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# DEPOSITIONAL ENVIRONMENT OF THE PRE-RIFT SEDIMENTS - GALALA HEIGHTS (GULF OF SUEZ, EGYPT)

by
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## ZUSAMMENFASSUNG

Der Sedimentstapel der beiden Galala-Gebirge einschließlich Gebel Ataqa im Norden und dem Nordrand des kristallinen Roten-Meer-Gebirges im Süden (Wadi Dakhal) setzt sich aus sechs verschiedenen Einheiten zusammen, die voneinander durch Ablagerungslücken sowie synsedimentäre strukturelle Veränderungen unterschieden sind. Hierbei liegen die Sedimente aber im wesentlichen konkordant übereinander. Die präkambrische Kristallinoberfläche wird von Karbon- bis Permsedimenten überlagert (erste Einheit), die in Zyklen gegliedert sind. In jedem Zyklus werden terrestrische und fluviatile Sande von marinen Sanden und Tonen, seltener auch fossilreichen Kalken überlagert. Die zweite Einheit besteht aus jurassischen, küstennah abgelagerten Sedimenten, meist Sanden und Silten; die wenigen Kalke sind fossilreich. Im südlichen Bereich ist als dritte Einheit ein fluviatiler bis küstennah abgelagerter Sandstein (Dakhal Formation) entwickelt, der wahrscheinlich in der untersten Kreide entstand. Im nördlichen Bereich setzt die Kreidesedimentation mit fluviatilen und untergeordnet auch marin beeinflußten Sandsteinen (Malha Formation) ein, überlagert von vollmarinen, fossilreichen Mergeln und Kalken des Cenoman und Turon. Diese vierte Einheit setzt sich bis in das Coniac hinein fort, danach herrscht in der Region Abtragung vor. Als fünfte Einheit werden Campan/Maastricht-Kreidekalke (St. Paul Formation) und Maastricht-Sande, -Kalke und -Mergel (St. Anthony Formation) in einem Becken abgelagert, welches nördlich des Südgalala-Gebirges gelegen ist. Eine Verlagerung des Beckenzentrums findet bei der Ablagerung der sechsten Einheit im Paleozän und Eozän statt. Im nördlichen Bereich ist nun eine Karbonatplattform mit kurzzeitig salinar beeinflußten Sandablagerungen entwickelt, während weiter südlich Beckensedimente nachweisbar sind.

Die Sedimentation ist eng mit der Struktur verknüpft: Störungszonen waren zu verschiedenen Zeiten aktiv und zeigen einen direkten Bezug zum Sedimenttypus. Diese Zonen folgen im wesentlichen dem Nordrand des Arabisch-Nubischen Kontinentes und lassen sich in Ansätzen sowohl im Westen als auch im Osten weiter verfolgen.

## ABSTRACT

The depositional environment of the sediments exposed between Gebel Ataqa in the north and the crystalline basement of the Red Sea Hills in the south is unraveled and correlated to synsedimentary structural unrest documented in fault zones which follow the contour of the northern margin of the Arabian Nubian continent.

The six major units 1. Carboniferous/Permian, 2. Jurassic, 3. Lower Cretaceous, 4. Cenomanian/Coniacian, 5. Campanian/Maastrichtian and 6. Paleocene/Eocene are separated from each other by hiatuses and morphology of depositional environment. Each unit can be differentiated by a characteristic set of sediment fabric and litho- and biofacies .

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## 1. INTRODUCTION

Field work 1986 concentrated on sections in the Galala heights, the Wadi Araba in between them and Wadi Dakhal to the south (Fig. 1). Former investigations were carried out by K. BANDEL and J. KUSS in March 1985 in the Western Desert: Bahariya, Farafra, Dakhla and Kharga as well as Gebel Ataqa. The study as a whole can be seen as extension and elaboration of the field work and its analysis of the Wadi Qena area by K. BANDEL, J. KUSS and N. MALCHUS in October 1984 (BANDEL et al. 1987). We could also rely on a number of studies undertaken by several colleagues of the SFB 69. In addition we were able to use data provided to us by E. KLITZSCH (personal communications). Many fossiliferous strata are exposed in the area of the Galala mountains along their slopes to the Gulf of Suez and the Wadi Qena, which allow a correlation of many of the sandstone units with rare fossils found in extensive exposures in Southern Egypt and Northern Sudan. Another major aim of the study is to unravel the stratigraphic positions, depositional environments and paleogeographic positions of the sediments deposited before the onset of the Red Sea rifting, which led to the recent configuration of the region around the Gulf of Suez.

## 2. GEOLOGICAL FRAME

In the Eastern Desert of Egypt, the sedimentary sequence overlies Late and post-Precambrian metasediments and magmatic intrusions which form outcrops between the river Nile and the Red Sea. The overlying Phanerozoic strata are represented by unfolded sediments extending onto the consolidated Arabian/Nubian craton. Intense block faulting combined with multiple reworking of older strata complicate the differentiation of the post-Cambrian sediments: From Cambrian to Middle Cretaceous time, Egypt and Northern Sudan were part of a broad shelf system with shallow water marine environments, interfingering with fluviatile and terrestrial environments. Sediments of different ages show very similar lithologies. They consist of sandstones, siltstones and few shaly or limy intercalations. The components of the sandstones may have been redeposited several times and in different environments. The interbedded siltstones and claystones have also been deposited in different environments and often contain palaeosoils. The temporary existence of carbonate platform sediments with high terrigenous input can be proved in the NE part of Egypt during uppermost Cretaceous and Paleocene/Eocene times (KUSS 1986a, BANDEL et al. 1987).

The reconstruction of Paleozoic transgressions is very difficult, because outcrops are only found near uplift positions of the Precambrian basement. As pointed out by KLITZSCH (1984), sediments of shallow marine transgressions came from NW directions during most of the Paleozoic, interfingering with fluvial deposits coming from southeast.

The stratigraphic data of Paleozoic sediments are obtained by trace fossils (SEILACHER 1983), spores and pollen and few macro-fossils. Cambrian sandstones were found on Sinai and have possible equivalents also in NE-Egypt. According to KLITZSCH & LEJAL-NICOL (1984), Silurian shallow marine strata were found south of Wadi Howar (Sudan), at different localities in the Kufra basin (reaching into Libya) and near Gebel Uweinat. Fluvial-continental sandstones above indicate a Late Devonian age (proved by plant remains).

Carboniferous faunas were recorded by ABDALLAH & ADINDANI (1963) from Wadi Araba/Abu Darag and from the Um Bogma area (Sinai) by KORA & JUX (1986). Age-equivalent sandstones are known from Sinai, Northern Galala, Northern Wadi Qena (BANDEL et al. 1987), Gebel Uweinat (KLITZSCH 1984); they contain plants and *Cruziana*-traces in some horizons and represent the nearshore facies-equivalents of dolomitic limestones with brachiopods, corals and foraminifera.

Permotriassic to Lower Jurassic sediments with arcosic sandstones and paleosoils are known from few outcrops in the Gulf of Suez area (SADEK 1929; ABDALLAH & ADINDANI 1963). Towards the end of the Paleozoic, the area along the Egyptian-Sudanese border became the axis of an E-W trending uplift, from which Paleozoic and Precambrian strata were eroded and transported southward into a strictly continental basin (KLITZSCH 1984). During Jurassic the epirogenetic uplifts were reversed, northernmost Sudan and Southern

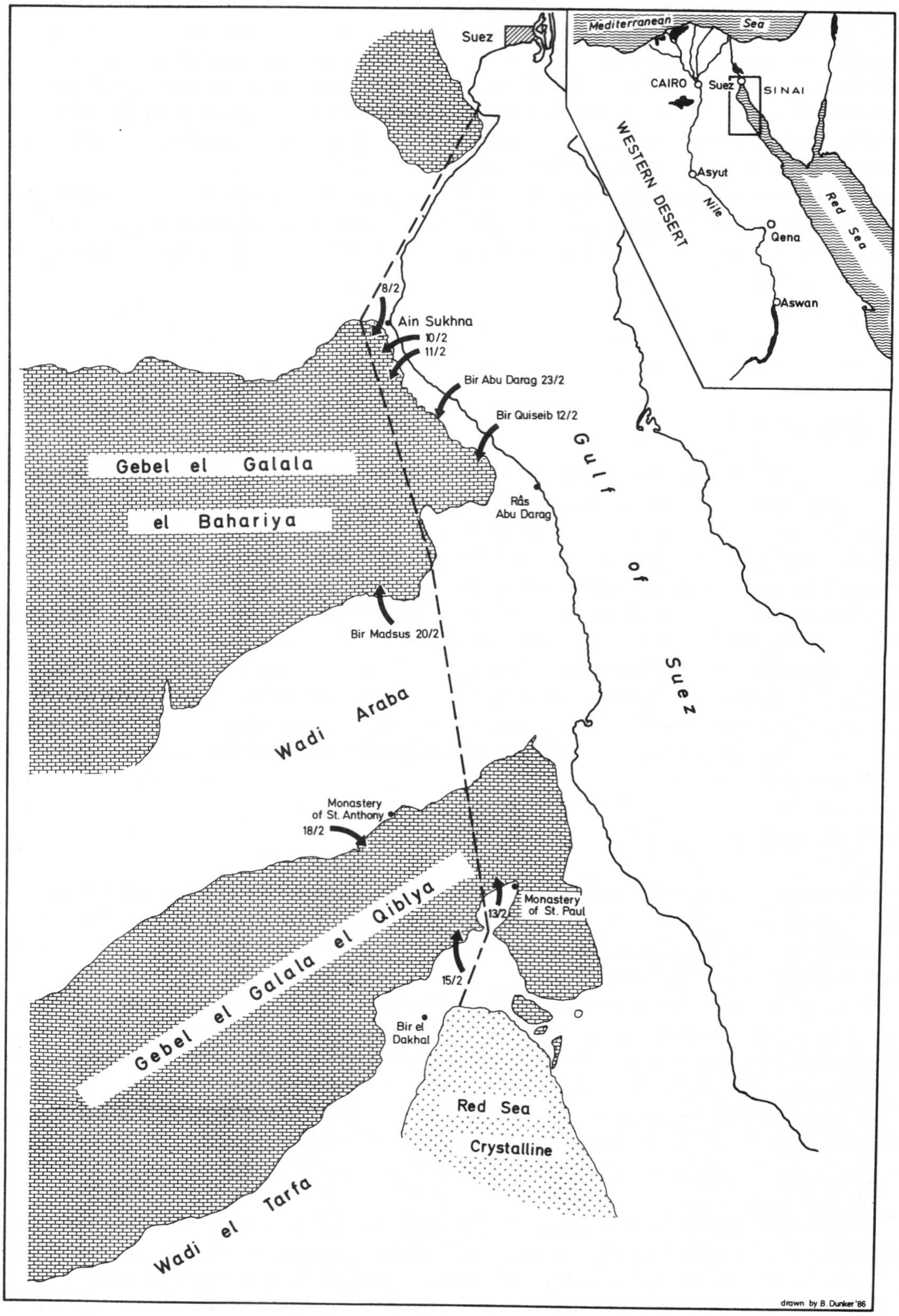


Fig. 1: Location map of the studied area of Northeast Egypt; arrows point to regions where sections were taken, stippled line indicates the profile-line of Fig. 11.

Egypt underwent subsidence and Egypt was flooded by the sea coming from the NE (BANDEL 1981; see here for more lit.). The Late Paleozoic to Early Jurassic strata were partly eroded again during Lower Cretaceous times. From the end of Jurassic wide shallow basin structures (with several 100 km in diameter), situated in the western part of Egypt e.g. Nile basin, Dakhla basin, Dakhla basin, Kufra basin subsided. Later these basins expanded towards the east and were filled with Cretaceous to Early Tertiary sediments, which generally are characterized by northward transported fluvial and deltaic sediments, interfingering with shallow marine strata, controlled by several transgressions and regressions.

The described sections were taken south of Suez, along the two Galala plateaus, which are separated by the E-W running Wadi Araba (Fig. 1). The stratigraphic and facial interrelations of the Paleozoic and Mesozoic sediments from the N-S running Wadi Qena (which follows south of the Southern Galala) were documented by BANDEL et al. (1987).

## 3. THE PALEOZOIC SEDIMENTS

### 3.1 General aspects

Marine Carboniferous strata had been discovered at the Northern Galala by SCHWEINFURTH (1886). The Carboniferous fauna was first described by WALTHER (1890); CUVILLIER (1937) gave a description of the marine Paleozoic strata in the Suez area and determined a Visean age based on the megafauna. OMARA & VANGEROW (1965) pointed out that none of CUVILLIER's twelve species could be accepted as index fossils of the Lower Carboniferous and determined some of these forms as Upper Carboniferous species; they confirmed their stratigraphic revision by cooccurring foraminifera of Wetsphalian age. MAMET & OMARA (1969) assigned the stratigraphic range of the marine Carboniferous sediments of Wadi Araba to the Westphalian C (Early Stephanien). The similarities of these layers (especially from Wadi Araba) with dolomitic limestones of the Um Bogma area (Sinai) led SAID (1962) to the correlation of both Paleozoic marine strata; he postulated a single lithostratigraphic unit, the Um Bogma "series" which in the type locality however is of Visean age (OMARA 1965). The boundary between both stages is assumed to occur in beds of Nubian facies (SCHÜRMANN et al. 1963).

### 3.2 Cycles and exposure

The exposures of Carboniferous rocks in the Eastern Desert are characterized by sandstones, shales and dolomitic limestones. Marine trace- and body fossils as well as terrestrial plants occur from close to the base of the sequence at Wadi Dakhal in the south to close to its top in the northern North-Galala (Fig. 3, section 10/2). The style of deposition, the type of sediment and the trace fossil assemblages (Pl. 1, Figs. 1-6) are similar from base to top. The total thickness of the sequence varies from 250-300 m in the northern outcrops (northern edge of the Northern Galala plateau) to 25 m in its southernmost occurrence at Wadi Dakhal, where the crystalline basement of the Red Sea Hills formed a palaeorelief resulting in a nearshore on-lap contact of the sediments. The sedimentary sequence is characterized by cyclic sedimentation with several fining-up sequences of 10-20 m thickness. A conglomeratic sandstone commonly forms the base of one cycle, while the top is dominated by limestones/dolomites and clays. A generalized complete cycle, compiled from several sections taken along the two Galala plateaus, is reconstructed in Fig. 2.

The base of each cycle is formed by coarse, large-scale cross-bedded sandstones, gravel beds and characteristic grading within single cross-sets (1). Layers with mud balls, rib-up clasts and driftwood are intercalated. These predominantly fluviatile sandstones are overlain by sandstones deposited in the intertidal environment as indicated by mm-scaled flaser- and herringbone cross-bedding (2). Root horizons often mark the top of this unit (3). Flaser-bedded silt and sandy intercalations with few bioturbations follow (4). Platy sandstone intercalations above (5) occasionally preserved the imprints of strong

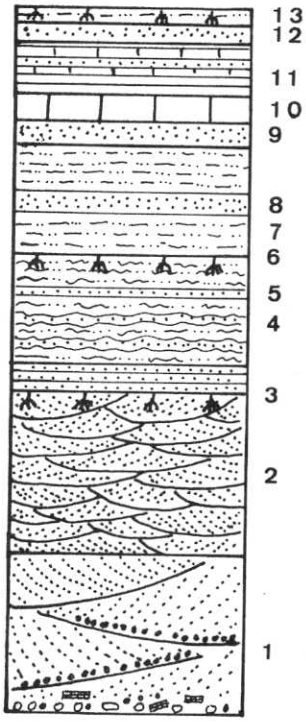


Fig. 2: Idealized Carboniferous cyclic sequence with 13 different units, which are explained in detail in the text.

currents as well as *Cruziana*-like trails, *Chondrites*- and *Planolites*-like burrow systems and the resting and crawling tracks of bivalves (Pl. 1, Fig. 1). The silty unit may end with soil and root horizons intercalated with bleached beds and iron-oxidic crusts (6). The overlying silty shale (7) contains layers with numerous terrestrial plant remains as well as such with marine bivalves and brachiopods. The sandstone beds above are bioturbated and their surface may be covered with a diversified fauna and trace fossils (8). *Zoophycos* may occur here, but it is more characteristic for sandstones (9) which occur just below the basal limestone bed. This lower carbonate bed (10) is commonly the thickest and contains crinoid debris. Within the bedded sandy-dolomitic sequence on top of this limestone (11), depositional environment shifted between the intertidal zone and the off-shore area where calcareous beds with brachiopods, bryozoans, crinoids, worm tubes, gastropods, bivalves and single corals were formed. Sandstones of the cycle 12 contain many trace fossils like those present below the carbonates (8, 9) (Pl. 1, Figs. 3, 5).

While the Carboniferous cycle up to here (11-12) represents a transgressive sequence (with increasing marine influence), the topmost unit (13) has a regressive character, indicated by soil- and root horizons and silt pebbles.

Within the Carboniferous/Permian deposits, several complete cycles can be recognized, e.g. near Bir Madsus (south scarp of Northern Galala), at Wadi Bikheit (northern rim of Wadi Araba) and west of Bir Abu Darag, Bir Aheimar and Bir Ma Sweilim (all at the eastern escarpment of Northern Galala). More common are incomplete cycles in which the upper portions are missing (Fig. 3, Bir Madsus). The most frequent cycles only consist of units 1 and 2 of the ideal sequence as shown in Fig. 2, respectively fluvial and intertidal sands, with more than 20 m thickness. Many of the massive sandstones represent channel deposits, erosively cutting into the underlying bed. In the central portion of Bir Aheimar - Bir Ma Sweilim sections, one of the channel systems, attaining a maximum thickness of 80 m, thins to 10 m only 3 km to the south, thickening again to 40 m further 2 km to the south. Lateral variation of thickness is very extreme within short distances. Thus the unit of coarse sandstones of more than 100 m thickness at the monastery of St. Paul, topped by a characteristic marine sequence with badly preserved crinoid ossicles and *Zoo-*

*phycos* in the silty shales below, cannot be correlated with beds exposed in any other Carboniferous/ Permian section of the Wadi Araba, the Northern Galala, or with the sandstone sequence of more than 25 m at Wadi Dakhal (Fig. 1).

### 3.3 Carbonate beds

The marine horizons from Wadi Quiseib and Ain Sukhna at the top of the Carboniferous cycles were investigated in detail with respect to their microfacies:

#### 3.3.1 Dolomitic limestones

The dolomitic limestones from unit 10 (Fig. 2) are 120 cm to 280 cm thick; the overprinting dolomitization is marked by an intense, step-like interfingering of single dolomite rhomboeders. The intercrystalline pores are either filled with later calcite or with reddish/black, iron-rich opaque minerals. Also, less common, conspicuously zoned dolomitic ankerites occur; the dolomite crystals show uniform, unzoned, yellowish/brownish colours. Most of the former 70-90 µm thick dolomite crystals are affected by a later dedolomitization; this neomorphous transformation resulted in the enlargement of the crystals to 400-550 µm and the growth of numerous hematite cubes as relicts of ferroan dolomites (ELMORE et al. 1985).

Multiple diagenetic transformations resulted in a bad preservation of all fossils with original low-Mg-calcite or aragonite composition. Calcite brachiopod shells and their spines, debris of echinoids, few archaediscid foraminifera (with iron-rich fillings of the chambers) and phosphatic shells remained basically unaltered. The limestones contain up to 30 % angular to subangular quartz grains (180-270 µm in diameter) and less than 1 % glauconite.

#### 3.3.2 Limestones

The limestones of a complete unit 11 form 40 to 180 cm thick beds and consist of biosparites which have subsequently been recrystallized. Terrigenous angular to subangular quartz grains of 50-140 um size as well as sand-sized rock fragments compose up to 35 % of the rock material. Well rounded and 350-500 um large sand grains in contrast, are surrounded by sparitic cements which have probably originally been fibrous aragonites. Several of these larger sand grains were fixed together by micritic crusts, which include small encrusting foraminifera of the *Nubecularia*-type. The larger agglutinated components resemble grapestones (Pl. 3, Fig. 6).

Well rounded quartz grains and many fragments of biota (especially punctate brachiopods) are also encrusted by thin micritic films. Small tubes (15-20 µm thick) of possible algal or bacterial microborings are visible at the inner linings of these crusts. Borings grade into undifferentiated micritic crusts (220-480 µm thick), containing also encrusting foraminifera with characteristic half-moon-shaped chambers. Some of the originally dark, micritic crusts have been transformed into reddish layers of iron oxide. These ferruginous crusts were recognized at other components, while neighbouring ones of the same type are free of iron crusts. This indicates multiple redeposition of the components after prior diagenesis with iron-impregnation and subsequent erosion.

Besides, also thicker cylindrical borings (80-190 µm) which deviate from the pattern of brachiopod punctae were possibly formed by sponges.

All former micritic areas of bioclasts are replaced by microsparitic crystals. Former sparitic portions of shells have been recrystallized and crystal size increased. Many of the former peloids can only be recognized as ghost structures. Most of the echinid fragments are surrounded by thin crusts of syntaxial rim cements, which indicate an early cementation, possibly with fresh-water subsurface diagenesis (FLÜGEL 1982). Single reddish/brownish dolomite rhomboeders were formed during late diagenesis.

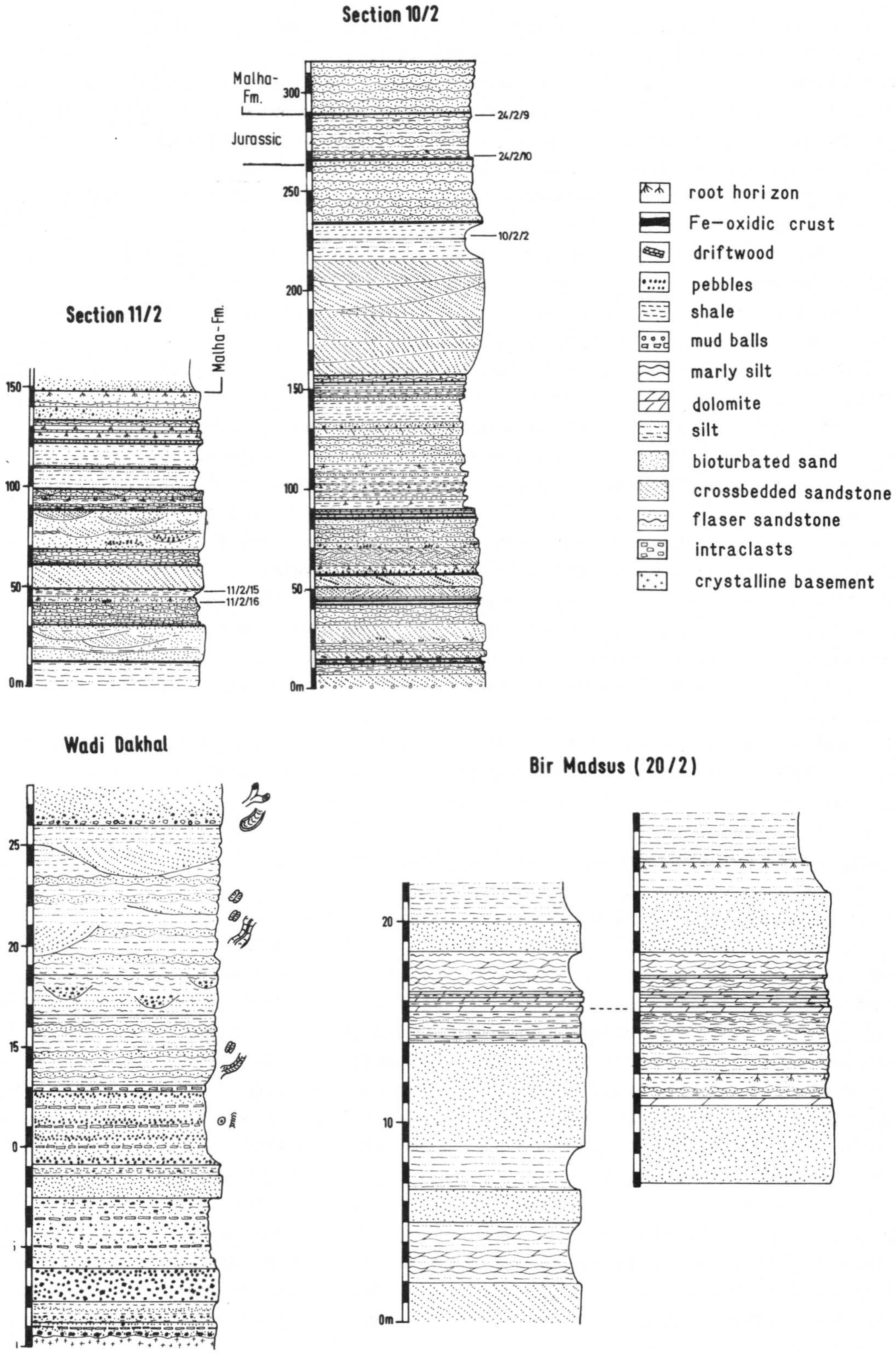


Fig. 3: Carboniferous sections taken from different localities of Northern Galala (Fig. 1); detailled sections of Bir Madsus were taken with few 100 m distance and illustrate the lateral facies changes.

The following biota could be recognized: The most frequent components are spines, arm plates and undifferentiated plates of echinoderms (up to 50 %). Remains of bivalves, gastropods and brachiopods represent between 35 and 40 % of the total fossil contents. Ostracodes and calcisphaerulids make up 3-5 %, while the first are more common. Some layers are rich with colonies of bryozoans. Solitary rugose corals of the genus *Amygdalophylloides* (det. HERBIG) indicate Late Carboniferous to Early Permian age; the genus is most common in Westfalian strata. Fragments of silicified wood, of phospatized fish remains and fragments of the alga *Koninckopora* make up 5-7 %. Several species of benthic foraminifera are present:

The most frequent foraminifera belongs to the genus *Hemigordius* (Pl. 3, Figs. 2-5) with typical trochospiral coiling of the first chambers and a planispiral (involute) coiling of the later whorls. Another type could be determined on species-level: *H. harltoni* described in detail by GROVES (1984). This form indicates a Pennsylvanian age in Northern American rocks.

Attached foraminifers of the genus *Calcitornella* and *Calcivertella* occur which show a nonseptate enrolled inner portion and irregularly uncoiling and winding across the attachment; the actual forms of the tests largely depend on the surface of the attachment. The calcareous, porcellaneous *Calcitornella* cf. *elongata* as well as *Calcivertella* sp. are characteristic of deposits of Pennsylvanian age (TOOMEY 1983). *Monotaxinoides* sp. with a discoidal slightly concave test shows planispiral coiling, non-septate second chambers and a calcareous wall. *Pseudobradyina* cf. *pulchra* with originally fine-granular, calcareous walls is commonly iron-impregnated. Its planispiral, involute test with few coiled chambers increases rapidly in size. According to TOOMEY (1983) it indicates Pennsylvanian age in North America. *Endothyranella* sp. (Pl. 3, Fig. 9) with a fibrous, fine-granular calcareous wall and a partially involute enrolled test, and *Endothyra* sp. (Pl. 3, Fig. 9) with bigger, thick-walled tests and a typical first trochospiral and then planispiral coiling are rare. *Tetrataxis* sp. (Pl. 3, Fig. 10), *Glomospira* sp., some sand-agglutinated forms (? *Bigenerina* sp.) and few bad preserved fusulinids, possibly belonging to *Pseudostafella* were recognized.

The washed-out foraminifera described by OMARA from age-equivalent pelitic deposits in several publications (e.g. OMARA & KENAWY 1966) yielded abundant arenaceous forms, which are stratigraphically less important in contrast to those from limestones and dolomites.

### 3.4 Depositional environment of the Carboniferous limestones

The depositional environment of the limestones present in several cycles of units 10 and 11 (Fig. 2) can be reconstructed, considering diagenesis as well as biotic and abiotic components. The first thick dolomitic limestones (unit 10) represent near shore carbonates with a characteristic fauna. These carbonates were affected by dolomitization and dedolomitization during early diagenesis. Fluviatile sandstones below contained much ferruginous material which could be mobilized and transported by pore solutions to the limestones, resulting in the formation of ferroan dolomites. Dolomites were also precipitated as late cements, according to a model presented by LAND & DUTTON (1978). Dedolomitization was probably caused by oxidizing fluids with high calcium contents which migrated through the pore spaces, as interpreted among others by ELMORE et al. (1985). As pointed out by AL-HASHIMI & HEMINGWAY (1973), ferroan dolomites are preferentially replaced by calcite under such near-surface and oxidizing conditions.

The slightly dolomitized younger limestones of unit 11 were protected by silt-, marl- and clay beds of unit 10, which prevented the exchange of fluids rich in iron from the sandstones below. Here aggregate grains and micritization of components indicate subtidal and intertidal shallow-water environments with restricted water circulation (about 10-30 m water depth). MONTY (1967) described comparable algal lumps formed in shallow waters of the tidal zone; grapestones are also known from algal mats of sea-marginal hypersaline pools (FRIEDMAN 1978):

Limonitic crusts were preserved in unit 11 and formed as inorganic ferruginous precipitations on organogenic substrates comparable to algal mats or algal encrustations within the shallow subtidal to supratidal environments (FÜRSICH 1971).

Limestones thus, as well as the high input of clastics, the occurrence of glauconites and the presence of multiple reworking indicate a shallow subtidal (sometimes supratidal ?) depositional environment.

### 3.5 Flora

A well preserved flora was found at Bir Quiseib in a small quarry besides the track near the entrance of the wadi, south of section 11/2 (Fig. 3) in a horizon which definitely overlies the Carboniferous limestones and dolomites. Here abundant branches of Coniferophyta (Pl. 2, Fig. 1) occur together with less common leaves of Pteridophylla (Pl. 2, Fig. 4). A. LEJAL-NICOL briefly described the flora in an appendix at the end of this work. The stratigraphic range is uppermost Carboniferous - Early Permian.

### 3.6 Paleogeographic setting of Carboniferous/Permian sediments

The base of the Carboniferous/Permian sequence is exposed in the outcrops at Wadi Dakhal and at the eastern escarpment of the northern Wadi Qena depression (Fig. 3). Here it overlies a smooth paleorelief on top of the crystalline basement; reworked soil held angular rock fragments as well as rounded quartz pebbles. Many of these pebbles show wind-polished facets indicating that fluvial pebbles were sandblasted during arid periods, prior to the Carboniferous transgression. Trace fossils produced by arthropods of a quite characteristic shape were found in the section near Bir Dakhal and 5 km to the south of the upper Wadi Um Arta as well as between Wadi Qena and Wadi El-Tarfa near Bir Ghalla (BANDEL et al. 1987). The Carboniferous age of the sediments at Wadi Dakhal cannot be doubted, since E. KLITZSCH (pers. comm.) found *Lepidodendron* at this locality and the trace fossil assemblage does not change farther up in the section. At the Bir Madsus area, the same *Cruziana*-like arthropod trails and resting tracks were discovered well above brachiopod-rich marine carbonates within the last tow cycles of the about 200 m thick sequence exposed there, also together with *Lepidodendron* (Pl. 2, Figs. 5-7). The coast line of the Carboniferous/Permian sequence was situated south of the Um Bogma (Sinai), south of Bir Dakhal to the east of the Red Sea Hills and south of central Wadi Qena. This occurrence indicates a coast line running in a general direction which followed the margin of the Arabian-Nubian continent from NE to SW. The presence of an ancient "Ras Gharib Gulf" as reconstructed by SAID & SHUKRI (1955) and taken up by SAID (1962) cannot be confirmed.

## 4. JURASSIC SEDIMENTS

### 4.1 General aspects

Along the eastern escarpment of the Galala el Bahariya, several sections of the Jurassic sediments have been measured (Figs. 1 and 4), based on the reports of SADEK (1926) and KOSTANDI (1959). ABDALLAH et al. (1963) described Mesozoic rocks from the western Gulf of Suez; it is assumed that their Permo-Triassic Quiseib Formation (43,5 m thick) - which does not contain index fossils - coincides with the lowermost member of the Jurassic section described here. The first mentioned authors determined Bajocian-Bathonian faunal assemblages, which they compared with faunas from Somalia and Ethiopia.

A brachiopod fauna of Bathonian age was found (and is still in work), which can be compared with that of the Jurassic sediments of Jordan (BANDEL 1981). The most important constituents of the limestones from outcrops east and south of Ain Sukhna were determined by microfacial studies. Two main microfacies types can be distinguished.

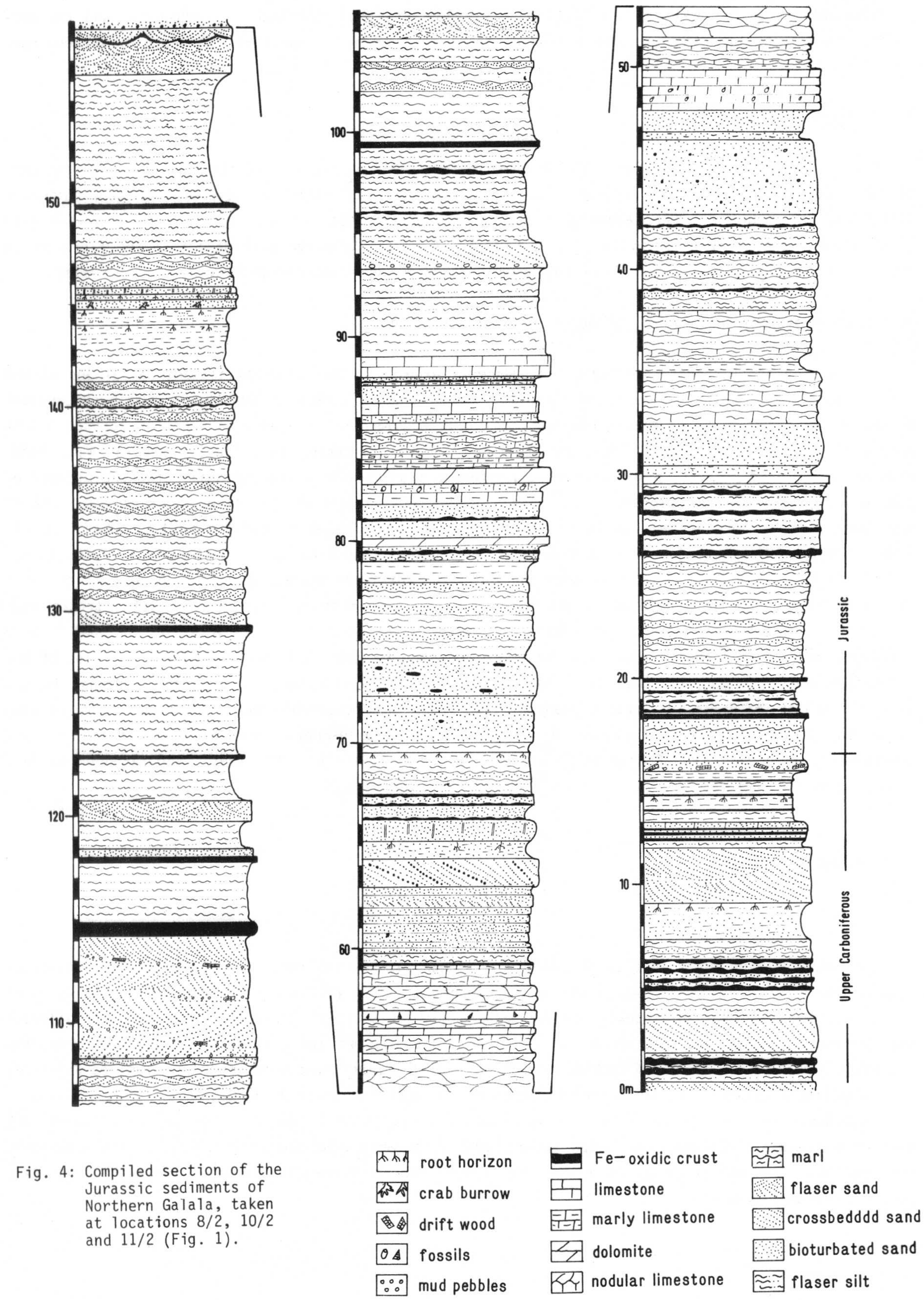


Fig. 4: Compiled section of the Jurassic sediments of Northern Galala, taken at locations 8/2, 10/2 and 11/2 (Fig. 1).

## 4.2 Limestone facies

### 4.2.1 Biomicrites / rudstones

They are composed of lumachelle layers with bivalve debris, some lamellibranchs are preserved with both valves together. Besides, only few brachiopods and bioclasts occur. All pores and fossil cavities are filled by micrite. It is supposed that these facies-types were formed in shallow, unagitated water.

Conspicuous secondary pores were formed by selective leaching of most of the primary aragonitic molluscan shells; tests with thin micritic crusts remained more stable (Pl. 3, Fig. 13). The new-created pore spaces were gravitatively filled with silt-sized crystals and few sand-sized, irregularly recrystallized particles of unknown origin, which also occur (with lower frequency) within the "autochthonous" micrites of the sediment. The remaining open cavities were filled with coarse blocky cements. Some of them show clear radiaxial-fibrous structures (Pl. 3, Fig. 14), which are indicators for phases of emersion or vadose zones (BRANDNER 1978). Thin crusts rich in clayey, opaque particles penetrate the outer parts of primary components. These crusts are intraformational linings, running irregularly subparallel to bedding planes with distances of 15 - 20 cm. They have formed during influxes of fresh water, which is confirmed by the presence of big, well rounded quartzes within only these thin zones. The diagenetic environment changed for a short time from marine cementation to a fresh water vadose zone with leaching, internal sedimentation and subsequent blocky cementation. The overlying mollusc debris is again dominated by marine cementation.

### 4.2.2 Biosparites

#### 4.2.2.1 Gastropod grainstones

The most frequent components are gastropods (60-70 %), followed by echinoderm fragments (22-30 %) and micritic extraclasts (15-18 %). All these are rounded to well rounded and most of them have thin, brownish oolitic crusts.The walls of most gastropods (which are filled with micrite) are micritized, with different intensities, up to total micritization. They lived originally in muddy, quiet water environments, and were later reworked within a high-energy environment and mixed together with other debris. The thin oolitic crusts with concentric lamellae were formed within agitated water. Besides, several oncoids (which are described in the next chapter), a few remains of algae and *Lenticulina* sp. were observed.

#### 4.2.2.2 Mixed grainstones

These limestones show areas with a micritic matrix, but more frequently the former matrix is washed-out. Micritic intraclasts give evidence for the interfingering of a biomicrite and a grainstone facies. 13-15 % quartz and rock fragments indicate strong terrigenous influence; clasts can be covered by irregular, micritic crusts which resemble to thin algal lumps.

Two different types of oncoids occur: one with a thin sparitic network of *Bacinella* cf. *irregularis*, a problematic algal remain described by RADOICIC (1966) (Pl. 3, Fig. 12). Similar *Bacinella*-oncoids (with five different sub-groups) were described by MALCHUS & KUSS (1987) from Bathonian limestones of southern Portugal. The second type is composed of dark micritic crusts without any concentric structures and with definite transitions to peloids and cortoids. Both oncoid types and grapestones compose up to 25 % of the limestones. Less than 5 % are normal ooids with tangential structures and few oo-grapestones. Fragments of lamellibranchs (3-5 %), brachiopods (1-2 %) and echinoderms with syntaxial overgrowth (up to 6 %) occur together with the foraminifers *Lenticulina* sp., *Rheophax* sp., *Placopsilina* sp., *Pseudocyclammina* sp. and different textulariids (Pl. 3, Fig. 20).

Algae are frequent, but not well preserved. At least three different species of the genus *Rivularia* are present; based on the classification of DRAGASTAN (1985) only *Rivularia haematites* could be determined

(Pl. 3, Fig. 16). Other cyanophytes micritized the substrates by boring, creating a micritic microfabric. They may belong to the genus *Schizothrix*. Some of these crusts are represented by other cyanophytes such as *Phormidium*, similar to fresh water oncolites (LEINFELDER 1985). Fragments of dasycladacean algae of the genus *Cylindroporella* are frequent but bad preserver; they are very similar to *C. arabica*, described by ELLIOTT (1957) from the Upper Jurassic of Iran (Pl. 3, Fig. 18).

All three types of microfacies represent shallow water environments. Their distribution is limited to the proximity of the shore line, corresponding to WILSON's (1976) facies zones 6 - 7. Neither sediments of a platform slope nor of the uppermost inter-supratidal areas are present. Terrigenous material, vadose diagenesis and the occurrence of possible fresh water cyanophytes support this reconstruction.

The correlation of this near-shore facies across the Gulf of Suez and the Sinai with the well studied sections in the Negev (GOLDBERG & FRIEDMAN 1974) and in Jordan (BANDEL 1981) has still to be elaborated.

### 4.3 The Khashm el Galala section

The Jurassic section exposed at Khashm el Galala is about 250 m thick (Fig. 4) and can be subdivided into 3 members.

The lower member, 30 to 35 m thick, is well exposed at the Khashm el Galala and to the south of it, topping the Paleozoic sequence as far to the south as Bir Aheimar (Fig. 3, section 10/2; Fig. 4 - left). This lower member consists of intercalations of cross-bedded coarse sandstones, flaser-bedded silt and sand, destratified silt, sand and clay with root horizons and soil, dolomitic sands and ferruginous to dolomite beds. The sandstone beds rapidly change in thickness, sand commonly fills channels, cut into flaser-bedded sandstone sequences. Bleached and weathered beds cooccur with ferruginous crusts and intraformational conglomerates and may show root horizons. Some dolomitic beds (especially towards the south) have laterally been disintegrated into pebbles, due to synsedimentary erosion, which now form layers within the sandstones. The sandstone units may either show cross-bedding, mud pebble layers and contain driftwood, or may be destratified due to bioturbation. The sedimentological characteristics point to a deposition in near-shore environments of a intertidal, rarely shallow subtidal foreshore to backshore facies.

The second member of the Jurassic sequence, about 60 m thick, is found only in the Khashm el Galala. It was studied in the slopes east of Ain Sukhna hotel as well as about 3 km to the south of it. Farther to the south, this member as well as the upper member were eroded before the beginning of Cretaceous sedimentation. It consists of intercalations of fossiliferous marls, limestones and dolomites with sandstones and silty shales. Some fossiliferous beds contain conspicuous bivalves like oysters, mytilids and pholadomyids and some gastropods. A very characteristic trace-fossil assemblage is dominated by crab burrows of the *Ophiomorpha* type. Most of the sandstones and calcareous beds are intensely bioturbated, indicating a deposition in shallow near-shore environments. This is also proved by microfacies investigations of several limestone beds which, as described in chapter 4.2, indicate a sedimentation under shallow subtidal conditions.

The upper member is about 70 m thick and consists of intercalated silt and sand with flaser structures. The grey colour changes into purple only a few metres above the central member. Several originally calcareous beds, some with a fauna of small gastropods and bivalves have been transformed into iron oxide during diagenesis. Trace fossils are present throughout, with an increasing density towards the base, respectively the central member. Only rare bioturbations occur within the flaser-bedded units. Thicker sandstones show vertical burrows with pipe-rock structures and large *Ophiomorpha*-like crab burrows. Root horizons are more common in the upper portion of this member and driftwood as well as rib-up clasts were found at the base of the sandstones. The sediments of the upper member have been deposited predominantly in a tidal environment.

Aside from the fossiliferous beds of the central member, which clearly indicate Middle Jurassic (Bathonian) age, the Jurassic rocks can be distinguished from the Paleozoic sequence below by a characteristic different trace-fossil assemblage, predominated by crab burrows. Dolomitic and ferruginous horizons tend to be more massive and more destratified by bioturbation than beds of similar lithology in the Paleozoic sequence below. Otherwise, sandstones and flaser-bedded silts and sands are very similar in both sequences.

## 5. CRETACEOUS SANDSTONES

### 5.1 Dakhal Formation

The Carboniferous/Permian sequence in Wadi Dakhal (Fig. 1) is overlain by a massive, about 80 m thick cliff-forming sandstone unit, which in its uppermost beds contains some fossils. About 10-15 m below the overlying marine Cenomanian beds (and a series of 12 m of silt and shale with bivalves), a sandstone bed holds abundant well preserved trunks of tree ferns (? *Weichselia*). On remains of *Weichselia*, E. KLITZSCH (pers. comm.) considered these sandstones down to their base to be of Early Cretaceous age.

The unfossiliferous bulk of this sandstone unit contains trace fossils which are only visible on freshly eroded surfaces in the base of the wadis. Here burrows are visible that form a network of tubes with diameters of 0.6 to 2 cm, which show a characteristic lining of fecal pellets as found in some crustacean burrows. These sands may well be of Cretaceous age, as suggested by KLITZSCH, since crustacean burrows of this type are common only from Jurassic times onward. Massive sandstones found in the northern and central Wadi Qena, had been placed into the Ordovician by JUX & ISSAWI (1982). This view cannot be supported since these sandstones rest on Carboniferous/Permian sediments. BANDEL et al. (1987) observed that these Cretaceous sandstones which they named Dakhal Formation, conformably overlie similar Paleozoic sandstones and, like these, have partially been eroded prior to the deposition of the Upper Cretaceous sediments in the southern Wadi Qena. Thus an Early Cretaceous age is very likely for the massive quartz sands at Wadi Dakhal and northern Wadi Qena.

At St. Paul monastery, Cenomanian deposits overlie Paleozoic sandstones without the intermediate sandstones. At the southern escarpment of the Northern Galala, in contrast, a characteristic pink to purple sandstone unit is found below the fossiliferous Cenomanian sediments (Fig. 5). This unit was termed Malha Formation by ABDALLAH & ADINDANI (1963). A similar sandstone of about 20 m thickness was found on top of white to brown sandstones in the southern escarpment of the Gebel Tih, north of the road from St. Catharine monastery to Nuweiba on the Gulf of Aqaba. Here the basal sandstones are very immature and arcosic, they become better sorted towards the east and increase in thickness from about 35 m to over 150 m in this direction. Such sandstones have been preserved in a graben zone, some are known from regions as far as Dahab on the Gulf of Aqaba. While the lower sandstones here resemble Dakhal sandstones from Wadi Qena in many aspects (including single, large, well rounded quartz pebbles, scattered on cross-beds, convoluted beds and twisted up cross-bed ends), the violet series on top of it resembles the violet Malha sands of the Northern Galala on one side, and the Hatira or Kurnub sandstones of the Negev and Jordan on the other side (BANDEL & HADDADIN 1979; BANDEL 1981). WEISSBROD (1979) suggested that the Cretaceous Hatira Formation at the road to St. Catharine is underlain by the Amir Formation (also Lower Cretaceous sandstones) and Netafim Formation (Upper Cambrian sandstones). We were not able to recognize these formations.

### 5.2 Malha Formation

The Malha Formation was measured in four sections, one at the Khashm el Galala (Fig. 5, section 8/2) and two intermediate sections in the east escarpment of the Northern Galala (Fig. 5, sections 10/2 and 11/2); the fourth section was taken near Bir Madsus at the southern slope of the Northern Galala (Fig. 1), where the Malha sandstones have a thickness of 30 to 60 m. They consist of rather coarse sand which, in the fraction of fine gravel (with pebbles up to 0.7 mm) is commonly very immature; grains are angular to slightly rounded. Conglomeratic layers become more dominant towards the south and marine

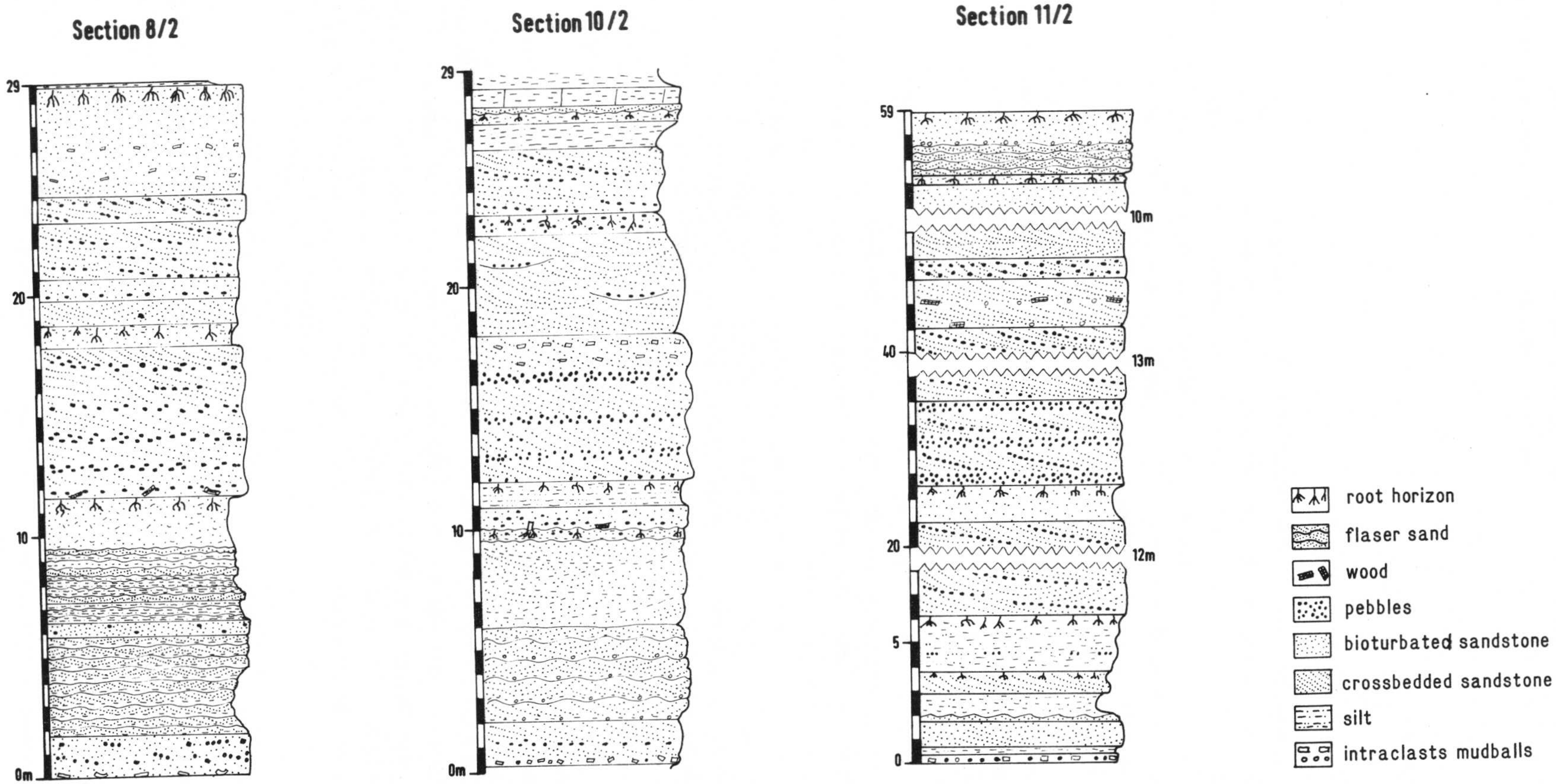


Fig. 5: Three sections from Lower Cretaceous Malha Formation exposed at the Northern Galala (Fig. 1).

influence more evident to the north. In the southernmost occurrence of the Malha sandstones eroded fragments of ferruginous beds with small molluscs from the upper member of the Jurassic sequence occur, which indicate that sediment transport was from north to south (or from E-NE), when the area of the Southern Galala was a high. The sea may have come from W or NW, as was the case during deposition of Kurnub (= Hatira) sandstone in Palestine (BANDEL & HADDADIN 1979).

The sandstones of the Malha Formation, in contrast to the upper massive Dakhal sandstone in the Wadi Dakhal and Wadi Qena area, are connected to the Upper Cenomanian sediments in their depositional history. They could well be just a little older than the sediments of the marine Cenomanian, and thus younger than the Dakhal sandstones. Their relation and source are connected to the NE where they grade into similar deposits found in the Negev and in Jordan, while the Wadi Qena sandstones have their source in the south and may be correlated with Lower Cretaceous marine deposits as they are known from the Western Desert (BÖTTCHER, 1985).

## 6. UPPER CENOMANIAN - LOWER TURONIAN SEDIMENTS

It is not difficult to connect Cenomanian/Turonian deposits of the Wadi Qena area (BANDEL et al. 1987) with those exposed at the Galala heights, since all these occurrences are quite fossiliferous and ammonites are fairly common. Two formations, a lower Atrash and an upper Tarfa Formation were distinguished along the whole Wadi Qena, and both can be traced into the Wadi Dakhal and the two Galalas. The Tarfa Formation was correlated with the terrestrial El Kanaeiss and Wadi Batar Formations in the south of Wadi Qena (ABD EL RAZIK 1972). A correlation to the Galala Formation errected by ABDALLAH, ADINDANI & FAHMI (1963) at Khashm el Galala, or to the Raha- and Wata Formation of GORAB (1961) found just opposite to the Galala heights on the Sinai, was not possible. Observations at the St. Paul monastery show that the section here is much reduced in thickness, while farther north, limestones and dolomites predominate.

The sands coming from the Red Sea Hills during Atrash times and from the Nubian continent in the south during Tarfa times (BANDEL et al. 1987) in their majority did not cross the Red Sea Hill devide and a small amount of sediments reached the area of Wadi Dakhal (Fig. 1). The region near St. Paul monastery was covered by the sea only during Late Cenomanian time. During Early (Upper ?) Turonian channel sands which are characteristic just to the west and the east of the Red Sea Hill devide did not reach the area south of Southern Galala and east of the Red Sea Hills, which at least to Wadi Dakhal was covered by the sea. This morphological high prevented sands from the Nubian continent to be transported to the east so that Cenomanian/Turonian sections in the Galala slopes are dominated by limestones, dolomites and marls (Fig. 6). A comparison of the upper calcareous portion of the section at Bir Dakhal and that of almost the same stratigraphic position at Bir Quiseib (Fig. 6) clearly shows the difference. The section at Bir Dakhal reaches the Campanian in the upper 15 m and is topped by Maastrichtian chalk of the St. Paul Formation. At Bir Quiseib, no Campanian or Maastrichtian was encountered and the section is incomplete at the top, due to faulting and erosion, before Paleocene limestones follow. But the comparison shows that Cenomanian/Turonian silts and sands grade into marls and limestones toward the north (visible at a section from Bir Dakhal to Bir Quiseib).

A very conspicuous dolomitic bed is developed at the 85 m level in the section at Bir Dakhal (Fig. 6), which probably can be correlated with that exposed to the east of the St. Anthony monastery. There it is overlain by a sequence of limestones, shales, silts and bioturbated sandstones more than 50 m thick, which is less sandy than the about 40 m of Cretaceous sediments at Bir Dakhal. The conspicuous dolomite bed was probably formed during the same terrestrial period like the dolomitic beds which end the Bir Quiseib section (Fig. 6). Within these dolomites a conspicuous dolomitic limestone sequence occurs containing abundant radial-fibrous ooids (Pl. 4, Fig. 8). The same horizon was found further north and it is assumed that its sandy equivalents are present also further south (KUSS 1987b).

The Cenomanian/Turonian (possibly also some Coniacian) sequence at the monastery of St. Anthony starts with mixed siliciclastic limestones and dolomites. Within these 150 m thick column, several

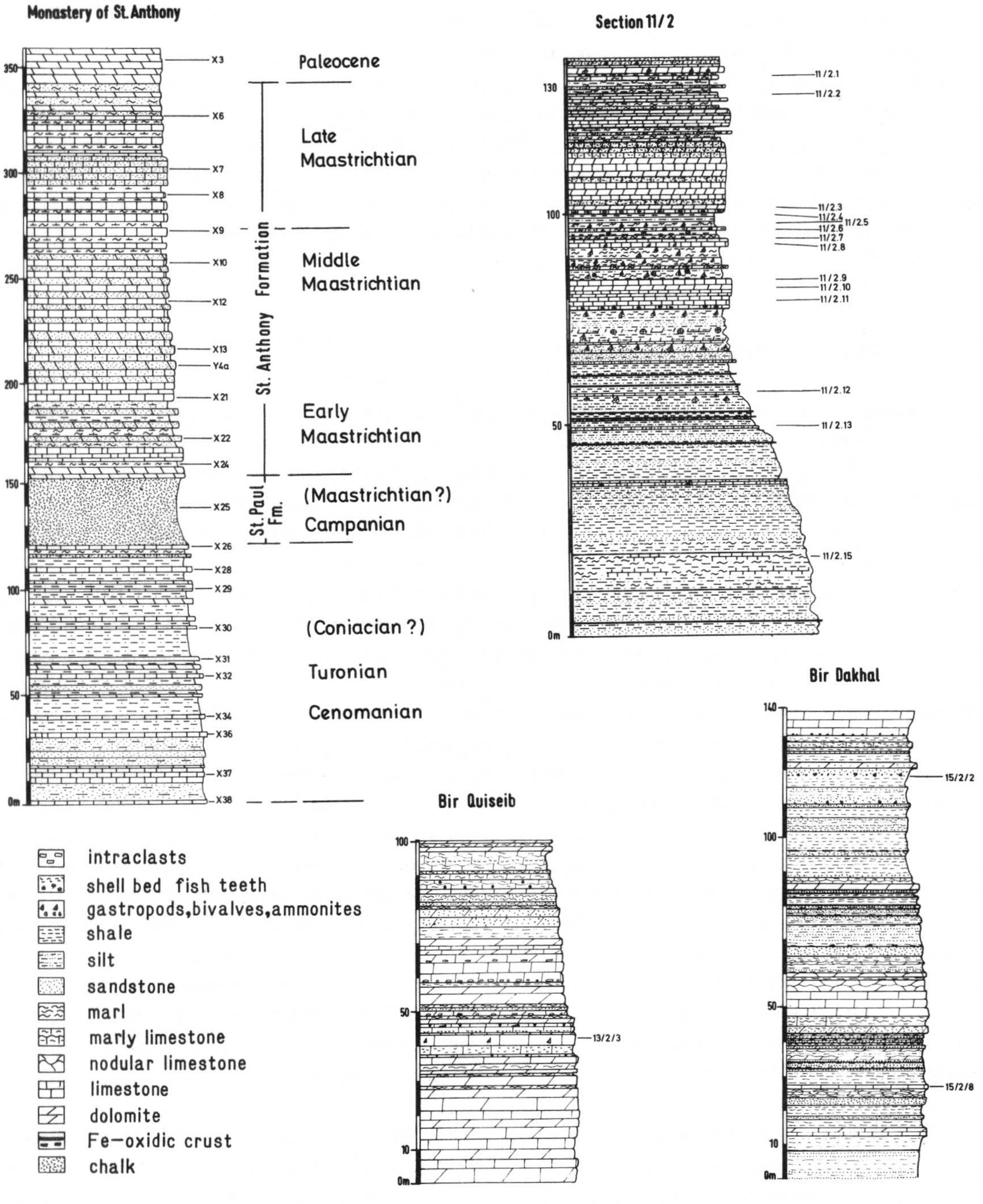


Fig. 6: Four sections of the Upper Cretaceous sediments exposed at Northern and Southern Galalas (Fig. 1). The new described St. Anthony Formation is indicated in the section of monastery of St. Anthony together with St. Paul Formation, which was described by BANDEL et al. (1987).

smaller cycles can be subdivided, each representing a coarsening upward sequence of marls or marly limestones at the base, to silts and sandstones at the top; pure limestones are rare. Frequent nodular sandy limestones with *Ophiomorpha*-burrowings occur (KUSS 1986a). In the Wadi Qena area (about 50-70 km to the south), the number of sandstones increases, oyster beds are more prominent and few horizons with ammonites occur. In the northern Wadi Qena a grainstone-facies with cortoids and *Trinocladus* sp. is developed within a locally 3-8 m thick coral-bearing limestone unit (Pl. 4, Fig. 2); this alga, together with *Boueina* sp. and *Dissocladella undulata* forms a characteristic Upper Cretaceous microflorule (KUSS 1986b).

A second shallow water limestone-type at the St. Anthony section is made up of miliolids (*Cyclogyra* sp.), codiacean fragments, dasyclads, lumps and strongly micritized cortoids. A lagoonal environmment is assumed for this low diversity fauna and flora; the original interparticle micrite is almost completely replaced by coarse, blocky cements, formed during meteoric diagenetic processes (Pl. 4, Fig. 7). This facies type is also present in sections of the Northern Galala, where it is intercalated with usually dolomitic, shallow water oolitic limestones.

While at St. Paul monastery only the lower 30-50 m of the Cenomanian/Turonian sequence are observed between Paleozoic sandstones and Maastrichtian chalks of the St. Paul Formation, only 5 km to the south the section is similar to that of Bir Dakhal with 150-160 m of mainly Cenomanian/Turonian sediments underlying the Maastrichtian chalks and overlying the Paleozoic/Lower Cretaceous sandstones.

## 7. CAMPANIAN (MAASTRICHTIAN) CHALKS OF THE ST. PAUL FORMATION

The sea withdrew from the whole studied area after deposition of Upper Turonian (Coniacian ?) sediments, to return at Campanian time. The terrestrial erosional phase which must have been present for an extended period did not leave more conspicuous signs in the sedimentary column than any of the shorter terrestrial phases, which occurred during deposition of the Turonian sediments. While sediments cannot provide good evidence for this large gap in the depositional history, the fauna can give this information. Ammonites (determined by J. WIEDMANN) were good indicators in the south, at Wadi Qena (BANDEL et al. 1987) and meanwhile, also an oyster chronology was established (MALCHUS in prep.). It was possible to recognize a hiatus between the Turonian and the Campanian, 5 km to the south of St. Paul monastery and at Bir Dakhal, and to correlate these sections with sections in the Wadi Qena and on the Sinai. In all sections from the northern Wadi Qena depression in the south to Wadi Dakhal, and the southern and northern slopes of Southern Galala, the characteristic phosphate formation (Abu Had Formation: BANDEL et al. 1987) was not found, and possible equivalents are only several ten meters thick. Within these equivalents, some shales and chert beds (with radiolarians) as well as layers rich in phosphatic pebbles and fish teeth provide a basis for comparison with the thick sequences seen at Gebel Abu Had at the southern end of Wadi Qena (BANDEL et al. 1987). In the two Galalas, like on the southwestern Sinai, a similar rock sequence was described for sediments, which were deposited during Coniacian time (LEWY 1975; BARTOV & STEINITZ 1977).

A dramatic change in the style of deposition, the type of sediment and the ease in which sedimentary history is unraveled in the area, came about during Campanian times. Synsedimentary movements were observed in the Abu Had sections (BANDEL et al. 1987) which occurred at the time of deposition of the Duwi member of the Abu Had Formation in the area of Wadi Qena. Similarly, the enrichment of phoshate in phosphatic sand and pebble offshore bars is connected to structural unrest in Egypt as well as in the Negev area and in Jordan (BANDEL & MIKBEL 1986).

Structural movements occured prior to the deposition of Camp./Maastr. St. Paul Formation, which displaced Cenomanian/Turonian sediments for at least 100 m between St. Paul and Bir Abu Kheleifi, only 5 km to the south of the monastery (Fig. 1). Thus the St. Paul Formation at St. Paul monastery rests on a sequence of 30-50 m thick Cenomanian deposits and near Bir Abu Kheleifi on 140-160 m thick Cenomanian/Turonian sediments. The St. Paul Formation can be correlated with Campanian/Maastrichtian chalks overlying the Abu Had Formation (phosphatic formation) in the area of Khour el Malik, to the west of Mut and at the southern Farafra depression (both in the Western Desert), and to chalks of the Mount Scopus Group in the Sinai (BARTOV & STEINITZ 1977). BARTHEL & HERMANN-DEGEN (1981) included these chalks in the Western Desert along with the phosphate-rich beds into the Dakhla Formation (Qur-el-Malik member). In contrast to the chalks of the St. Paul Formation, Western Desert equivalents are not well bedded and do not reach a greater thickness than about 50 m.

Chalks of the St. Paul Formation are well-bedded, with beds becoming less thick towards the top of the Formation. Its equivalents in the northern Wadi Qena measure a few meters in the south and 30 m in the north (BANDEL et al. 1987). At the southeastern edge of South Galala they become thicker than 200 m. St. Paul Formation could not be proved at the Northern Galala.

## 8. ANTHONY FORMATION

At the monastery of St. Anthony, the chalks of the St. Paul Formation are overlain by a 190 m thick unit of chalky limestones, marls and sandstones which underlie the Paleocene strata. This unit was found only near St. Anthony monastery (and the escarpment just to the east of it), but not in the Wadi Qena, St. Paul monastery or the Northern Galala. It was dated by planctonic and larger benthonic formaminifera (KUSS 1986a) as Maastrichtian. Ammonites (det. WIEDMANN) confirm these dates. The first thick *Orbitoides*-bearing limestone overlies marly wackestones, which contain planctonic organisms. These *Orbitoides*-packstones contain a rich microfauna and -flora (Pl. 4, Fig. 1). Besides, large well-rounded extraclasts of the underlying wackestones, intraclasts, ruditic bioclasts, algae, orbitoids and other foraminifera occur. *Omphalocyclus* sp. occurs together with *Orbitoides* sp., *Siderolithes vidali* and *Siderolithes calcitrapoides*, representing a typical euhaline Maastrichtian shallow water assemblage. The sandstones above with marly silty layers and dolomitized limestones yield one less altered horizon of densely packed sandy wackestones. Abundant planctonic organisms and debris occur within a silty clayey matrix. Shallow water indicators are totally absent. A Late early to Middle Maastrichtian age is indicated by globotruncanids of the *stuartiformis*- and lower *gansseri*-zone (KUSS, 1986a). A facies change from sandy limestones to

Fig. A and B show the central portion of St. Anthony Formation, exposed a few km to the east of St. Anthony monastery.

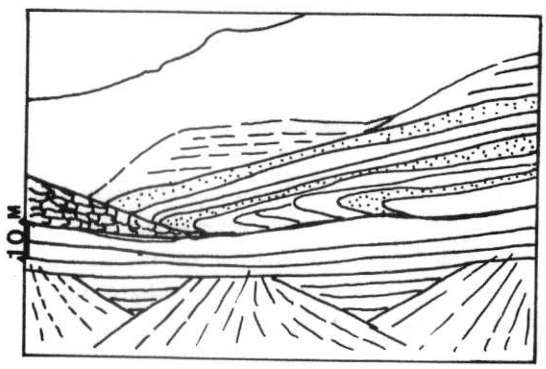


Fig. A: Sketch of a mass that has slid onto a bed with slump boulders. The front of the mass is folded; the allochthonous mass measures up to 20 m in thickness.

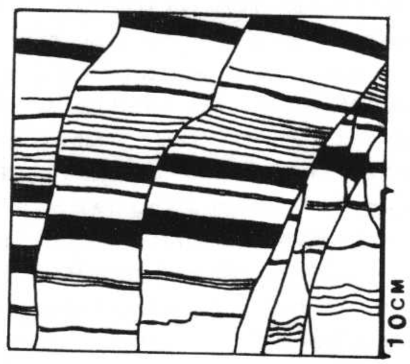


Fig. B: A rock surface showing step-like fracturing and sliding in small scale (20 cm high).

sandstones intercalated with sandy limestones which contain *Exogyra overwegi* is noted. While these sandy sediments may have been deposited in a near-shore environment, the thick sequence below represents deposits of an open, shallow sea. A similar Maastrichtian sequence of marine sediments is only exposed a few km to the west of St. Anthony monastery at Wadi Ashkar and to the east near Gebel Thilmet, but neiter on the southern slope of the Southern Galala nor in the Northern Galala. Equivalent sediments with *Exogyra overwegi* were found only in the escarpments bordering Dakhla, Kharga and Farafra oases to the east and north (Western Desert). The facies of the Maastrichtian rocks from the north escarpment of the Southern Galala clearly indicates that during most time these sediments were deposited in an environment distal to the shoreline. Towards the north, a slope must be reconstructed leading into a basin, because several well bedded portions of the St. Anthony Formation moved towards the north and formed slump masses and intraformational folds (Fig. A). Individual limestone beds in their fine composition also show a movement towards a general northerly direction, indicated by small-sized fragmentation and inclination of the beds (Fig. B). Since St. Anthony Formation as well as St. Paul Formation below are not preserved in the Northern Galala, subsequent erosion before deposition of the Paleocene beds is assumed. The same is probably also true in the south, where remnants of the St. Paul Formation are overlain by Tertiary chalks.

## 9. PALEOCENE/EOCENE ROCK SEQUENCE

### 9.1 Paleogene stratigraphic setting

The Cretaceous sediments are overlain by a limestone-dominated sequence of Paleocene to Eocene age. During this time, 300-440 m of sediments were deposited in the Gulf of Suez area at the two Galala plateaus. The Tertiary sea extended to the Abyad Plateau in Northern Sudan, where 25-40 m of limestones are exposed (BARAZI & KUSS 1987). They represent the southernmost outskirts of Paleocene to Eocene transgressions which reached farther onto the Nubian continent than any Cretaceous marine ingression.

At the southern and middle Wadi Qena, the Paleocene and Eocene Serai Formation consisting of chalks, chalky limestones, flint nodule layers and beds, overlies 30 m of Upper Cretaceous St. Paul chalk (BANDEL et al. 1987). Farther to the north at Wadi Dakhal (Fig. 1), the chalk of the St. Paul Formation is about 150 m thick and overlain by bedded limestones showing large synsedimentary slump folds, which had slid in southerly directions. In the measured sections near St. Paul and St. Anthony monasteries, conspicuous slumps occur, some of which with well visible folds dipping to the south. Farther to the south (in the direction of Wadi Qena), these slumped and bedded limestones grade into flint-bearing chalks and chalky limestones/marls of the Serai Formation; the latter has the same lithological characteristics as the Thebes Formation. ISMAIL & ABDALLAH (1966) summarized the section at St. Paul monastery in a very generalized way and pointed out that the conglomeratic limestones were formed after Landenian, respectively the *Globorotalia velascoensis* zone which belongs to the lowermost part of Early Eocene (CAVELIER & POMEROL 1986).

At Bir el Dakhal, this series is underlain by several marls and sandstone beds. Marls in the same position (between the St. Paul Formation and Lower Tertiary deposits) contain foraminifera of Middle Paleocene age (det. LUGER) at northern Wadi Qena and at St. Paul monastery. In the area of the two Galala plateaus, the oldest Tertiary beds overlying the Cretaceous are of Middle Paleocene age. To the north of Wadi Dakhal in the Southern Galala, thick-bedded, often conglomeratic limestones overlie the basal Tertiary units of Middle Paleocene to Early Eocene age.

Deposition of Tertiary sediments on the Northern Galala at the same time occurred within open marine and restricted lagoonal environments. Here "loferite"-like bedded limestones and dolomites were formed on a carbonate platform gently dipping to the south. This is indicated by huge slumpings and synsedimentary movements in the area of Wadi Araba and to the south. A transition zone to a gently south-dipping slope is assumed in this region. Slumping was still observed at the southern escarpment of the North Galala near Bir Madsus, having affected some Paleocene beds.

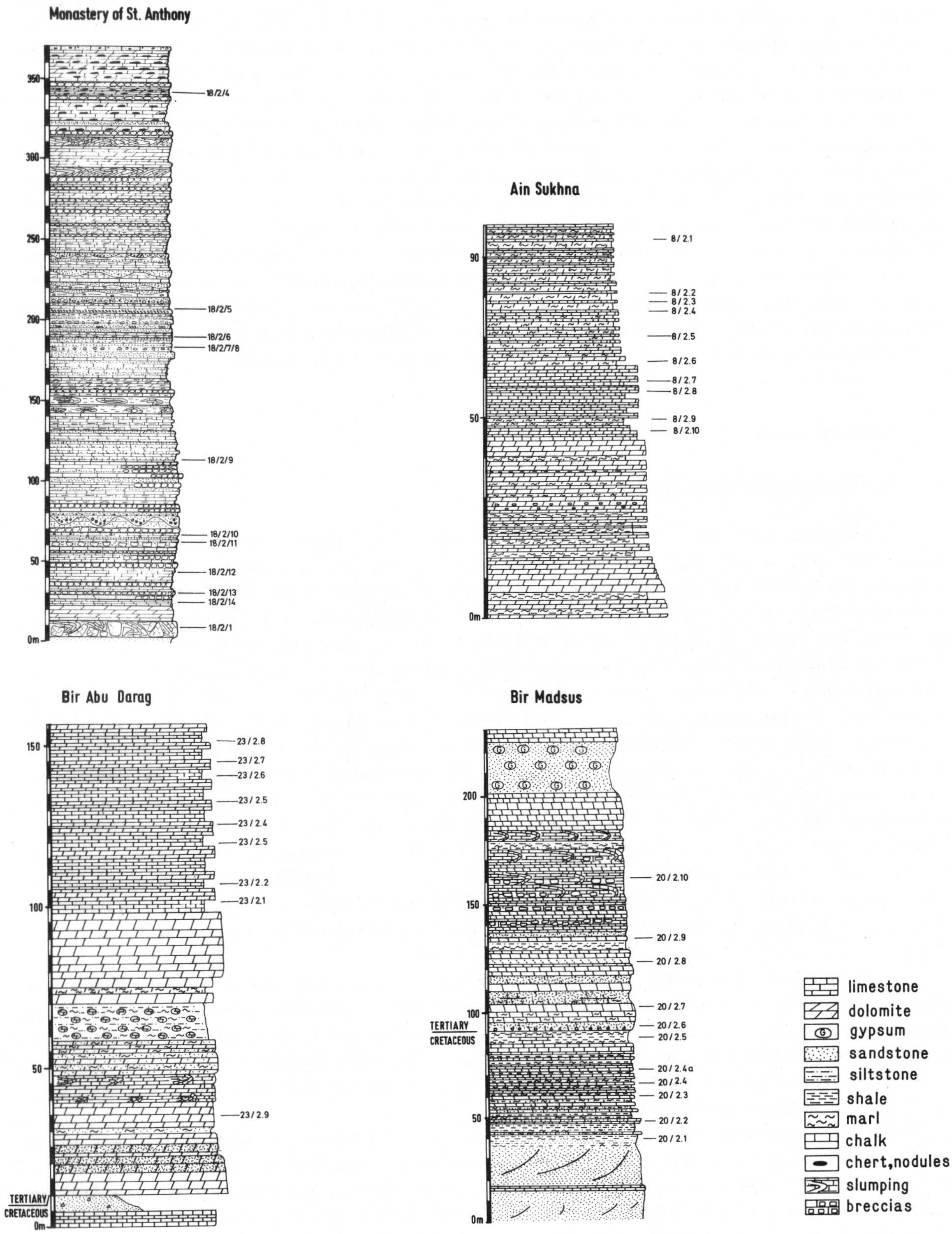


Fig. 7: Four sections of Lower Tertiary strata taken at Northern and Southern Galalas (Fig. 1).

At the monastery of St. Anthony, Maastrichtian beds with *Exogyra overwegi* are overlain by 10-15 m thick sandy limestones and marls of Paleocene age, which have a very similar lithology as the Campanian-Maastrichtian beds below. The top of these Paleocene deposits has been chipped and deformed by movements of the massive slump beds above, while they themselves do not contain any characteristics of slope deposition. Thus they represent a time marker which dates the onset of unrest and the beginning of tilting of the Southern Galala platform with a dip towards the south during Paleocene time. This direction of dip contrasts with the northerly dip direction which existed during Campanian time. At the end of Paleocene time, a basin formed in the south where chalks and flints of the Serai Formation were deposited in Wadi Qena, the upper Nile (e.g. Valley of the Kings at Thebes) and the southwestern Sinai.

The section near St. Anthony monastery shows a strong influence of quartz sand in the central portion of the Tertiary sequence and some sandy beds in the lower and upper portions. In the Lower Eocene sediments at the southern escarpment of the Northern Galala, 40 m of sandstones are intercalated between basal "loferites" (Late Paleocene to Early Eocene) and nummulitic limestones above (Middle Eocene). This sandstone unit in its characteristic lithology is found in the whole upper escarpment of the Northern Galala and it reappears at the surface at Gebel Ataqa. In the Northern Galala, a greenish-glauconitic, marine lower portion is found which is not developed at Gebel Ataqa. Here the whole sandstone is red and contains layers with nodular gypsum which definitely formed during deposition. At Bir Madsus, a conglomerate near the base of the sandstone contains quartz pebbles, eroded oysters, ferruginous clasts and limestone pebbles. At Gebel Ataqa a conglomeratic bed at the top of the unit contains limestone pebbles and fragments of reworked chert. These sands give evidence for withdrawal of the sea from the Northern Galala platform at Middle Eocene time and the deposition of sand in lagoonal, salinal environments; they were connected to the open shore at the south of Wadi Araba. At this time, limestones, sandstones and flint-bearing chalks were eroded in the north, perhaps also in the east. EL-AKKAD & ABDALLAH (1971) described these deposits at Gebel Ataqa and presented evidence for their Middle Eocene age.

## 9.2 Microfacies and paleontology of the Tertiary deposits

Microfacies analysis was carried out on limestones of all figured sequences in the Southern Galala and in the "loferites" of the Northern Galala, supplemented by some data from Wadi Dakhal (Fig. 7). The nummulitic limestones of the Middle Eocene are dolomitized to such a degree that no data could be provided for microfacies analysis.

### 9.2.1 The Northern Galala

Deposits of Tertiary age overlie the erosional surface of Turonian to Coniacian sediments on the Northern Galala. The lowermost Tertiary deposits consist of sandstones, usually with a limestone matrix (often dolomitized), and sand-dominated limestones with few fossils of Paleocene age when viewed in thin-sections. Within this unit, coarse, graded sandstones form channels, cutting into limestone-dominated layers. These were found in the lower parts of several sections from both Galala plateaus. Their thickness varies between 3 and 15 m and they were classified as "micritic sandstones" (MOUNT 1985). The stratigraphic subdivision of the following units is based on alveolinids which first occurred in Late Paleocene time with *A. primaeva* (Fig. 10).

The lower sandy units are overlain by limestones of open marine as well as restricted lagoonal environments of Late Paleocene to Early Eocene age. They are made of limestone-dolomite-marl sequences with cyclic alternations. A compiled complete sequence of a cycle documents onlap-offlap conditions. It starts off on an erosive surface with sediments formed in the subtidal stage (trangressive) stage. Next are deposits of the intertidal stage (regressive) overlain by beds formed in the supratidal area, which are transformed by the following terrestrial phase or may be covered even by lake deposits (Fig. 8). 15 to 20 cycles were counted at different sections of the Northern Galala plateau; the thickness of each cycle varies between 2.8 and 4.5 m; the complete cyclic sequence measures between 30 and 80 m. The cyclicity of the carbonate beds can be viewed best from the distance on weathered surfaces: Micritic dolomites have a white crust on weathered surfaces; limestones and marls appear dark.

The thick basal limestones are often dolomitized; intraformational reworking, bioturbation and only rare algae and gastropod remains occur. They appear to be of low diversity. These subtidal wacke-/mud-stones grade into peloidal calcarenites of the intertidal regime. They usually exhibit various types of open-space structures with a conspicuous laminoid fabric. The supratidal sediments consist of reworked dolomitic limestones with intercalated stromatolitic layers. Soils and fragments of hardgrounds may be abundant in vein and cavity fillings.

Shallowing-upward cyclothems of the characterized type are repeated several times. They form a succession of shallow-water deposits of the type first described by FISCHER (1964) from Upper Triassic limestones. His interpretation that such cycles indicate small-scale transgressive/regressive events can also be applied to the limestones described above. A short description follows with microfacies criteria of the main facies types:

9.2.1.1 The intertidal/supratidal stage

a. L-F-dolomites

The micritic, laminated loferites (L-F-dolomites) are characterized by irregularly bordered voids which are filled with calcite. Peloidal dolomitic limestones show an abundance of fenestral fabrics (birdseyes, stromatactis; Pl. 5, Figs. 1, 4), some are subsequently filled by geopetal mud as well as by vadose silts. Two different types of lamination which are intercalated with eath other (each layer is 0.5-1.5 cm thick) can be distinguished:

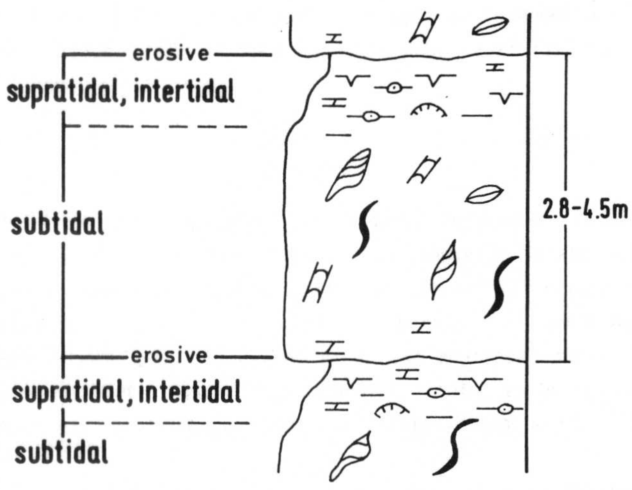


Fig. 8: Compiled sequence of a transgressive - regressive cycle within the Early Tertiary limestones of Northern Galala. Subtidal stage with molluscs, algae and bioturbations, the intertidal and supratidal stages with loferites and open space structures.

The first consists of loosely packed, graded pelsparitic layers with many intraclasts and abundant bedding-parallel stromatactis cavities (Pl. 5, Fig. 4, upper part). Their horizontal, bedding-parallel voids are filled with internal crystal silts. Burrowing tubes similar to those produced by crabs and other crustaceans irregularly cross laminations. They are larger than the fenestrae, have sharp boundaries and have been gravitatively filled with sand/gravel-sized clasts of reworked dolomites.

The second type of lamination consists of thin micritic and very fine layers with a dark and fine-grained dense matrix between the peloids; birdseyes predominate here together with rare small-sized stromatactis (Pl. 5, Fig. 4, lower part).

This second type is interpreted to have formed thin organic algal films, fixing the fine-grained particles together. The peloidal components within the first type were deposited in agitated water, without binding and fixing of algae. The rocks of the L-F-dolomites are conspicuously porous, because voids of tubes and fenestrae have not totally been filled by calcitic cements. An intertidal environment is very likely for the formation of both L-F-dolomites similar to those studied by FLÜGEL (1982).

b. Reworked dolomites with stromatolites

Reworked, irregular, polygonal dolomitic clasts, flat pebbles formed from laminated dolomites, and irregular lumps are embedded in a fine crystalline dolomitic matrix. Such beds are intercalated with stromatolitic algal mats (Pl. 5, Fig. 2), which are also strongly dolomitized. Irregular hardground surfaces with karstic morphology appear in these beds and indicate that periods of exposure and erosion alternated with periods of deposition. The reworked dolomite beds were probably storm-generated. The whole rock type described here was formed within the supratidal area. A model for the formation of such type of sediment with dolomitic surface clasts, storm layers and fine crystalline dolomites could be found in several descriptions of similar lithologies, e.g. JAMES (1984).

c. Palustrine limestones

Palustrine limestones were formed when limy deposits had been transformed by terrestrial, pedologic processes (FREYTET & PLACIAT 1982). The following microscopic aspects are significant:

The limestones commonly contain micritic, dark nodules with irregular sparitic voids, which have probably been excavated by small roots. Clear tubes of roots are present outside the nodules forming larger voids in the matrix. Brown films cover the inner walls of both voids. Percolating surface waters may have generated big, horizontally elongated voids (Pl. 5, Fig. 6), which were filled with fine-grained internal sediment, totally or partly, and remaining cavities were later filled with sparitic calcite. Mottled rock surfaces are due to irregular zones of impregnations by ferruginous oxides. They create globular halos, which are best visible at low magnification. The limestones also contain ferruginous quartz nodules which are clearly distinct from the surrounding rock. They have sharp boundaries and probably represent reworked clasts of terrestrial origin. HERBIG (1987) described similar rock fabrics (disorthic features) from Lower Tertiary palustrine limestones of Morocco.

Palustrine limestones, formed during long-lasting regressive phases, are rarely preserved as concrete layer and are usually disintegrated into fragments. Therefore, complete cycles with these limestones preserved are rare. Usually, fragments of palustrine limestone beds are integrated into the base of the next cycle.

The terrestrial conditions during the most regressive phase are also expressed by planar fissures and by desiccation breccias. The resulting cracks and cavities were filled by coarse fragments mainly derived from the surrounding substrate and material coming from above, like soil. In some cycles, terrestrial influence directly altered the subtidal carbonates, and palustrine limestones are not developed.

9.2.1.2 Subtidal stage

The subtidal cycles are characterized by thick-bedded limestones or dolomites. These can be differentiated into wackestones and biosparites.

a. Wackestones

The wackestones contain a fauna and flora of low diversity but with large numbers of individuals present in some layers only. They represent restricted, possibly lagoonal environments. The most frequent dasycladacean alga is *Belzungia silvestrii*. It occurs together with the foraminifera *Orbitolites complanatus*. A second yet undescribed dasycladacean (Pl. 6, Fig. 1) is only found in a few layers, but is common here together with *Alveolina primaeva*. Also fragments of bivalves and other organisms occur.

Irregular dark crusts of variable thickness often crisscross limestones of this facies-type when seen in thin-section (Pl. 5, Fig. 8). These crusts are not connected with changes of the fabric. Crusts formed when the sediment was not indurated, so they could be washed out and redeposited as clasts. Such clasts larger than 1 mm are common in some layers. BARTHEL (1974) described similar ones as black pebbles from recent carbonates as well as from the Mesozoic. He interpreted them to have been originated by infiltration of organic substances into porous carbonates and subsequent reworking and redeposition. Probably iron-manganese salts were precipitated within the semi-indurated carbonate mud, forming later the crusts. Since small traces of roots are also found connected with crusts, the formation of these crusts took place at the shore, where fresh- and salt-water mix. The limestones have not necessarily been very long under terrestrial influence; other diagenetic features indicate normal marine conditions.

b. Biosparites

A second limestone-type of the shallow subtidal portion of a cycle consists of well-sorted, micritized bioclasts and small micritized miliolids along with intraclasts, a few gastropods, dasycladaceans, udoteaceans and a few alveolinids. Among the greater agglutinated foraminifera, *Broeckinella arabica* (Pl. 6, Fig. 5) represents a typical Paleocene shallow water form of the southern Tethys (DROBNE & HOTTINGER 1971). Also *Fallotella kochanskae*, *Coskinolina* sp. and *Alveolina primaeva* (Pl. 6, Figs. 6, 15) are present. Distinct beds contain *Miscellanea meandrina* (det. U. LEPPIG) occurring together with *A. primaeva* in Turkey (SIREL & OZCAN 1975). Some biosparites have two types of sparitic cement. Nearly all components are covered by thin crusts of microsparites, often mixed together with relicts of brownish crusts which may represent solution residues. The remaining cavities are only partly filled with blocky calcites, so that a conspicuous intercrystalline porosity remains. A mixed marine/phreatic diagenetic environment could be responsible for the diagenesis of these carbonates. This type of rock occurs also in beds of the sections of the Southern Galala, but here as clasts and pebbles of reworked layers, redeposited in the nodular limestone facies.

9.2.1.3 Shoals and offshore bank deposits

Well sorted, coarse grained limestones are intercalated with the cyclic carbonates at Northern Galala and the nodular limestones at both Galalas. They are more prominent in southern sections than in northern ones, vary in thickness from 0.25 to 1.7 m and represent high energy shoal beds of lenticular shape, which laterally interfinger with low energy lagoonal deposits. Crossbedding structures are well visible on weathered surfaces, and thin-sections reveal good sorting, imbrication of the components and the absence of carbonate mud. Both subfacies represent shallow shoal units, which can be subdivided in two facies types: The first are rudstones/boundstones with coarse, gravel-sized components; they represent the central shoal. The second are skeletal grainstones/rudstones; such shoals are common in middle shelf settings (WILSON & JORDAN 1983).

a. Shoal core facies (rudstones/boundstones)

Where both shoal facies are found together in one shoal deposit, the central type appears more massive, while the second type is banked. Thus, such lenticular bodies resemble small reefs flanked by reef debris when seen in the field. SNAVELY et al. (1979) also found shallow water platform deposits without reefs and their organisms in Tertiary limestones of the Duwi area (near Safaga, Red Sea).

Most prominent components of the shoal core facies are thick crustose lumps. They are composed of a nucleus consisting of small bioclasts, foraminifera, ball-like red algae, corals and unidentified debris. These nuclei are surrounded by cm-thick crusts consisting of two types of red algae: *Lithothamnium* sp. forms consecutive crusts together with *Solenomeris o'gormanni*, which may hold few encrusting foraminifera beween them (Pl. 5, Fig. 7). Besides, few 3-7 cm large colonies of corals and bioclasts are surrounded by thin crusts of *Lithothamnium* sp. only. The latter and crustose lumps may have become fractured and are surrounded by algal encrustations again. *Alveolina primaeva* is present within the older of the shoal cores, which therefore are of late Paleocene age. Younger shoals contain *Alveolina pasticillata* of Early Eocene (Ypresian) age (Pl. 6, Figs. 15, 17).

b. Shoal flank facies (grainstones/rudstones)

The shoal flank facies laterally interfingers with the central shoal facies or occurs without this first subfacies type in thinner beds. A typical aspect of this facies-type is shown in Fig. 3 of Pl. 5. Brownish irregularly shaped extraclasts together with alveolinids dominate. The brownish extraclasts represent reworked clasts, which were still plastic when they were eroded from lagoonal and shallow subtidal areas. These reworked algal packstones usually show monotypic accumulations of *Ovulites morelleti*; extraclasts are formed by clasts of pelsparitic grainstones and slightly dolomitized bioclastic packstones. These extraclasts are between 200 µm and 1.2 cm in diameter, with a maximum at 500 µm. Grainsize of alveolinids (which are very important constituents of all shoal sediments) vary within the same ranges. Since they have the size and round shape of large sand grains, they can be transported and reworked like these. Alveolinids were vagil benthonic organisms and not reef forming, as was assumed by SAID (1960). Their accumulation in shoals is due to mechanical concentrations. Along with extraclasts and alveolinids, *Opertorbitulites* sp., *Alveolina primaeva*, *Ranikothalia* sp., *Nummulites* sp. and other rotaliid foraminifera occur. Because of their flat shape, they are usually oriented parallel to bedding. While well-rounded quartz grains are common, fragments of red algae are rare. The surface of many components is stained by iron-oxides, indicating that they have been deposited within the shoals after an individual history of formation.

9.2.2 The Southern Galala

Deposition of Tertiary sediments on the Southern Galala took place on top of the eroded Upper Cretaceous deposits, ranging from Late Turonian to Late Maastrichtian. As in the Northern Galala, the lowermost Tertiary deposits consist here of sandstones channelling into limy deposits. A few marls of Middle Paleocene age were found in this series. The sandy unit is overlain by thickbedded, often conglomeratic limestones with conspicuous slump structures. Though these were already proven in the southernmost escarpment of the Northern Galala (at Bir Madsus) as a sequence of 25 m thickness, they are very typical in sections at the Southern Galala. These sediments were deposited on the slope and consist mainly of sand-sized particles, forming beds of a thickness between 10 cm and several meters. Most of these beds moved downslope with a semiconsolidated to consolidated consistence. Those carbonates sliding relatively gradually, were transformed into nodular limestones; others moved less gradually and formed chaotic beds of variable thickness reaching several ten meters. Sliding masses eroded into the underlying units and distorted them locally. While moving down the slope, stratified carbonate beds disintegrated into 10 to 50 cm large fragments due to tension fractioning. Quartz sand remained unconsolidated during downslope movements. Where beds had been solid before the downslope movement, fracturing formed breccia slumps. Whole piles of bedded nodular limestones form slump units which may or may not be folded. Synsedimentary folds clearly show downslope movements with generally southerly directions.

The thickness of this sequence rises up to more than 200 m (near the monastery of St. Anthony) and to 160 m at the monastery of St. Paul. The nodular limestones interfinger at the southern edge of Southern Galala with bedded chalky limestones and marls of the Serai Formation, which also show conspicuous slumping structures, dipping to the south. Nodules of slumped shallow water limestones could be found within

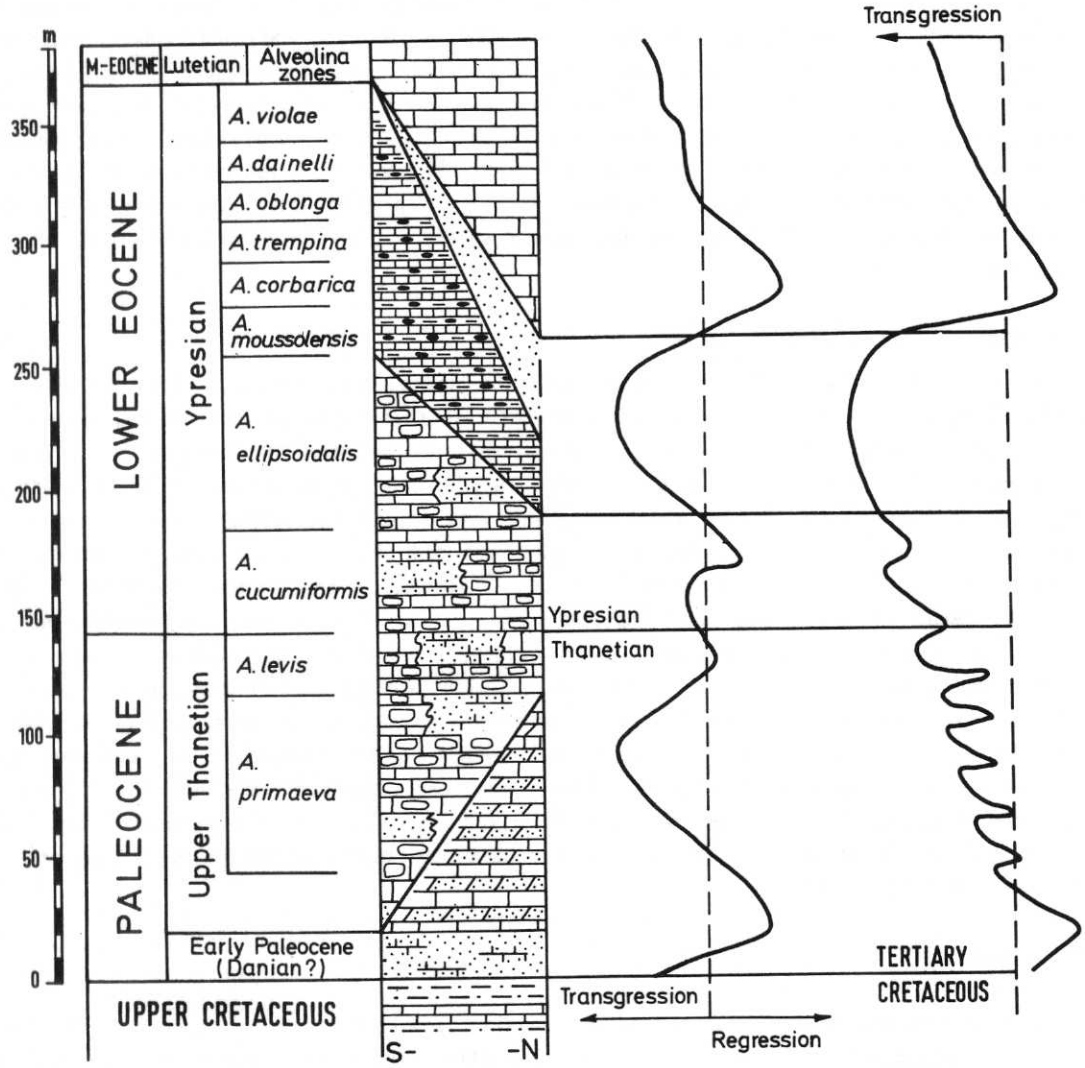


Fig. 9 : Summarized succession of Late Paleocene to Early Eocene deposits with variations of facies and thickness from the Southern Galala (S) to the Northern Galala (N). *Alveolina*-zonation is based on CAVELIER & POMEROL (1986). The global transgressive - regressive phases are firgured on the left curve (VAIL & HARDENBOHL, 1979). The right curve was drawn with the data of this work.

the latter. As the Serai Formation was also found on top of the nodular limestones, diachronous transitions between both are very likely.

The nodular limestones contain *A. primaeva* at the base and *A. oblonga* near the top, corresponding to a Late Paleocene to middle Early Eocene age. The latest Early Eocene was proved with alveolinids of the *A. violae* zone, occurring within silicified slump nodules.

Slump deposits (nodular limestones)

Three different types of deposits can be distinguished which are represented by various microfacies types: a) Nodular limestones with two different microfacies, b) sand-dominated units, and c) a facies transitional to basin deposits of the Serai Formation with two different lithological types.

a. Most limestone nodules are made of biosparites, often containing high amounts of *Alveolina*-tests; besides small miliolids, micritized bioclasts and intraclasts are frequent. Several units hold discocyclinids (Pl. 6, Fig. 8). These nodules were formed in shallow subtidal areas.

More distorted slide masses commonly show large and irregular fragments of carbonates which have been altered by a strong diagenesis within the vadose zone. Surfaces have strongly been eroded and acquired a karstic morphology before they were incorporated into the breccia-slumps. Many of these breccia clasts have a chalky porous structure, and former fabrics are destroyed, probably due to the influence of salty fluids.

Both facies types represent sediments which were redeposited from the region of intertidal and subtidal areas of the Northern Galala. The second type which is strongly affected by vadose diagenesis, was only found in the northerly outcrops of the Southern Galala, whereas the biosparites were proved also within the southern sections.

b. The sandstone units are usually unfossiliferous. A few sandy layers with a higher limestone content contain dasycladaceans and thick-walled rotaliids, among them *Cuvillierina* sp. and *Miscellanea* sp. together with alveolinids which indicate the *Alveolina cucumiformis*-zone (HOTTINGER 1980) respectively Early Ypresian.

These sand-rich layers (also with slumping structures) occur as cyclic units alternating with limestone-dominated, nodular carbonates within thick units of several sections (fig. 7). They possibly reflect deposition in different water depths, which may correlate with the shallow water cycles observed in the Northern Galala (chapter 9.2.1).

c. The transitional facies to basin deposits of the Serai Formation is characterized by intercalations of bedded and nodular limestones which are partly siliceous.

In section U, measured in the northern parts of Wadi Tarfa (BANDEL et al. 1987), a basinward facies of the Serai Formation occurs: The lower 85 m of chalky limestones with marly, flint bearing intercalations show the typical rhythmic Serai sequence, known from many other localities. A conspicuous bedded limestone unit of 15 m thickness is intercalated with a sharp, erosional base. Several layers containing alveolinids and rare dolomitized nummulitids accumulated 6 m above the base. To the top, these limestones grade into the bedded Serai Formation, in which predominantly planctonic organisms occur.

Several discontinuities are irregularly cutting the limestones, which are classified as biosparites and wacke-/packstones. Biosparites containing miliolids (orbitolitids and alveolinids which in some layers are imbricated), dasyclads, biodetritus and high amounts of terrigenous quartz. In the wacke-/packstones the same components occur, but they usually contain more quartz; bioclasts are more destroyed due to erosion (Pl. 6, Fig. 3). All these components were found neither in the underlying beds nor in the limestones above. Both types have sharp contacts to the normal pelagic limestones, mainly composed of globorotaliids. The shallow water limestones contain *Alveolina ruetimayeri* and *A. oblonga* of latest Ypresian.

The second type of flint-like nodules often occurs within the slumping structures of the Serai Formation. Thin-sections show that thin siliceous crusts surround a well preserved shallow water fauna and flora within a miliolid grainstone facies (Pl. 5, Fig. 9). Besides dasycladacean algae, different alveolinids are present, ranging from *A. ellipsoidalis* to *A. violae* zone.

Both limestones were interpreted as distal outlayers of different slump events. Their components were originally formed by animals and plants living in shallow water facies further to the north.

## 9.3 Paleoecology and paleogeography of the Paleogene sediments

Considering all the mentioned vertical and lateral facies characteristics and changes, the following model is proposed (Fig. 10). In the north cyclic shallow water carbonates with loferites represent coastal tidal flats with changing input of terrigenous sandy material. They interfinger with shoals and offshore bank deposits. The nodular limestones continue to the south and are diachronously interfingering with Serai limestones in the sections further to the south.

According to KUSS (1986b) and BARAZI & KUSS (1987), sediments were deposited on a widespread shelf-area in northeastern Egypt during Late Paleocene/Early Eocene times. Sedimentologic, microfacial and paleontologic data suggest that deposition occurred on a shallow shelf system in the north, with a gentle slope to the south. Shallow platforms were developed on uplifted fault blocks, where organic carbonate deposition flourished. Uplift movements are also indicated by evaporitic, sebkha-like deposits, intercalated in near-shore strata at Gebel Ataqa and to the southern rim of Northern Galala. To the south, a "dolomitic restricted" facies follows, grading into "epeiric clear-water sedimentation" and farther south into the "outer shelf" facies of the Serai Formation. This mosaic of different carbonate facies coincides with the general shelf model proposed by IRWIN (1965), who first mentioned the term platform ramp for a shelf-margin profile without reef structures and gradual deepening.

Deposition on the slope was controlled by the accumulation rates from the shallower environments. It is supposed that cyclicity which occurs in both facies belts (Northern and Southern Galala), was caused by tectonic uplifts in the northern areas. STROUGO & HAGGAG (1983) described a thin sequence of Paleocene deposits near Abu Roash (near Kairo) which underwent up-arching towards the end of Cretaceous; the tectonic movements reversed during Paleocene, resulting in limestone pebbles and coarse debris deposited on a slope. They postulated a slope-rise to have taken place during Upper Paleocene (*pseudomenardii*-interval) which is well comparable with the sequence described, where Tertiary sediments not older than Late Paleocene (*A. primaeva*) could be found.

Uplift, cyclic sedimentation and in-situ reworking are confirmed by multiple slumping events, disintegrating the shallow-water deposits downslope. In the studied area, they are active from Late Paleocene (*A. primaeva*) to latest Ypresian (*A. violae*), which corresponds well with observations of STROUGO (1986) from several other places in Egypt. SNAVELY et al. (1979) described periods of localized uplift and probably rapid sea-level drop from the Duwi trough (about 250 km to the south), which began during Early Eocene.

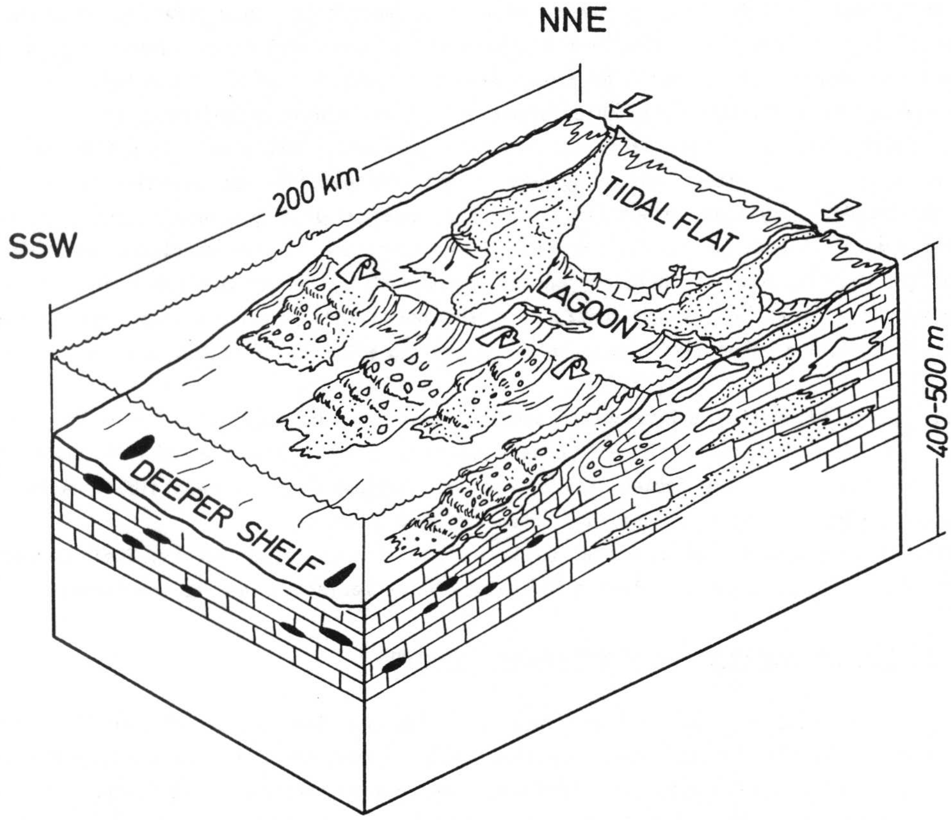


Fig. 10: Depositional model of the Late Paleocene to Early Eocene sediments between Gebel Ataqa (in the north) and Wadi Dakhal (in the south).

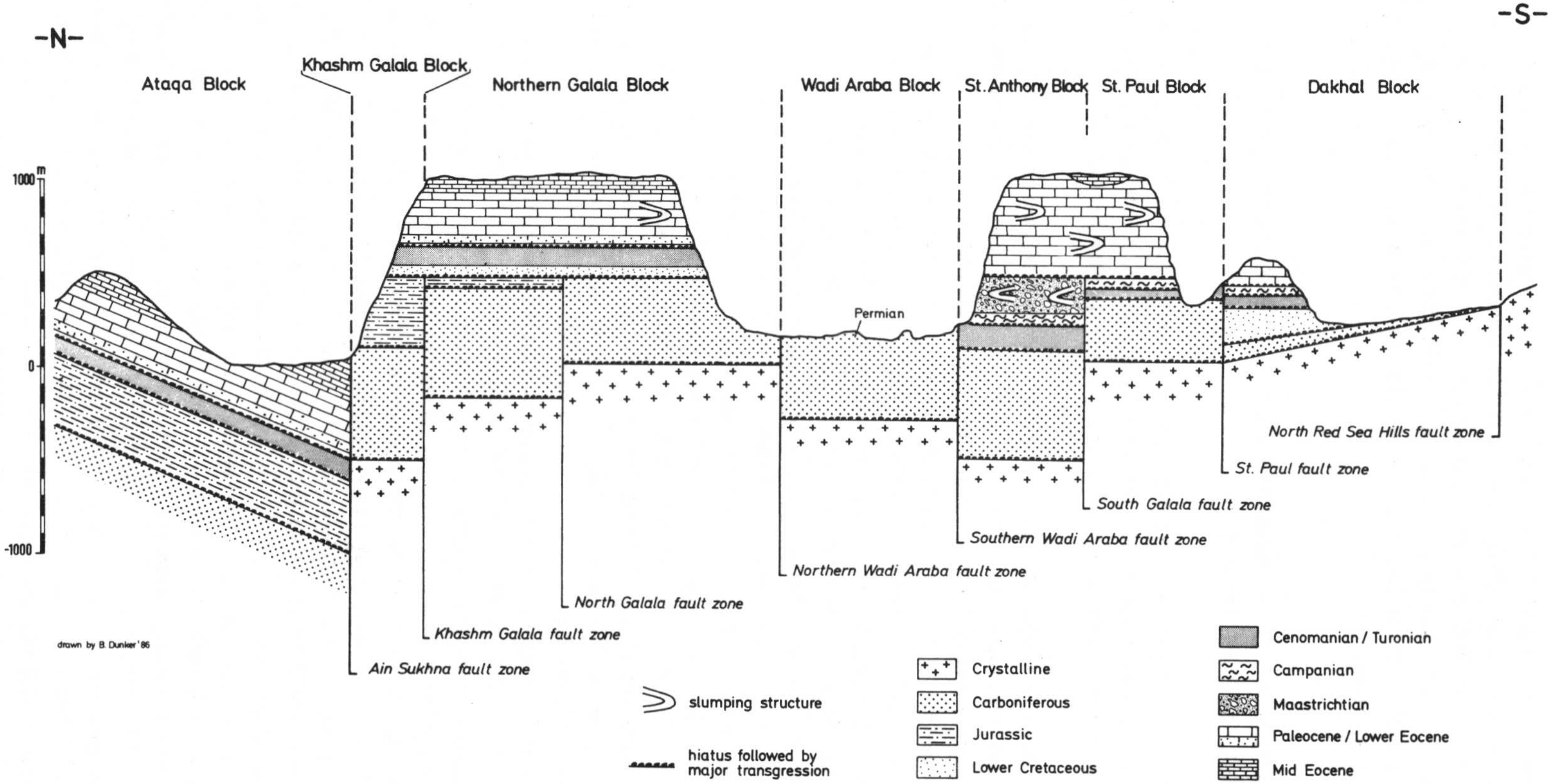


Fig. 11: Profile indicating a structural, stratigraphical and facial reconstruction of the sedimentation between Gebel Ataqa in the north and the Red Sea Hills in the south (see stippled line of Fig. 1).

Eustatic changes of the sea level at Paleocene to Eocene time are noted world-wide (VAIL & HARDENBOL 1979), and they may well be responsible for rhythmic sedimentation observed in the two Galalas (Fig. 9). The sea-level drop at the end of the Lower Eocene caused widespread regressions, which are documented here by evaporitic Middle Eocene sandstones. Though the dolomitic limestones above yielded no index fossils, they were correlated with the Middle Eocene transgression. In the sediments of the basin, the oscillations of the sea-level are not clearly imprinted on the deposits. But limestones of late Early Eocene age from the southern escarpment of the Southern Galala indicate a more shallow environment of formation, compared with the beds below.

## 10. A STRUCTURAL, STRATIGRAPHICAL AND FACIAL RECONSTRUCTION OF THE SEDIMENTATION BETWEEN GEBEL ATAQA IN THE NORTH AND THE RED SEA HILL RANGE IN THE SOUTH

Several structural units (blocks) can be differentiated from the south to the north: Dakhal Block, St. Paul Block, St. Anthony Block, Wadi Araba Block, Northern Galala Block, Khashm Galala Block and Ataqa Block, each ending with a major fault zone which was active during different times of the depositional history of each area.

The southernmost Dakhal Block situated in the Wadi Dakhal area where the strata dip towards the north, is cut off by faults in the south, separating it from the high range of the northern Red Sea Hills. It is the only block sytem discussed here in which the crystalline basement is exposed and covered by Early Carboniferous sands and silts. These, in turn, are unconformably overlain by massive Early Cretaceous sandstones (quarried quartz-sand, Dakhal Formation: BANDEL et al. 1987), the characteristic unit of the Dakhal Block. The Lower Cretaceous sandstones are overlain by Cenomanian-Turonian marine sands, silts, marls and limestones in the north and west, covered by Campanian/Maastrichtian chalks; the latter are overlain by a thick series of Tertiary bedded sandy limestones, marls and chalky limestones.

The St. Paul fault system limits both the Dakhal Block to the north and the St. Paul Block to the south. Tectonic activities can be confirmed along this zone as regards pre-Tertiary sediments. Displacements which were active during the Cretaceous, resulted in the removal of Lower Cretaceous sandstones (Dakhal Formation) from the St. Paul Block before deposition of Cenomanian beds, and also in the erosion of a thick Turonian series before deposition of Campanian chalk and marl. In contrast, the Carboniferous sequence is much thicker here than in the Dakhal Block. This fault zone was active at different times prior to the onset of Paleocene sedimentation.

The St. Paul Block is characterized by the Campanian/Mastrichtian chalk-marl unit (St. Paul Formation: BANDEL et al. 1987) which is well exposed near the monastery of St. Paul. The lithological differences between it and the Dakhal Block are of similar significance as regards the northward following St. Anthony Block, which is separated from the St. Paul Block by the South Galala fault zone. This fault system lies hidden underneath the thick Tertiary sequence which forms most of the mountains up to the Wadi Araba. The sedimentary cover on the crystalline basement of the St. Anthony Block is probably thicker than in the blocks farther to the north (except the Ataqa Block). A Carboniferous sequence is present here, which reached into the Permian strata exposed on top of the Wadi Araba Block (pers. comm. E. KLITZSCH); the latter has moved upward (in relation to the St. Anthony Block), evidenced by fault-folding of the Cenomanian-Turonian and ?Coniacian sediments. Their thickness here is comparable to that of the Dakhal and the Galala blocks; their lithofacies is characterized by an increasing number of carbonates, becoming thicker and more intensely dolomitized farther to the north. The Campanian/Maastrichtian chalks here are overlain by a thick series of Maastrichtian sandy limestones, marls and sandstones.

The Southern Galala fault zone ceased its activity (like the St. Paul fault zone) before the deposition of Paleocene sediments. It must have been active before the onset of Cenomanian deposition, so that at least 150 m of Carboniferous sandstones are preserved south of it, and more than 400 (possibly 600) m of Carboniferous-Permian sandstones and silts are probably preserved in the northern part - definitely so in the Wadi Araba Block. After Turonian/Cenomanian deposition, the fault was active during Campanian as well as after the Maastrichtian deposits had formed. From this time on, the three blocks mentioned above acted as one single block system, with a dip towards the south during deposition of Paleocene and Eocene sediments. Inclination continues into the Galala Block, leading to the conclusion that both southern and northern Wadi Araba fault zones may have been inactive at Paleocene to Eocene times. Both fault zones were active again after marine deposition stopped in the Eocene. The transect crossing the Wadi Araba Block (near St. Anthony monastery) only exhibits Paleozoic rocks with a sequence going up from Lower Carboniferous to Lower Permian strata. Whereas this block dipped towards the south in older Tertiary times, the dip was directed towards the north in Campanian and Maastrichtian times, connected with the St. Anthony Block.

The northern Wadi Araba fault zone was active before the onset of Cretaceous deposition, so that the upper portions of the Carboniferous/Permian sequence were eroded from the Northern Galala Block. Prior to the advance of the Cenomanian sea, sands coming from the north were deposited and formed a characteristic violet to pink sandstone unit (Malha Formation). It is preserved within the Northern Galala Block as well as within the Khashm Galala Block, and most probably also farther to the north with increasing evidence of marine influence in that direction. Its southernmost outcrops were observed near the northern Wadi Araba fault zone.

The Northern Galala Block consists of two different tectonic units with a minor fault zone between them (Northern Galala fault zone). It was active before the unset of deposition of the Malha sandstone. Thus, a thin cover of Jurassic beds is present on top of the Carboniferous sequence in the north, but could not be confirmed in the south. During the activity of the Northern Galala fault zone, the more prominent fault zone at the southern end of the Khashm Galala Block was active, displacing Jurassic rocks in the north for several hundred metres against Carboniferous sediments in the south.

The characteristic deposits of the Khashm Galala Block are Jurassic sandstones, laminated siltstones and few limestone/marl layers. To the north, this series is down-faulted at the Ain Sukhna fault zone. The Ataqa Block is the northernmost block with beds dipping towards the south.

## 11. ACKNOWLEDGEMENTS

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## 12. LITERATURE

ABD EL RAZIK (1972): Comparative studies on the Upper Cretaceous-Early Paleogene sediments on the Red Sea Coast, Nile Valley and Western Desert, Egypt. - 8th Arab. Petrol. Congr. Algiers, pap. 71 (B,3), 23 p., Algiers.

ABDALLAH, A.M. & ADINDANI, A. (1963): Stratigraphy of Upper Paleozoic rocks, Western side of the Gulf of Suez. - Mem. Geol. Surv., 25, 18p., Cairo.

ABDALLAH, A.M., ADINDANI, A. & FAHMY, N. (1963): Stratigraphy of the Lower Mesozoic rocks, western side of Gulf of Suez. - Geol. Survey Egypt, paper 27, 18p., 4 Figs., 2 Maps, Cairo.

AL-HASHIMI, W.S. & HEMINGWAY, J.E. (1973): Recent dolomitization and origin of rusty crusts of Northumberland. -Journ. of Sed. Petrol., 43, 82-91, Tulsa.

BANDEL, K. (1981): New stratigraphical and structural evidence for lateral dislocation in the Jordan Rift Valley connected with a description of the Jurassic rock column in Jordan. - N. Jb. Geol. Paläont. Abh., 161, 271-308, 12 Figs., Stuttgart.

BANDEL, K. & HADDADIN, A.A. (1979): The Depositional Environment of Amber-bearing rocks in Jordan.- Dirasat, 6, 39-62, Amman.

BANDEL, K. & MIKBEL, S. (1985): Origin and deposition of phosphate ores from the Upper Cretaceous at Ruseifa (Aman, Jordan). - Mitt. geol.-paläont. Inst. Univ. Hamburg, 59, 167-188, 20 Figs., Hamburg.

BANDEL, K., KUSS, J. & MALCHUS, N. (1987): The sediments of Wadi Qena. - Journ. of Afric. Earth Sc. (in press).

BARAZI, N. & KUSS, J. (1987): Southernmost outcrops of marine Lower Tertiary carbonate rocks in NE-Africa (Gebel Abyad, Sudan). - Geol. Rdsch. 2 Plts., 1 Fig. (in press) Mending.

BARTHEL, K.W. (1974): Black pebbles, fossil and recent, on and near coral islands. -Proc. Soc. Intern. Coral Reef Sympos. - 395-399, 5 Figs., Brisbane.

BARTOV, Y. & STEINITZ, G. (1977): The Judea and Mount Scopus Groups in the Negev and Sinai with Trend Surface Analysis of the Thickness Data. - Israel Journ. of Earth-Sc., Vol. 26, 119-148, 3 Tab., 15 Figs., Jerusalem.

BÖTTCHER, R. (1985): Environmental model of the shallow marine Abu Ballas Formation (Aptian, Nubian Group) in South-Western Egypt. - N. Jb. Geol. Paläont., Abh., 169/3, 261-283, Stuttgart.

BOUKHARY, M. & ABDELMALIK, W. (1983): Revision of the stratigraphy of the Eocene deposits in Egypt. - N. Jb. Geol. Pal. Mh., 1983/6, 321-337, 4 Figs., Stuttgart.

BRANDNER, R. (1978): Tektonisch kontrollierter Sedimentationsablauf im Ladin der westlichen nördlichen Kalkalpen. -Geol. Paläont. Mitteilungen Innsbruck 8, 317-354, 22 Figs., Innsbruck.

CAVELIER, C. & POMEROL, C. (1986): Stratigraphy of the Paleogene. - Bull. Soc. geol. France, 1986 (8), t.II, no.2, 255-265, 1 Tab., Paris.

CUVILLIER, J. (1937): Presence du Carbonifere marin dans l'Ouadi Abu Darag (Desert arabique).- C. R. Acad. de Sciences, Paris, tome 204/24, 1834-1835, Paris.

DRAGASTAN, O. (1985): Review of Tethyan Mesozoic Algae of Romania, In: TOOMEY, D.F. & NITECKI, M.H. (Eds.): Paleoalgology, 101-161, 28 Plts., 3 Figs., Springer, Heidelberg.

DROBNE, K. & HOTTINGER, L. (1971): Broeckinella und Saudia (Foraminiferida) aus dem nordwestlichen Teil Jugoslawiens, ihre Morphologie und ihre stratigraphische Verbreitung. - Razprave, IV, razr. SAZU 14, 215-238, 5 Figs., 4 Plts., Ljubljana.

EL-AKKAD, S. & ABDALLAH, A.M. (1971): Contribution to Geology of Gebel Ataqa area. -Ann. of Geol. Surv. of Egypt, 1, 21-42, 2 Figs., Cairo.

ELLIOTT, G.F. (1957): New calcareous algae from the Arabian Peninsula. - Micropaleontology, 1, 125-131, 1 Pl., New York.

ELMORE, D.R. DUNN, W. & CRAIG, P. (1985): Absolute dating of dedolomitization by means of paleomagnetic techniques. - Geology, 13, 558-561, 5 Figs., 1 Tab., Boulder.

FISCHER, A.G. (1964): The Lofer Cyclothems of the Alpine Triassic. - Kansas geol. Surv. Bull., 169, 107-149, 38 Figs., Lawrence/Kansas.

FREYTET, P. & PLAZIAT, J.-C. (1982): Continental Carbonate Sedimentation and Pedogenesis - Late Cretaceous and Early Tertiary of Southern France. - Contrib. to Sediment., 12, 1-213, 49 Plts., 59 Figs., Stuttgart.

FRIEDMAN, G.M. (1978): "Solar Lake": A sea-marginal pond of the Red Sea (Gulf of Aqaba or Elat) in which algal mats generate carbonate particles and laminites. In: KRUMBEIN, W. (Ed.): Environmental biogeochemistry and geomicrobiology, vol.1, The Aquatic Environment, 227-235, 4 Figs., Ann. Arbor: Ann Arbor Science.

FÜRSICH, F.T. (1971): Hartgründe und Kondensation im Dogger von Calvados. - N.Jb. Geol. Paläont. Abh., 138, 313-342, Stuttgart.

GOLDBERG, M. & FRIEDMAN, G. (1970): Paleoenvironments and Paleogeographic Evolution of the Jurassic in Southern Israel. - Geol. Survey Israel Bull., 66, 39-49, 2 Figs., 2 Pls., Tel Aviv.

GROVES, J.R. (1984): Biostratigraphic significance of Hemigordius harltoni CHUSHMAN and WATERS, 1928 (foraminiferida). - Journ of Paleont., 58/1, 78-81, 2 Figs., Tulsa.

HERBIG, H.-G. (1987): The Upper Cretaceous to Tertiary Hammada west of Errachidia (SE Marocco): A continental sequence involving paleosol development. - N. Jb. Geol. Paläont.Abh. (in press), Stuttgart.

HOTTINGER, L. (1960): Recherches sur les Alveolines du Paleocene et de l'Eocene. - Mèm. Suisses Paléont., 75/76, 243 p., 117 Figs., 1 Tab., 18 Plts., Basel.

IRWIN, M.S. (1965): General theory of epeiric clear water sedimentation. - Bull. Amer. Ass. Petrol. Geol., 49/4, 445-459, 12 Figs., Tulsa.

ISMAIL, M.M & ABDALLAH, A.M. (1966): Contribution to the stratigraphy of St. Paul monastery area by microfacies. -Bull. Fac. Sc. Alexandria, Vol. VII, 325-334, 5 Plts., Alexandria.

JAMES, S. (1984): Shallowing-Upward Sequences in Carbonates. In WALKER, R.G. (Ed.): Facies models. - Geosc. Canada Ser.,1, 213-228, 19 Figs., Hamilton.

KLITZSCH, E. (1984): Northwestern Sudan and bordering areas: geological development since Cambrian time. -Berliner geowiss. Abh., (A) 50, 23-45, 3 Figs., 3 Plts., Berlin.

--- (1986): Plate tectonics and cratonal geology in northeast Africa (Egypt, Sudan). -Geol. Rdsch.,75/3, 755-768, 7 Figs., Stuttgart.

--- & LEJAL-NICOL, A. (1984): Flora and fauna from strata in Southern Egyot and Northern Sudan. -Berliner geowiss. Abh., (A) 50, 47-59, Berlin.

KORA, M. & JUX, U. (1986): On the early Carboniferous macrofauna from the Um Bogma Formation, Sinai. - N. Jb. Geol. Pal. Mh., 1986/3, 85-98, 5 Figs., Stuttgart.

KOSTANDI, A.B. (1959): Facies maps for the study of the Paleozoic and Mesozoic sedimentary basin of the Egyptian region. - First Arab. Petrol. Congr., Cairo, 2, 54-62, Cairo.

KUSS, J. (1986a): Facies Development of Upper Cretaceous - Lower Tertiary sediments from the monastery of St. Anthony/Eastern Desert, Egypt. - Facies, 15, 177-194, 3 Plts., 3 Figs., Erlangen.

KUSS, J. (1986b): Upper Cretaceous calcareous algae from the Eastern Desert of Egypt. -N. Jb. Geol. Paleont. Mh., 1986/4, 223-238, 6 Figs., 2 Tabs., Stuttgart.

KUSS, J. (1987a): Upper Cretaceous - Lower Tertiary calcareous algae of Egypt. - (in prep.)

KUSS, J. (1987b): Faziesverzahnung der karbonatischen Ober-Cenoman/=ber-Turon Ablagerungen im Bereich des Wadi Qena - Northern Galala (Eastern Desert, Egypt).- (in prep.)

LAND, L.S. & DUTTON, S.P. (1978): Cementation of a Pennsylvanian deltaic sandstone; isotopic data. - Journ. of Sed. Petrol., 48, 1167-176, Tulsa.

LEWY, Z. (1975): The Geological History of Southern Israel and Sinai During the Coniacian. - Israel Journ. of Earth-Sc., Vol. 24, 19-43, 15 Figs., Jerusalem.

LEINFELDER, R.R. (1985): Cyanophyte Calcification, Morphotypes and Depositional Environments (Alenquer Oncolite, Upper Kimmeridgian ?, Portugal). - Facies, 12, 253-274, 2 Plts., 3 Figs., 2 Tabs., Erlangen.

MALCHUS, N. & KUSS, J. (1987): Microfacies studies of carbonate rocks from Jurassic outcrops in the north of Faro, Southern Portugal. - 4 Plts., 7 Figs. (in press).

MAMET, B. & OMARA, S. (1969): Microfacies of the Lower Carboniferous dolomitic limestone formation of the Um Bogma terrane (Sinai, Egypt). - Contr. Cushman Found, Foram. Res., 20/3, 106-109, 1 Fig., Bridgewater.

MONTY, C. (1967): Distribution and structure of recent stromatolitic algal mats, Eastern Andros Island, Bahamas. -Ann. Soc. Geol. Belgique, 90/3, 55-100, 19 Plts., 13 Figs., Bruxelles.

MOUNT, J. (1985): Mixed siliciclastic and carbonate sediments: a proposed first-order textural and compositional classification. - Sedimentology, 32, 435-442, 3 Figs., Oxford.

OMARA; S. (1965): A micropaleontological approach to the stratigraphy of the Carboniferous exposures of the Gulf of Suez region. - N. Jb. Geol. Paläont. Mh., 1965/7, 409-419, 2 Figs., Stuttgart.

--- & VANGEROW (1965): Carboniferous (Westphalian) foraminifera from Abu Darag Eastern Desert, Egypt. - Geol. en Mijnbouw, 44, 87-93, 1 Plt., 1 Tab., 4 Figs., Dordrecht.

RADOICIC, R. (1966): Microfacies du Jurassique des Dinarides externes de la Yougoslavie. - Geologija, 9, 377 p., 165 Plts., 11 Figs., Ljubljana.

SADEK, H. (1929): The Geography and Geology of the District between Gebel Ataqa and El Galala Li Bahariya (Gulf of Suez). - Geol. Surv. Egypt. Cairo.

SAID, R. (1960): Planktonic foraminifera from the Thebes Formation, Luxor, Egypt. -Micropaleontology, 6, 227-286, London.

SCHÜRMANN, H.M.E., BURGER, D. & DIJKSTRA, S.J. (1963): Permian near Wadi Arba Eastern Desert of Egypt. - Geol. en Mijnbouw, 42, 329-336, Dordrecht.

SCHWEINFURTH, G. (1886): Sur la decouverte d'une Faune paleozoique dans le gres d'Aegypte.- Bull. de l'Institute egyptien, II Ser., No.6, 283-342, Cairo.

SEILACHER, A. (1983): Upper Paleozoic trace fossils from the Gilf Kebir - Abu Ras area in southwestern Egypt. - Journ. Afr. Earth Sc., 1/1, 21-34, 31 Figs., Oxford.

SIREL, E. & OZCAN, A. (1975): The occurence and age of Miscellanea meandrina (CARTER) in Turkey. - Bull. Geol. Soc. Turkey, 18, 193-194, Istanbul.

SNAVELY, P.D., GARRISON, R.E. & MEGUID, A.A.A. (1979): Stratigraphy and regional depositional history of the Thebes Formation (Lower Eocene), Egypt. - Ann. Geol. Surv. Egypt, IX, 344-362, 5 Figs., Cairo.

STROUGO, A. & HAGGAG, A.Y. (1983): The occurence of deposits of Paleocene age at Abu Roash, west of Cairo Egypt. - N. Jb. Geol. Paläont. Mh., 1983(11), 677-686, 4 Figs., Stuttgart.

--- (1986): The Velascoensis Event: A Significant Episode of Tectonic Activity in the Egyptian Paleogene. - N. Jb. Geol. Paläont. Abh., 173/2, 253-269, 10 Figs., Stuttgart.

TOOMEY, D.F. (1983): The Palecology of a "Middle Limestone Member" (Leavenworth) of an Upper Carboniferous (Stephanien) Cyclothem, Midcontinent, U.S.A. - Facies, 8, 113-190, 15 Plts., 15 Figs., 8 Tabs., Erlangen.

VAIL, P. & HARDENBOL, J. (1979): Sea-level changes during the Tertiary. - Oceanus, 22, 71-79, Woods Hole/Mass.

WALTHER, J. (1890): Über eine Kohlenkalk-Fauna aus der ägyptisch-arabischen Wüste. - Z. dt. geol. Ges., 42, 419-449, 2 Figs., 6 Plts., Stuttgart.

WEISSBROD, T. (1970): "Nubian Sandstone": Discussion. - Amer. Ass. of Petrol. Geol. Bull., 54/1, 526-530, 1 Pl., Tulsa.

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## UPPER CARBONIFEROUS PALEOFLORA FROM BIR QUISEIB (NORTHERN GALALA, GULF OF SUEZ)

- see Chapter 3.5 and Plate 2 -

The flora from Bir Quiseib (section 11/2) is an association of the following taxa:

- Lycophyta: *Sigillaria ichtyolepis* CORDA 1845
  *Sigillaria brardi* BRONGNIART 1828
  *Syringodendron* sp. STERNBERG 1820
  *Lepidodendron posthumi* JONGMANS & GOTHAN 1935
  cf. *Lepidocarpon* sp. SCOTT 1902
  a new species of *Tunguskadendron* THOMAS & MEYEN 1984
- Sphenophyta: *Equisetites* STERNBERG 1833
- Pteridophylla: *Sphenopteris* aff. *souichi* ZEILLER 1888
- Coniferophyta: *Lebachia* aff. *hypnoides* FLORIN 1940
  *Walchia* sp. STERNBERG 1825
  a new species of *Lebachia*

The stratigraphic position of the determined flora is drawn in the following scheme:

| | NAMUR. | WESTPH. | STEPH. | PERM. |
|---|---|---|---|---|
| *Sigillaria brardi* | | ----- | ----- | ----- |
| *Sigillaria ichtyolepis* | | | ----- | |
| *Syringodendron* sp. | | ----- | ----- | ----- |
| *Lepidodendron posthumi* | | | ----- | ----- |
| *Lepidocarpon* sp. | | ----- | ----- | ----- |
| *Equisetites* sp. | | | ----- | ----- |
| *Sphenopteris* aff. *souichi* | ----- | ----- | ----- | |
| *Lebachia* aff. *hypnoides* | | | ----- | ----- |
| *Walchia* sp. | | | ----- | ----- |

It appears that most of these species have an Upper Carboniferous respectively Stephanian age.

### Paleoclimatology and paleoecology:

This flora indicates that the climate was warm and sunny (with rainy periods) during deposition of Upper Carboniferous sediments. The organisation of the Lycophyta (Sigilliaria, Lepidodendron, Syringodendron) shows that they grew in swamp areas (parichnos and ligula) during a warm and humid climate, associated with Pteridophylla and Sphenophyta. The Coniferophyta with short leaves and a thick cuticula were adapted to a more temperate climate and grew in topographic higher positions, possibly on the slopes of mountains.

### Palaeophytogeography:

This flora is an association of mostly Euramerican forms (*S. brardi*, *S. ichtyolepis*, *Equisetites*, *Lebachia*, *Walchia*) and only one Cathaysian form (*L. posthumi*) with a new species of a Siberian genus.

These observations confirm that during Upper Devonian and Lower Carboniferous, the affinities of the floras of NE Africa were mostly Gondwanian, whereas during Upper Carboniferous the floras were mainly Euramerican.

Plate 1: TYPICAL PALEOZOIC TRACE FOSSILS

Fig. 1: *Chondrites*-like tracefossils (21/2,3; x 2).

Fig. 2: *Nereites*-like fecal burrow filled with pellets following bedding plane (24/2,10; x 1,8).

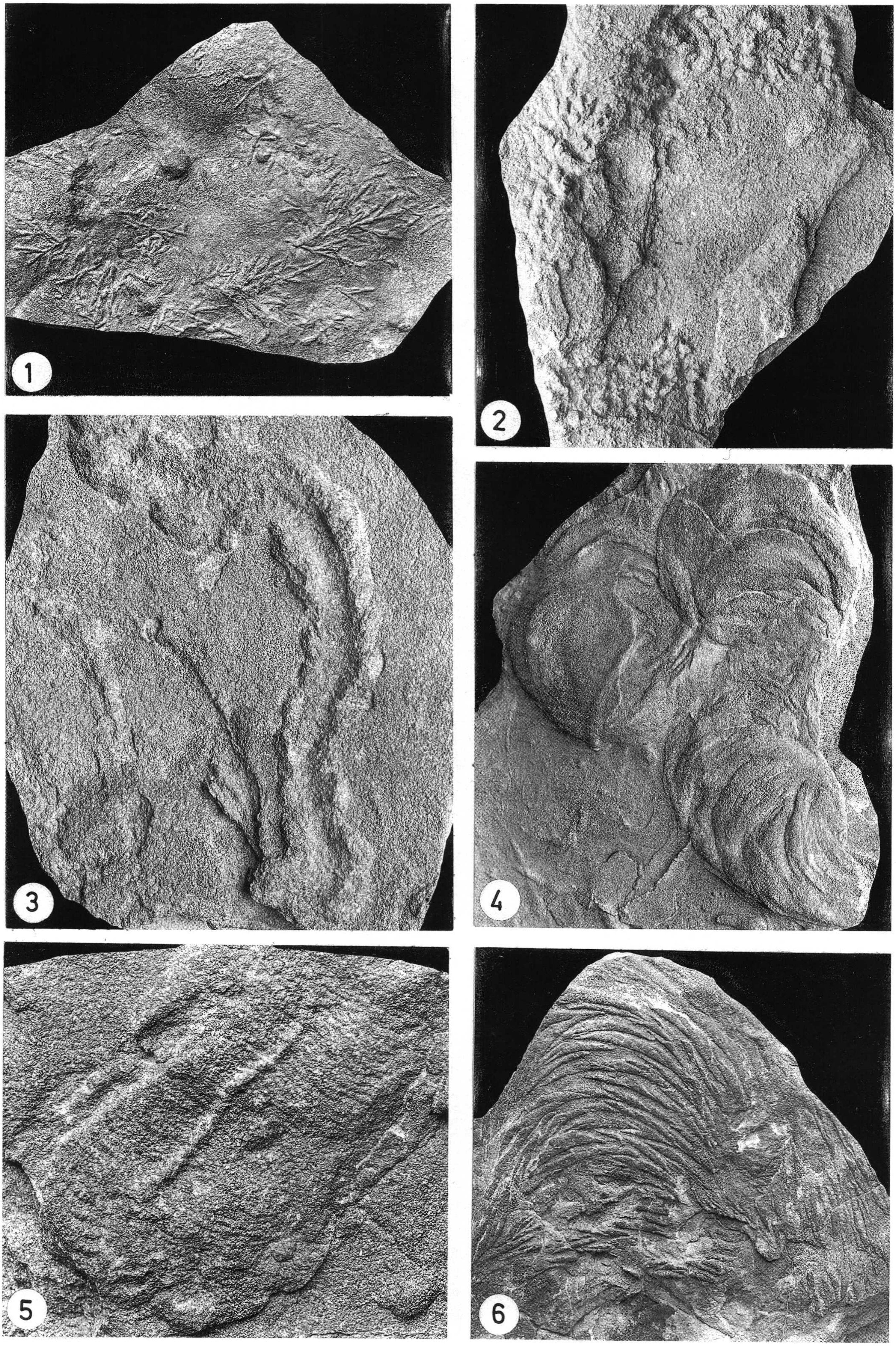
Fig. 3: *Scolicia*-like borrow following bedding plane (11/2,16; x 1,3).

Fig. 4: *Zoophycos*-like feeding structures (16/2,8; x 1,3).

Fig. 5: Complex feeding burrow that follows bedding plane (21/2,1; natural sizes).

Fig. 6: *Lophoctenium*-like spreiten burrow; similar to *Zoophycos* feeding burrows (10/2,1; x 1,8).

# Plate I



# Plate 2

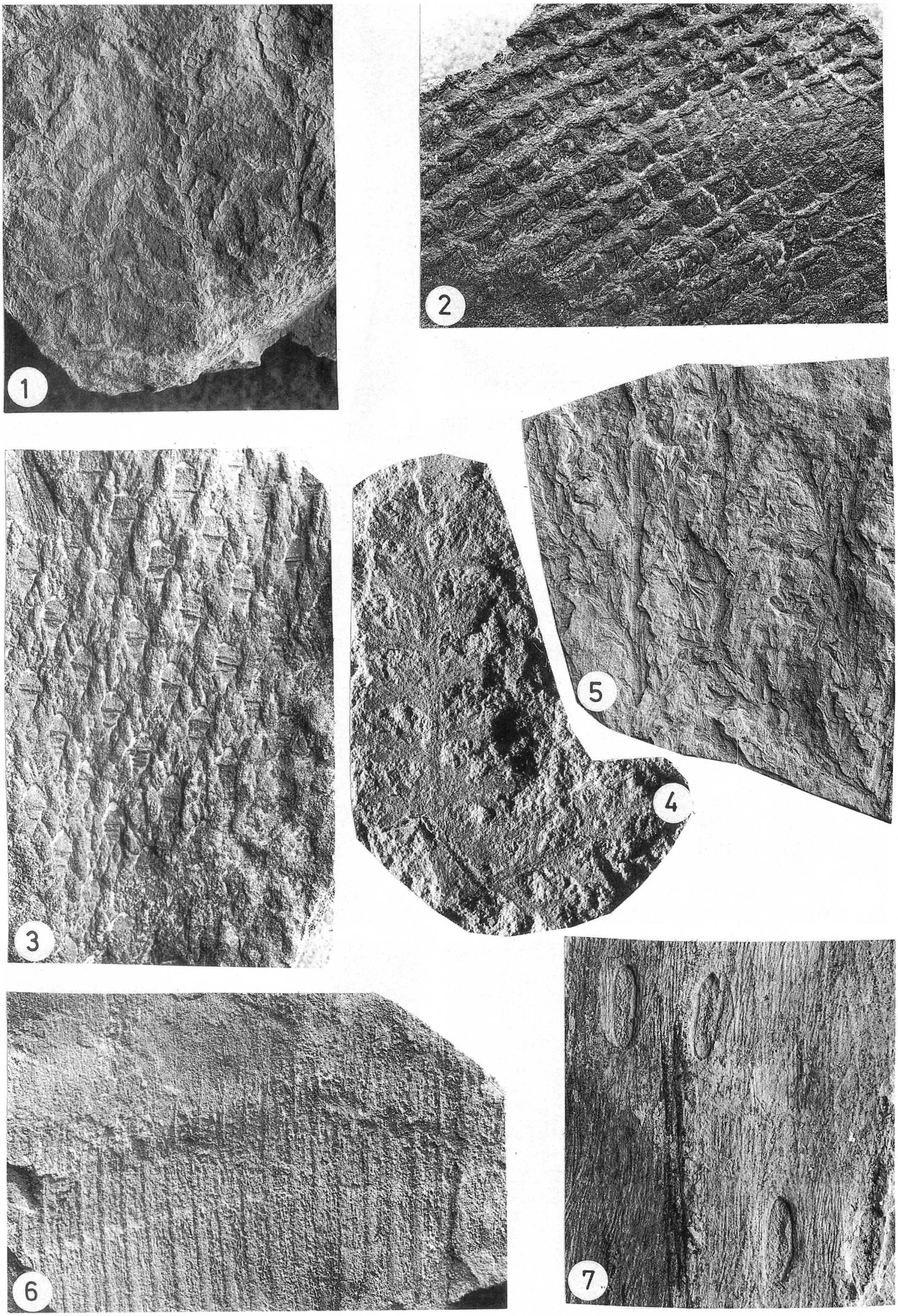


Plate 2: PLANT REMAINS FROM THE CARBONIFEROUS OF THE NORTH GALALA

Fig. 1: *Lebachia* aff. *hypnoides* FLORIN 1940.
(B 502), X 2,3.

Fig. 2: *Sigillaria ichthyolepis* CORDA 1845.
(B 508), X 1,3.

Fig. 3: *Lepidodendron posthumi* JONGMANS & GOTHAN 1935.
(B 509), X 2,3.

Fig. 4: *Sphenopteris* aff. *souichi* ZEILLER 1888.
(B 513), X 5,5.

Fig. 5: *Walchia* STERNBERG 1825.
(B 511), X 2,5.

Fig. 6: *Equisetites* STERNBERG 1833.
(B 512), X 3.

Fig. 7: *Syringodendron* STERNBERG 1820.
(B 516), X 1,5.

PLATE 3: MICROFACIES TYPES AND THE MOST IMPORTANT CONSTITUENTS OF THE CARBONIFEROUS AND JURASSIC LIMESTONES

Fig. 1 - 11: Carboniferous limestones

Fig. 1: Carboniferous limestone of unit 11: Grainstone with many sections of *Hemigordius harltoni*, few of *Pseudobradyina* sp., brachiopod and bivalve shells, echinoid fragments, plenty of small, subangular quartzes and one big, rounded quartz grain with thin micritic crust (21/2,2; x 14,5).

Fig. 2-4: Different nearly axial sections of *Hemigordius harltoni* with typical streptospiral coiling of the first chambers; subsequent growth continues planispiral (All 21/2,2; 2: x 90; 3: x 95; 4: x 85).

Fig. 5: Carboniferous dolomitic limestone of unit 10; only peloid remains and few foraminifera could be indentified. Intercrystalline dolomite pores were filled by black opaque minerals (10/2a; x 21).

Fig. 6: Grapestone composed of big, well-rounded quartz grains, with a thin outher micritic crust and recrystallized cements in the inner parts (21/2,2; x 18).

Fig. 7-8: Vertical (7) and oblique (8) sections of *Pseudobradyina* sp. within unit 11 (Both from 21/2,2; 7: x 110; 8: x 105).

Fig. 9: Axial and oblique sections of *Endothyranella* sp. (21/2,2: x 110).

Fig. 10: Nearly axial section of *Tetrataxis* sp. (10/2: x 55).

Fig. 11 - 20: Jurassic limestones

Fig. 11,12: Mixed grainstone with many intraclasts, oncoids, cortoids, quartz grains lumps and few algae (25/2, 1; x 8,5).

Fig. 13: Lumachelle packstone composed of bivalves, which were leached out and internal filled by sandy-micritic sediment indicating, that top lies left (25/2,2; x 5,5).

Fig. 14: Leached out cavity (vadose) in the lower half below a shell fragment, which was gravitatively filled by internal sediment; the remaining pore-space was filled with radial-fibrous cements (25/2,2; x 17,5).

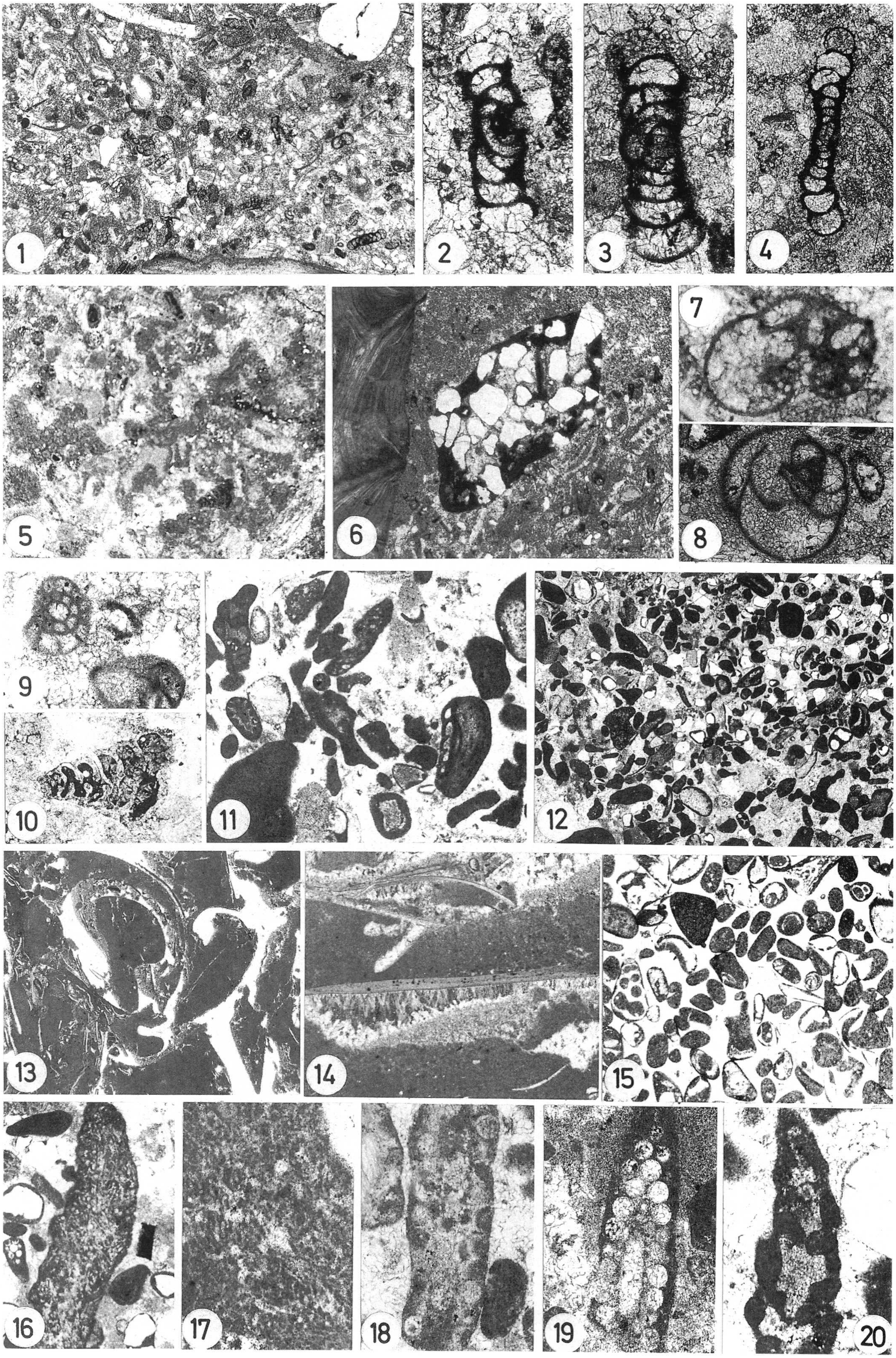
Fig. 15: Gastropod grainstone with cortoids, oncoids and peloids; few components with ooid-crusts (25/2, 1c; x 18,5).

Fig. 16, 17: *Rivularia* sp., section of a small stem (17); 18 shows details of interal structure (both 25/2, 16; 17: x 65; 18: x 115).

Fig. 18, 19: Axial (19) and oblique (20) sections of *Cylindroporella* cf. *arabica* (both 25/2,1; 19: x 60; 20: x 65).

Fig. 20: Nearly axial section of *Rheophax* sp. (25/2, 1; x 59).

Plate 3



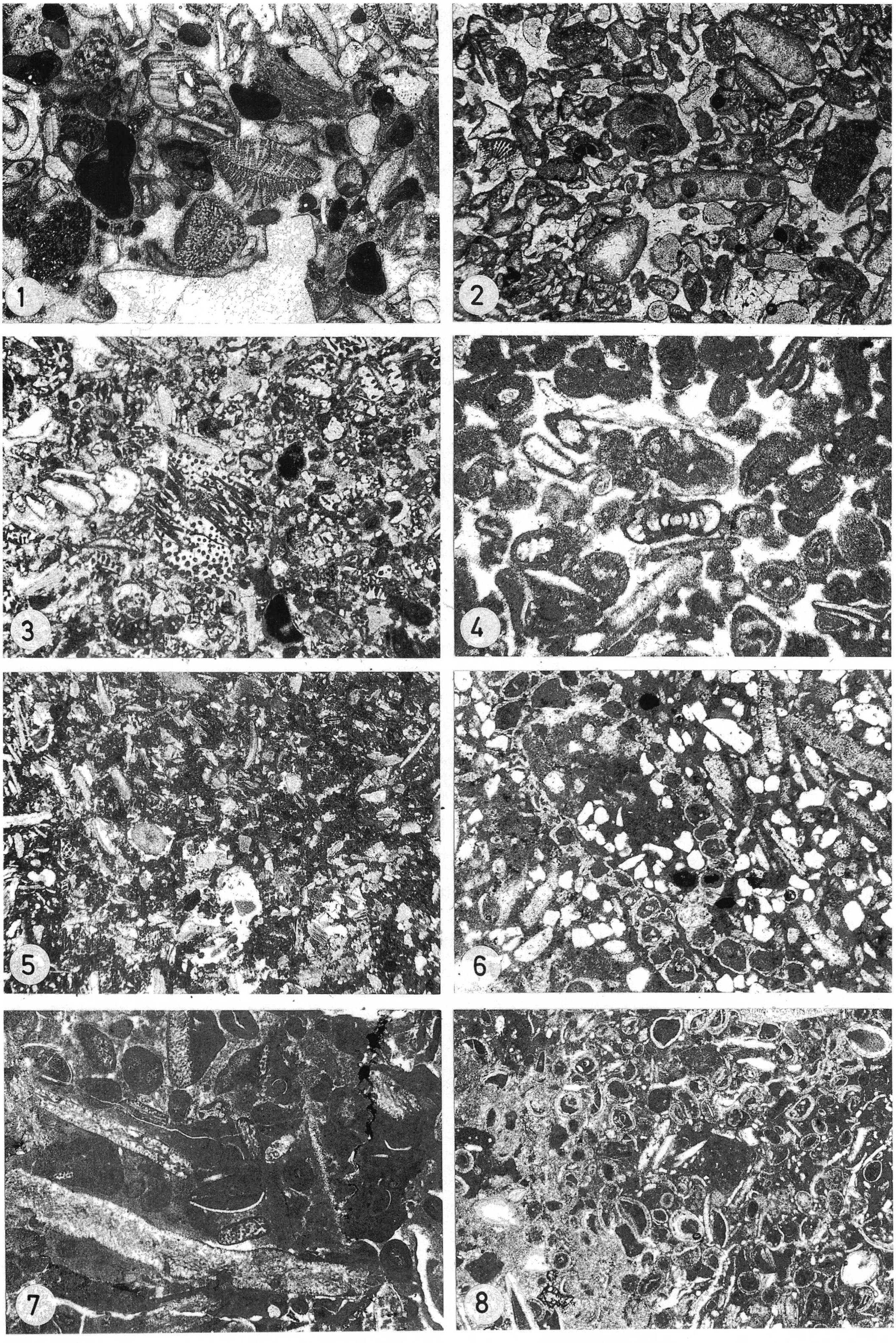

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Plate 4: DIFFERENT MICROFACIES-TYPES FROM UPPER CRETACEOUS

Fig. 1: Bioclastic packstone of Uppermost Campanian/Lower Maastrichtian limestones,which occur only at monastery of St. Anthony. Different sections of Orbitoides media are visible together with extra-clastis (lower left shows reworked biomicrite with planctonics), coated grains, oncoids and oyster-debris (Y5; x 9).

Fig. 2 - 8 represent thin-sections of Upper Cenomanian/Lower Turonian limestones, sampled at the two Galalas as well as at the Northern Wadi Qena.

Fig. 2: Algal biosparite with a random-section of *Neomeris* sp. (centre), intraclasts, cortoids, echinoids and bioclasts (L3,8; x 23).

Fig. 3: Algal packstone mainly composed of broken fragments of *Boueina* sp.; this facies type is very frequent in cavities of big snails, which occur in marly-silty layers (W3,8; x 28).

Fig. 4: Oolithic grain-/packstones with *Nummoloculina* sp. (centre), radial-fibrous ooids and bioclasts, which are often encrusted by thin micritic seams (20/2,4; x 58).

Fig. 5: Bioclastic wacke-/packstone with oyster-debris and echinid-fragments, which show conspicuous even rim cements (20/2,1; x 10).

Fig. 6: Wackestone with a high portion of quartz (micritic sandstone), fragments of echinids and circular section of a bryozoan (?) colony (N9; x 30).

Fig. 7: Oncolithic wackestone with algae of *Boueina pygmaea* (arrow) and *Boueina* sp. (X 17; x 12).

Fig. 8: Oolithic packstone (see Fig.4) with radial-fibrous ooids (partly broken), ostracodes and shell-fragments (11,2,10e; x 15).

Plate 5: DIFFERENT FACIES TYPES FROM UPPER PALEOCENE/LOWER EOCENE LIMESTONES OF THE TWO GALALA PLATEAUS

Fig. 1: Laminated loferitic dolomite (LF-dolomite) with a layered texture caused by rhytmic alternations of thin algal-laminated bindstones with stromatactis-dominated peloidal layers. Arrow shows burrowing tube with internal filling; thinsection of Fig. 4 shows a close-up view (23/2,4; x 1,3).

Fig. 2: Reworked dolomite with lumps, dolomitic shells in the lower part, overlain by a stromatolitic algal mat with shell fragments and burrowing tubes which is again overlain by a dolomitic reworking horizon (8/2,7; x 1,3).

Fig. 3: Rudstone of the shoal/offshore bank facies with plenty of alveolinids (white dots) and dark, irregluar shaped extraclasts of algal packstones - magnified thinsection in Fig. 5 (13/2,3; x 1,3).

Fig. 4: Enlarged section of Fig. 1: Lower (dark) part with micritic, laminated bindstone with birdseyes; upper part is a peloidal grainstone with conspicuous stromatactis fabrics, which are partly filled with vadose silts (23/2,4; x 12).

Fig. 5: Close up of Fig. 3: The central part shows an irregular extraclast composed of monotypic accumulations of *Ovulites morelleti* and very few other dasyclads. Besides different alveolinids, quartz grains and few rotaliids occur (13/2,3; x 8).

Fig. 6: Palustrine limestone with top oriented to the left. Plenty of micritic dark nodules (with small irregular, sparitic voids) lay between big, horizontally elongated voids, partly filled with internal sediment, vadose silts and sparitic calcites. Irregular elongated sparitic voids are due to small roots (8/2,5; x 15).

Fig. 7: Big crustose lump of the shoal facies (rudstone). Multiple alternations of *Solenomeris o'gormanni* and *Lithothamnium* sp. encrust a grapestone. Fragments of reworked red algal crusts, alveolinids and quartz grains occur moreover (8/2,7; x 8).

Fig. 8: Shallow subtidal deposited wackestone with different sections of *Belzungia silvestrii* and *Orbitolites* sp.. Small black pebbles and thick dark crust (with small root-like voids) document infiltration of organics (plant growth), which overprinted the limestones in an early diagenetic stage (8/2,7; x 8).

Fig. 9: Part of a flint nodule within the Thebes Formation. The arrow points to white lining, representing the outer silicified covering with decreasing silification to the centre (left) of the shallow water nodule and a sharp boundary to the pelagic micritic wackestone of Thebes Formation (right) (16/2,3; x 8).

Plate 5

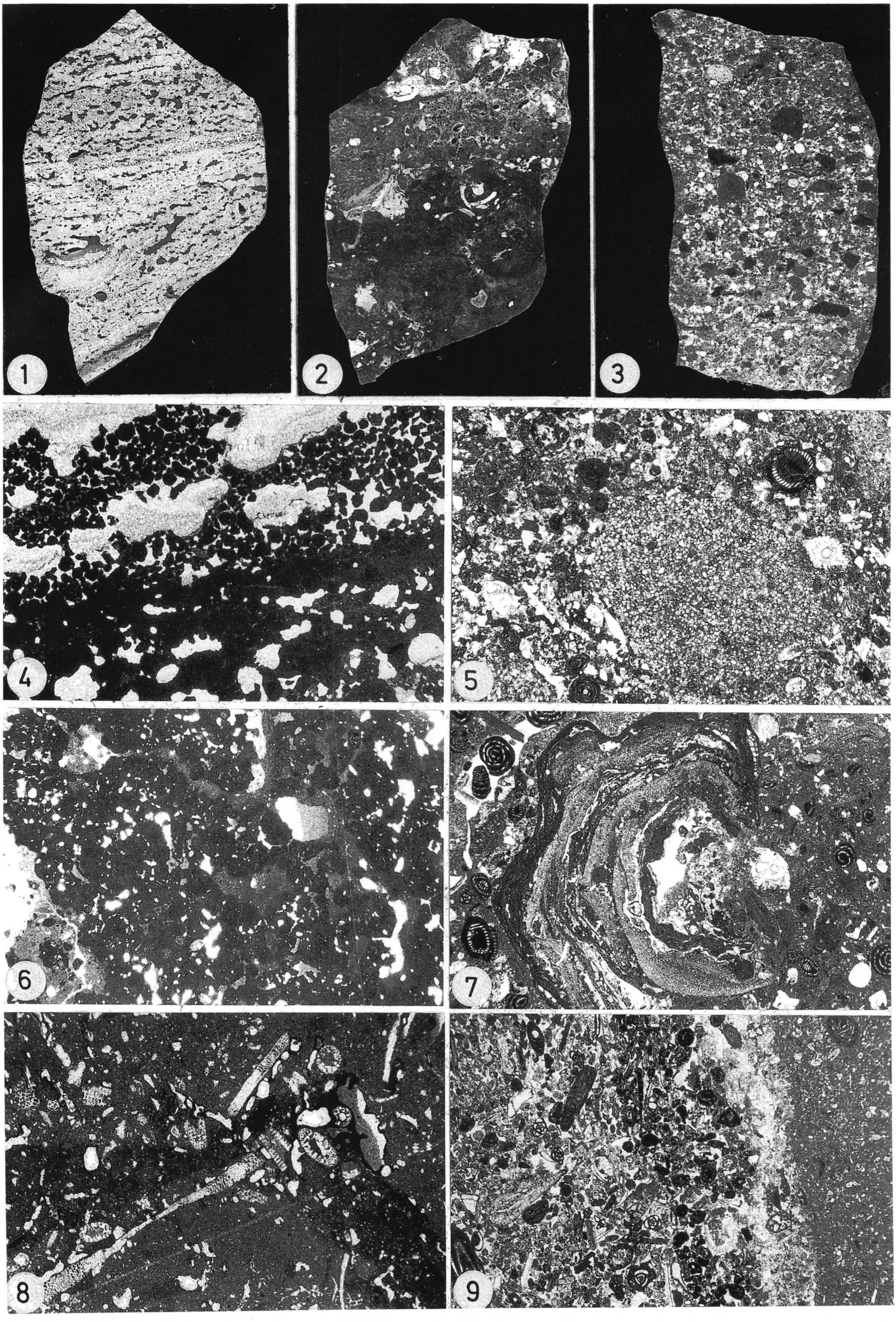

Plate 6

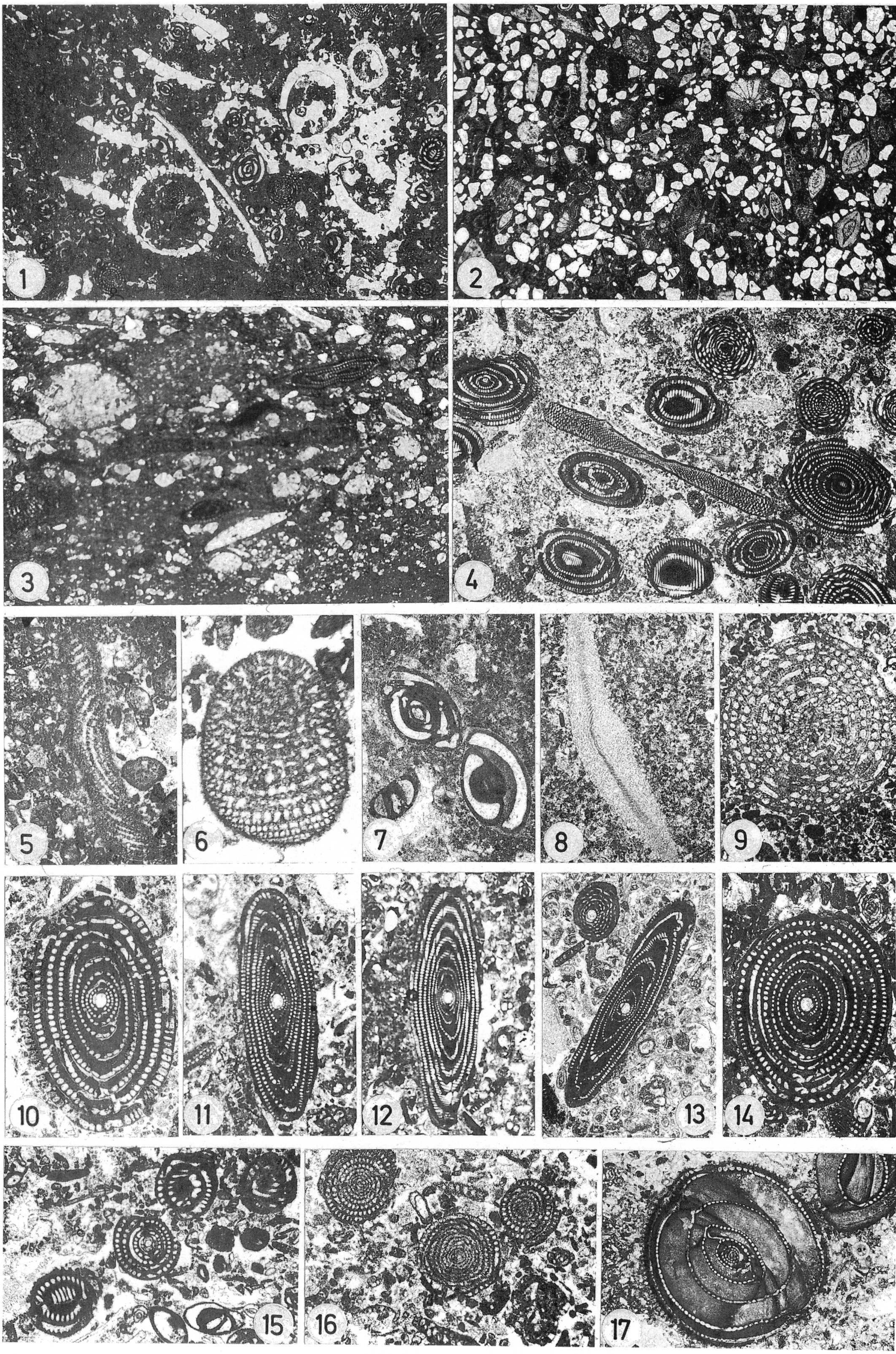

Plate 6: PALEOCENE/EOCENE FACIESTYPES AND TYPICAL FORAMINIFERA

Fig. 1: Subtidal wackestone with *Alveolina primaeva*, *Quinqueloculina* sp. and several sections of dasyclads with unknown systematic position (20/2,13; x 7,6).

Fig. 2: Micritic sandstone with nummulitids and rotaliids (13/2,11; x 7,4).

Fig. 3: Sandy limestone-intercalation with alveolinids and rotaliids within the pelagic Thebes Formation (U3; x 6).

Fig. 4: *Alveolina*-grainstone with *Alveolina dolioliformis*, *Alveolina* sp. and *Orbitolites* sp. (13/2,5; x 8,5).

Fig. 5: *Broeckinella arabica*, within grainstone-facies (20/2,8; x 13,5).

Fig. 6: *Fallotella kochanskae*; oblique section showing the typical pilars (20/2,8; x 19).

Fig. 7: Sections of different miliolids, the upper left possibly of *Idalina* sp. (20/2,5; x 22).

Fig. 8: Nearly axial section of *Discocyclina* sp. (18/2,6; x 12,2).

Fig. 9: Axial section of *Miscellanea meandrospira* (20/2,9; x 9,5).

Fig. 10: Axial section of *Alveolina decipiens* (13/2,$5_{(2)}$; x 16).

Fig. 11: Axial section of *Alveolina oblonga* (U6; x 13,5).

Fig. 12: Axial section of *Alveolina ruetimeyeri* (U10; x 7,5).

Fig. 13: Axial and vertical sections of *Alveolina frumentiformis* (16/2,3; x 10).

Fig. 14: Axial section of *Alveolina dolioliformis* (13/2,3; x 15).

Fig. 15: Different sections of *Alveolina* (*Glomalveolina*)*primaeva*a together with other miliolids and dasyclads (20/2,7; x 13,5).

Fig. 16: Several sections of *Alveolina lepidula* (18/2,12; x 18).

Fig. 17: *Alveolina pasticillata* with eroded senile chambers (18/2,9; x 14).