Age and paleoenvironment of the Nukhul Formation, Gulf of Suez, Egypt: Insights from palynology, palynofacies and organic geochemistry

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ABSTRACT

Palynological results of a detailed study carried out on 56 samples retrieved from two selected wells (GH 404-2A and SA-E6A) of the Hilal and Shoab Ali fields within the southern part of the Gulf of Suez, Egypt, are presented. This study is mainly focused on the poorly dated Nukhul Formation, for which very little information from palynology is available despite its importance from a petroleum viewpoint. The assemblages discovered in our study are moderately preserved and reveal a sparse but significant record of spores and pollen and dinoflagellates together with highly diverse fungi and algal taxa, e.g. *Botryococcus* and *Pediastrum*.

A latest Oligocene–Early Miocene (Chattian–Aquitanian) age has been suggested for the Nukhul Formation, based on compiling palynostratigraphic and ecologic data obtained from palynomorphs that have previously been assumed to be representatives for this period on a regional scale. In addition, the Oligocene/Miocene Boundary (OMB) could be lithostratigraphically defined within the studied formation, most likely at the boundary between the lower Shoab Ali Member and upper Ghara Member. A fungal/algal 'event' within the interval from 11,370–11,430 ft in the GH 404-2A Well may be associated with a strong regressive phase. Such a regression was previously observed in the Nile Delta and other locations around the Red Sea province, and may be assigned to the global Mi-1 glaciation event at the OMB. However, not only glacial-driven eustacy but also tectonic activity related to the Gulf of Suez rifting may have contributed in forming such an event.

Palynofacies investigations were carried out under both transmitted and fluorescence microscopy and the results were partly supplemented by existing organic geochemical analyses (GH 404-2A Well) involving Rock-Eval pyrolysis and total organic carbon (TOC) measurements. The analysis was used to interpret the depositional regime, paleoenvironment and thermal maturation history of the studied succession. These results support the temporary existence of shallow, pond- or lake-like aquatic habitats during deposition of the lower Shoab Ali Member that evolved into a shallow-marine environment with the onset of the deposition of upper Ghara Member of the Nukhul Formation.

INTRODUCTION

In 1886, the first discovery and production of oil took place in the Middle East at Gemsa, Gulf of Suez, Egypt (Figure 1). Since then, large-scale commercial exploration and production have been ongoing. The major research focus in the Gulf of Suez has so far been on structural geology, geophysics, organic geochemistry, sedimentology, stratigraphy and micropaleontology (e.g. Souaya, 1966; Scott and Govean, 1985; Richardson and Arthur, 1988; Montenat et al., 1988; Evans, 1988, 1990; Hughes et al., 1992; Patton et al., 1994; Schütz, 1994; Bosworth and McClay, 2001; Alsharhan, 2003; Youssef, 2011).

Previous palynological studies of subsurface Miocene sediments of the Gulf of Suez, Egypt, are very scarce and have concentrated almost solely on taxonomy, biostratigraphy and fragmental interest on paleoenvironmental analyses (Mahmoud, 1993; Ahmed and Pocknall, 1994; El Beialy and Ali, 2002; El Beialy et al., 2005; Soliman et al., 2012). These studies have demonstrated the
Figure 1: (a) Location map of Egypt showing the Gulf of Suez.
(b) The Gulf of Suez forms the northwestern extension of the Red Sea Rift System.
(c) The studied wells (red circles), Gulf of Suez, Egypt (based mainly on GUPCO, 1983, 1986).
presence of abundant palynomorphs, especially dinoflagellate cysts (dinocysts) in the Miocene Rudeis and Kareem formations, but only few data exist from the Nukhul Formation of the Gharandal Group, as well as the overlying Ras Malaab Group (Table 1).

Most wells in the Gulf of Suez are drilled on structural highs in carbonate facies, where the Nukhul Formation tends to be thin. As a result, sufficient information for detailed palynological analysis from thick Nukhul succession is not available (cf. Schütz, 1994). Up to now, no palynological investigations have addressed palynofacies, organic thermal maturation or the source rock potential of these formations in general, and for the Nukhul Formation in particular, although it has a good hydrocarbon potential as a source and reservoir rock (Schlumberger, 1984; Hughes et al., 1992; Winn et al., 2001). Hence, the present study deals with palynofacies, biostratigraphic analyses and organic geochemistry of 56 cutting samples from the poorly studied Nukhul Formation.

Since most micropaleontologic studies carried out on the Nukhul Formation, especially on its lower Shoab Ali Member, proved to be nearly barren of foraminifera and nannoplankton, we believe that a study of spores, pollen and dinocysts has a great potential to supplement already existing data.

The aims of this work are to: (1) review the geologic history and refine the microfloral framework of the sedimentary succession within the Nukhul Formation of the Gulf of Suez; (2) discuss the distribution of palynomorphs present in the Nukhul Formation. Very few palynological studies have been performed, and it is the intention to correlate these palynomorph assemblages with their counterparts in Africa, the Tethyan realm and worldwide; (3) focus on the spores and pollen events around the Oligocene/Miocene Boundary (OMB) as an independent proxy for biostratigraphic control; (4) interpret the presence of a fungal/algal ‘proliferation’ (El Atfy et al., in press) that may be related to eustatic and/or tectonic events at or near the OMB; (5) focus on the palynomorphs and kerogen distribution, supplemented in part with total organic carbon (TOC) and Rock-Eval measurements; (6) review the use of non-pollen palynomorphs (e.g. fungal palynomorphs and freshwater algae) as paleoecologic indicators within the studied sequences; and (7) portray the climatic and paleoenvironmental conditions that prevailed during the deposition of the Nukhul Formation, in comparison with global events.

**MATERIAL AND METHODS**

Fifty-six cutting samples have been collected from the Nukhul Formation encountered in the GH 404-2A (27°50’54.65”N, 33°42’23.30”E; Hilal Field) and SA-66A (27°50’53.25”N, 33°51’51.07”E, Shoab Ali Field) wells, southern Gulf of Suez, Egypt (Figure 1). The samples were prepared for palynological analysis following the standard extraction HCl-HF treatment (e.g. Traverse, 2007), without any oxidation. As the studied samples were retrieved from oil-based mud boreholes, the presence of hydrocarbon contamination was noted and as a result we used Triton X-100 as an emulgating agent to get rid of these contaminants (cf. Nørgaard et al., 1991). The organic residue was stained with Safranin solution, sieved using a 10 µm mesh and mounted on slides using ICI’s Elvacite acrylic resin. For TOC and Rock-Eval pyrolysis preparation, see GUPCO (unpublished report, 1984).

Most of the studied samples were poorly fossiliferous and the palynomorph recovery was generally low, with the exception of the presence of some stratigraphically diagnostic taxa. The latter are studied in detail in terms of their age and paleoenvironmental significance. It should be mentioned that a relatively massive fungal proliferation (El Atfy et al., in press) was recorded only in the GH 404-2A Well and not in both studied boreholes. Relevant slides and residues are deposited in the palynological collection of the Section ‘Palynology and Microvertebrates of the Paleozoic’ at the Senckenberg Research Institute and Natural History Museum, Frankfurt am Main, Germany.

**GEOLOGIC SETTING**

The Gulf of Suez is a shallow and narrow body of water, which developed as a failed intracontinental Cenozoic rift system, approximately 300 km long, forming the northern extension of the Red Sea, and covering an area of about 25,000 sq km at an average water depth of 55–100 m.
Table 1
Classification of the Miocene time and rock units in the Gulf of Suez and Sinai, Egypt (based on different authors)

| Moon and Sadek (1923) | Macfadyen (1930) | Stainforth (1949) | EGPC (1964) | Said and El-Heiny (1967) | NSSC (1976) |
|-----------------------|------------------|-------------------|-------------|--------------------------|-------------|
| Gypsum V | Gypsum V | Evaporite Series | Zeit Formation | Evaporite V | Zeit Formation |
| Nullipore rock | Nullipore Rock | Burdigalian–Helvetian | Feiran Member | Feiran Member | Burdigalian–Helvetian |
| Gypsum IV | Gypsum IV | Upper Globigerina Marls | Belayim Member | Evaporite IV | Belayim Formation |
| Gypsum III | Gypsum III | Lower Globigerina Marls | Baba Member | Evaporite III | Baba Member |
| Gypsum I | Gypsum I | Lower inter- gypsum Marl and Gypsum II | Shagaar Member | Evaporite II | Shagaar Formation |
| Miocene Marl | Miocene Marl | Lower inter-gypseous Marl and Gypsum II | Markha Member | Evaporite I | Shagaar Member |
| Miocene Grits | Miocene Grits | Lower Globigerina Marls | Ayun Musa | | |
| Miocene Clays | Miocene Clays | Kareem Formation | Shagaar Member | | |
| Oligocene–Miocene (Pre-Aquitanian–Aquitanian) | Oligocene–Miocene (Pre-Aquitanian–Aquitanian) | Rudeis Formation | Aypie Member | | |
| Nukhul Formation | Khoshera | Nukhul Formation | Basal Miocene | | |
| Chattian–Aquitanian | Sudr | Nukhul Formation | | | |
| Nukhul Formation | Nebwi | Nukhul Formation | | | |
| Nukhul Formation | Ras Matarma | Nukhul Formation | | | |
| Oligocene–Miocene (Pre-Aquitanian–Aquitanian) | Oligocene–Miocene (Pre-Aquitanian–Aquitanian) | Gharandal Group | Aypie Member | | |
| Nukhul Formation | Khoshera | Gharandal Group | Aypie Member | | |
| Nukhul Formation | Sudr | Gharandal Group | | | |
| Nukhul Formation | Nebwi | Gharandal Group | | | |
| Nukhul Formation | Ras Matarma | Gharandal Group | | | |

Legend:
- Conglomerate
- Sandstone
- Shale/clay
- Evaporite
- Limestone
- Marl
- Mixed lithology

Continued
### Table 1 (continuation)

| El-Heiny (1982) | Saoudi and Khalil (1984) | Hosny et al. (1986) | Phillip et al. (1997) | Al-Husseini (2012) | Present Study |
|----------------|--------------------------|---------------------|-----------------------|-------------------|---------------|
| **Oligocene–Miocene Nukhul Formation, Gulf of Suez** |
| **Serravalian–Tortonian** |
| Zeit Formation | Zeit Formation | Zeit Formation | Belayim Formation/ Hammam Faraun Member | Zeit Formation |
| **Langhian** |
| Belayim Formation | Belayim Formation | Belayim Formation | Belayim Formation | Not Studied |
| **Kareem Formation** |
| Kareem Formation | Kareem Formation | Shagaar Formation | Lower Rudeis Formation | Not Studied |
| **Rudeis Formation** |
| Rudeis Formation |
| **Burdigalian** |
| Nukhul Formation | Nukhul Formation | Nukhul Formation | Nukhul Formation | Nukhul Formation |
| **Conference** |
| October Member | Gharmul Member | Ghara Member | Nukhul Member | Nukhul Member |
| **Shoab Ali Member** |

**Note:** The table and diagram illustrate the classification of the Miocene time and rock units in the Gulf of Suez and Sinai, Egypt, based on various authors.
Rifting occurred during the separation of the African and Arabian plates from the latest Oligocene up to the Early Miocene (ca. 24.0–15.5 Ma). Sedimentation rate and tectonic subsidence during the initial phase of rifting was slow and the corresponding stratigraphic sequences show an upward transition from continental volcanics and red beds of the Abu Zenima Formation to marginal marine clastics of the Nukhul Formation (e.g., Patton et al., 1994). Moreover, Bosworth and McClay (2001) assumed that the onset of rifting is no younger than 27–25 Ma for the latitude of Hurghada, 25–23 Ma for the central Gulf and ca. 23.5 Ma for the northern Gulf and Cairo-Suez relay Zone.

The Gulf of Suez is a complex graben/half-graben system located between two basement uplifts: Sinai and the Eastern Desert Mountains of Egypt (Schütz, 1994). The rift is not connected with the Mediterranean Sea, but terminates north of Suez (Colletta et al., 1988) and the structure of this northern portion has been considerably affected by Syrian Arc tectonics (Tewfik, 1988).

The Suez Rift is divided into three sectors of different dip regimes; a northern sector with a main SW tilt, a central sector with a NE tilt and a southern one with a SW tilt, separated by two shear/accommodation zones (Sestini, 1995). Longitudinally, the prominent structures are three major NNW-SSE uplifted basement belts separated by two structural lows, which are associated with large Bouguer gravity anomalies (the Darag, Lagia, Belayim, Gharib, East and West Zeit, Gemsa, South Ghara and Qaa Plain sub-basins, Meshref, 1990).

The Suez Rift is considered to have developed as an element of the two complementary shear fractures of Aqaba and Suez that resulted from early Tertiary persistence of NW-SE compression (Meshref, 1990). Rifting of an incipient graben commenced in the latest Oligocene to the Early Miocene (35–24 Ma), at the same time as early Red Sea rifting (Sestini, 1995). Rapid tectonic subsidence in middle Burdigalian–Langhian was followed by strong block faulting and uplift of rift shoulders, about 19–17 Ma (Evans, 1988; Patton et al., 1994), although the intensive tectonic movements continued until post-Miocene (Sestini, 1995).

Geomorphologically, the Gulf of Suez represents a rejuvenated, slightly arcuate NW-SE topographic depression, traditionally known as the Clysmic rift or Clysmic gulf, named after the ancient Roman settlement of Clyisma that occupied the present site of city of Suez (Hume, 1921; Robson, 1971). It extends northwestward from 27°30’N to 30°00’N, and its width varies from about 50 km at its northern rim to about 90 km at its southern end, where it merges with the Red Sea (Bosworth and McClay, 2001).

The Gulf of Suez is the location of an extensive hydrocarbon play and has excellent hydrocarbon potential. Petroleum exploration around the Gulf of Suez started just over 100 years ago with commercial scale hydrocarbon production. As an oil province, it is ranked seventh in terms of production among the major grabens or rift basins of the world (Clifford, 1986; Schlumberger, 1995), with a prospective sedimentary basin area measuring ca. 19,000 sq km. This province has more than 1,000 exploration wells, with 240 oil discoveries in more than 80 oil fields and it is also considered as the most prolific oil province rift basin in Africa and the Middle East (Schlumberger, 1984; Alsharhan, 2003). The Gulf of Suez is the main oil province in Egypt, with oil being produced from Paleozoic, Mesozoic and Cenozoic rocks; however, the Miocene sediments are the most prolific reservoirs both in onshore and offshore fields (El Ayouty, 1990). Particularly, the Nukhul Formation produces hydrocarbons in 17 oil fields as a result of being the first porous interval above the Eocene source rocks with the overlying Rudeis shales forming a seal (Schütz, 1994).

LITHOSTRATIGRAPHY

The lithostratigraphic column in the Gulf of Suez was subdivided into three megasequences: a pre-rift succession (pre-Miocene or Paleozoic–Eocene), a syn-rift succession (Oligocene–Miocene), and a post-rift succession (post-Miocene or Pliocene–Holocene), with clear variations in their lithology, thickness, areal distribution, depositional environment, and hydrocarbon importance (e.g., Alsharhan, 2003).
The pre-rift succession is generally comparable to that of northern Egypt until the Eocene with a sedimentary record of 1,000–1,200 m in thickness. It comprises the Nubian Complex (Paleozoic–Lower Cretaceous) at the base, overlain by a Cenomanian–Eocene sequence (Sestini, 1995).

The Abu Zenima Formation (Hantar, 1965; Sellwood and Netherwood, 1984) represents the earliest syn-rift deposits in the Gulf of Suez (Bosworth and McClay, 2001) and it is mainly composed of a succession of red and white sandstones and siltstones with a basal conglomerate bearing basalt pebbles. It was dated as Late Oligocene–Early Miocene (Chattian–Aquitanian), mainly based on K-Ar dating of basalt flows (Plaziat et al., 1998). It is overlain unconformably by the Nukhul Formation but probably has no stratigraphically equivalent outcrops in the southern Gulf of Suez, although some authors working on subsurface sections correlate the basal member of the Nukhul Formation (Shoab Ali Member) with the Abu Zenima Formation (Bosworth and McClay, 2001).

The Miocene syn-rift deposits exhibit numerous changes in thickness and facies resulting in a multiplicity of rock unit names (Table 1), based both on offshore and outcropping sections, which leads to some confusion and inconsistencies (Montenat et al., 1988; Al-Husseini, 2012). Moreover, a notable lateral facies variation characterizes the Rudeis, Kareem and Belayim formations, due to the Middle Miocene tectonic rearrangement and differential subsidence (Sellwood and Netherwood, 1984).

The Miocene deposits in the Gulf of Suez and in western Sinai have been investigated and classified by numerous studies (e.g. Egyptian General Petroleum Corporation Stratigraphic Committee – EGPC, 1964; National Stratigraphic Sub-Committee – NSSC, 1976; Saoudi and Khalil, 1984), and as a result, the Miocene sequence of the Gulf of Suez is commonly subdivided into the Lower Miocene Gharandal Group, and the Middle–Upper Miocene Ras Malaab Group. Both groups are important regarding hydrocarbon exploration and generation; the lower contains the richest source rocks in combination with excellent reservoirs (e.g. Nukhul Formation), deposited under favorable structural conditions, whereas the upper one provides the most efficient seal for both Miocene and pre-Miocene reservoirs (Alsharhan, 2003).

**NUKHUL FORMATION**

The term “Nukhul Formation” was introduced by Waite and Pooley (1953) for the basal marine Miocene beds in Sinai and the Gulf of Suez regions, and then formally established by EGPC (1964); it is synonymous to the Basal Miocene of Hume et al. (1920) and the Burdigalian of Macfadyen (1930). Its type section is 60 m thick at the tributary south of Wadi Nukhul, central Sinai coastal Gulf of Suez, Egypt (Figure 1). In subsurface sections, it is absent on structural highs and reaches more than 700 m in the depocenters of the rift sub-basins (Richardson and Arthur, 1988). The Nukhul Formation was subdivided by Waite and Pooley (in Schlumberger, 1984) by means of electric well logs at some localities in west central Sinai into four informal members designated from top to bottom as: Khoshera, Nebwi, Sudr, and Ras Matarma members. Their lithology comprises alternating gray shale with calcareous sandstones commonly grading into sandy limestones (NSSC, 1976).

The Nukhul Formation (a subdivision of the Gharandal Group) is, in general, poorly dated due to the scarcity of diagnostic fossils (El-Heiny and Martini, 1981; Evans, 1988; Soliman et al., 2012). Nevertheless, it has repeatedly been assigned to the Early Miocene (e.g. Souaya, 1966; NSSC, 1976; Andrawis and Abdel Malik, 1981; El-Heiny and Martini, 1981; Hermina and Lindenberg, 1989; El Heiny and Morsi, 1992; Evans, 1988; Phillip et al., 1997; El-Deeb et al., 2004; El-Barkooky et al., 2006; Issawi et al., 2009; Youssef, 2011) based on foraminifera, and calcareous nannoplankton and its stratigraphic position. Palynologically, mainly based on dinocysts, it has previously been dated as Aquitanian (Ahmed and Pocknall, 1994) and Aquitanian to mid-Burdigalian (Soliman et al., 2012). Recently, Hewaidy et al. (2012) determined the Nukhul Formation as Late Oligocene–Early Miocene, based on planktonic and benthonic foraminifera from a surface exposure in Sinai; the same age was postulated from Wadi Dib-1 Well, Gulf of Suez according to a tuned orbital-forcing glacio-eustatic time scale by Al-Husseini (2012). The Nukhul Formation overlies pre-rift rocks ranging
from Precambrian to Middle–Late Eocene, depending upon the degree of uplift and erosion prior its deposition (Richardson and Arthur, 1988).

Saoudi and Khalil (1984) divided the Nukhul Formation into four members. The lower Shoab Ali Member consists of continental clastics, which were deposited in the oldest syn-rift grabens of the Gulf of Suez. Above the Shoab Ali Member, the October, Ghara and Gharamul members are laterally equivalent and characterized by different facies. The four members are well defined in many localities in the Gulf of Suez and Red Sea basins but they are differently named (cf. Al-Husseini, 2012).

For our study, we adopted the nomenclature of Saoudi and Khalil (1984) and EGPC (1996) by using the terms Shoab Ali and Ghara members. The description of these units is lithologically similar to the one in the vicinity of our study area. The lithologic change is also seen from the gamma-ray (GR) and sonic well logs (Figures 2a, b). The description below discusses in descending stratigraphic order their correlation and detailed characteristics for each unit.

Ghara Member

The Ghara Member was originally described from the well GS 391-2 by Saoudi and Khalil (1984). It is mainly composed of white, hard anhydrite layers interbedded with sandstones, gray marls, calcareous shales and limestones. In the southern Gulf of Suez, it contains up to four distinct evaporitic layers, semi-regionally developed as marker beds (Saoudi and Khalil, 1984). The studied interval of the Ghara Member in the GH 404-2A Well ranges from 10,840 to 11,400 ft (560 ft thick) and from 4,842 to 5,098 ft (256 ft thick) in SA-E6A Well. It conformably overlies the Shoab Ali Member and is unconformably overlain by the Lower Rudeis Member or its correlative Mheiherrat Formation.

According to the correlations proposed by Al-Husseini (2012), the Ghara Member is stratigraphically equivalent to (1) Subgroup Ae, lacustrine deposits, early evaporites or Nukhul evaporite (Montenat et al., 1986, 1988), (2) lower evaporites or earlier sulfate deposits (Orszag-Sperber et al., 1986), and (3) Yanbu Formation in Saudi Arabia (Hughes and Filatoff, 1995; Hughes et al., 1999; Hughes and Johnson, 2005).

Shoab Ali Member

The Shoab Ali Member was originally described from the GH 385-1 Well by Saoudi and Khalil (1984). It is mainly restricted to the southern part of the Gulf of Suez and is mainly composed of sand and sandstone; the sand is generally loose, colorless, pink or yellow, fine- to medium-grained, becoming coarser near the bottom. The sandstone is well- to fairly well-sorted, sub-rounded, with streaks of reddish brown shale, which is mostly barren of fauna. These sands and sandstones are porous and constitute an excellent petroleum reservoir (Hughes et al., 1992).

The studied interval in the GH 404-2A Well ranges from 11,400 to 11,774 ft (374 ft thick) and from 5,098 to 5,700 ft (602 ft thick) in SA-E6A Well. It unconformably overlies the ?Thebes Formation, and is conformably overlain by the Ghara Member.

Similar to the correlations proposed by Al-Husseini (2012) the Shoab Ali Member is a synonym to (1) the Abu Zenima Formation of Hantar (1965) and Bosworth and McClay (2001), (2) Subgroup A1 of Group A, basal Red Series or Red Bed Series (Montenat et al., 1986, 1988) or Subgroup Ar (Thiriet et al., 1986), (3) Lower part of Umm Abbasa Formation and Rosa Member of the Ranga Formation (Cindy, 1963; Issawi et al., 1981; NSSC, 1976; Montenat et al., 1998), and (4) Lower part of the Abu Ghusun Formation (Mazhar and Labib, 1966). Outside Egypt and within the Red Sea province it can also be correlated with the Al Wajh Formation from Saudi Arabia (Hughes and Filatoff, 1995; Hughes et al., 1999; Hughes and Johnson, 2005), the Hamamit Formation (Bunter and Abdel Magid, 1989) from Sudan, and the Dogali Formation (Savoyat et al., 1989) from Eritrea.
Figure 2: Lithologic and electric logs of the Nukhul Formation within the studied wells, Gulf of Suez, Egypt (modified after GUPCO, 1983, 1986).
Plate 1: Microscope photographs of palynomorphs and palynofacies assemblages from the Nukhul Formation, GH 404-2A and SA-E6A Wells, Gulf of Suez, Egypt. An England Finder reference (e.g., EF W43-4) follows the sample (e.g. GH04-172) and slide number (e.g. 1, 2) for each specimen. Photomicrographs are in bright field, magnifications are in the form of bar scales. 

(1) Verrucatosporites usmensis; GH04-172-1_EF W43-4. (2) Verrucatosporites favus; GH04-181-1_EF O51-1. (3) Magnastriatites howardi; GH04-168-1_EF K42-4. (4) Foveotriletes margaritae; SA-E6A-23-1_EF V52-2. (5) Cyperaceaepollis neogenicus; GH04-158-1_EF X52-1. 

See facing page for continuation.
BIOSTRATIGRAPHY AND AGE ASSIGNMENT

The Miocene successions of the Gulf of Suez have been examined biostratigraphically in a number of investigations, beginning with the foraminiferal analysis of Moon and Sadek (1923), and followed by a series of foraminiferal, nanoplankton and very few palynological contributions (e.g. Souaya, 1966; Andrawis and Abdel Malik, 1981; El-Heiny and Martini, 1981; Mahmoud, 1993; Ahmed and Pocknall, 1994; El Beialy and Ali, 2002; El Beialy et al., 2005; Youssef, 2011; Soliman et al., 2012). These studies were mainly concentrated on the Rudeis, Kareem and Belayim formations. In contrast, the Nukhul Formation was somewhat neglected except for a few studies (e.g. Hewaidy et al., 2012).

Poor preservation, scarcity of fossils, reworking and other known problems related to cutting samples make our interpretation of microfossils partially inconsistent. However, most of the studied samples yielded moderately preserved palynological assemblages consisting of few records of spores and pollen, dinocysts in fluctuating relative abundances. Microforaminifer linings occurred only very sporadically. Non-marine colonial algae (e.g. Botryococcus and Pediastrum), and fungal palynomorphs were recovered in relatively high percentages (up to 90% of the total palynological content) in some samples. Reworked Cretaceous and Eocene palynomorphs constitute a minor proportion among the recorded yield.

This study deals mainly with the spores and pollen record, as the stratigraphic-relevant dinocyst taxa are mostly fragmentary and badly preserved. As a result we exclude them from our biostratigraphic discussion.

Magnagnostiates howardi (First Appearance Datum, FAD in sample GH04-173-1, Plate 1.3) was described by Germeraad et al. (1968) from the Oligocene M. howardi Pantropical Zone; later on Salard-Cheboldaef (1978, 1979) reported it from the Oligocene of Cameroon. It was used as a marker species from the Late Eocene–Oligocene Zone E (Kaska, 1989) and the Oligocene sediments of Sub-zone 3b (Stead and Awad, 2005) from Sudan. M. howardi was previously reported from the Nukhul Formation of the Gulf of Suez by Ahmed and Pocknall (1994). Younger records from Egypt include the Burdigalian Lagia Shale of Sinai (Wescott et al., 2000).

The investigated samples have also provided a rare record of Margocolporites vanwijhei (FAD in sample GH04-172-2, Plate 1.4), which was described by Germeraad et al. (1968) indicating a Middle Eocene–Pleistocene range in the Caribbean, Late Eocene–Oligocene in Borneo, and Eocene–Oligocene range in West Africa. Takahashi and Jux (1989) reported it from the Eocene–Oligocene of Nigeria. Extra-tropic records include those from the Late Oligocene in New Zealand (Pocknall, 1982) and the Early–Middle Oligocene of South Australia (Truswell et al., 1985). Moreover, Yamamoto (1995) found M. vanwijhei in Brazil in the São Paulo Formation, Resende Formation and in the Itaquaquecetuba Formation, thus suggesting an Oligocene age for these formations. Younger records include the Miocene of Amazonia (Hoorn, 1993, 1994a, 1994b) and the Burdigalian (Lagia Shale) of Sinai and (Rudeis Formation) of the Gulf of Suez, Egypt (Wescott et al., 2000; El Beialy et al., 2005).

Plate 1 (continued): (6) Polypodiaceoisporites pseudosilatus; SA-E6A-23-2_EF Y35-1. (7) Fagraecopollis reticulatus; SA-E6A-20-1_EF M34-1. (8) Praedapollis africanus; GH04-168-1_EF W51-4. (9) Racemonocolpites racematus; GH04-174-1_EF U34-4. (10) Racemonocolpites hians; GH04-174-1_EF K45-3. (11) Florschuetzia cf. trilobata; GH04-163-1_EF W46-1. (12) Retitricolpites striatoides; GH04-164-2_EF U34-3. (13) Subtriporopollenites reticulatus; SA-E6A-29-1_EF F57-4. (14) Margocolporites vanwijhei; SA-E6A-21-1_EF O35-1. (15) Zonocostites ramonae; SA-E6A-23-2_EF M51-1. (16) Foveotricolpites gonzalezii; GH04-174-1_EF J49-2. (17) Tetralcolporites firmus; GH04-159-1_EF G38-1. (18) Foveotricolpites perforatus; GH04-176-1_EF K48-2. (19) Multiareolites formosus; SA-E6A-19-1_EF J41-4. (20) Acaciapollenites myriosporites; GH04-168-1_EF R60-1.
Fagraeapolpis reticulatus (FAD in sample SA-E6A-20-1, Plate 1.7) was first described from the Late Oligocene–Early Miocene of Nigeria (Takahashi and Jux, 1989), which is regarded by Eisawi and Schrank (2008) as a junior synonym of Spirosyncolpites brunii. This species was recorded from the Middle Eocene to the earliest Middle Miocene of Cameroon and Nigeria (Legoux, 1978), and from the Oligocene to Early Miocene of Tunisia (Torricelli and Biffi, 2001).

Praeadapollis africanus (FAD in sample GH04-173-1, Plate 1.8) is reported from the Neogene of Nigeria (Legoux, 1978). It has subsequently been recorded in the Late Oligocene–Early Miocene of Cameroon and Tunisia (Salard-Cheboldaeff, 1978; Torricelli and Biffi, 2001). In Sudan, it was reported from the Lower Paleogene from the Zone E of Kaska (1989).

Foveotricolporites gonzalezi (FAD in sample GH04-174-1, Plate 1.16) and Retitricolporites striatoides (FAD in sample GH04-164-2, Plate 1.12) were described from the Oligocene and Oligocene–Miocene of Cameroon (Salard-Cheboldaeff, 1978, 1979). The former is also reported from the Oligocene–Miocene (Eisawi and Schrank, 2008) from Sudan, while the latter appears closely similar to Rhoipites baculatus, which was recovered from the same interval in Argentina by Barreda (1997).

Racemonocolpites racematus (FAD in sample GH04-174-1, Plate 1.9) was reported from the Oligocene of Cameroon (Salard-Cheboldaeff, 1979). Racemonocolpites hians (FAD firstly reported from sample GH04-174-1, Plate 1.10) occurs mainly in the Oligocene–Miocene (Legoux, 1978) and from the Oligocene–Early Miocene within Zone VI of Eisawi and Schrank (2008).

Polypodiaceoisporites pseudopsilatus (Firstly reported from sample SA-E6A-23-2, Plate 1.6) shows its first appearance in the Chattian and near the OMB of northern South America (Ochoa et al., 2012).

Cyperaceaepollis neogenicus (FAD in sample GH04-173-1, Plate 1.5) was considered as a good marker for the Early Miocene in Argentina (Barreda and Palamarczuk, 2000, Guerstein et al., 2004).

Acaciapollenites myriosporites (FAD in sample GH04-168-1, Plate 1.20) was recorded from the Late Miocene and Late Oligocene–Early Miocene in the Angola Basin by Partridge (1978), Tunisia (as Polyadopollenites sp.) by Torricelli and Biffi (2001), and Patagonia by Barreda and Palazzesi (2007). In Australia, it witnessed its earliest occurrence from the Early Miocene (Stover and Partridge, 1973), then from the Oligocene (Truswell et al., 1985). In Egypt, the basal occurrence of Acaciapollenites was attributed to the Early Eocene of the Western Desert (Guinet et al., 1987), younger records include that from the Pliocene–Pleistocene of the Nile Delta (Saad et al., 1987) and the Burdigalian of the Gulf of Suez (El Beialy et al., 2005).

In summary, our palynological record reveals no strong evidence for an accurate dating, due to scarcity of palynomorphs. Based on local and regional correlations with similar assemblages from

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the above-mentioned published records and other additional data from Palynodata and White (2008; Figure 3), the youngest age for the lower Shoab Ali Member could be Late Oligocene and the upper Ghara Member as Early Miocene. In addition, Shahin (1998) assumed that the OMB coincides with Tayiba/Nukhul contact in Sinai. When applying the correlation that the Tayiba (or Abu Zenima) Formation is a synonym to the Shoab Ali Member (Al-Husseini, 2012), such a conclusion supports drawing the OMB at the contact between Shoab Ali and Ghara Member. At this position, a clear change from a terrestrial/fluvial system into a lacustrine or shallow-marine setting may serve as a lithologic indicator for the OMB.

In addition, the presence of a well-developed fungal/algal proliferation at the boundary between the two members (El Atfy et al., in press) is in agreement with the conclusion of Hewaidy et al. (2012) who established a Late Oligocene–Early Miocene age for the Nukhul Formation and the OMB to be located within the Nukhul Formation.

**PALYNOFACIES ANALYSIS**

The concept of palynofacies was introduced by Combaz (1964) embracing all acid-resistant organic matter in sediments. It was subsequently studied by many authors who demonstrated its usefulness as a tool for environmental and depositional interpretations (e.g. Batten, 1973, 1981, 1982; Fisher, 1980; van der Zwan, 1990; Davies et al., 1991). These studies have resulted in a wide range of terminologies of palynodebris classifications (e.g. Bujak et al., 1977; Habib, 1979; Venkatachala, 1981; Masran and Pocock, 1981; Parry et al., 1981; Boulter and Riddick, 1986; Lorente, 1990; Powell et al., 1990; Batten, 1996).

The use of palynofacies analysis in both commercial (i.e. hydrocarbon exploration) and academic contexts has been steadily increasing during the last decades (Roncaglia and Kuijpers, 2006). It addresses the ever increasing need for a better understanding of biologic, paleoenvironmental, depositional, and maturation changes within a definite type of sediments, especially when it has a petroleum potential and other types of microfossils are lacking, as is the case in the Nukhul Formation.

The main aims of studying palynofacies here are to: (1) aid in the characterization and correlation of the recorded palynodebris; (2) establish the kerogen type and hence the type of expected expelled hydrocarbon; and (3) provide more information about the prevailing environmental conditions during the deposition of the Nukhul Formation. Despite the low palynological yield, which may be related to the prevalence of sandy facies of the studied samples within the Nukhul Formation.
Formation of the GH 404-2A and SA-E6A wells, the presence or absence of different organic components can be directly related to lithofacies pattern (Batten, 1981). Based on the quantitative and qualitative variations in the kerogen content and the distribution of amorphous organic matter (AOM) versus phytoclasts and palynomorphs, three associations in two main types of palynofacies (PF-Ia, PF-Ib, and PF-II), linked to depositional and paleoenvironmental variations are recognized and described as follows:

**Palynofacies Type PF-Ia (AOM-Phytoclasts-Palynomorphs)**

The PF-Ia is the dominant palynofacies type in most of the Ghara Member samples (GH 04-155 to 171, except for samples 159–161, 165, 167 and 169) within the interval 10,860–11,340 ft in the GH 404-2A Well and within the interval 4,900–5,310 ft as well as sample SA-E6A-31, except for sample SA-E6A-25 in the SA-E6A Well of the studied formation. This palynofacies type is dominated by AOM (normally more than 50% of the bulk organic particles), followed by phytoclasts with percentage usually no higher than 40% and thirdly by palynomorphs which show percentages from 0–10% from the bulk organic particles.

Plate 2: (1) A mixture of granular yellow (green arrows) and yellowish-amber to brown masses (red arrow) of AOM, mineral inclusions especially of pyrite are very clear (black arrows); SA-E6A-16-1_EF M34-1. (2) A fluorescence view over the same AOM mixture, showing a very weak fluorescence potential and no difference witnessed between the two different colored AOM assemblages; SA-E6A-16-1_EF M34-1. (3) An isolated yellowish brown AOM clot with highly mineral (pyritic) inclusion (white arrow); SA-E6A-19-1_EF J56-1. (4) A similar view under the fluorescence light showing a weak fluorescence potential of the AOM mass; SA-E6A-19-1_EF J56-1.

See facing page for continuation.
AOM consist mainly of yellowish amber to brown heterogeneous, more-or-less fluorescent, coagulations having a fluffy appearance, irregularly-shaped masses with no cellular details and often uniformly granular with inclusions (Plates 2.1 to 2.8). AOM is the dominant organic component of sediments that accumulated under dysoxic to anoxic conditions, and increases mainly with more nutrient supply and decreasing oxygen in water (Tyson, 1995). AOM are gel-like and exhibit a “clotted” appearance (Tyson, 1984), and are believed to be alteration products rather than primary material (Staplin, 1969; Rogers, 1979). Also, Rogers (1979) demonstrated that bacterial and thermal degradation can produce amorphous material from other palynodebris.

According to Pieńkowski and Waksmundzka (2009), AOM can be subdivided into AOMT (AOM of terrestrial derivation, dark and usually opaque, occurring most often in rounded fragments) and AOMA (AOM of aquatic origin, mostly light and translucent). AOMT embraces heterogeneous, fluorescent amorphous organic matter, humic gel and resin (Tyson, 1995). AOMA is usually of planktonic/bacterial origin (Batten, 1996) and it shows “spongy”, translucent structure and diffusive outlines. AOMA dominate in the marine deposits (Tyson, 1995), however, it is not determined if it is of plant or animal origin.

Plate 2 (continued): (5) Mixed assemblage of yellow algal debris with a notable occurrence of Botryococcus colonies (blue arrows), biodegraded phytoclasts (black arrows) including some interinite fragments or opaque phytoclasts (brown arrow), and some fluffy yellowish brown AOM (red arrow); SA-E6A-20-1_EF N59-1. (6) A variable fluorescence potential among the component of the same palynofacies with a very strong fluorescence for algal debris and Botryococcus colonies (arrows), weak fluorescence for the AOM as noted before; SA-E6A-20-1_EF N59-1. (7) Yellow (green arrows) and yellowish brown (red arrow) AOM coagulations as seen under transmitted light, with some mineral inclusions; SA-E6A-37-1_EF J42-3. (8) A clear difference in fluorescence potential between the two types of AOM supporting a mixing between algal (white arrow) and fungal or bacterial (pink arrow) origin; SA-E6A-37-1_EF J42-3.
Since the investigated sediments are mostly immature to marginally mature (Figure 5c), in addition to the presence of fungi and some cyanobacteria-like organisms, this palynofacies type is most likely due to bacterial and possibly fungal degradation. The yellowish-amber color suggests derivation from terrestrial material as opposed to the gray-colored marine variety (Masran and Pocock, 1981; Venkatachala, 1981). Recently, Boussafr et al. (2012) described a similar AOM facies from modern lacustrine sediments, material that has a close appearance and morphologic character to our AOM (Plates 2.3 and 2.4).

Routine fluorescence microscopic examinations indicate that the recorded AOM masses show different degrees of weak fluorescence (Plates 2.2, 2.4 and 2.6) supporting a bacterial or fungal origin. However, some AOM show a moderate fluorescence (Plate 2.8), the evidence that could support an algal origin. As a result we can assume a lacustrine or lake origin of deposition rather than a pure AOMT type.

**Palynofacies Type PF-Ib (AOM-Funginite/Alginite-Phytoclasts)**

This palynofacies type is dominated by AOM, algal masses forming about ≥ 50%, and fungal palynomorphs (Funginite; 30–40%), with a very minor percentage of biodegraded structured phytoclasts (ca. 10%, Plates 3.1 to 3.8, and 4.1 to 4.4). PF-Ib is the dominant facies type in the

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**Plate 3:** (1) Mixed assemblage of fungal spores (pink arrows), biodegraded phytoclasts (black arrows) with minor AOM particles; GH04-174-1_EF P30-2. (2) A similar assemblage under fluorescence; GH04-174-1_EF P30-2B. (3) A typical PF-Ib assemblage with a multicellate fungal spore (pink arrow), algal mass (blue arrow), biodegraded phytoclasts (black arrows), fluffy AOM particles (green arrow) and some mineral inclusions (white arrow); GH04-174-1_EF G36-2. (4) A strong fluorescent algal mass in comparison with other phytoclasts of the same assemblage; GH04-174-1_EF G36-2B.

*See facing page for continuation.*
investigated samples GH04-172 to 174 (11,370–11,430 ft) from the lower part of the Ghara Member and at the boundary between Ghara and Shoab Ali members in the GH 404-2A Well. Accordingly, it is present in the SA-E6A Well within the interval 5,330-5,370 ft, and the sample SA-E6A-25.

Fungal palynomorphs are represented mainly by fungal spores with both aquatic and terrestrial affinities. Furthermore, fungal hyphae (Plates 3.7 and 3.8) and fructifications showing higher diversity in comparison with other palynomorphs reported from the same interval within the Nukhul Formation. In addition, palynodebris attributed to algae (e.g. Plates 3.3 and 3.4) or algal masses (?Chlorophyceae), together with other palynomorphs of unknown origin that may be considered to be cyanobacteria, constitute a major component in PF-Ib beside fungi.

Some of the recorded palyno-particles, which were thought to be of algal origin reveal a very weak fluorescence, meanwhile other particles show strong fluorescence results. Such an observation could be interpreted as mixing of algal, fungal and/or bacterial proliferation at the OMB as a result of a regional regressive phase covering the studied area at the time of deposition (El Atfy et al., in press). This environmental variation could explain the absence of the zonal marker Cicatricosisporites dorogensis, which is considered an alluvial plain element (as not the case here) according to Germeraad and de Haan (unpublished report, 1964).

Plate 3 (continued): (5) Fungal spores (pink arrows), biodegraded phytoclasts (black arrows), angiosperm pollen (gray arrow) and AOM masses (green arrow); GH04-174-1_EF G48-0. (6) A similar assemblage under fluorescence; GH04-174-1_EF G48-0. (7) Melanized fungal hyphae (pink arrow), biodegraded phytoclasts (black arrow), and minor AOM clots (green arrow); SA-E6A-27-1_EF M32-1. (8) Non-fluorescent melanized fungal hyphae (pink arrow), biodegraded phytoclasts, and minor AOM clots; SA-E6A-27-1_EF M32-1.
Plate 4: (1) Mixed assemblage of *Pediastrum* colony (blue arrow), biodegraded intertinite phytoclasts (black arrow) and minor AOM clots (green arrows); SA-E6A-25-1_EF R37-4. (2) A fluorescence image showing a strong fluorescence for *Pediastrum* colony, and some fluorescence for AOM clots; SA-E6A-25-1_EF R37-4. (3) Mixed assemblage of AOM masses (green arrows) and *Botryococcus* colony (blue arrow); SA-E6A-20-1_EF Y43-3. (4) A fluorescence image showing a strong fluorescence for *Botryococcus* colony and with non-fluorescent AOM masses; SA-E6A-20-1_EF Y43-3. 

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**Palynofacies Type PF-II (Phytoclasts-AOM-Palynomorphs)**

The bulk of this facies is made up of phytoclasts (≥ 60%) followed by AOM (5–30%) with a minor percentage of palynomorphs (10–15%). This palynofacies (Plates 5.1 and 5.2) prevailed during the deposition of the lower part of the Shoab Ali Member covering the interval from 11,460–11,760 ft, except for samples GH 04-175, 179 and 180, 159 to 161, 165, 167 and 169 of the GH 404-2A Well, and within the interval 5,450–5,700 ft of the SA-E6A Well.

Phytoclasts herein refer to moderately preserved structured terrestrial plant fragments mainly composed of cuticles, tracheid, xylem ray tissues and opaque, often biodegraded phytoclasts. The relative amount of phytoclast provides information about distance from the shore and terrestrial influx in the sediments (Tyson, 1995). Cuticles (Plates 5.1 to 5.6) are the most abundant components and are thin, platy epidermal fragments from non-woody plant organs such as leaves and roots (Rich, 1989; Tyson, 1995). Also, cuticles are most common in lagoon and deltaic (marsh) deposits, and are especially characteristic of facies resulting from the settling out of flotation and suspension loads under low energy conditions (Fisher, 1980).

Cuticles are usually well-preserved and show a clear structure of epidermal cell outlines (e.g. Plates 5.3 and 5.4) and are known to have a high fossilization potential (cf. Thomas and Spicer, 1986; Kerp,
Plate 4 (continued): (5) Botryococcus colony; showing a characteristic lustrous yellow color; SA-E6A-37-1_EF T48-3. (6) A highly fluorescent Botryococcus colony; SA-E6A-37-1_EF T48-3. (7) Ovoidites grandis; SA-E6A-17-2_EF Z39-1. (8) A highly fluorescent Ovoidites grandis; SA-E6A-17-2_EF Z39-1. (9) A highly fluorescent Pediastrum colony; SA-E6A-34-1_EF S36-4.

1990). Their buoyant nature ensures easy dispersal by water and they can be deposited either in low-energy conditions by settling out of suspension (Parry et al., 1981) or pushed under heavier particles of sand in high energy conditions (Boulter and Riddick, 1986).

Wood (Plates 5.9 and 5.10) is especially abundant in alluvial plains and in deltaic environments. Coarse phytoclast material (> 1 mm) is usually dominant in high, first- and second-order, headwater streams (Minshall et al., 1985). Fungal hyphae and other plant fragments including tracheid (Plates 5.7 and 5.8) and resins are present in small abundances.

Routine fluorescence microscopic examinations show that most of the recorded phytoclasts show no or only weak fluorescence except for some particles as illustrated in Plate 5.

**Interpretation of Palynofacies Data**

In our work, we use ternary diagrams for illustrating our data due to their good potential in showing a clear separation among various groups of samples or assemblages (Tyson, 1995). The AOM-phytoclast-palynomorph (APP) of Tyson (1993) graphic presentation is applied in our study.

Data plotted on the APP ternary diagram enabled us to separate the respective samples into clusters, which refer to two main palynofacies types (Figure 4a). PF-la samples lie mainly within the fields IX, VIII and VII which reflect distal dysoxic-anoxic shelf setting, based on their high content of AOM that dilutes all other organic particles and hence classified as type II to II >> I kerogen of Tyson (1995; oil prone).
Plate 5: (1) A mixed assemblage of phytoclasts (cuticle, black arrow) and brown fluffy AOM (green arrow); SA-E6A-31-1_EF W34-4. (2) A similar assemblage under fluorescence showing potentially fluorescent cuticle (arrow) and non-fluorescent AOM (arrow); SA-E6A-31-1_EF W34-4. (3) Dispersed leaf cuticle phytoclast showing irregular, hexagonal cellular structure, (probably of gymnospermous origin); SA-E6A-38-2_EF X49-4. (4) A highly fluorescent cuticle; SA-E6A-38-2_EF X49-4.

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The APP ternary plot of Tyson (1993) hosts PF-Ib facies samples on the field VII representing a distal dysoxic-anoxic shelf which yields moderate to common microplankton and can be classified as type II oil-prone kerogen. However, this interpretation partially contradicts the microscopic results of PF-Ia and PF-Ib samples that show a sparse occurrence or complete absence of marine microplankton and is mainly dominated by funginite or fungal palynomorphs. Such assemblages are well known to be of terrestrial origin and their abundance is generally higher in lagoon facies (Oboh, 1992) or lacustrine AOM (Boussafir et al., 2012) as shown in details in the paleoenvironment part. Therefore, we prefer to supplement our results by additional plotting on the revised model of Tyson (1993) which was established by Roncaglia and Kuijpers (2006) as AOM-PS-FDAO (APF) as shown in Figure 4b. PF-Ia and PF-Ib samples lie within the category that suggests a suboxic-anoxic environment.

PF-II samples lie in the fields II, IVa, III and V with a clear transition from marginal dysoxic-anoxic basin into mud-dominated oxic shelf, classified as type III or II kerogen (mainly gas prone, could be related to the high phytoclast input) as deduced mainly from the APP plot of Tyson (1993). By plotting on the APF diagram of Roncaglia and Kuijpers (2006; Figure 4b), those samples show proximal suboxic-anoxic shelf with high AOM content and refer to suboxic-anoxic conditions.
Plate 5 (continued): (5) Dispersed leaf cuticle phytoclast showing a clear cellular structure; SA-E6A-33-1_EF J36-2. (6) A highly fluorescent cuticle; SA-E6A-33-1_EF J36-2. (7) Bio-structured tracheid phytoclast shows a clear scalariform (ladder-like) structure; SA-E6A-37-1_EF F39-4. (8) Non-fluorescent tracheid; SA-E6A-37-1_EF F39-4. (9) A brown probably uncharred woody phytoclast with clear striate structure; SA-E6A-38-2_EF Q38-4. (10) Non-fluorescent wood; SA-E6A-38-2_EF Q38-4. (11) Bio-structured phytoclast with no clear internal structure as a result of jellification obliteration; SA-E6A-31-1_EF J47-2. (12) A fluorescent phytoclast; SA-E6A-31-1_EF J47-2.

GEOCHEMICAL INTERPRETATIONS

Palynofacies data are supplemented by organic geochemical results adopted from GUPCO (unpublished report, 1984). Herein, Rock-Eval pyrolysis was undertaken on 18 cutting samples from the Nukhul Formation (GH 404-2A Well) using Rock-Eval 6 analyzer to determine their TOC, hydrogen index (HI) and other parameters to assess kerogen type, organic richness, the level of thermal maturation and petroleum generative potential. All geochemical results used for our compilation are listed in Table 2. In addition, vitrinite reflectance ($R_o$) was measured for three samples by the Exploration Logging International Inc. EXLOG (Table 2).

TOC and Rock-Eval Pyrolysis

Geochemical results of 18 samples from the depth interval 10,850–11,750 ft penetrating the Nukhul Formation within the GH 404-2A Well were obtained. Pyrolysis gives rise to two parameters. The first, $S_1$ corresponds to the organic matter that can be vaporized, and the second, $S_2$ refers to the pyrolysis products formed from the kerogen during Rock-Eval analysis.

TOC is expressed as the relative dry weight percentage of organic carbon in the sediment (Jarvie, 1991), but is not a direct measure of the total amount of organic matter in a sediment. For a rock to be a source of hydrocarbons it has to contain sufficient organic matter for significant generation and expulsion (Batten, 1996), for many years this was taken as 0.5% TOC for shales and somewhat less (0.3%) for carbonates. Whereas, Peters and Cassa (1994) noted that the minimum organic content of a source rock needs to be within the range of 1–2%.
The interpretation of Rock-Eval pyrolysis in particular covers several aspects of petroleum geochemistry, and undertaken using industry-standard cross-plots (Cornford, 2005). One of the simplest and the more common, plots the HI (mg of hydrocarbons evolved during kerogen breakdown, divided by wt% TOC, x 100 = S2/TOC × 100 mg HC/g TOC) against the oxygen index OI (mg of CO2 derived from thermal alteration of oxygenated organic compounds = S3/TOC × 100 mg CO2/g TOC), another one is plotting the HI versus Tmax (the temperature at which the maximum generation of the products of pyrolysis occurs), as illustrated in Figures 5a to 5c.

Raw data plotting on the so-called pseudo-van Krevelen diagram shows that all samples are mainly of kerogen III type, and hence has the ability for gas prone production (Figure 5a). Plotting the readings of HI versus Tmax data on the modified diagram of Espitalié et al. (1985) shows that samples of the Nukhul Formation fall mainly in the field of mature kerogen with poor to fair petroleum potential as deduced from plotting the S2 versus TOC (Figure 5b).

TOC values are generally low, ranging between 0.42% and 0.78% (Table 3). TOC normally correlates well with the relative abundance of AOM in the sediment (Tyson, 1995), meaning that high TOC values usually indicate the presence of well-preserved AOM (cf. Hermann et al., 2011). Therefore, the low TOC values of the AOM-rich intervals (Ghara Member) are unexpected, but may be attributed to the sandy lithofacies. The generating source potential S1 and S2 values range from 0.30 to 0.70 mg/g and 0.91 to 1.97 mg/g, revealing poor to fair petroleum potential.

HI readings range from 165 to 302 mg HC/g TOC, indicating a type III kerogen and hence it can produce gas or non-hydrocarbon as interpreted from Peters and Cassa (1994).

Tmax results range between 435°C and 441°C which display a thermal maturation on the upper limit of the oil window (Tmax of top oil window 435–445°C after Peters, 1986), only one of measured samples is slightly below the oil window with a value of 433°C.

Production index (PI) ranges from 0.20 to 0.30, displaying an immature setting. PI gradually increases with depth for fine-grained rocks. Reservoir rocks show anomalously high PI values.
Table 2
Geochemical results in GH 404-2A Well (based on GUPCO, unpublished report, 1984)

| Sample | Depth (feet) | TOC (wt. %) | Petroleum Potential (Quantity) | Rock-Eval Pyrolysis | Kerogen Type (Quality) | Thermal Maturation |
|--------|-------------|-------------|--------------------------------|----------------------|------------------------|-------------------|
|        |             |             |                                | S1 (mg/g) | S2 (mg/g) | HI (mg HC/g TOC) | S1/S2 | R0(%) | Tmax(ºC) | PI [S1/(S1+S2)] |
| 1      | 10,850      | 0.72        | 0.49                           | 1.36       | 189       | 1.89            | 0.58   | 439   | 0.26     |
| 2      | 10,900      | 0.60        | 0.50                           | 1.27       | 212       | 1.72            | 0.58   | 436   | 0.28     |
| 3      | 10,950      | 0.58        | 0.44                           | 1.16       | 200       | 1.53            | 0.58   | 433   | 0.28     |
| 4      | 11,000      | 0.66        | 0.52                           | 1.67       | 253       | 1.45            | 0.58   | 435   | 0.24     |
| 5      | 11,050      | 0.45        | 0.42                           | 0.99       | 220       | 0.97            | 0.58   | 435   | 0.30     |
| 6      | 11,150      | 0.42        | 0.51                           | 1.27       | 302       | 0.88            | 0.58   | 436   | 0.29     |
| 7      | 11,200      | 0.66        | 0.39                           | 1.09       | 165       | 0.76            | 0.58   | 436   | 0.26     |
| 8      | 11,250      | 0.72        | 0.49                           | 1.57       | 218       | 1.22            | 0.55   | 436   | 0.24     |
| 9      | 11,300      | 0.74        | 0.59                           | 1.66       | 224       | 1.39            | 0.58   | 439   | 0.26     |
| 10     | 11,350      | 0.78        | 0.70                           | 1.97       | 253       | 1.60            | 0.58   | 436   | 0.26     |
| 11     | 11,400      | 0.63        | 0.41                           | 1.22       | 194       | 0.95            | 0.58   | 439   | 0.25     |
| 12     | 11,450      | 0.67        | 0.48                           | 1.58       | 236       | 1.16            | 0.58   | 440   | 0.23     |
| 13     | 11,500      | 0.55        | 0.36                           | 1.41       | 256       | 2.27            | 0.58   | 437   | 0.20     |
| 14     | 11,550      | 0.56        | 0.39                           | 1.17       | 209       | 2.39            | 0.62   | 438   | 0.25     |
| 15     | 11,600      | 0.51        | 0.34                           | 1.14       | 224       | 2.24            | 0.58   | 439   | 0.23     |
| 16     | 11,650      | 0.45        | 0.30                           | 0.91       | 202       | 1.60            | 0.58   | 441   | 0.25     |
| 17     | 11,700      | 0.46        | 0.33                           | 0.93       | 202       | 1.72            | 0.58   | 435   | 0.26     |
| 18     | 11,750      | 0.44        | 0.30                           | 0.99       | 225       | 2.02            | 0.58   | 438   | 0.23     |

Compared to adjacent fine-grained rocks, also for $T_{\text{max}}$ values of $< 435^\circ\text{C}$ and $T_{\text{max}}$ ranges between $435^\circ\text{C}$ and $445^\circ\text{C}$, PI values exceeding 0.2 and 0.3, respectively, are considered anomalous (Peters and Cassa, 1994). This is the case for the Nukhul Formation (mainly Nukhul Clastics/Shoab Ali Member) in the Hilal Field, which was formerly established as a reservoir (e.g. EGPC, 1996).

Data given in Table 3 summarize the petroleum potential (quantity), kerogen type (quality) and thermal maturation gradient, or extent of burial heating, in correlation with the interpretation of Peters and Cassa (1994). From the above discussion and based on TOC and Rock-Eval results, one can conclude that the Nukhul Formation is thermally immature and has a poor petroleum potential. Also, it is generally of type III kerogen, and hence is capable of expelling gas, with good reservoir characteristics.

**Vitrinite Reflectance (Ro)**

Vitrinite reflectance ($R_o$) has been the major calibration parameter for modeling the thermal history of sedimentary rocks. Urban and Allen (1975) established four phases of thermal maturation that can be discussed as follows:

- Immature: Early genesis dry gas and minor amounts of liquid hydrocarbon ($R_o$: 0.00–0.50%).
- Mature: Main phase of liquid petroleum generation ($R_o$: 0.50–1.30%).
- Dry gas: Organic metamorphism ($R_o$: 1.30–2.00%).
- Dry gas or barren ($R_o$: > 2.00%).
Figure 5: Kerogen plots of the GH 404-2A Well. (a) HI versus OI, (b) S₂ versus TOC, and (c) HI versus T_max.

Rₜ was tested for three representative samples using 4–20 vitrinite particles (GUPCO, unpublished report, 1984). Rₜ measurements reveal that their values are 0.58, 0.55, 0.62% indicating thermally immature to early phase of mature source rocks that lie within the main phase of liquid petroleum generation. In contradiction with palynofacies and other results, the matter could be attributed to the reservoir potential of the Nukhul Formation.
Table 3
Geochemical results (based on GUPCO, unpublished report, 1984) describing the petroleum potential, kerogen type and thermal maturation of the Nukhul Formation, GH 404-2A Well in comparison with the interpretation of Peters and Cassa (1994)

| Features                  | Results (GUPCO, 1984) | Interpretation (according to Peters and Cassa, 1994) |
|---------------------------|------------------------|------------------------------------------------------|
| Petroleum Potential       |                        |                                                      |
| (Quantity)               | TOC (wt. %)            |                                                      |
|                          | S1(mg/g)               |                                                      |
|                          | S2(mg/g)               |                                                      |
| 0.42–0.78                | 0.30–0.70              | Poor to fair petroleum potential                     |
| Kerogen Type             | HI (mg HC/g TOC)       |                                                      |
| (Quality)                | S1/(S1+S2)             |                                                      |
| 165–302                  | 0.76–2.39              | Type III/IV kerogen - gas or none expelled product   |
| Thermal Maturation       | Ro (%)                 |                                                      |
|                          | Tmax (°C)              |                                                      |
|                          | PI [S1/(S1+S2)]        |                                                      |
| 0.55–0.62                | 433–441                | Immature to early/peak mature sediments              |

PALEOENVIRONMENTAL INTERPRETATIONS

The Nukhul Formation was deposited in a variety of depositional settings, reflecting environmental heterogeneity due to differences in tectonic setting of the separated fault blocks (Alsharhan and Salah, 1997). Moreover, Carr et al. (2003) postulated two contemporaneous scenarios for the deposition of the Nukhul Formation: (1) an open-shelf (offshore) to shore-face environment, and (2) a structurally controlled estuary. Saoudi and Khalil (1984) previously assumed for the Shoab Ali Member and the Ghara Member a deposition in continental and restricted shallow-marine settings within the Gulf of Suez, respectively.

Deposits of the Nukhul Formation record the earliest phase of sedimentation in a rift basin environment. Its lower strata document a continental sedimentation in the young incipient rift while the upper portion chronicles the initial marine transgression into the basin (Richardson and Arthur, 1988). The sediments are of mixed clastic, carbonate and evaporitic composition containing brackish-water ostracods and oysters, coral patch reefs, shallow-marine foraminifers such as Miogypsinia, Elphidium crispum and Amphistegina sp., as well as many species of Pecten. In addition, reworked Cretaceous and Paleogene foraminifers are also locally abundant (Scott and Govean, 1985).

Fungal/Algal Proliferation

Considerable fungal and algal proliferation phases, with fungal remains reaching up to 50% of total palynomorphs, were recorded in the GH 404-2A Well (El Atfy et al., in press). The main phase of this proliferation lies within the interval from 11,370 to 11,430 ft, and may be considered to be a potential ecologic representative for the OMB within the area of study. This proliferation is likely situated near or at the OMB, which coincides with a lithologic transition from sandy facies of the Shoab Ali Member to evaporitic and partially calcareous, shallow-marine deposits of the Ghara Member.

Generally, fungi are known to respond rapidly to environmental stress and disturbance (Pugh and Boddy, 1988). Fossil records indicating excessive fungal activity may therefore provide confirming evidence of past ecosystem destabilization (Visscher et al., 1996). Moreover, a rich diversity of fungal remains may be interpreted as an indicator of prevalence of a humid climate in the area.
of deposition (Singh and Chauhan, 2008). Furthermore, a sufficiently distinct record of freshwater algae (mainly *Pediastrum* and *Botryococcus*) was reported in our samples of the Nukhul Formation, in coincidence with the fungal proliferation main phase between the upper part of the Shoab Ali and lower part of the Ghara members.

*Botryococcus* colonies are preserved in sediments from the Late Precambrian to Recent (Combaz, 1980; Guy-Olsson, 1998). They are mostly found in freshwater fens and lakes like the cosmopolitan *B. braunii* (Batten and Grenfell, 1996) but can also inhabit temporary ponds, pools, ditches, bogs and wet mud (West and Fritsch, 1927; Round, 1965; Davis et al., 1977; Graham and Wilcox, 2000). Moreover, De Deckker (1988) reported *B. braunii* in southeastern Australia from water bodies with salinities up to 20‰. Dulhunty (1944) noted that *Botryococcus* preferred quiet water bodies with relatively low sediment input and without sub-aerial plants but in close enough proximity to shore for rotting vegetation to be a supply of nitrogen for growth. Davis (1999) identified a rise in *Botryococcus* coincident with a decline in littoral vegetation as indicative of open water and high lake levels. Singh et al. (1981) considered *Botryococcus* as indicative of permanent open water but noted that *Botryococcus* peaks occurred at variable lake depths, deduced from coexisting pollen evidence. Accordingly, following early observations by Blackburn and Temperley (1936), the highest *Botryococcus* occurrences might be equally well explained by greater lake depths resulting in less aquatic vegetation, particularly in the littoral zone, and also adaptable to brackish water.

Recently, Cook et al. (2011) found that high frequencies of *Botryococcus*, together with *Pediastrum*, *Cosmarium*, *Hydrocharitaceae*, *Myriophyllum* sp. and *Myriophyllum muelleri* indicate that Lake Turangmoroke in Australia was of variable depth, clear and received regular freshwater input. In addition, living colonies of *Botryococcus* prefer elevated alkaline concentrations (Nichols, 1973), and the algae can be flushed out of the freshwater system in large quantities (Bayly, 1989; Traverse, 1990), and it is not surprising that accumulations of dead colonies are reported from alkaline bodies of water such as inland saline lakes or warm climate brackish water lagoons. In moderate abundances or in association with other terrestrial and freshwater indicators, as in our case, *Botryococcus* is a compelling indicator of freshwater paleoenvironments (Zippi, 1998; Larsson et al., 2010).

To summarize, sporadic occurrence of *Botryococcus* in the absence (or with sporadic occurrence) of other coccal green algae probably indicates specific conditions, likely relatively extreme environments (very cold, clear, oligotrophic, eventually dystrophic), which prevents the occurrence of e.g. *Pediastrum*, as it is the case in some samples from the Ghara Member. *Botryococcus* spp. probably could better tolerate even more extreme water biotopes, unfavorable for *Pediastrum* and most other coccal green algae. Those findings may correspond to water bodies with cool, clean, oligo- to dystrophic waters, which are particularly characteristic for cool periods and high altitude (Jankovská and Komárek, 2000).

Representatives of *Pediastrum* extend back to the Early Cretaceous (Evitt, 1969), they occur as frequent components of freshwater phytoplankton communities. Its presence in sediments has been largely recognized as one of the most reliable paleoecologic markers of freshwater environments, in particular, of standing water bodies such as ponds and lakes (Tell and Zamaloa, 2004; Sarmaja-Korjonen et al., 2006). The most common and widely distributed species *P. boryanum* occurs in a variety of waters (Jankovská and Komárek, 2000; Komárek and Jankovská, 2001) and was considered as characteristic of the littoral zone of oligotrophic waters (Woolfenden, 1993; Cook et al., 2011).

In Australia, Singh et al. (1981) found that elevated levels of *Pediastrum* coincided with deep water (at least 7 m), in addition, higher frequencies of *Pediastrum* were associated with rising lake levels in the early Holocene (Dodson, 1974). Furthermore, since *Pediastrum* species are easily transported palynomorphs, they are often found in sediments deposited on or beyond the marine continental shelf as part of allochthonous associations brought by fluvial systems from inland areas (Brenac and Richards, 1996). In fossil lake sediments, mass developments of algae reflect autochthonous vegetation in the water body, while most pollen grains are derived from land habitats (Jankovská and Komárek, 2000).
The alga *Ovoidites* (Zygnemataceae) has a strong association with freshwater fluvial-paludal (i.e. marshy) deposits. In particular, in-situ *Ovoidites* can be used to infer freshwater paleoenvironments where the water depth is shallow lacustrine, paludal or low gradient fluvial. Similar to any terrestrial or freshwater fossil, *Ovoidites* may be transported to a near-shore marine environment in substantial number (cf. Zippi, 1998). In our material the recovery of *Ovoidites* (Plates 4.7 and 4.8) is not as high as of *Botryococcus* and *Pediastrum*.

A global eustatic sea-level fall in the latest Chattian to early Aquitanian, indicated by Zachos et al. (2001) could have an influence on the depositional setting of our studied succession and may have resulted in shallow water at the onset of the deposition of the Ghara Member of the Nukhul Formation. The low diversity of dinocysts, accompanied by freshwater algae and fungi has been previously identified at the end of the Chattian, close to the OMB in Turkey (Sancay et al., 2006) and could support our interpretation.

Moreover, Zevenboom (1996) reported two major cooling phases and relative sea-level falls at the Lemme-Carrosio section in Italy (Global Stratotype Section and Point – GSSP) for the base of Neogene including the Oligocene–Miocene transition. These events have been calibrated with third-order sea-level cycles of Haq et al. (1988), and hence can be interpreted as global events. Seemingly, this coincides with the relative sea-level fall indicated at the probable OMB in the present study.

In the Nile Delta, El Beialy (1988, 1990) reported a high percentage of *Pediastrum* from the Sidi Salem Formation, close to the transition from the Middle–Late Oligocene to the Early Miocene, which was named by El Beialy (1988) as “Regression I”. A notable decrease within the marine palynomorphs and a major decrease in diversity were witnessed among the overall reported species. This regressive phase was also been recognized lithologically (Zaghoul et al., 1977; Kora, 1980). Clear lithologic changes in our successions strongly support a major regression around the OMB as a result of the Mi-1 glaciation event of Miller et al. (1991), and enable us to put the OMB at the border between the Ghara and Shoab Ali members within the Nukhul Formation.

Palynologically, the presence of some environmentally restricted marker species in our material (mainly within the Ghara Member) is potentially a strong proxy for paleoenvironmental interpretations especially if these taxa have recent analogs, as follows: *Magnastriatites howardi* (Parkeriaceae, Ceratopetris, a freshwater fern: Germeraad et al., 1968) which is a fern of open freshwater habitats and its presence always suggest some freshwater influence and the presence of some more open vegetation. Its intermittent and infrequent presence together with such a kind of litho- and palynofacies can suggest an aquatic setting like coastal swamps or floodplains (Germeraad et al., 1968; Oboh et al., 1992; Eisawi and Schrank, 2008).

*Verrucatosporites usensis* (Blechnaceae, Stenochlaena palustris: Germeraad et al., 1968; Polypodiaceae: Rull (1997), was considered by Germeraad et al. (1968) as a climbing fern and its presence indicate a swamp forest setting (cf. Oboh et al., 1992).

*Crassoretitiretes vanraadshoovenii* (Lygodium microphyllum: Germeraad et al., 1968) was considered as a climbing fern in humid marsh and swamp forests of West Africa (Germeraad et al., 1968).

*Margocolporites vanwijhei* (Caesalpinia type: Germeraad et al., 1968; Fabaceae: Jaramillo and Dilcher, 2001) prefers a coastal habitat, and is also related to *Caesalpinia* sp., a broad-leaved and evergreen temperate tree belonging to the Leguminosae (Utescher and Mosbrugger, 2007).

*Zonocostites ramonae* (Rizophoraceae, Rhizophora type: Germeraad et al., 1968), is a distinctive pollen type found in extant genera of mangroves notably *Rhizophora* (Germeraad et al., 1968), however it is rare in our material but its record in SA-E6A Well with other *Rhiopites* spp. and *M. howardi* could support the prevalence of coastal environment (Oboh et al., 1992; Jaramillo and Dilcher, 2001; Eisawi and Schrank, 2008).
In the studied Nukhul Formation, the palynomorph recovery and also organic matter is generally low which shows a somewhat low preservation potential may be attributable to sub-aerial oxidation. The lower Shoab Ali Member is characterized by the presence of partially degraded subordinate cuticles, few algae, and little AOM, suggesting deposition in distal facies tracks of a fluvial system (cf. Eisawi and Schrank, 2008). A shift toward a lacustrine environment, under possible shallow-marine anoxic to dysoxic conditions can be inferred from the APP and APF plots (Figures 4a, b), showing some AOM abundance and coincides with the freshwater fern Magnastriatites howardi, which is an indicator of aquatic habitats such as coastal swamps and floodplain (Germeraad et al., 1968). Additional evidence for humidity (may be on a local scale) beside the fungal record is derived from the fair occurrence of spores of Osmunda (Osmundacidites), Polypodiaceae (Polypodiisporites), which are indicative of more humid conditions (Wilkinson and Boulter, 1980).

Rull (1997) assumed an interpretation for the presence of Botryococcus and Pediastrum in marine sediments with some dinocysts and foraminiferal linings as follows: (1) It is possible that these genera effectively tolerate brackish conditions; (2) A tidal activity and transport by rivers mix fossils from different sedimentary environments. The two scenarios of Rull (1997) are possible in our case especially as tidal activity has been recorded in the Nukhul Formation before (Carr et al., 2003).

Foraminiferal linings show very few and scattered occurrences in the Ghara Member but they are typical of shallow water, coastal sediments (Stancliffe, 1989), and some mixing with continental fossils can be expected (Rull, 1997).

The characteristic taxa for the Ghara Member are some mangrove elements such as Zanocostites ramonae and Rhiopites spp., marine representatives (foraminiferal linings and dinocysts) and the freshwater algae mainly Botryococcus and Pediastrum. Altogether they reflect sedimentation in coastal and shallow-marine or lacustrine environments, close to mangrove vegetation (Figure 6). In addition, ferns (e.g. Verrucatosporites and Laevigatosporites) and reported fungal spores, representing another type of coastal swamps common in the Neo-tropics (cf. Rull, 2001). The relative absence of environmentally restricted taxa in the Shoab Ali Member, make it difficult to provide a clear interpretation of the depositional environment. From the obtained results as well as compilation of earlier studies, we assume a continental setting, strongly influenced by both tectonic and climatic forcing factors besides eustatic changes.
CONCLUSIONS

The palynological, palynofacies and organic geochemical analyses of 56 cutting samples from the Nukhul Formation in the GH 404-2A and SA-E6A wells, Gulf of Suez, Egypt, have led to the following conclusions:

(1) It is possible to separate the Nukhul Formation into the upper Ghara Member, which is mainly lacustrine/lake to shallow-marine, and the lower Shoab Ali Member, which was deposited under fluvial or continental conditions. The Ghara Member in our studied wells is not only represented by evaporite but also by some calcareous intervals.

(2) These two members are separated by a massive fungal/algal proliferation, which could serve as a representative signal for the Oligocene/Miocene Boundary (OMB) within the studied sequence.

(3) The Nukhul Formation generally yielded restricted and only in parts diverse palynomorph assemblages. Minor taxa show a limited biostratigraphic potential and may support the proposed position of the OMB. Most of the reported palynomorphs are Early Miocene although the presence of some taxa supports our opinion to extend the Nukhul Formation into Late Oligocene. The best examples to support such an interpretation are: *Fagraeapollis reticulatus, Praedapollis africanus*, *Foveotricolporites gonzalezi* and *Polypodiaceoisporites pseudopsilatus*.

(4) Geochemical analyses show that most of our samples belong to kerogen III type, showing good reservoir characteristics particularly within the Shoab Ali Member.

(5) Palynofacies analyses using both the transmitted and fluorescence microscopy enable us to differentiate the studied material into three of two main (PF-Ia, PF-Ib and PF-II) palynofacies assemblages. The results of this investigation reveal the following conclusions: (a) PF-Ia and PF-Ib are the dominant facies within the Ghara Member, showing a suboxic-anoxic environment. Kerogen III is established for these two assemblages despite the relative difference between their palynofacies content. (b) PF-II mainly dominates the Shoab Ali Member, with a clear transition to kerogen III or II type with more phytoclasts input supporting a somewhat continental suboxic-anoxic basin with low AOM.

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