a past maximum new changes can be observed.

Rank	Volatile % Content	Carbon % Content	Pollen opacity (1) %	Pollen colour (1)	Vitrinite reflectance %	Xray diffraction					Metamorphic facies	Characteristic mineral reaction	Possible °C temperature	
						"Peak" position			Shape factor					
						(2)	(3)	(4)	(3)	(4)				
Peat	50-75	10	70	yellow	0.6			4.1		0.5				
Lignite	40-50	60	70	yellow	0.6	3.6-3.8	4.1-4.4	3.95	0.25-0.50	0.1				
Sub-bituminous			70	yellow	0.6									
High volatile bituminous	31-40	60-69	70	yellow	0.6-1.0			3.8-3.9	3.92	0.25-0.50	0.2	zeolite	analcite + quartz → albite	200
Medium volatile bituminous	22-31	69-78	70-84	brown	1.0-1.2	3.5								
Low volatile bituminous	14-22	78-86	84-92	dark brown	1.3-1.7		3.5-3.6	3.56	0.50	0.6				300
Semi- anthracite	8-14	86-92	100	black	1.7- 3				0.75-1.0					
Anthracite	2-8	92-98	100	black	3-5					1.25	greenschist	kaolinite + quartz → pyrophyllite	400	
Meta- anthracite	0-2	98-100	100	black	5-6	3.37		3.51		1.8				
Graphite	0	100	100	black	6-17		3.35-3.36	3.36	30	25	amphibolite	chloritoid + andalusite → staurolite+quartz	500	

Table 8. Tentative relations of rank, rock metamorphism and temperature based on information from papers by Gutjahr (1966), Winkler (1967), McCartney and Ergun (1967), Griffin (1967), Kish (1968), Castano and Sparks (1970), and Landis (1971). Note (1) these values from Gutjahr (1966) refer only to specific conditions. Notes (2), (3) and (4), these values are from Griffin (1967), Landis (1971) and this paper (fig. 4) respectively. The shape factor is the height of the peak above background divided by the width of the peak at the half height level.

This type of geothermometry is not restricted to ranks below that of graphite. Differential thermal analysis, x-ray diffraction and electron diffraction successfully differentiated between a graphite formed at hornblende-hornfels facies and another formed at granulite facies (Hamilton, Ramsden and Stephens, 1970). According to Winkler (1967) the hornblende-hornfels facies ranges from 520°C to 640°C and the granulite facies ranges upward from 650°C. Both of these graphites are true graphites. Disordered graphites have been similarly studied by French (1964) and Landis (1971).

2.5.3

TECHNIQUES

CHEMICAL techniques of coal rank determination are well established. Volatile contents and fixed carbon determinations have long been routine methods of rank determination. Oxygen and hydrogen determinations are also commonly used for rank determination. Various sophisticated techniques are now available for the study of very small quantities of organic matter (see Brooks, 1971).

REFLECTIVITY is a well established technique for the determination of coal rank and thermal alteration (Chandra, 1965; Murchison, 1964; Francis, 1961; Taylor, 1961; Bennett and Taylor, 1970; McCartney and Ergun, 1967; Brown and Taylor, 1961). This technique has now reached the stage that the percentage reflectivity of polished vitrinite surfaces one micron in diameter can be measured to two decimal places. Reflectivity can be applied to spores as well as vitrinite but less exactly (Alpern, 1967). Refractive index can also be used but it is not as widely applicable as reflectivity (Forman, 1963).

COLOUR AND LIGHT ABSORPTION can be estimated by use of the Munsell colour system and by photometric measurements. Results can be compared with standards of pre-heated spores. Spores and pollen progressively darken as they are heated until they are ultimately opaque (Gutjahr, 1966).

X-RAY DIFFRACTION patterns of coals are rank dependent (French, 1964; Kisch, 1966; Griffin, 1967; Rutland, 1968; and Landis, 1971). Those of low rank have a very diffuse pattern with a very low wide peak poorly differentiated from the general background. As the rank increases the pattern tends towards that of graphite as illustrated in fig. 215. The peak obtained for peat is anomalous. It should be noted however that peat, unlike other coals, will not withstand the nitric acid treatment in Saxby's (1970) maceration technique. Resins give even better peaks than peat but the centre of these peaks is at 5.7 to 6.2 Å (Fronzel, 1967). When the peak centre corresponds to a spacing of more than about 3.95 Å the substance studied cannot be compared directly with coals although such substances most probably behave in an analogous fashion.

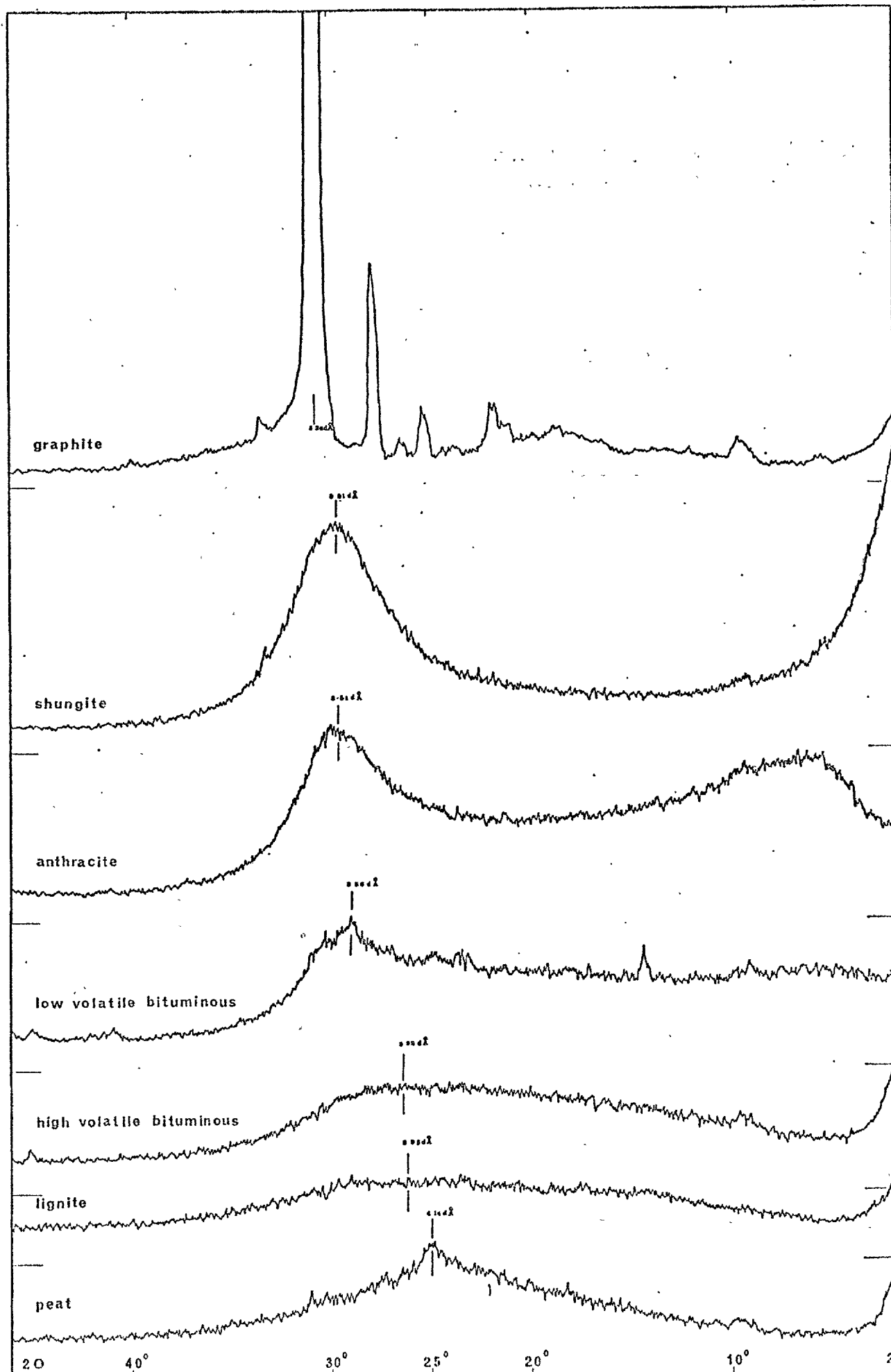


Fig. 215. X-ray diffraction patterns of coals of various ranks (obtained using cavity mounts).

Spores and pollen too behave in a slightly different way to coal. The $d\lambda$ spacing however is reduced and the diffraction peak is enhanced with heating. The X-ray diffraction patterns of heated spores examined by the writer formed a stepped rather than continuous series. The peak positions are at higher $d\lambda$ spacings than found with coals of similar ranks. From this it can be seen that to use the X-ray diffraction technique both the position of the peak as well as the peak shape (given by peak height divided by peak width at half peak height) need to be noted to avoid confusion with non-coaly matter. That is, the diffraction pattern depends on the type of phytoclastic material as well as its metamorphism, except perhaps with high ranks. It can be seen from table 8 that the results of different workers are not in full agreement. This probably reflects details of techniques as the X-ray diffraction technique depends very much on sample preparation and presentation. Each worker needs his own set of standards.

ELECTRON DIFFRACTION can be used on even smaller quantities than X-ray diffraction although it is not widely used on phytoclastic matter. Ramsden (in Hamilton, Ramsden and Stephens, 1970) was able to show by selected-area electron diffraction that the size of single crystal units in graphite flakes from Ceylon and N.S.W. Australia is related to the degree of order of the graphite. The degree of order is in turn related to the metamorphic environment in which the graphite developed. The technique when applied to coals requires special sample preparation. Details are given by McCartney and Ergun (1967).

DIFFERENTIAL THERMAL ANALYSIS was used by Stephens (in Hamilton, Ramsden and Stephens 1970) for differentiating graphite of different degrees of order. Presumably this can be used on lower rank material as well. Swaine (1969) has used this technique for discriminating between different types of carbonaceous matter as well as for determining the presence of small amounts of carbonaceous matter.

AUTOFLUORESCENCE of spores and pollen irradiated with ultraviolet rays is related to the physical-chemistry of the spores and pollen and is therefore related to its thermal history (van Gizel, 1971). This is a relatively new technique.

CATHODOLUMINESCENCE of spores, bombarded by a beam of electrons, is an effect somewhat analogous to UV autofluorescence (Muir and Grant, 1971). The technique, however, is still in the development stage. An advantage is that it can be used in conjunction with a scanning electron microscope. This means that not only are very small quantities necessary but the state of preservation of the material also can be clearly observed during the examination. The state of preservation of the spores partially depends on temperature (Correia, 1971) and its study can be considered a supplementary technique.

SECTION 5

PHYTOCLAST
GEOTHERMOMETRY

Contd.

2.5.4

APPLICATION TO JABAL DHAYLAN AREA

The first attempt at applying phytoclast geothermometry to ore deposits was made at Jabal Dhaylan. Outcrop samples were collected at one metre intervals across a mineralized area and its barren host rock in a line perpendicular to the trend of the mineralization. The samples were kindly examined by Dr. M. Muir at Imperial College who reported encouraging results but was not able to use these results to suggest the most suitable sampling interval. Thereafter many samples were collected at 2 m or 5 m intervals from similar lines crossing mineralization trends.

The phytoclasts were separated from samples by maceration techniques and were examined as smear mounts. No correlation was found between phytoclast colour and ore occurrence. This is probably due to the scarcity of phytoclasts, to weathering and to the nature of the mineralization.

The concentration of phytoclasts in these rocks is lower than would be expected with limestones (that recovered possibly represents 100 ppm). Because of this, contamination became a serious problem especially as the phytoclasts are generally only slightly altered and vary considerably in type. Beside laboratory contamination there is a possibility that some contamination came from modern dust that worked its way down into the porous craggy rock. The phytoclasts probably came from heterogeneous sources - from coral polyps, from floating seaweed fragments and from air-borne material. Much of the material is very fine ($< 10\mu$ diameter) so could have travelled a great distance. It is also difficult to observe the colour of this material because of interference effects. Grass pollen is common in the genuine phytoclasts.

Weathering has been intense at Jabal Dhaylan and is of the worst kind for phytoclast preservation. The occurrence of plattnerite indicates extreme oxidation under alkaline conditions (Takahashi 1960). This has probably removed much of the phytoclast material and may have caused some spurious colour changes (thinning reduces colour, oxidation intensifies it).

Leythaeuser (1973) noted that weathering can reduce the organic content of shales to a quarter of their former value.

The nature of the mineralization did not facilitate the investigation. Instead of a single large vein bearing the mineralization it tends to occur in a swarm, or weak stockwork, of small veins and heating would not be localized around the individual veins but spread out.

Dr. M. Muir kindly examined some of the smear mounts and identified alnus pollen, conifer tracheids, algal fragments, resin, wax bodies, non-descript coccoid mucilage and minute droplets of probable natural oil. Hydrocarbon materials are not obvious in the field at Jabal Dhaylan as they are in many other areas of Mississippi Valley type mineralization (e.g. Brown, 1967). The phytoclasts recovered range from colourless to yellow to brown and typically are yellow to pale brown. Those from the ore zones are also typically yellow to pale brown but very few were recovered from the ore zones possibly due to more intense weathering in these zones.

Although the results for Jabal Dhaylan are not useful in delineating thermal aureoles they do nevertheless indicate a low temperature of mineralization, certainly well below 200°C otherwise the pollen would have been darker and the wax and resin would be unrecognisable if still present.

2.5.5

APPLICATION TO OTHER AREAS

Phytoclasts in rocks of the greenschist facies of metamorphism should have a rank about equal to that of anthracite and should be dark brown to black. It was interesting therefore to find phytoclasts of light brown colour in Precambrian rocks of the Arabian shield designated as greenschist facies. It is most likely that these rocks have not been heated to the temperatures calculated for the greenschist facies of metamorphism. It is also possible that greenschist rocks can form at a lower temperature than previously assumed. Rexworthy (1972) gave the writer specimens from the Jabal Samran area to study and used the results to estimate the metamorphic gradient across that area.

During examination of Jackaman's (1971) thin sections of rocks from Wadi Wassat and Wadi Bidah the writer occasionally observed colourless biomorphs. There is no reference to such phantoms in the literature however recently Dr. S. Sengapta (pers. com.) of Imperial College found from laboratory experiments that they can form under certain conditions of heating under pressure. Dr. M. Muir (pers. com.) suggests they possibly may also be due to weathering.

As Jabal Dhaylan was found to be unsuitable for studying thermal aureoles using phytoclasts a brief attempt was made at applying the method to the Irish Carboniferous lead-zinc field. To the writer's surprise all the phytoclasts recovered from ore zones, wall rocks and distant country rocks are black, indicating a relatively high temperature. X-ray diffraction examination (by the writer) and reflectivity (by E. Scott of Newcastle University) supported the evidence of the visual examination. The phytoclasts are of about anthracite rank and therefore have been heated to at least 200°C and probably to about 400°C. This is higher than the temperatures tentatively indicated by fluid inclusion study (Morrissey, Davis and Steed, 1971) hence it appears that these deposits have been mildly metamorphosed.

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SECTION I

GENESIS OF INDIVIDUAL DEPOSITS IN THE RED SEA REGION

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3.1.1

DEPOSITS OF THE RED SEA DEEPS

INTRODUCTION

The sulphide deposits in the Red Sea deeps have caused widespread interest amongst economic geologists because of their extremely young age. Consequently a "Red Sea model" has been used in accounts of the genesis of many ancient sulphide deposits. The Red Sea deposits however do not offer a glib panacea to the knotty problems of genesis of stratabound sulphide deposits. Stratiform sulphide deposits of volcanic association generally occur in shallow water sequences and are associated with abundant carbonaceous matter (Stanton, 1972). There are actually very few deposits that formed in environments like those of the Red Sea deeps.

Despite the enormous international research on the Red Sea deposits (e.g. Degens and Ross, 1969) their genesis is still problematical. The sulphide facies furthermore, although the most valuable, occupies a small volume compared with the associated iron oxide facies. The deposits most like the Red Sea deeps deposits are probably the pyritic copper lenses in Cyprus and similar deposits discussed by Sillitoe (1972). James (1969) noted similarities between the Red Sea deposits and Lake Superior type iron ores but concluded that they actually bear little resemblance to any known base metal deposit. Since then the En Kafala manganese deposit has been studied by Bonatti et al (1972). This too has some similarities with the Red Sea deposits but lacks a sulphide facies.

The following discussion concerns essentially the sediments of Atlantis II Deep. Other sediments and associated brines in Red Sea deeps differ markedly from those in Atlantis II Deep but these are less well known and of less economic interest at present. It should be noted however that all writers (e.g. Degens and Ross 1969) agree that the deposits in Discovery Deep resulted from brines spilled over from Atlantis II Deep. Furthermore Gypsum Deep, not surprisingly, contains abundant gypsum, Kebrit Deep nearby contains abundant H_2S , and Chain Deep sediments have a high manganese content (Backer and Schoell, 1972).

SOURCES

Water

There seems to be little doubt that the source of the water in the brine is subterranean seawater. Craig's work (1969) on deuterium and oxygen isotopes has excluded volcanic emanations, connate water and fossil brines (formed at ultra-low sea-level) as possible sources. His conclusion was that sea water entered the rock strata near Bab el Mandeb and flowed at depth to Atlantis II Deep with a temperature "not much higher than 100°C at most". Recent results from Ross (1972) indicate that the minimum temperature of the water is 104°C and that based on heat flow and balance calculations it probably originated locally at a depth of probably 500 to 1500 m or a little more. This together with the discovery of fractures crossing Craig's suggested flow path (Ross, 1972) and the occurrence of other brine pools between Atlantis II Deep and Bab el Mandeb (Backer and Schoell, 1972) casts doubt on Craig's source hypothesis. Trace element studies by Brooks et al (1969) support the seawater hypothesis but Brooks et al favoured a more local entry point.

An alternative source, not discussed by Craig (1969), is the water expelled by gypsum on its conversion to anhydrite. Brooks et al (1969) found the evidence for or against this hypothesis to be inconclusive. This source must now be considered unlikely because recent evidence (Ross et al 1972) indicates that the evaporites consist of anhydrite.

Alkalis

All writers (e.g. Degens and Ross, 1969) agree that the source of the main major elements in the brines is evaporites. Na, K, Li and Cl presumably came from halite and sylvite (or possibly carnallite). White (1968) suggested that Ca comes from cation exchange and hydrolytic decomposition of plagioclase. Brooks et al (1969) suggested Ca is derived from dolomitization of calcite or from the leaching of gypsum and "sediment" at "elevated temperatures". They further suggested that dolomitization and reaction with "clays and silicate minerals" would remove Mg. Craig (1969) considered that oxygen isotope evidence mitigated against Mg fixation in either dolomite or silicates and suggested that membrane filtration through clays would remove Mg, SO₄, I and F.

Backer and Schoell (1972) noted that most of the brines in the Red Sea deeps, unlike those in Atlantis II Deep, are not depleted in Mg and SO_4 ; and they have a wide range of salt contents. This together with evidence on the origin of the water suggests a local source for the salt.

Bush (1970) has illustrated the chemistry by which water percolating through evaporites can evolve in subkha environments. This is complicated even at ambient temperatures. The more complicated reactions that undoubtedly take place at elevated temperatures and pressures are problems yet to be elucidated.

Volcanism

The Red Sea deeps are at the location of extensive volcanic activity and it is difficult to accept that this has not contributed some material to the brines, especially since these environments would prevent all but a minimum of loss of volcanic emanations. Atlantis II Deep itself is a basin within basalt and the geothermal sediments are separated from the basalt by no more than about four meters of carbonate sediment (Hackett and Bischoff, 1973). Craig (1969) claimed that the strontium isotope measurements made by Faure and Jones (1969) were consistent with derivation from marine sediments or from shells of marine organisms. Faure and Jones acknowledged discussion with Craig but actually concluded only that the strontium isotopes in the brine differed significantly from those of normal sea water. Faure et al (1971) and Carwile and Faure (1971) subsequently stated that the strontium isotope evidence favoured a volcanic source. Schoell and Stahl (1972) found from isotope studies that some of the carbon in the brines is probably juvenile. Ba is enriched 70 times in the hottest Atlantis II brine (Bonatti et al 1972) and may have a volcanic source. This has also been postulated for Ba occurrences at the East Pacific Rise (Bostrom and Peterson 1966) and at En Kafala in the Danakil Depression (Bonatti et al 1972).

The aqueous phase separating from magma can be rich in NaCl, KCl and CaCl_2 and it could be difficult to distinguish such volcanic components from evaporitic components, if other volcanic contribution was very minor.

Base Metals

Although there are references to various possible sources for the base metals in the deposits, the writer may be the first to suggest that different metals probably have different sources.

Analogies with deposits in the East Pacific Rise (Dasch et al 1971, Bostrom and Peterson, 1966; Cook 1972), the Karlsberg Ridge of the Indian Ocean (Dmitriev et al 1971), and the Danakil Depression (Bonatti et al 1972, Holwerda and Hutchinson 1968) suggest Fe, Mn, Ba and possibly some Cu, Co, Cr, Ni, Pb and Zn are derived from basalt. Miocene and Pliocene sediments associated with evaporites however contain, locally abundant, Zn, Pb, Cu, V, Mn and Fe. From this it would seem that the goethitic sediment, like that at En Kafala and the East Pacific Rise, could be essentially derived from hot basalt by the action of salt water; the montmorillonitic sediment could be essentially derived from shales associated with evaporites (few analogous sediments in open oceans have been reported) and the sulphide sediment could possibly be derived from sediments associated with evaporites or from the basalt. Ellis (1968) experimentally found that hot brines can effectively leach Fe, Mn, Cu and Pb from andesite or shale.

The glassy basalt from the Red Sea deeps analysed by the writer (discussed in section 1.3.6) has normal trace metal values but the crystalline basalt showed a concentration of Cu and Zn. This is probably a local concentration under a glassy skin of the basalt but such mobilization and local concentration could make these elements more available for leaching. It is surprising that only Cu and Zn were concentrated however there were no signs of contamination in the samples, and no other basalts in the same batch had abnormal base metal values. More Red Sea deeps basalt analyses, of course, are needed for more conclusive evidence.

LEAD ISOTOPE DATA ON BASE METAL SOURCES IN RED SEA DEEPS

Several sources are possible for the base metals in the brines, and as suggested above, it is possible that the different sediments deposited have base metals from different sources. It therefore seemed reasonable to the writer that "finger-printing" of the lead using isotopes could be used for correlation (or irrelation) of the lead in the possible source

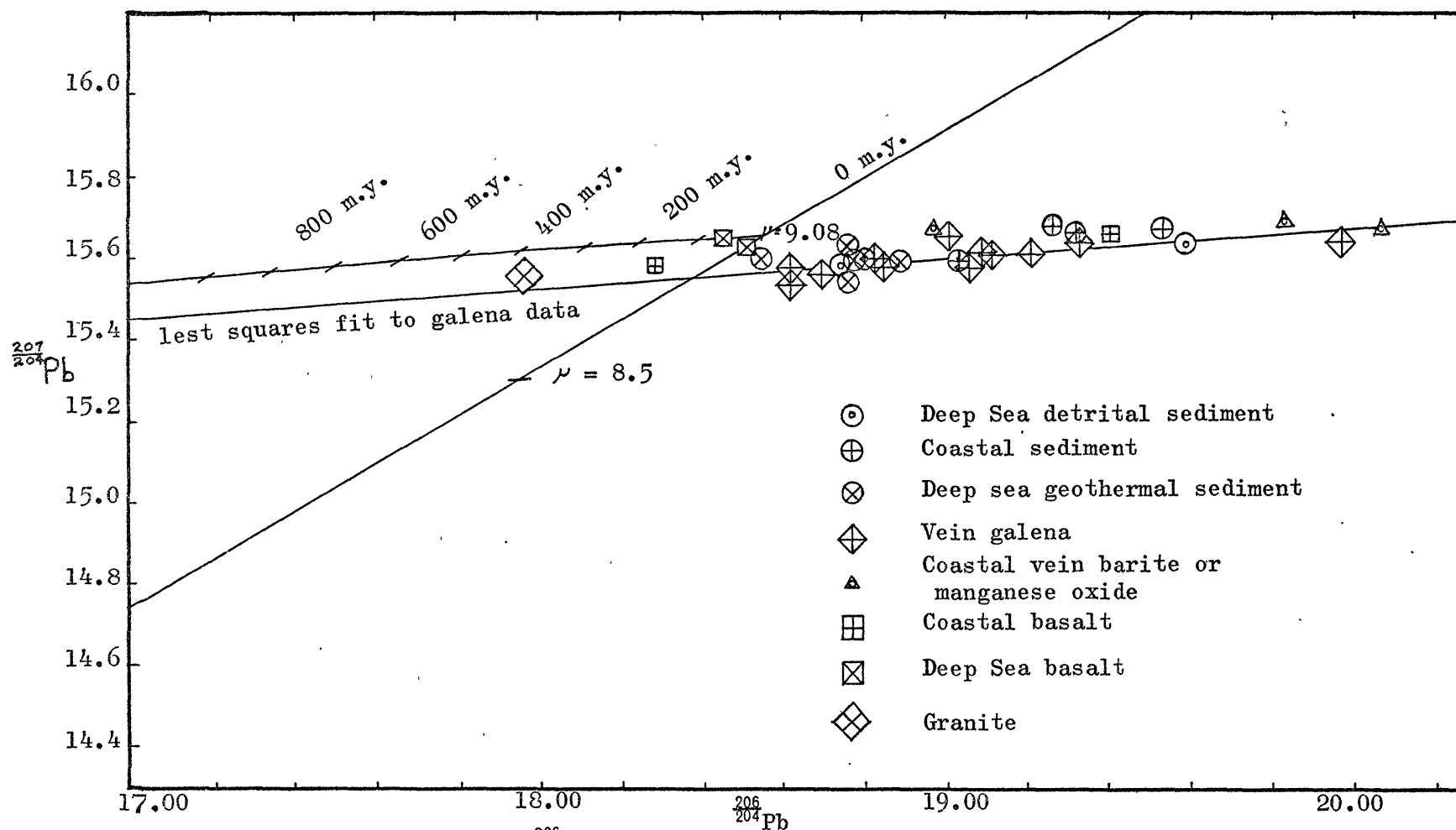


Fig. 216. Plot of $\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$ versus $\frac{^{207}\text{Pb}}{^{204}\text{Pb}}$ of specimens from Red Sea region by I. Swainbank of I.G.S.

rocks with the lead in the sediments of the deeps. The sources considered were Red Sea basalt, Miocene coastal sediments, coastal Pb-Zn deposits and Quaternary Red Sea sediments. Dr. I. Swainbank of the Institute of Geological Sciences kindly measured and studied the lead isotopes in relation to the source of the Red Sea deposits and the source of the Jabal Dhaylan deposits (discussed in section 3.1.2) and in relation to the metallogenic province of the Red Sea region (discussed in section 3.2.4). Fig. 216 is a graphic representation of his results which are also presented in more detail in Table 9.

Two previous studies were made on the lead of the Red Sea deeps. Delevaux et al (1967) found that lead in a vein at Umm Gerad (Saudi Arabia) and in veins in Egyptian Cenozoic deposits have similar isotopes to lead in the 56°C brine of Atlantis II Deep. All are J type leads giving negative ages and plot close to the secondary isochron that fits the new data. Cooper and Richards (1969) concluded that the leads from Atlantis II Deep geothermal sediments (facies not discriminated) and barren Red Sea sediments differ from each other, and that lead in Discovery Deep sediment can be considered as a mixture of Atlantis II Deep type lead with barren sediment type lead.

The new data indicate that the lead in the Red Sea deeps basalt differs significantly from that in the sulphide facies and montmorillonite facies of Atlantis II Deep hence it is concluded that the lead is dominantly non-volcanogenic. Kaplan et al (1969) and Cooper and Richards (1969) have noted that Pb closely correlates with Cu and Zn but not Fe and Mn hence it is unlikely that the main source of the Cu and Zn in the sulphide facies is volcanic.

Although the lead isotopes in normal Red Sea water were not measured it is very unlikely that such sea water lead would lie on the secondary isochron which fits the mineralization data. Normal sea water is also less radiogenic than the leads examined and can be excluded as a source (I. Swainbank pers. com.). The results of Cooper and Richards (1969) tend to exclude Quaternary Red Sea detrital sediments as a source for the lead in the geothermal deposits. The new data do not conflict with this but do indicate that the Quaternary detrital sediment lead and the geothermal sediment lead have the same source or very similar sources because the leads fit the same secondary isochron. Plio-Pleistocene

SPECIMEN	ISOTOPE RATIOS			
	$^{206}/_{204}$	$^{207}/_{204}$	$^{207}/_{206}$	$^{208}/_{204}$
<u>Red Sea Deeps</u>				
Iron Montmorillonite Facies (Atlantis II Deep)	18.77	15.59		
Sulphide Facies (Atlantis II Deep)	18.76	15.54		
Goethite Facies (Atlantis II Deep) A	18.55	15.59		
Goethite Facies (Atlantis II Deep) B	18.76	15.63		
Goethitic sediment from Nereus Deep	18.89	15.59		
Detrital Sediment V14-126	19.59	15.63		
" " V14-120	18.85	15.66	0.8307	38.24
" " V14-116	18.74	15.58		
<u>Coastal Galenas</u>				
Jabal Dhaylan JJM2	18.85	15.59	0.8270	38.42
" " JJM3 vein	19.21	15.60	0.8118	38.41
" " JJM3 crystals	19.33	15.64	0.8092	38.46
" " DY19/83.9	19.09	15.61	0.8171	38.36
" " DY19/86.9	19.12	15.60	0.8165	38.35
" " JJM10	18.83	15.60	0.8287	38.22
Maqna (Jabal Tayran)	19.97	15.64	0.7830	39.22
Merataba	18.62	15.57	0.8356	38.09
Rerga	18.62	15.53	0.8342	37.83
Wazra	19.06	15.58	0.8169	38.13
Umm Gerad	18.70	15.56	0.8322	37.89
Zug el Bohar (Egypt)	19.01	15.65	0.8230	38.49
<u>Coastal Sediments and Veins</u>				
Azlam barite	19.82	15.67	0.7900	38.96
Yahmum barite	18.97	15.67	0.8256	38.06
Azlam cupriferous sandstone	19.53	15.71	0.8041	38.71
CJ1 cupriferous sandstone (Jabal Dhaylan)	19.03	15.59	0.8249	38.26
Sharm el Sheik manganese oxide	20.62	15.68	0.7607	39.54
Plio-Pleistocene limestone (Jabal Dhaylan)	19.27	15.67		
Miocene reef limestone	19.33	15.65		
<u>Basalts</u>				
Deep Sea basalt glass	18.45	15.65	0.8478	37.91
Deep Sea phyric basalt	18.51	15.62	0.8437	38.20
Qalib	18.28	15.86	0.8532	38.09
JJM11 (Jabal Dhaylan)	19.41	15.66	0.8074	39.28
Azlam	18.38	15.60	0.8490	38.36
<u>Granite</u>				
Felspar from JJM3 granite (Jabal Dhaylan)	17.96	15.57	0.8666	37.59

Table 9. Lead Isotope Ratios obtained by I. Swainbank of IGS from specimens from Red Sea region.

limestone from Jabal Dhaylan contains lead distinctly different from that in the deeps. The lead in galena veins at Jabal Dhaylan and other localities fits the secondary isochron and the new data agree remarkably well (in so far as they can be compared) with the data of Delevaux et al (1967). It appears therefore that most of the lead in the coastal deposits, the Red Sea deeps detrital sediments, and the iron montmorillonite facies of Atlantis II Deep, is related to the lead in the sulphide facies of Atlantis II Deep.

Two of the goethitic sediment samples (one from Atlantis II Deep and one from Nereus Deep) have similar lead to that in the iron-montmorillonitic and sulphide sediments. The third (from Atlantis II Deep) is less radiogenic than the sulphide sediment and in this respect appears to have a much higher volcanic contribution. The Nereus Deep sample is more radiogenic than the two from Atlantis II Deep. This conforms to the trend of the three detrital sediment samples from the deep in that radiogenic contamination of the samples increases with latitude (discussed further in section 3.2.4).

As the source of the lead in the coastal rocks is of the same nature as the source of the lead in the sediments of the deeps it is likely to be Miocene or possibly Pliocene sediments. The contribution of lead to the coastal deposits from the basement and Tertiary basalt is likely to be minor. Basalt at Qalib and felspar separated from granite near Jabal Dhaylan contain less radiogenic lead than the lead deposits and associated sedimentary rocks. Furthermore one would not expect a significant contribution of lead from sialic crustal basement sources to the Red Sea deeps if the brine has a relatively local source.

The conclusion is therefore that lead in the sulphide and iron montmorillonite facies appears to be derived from Cenozoic sediments and that as it closely correlates with zinc and copper, all of these probably had essentially the same source. The lead in the goethite facies is more problematical. Some of this appears to have a volcanic source but some of it appears to have been derived from Cenozoic sediments like that in the sulphide facies. It is possible that the lead had two sources. It must be noted however that lead does not correlate well with iron and manganese hence it is not very reliable as a source indicator for them.

Furthermore in the sedimentary pile the hot brines could have mixed some of the lead of the sulphide and iron montmorillonite facies with that in the goethite facies. In general basic volcanism and alteration of basic rocks yield copious quantities of secondary iron minerals. Such could well be the source of the iron in the goethite facies as it appears to be in iron concentrations at En Kafala and at the East Pacific Rise where copper, and especially zinc, are very scarce.

TRANSPORTATION

Most writers (e.g. Degens and Ross, 1969) agree that the base metals are transported as solutes in brine which moves because of density differences caused by heating.

With increasing temperature Ca SO_4 and Fe CO_3 become less soluble and would presumably be deposited thereby reducing the density of the brine but perhaps simultaneously reducing the efficiency of the transportation system by clogging the channelways. If Ca SO_4 and Fe CO_3 are not deposited and more NaCl is dissolved from the surrounding rocks the density of the brine would increase and thus also reduce the efficiency of the transportation system. The brine however is gushing into Atlantis II Deep at a great rate (see section 1.1.2). The cause for this is not known and the writer has not found references to the points raised above. A possible solution to the problem is that the water becomes fully heated after leaving the evaporites so that the sodium content cannot further increase with temperature. Sea level fluctuations may also have affected brine flow (see next discussion).

The solubility of halite and the high rate of discharge of the brines suggests they are moving through relatively large subterranean passages. Some of the coastal evaporites contain halite intimately mixed with smectite hence it seems likely to the writer that some of the montmorillonite and other sediment in the deeps could have been transported in a colloidal suspension and altered to its present state while in suspension.

DEPOSITION

General

Many writers consider the hot brines in the Red Sea deeps to be ore-bearing fluids presently precipitating base metal sulphides but this is somewhat erroneous. The most recent precipitate in Atlantis II Deep is iron-

montmorillonite, not base metal sulphide (Bischoff, 1969). The present brine in the bottom of Atlantis II Deep contains only 0.3 ppm Cu. The lead content is twice that of the copper but galena is apparently absent even from the sulphide facies. The deposits are distinctly layered and few writers have suggested a cause for this layering. Bischoff suggested that each facies represents a dominance of one process of mineral precipitation over another. According to Bischoff fluctuating temperature and brine level could explain some of the variation but most is due to the chemistry of the brine.

The layering may also be related to depositional rates of the various sediments. Unfortunately such rates have not been published. Instead Ku (1969) and Ku et al (1969) have given rates for carbonate detrital sediment and silicate detrital sediment and for geothermal sediments as an undifferentiated group. Furthermore the various methods used (C^{14} and uranium series isotopes) gave inconsistent results. Because sampling was not made with respect to facies, algebra has to be used to estimate the rate of deposition of the different facies. Using C^{14} dates (Ku et al 1969) on core 118, goethitic sediment appears to have been deposited at the rate of 5 to 10 cm/1000 years (assuming the same rate for detrital sediments - see section 1.1.4). Using Th^{230} isotope data (Ku, 1969) on a section of core 85, the rate for goethite is 109 cm/1000 years! On the same section the rate, using Pa^{231} isotope data (Ku, 1969), is 66 cm/1000 years. The rate of goethite precipitation for core 161 using Th^{230} isotope data is 257 cm/1000 years (disregarding a short interval of manganite). With core 120, Th^{230} data give a rate of 35 cm/1000 years for goethite and 29 cm/1000 years for iron montmorillonite. The present data (Ku (1969) and Ku et al (1969)) are useless for calculating the rate of sulphide deposition. The above calculations obviously are faulty because of experimental error nevertheless it is very likely that the rates of deposition were not steady. Some (slowly deposited) cores of goethite contain much detritus, others have none. There is no evidence from the grain size or morphology (e.g. coarse flocs) of the goethite for very rapid deposition. Some of the deepest sediment described by Bischoff (1969) consists of intermixed detritus and goethite, hence it is likely that the early phase of brine deposition was relatively slow.

From initial sedimentation 25,000 years ago to 12,900 years ago detrital carbonate was deposited in Atlantis II Deep. Since 12,900 years ago the present phase of geothermal activity has been continuous but the sulphide sediment was deposited in the earliest part of the phase (Hackett and Bischoff, 1973). The sulphide facies is therefore probably related to the close of the last glacial period, when the Red Sea was a barred basin. Aragonitic hard layers (discussed in section 1.1.3), found throughout the main Red Sea Trough, are thought to have formed in periods of lowered sea level and high evaporation. In places these contain iron oxide with traces of anhydrite, gypsum and sphalerite (Olausson, 1971). An earlier phase of geothermal activity in the northern part of the Red Sea near St. John's Island can also be related to glaciation. Herman and Rosenberg (1969) found a goethitic bed that was deposited in a cold period near the beginning of an interstadial period which began about 100,000 years ago (incidentally there appears to be a printing error in their conclusion indicating this bed is 60,000 years old). This bed has a low Cu and Zn content but was not deposited in a deep. Hugh Dunsmore (Imperial College, pers. com.) has found a relationship between mineralization of Mississippi Valley type deposits and glaciation. He suggests that eustatic readjustment at the close of glaciation could cause renewed water movement which would mobilize hot metalliferous brines. The present brines appear to have been more metalliferous in the earlier phase of deposition in Atlantis II Deep than now, hence during times of relative immobility they could have become highly charged with metals. Exhaustion of source metals, changes in flow rate (partially governed by sea-level changes) and possibly changes in temperature could account for the present relatively low metal content of the brines.

Goethitic Facies

The marked depletion of dissolved iron in the 44°C brine of Atlantis II Deep suggested to Bischoff (1969) that the process of precipitation of goethitic material is currently active. The amount being deposited must be minute however because the dominant material now settling out of the brines is iron-montmorillonite.

Mixing of ordinary sea water with the 56°C Atlantis II Deep brine is sufficient to precipitate iron hydroxide (or oxyhydroxide) but such a mechanism, although elegantly simple, does not necessarily account for all or even most of the goethitic sediment. Bischoff (1969) gave possible equations for this precipitation but noted that as oxidation proceeded the pH would drop (this would tend to cause the ferric oxyhydroxide to dissolve). For a steady-state model Craig (1969) calculated that the supply rate of oxygen would be too low and concluded that some of the iron would need to be recycled or be precipitated as carbonate (both are improbable).

The writer considers (section 2.2.5) that much of the goethite is biogenically precipitated. The chemistry of the activities of marine iron bacteria, especially of a species compatible with the observations, are not sufficiently well known to account for the problems discussed above. Nevertheless in other situations the chemistry can be adequately explained with biological precipitation (Zajic, 1969, Harder, 1919). Iron bacteria typically live in oxidizing environments where ferrous iron is available e.g. at the interface of a reducing environment. Under these conditions the biogenically precipitated iron is often mixed with inorganically precipitated iron. That in the Red Sea appears to be dominantly biogenic.

The precipitation of iron as an oxide rather than a sulphide is probably due to the episodic supply of sulphide ions however it is also possibly due to competition with zinc for available sulphide ions.

The preferential deposition of Fe over Mn is probably due to Eh and to the higher solubility of Mn salts.

Iron Montmorillonite Facies

The main components of this facies are iron oxide and silica hence deposition is probably occurring because of instability of these substances in the brine due to dilution or cooling. The solubility of silica suggests the brine now depositing montmorillonite is probably hotter than that which deposited goethite. The deposition may also be due to settling of an altered suspension but the low alumina content does not favour

this possibility. At En Kafala in the Danakil Depression goethite and smectite were deposited in shallow water simultaneously (Bonatti et al 1972). This probably would also have occurred with the Red Sea brine if it had been deposited in shallow water.

Minor Facies

Minor facies consist of mainly monomineralic layers of manganite, manganosiderite, magnetite and anhydrite. Most writers (e.g. Degens and Ross, 1969, Tooms 1970) consider the manganite facies to be an enigma as it suggests oxidizing conditions and the brines are now in a reduced state. This facies (discussed further in section 1.1.4) however occurs within goethitic sediments which presumably required at least mildly oxidizing conditions.

The manganosiderite facies presumably formed under reducing conditions when CO_2 was available. If depositional rates allowed, the CO_2 may have been supplied by detrital CaCO_3 . A more likely possibility is that CO_2 was supplied as a fumarolic product of the extensive volcanism along the axial trough.

A fumarolic source of sulphate ions may also explain the occurrence of anhydrite but a more likely possibility is that the sulphate ions were derived from sea water. This could have been due to increased availability when the Red Sea was a barred basin (see section 1.1.3) or perhaps to the activity of sulphate reducing bacteria being suppressed or displaced by rising hot brine.

The magnetite was presumably formed by thermal alteration of goethitic material with which it is associated in places near basalt Hackett and Bischoff, 1973).

Sulphide Facies

When the sulphide facies formed, the base metal content of the brines was presumably higher than that of the present brine. The main problem with the sulphide facies is that a source for the sulphur is not readily apparent. Only three different sources appear to be possible viz. volcanic, bacterial reduction of sulphate, and non-biogenic reduction of sulphate.

Sulphur isotope evidence indicates that the sulphide in Discovery Deep has been bacterially precipitated within the sediment, but the sulphide in Atlantis II Deep is more typical of hydrothermal deposits (Kaplan et al 1969). This evidence further appears to exclude volcanic and bacterial reduction sources.

Inorganic reduction of sulphate through reaction with organic matter requires a relatively high temperature perhaps as high as 300°C to 350°C (Kaplan et al 1969) but the evidence appears to be against such high temperatures. White (1968) suggested that the K/Na and Mg/Ca ratios in the Atlantis II brine are indicative of temperatures generally below 150°C. Craig (1969) concluded from deuterium and oxygen isotope studies that the brine temperature could not have been much higher than 100°C. It is possible however that the brines of the past were hotter than those at the present although there is no evidence to corroborate this. Remote possibilities are that sulphate reduction occurred at great depth or that heat from a basalt intrusion caused the reaction where anhydrite and organic matter were abundant together. The lack of a sulphide facies in Discovery Deep further suggests that the sulphides in Atlantis II Deep were not formed by a simple cooling or dilution mechanism.

The main argument against a bacterial reduction source is that the $\delta^{34}\text{S}$ has a narrow range of positive values (+ 3.1 to + 9.8 per mil, Kaplan et al 1969). Bischoff (1969) suggested that in the water column where there is an almost infinite reservoir of SO_4 ions bacteria might give a much narrower range of $\delta^{34}\text{S}$ values than they would in sediment. Watson and Waterbury (1969) suggested that these isotope results could be expected from sulphate reducing bacteria operating in a very thin layer where the diffusion rate could allow the exhaustion of the incoming sulphate ions, but this would probably result in isotope values similar to those in sea water ($\delta^{34}\text{S} \approx 20.2$ per mil). It must also be noted that sulphate reducing bacteria actually do live in a thin layer immediately above the Atlantis II Deep 44°C brine (Watson and Waterbury 1969). It should further be noted that Saxby (1972) found that the sulphides are so intimately associated with organic matter that he found it difficult to remove the sulphides. Also, Stephens and Wittkopp (1969) found iron and zinc sulphides filling the interiors of microfossils.

Whether the sulphide ions were introduced with the metalliferous brines or with separate brines, the source appears to have been anhydrite or sea water that was reduced biogenically or non-biogenically through interaction with organic matter. Further sulphur and carbon isotope study might resolve the problem. Perhaps bacteria can reduce sulphate to sulphide with relatively little fractionation of the isotopes; or perhaps the reaction can take place inorganically at low temperatures when anhydrite is reduced by organic matter that contains the enzymes used by bacteria for the reaction or some other suitable catalyst (e.g. adenosine triphosphate). Barton (1967) found theoretically that methane could very slowly reduce aqueous sulphate to sulphide at low temperature. This is yet to be demonstrated and it is possible that the kinetic isotope effect will cause a broad range of $\delta^{34}\text{S}$ values in such a product. At very rapid metabolic rates the bacteriological fractionation of sulphur isotopes approaches zero (Harrison and Thode, 1958). Metabolic rate increases with temperature hence it is possible that some thermophilic sulphate reducing bacteria cause little fractionation. The fractionation effect may even be related more to the temperature of the reduction process rather than the manner by which it is produced.

G E N E S I S O F I N D I V I D U A L
S E C T I O N I D E P O S I T S I N T H E C o n t d .
R E D S E A R E G I O N

3.1.2 COASTAL LEAD-ZINC DEPOSITS ASSOCIATED WITH LIMESTONES

CLASSIFICATION

The coastal lead-zinc deposits (and zinc occurrences) at Ash Sharma, Azlam, Jabal Dhaylan, Abu Anz, Bir Ranga, Wadi Wizr, Talet Eid, Gebel el Russas, Zug el Bohr, and Um Gheig all appear to be of the same type and they have many common characteristics. The data on most of the deposits however are so scanty that conclusions on genesis cannot be drawn with confidence. By analogy however it will be tentatively assumed that the relatively well known deposits at Jabal Dhaylan and Um Gheig are characteristic of this population and the remaining occurrences have the same mode of origin.

The deposits can be considered as Mississippi Valley types because they share the following characteristics with typical Mississippi Valley types (using data drawn from Ohle 1959, Heyl 1969, Brown 1970b, Stanton 1972).

1. The mineralization is commonly limited to a restricted stratigraphic interval.
2. The host rocks are mostly dolomite and limestone commonly in or near a reef facies. A small percentage however are found in sandstone.
3. The host rocks generally were deposited in shallow marine environments near the margins of basins and commonly on the flanks or crests of positive structures.
4. The mineralization is generally in the form of replacement bodies, breccia and open-space fillings, and veins.
5. Stratigraphic reconstructions indicate that the ore was deposited at shallow depths, probably less than one kilometre.
6. The deposits are mineralogically simple and consist of galena, sphalerite (low Fe content) and barite as the main ore minerals associated with marcasite, pyrite, fluorite (this may be locally abundant or absent), dolomite, ankerite, calcite, quartz, hydrocarbons (locally) and jasper. Copper, cobalt, nickel and other elements are usually very minor constituents. Precious metal content of the galena is generally low.

7. Fluid inclusion evidence indicates that the ore was deposited from concentrated Na-Ca-Cl brines, similar to connate brines in oil-fields, at low temperatures (70°C to 200°C).
8. Alteration haloes, where detected, are subtle and pervasive.
9. Igneous rocks with an obvious relationship to the mineralization are scarce, but where present they are alkaline and usually basic.
10. Deformation and metamorphism have usually been only slight in the mineralized areas.

The mineralization at Jabal Dhaylan is most intense in Miocene limestone but also occurs in Plio-Pleistocene limestone and Oligocene arkose (although no primary lead-zinc mineralization is known in the arkose the lead values exceed the zinc values). This together with the vein and replacement habit of the mineralization indicates the mineralization is epigenetic. Watling's mercury analyses add further support to this conclusion (JJM3, section 1.2.3). The Egyptian deposits occur mostly in Miocene limegrit. Traces also occur in Eocene limestone. The fissure filling and stockwork habit of this mineralization accords with the general epigenetic replacement origin attributed to it by Herbst et al (1965), El-Kholy et al (1970) and El Shazly (1959).

ISOTOPE DATA ON THE SOURCE OF THE JABAL DHAYLAN LEAD

There are several possible sources for the lead at Jabal Dhaylan. The writer considered that the range of possibilities might be reduced by using isotopes to match the lead in the ore with that in the possible source rocks. All care was taken to avoid contamination of the possible source rocks by lead (especially from secondary dispersion). The isotope study was kindly made by Dr. I. Swainbank of I.G.S. (London) and the results are shown graphically in fig. 216 (this study is further discussed in sections 3.1.1 and 3.2.4).

The galena lead isotopes all plot on the one secondary isochron. They are all J types - another feature typical of Mississippi Valley type ores (Brown, 1970b). The Jabal Dhaylan galena leads have a wide range of $\frac{206}{204}\text{Pb}$ values and are relatively radiogenic. This suggests the lead has come from at least two sources and the mixing has not given a homogeneous product.

The trace lead in feldspar of the basement granite at JJM3 and in the Pliocene basalt at Qalib lie on the same secondary isochron but have relatively low $\frac{206}{204}$ Pb ratios. This suggests that some of the galena lead may have come from these sources but most of the lead has another source or sources.

The trace lead in the basalt at JJM11 is very similar to that in the ore lead but K/Ar dating by I.G.S. indicates that this basalt is older than the mineralization and it does not appear to be sufficiently widespread to have contributed much lead by leaching.

The trace lead in the Plio-Pleistocene limestone conversely has a relatively high $\frac{207}{204}$ Pb ratio and therefore also appears to be an unlikely source for the lead in the galena. Although not measured, sea water lead is also an unlikely source as it is generally less radiogenic than the galena lead and would probably not fit the secondary isochron anyway. The remaining possible sources are Miocene limestone, Miocene shale, Miocene evaporites, or Oligocene arkose.

The writer tentatively advocates Miocene limestone and evaporites as the main source rocks. It seems likely that some contribution would also have come from the other sedimentary rocks and the basement. The close association of Mississippi Valley deposits with Phanerozoic carbonate rocks, especially reef facies, suggests that these rocks are essential to the ore-forming process. Arkoses do not show the same temporal relation to mineralization. The arkose at Jabal Dhaylan furthermore is not derived from a strongly mineralized terrain and is thin and lenticular. An angular unconformity indicates it was partially eroded before the deposition of the Miocene limestone. The shale at Jabal Dhaylan is also thin and lenticular and is not known to be highly carbonaceous.

CONCENTRATION, DIFFERENTIATION AND TRANSPORT

Concentration of the lead and zinc could have occurred during the deposition of the source rocks, and during mobilization, transport and deposition. It is difficult to discuss these processes separately.

As the ore occurs in veins in Pliocene and older rocks it cannot itself be any older than Pliocene and is therefore some of the youngest known

Mississippi Valley type lead-zinc mineralization. Despite the very young age of the mineralization the source area does not appear to be exposed and cannot be identified. The transport history therefore is conjectural.

The transporting medium presumably was a sodium chloride rich brine. The present subsurface waters are salty; the rocks are salt veined and the sea very probably covered these shallow water rocks at the time of mineralization.

There is no evidence that the sediments were abnormally enriched in lead and zinc during deposition however some metal differentiation probably occurred by the incorporation of lead and zinc into limestone. Solid solution and substitution can allow rhombohedral high-magnesium calcite and dolomite to incorporate relatively high proportions of Mn, Fe, Zn, Co and Cd which have similar ionic radii (D. Shearman and H. Dunsmore, pers. comms.). Similarly, orthorhombic aragonite can incorporate Sr, Ba and Pb. Typical enrichment factors found by H. Dunsmore (pers. com.) by comparing Recent with equivalent ancient limestones are 10 for Sr^{++} and 5 for Mg^{++} . Assuming a conservative enrichment factor of 3 for Pb and Zn he calculated that at least 60 ppm combined Pb and Zn are available during diagenesis for transport. (Typical Recent limestone may have 30 ppm Pb, 60 ppm Zn and typical ancient limestone may have 10 ppm Pb, 20 ppm Zn). Cu and Ag do not find appropriate places in these lattices and hence are not easily incorporated in limestones. A similar mechanism might operate with evaporitic carbonates and probably sulphates (e.g. orthorhombic anhydrite with Sr, Ba and Pb) and chlorides. The base metal content of sea water is extremely low and evaporites are not generally known to contain high base metal contents but even to get a few centimetres of aragonite about a kilometre depth of sea water has to be evaporated. This involves a considerable concentration. Bush (1970) has shown how sequences of solution and deposition can effect remarkable changes in evaporitic major element contents. Presumably the base metal contents could be similarly affected but at present it is not known how this will cause concentration or dilution in evaporitic solutions. Further research is obviously needed.

Strakhov (1967) and H. Dunsmore (pers. com.) have found that Mississippi Valley type deposits generally occur in host rocks formed in arid and equatorial regions. This of course also applies to the Red Sea lead-zinc deposits. The reason for this climatic control is unknown but could be related to the need for evaporites for brine and possibly some lead-zinc generation. It also might be related to the climatic control on limestone reef formation needed for traps and some metal generation. Tropical regions have the highest biogenic production rates and this might be another reason for climatic control as organic matter is necessary for the reduction of sulphate ions to sulphide ions. H. Dunsmore (pers. com.) also pointed out that evaporites are the obvious source for the sulphide ions. He further pointed out that in climates of relatively high temperature more metals may be incorporated into the limestone during deposition. (An increase in temperature of 10°C causes a 5% weight increase in the MgCO_3 incorporated into the CaCO_3 lattice). It is also possible that high salinities as in arid environments, may cause base metal concentrations to be relatively high in limestone. It is known that the strontium content of aragonite increases with the salinity of the solution from which it is deposited (H. Dunsmore, pers. com.).

During diagenesis, aragonite and high-magnesian calcite convert to low-magnesian calcite. This cannot readily accommodate the base metals hence they become available to the brine transporting medium (Shearman and Dunsmore, pers. com.). Further differentiation of metals can be caused by the brines themselves. For example, Nriagu and Anderson (1971) calculated that for sulphur-poor brines "enough Pb and Cd, but not enough Hg and Cu, can be transported in the form of chloride complexes to account for the formation of ore deposits".

Organic matter in limestones may be another (neglected) source of base metals. Presumably most of the organic matter is lost during diagenesis. This can supply base metals, sulphur and magnesium as it is well known that base metals and sulphur can be considerably concentrated in organic matter. Furthermore, Friedman et al (1973) pointed out that about a third of the magnesium in Recent magnesian limestones actually occurs in the organic matter associated with the high-magnesian calcite.

The motivation of the brine was probably dominantly tectonic. The Jabal Dhaylan area consists of a series of horsts and grabens, and dips range from 25° ENE (Oligocene) to 30° WSW (Pliocene). Furthermore, sediments and geomorphology indicate independent block movement as well as eustatic sea level changes. Dadet et al (1970) concluded that the piercement of the Jabal Karkuma diapir (50 km NNW of Jabal Dhaylan) began about the end of the Pliocene and continued through the Pleistocene. The initiation of this movement could have been caused by basalt intrusion (although no basalt is exposed). The basalt at Qalib was extruded about 1.78 m.y. ago (I.G.S. determination discussed in section 1.3.5) which is toward the end of the Pliocene and it is quite possible that salt movement, magma movement, fault block movement and ore fluid movement are all related and penecontemporaneous. It is also possible that fluid movement was affected by thermal convection due to a compressed geothermal gradient associated with a high heat flow. This is probably the motivation of the hot springs presently active in the southern part of the coast.

DEPOSITION

Interaction with sulphide ions most probably caused the deposition of the ore minerals. Less probably; the ore could have been deposited from a complexed solution due to the breakdown of the complex by reduction in temperature, by dilution or by changes in Eh, pH conditions. The source of the sulphide ions is considered in the following discussion.

Space and access were necessary at the site of deposition for the accumulation of ore. Space presumably was provided by cavities in porous reef rocks, fracture zones, and porous arkose beds; less space was provided by replacement of limestone. Access was presumably through a combination of permeable beds and fault zones. The Jabal Dhaylan horst is fault-bounded and although these faults are not known to be mineralized they probably provided the final channel-ways to the site of deposition.

As there is very little organic matter and no apparent residual organic matter in the host rocks the sulphide ions were probably generated elsewhere and probably migrated independently of the metalliferous brine into the host traps.

The temperature of deposition was relatively low, possibly around about 100° C. This is indicated by the colour of the phytoclasts in the host rock; the mercury chemistry (by J. Watling), the lack of fluorite but occurrence of marcasite and gypsum, and by the trace element contents of the galena and barite. There is no evidence that the depth of burial was great. The formations are relatively thin and erosion has not been extensive. The source of the heat was probably geothermal assisted by regional high heat flow.

	GENESIS OF INDIVIDUAL	
SECTION I	DEPOSITS IN THE	Contd.
	RED SEA REGION	

3.1.3 SULPHATE, SULPHIDE AND SULPHUR OCCURRENCES

Sulphate occurs mostly as anhydrite, and the anhydrite in the Red Sea region is typical of that deposited from sea water in sabkha environments (D. Shearman pers. com.). The gypsum presumably is sedimentary or is derived from the anhydrite by hydration. Barite genesis is discussed later in section 3.1.4.

As a volcanic source of sulphide ions is generally lacking in the Red Sea region it appears that the sulphide is derived from the abundant anhydrite by reduction with organic matter. The main problem here is whether this occurred at high temperatures non-biogenically or at low temperatures biologically. (This problem is discussed in more detail in section 3.1.1). Some sulphide ions presumably were produced by decaying organic matter but such sulphide is generally minor compared with that produced by sulphate reduction. Sulphides produced from bacterially generated H_2S generally are not biomorphic (e.g. the mackinawite produced by the writer, fig. 104) hence it is difficult to characterise the source of the sulphide. Isotopes are commonly used for distinguishing between biogenic and non-biogenic sulphides but even this method has problems as discussed in the previous section on the Red Sea deeps (3.1.1). A program and samples were prepared for sulphur isotope study of the coastal sulphur species but unfortunately facilities for this did not eventuate. The sulphides in the Arabian - Red Sea coastal plain occur with Mississippi Valley type lead-zinc mineralization as at Jabal Dhaylan, with barite-jasper veins as at Umm Gerad, and with native sulphur as at Azlam. The mineralization at these occurrences and at analogous occurrences elsewhere in the Red Sea region appears to have formed at relatively low temperatures hence it is more probable that the sulphide was derived from sulphate by biogenic reduction than by non-biogenic reduction.

The sulphur occurrences in the Red Sea region (other than those associated with volcanoes in Ethiopia) are invariably associated with gypsum or

anhydrite, and the Egyptian occurrences, which are the largest, are also associated with oil or bitumen. The sulphur occurrences also commonly lie in fault zones. The most probable origin appears to be partial oxidation of sulphide ions (e.g. H_2S) produced by bacterial reduction of anhydrite. The partial oxidation may have been inorganic or biogenic. The distinction between the two is difficult to make because bacteria can precipitate crystalline sulphur (Kuznetsov et al 1963). Schnellmann (1959) considered that it was likely that sulphate reducing bacteria produced sulphur directly. Although bacteria are known that will do this (Zajic 1969) they are rare. The reduction of SO_4 in anhydrite results in the replacement of the anhydrite by aragonite, with a volume reduction of 25% (Schnellmann, 1959). The aragonite at Gemsa associated with the sulphur is cavernous, eucrystalline and occupies an "anomalous stratigraphic position" hence Schnellmann considered it to have resulted from the biogenic production of sulphur. Similar cavernous aragonite layers associated with traces of sulphur occur at Jabal Yahmum and Sharm al Badu (in two hills both named Tuwyil al Kibrit).

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SECTION I	D E P O S I T S I N T H E	Contd.
	R E D S E A R E G I O N	

3.1.4 COASTAL BARITE-JASPER VEINS

COMMON CHARACTERISTICS

The coastal barite-jasper veins have the following common characteristics:-

1. They occur as distinct steep to vertical fissure fillings in faulted areas. Some replacement however appears to have occurred at Jabal Dhaylan (discussed in section 1.2.3). The veins are mechanically but not chemically related to the host rocks.
2. They occur close to the unconformable contact of Cenozoic and Precambrian rocks, and occur either side of the unconformity (on both sides at Jabal Yahmum).
3. All are of Cenozoic age.
4. They possess a common simple mineralogy viz. barite, jasper, manganese oxides, galena and secondary zinc and copper minerals. The mineral proportions however have wide ranges. Some veins contain either barite or jasper but most contain both. Base metals are minor to trace constituents.
5. The galenas in the veins all have similar low trace element contents (section 3.2.3).
6. Lead isotopes indicate that the galenas are all J types and all lie on the same secondary isochron (I. Swainbank, section 3.2.4).
7. The barites have high Sr contents (in the range of 3 to 5%) which is normal for low temperature barite (J. Barrett, pers. com.)

With these characteristics in common it seems reasonable to assume that all the veins have similar origins. Furthermore, as the barite at Jabal Dhaylan is an integral part of the lead-zinc mineralization there, it seems reasonable to assume that the genesis of the barite veins is related to the genesis of the lead-zinc mineralization. At Jabal Dhaylan however barite is minor compared with lead and zinc mineralization and is known at only three of the lead-zinc prospects. The Jabal Dhaylan barite furthermore is minor compared with several of the other barite vein

deposits along the coastal plain e.g. Umm Gerad. Also there are no distinctive occurrences of jasper at Jabal Dhaylan. Jasper is generally absent from barite veins in limestones but occurs with barite in most of the veins located in Precambrian basement rock.

The barite-jasper veins appear to be epigenetic deposits precipitated partly by the aid of bacteria at low pressure and temperature in open spaces produced by faulting near a major unconformity.

SOURCES

The source of the barium and base metals is problematical. Some of the veins are remote from Cenozoic basalt hence direct magmatic influences appear to be minimal. The remaining possible sources are the basement and the Cenozoic sediments. Granite occurs within a few kilometres, at most, from the barite veins and it is appealing to suppose that deep circulating solutions moving along deep faults in the basement may have leached the barium from granite. The lead isotopes indicate that some of the lead (hence presumably some of the barium too) has come from the basement but most has come from the sediments (lead isotopes, studied by I. Swainbank of I.G.S., are discussed in detail in section 3.1.1, 3.1.2 and 3.2.4). Furthermore, the basement rocks are crystalline hence the faults are probably tight at depth. Several large faults in granite near the coast are not apparently mineralized, and barite mineralization is rare outside the coastal region. For example, faults of the Great Najd System in the shield contain minimal barite. Even in the areas of barite mineralization at Aqaba, Jabal Dhaylan, Yahmum and Wazra the basement has inferior or no barium mineralization. Of the sedimentary rocks evaporites and possibly carbonaceous rocks are probably the most likely sources. Repeated solution and deposition of salt should increase the barite content because of its great insolubility. As barium is so readily precipitated by sulphate ions and as sea water is undersaturated in this element it is difficult to see how it would be concentrated in shale, sandstone or limestone except by microorganisms or from an abundant local detrital source (none is known). From the time of Lindgren (1933) it has been known that certain microorganisms can segregate minute barite crystals in their tissues and these can be released on alteration or

solution of the organisms remains. Goldberg (1961) and Wolgemuth and Broeker (1970) found that barium is related to opal in marine sediments, and they postulated a biogenic source for both. Occurrences of barite are perhaps better known in association with evaporites than with carbonaceous rocks. Lindgren (1933) considered the great German barite deposits to be derived from Permian evaporites, and Davidson (1966) claimed that barite underlies virtually every evaporitic sequence in Europe and Asia.

MOBILIZATION

The mobilization of barium is problematical. It presumably was in the form of the very insoluble sulphate or carbonate as its source. Oil field brines commonly carry relatively high trace barium contents (Davidson 1966) and such brines probably resemble the solutions from which the barite veins were formed. Chloride complexes may have dissolved the barite, or the barite may have been reduced to a soluble form by sulphate reducing bacteria at relatively low temperature (such biogenic reactions with barite have been documented by Puchelt and Nielsen 1967). A determined effort was made to study fluid inclusions in the barite but despite much time and specimen examination it was found that although most of the barites have abundant inclusions no fluids were found in them and no reaction was observed on a hot stage. It is concluded therefore that probably the first effect weathering has on barite is the imperceptible release of the fluids enclosed in the barite.

The motivation of the barium-containing solutions could have been due to thermal convection associated with high heat flow in the Red Sea region, to tectonic causes, or to simple gravitation. The structural control of the deposits is suggestive of tectonic motivation, and the fluids initially may have been imbibed rather than injected. Fracture zones and unconformities probably formed the main channelways for the solutions.

DEPOSITION

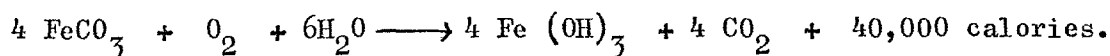
Deposition of the barite could have been caused by an influx of sulphate ions, by a decrease in temperature, by a decrease in pressure by salinity alteration of the ore solution by mixing with connate water, by an increase in Eh, or more probably by a combination of these. A low temperature of

deposition (probably less than 100°C) is suggested by :-

1. High Sr contents of barites (according to John Barrett, Imperial College, pers. com., low temperature high Sr barites contain 3 to 10% Sr and high temperature low Sr barites contain 0.1 to 3% Sr).
2. Low trace metal contents of galenas (discussed further in section 3.2.3).
3. The occurrence of opal and chalcedony in some veins.
4. The presence of remains indicating bacteria lived during deposition of the goethite.
5. The pale colour of the phytoclasts in the jasper.

Textural evidence (section 1.2.4) indicates that the order of deposition was essentially barite followed by goethite followed by silica, with galena spanning the barite-silica period and manganese oxide before and after the silica. Pb, Zn, Cu, Ag presumably came from essentially the same source as the Ba. The Fe, Mn and Si may have been derived from red beds (arkose, sandstone and shale) associated with the evaporites and possibly from the evaporites and basement as well.

The paragenesis may be explained with the following theory. Firstly, barium was precipitated to exhaustion by reaction with sulphate ions supplied by interaction with independent groundwater. Minor oxidation probably occurred as well but would have supplied little sulphate because the sulphide content of the ore brine was low (otherwise iron sulphide would have precipitated). Dilution, pressure decrease, and temperature decrease would have further aided the precipitation of the barite. Following the exhaustion of the barium, iron bacteria, probably derived from connate water, began to multiply and to precipitate the iron (as ferric oxyhydroxide ?). Sphaerotilus thrives in an environment of low concentrations of oxygen, organic matter and ferrous (and/or manganous) salt, and where biological competition is limited. The equation of reaction often quoted for this bacterium and others including Gallionella (a salt water iron bacteria) is



Iron bacteria usually require acid to mildly acid conditions. It will be noted from the above equation that if the iron is present as sulphate or chloride the pH of the environment will drop as the reaction proceeds. A fall in the pH will then cause the precipitation of silica. This

explains the relatively even distribution of the iron oxide biomorphs in the jasper. It further explains the lack of calcium salts in the veins, and the lack of jasper in barite veins in limestones (the pH of the environment would inhibit the activity of iron bacteria and would not cause the precipitation of silica). Further information on the possibilities of bioepigenetic ore deposition has been given in section 2.3.2. The lead and copper sulphides were probably deposited from small proportions of sulphide ions in the barium-rich chloride solutions. Some may have been precipitated however by sulphide locally made available by sulphate-reducing bacteria which would have been able to operate at times when iron bacteria had temporarily exhausted the oxygen supply. Presumably however sulphide ions were never available in sufficient abundance to precipitate iron as well as lead and copper. Under strongly oxidizing conditions manganese was precipitated.

The growth of the bacteria would have been controlled by the flow rate and nutrient supply. The shape of the colonies would be controlled by flow rate, water level and oxygen distribution. With high flow rates and axially disposed oxygen concentrations, as in narrow channelways, colonies would have been closely attached to the walls. With low flow rates, low water level and vertical variations in oxygen concentrations, as in wide channelways, colonies could stretch from wall to wall and build upwards from a floor if the direction of flow permitted.

Brobst (1966) found that the Umm Gerad jasper has "highly anomalous" contents of Be (0.2 ppm) B (5 ppm) and Y (4 ppm). He considered that the veins had a hypogene origin and that the composition of the jasper possibly reflects "the residual composition of the ore forming solution". Be, B and Y are lithophile elements and could have a biogenic or evaporitic origin. B and Y are concentrated by algae and occur in some asphalts, Be is concentrated in coal ash. Boron also occurs in continental evaporites and in seaweeds (Lindgren 1933, Zajic 1969).

The Cenozoic barite-jasper veins in the Red Sea region of Saudi Arabia probably have similar origins to those in the African coastal region except for the En Kafala deposit in Ethiopia. The source of the Ba in the En Kafala deposit however is probably evaporitic. Some of the Egyptian and Sudanese deposits contain calcite but this may be secondary.

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3.1.5 COPPER OCCURRENCES IN THE COASTAL REGION

Copper mineralization is sparse in the coastal region and it is difficult to determine the origin of the occurrences because of their small sizes and because of the extensive weathering.

The occurrence at Prospect JJM6, Jabal Dhaylan, is probably the most characteristic. The secondary copper there occurs as cement in Oligocene arkosic sandstone and is associated with some organic matter. That at Azlam appears to be identical with the JJM6 occurrence. At Prospect JJM10, Jabal Dhaylan the copper occurs mostly in limestone and is replacing galena. There however it is adjacent to arkosic sandstone from which it has probably migrated over a distance of perhaps a few metres. The Ras Benas deposit appears to be similar to that at JJM10 and is associated with arkose and gypsum even though it occurs in dolomitic limestone. The Pre-Cenozoic deposits at Um Bogma and Aqaba (Timna) in Sinai and at Chercher in Ethiopia also occur in sandstone. The occurrence at Prospect CJ1 Jabal Dhaylan is in Quaternary conglomerate as cement but despite its young age no source was obvious. The source however is apparently extrinsic as the pebbles and cobbles in the conglomerate contained no obvious copper mineralization. Marchessau (1969) suggested the source was from an area in the Precambrian basement but no such likely source area was found by either Chaumont (1970a) or the writer. The isotopes of trace lead in copper ore specimens from Azlam and CJ1 lie on the same secondary isochron as those from galena at Jabal Dhaylan and from galena in coastal barite veins (see section 3.2.4). This suggests the copper is derived from Cenozoic sediments.

On the limited evidence available the writer concurs with Bouladon (1968) who considered that the copper mineralization at Jabal Dhaylan is of the Red Bed type (as in the Colorado Plateau). The characteristics of this mineralization (from Stanton 1972, Lindgren 1933) are :-

1. The host rocks are continental and marginal-marine clastic sediments deposited under fluvial-deltaic conditions in arid to semi-arid environments.
2. The orebodies are usually tabular masses related to coarse zones in the enclosing sediment or similar sedimentary structures such as fossil stream channels.
3. The orebodies although related to sedimentary structures often transect such structures.
4. The mineralization usually occurs as cement and less commonly as veins. The copper mineralization rarely is associated with gangue minerals. Of these gypsum appears to be the most common.
5. A close association of copper mineralization and organic matter is very common.
6. Localizing controls on mineralization appear to be Eh and ground water conditions.

The origin most commonly advocated for this type of mineralization is leaching of sediments by brines or connate fluids, transport in permeable zones as sulphates and deposition as sulphides from bacteriological reduction of sulphate ions. It should be noted also that organic matter, especially of low rank, has a very high absorbing power on copper in solution. The above epigenetic type of origin appears to be in agreement with the limited information now available on the Cenozoic deposits of the coastal region.

It is noteworthy that Red Bed mineralization can occur with copper, uranium or vanadium. Because of this, the writer examined likely clastic rocks at many localities between Jeddah and Aqaba with a scintillometer but found no occurrences with radioactivity above that of basement granite. Although this is somewhat discouraging it must be remembered that uranium mineralization can be very localized (Stanton 1972 recorded an orebody consisting of little more than two large replaced logs containing uranium to the value of \$ 350,000). It is encouraging to note that traces of vanadinite have been found by BRGM geologists at Wazra. Ross et al (1972) also found vanadium associated with copper and zinc in pyritic silty clay resting unconformably on Miocene evaporites in a drill hole in the median valley of the Red Sea, and traces of uranium and vanadium have also been found in the Egyptian phosphate deposits at Safaga.

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3.1.6 MANGANESE OCCURRENCES IN THE COASTAL REGION

CLASSIFICATION

Although there is a dearth of manganese sulphides, silicates and carbonates in the coastal region, manganese oxide minerals are very common. The manganese mineralization has a wide variety of occurrences but few of these are sufficiently concentrated and large enough for profitable extraction. The occurrences are also difficult to classify because some types (which also may be widespread throughout the world) probably are so rarely of economic value that they are very seldom mentioned in the literature. Another cause for this difficulty is that some of the occurrences are very unusual. The Black Mountain manganese oxide occurrence, for example, is in a volcano of salt activated by basaltic magma intruding the Danakil evaporites. The occurrences have been broadly classified into "Mississippi Valley" types (lead-zinc-barite-limestone association although not generally recognized as a type of manganese mineralization), Moroccan types (limestone association), Nikopol types (conformable, sandstone association), Wadi Malik types (veins, sandstone association) and En Kafala types (basalt-limestone-evaporite association).

"MISSISSIPPI VALLEY" TYPE

The Mississippi Valley type lead-zinc mineralization at Jabal Dhaylan is accompanied by manganese oxides to a greater extent than iron oxides. The lead-zinc mineralization is almost invariably associated with manganese, and manganese oxide spotted limestone has been used as a guide to mineralization there (P. Chaumont, pers. com.). This manganese mineralization probably had the same source or similar sources as the lead-zinc mineralization and probably was transported in the same ore solutions. Manganese oxide occurs as biogenic concentrations in limestones (section 2.3.4), and some manganese may have ionically substituted for calcium in high-manganesian calcite and been released on diagenesis. Clastic rocks such as arkoses may also be a source for manganese. The deposition of manganese is generally attributed to changes in Eh pH

conditions, and this also explains its separation from iron. Iron is deposited at lower Eh and lower pH than manganese hence manganese usually has a longer residence time in solution and migrates further than iron (e.g. Stanton, 1972, Mart and Sass 1972, Bonatti et al 1972). This explains why manganese oxide is so widely distributed in the coastal zone and why it is found more commonly in limestones than iron. Manganese oxides occur in barite veins in limestone at Maqna but jasper is absent (presumably because of the pH of the environment). It also accompanies barite-jasper mineralization at Yahmum, Rerga, Wazra and Mersa Gulbub (Ethiopia) where it was most probably transported to by the same solutions that transported the barite and jasper. Manganese oxides also accompany the Mississippi Valley type mineralization in the Egyptian deposits.

MOROCCAN TYPE

Mississippi Valley type accessory manganese mineralization appears to merge into Moroccan type manganese mineralization and the two types would probably be synonymous if the Moroccan type contained more lead and zinc and less manganese. The Moroccan type typically occurs as irregular to lenticular manganese oxide concentrations commonly in or near reef facies in limestones or dolomites sandwiched between terrigenous rocks, and resting above an old stable basement (Stanton, 1972). The manganese oxides may contain up to 7% BaO, 6.5% PbO, 0.4% CuO and 0.1% ZnO. The environment of deposition is non-volcanic and generally contains some evaporites. Occurrences at Jabal Dhaylan (beyond the main lead-zinc mineralization), Wazra, Sharm al Badu and Jabal al Buwana are of this type. The source of the manganese was probably the basement and fringing sediments resting on it. The release of the manganese was probably due largely to biological degradation. Surface and ground waters probably carried the manganese in an organic form and deposited it when the organic complex broke down through biological activity or through a change in pH. The occurrence of the manganese at Sharm al Badu and Jabal al Buwana as replacements in coral suggests that slightly acid water carried the manganese and then dropped it when solution of the limestone raised the pH.

The Um Bogma manganese deposits appear to be classic Moroccan types. They occur in Lower Carboniferous dolomite just outside the Red Sea province (discussed in section 3.2.4) and so are beyond the scope of this

study. Some discussion on their origin nevertheless has been given in section 1.2.5. It is relevant to note that Mart and Sass (1972) considered that the ores have been reconcentrated to some extent by diagenesis (when an Eh pH gradient caused some separation of Fe from Mn). Most manganese oxide deposits in the coastal region appear to have been very susceptible to diagenetic influences. Very minor occurrences of manganese oxides associated with calcite crystals in veinlets and geodes in limestones are very common in the coastal region, especially between Wadi Thalba and Jabal Dhaylan. These are probably due to localized diagenetic readjustment.

NIKOPOL TYPE

Nikopol type manganese mineralization consists essentially of extensive conformable manganese oxides associated with shallow water sandstones and shales deposited on a stable platform. Concentration of manganese is by sedimentation and diagenesis. Stanton (1972) classified the Timna deposits as belonging to this type. Like the Um Bogma deposits however they are beyond the scope of this study. The Sharm el Sheik deposit may be of this type but little is known of the deposit and it is probably small. The Wadi Thalba deposits also may be of this type even though it is not yet known to be extensive (this is not intended to imply that size is a criterion of ore type). It will be noted from fig. 45 that the manganese occurrences in the Thalba area occur well within a kilometre of the Precambrian basement basin rim. It is difficult therefore to avoid the proposition that the manganese was supplied from the basement by weathering (with biological agents, as the separation of Mn from Fe is conspicuous). The diagenetic biological concentration of the manganese oxide (discussed in section 2.3.4) presumably occurred within the sand where Eh pH conditions were appropriate. A rise in pH could account for the deposition of manganese oxide and the solution of quartz. Deposits probably of a similar origin are those at Jabal Dhaylan (Wadi Zarufi area) and Wadi Araba in Egypt, at Chercher in Ethiopia and Paloich and Wabuit in Sudan.

WADI MALIK TYPE

There are few detailed descriptions of manganese oxide veins in clastic rocks in the literature so for the present purpose the veins at Wadi Malik (which have yielded more than 2000 tons of ore) will be taken as the representative type. These veins are steeply inclined and occur in Miocene sandstone and conglomerate resting on Precambrian basement. The vein minerals consist of manganese oxides with lesser iron oxides, calcite, quartz and barite. Minor bedding replacement may also occur with this type of mineralization. Wadi Malik type deposits occur in the Halaib area of Sudan and at Wadi Araba, Elba, Araqwab, Hubal, and Mated Onqwab in Egypt. A small but similar vein occurs also at Wadi Thalba. This type of mineralization is probably related to Nipokol type mineralization, the main difference being that during diagenesis the manganese was concentrated in fractures and irregular zones rather than bedding zones. Otherwise the main differences appear to be that Nikopol ore contains less Ba ($< 0.5\%$) than Wadi Malik ore, and glauconite is associated with Nikopol ore but not Wadi Malik ore.

EN KAFALA TYPE

The Black Mountain and En Kafala manganese occurrences in Ethiopia were probably derived from volcanic rocks and were deposited from volcanically heated brine solutions when they were oxidized through mixing with shallow sea water or through exposure to air (for further discussion see section 1.2.5).

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3.1.7 IRON OCCURRENCES IN THE COASTAL REGION

The only large concentration of iron in the Saudi Arabian coastal region is that at Wadi Fatima. Hypogene occurrences containing iron oxides have already been discussed and none of these (except perhaps the Um Gerifat ochre deposit in Egypt) can be considered as a potential source of iron ore. Beach sands and fossil beach accumulations also appear to be insignificant as sources of iron ore. The lack of large scale concentration is probably due to topography, relatively quiet sea conditions, coral growth, small catchment areas, and low rainfall. Laterite deposits too are of very limited extent in Saudi Arabia and appear to have been largely eroded away except where protected by basalt caps (e.g. As Sarat). Cenozoic laterites however appear to be widespread in Ethiopia and Sudan but these are generally beyond the Red Sea metallogenic province.

Al Shanti (1966) considered the probable source of the Wadi Fatima iron ores to be a ferruginous horizon in the nearby Precambrian Fatima Series. The abundance of fossil wood and leaf material in the ore and associated rocks suggests the iron was mobilized from nearby well vegetated low lying land or swamp by the action^{of} biological agencies and was then transported possibly as colloidal iron oxide (ferric oxyhydroxide) unattached or attached to organic matter, or as iron bicarbonate or humate. Deposition was presumably through a combination of pH increase (as the fresh water mixed with salt water), Eh increase through the action of blue-green algae, and through particle trapping in the sticky surface of the blue-green algal colonies which rolled around to form the onkolites. Sizing of the onkolites may have been due to growth limitation by excessive abrasion and mechanical disintegration of the larger onkolites, or to current action and redeposition. In either case, plenty of organic matter was deposited with the iron in the form of macerated tree material and algal mucilage. During diagenesis mildly reducing conditions resulted from the action of sulphate reducing bacteria and other forms of plant decomposition. This enabled the formation of chamosite and pyrite from the goethite. Siderite later formed (mainly as cement) probably by mutual oxidation - reduction of organic matter

and iron oxide.

The Wadi Fatima iron ore is a typical Minette type and also appears to be similar to many other deposits in the Arabo-Nubian region such as those at Wadi Husainiya in Iraq near the Saudi Arabian border, Wadi Halfa in Sudan and the Baharia Oasis in Egypt. These deposits however are generally older (Jurassic and Cretaceous) than the Oligocene Wadi Fatima deposits. In Saudi Arabia the only other occurrence like that at Wadi Fatima so far found appears to be that at Wadi Khulab. Oolitic limestones and red bed sandstones however occur in the northern part of the coastal region.

PART III	Interpretative Geology - Metallogensis -	Section 1	Individual Deposits
		Section 2	Metallogenic Province
		Section 3	Tectonism

SECTION 2

T H E M E T A L L O G E N I C P R O V I N C E

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SECTION 2 THE METALLOGENIC PROVINCE Contd.
 OF THE RED SEA.

3.2.1

INTRODUCTION

As far as possible this study has dealt with all the mineral occurrences in the geographical province of the Red Sea region. Because this region is well defined tectonically and the formations deposited in it appear to have been tectonically controlled, it was assumed that the enclosed mineral occurrences were also ultimately tectonically controlled and therefore are related in some way to each other. That is, it was assumed the geographic - tectonic province is probably also a metallogenic province. Although relationships are easily found for many of the coastal Pb, Zn, Ba deposits reservations were held about some other deposits such as the Wadi Fatima iron deposits and the geothermal deposits in the Red Sea deeps. Much effort was made to find the relation of these deposits to the others in this region.

The previous section dealt with the genesis of the individual deposits. This section deals with their relation to each other and the region as a whole. The following section attempts to deal with the genesis and metallogenesis of the region.

Unfortunately there is no universally accepted definition of the term "metallogenic province" and different writers emphasise different aspects. The concept appears to have been originated by de Launey (1913, quoted by Petrascheck 1965) who considered that "each metallogenic province belongs to a definite regional type depending on the tectonics". Spurr (1923) strengthened the idea that the deposits are inter-related by stating that "all the metals possess a unity, a blood relationship (sic), distinctive of that province". The most rigorous definition is that by Petrascheck (1965) - "The metallogenic province is the entity of mineral deposits that formed during a tectonic-metallogenic epoch within a major tectonic unit and which are characterized by related mineral composition, form of the deposits, and intensity of mineralization". He distinguished metallogenic provinces from "metal provinces" which are regions "in which mineralization featuring a definite metal or a definite set of metals repeatedly appears".

	THE METALLOGENIC PROVINCE	
SECTION 2	OF THE RED SEA	Contd.

3.2.2 SUB-PROVINCES OF THE RED SEA REGION

The main tectonic elements of the Red Sea region are the Arabian plate, the Sinaian plate, the Nubian plate and the Afar Triangle together with the seaward plate fringes and the neo-oceanic crust. The sub-provinces corresponding to these are the Arabian coastal plain, the Sinaian coastal plain, the Nubian coastal plain, the Afar Triangle and the submarine regions.

The main mineralization in the neo-oceanic sub-province is that in Atlantis II Deep which can be taken as the type area for this sub-province. Little is known of mineralization in the other sub-marine sub-provinces other than the occurrence of some Cu, Zn and V in Pliocene black shale resting on Miocene evaporites.

Little mineralization is known in the Sinaian coastal plain along the Gulf of Suez other than manganese at Sharm el Sheik (the Um Bogma district is outside the Red Sea Province). The only mineralization known in the Afar Triangle sub-province is volcanic sulphur and manganese associated with basalt and evaporites (e.g. En Kafala). Most of the mineralization known in the Red Sea province therefore occurs in the Nubian and Arabian coastal sub-provinces, and these are very similar. The mineralization in these sub-provinces is of the Mississippi Valley Pb, Zn, Ba type (with associated sulphur mineralization), Colorado red bed Cu type, Moroccan Mn type, Mikopol Mn type (?) and Wadi Malik Mn type as discussed in section 3.1.6). The occurrences within these groups are all very similar in mineralogy, host rock association, and structure, and, of course, all are of Cenozoic age. Although somewhat similar Cu, Mn, Fe mineralization is known in the regions surrounding the Red Sea province (e.g. Um Bogma) none of the Mississippi Valley type mineralization occurs immediately beyond this province.

The Wadi Fatima deposits are enigmatic. Apart from a doubtful occurrence at Wadi Khulab (discussed in section 1.2.4) no other occurrences of similar Minette type ore are known in the Red Sea province but they are known in the region west of the Red Sea province (as discussed in sections

1.2.6, and 1.2.7). It seems therefore that the Wadi Fatima ore probably belongs to a different province but one which joined or overlapped the Red Sea province in the Oligocene. The En Kafala manganese deposit also is problematic. Its shallow water conglomerate limestone reef association and high barium content suggest affiliation with coastal Moroccan types but its association with goethite, smectite and basalt suggest some affiliation with the geothermal deposits of the Red Sea deeps. En Kafala type deposits are not known in the region immediately outside the Red Sea province.

The geothermal deposits of the deeps are unlike any other deposits in the coastal sub-provinces in many respects and indeed it seems that there are no deposits quite like them anywhere beyond the Red Sea. This does not necessarily mean that the neo-oceanic sub-province is not part of the Red Sea province. Although Petrascheck (1965) defined a province as consisting of an entity of like deposits the province may also contain other deposits. The coastal sub-provinces for example contain several types of deposits. The different types of coastal deposits nevertheless contain some features in common viz.

1. they are of a similar age
2. they have similar tectonic settings
3. they share common elements e.g. Pb, Zn, Cu, Mn, Fe, Ba though in very different proportions and they have certain geochemical similarities (discussed further on)
4. they are low temperature deposits
5. they have similar host rock associations (limestone and sandstone mainly)

These common features are therefore considered as being the main criteria for determining whether or not a deposit is of the Red Sea province.

The geothermal deposits of the deeps are in an environment transitional between a coastal margin and an ocean. Oceanic type deposits such as those of Cyprus or the East Pacific Rise are genetically related to basic magma whereas coastal margin deposits such as Mississippi Valley type deposits are genetically related to shallow-marine sediments. Compared with oceanic type deposits the Red Sea geothermal deposits have in common:-

1. basic volcanic setting
2. deep water environment
3. geothite as the main deposit or as an associate.

Compared to the coastal deposits they have in common:-

1. association with evaporites
2. they share common elements (especially Zn which is relatively rare in true oceanic deposits)
3. they were deposited from metalliferous brine (indirect evidence suggests the coastal Mississippi Valley type deposits were also)
4. they have similar sources
5. they occur in sedimentary formations (the Cyprus pyrite - chalcopyrite ore occurs in pillow basalt interstices).

It therefore appears that the Red Sea geothermal deposits, although having some transitional features are essentially affiliated with the coastal deposits and are still part of the Red Sea province. This is further substantiated by isotope evidence discussed later in this section. Future deposits forming in the axial zone presumably will gradually acquire a full oceanic character as spreading proceeds and the influence of the continental margins ⁱⁿdiminishes.

SECTION 2 THE METALLOGENIC PROVINCE Contd.
 OF THE RED SEA

3.2.3

TRACE ELEMENT STUDIES

Consanguinity of deposits can be shown by geochemical similarities and particularly by trace elements in certain minerals. Although some variation can be expected within a province similarities should be evident at district scale (Petrascheck, 1965). Manganese oxide minerals are commonly related to weathering and tend to act as absorbants hence they are not reliable for primary trace element studies. The only remaining common ore minerals in the Red Sea metallogenic province then are galena and barite.

GALENAS

Galena is commonly slightly weathered and replaced by secondary lead and copper minerals in the coastal region. Nevertheless all the polished sections studied by the writer are homogenous and free of primary inclusions. Extremely high Zn values in trace element contents of Egyptian coastal galenas reported by El Kholy et al (1970) appear to reflect contamination (table 10). To avoid this with the Arabian galenas the writer crushed the specimens to a grain size of less than 1 mm and then carefully selected and hand-picked the galena fragments under a microscope. To further remove contamination e.g. by powdery hydrozincite the hand picked grains were washed with warm dilute HCl. This should not have affected the trace element contents and the results indicate it did not do so. The results (table 10) consistently show low Zn contents, in contrast to the Egyptian analyses. Both sets of analyses were made by the optical spectrograph method and the differences shown by both sets of analyses on galena from the same deposit (Zug el Bohr) are rather surprising. Some differences may be due to analytical technique but the results suggest that sample preparation is probably the main cause of the differences. The high Zn, Cd and Mn of the Egyptian analyses are probably due to sphalerite contamination.

Locality	Ag	Bi	Cd	Cu	Ga	Mn	Mo	Ni	Sb	Sn	Tl	Zn
<u>EGYPTIAN RESULTS</u>												
Um Gheig	15	<10	150	10	<3	33	15	<20	30	75	<30	8000
	10	<10	170	15	3	13	20	<20	30	50	<30	5000
Gebel el Rusas	25	<10	70	10	<3	45	20	20	20	40	45	26000
	30	20	75	10	<3	60	15	25	15	80	40	18000
Abu Anz	10	<10	180	80	5	600	20	50	35	120	<30	67000
	5	<10	200	70	<3	350	15	50	30	80	40	38000
Bir Ranga	20	10	85	15	3	40	25	35	15	100	40	32000
	30	<10	65	10	3	50	15	40	25	90	45	28000
Zug el Bohr	10	<10	65	60	5	50	5	70	15	<40	<30	12000
	15	<10	35	70	3	40	10	40	20	50	<30	31000
<u>IMPERIAL COLLEGE RESULTS</u>												
Zug el Bohr	0.2	<5	<20	4	<2	2	<2	-	<50	<5	-	<50
Zug el Bohr (duplicate)	0.5	<5	<20	2	<2	2	<2	-	<50	<5	-	<50
Zug el Bohr (unwashed)	<0.2	<5	<20	2	<2	2	<2	-	<50	<5	-	<50
Maqna	50	<5	<20	2	<2	2	<2	-	<50	<5	-	<50
Jabal Dhaylan	<0.2	<5	<20	20	<2	2	<2	-	<50	<5	-	50
Merataba	500	10	<20	60	<2	2	<2	-	600	<5	-	<50
Rerga	85	<5	<20	20	<2	2	<2	-	<50	<5	-	<50
Wazra	30	<5	<20	40	<2	2	<2	-	<50	<5	-	<50
Umm Gerad	50	<5	<20	100	<2	2	<2	-	<50	<5	-	<50

Table 10. Trace element contents (in ppm) of galenas from Egyptian and Arabian coastal deposits. Results from Egypt by El Kholy et al (1970) and for Arabia by E. Banargee, analyst, Applied Geochemistry Department, Imperial College (job SJ73/04).

Despite differences caused by sphalerite contamination the Egyptian analyses are very uniform and the trace element contents (especially of Ag and Bi) are quite low. This is consistent with a low temperature of formation (El Kholy et al 1970, El Shazly, Webb and Williams, 1957). The Zug el Bohr and Jabal Dhaylan analyses are very similar hence the

Jabal Dhaylan analyses would presumably be very similar to the Egyptian analyses had they all been made under the same conditions. All these deposits with very low Ag contents are Pb-Zn rich and Ba poor. The remaining analyses also form a group having somewhat higher (but still low) Ag and Cu contents, and represent galenas from the barite veins where Zn is particularly scarce and Pb and Cu are only minor. The slightly high Ag and Cu contents may be common for galenas in barite veins because discussion on the paper by El Shazly, Webb and Williams revealed that this applies also in the English Lakes District. It is possible that some of the copper may be due to contamination as secondary copper minerals have delicately replaced galena along cleavages and it is impossible to be sure of the removal of this by microscopic hand picking and acid washing. Contamination might also account for some of the Ag in the Merataba galena. Boleite was found associated with this galena though after careful search no primary silver minerals were found. It can be seen from the results that all the galenas studied characteristically have low trace element contents, in the range suggested by El Shazly, Webb and Williams (1957) to represent a low temperature of formation, and that within the group as a whole the mode of galena is further reflected in subgroups of low and very low Ag content. The close similarities in composition of these galenas is further evidence that they are co-provincial.

BARITES

The trace elements in several of the coastal barites are shown in table 11. Some variation appears in the results because of the different methods used. The writer used XRF but no barite standard was available (W1 basalt was used instead) hence the results are not meaningful in absolute terms. They do nevertheless show the similarity of the various barites to each other. Standardization factors for Cu and Pb are probably near 1 but for Cr, Ni and Y much smaller factors are necessary. When this is taken into account the XRF results are generally in fair agreement with the optical emission spectrography results. Sr contents were too high for XRF determination without a barite standard however J. Barrett (Imperial College) noted from optical spectrography that all the Sr contents are in the range of 3 to 5% (pers. com.).

Locality		Ca	Cr	Cu	Fe	Mg	Mn	Ni	Pb	Rb	Sr	Y	Zn
Um Gerad	A	100	<5	3	18	<3	6	<2	800		3-5%	<5	
	B		20x	9x				5x	100x	65x		80x	10x
	C	<0.1%	<5	<2	<0.1%	<100	20	<2	150		>1%	<5	
Merataba	A	1300	<5	4	200	45	<5	<2	40		3-5%	<5	
	B		0x	7x				25x	90x	75x		80x	15x
Jabal Dhaylan	A	130	<5	2	80	3	230	<2	400		3-5%	<5	
	B		20x	10x				25x	100x	70x		75x	20x
Yahmum	A	300	<5	8	1000	170	<5	<2	80		3-5%	<5	
	B		20x	10x				25x	95x	70x		80x	15x
Wazra	A	70	<5	8	1000	20	<5	<2	3000		3-5%	<5%	
Um Bogma	A	80	<5	6	50	90	6	<2	40		3-5%	<5%	
	B		20x	8x				35x	80x	65x		75x	15x

Table 11. Trace element contents in ppm and % of barites from Arabian coastal deposits. Analyses A by John Barrett (I.C.) from optical emission spectrography, B by writer from XRF (x is unknown factor due to lack of standard), C by Brobst (1966) from optical emission spectrography.

The results at first may appear to show a wide range of values but compared with the ranges possible the results actually have a relatively narrow range. Ca and Pb can substitute for Ba in barite up to values of 6% and 18% respectively hence values of 1300 ppm Ca and 3000 ppm Pb are relatively low. J. Barrett classified these barites as high Sr barites which probably formed at low temperatures (see section 3.1.4). To a more limited extent Mg, Fe and Mn can also substitute for Ba hence respective values of 170 ppm, 1000 ppm and 230 ppm are relatively low. The ranges of the values for these elements are therefore rather restricted and suggest consanguinity. Values for Cr, Cu, Ni, Rb, Y and Zn are fairly similar, however the range for these elements in barite is probably rather narrow anyway.

The relative similarity of the trace element contents of the barites suggests they are consanguineous and hence co-provincial. This however proves to be an unreliable indication by itself because the Um Bogma barite which has a similar suite of trace elements to the Arabian barites is known to come from another older province (despite many similarities of the deposit with some in the Red Sea province). Furthermore the En

Kafala barite is quite different to those in table 11. Bonatti et al (1972) recorded that it contains 14.9% Sr SO_4 , 0.9% Ca SO_4 and 0.1% Pb SO_4 .

It seems therefore that barites within this sub-province are consanguinous but may or may not differ considerably from barites in adjacent sub-provinces. El Shazly, Webb and Williams (1957) found that 33 barites they studied from mineralized areas and non-mineralized areas, of syngenetic and epigenetic origins, have similar trace element contents (with slight variations in Mg, Ag and V). It would seem therefore that differences in trace element contents reflect different sub-provinces but similarities are not necessarily confined to sub-provinces.

SECTION 2	THE METALLOGENIC PROVINCE OF THE RED SEA	Contd.
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3.2.4 LEAD ISOTOPE STUDIES

Although lead isotopes do not seem to have been used before for defining metallogenic provinces it seemed to the writer that they would be probably suitable for such work because lead has several isotopes whose relative proportions are not readily affected by ephemeral events, but reflect the source of the lead, its age and its radiogenic contamination with lead encountered during migration. The lead isotope study was kindly made by Dr. Ian Swainbank of I.G.S. and his results are shown in table 9 and graphically in fig. 216.

These new results, on samples taken from as many lead occurrences as possible in the province, are in agreement with the limited studies of previous workers (Delevaux et al 1967, Cooper and Richards 1969, El Kholy et al 1970). The leads are all J types and plot on one secondary isochron, and hence can be considered consanguineous. The data are best explained in terms of a homogeneous source modified by a crustal phase. It was also found that even trace leads associated with copper mineralization (at Azlam) and manganese mineralization (at Sharm el Sheik) are consanguineous. All of these leads are different to the leads in the areas to the east (El Kholy et al 1970), west (Delevaux et al 1967) and north (Swainbank 1973) of the Red Sea province. It is noteworthy that Swainbank (1973 and this study) found the Sharm el Sheik trace lead (in a manganese deposit) to be of the Red Sea province but that in the manganese deposits of Um Bogma and Timna to be alien.

The Red Sea geothermal deposits i.e. the sulphide, goethite and montmorillonite facies, have similar leads to the coastal deposits and these, being J types, are typical of Mississippi Valley type deposits (Brown 1970b). This contrasts with results on goethitic sediments from the East Pacific Rise studied by Dasch et al (1971) who found that the lead isotopes in that goethite are of magmatic affiliation rather than marine affiliation. It must be conceded however that in some Red Sea goethite samples a minor magmatic lead contribution is present even though the isotopes are generally more like those in the coastal sediments. The

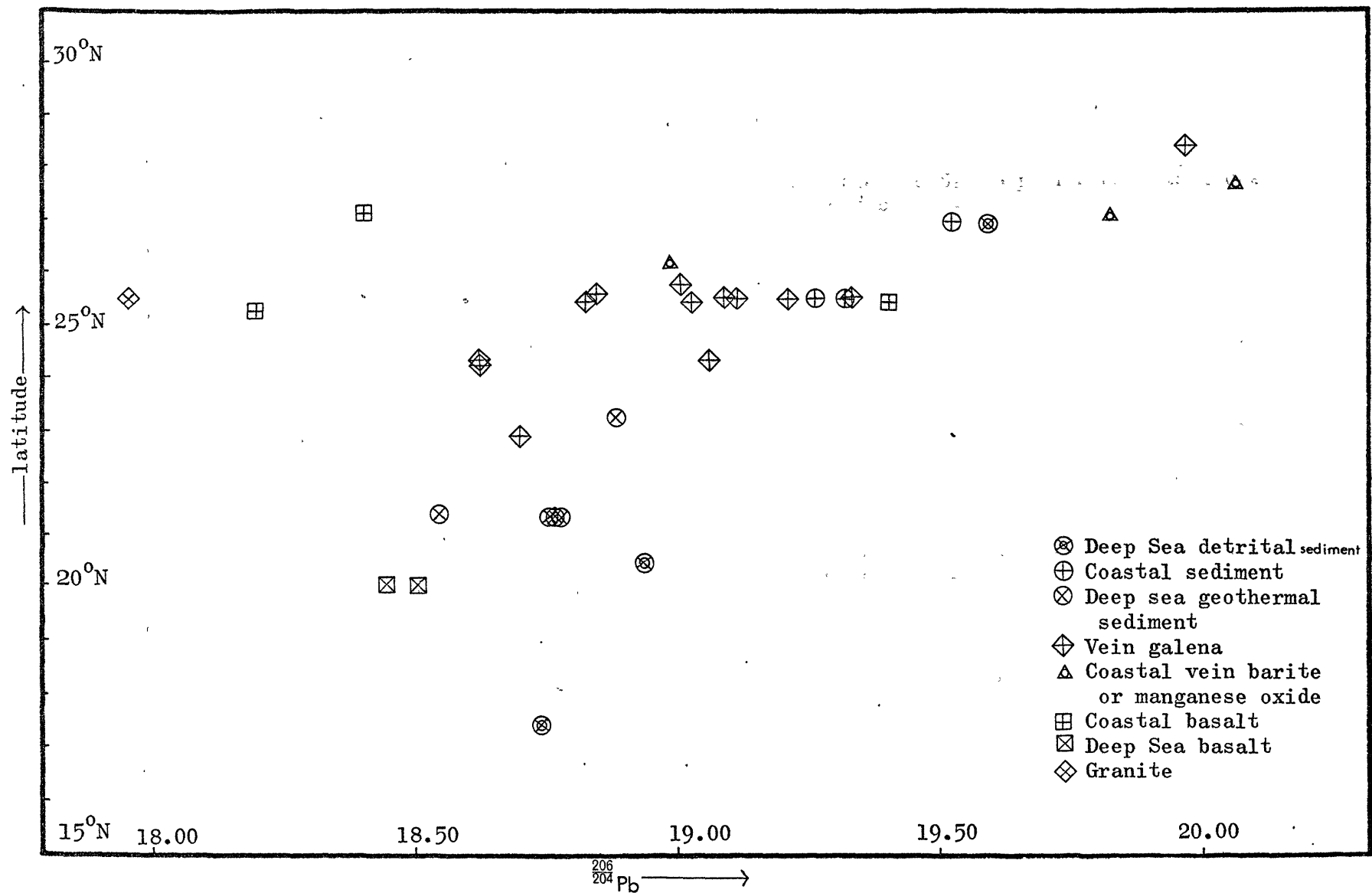


Fig. 217. Plot of radiogenicity ($\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$) of specimens from the Red Sea region versus latitude (using results of I. Swainbank of I.G.S. shown in table 9).

lead in the sulphide facies however is unlike that in the Sudan Deep basalt and is similar to that in the coastal plain, hence it appears from this evidence that the sulphide facies in Atlantis II Deep is actually part of the Red Sea province.

Lead isotopes unfortunately do not necessarily characterize a province even though it seems that a province should contain consanguinous leads. The lead in the Precambrian granite at Jabal Dhaylan also fits the same secondary isochron as the Mississippi Valley type deposits but this lead is obviously older and has probably been a partial source for the lead in the Jabal Dhaylan lead-zinc deposits. The leads in the Jabal Dhaylan basalts also lie on the same secondary isochron. These are of Cenozoic age and although they are tectonically related to the Red Sea metallogenic province they obviously are not part of it.

Although the silver content of the galenas appears to be unrelated to the degree of radiogenic contamination of the lead there is an interesting relation of latitude with $\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$ ratios. This is shown in fig. 217. It can be seen that the radiogenic component of the lead increases in a northerly direction for both the coastal and deep sea specimens. This is more apparent when leads in igneous rocks (which are not really part of the Red Sea metallogenic province) are ignored. This grouping in such a trend appears to be further evidence of consanguinity. The results also indicate that crustal contamination is stronger in the north and has been so probably from the Miocene to the present.

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SECTION 3

M E T A L L O G E N E S I S A S A N A S P E C T O F T E C T O N I S M

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SECTION 3 M E T A L L O G E N E S I S A S A N
A S P E C T O F T E C T O N I S M

3.3.1

INTRODUCTION

Previous sections have dealt with the genesis of individual deposits and groups of deposits within the Red Sea metallogenic province. This section is concerned with the genesis of the metallogenic province as a whole. In this respect the mineral deposits are regarded simply as rock units which are an intrinsic part of the crust in this region and which are an inevitable result of its evolution. After discussion of the broad evolution of the Red Sea crust and the relationship of plate tectonics to Red Sea mineralization a comparison will be made of the Red Sea province with other similar platform regions.

Tectonism obviously has had the most profound control on the development of the Red Sea crust and is given first consideration in the following discussion however it must be noted that crustal evolution is also controlled to a large extent by climate (including sea level fluctuation) and biological activity.

3.3.2

TECTONISM IN RELATION TO MAGMATISM

As so little is known about the basement below the Red Sea the early development of this region is obscure. It does appear however that the initial rifting occurred in Pre-Tertiary time (in the early Phanerozoic according to Dubetret 1970) and some spreading had occurred probably before the Oligocene. Most writers agree that in the Miocene thick shallow water evaporites accumulated, and the present phase of spreading is at least as old as the Pliocene. Although Gass (1972a,b, 1970a,b) considered that the "proto" ocean of the Red Sea formed only in the last 25 million years, his theory on the tectonic-magmatic development of the Red Sea is the most comprehensive and self-consistent so far published.

According to Gass's theory the development of the Red Sea began with thermal instability at the lithosphere/asthenosphere boundary in the Afar area and in an area "to the north" (probably around 20°N). Due to radioactive heating, the crusts tends to act as a heat blanket and once an instability is initiated it develops and grows quickly into an exaggerated thermal perturbation or rising lithothermal system. "The effect of such a system, progressively penetrating into lithosphere, would be to steepen the regional thermal gradient, cause extensive partial fusion at a high level in the lithosphere and also the depression of the main peridotite phase boundaries, thereby producing an increase in volume most easily relieved by uplift" (Gass 1972a). For a crust 30 to 50 km thick the corresponding uplift is about 3 km. Magma initially generated under these conditions would develop at a depth of some 60 km and would be alkalic. Some of this magma would be injected upward along major fracture zones generated by the geodoming and would thereby distort upward the initially flat isothermal surfaces. This would allow magma to equilibrate higher in the mantle to give "intermediate" basalts and per-alkaline silicic differentiates. Repeated injection would further steepen the isotherms so magma could equilibrate even higher possibly at 3 to 5 km (with corresponding lower pressure) to give neo-oceanic tholeiites. Amalgamation of lithothermal systems causes splitting and then spreading

as dykes intrude dykes. This theory accounts for doming and rifting in the Red Sea regions as well as the Eocene Trap volcanism, Aden Series volcanism and neo-oceanic Red Sea volcanism.

SECTION 3	METALLOGENESIS AS AN ASPECT OF TECTONISM	Contd.
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3.3.3 SEDIMENTATION RELATED TO TECTONISM

Post-Miocene sea-floor spreading has added about 50 km of neo-oceanic crust to the width of the Red Sea. As most continental rift zones are around about 50 km wide, 100 to 200 km of the width of the Red Sea has to be accounted for by extensive stretching of the crust or by pre-Miocene sea-floor spreading, or by both. Extensive stretching alone appears to be inadequate and does not account for matching disjuncts either side of the sea hence it is concluded that a pre-Miocene episode of sea-floor spreading has occurred. Geophysical evidence discussed in section 1.1.6 weakly supports this.

Paleozoic and Mesozoic sediments dip away from the Afro-Arabian shield (geodome) and Paleozoic sediments are very rare in the Red Sea region hence it appears that the Red Sea region was emergent in the late Paleozoic and early Mesozoic. Much of the sediment recording the early history of the Red Sea has been eroded away or buried below the thick Miocene sediments. In the Jurassic, the region slowly sank so that clastic sediments accumulated followed by shallow marine limestone and evaporites. From the late Jurassic to the late Oligocene epeirogenic uplift produced predominantly regressive marine and continental sequences. The main period of uplift reached its zenith in the Eocene according to Gass (1970b) on the basis of petrology and in the Oligocene according to Lowell and Genik (1972) on the basis of sedimentation (or rather, the lack of sedimentation). Trap volcanism was extensive in the Eocene and intrusions are abundant in Triassic to Jurassic formations near Jizan hence it is probable that the early phase of sea-floor spreading occurred in upper Mesozoic to lower Tertiary time.

The Oligocene was an epoch of fault block movement, erosion, red bed sedimentation, volcanism and a minor marine incursion. Widespread slow sinking of the Red Sea region began in early Miocene with the formation of carbonaceous marls and then evaporites throughout the whole Red Sea

region. Following this, slow uplift began in the coastal regions and is still active but simultaneously with this coastal uplift the mid-sea region has sunk.

Epeirogenic movements have not been uniform over the entire Red Sea region and small relative differences between the northern and southern parts have profoundly affected sedimentation through connection with the Mediterranean and Arabian seas. Although the Mesozoic was a time of general uplift a slight lowering of the northern part of the region allowed the Mediterranean sea to extend as far as Jeddah in late Cretaceous to early Eocene time. This was cut off by uplift but another minor temporary incursion again as far south as Jeddah occurred in early Oligocene time. Later in Oligocene or early Miocene time most of the present Red Sea region was connected with the Mediterranean sea. During late Miocene to early Pliocene time however a swell in the north and subsidence in the south closed the Mediterranean connection and opened the present Arabian sea connection.

It can be seen that in general the sequence of deposition of the sedimentary rocks more or less follows that which could be expected from the lithospheric activity outlined by Gass (1972a). Up-doming produces regression and erosion. Faulting produces boulder beds, fanglomerates, red beds and restricted continental deposits. Uplift of rift margins results in diversion of clastic sedimentation away from the rift zone. Rift subsidence through stretching with restricted fluvial sedimentation and an arid climate, gives rise to marl, limestone and evaporite sedimentation. Slow subsidence in clear warm water favours reef growth. Corals (e.g. the present madreporians in the Red Sea) tend to grow vertically upward to mean sea level and upon reaching wave base then spread laterally so that relatively large flat areas of reef and marl result. This also favours evaporite deposition. Fringing reefs strive to keep pace with subsidence and when they do so in hot arid regions, they can maintain an ideal environment for evaporite accumulation in the back-reef area and beyond. According to Nesteroff (1955) subsidence is the most important factor in the formation of the Red Sea corals. Thick evaporite sequences fringe the Atlantic Ocean, and Lowell and Genik (1972) claimed that "a thick accumulation of salt is a natural consequence of sea-floor spreading at low latitudes".

This subsiding tectonic environment which favours limestone, marl and evaporite accumulation can also favour hydrocarbon accumulation. Four of the fields in the northern Red Sea-Gulf of Suez area are classed by Halbouty et al (1970) as "giants". They further stated "Graben or half-graben rift basins include some of the richest hydrocarbon yields within the cratonic basin category, particularly where extensive evaporite caprocks (1) cover the source beds and reservoir-rock sequence in a basin, and (2) tend to (a) limit leakage of hydrocarbons, and possibly (b) disseminate the heat from moderately high heat flow centres (below the incipient rifts). Under such conditions, the evaporite seal acts as a 'lid on the kettle'". Silled basins provide optimal conditions for the preservation and accumulation of organic matter. Warm climates favour biological activity and high salt contents of water in restricted basins can further act as a trap by killing organisms introduced from less saline environments (as happens now in the southern Red Sea according to Friedman 1972). The high heat flow in rift regions can free the oil from its source material and assist in its migration. The associated evaporites in these regions can provide diapiric and decollement structures as well as seals for traps. High heat flow and igneous intrusions however can also detract from the oil potential in trailing margin environments.

The situation, however, is not as simple as implied in this discussion. Rifting produces horsts as well as grabens (e.g. by differential subsidence) and the Red Sea region is one of compound rather than simple rifting (Lowell and Genik 1972).

3.3.4

RELATION OF MINERALIZATION TO TECTONISM

Mineralization is directly related to tectonism by meso- and mega-scale structures. It is also indirectly related to tectonism because it is related to sedimentation and magmatism and these are related to tectonism.

Many writers have noted the association of basic alkalic igneous rocks to Mississippi Valley type mineralization and controversy has arisen because some have subsequently attributed the origin of these deposits to magmatic hydrothermal solutions. It can be seen however that within the Red Sea province the two are extrinsically related i.e. mineralization is not an effect of magma intrusion but both mineralization and intrusive activity are the effects of deep seated mantle activity. Heating associated with volcanism nevertheless probably assisted with the formation of the Atlantis II Deep geothermal deposits and some of the material in the deposits may actually have been leached from volcanic rocks. (Mineralization in igneous rocks is outside the scope of this discussion).

The relation of mineralization to sedimentation is fairly obvious. Sedimentary rocks have produced the source, solutes for the mobilization and transporting medium, hydrocarbons for sulphate reduction and environments for mineral deposition.

The relation of mineralization to meso-scale structures is also obvious. Most of the barite mineralization occurs in fissures related to faulting close to the basement or just within it. Mississippi Valley type lead-zinc mineralization is related to horsts and even sulphur mineralization is related to faults. Further structural effects that possibly assisted mineralization are fissures which acted as channel-ways and tilting (and vertical movement) which promoted fluid migration.

The relation of mineralization to mega-structures can be discussed with respect to plate movement. The copper-manganese district of Timna (Sinai - Palestine) has been split along the conservative margins of the Sinaian

and Arabian plates so that part of it is now displaced about 100 km to the north at Fenan in the Dead Sea rift region of Jordan. Deposits separated by constructive margins are less easily related. The Wadi Sawawin iron district appears to be part of a Lake Superior type iron ore province which occurs on the corresponding side of the Red Sea in Egypt.

Abdel-Gawad (1970) found on satellite photographs that two shear zones in Saudi Arabia are remarkably similar to a pair of similar structures in Egypt and he used this and other evidence to show the Red Sea has a dilational displacement of about 150 km in the north (fig. 218) and 225 km

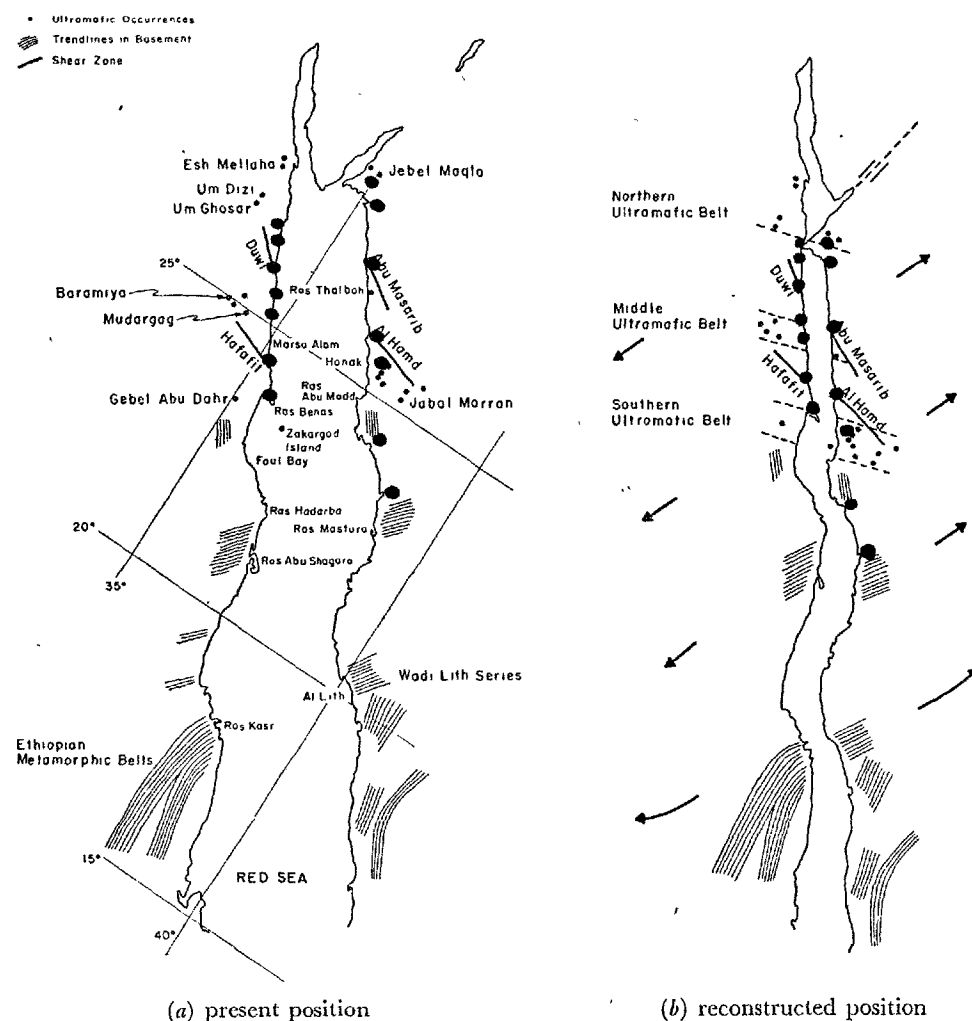


Fig. 218. Displacement of shear zones across the Red Sea from Abdel-Gawad (1970). Additional red dots indicate Pb, Zn, Cu or Ba mineralization.

to the south between Ras Kasr and Al Lith. From field work, the writer wholly concurs that the two shear zones in north-western Saudi Arabia are large and spectacular structures. Furthermore the writer found from maps that the difference in strike direction (about 6°) between the Arabian and Egyptian shear zones is in approximate agreement with the angular displacement (7.6°) of the Arabian plate from the Nubian plate calculated by McKenzie et al (1970) using coastline morphology, magnetic lineations and transform faults (in the Gulf of Aden). Abdel-Gawad's Abu Masarib shear has been termed the Wadi As Sirr rift valley by the writer (section 1.2.1) and the Al Hamd shear occurs at Yahmum (fig. 46). The Azlam mineralized area and a basalt dyke (Az1) are associated with the Abu Masarib shear, and the Yahmum mineralization occurs in fractures connected with the Al Hamd shear. These areas of mineralization were found by the writer in 1970 without knowledge of Abdel-Gawad's work and of course the mineralized areas were unknown to Abdel-Gawad at that time. Corresponding mineralization occurs in the Egyptian shear zones which are considered to be the extensions of the Arabian shear zones. This further supports Abdel-Gawad's hypothesis of dilation. The four mineralized areas in these four shear zones are of Neogene age hence they post-date the initial spreading of the Red Sea. Presumably the controlling precursors of the mineralization were once contiguous.

The early phase of plate separation appears to favour the types of mineralization found in the coastal regions. Following this early phase, mineralization of Atlantis II Deep type became significant and later as plates reach an advanced stage of separation, when the spreading centre is no longer chasmal and associated with shallow-water sediments, new mineralization presumably will develop with similarities to that at the East Pacific Rise and the Mid Atlantic Ridge.

SECTION 3 M E T A L L O G E N E S I S A S A N
A S P E C T O F T E C T O N I S M

Contd.

3.3.5 COMPARISON OF THE RED SEA METALLOGENIC
PROVINCE WITH OTHER PROVINCES

Provinces are usually classified into three types viz. mountain belts, shields and platforms (or stable shelves). According to Stanton (1972) platform provinces typically contain the following types of mineralization

1. Mississippi Valley type Pb Zn Ba mineralization associated with oil and evaporites.
2. Nikopol and Moroccan type Mn mineralization.
3. Colorado red bed type Cu U V mineralization.
4. Minette type iron mineralization.
5. Witwatersrand type Au U mineralization.
6. Kupferschiefer Cu Pb Zn type mineralization and Central African Copperbelt type Cu Co mineralization.
7. Northern Hemisphere bog type Fe mineralization.

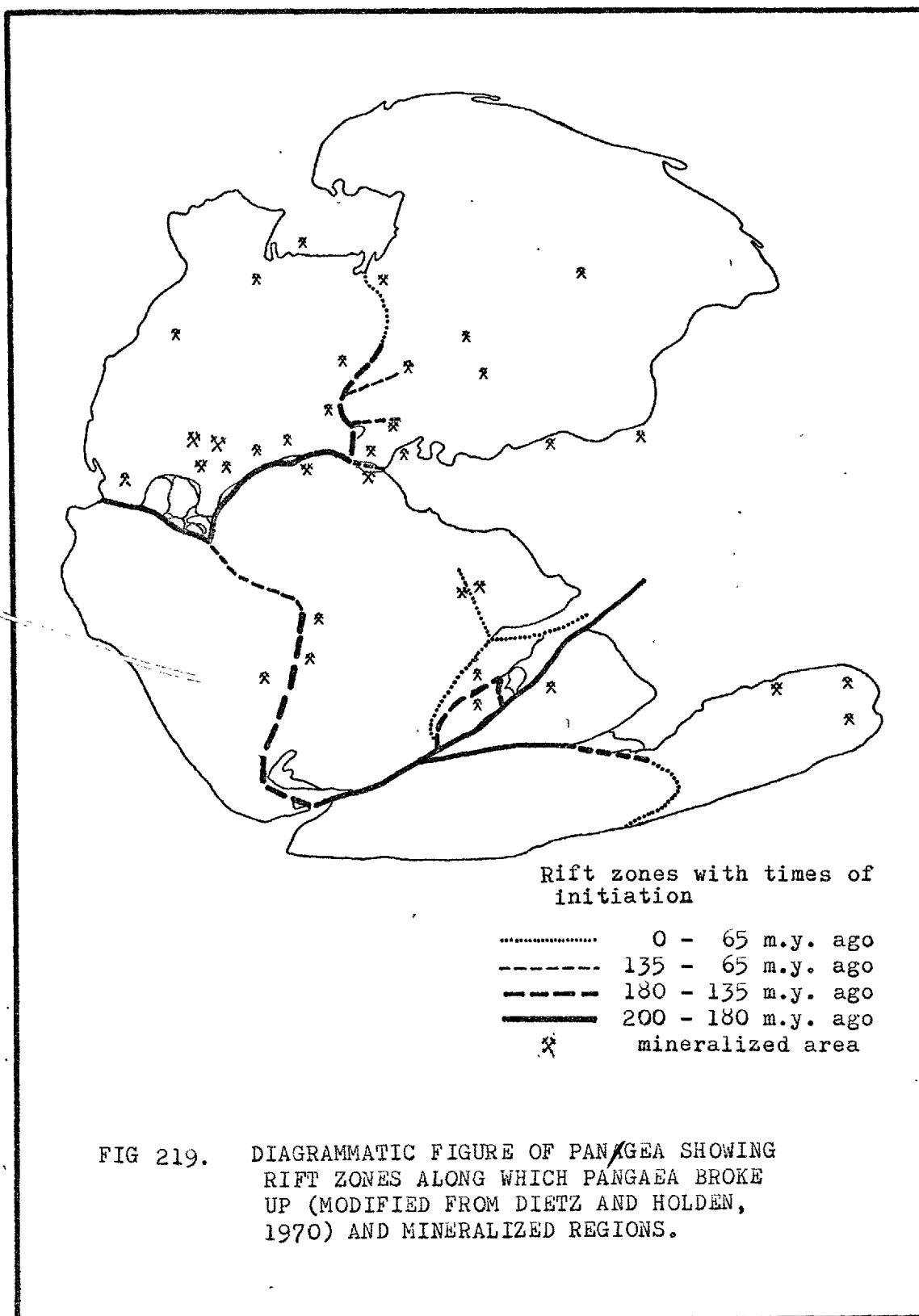
Witwatersrand and African Copperbelt type mineralization are characteristically Precambrian and are therefore beyond the scope of this discussion. Other than Kupferschiefer type mineralization (which may yet to be found in the marine sub-provinces) and Northern Hemisphere type bog iron ore mineralization (which is climatically inappropriate) all the remaining types have been found in the Red Sea province. Phosphate deposits in Upper Cretaceous to Lower Eocene shallow marine sediments occur in the Egyptian Red Sea coast and in the North African region, and deposits of this type should perhaps be included in Stanton's list. To some extent the North African province can be considered an extension of the Red Sea province as it also contains Mississippi Valley type Pb Zn mineralization and Moroccan type Mn mineralization in a similar tectonic environment.

Mississippi Valley type Pb Zn mineralization is usually associated with barite and jasper but commonly the barite mineralization occurs separately to the lead-zinc mineralization. This applies to the Red Sea, Sardinia, English Pennines, Kazakhstan and North America (e.g. Brown 1967, Lindgren

1933). It is probable that barite mineralization is even more widely associated with lead-zinc mineralization than the literature implies because barite is less valuable than sphalerite and, where uneconomic, probably receives little mention in published papers. A similar situation probably also applies with Moroccan type manganese mineralization.

Turneau (1955) noted that Mississippi Valley type Pb Zn Ba F mineralization is the most characteristic type of mineralization in stable platform regions and he further noted that the Provinces of Mississippi Valley, Slave Lake (Canada) and North Africa are very similar to each other. They are also very similar to the Red Sea province so this leads to the question of whether or not this type of mineralization is a normal part of the geological development related to the early stages of plate separation as it appears to be in the Red Sea (Hamilton, 1973).

When a world-wide selection of Mississippi Valley type Pb Zn Ba mineralized areas are plotted on the reconstruction of Pangaea from Dietz and Holden (1970) it can be seen that most deposits occur near divergent plate margins (fig. 219). Furthermore as far as the age of the deposits can be ascertained the time of mineralization generally agrees with the time at which the plates began to separate. Dr. R. Sillitoe (pers. com.) considers that this may be due to increased heat flow that accompanies plate separation. This should lead to a compressed geothermal gradient which would promote diagenesis (and displacement of Pb, Zn from aragonite and high-magnesian calcite) and brine leaching. Slight heating would also mobilize hydrocarbons from source organic matter and may foster formation fluid circulation. This agrees with fluid inclusion data which indicate the ore bearing fluids were generally too warm to have been heated simply by normal heat flow. Stretching and uplift associated with geodoming may also have promoted fluid movement. Mississippi Valley type deposits however will form by this mechanism only in suitable environments where there are undeformed platform type formations deposited in an arid climate. As these formations developed in the Red Sea under conditions of rifting it is very tempting to consider that the deposition of these formations (non-deformed limestones, evaporites etc) in other



Mississippi Valley type areas also occurred under conditions of rifting. Unfortunately present models (e.g. Dietz and Holden 1970) require the dates of plate separation to be generally much younger than the age of the hosts of the Mississippi Valley type mineralization. Nevertheless plate separation in the Red Sea appears to be a two stage (or possibly multi-stage) process and it may be that the separations of the other plates of Pangaea were also multi-stage processes. There are some deposits (e.g. Pine Point) that cannot be related at present to the disintegration of Pangaea so it must be noted that while divergent plate regimes appear to be favourable for the development of Mississippi Valley type mineralization they apparently are not absolutely necessary for it.

C O N C L U S I O N

The mineralization so far known in the coastal plain of Saudi Arabia is relatively sparse and weathered, and it was found that it is best understood using a gestaltist approach. From this study it has emerged that the mineral occurrences can be grouped into various types and these can be related to the geological development of the Red Sea metallogenic province as a whole. The study of minor economically unattractive mineral occurrences has contributed much to the understanding of the larger deposits which can now be seen as normal consequences of crustal development rather than as geological freaks. The study therefore has included not only an analysis of the processes of mineralization but also a compilation of all known mineral occurrences in the Red Sea region (excluding the shield areas) and a synthesis of the whole geological development of this region.

The various aspects of this study are related not only geographically but also tectonically and to some extent biologically. The shield region forms the frame to the coast and has made a minor direct contribution to the coastal mineralization. The coastal regions form a frame to the Red Sea deeps and have contributed to the mineralization there. Iron precipitating microorganisms appear to have contributed significantly to the mineralization in the deeps, the coastal regions and the shield.

Although phytoclast alteration was not successfully developed as a prospecting tool for the Jabal Dhaylan area, it was found to be useful as a geothermometer and as such was applicable at Jabal Dhaylan (and elsewhere) even though fluid inclusion geothermometry was not (due mainly to weathering).

This study has brought the role of microorganisms in processes of mineralization into clearer perspective. Iron precipitating microorganisms appear to have played a direct and important role in the formation of iron deposits in the Red Sea metallogenic province and in the Arabian shield. Sulphate reducing bacteria also appear to have been the means of precipitating some sulphides and possibly native sulphur but isotope study is still necessary for evaluating their full significance. Lead and zinc appear to have been derived from biogenic marine limestone but

the relation of lead-zinc mineralization to microbial action is tenuous.

Perhaps the most important discovery made in this study is that some epigenetic veins can have a biogenic origin. The concept of biogenic veins previously was almost inconceivable. Although the barite-jasper veins in Saudi Arabia are of little economic value in themselves it is now possible to look at other types of low temperature epigenetic deposits in terms of biogenesis.

Although it is generally accepted that the simplest theory that fits a set of facts is the best, seven paradoxes of the Red Sea region which have emerged from this study illustrate the dangers of readily accepting such theories.

1. The sulphide deposits in the Red Sea deeps occur in a volcanic environment which at first would appear to have provided an adequate source for the ore metals but the metals appear to have been derived from shallow-water sediments.
2. Circumstances appear to favour bacterial reduction of sulphate as a source of sulphide ions for the deposits of the deeps and while this is supported by laboratory culturing and isotope studies for Discovery Deep the source of the sulphur in the more significant sulphides of Atlantis II Deep is problematical, despite its young age, and isotope studies suggest the sulphur is non-biogenic and non-volcanic.
3. Iron oxide (or oxyhydroxide) is simply precipitated from the geothermal brines of Atlantis II Deep by mixing with normal sea water yet much of this sediment is biomorphic.
4. Jasper associated with barite and trace sulphides in veins is most probably both epigenetic and biogenic.
5. Two areas of mineralization in the Saudi Arabian coastal region have homologues in the Egyptian coastal region with which they were probably contiguous but these areas were mineralized after continental separation.
6. The Precambrian was previously thought to be relatively desolate but the writer found abundant evidence of microorganisms in Saudi Arabian rocks of this age.
7. Rather than being hostile to life some volcanic environments in which mineralization has occurred have encouraged it presumably with warmth and nutrients e.g. at Wadi Wassat, Wadi Bidah and Wadi Sawawin.

Although this study has dealt with the Red Sea region of Saudi Arabia it can be seen that the wider implications resulting from it are useful to people elsewhere studying metallogenesis in relation to microorganisms, province groupings and tectonism.

A P P E N D I X

PROSPECTING SUGGESTIONS

Although the Red Sea coastal region of Saudi Arabia presently appears to offer little attraction to large metalliferous mining companies it must be noted that the region is still relatively unexplored and viable deposits of lead-zinc, sulphur, phosphorous, and manganese as well as oil fields are known in the corresponding western coastal region of the Red Sea.

This study indicates that the Red Sea metallogenic province has potential for Mississippi Valley type lead-zinc-barite mineralization, Colorado red bed type copper-uranium-vanadium mineralization and various types of manganese mineralization in sedimentary rocks. Further Mississippi Valley type lead-zinc mineralization might be discovered through detailed mapping and geochemical surveying as discussed in section 1.2.3. The potential for discovery possibly increases northwards where hydrocarbon accumulations appear to be more common. The prospect near Maqna appears to warrant further investigation. The writer found no significant radioactivity along the coastal zone but this should not be discouraging because Colorado type uranium mineralization is generally very localized and could easily be missed on a reconnaissance survey. Again the northern parts of the coastal region are possibly more favourable because of the incidence of hydrocarbon accumulations and because radiogenic contamination of galenas (section 3.2.4) increases northward. Traces of copper and vanadium mineralization are known but are easily missed because they occur in friable easily eroded sandstone. Manganese mineralization offers some hope. The occurrence found by the writer in the Wadi Thalba area deserves further examination. It should be noted that Nikopol and En Kafala types of manganese mineralization can be profitably exploitable even if they are only one metre thick provided the mineralization is laterally extensive. The trace of silver at Merataba is anomalous to the Red Sea metallogenic province and begs explanation. This area is worth re-investigating as the Precambrian rocks (especially the amphibolite) may contain further silver mineralization which contributed to the Cenozoic mineralization. Several barite veins were observed by the writer but only one ancient mine. It is very probable that more ancient mines occur in this vicinity however, like

the one found by the writer, they have been covered and hidden by the local inhabitants.

Favourable areas for prospecting may be found by reconstruction of the pre-drift edges of the Red Sea coastlines (as in section 3.3.4). From this the following possibilities are thought to be worth further study:-

- P in post-Paleozoic sedimentary rocks between Duba and Al Muwaylih
(Note traces have been found in Maestrichtian rocks near Jeddah
and in Triassic-Jurassic rocks near Jizan - section 1.2.2).
- Pb-Zn in Cenozoic sedimentary rocks between Jabal Yahmum and Qal at al
Azlam
- Sn in Precambrian rock between Al Wajh and Wadi Thalba
- Be in Precambrian rock between Al Wajh and Jabal Dhaylan
- Ti (ilmenite) in Precambrian rock in the Umm Lajj district
- Au in Precambrian rock in the Bayish district

If the deposition of base metal sulphides in the Red Sea deeps is partially controlled by glacial activity, as this study indicates is possible, then prospecting should take account of the age of the sediments and should not be restricted to areas where hot brines are now present.

Archaeological information from both field and literary sources should be of great assistance to the prospector. The several ancient mines found at Jabal Dhaylan by the writer are possibly mentioned somewhere in records of ancient literature. If these records are found we might then know the extent of ancient mining, the metals sought, reasons for abandonment and localities suitable for further prospecting. Excavation at the ancient dwellings found by the writer may further assist with this. The geography has probably changed since mining ceased at Jabal Dhaylan - which most probably belonged to a much earlier period than the millenium old mining phase at Mahd adh Dhahab. Exposures were probably more extensive in the days of mining and there were certainly more people around who could have noticed metalliferous outcrops before the sands came. Early people presumably were not seeking lead but copper or silver (which was worth more than gold in early Roman times) neither of which are in much evidence today. Some ancient mining sites contain more circumjacent slag than the volume of extant ancient excavations (e.g. Merataba and Wazra) would

suggest so the volume of slag heaps is not necessarily a reliable indication of the extent of mining activity.

Finally the writer wishes to make a plea for the consideration of the local Bedoui in the prospecting and mining business. There are possibly some small prospects, which are of no interest to large companies, that can be profitably worked by small syndicates of local people with a low capital outlay. Although the lead-zinc mineralization at Jabal Dhaylan would hardly interest a big company it still may be possible (since simple open-cut methods could be used) for local people to profitably excavate the high-grade material and transport it to the metallurgical plant at Um Gheig in Egypt where similar ore is presently treated. It is remotely possible that by such activity further concealed mineralization might be located that has so far escaped detection. Possibilities of local mining of manganese ore should also be considered. Serious thought should also be given to fostering local use of gypsum which is abundant throughout the coastal plain and crops out even close to townships. Incidentally, as gypsum has many uses it may also be of interest to large companies.

It is known that peridotites are associated with Cenozoic basalt in Saudi Arabia. Mining of these gemstones, if feasible, should be well suited to local labour. Peridotites are mined from such rocks at Kod Ali and elsewhere in Ethiopia. The economic potential of the Umm Gerad barite deposits has been discussed by Al Shanti (1970) and the potential of the Wadi Fatima iron deposits has been the concern of government for some years e.g. (Zeidler 1970). As the barite specimens analysed for this study contain a few percent of strontium, witherite or strontianite may yet be found associated with the barite mineralization.

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