The Oligocene Palynology and Palaeoclimates of Northern Egypt as Recorded in the Dabaa Formation

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Abstract

In this chapter, we present the story of the Oligocene palynostratigraphy and floral composition in Egypt as told by the preserved palynomorphs in the Dabaa Formation. It is a step in establishing a regional biostratigraphic and vegetational framework for the Oligocene, thus making future integration of regional and global schemes possible, especially when more independent biostratigraphic information becomes available. Synthesis of published materials from surface and subsurface sections points to significant floral changes during the Oligocene time leading to the development of a number of different vegetation belts in this part of North Africa. These include coastal mangrove forests, back-mangrove wetland ecosystems such as freshwater marshes, lakes, and swamps, as well as inland forests and open woodland-savannah habitats. The prevailing climate is interpreted to have been warm and humid with local or seasonal dry conditions.

Keywords

Oligocene Palynology • Palynostratigraphy • Biostratigraphy • Dabaa Formation • Egypt • North Africa

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1 Introduction

The Paleogene palynology of Egypt, especially the Oligocene, is less studied than other periods. Only a small number of contributions have been carried out on Cenozoic strata in general (Kedves, 1971, 1985; El-Sabrouty, 1984; El-Bassiouni et al., 1988; Takahashi & Jux, 1989a) and on the Oligocene particularly (e.g. El Beialy et al., 2019). Most of these previous studies have primarily focused on taxonomy, palynostratigraphy, and palaeoenvironmental interpretations in addition to palynofacies (El Beialy et al., 2016). Only recently studies that tackled vegetation analysis have emerged (El Atfy et al., 2021, 2022). Previous palynological investigations on the Oligocene sedimentary successions of the north Western Desert of Egypt include Kedves (1971, 1984, 1985), El-Sabrouty (1984), El-Bassiouni et al. (1988), Takahashi and Jux (1989a), El Beialy et al. (2019), Mohamed et al. (2020). Those targeting the Nile Delta include exclusively El-Beialy (1988a, 1990a, b), despite dealing with the Oligocene as part of their studied successions with no detailed data. To the best of the authors' knowledge, no other palynological data exist for the Oligocene from any other location in Egypt (Fig. 1). The goal of this chapter is to review the previous contributions and shed more light on the palynostratigraphy, vegetation, and palaeoclimate of the Oligocene deposits in Egypt with a special emphasis on the Dabaa Formation.

2 Oligocene in Egypt: Geology and Facies Distribution

Fluvio-deltaic and shelf progradation were the dominant depositional schemes in the Oligocene in northern Egypt (Fig. 2). The innermost (southern) basins were completely filled by the end of the Eocene. Oligocene thickness variations show an irregular bottom topography in the north's remnant basin, which was continuous and open. In the

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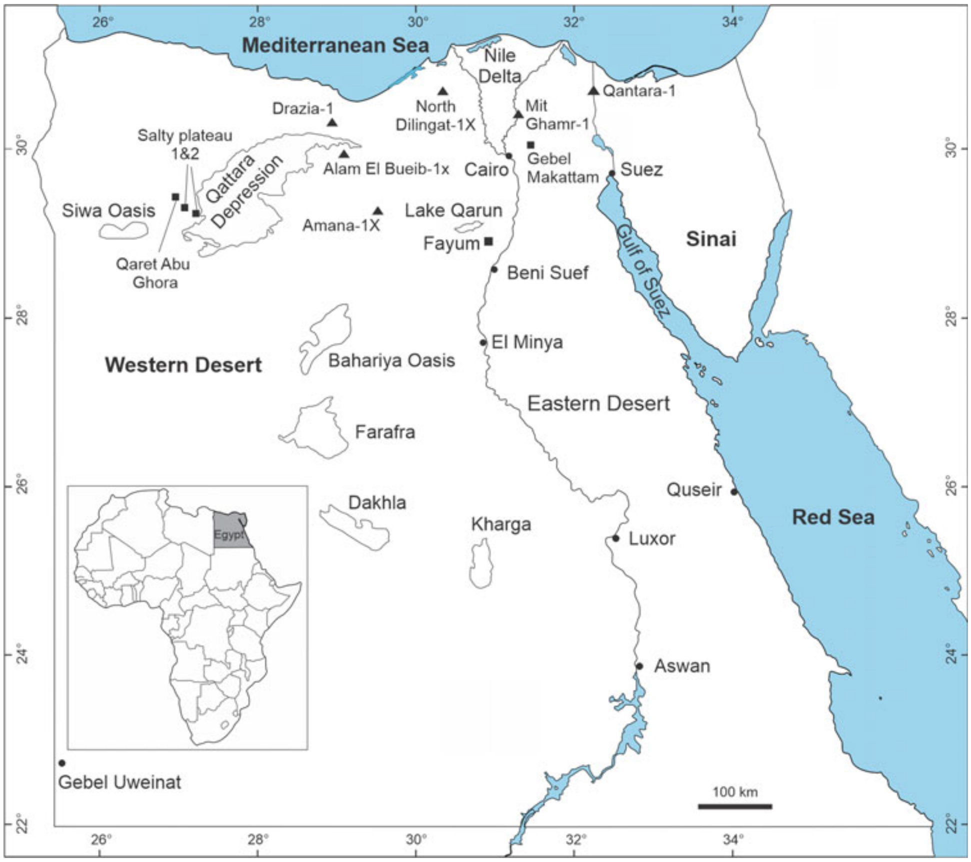


Fig. 1 Location map of Egypt showing the previously studied Oligocene locations and boreholes; triangles represent subsurface occurrences while quadrangles denote surface exposures

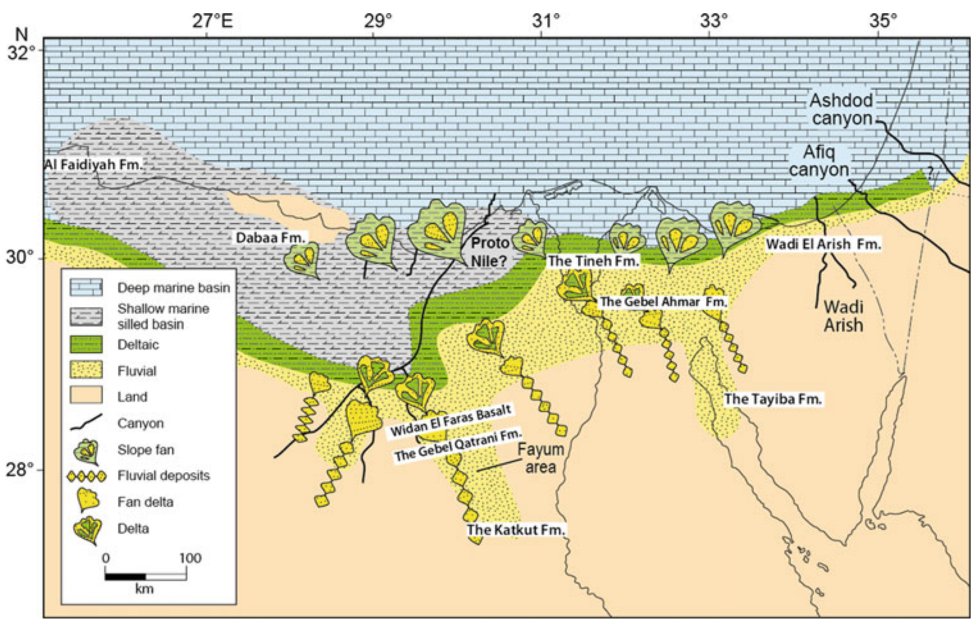


Fig. 2 Egypt during the Oligocene time, simplified and redrawn after Salem (1976) and Kuss and Boukhary (2008)

Oligocene basin, these morphologic irregularities were acquired from the preceding structures and/or compaction of the underlying sediments. Terrigenous clastic deposits from the Oligocene continued to prograde north-westward over an immature shelf with passive ridges extending northeast–southwest (Salem, 1976). Irregular bathymetry is an "immature shelf", regardless of where it originated (Emery, 1968). The shelf was prograded by Oligocene sedimentation filling the inner lows first. The northern flank of the next ridge became the shelf's outer slope when the shelf edge had prograded distant enough northwest to reach it. Submarine highs were also seen in the northwest during the same period. This is corresponding to a shelf with structural dams, as outlined by Curray (1965). It is difficult to determine the features and types of depositional processes that operate in the region because of the immature Oligocene shelf as well as lack of data. Salem (1976) assumed that high constructive delta systems dominated the scene due to the high content of mud. Many slope cones (turbidites) are likely to be present in the Oligocene sedimentary section. Oligocene slope fans are most likely to be found in the area directly south of the existing coastline (Fig. 2).

The Oligocene sedimentary facies comprise sandstones, siltstones, and clays of the deltaic Gebel Qatrani, the fluviatile Gebel Ahmar, and the lacustrine Nakheil formations as the most significant Oligocene outcrops in the Fayum area and its surroundings (Fig. 2). Osman (2003) proposed that the Oligocene shoreline passed north of the Fayum Depression, skirted the southern slope of the El Arag, and continued into the Siwa Oasis in his study of the El Arag area in the north Western Desert. Moreover, El Heiny and Enani (1996) termed the Oligocene (Chattian) section north of the Nile Delta and offshore to the northwest of the Sinai Peninsula the Tineh Formation. The latter is made up of prodeltaic sediments that interfinger with open-marine basinal deposits further offshore (Dolson et al., 2002). As ancient river systems developed in the Red Sea and Uweinat regions, they drained ancient uplifted areas where these facies were distributed regionally throughout the early Oligocene. Issawi and McCauley (1993) proposed that these palaeo-drainage systems shifted position during the Oligocene–Miocene interval. Even though the exact pathways, timing, and position of several valleys and canyons are still being debated (Dolson et al., 2002), it is undeniable that these features are critical to the far-piece transport of clastics as well as several shelf dissections and incision processes.

Many episodes of volcanic activity occurred in Egypt during the Cenozoic. The most widespread volcanic activity was during the Paleogene. Isotopic age determinations indicate that several successive volcanic pulses occurred in the late Eocene with subsequent extensional phases ranging from late Oligocene to middle Miocene (Ibrahim, 2008). The Red sea opening was immediately conjugated with this volcanic activity. It is strongly believed that the Red Sea gained its physiognomies and isolation from the Tethys as a result of the Cenozoic tectonics, starting with the late Eocene–Oligocene-early Miocene uplift and expansion of a major fault system (Ibrahim, 2008). These faults trending NW–SE show a substantial displacement along fault planes dipping away from the Red Sea. The basaltic volcanic rocks are extensively distributed in the northern part of Egypt underneath the Nile Delta and the Western Desert (Said, 1981; Williams & Small, 1984). Also, some isolated outcrops along the Fayum-Abu Rowash, Cairo-Suez, and Tihna-El Bahnasa stretch. The volcanic rocks of the Western Desert vary in composition and belong to more than one phase of volcanic activity (Meneisy & Abdel-Aal, 1984).

During the Oligo-Miocene, normal faults of a northwest trend echoed the opening of the Red Sea and Gulf of Suez (Meshref, 1990). The main trend affecting the area is a roughly east northeast–west southwest and is represented by a normal fault. This trend is reflected in the elongation of the scarp, and of Lake Qarun, to the south. This trend, called the "Tethyan" trend, has affected most of the African continent since earlier Mesozoic times (Pavoni, 1993). It could be the echo of the opening of the South Atlantic Ocean.

The major faults affecting the area are mostly of the normal type and have different trends, such as a roughly north–south or north northwest–south southeast trend, especially east of the Qatrani scarp. The intersection of the latter with the east northeast–west southwest master fault gives rise to the sudden change of the scarp direction and the drainage pattern. A group of northwest–southeast trending faults affects the basaltic rocks west and east of Widan El Faras. These synbasaltic trends could be due to the opening of the Red Sea and the Gulf of Suez. A northeast–southwest right-lateral strike-slip fault west of the Gebel Qatrani scarp brings the Gebel Qatrani Formation side by side with the basaltic rocks (Ibrahim, 2008).

3 Lithostratigraphic Framework of the Oligocene in Egypt

The Oligocene deposits overlie disconformably late Eocene sediments and they are only recorded in the northern part of Egypt. They were classified into two well-defined facies: fluvial facies of sands and gravels with minor deltaic facies and marine facies of shales and negligible limestone interbeds in the subsurface of northern Egypt. The volcanic, geyser activity and tectonism affected the Red Sea region and the belt of highs amid the stable and unstable shelves during the Oligocene. This governed the distribution of these sediments to a large extent (Said, 1962). Oligocene fluviatile sediments crop out along a narrow belt extending from Suez to Fayum via Cairo and onward into the Western Desert (Fig. 2). Small and isolated outcrops of this facies are also known from west of Beni Suef and the Bahariya Oasis. Generally, these deposits are difficult to date and they are classified as Oligocene based on stratigraphic correlation, and most lack reliable biostratigraphic data (Ibrahim, 2008).

3.1 The Gebel Qatrani Formation

Considerably, thick deposits of Oligocene sands occur along the northern and western scarps of the Fayum depression. The Gebel Qatrani Formation is the subject of the classical work of Beadnell (1905), Said (1962), Bowen and Vondra (1974), and Bown et al. (1982). It is mainly controlled by the presence or absence of the Widan El Fars basalt and by the development of large east–west trending normal faults. Beadnell (1905) used the term "Fluvio-marine series" however, Said (1962) favours the term Gebel Qatrani Formation or "Qatrani Formation" for the variegated sandstones, gravelly sandstone, sandy mudstones, limestones, and shales that form Gebel Qatrani and extended northeast to greater Cairo area.

In its type section, the Gebel Qatrani Formation is about 340 m thick; however, it thins considerably to the west. It is readily distinguished from the underlying dark green and grey Dir Abu Lifa Member of the Qasr El Sagha Formation by the dominance of brightly coloured-variegated sandy mudstones, sandstones, and gravelly sandstones. The overlying alluvial Miocene Khashab Formation, the Gebel Qatrani is separated by an erosional unconformity, followed in places by up to 25 m of Widan El Faras Basalt (Bowen & Vondra, 1974), a dense, iron-rich, cliff-forming unit capping the Gebel Qatrani. In areas where the Widan El Faras Basalt is absent, the Gebel Qatrani Formation is overlain by the Khashab Formation with an erosional unconformity (Ibrahim, 2008).

3.2 Widan El Faras Basalt

The Widan El Faras Basalt is a dark, generally densely aphanitic iron-rich extrusive basalt. It seems to be composed of a single flow in areas where it is thinnest; however, weathered and sometimes charred contacts within the basalt, as well as thin coarse sandstone interbeds (some containing mixed quartz sand and basaltic debris), demonstrate the existence of no less than two or three distinct flows over a great area of exposure of the basalt. These shallow scours show that unconformities of unknown magnitude are contained in the Widan El Faras Basalt. The basalt's upper contact with the overlying Khashab Formation is also erosional and, in the area between Widan El Faras and Tel Homar, is marked by at least 9 m of relief (Bowen & Kraus, 1988).

3.3 The Gebel Ahmar Formation

A typical example of the sands and gravels of the Cairo-Suez district is exposed in Gebel Ahmar, east of Cairo (Barron, 1907; Shukri, 1954). Lateritic soils cap the deposit in places, inducing red colouration to the underlying sediments. Shukri (1954) assumes that fluids ascended along faults caused the colouration and silicification of the Oligocene sands and gravels of Cairo-Suez district. The Oligocene Gebel Ahmar Formation detours Gebel Mokattam to the north, where it blankets the desert between Cairo and Suez. The E–W faults also displaced the continental Miocene sediments in the north counter to the Oligocene Gebel Ahmar Formation to the south. Within this huge surface area, prominent hills rise 60–70 m above the desert plain. The Gebel Yahmoum El Asmar, Gebel El Khashab (Petrified Forest), and Gebel Ahmar overlook Nasr City to the northeast of Cairo (Issawi et al., 2009).

The Gebel Ahmar Formation (at its type section) is composed of 40–100 m thick, coarse-grained, cross-bedded, vividly coloured, frequently friable sands, with a hard quartzitic dark brown bed at the top. At Gebel Ahmar, geyser action is well seen in the dark red and dark brown silicified tubes which cut erratically through the sands rising several metres in castle-like forms. Regrettably, this national park—as it should be—was damaged under the construction of the Arab Contractors Medical Centre and Elmokawelon Arena; even the designation transformed into Gebel Akhdar (Issawi et al., 2009)

3.4 The Nakheil Formation

The Nakheil Formation is recorded from the synclinal troughs of the coastal areas of the Red Sea south of Quseir city. Lithologically, the Nakheil Formation is 60 m thick of very coarse breccia and angular blocks of limestone and chert concretions that originated from the underlying Thebes Formation. Fine lacustrine sediments made of fine carbonates, clays, and sandstones alternate with the coarse breccia. No in-situ fossils were recorded except those reworked from the older beds (El Akkad & Dardir, 1966).

3.5 The Katkut Formation (Serir Deposits)

The Katkut Formation or Katkut Gravels was first presented to represent the post-Eocene strata west of the Nile Valley by some authors (e.g. El Hinnawi et al., 1978; Issawi & Osman, 2008; Issawi et al., 2009), encompassing all the reddish-brown coarse clastics that unconformably overlie the early Eocene limestone beds. These are allocated to late Oligocene to early Miocene age (Abu Seif, 2015; Mahran et al., 2013). These fluvial sediments resemble the undistinguishable gravel unit west of Nagh Hammadi (Klitzsch et al., 1987), to Higaza Formation east of Qena (Philobbos & Abdel Rahman, 1990), and to pre-Eonile gravels in the extent between Aswan and Nagh Hammadi, west of the Nile (Lansbery, 2011). Mahran et al. (2013) and Abo Seif (2015) documented three informal lithologic subdivisions of the Katkut Formation; gravel, sands and silts, gravel, and sands.

The Katkut Gravels are developed on the plateau surface to the southwest of Sohag opposite El Minshah village with a greater thickness ca. 50 m. The gravels are embedded in a highly ferruginous red matrix, stretching for 30 km on the western Limestone Plateau and also on its eastern slope (Issawi et al., 2009).

3.6 The Tayiba Formation/Red Beds

The Tayiba (also referred to as the Abu Zeneima Formation in some works) overlies unconformably the late middle Eocene Tanka Formation in the Abu Zeneima-Feiran area in Sinai and the Gulf of Suez. It comprises a 5-m-thick basal conglomerate, a 15-m-thick sequence, and an upper 15-m-thick porcelaneous limestone bed, coarsening-upward siltstone beds (Ibrahim, 2008).

3.7 The Tineh Formation

The Tineh Formation represents the early Oligocene-lower Miocene sediments in the central and eastern Nile Delta regions. The Qantara-1 well penetrated a thickness of 1086 m of the Tineh Formation, which consists of dark grey shales interbedded with sandstone bands representing prodeltaic sediments, interfingering with open-marine basinal facies further offshore (Dolson et al., 2002). The penetrated late Oligocene-early Miocene section in this well includes planktonic foraminifera deposited under outer neritic to upper bathyal conditions (El Heiny & Morsi, 1992). The Tineh Formation was deposited on a submarine high during the late Oligocene time as evidenced by the absence of its lower part and the relatively thin units encountered in this formation (El Heiny & Enani, 1996).

3.8 Wadi El Arish Formation

The Wadi Arish Formation was so-called after Wadi Arish, a large valley that drains the Central Sinai Peninsula in the south and laid-off into the Mediterranean Sea to the east of El Arish town, it was divided into three members (Kuss & Boukhary, 2008). The lower member is best detected at the eastern part of Gebel Risan Eneiza, 20 km south of El Arish. The member unconformably overlies the karstified Albian limestones of the Risan Eneiza Formation. The Lower member of the Wadi El Arish Formation is 42-m-thick, coarse-grained sandstone intercalated with gypsum layers, including larger foraminifera and coralline algae (Kuss & Boukhary, 2008).

The middle member is 8 m thick consisting of clay and marls with thin intercalations of nodular limestones or rhodoliths. The upper member is 26.5 m thick, well-bedded, or massive limestones, including larger foraminifera and algal rhodoliths. The foraminiferal assemblage, identified by Kuss and Boukhary (2008) from the Wadi El Arish Formation includes *Nephrolepidina*., *Lithothamnium* sp., *Lithoporella melobesoides* (Foslie), *Sporolithon* sp. and *Neogoniolithon* sp.

3.9 Dabaa Formation

The Dabaa Formation is defined by Norton (1967) as a subsurface rock unit in the north Western Desert. The name was also used by Abdallah (1967) to designate Pliocene pink limestone of the Mediterranean littoral zone, but this is a homonym (Hantar, 1990). In this chapter, we employ the term Dabaa Formation in the sense of Norton (1967). It consists mainly of grey shale and thin limestone interbeds and has a uniform thickness in the north Western Desert (Norton, 1967). The Dabaa Formation has a thickness of 442 m at its type section (Dabaa-1 well) and its depositional environment is mainly inner shelf to littoral that changes to estuarine near its top (Issawi et al., 2009).

The Dabaa Formation has been dated to the Oligocene (Issawi et al., 2009), though oil geologists account it to be late Eocene–Oligocene. Alike was anticipated by Hantar (1990) who measured the lower 33 m of the Dabaa Formation, in its type section, to reveal upper Eocene, and the upper 209 m to be Oligocene. Palynologically, it was dated as Rupelian (early Oligocene) or late Eocene–Oligocene, based on dinocyst evidence from surface material in the vicinity of the Qattara Depression (El Beialy et al., 2019).

3.10 Al Faidiyah Formation

In Libya across the borders from Egypt, El Deftar and Issawi (1977) mapped a 55 m section of alternating limestone, marly limestone and clay which they named the Al Faidiyah Formation. Its faunal content included both macro- and microfossil assemblages including *Brisoposis frassi* Fuchs, *Tellina lacunosa* Chemnitz, *Cardium gallicum* Mayer, *Lepidocyclines,* and *Operculines* that suggest Chattian to Aquitanian age. It makes the basal cliffs overlooking the Mediterranean near Tobruk, hence there is a great possibility that the section extends into Egypt. The vertical cliffs of the Sallum area have to be checked for the Al Faidiyah Formation especially when the unit is overlain by Al Jaghbub in Libya, a unit also known in west Egypt (Ibrahim, 2008).

It is worth mentioning that palynological dating based mainly on dinocyst evidence on the Al Faidiyah Formation has been introduced by El-Mehdawi and El Beialy (2008) from surface samples from Al Jabal Al Akhdar, NE Libya. They established a late Oligocene to early Miocene age, rather than early Miocene which could reveal that Al Faidiyah Formation is younger than the Dabaa Formation and other Oligocene equivalent rock units in Egypt.

4 Terrestrial Megaflora

The Oligocene plant-bearing strata in North Africa are generally rare. They produce mostly petrified wood indicative of a diverse terrestrial floral community (e.g. Dupéron-Laudoueneix & Dupéron, 1995; Jacobs et al., 2010; El-Saadawi et al., 2020). Based on the spatial variation within such community, the Oligocene vegetation in North Africa is believed to have been tropical forests along the Tethys coastal line, transitioning to an inland zone of mixed woodland and grassland (Boureau et al., 1983; Louvet, 1971).

In Fayum, Egypt, an area well-known for its palaeobiodiversity, a plethora of vertebrate fossils as well as a diverse early Oligocene macrofloral community are preserved within the Gebel Qatrani Formation. Floral remains include wood, fruits, seeds, and leaves (Blanckenhorn, 1921; Kräusel & Stromer, 1924; Kräusel, 1939; Bown et al., 1982; El-Saadawi et al., 2014, 2020; Stull et al., 2020). Louvet (1971) interpreted the floral community of the Gebel Qatrani Formation to represent a portion of a tropical forest formed on the northern coast of Africa. However, to the authors' knowledge, no more data are available outside Fayum detailing the megafloral assemblages of Egypt.

5 Floristic Composition, Vegetation, and Palaeoclimatic Inferences

In Egypt, Oligocene spores and pollen were initially reported from the regions of Abu Rauwash and Moquattam (Kedves, 1971, 1985). This was followed by studying the spores and pollen of the late Eocene-early Oligocene Qasr El Sagha Formation in the Fayum area (Takahashi & Jux, 1989a). Recently, El Atfy et al., (2021, 2022) studied the Dabaa Formation sporomorphs from surface exposures near the Qattara depression and from the Amana-1X well, respectively, both located in the north Western Desert (Fig. 1).

The palynofloral elements known from the Dabaa Formation in the north Western Desert of Egypt display taxonomic resemblances to the hitherto-described Oligocene macroflora, and palynoflora from the Oligocene of Egypt (e.g., Kedves, 1985; El Atfy et al., 2021, 2022 and discussion therein). These elements include, for example, fern and lycopsid spores such as *Crassoretitriletes vanraadshooveni*, *Magnastriatites howardi*, and *Verrucatosporites usmensis*, gymnosperm pollen like *Ephedripites* in addition to diverse angiosperm pollen species such as *Aceripollenites striatus, Bombacacidites nacimientoensis, Chenopodipollis multiplex, Echiperiporites estelae, Monoporopollenites annulatus,* and *Peregrinipollis nigericus*. The presence of these and other species (Fig. 3) within the Dabaa Formation indicates mixed tropical habitats including coastal mangroves with back swamps, ever-wet inland forests, and open woodland-savannah ecosystems. Temperate indicative species may have also been present (El Atfy et al., 2021, 2022). A distinct vegetational pattern could be outlined during the deposition of the Dabaa Formation in the north Western Desert as shown in Fig. 4. This vegetational pattern has been described so far from only the Dabaa Formation based on palynomorphs. Therefore, further palynological investigations are required to confirm its extent, although macro-palaeobotanical data point to an extension of this pattern across all of northern Africa during this period.

In terms of palaeoclimate, the Dabaa Formation preserves a terrestrial palynomorph assemblage indicative of predominantly warm and humid conditions (El Atfy et al., 2021, 2022). This is evident by the abundance of angiosperm pollen species of the families Arecaceae, Malvaceae (subfamily Bombacoideae), Dipterocarpaceae, and Ctenolophonaceae. The abundance of precipitation and humidity are also inferred from the presence of fern and lycopsid spores which are known to be water-loving and thrive under wet conditions. Local or seasonal dry and arid conditions may have existed based on the reported species of *Chenopodipollis, Monoporopollenites,* and *Ephedripites* (El Atfy et al., 2021) from surface samples collected from the vicinity of the Qattara Depression.

6 Marine Palynomorphs, Primarily Dinoflagellates

The analysed samples from the Dabaa Formation in the north Western Desert (namely from the Qattara profiles and the Amana-1X well) have high to moderate recovery of well-preserved dinoflagellate cyst assemblages (Fig. 5) besides sporomorphs (Fig. 3) and palynofacies particles as

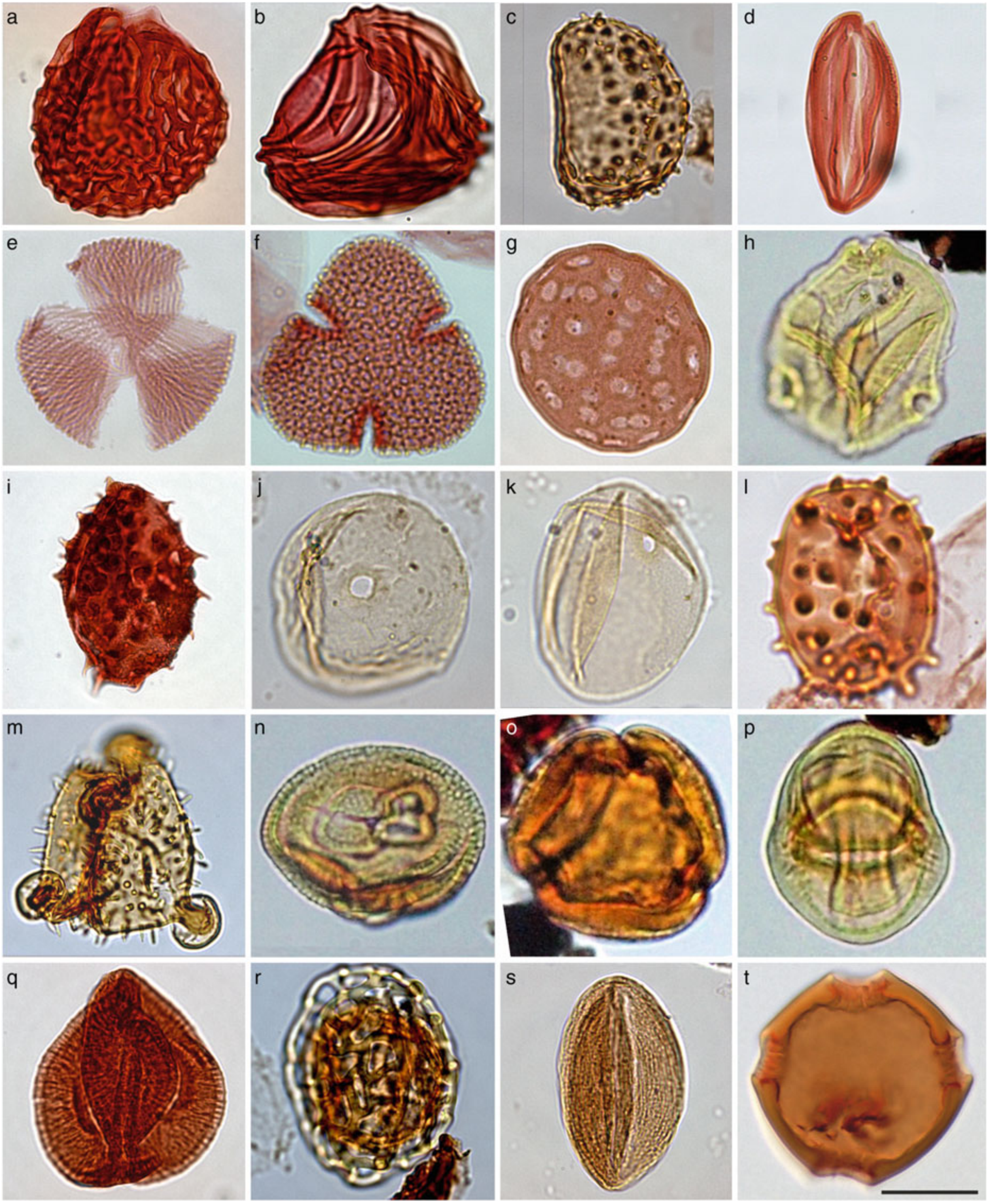


Fig. 3 The most significant Oligocene spores and pollen assemblage from the Dabaa Formation, Qattara area, north-Western Desert, Egypt. All photomicrographs are in bright field illumination. An England Finder reference follows the sample number for each specimen. **a–c**; fern and lycopsid spores **a** *Crassoretitriletes vanraadshooveni* Hopping & Muller; P1-2_X37. **b** *Magnastriatites howardi* Germeraad, Hopping & Muller; P1-1_O35. **c** *Verrucatosporites usmensis* (van der Hammen) Germeraad, Hopping & Muller; P1-1_U39. **d** Gymnosperm; *Ephedripites* sp.; P1-2_O31. **e–t**, angiosperms; **e** *Aceripollenites striatus* (Pflug) Thiele-Pfeiffer; P1-2_L30. **f** *Bombacacidites nacimientoensis* (Anderson) Elsik; P1-1_035. **g** *Chenopodipollis multiplex* (Weyland & Pflug) Krutzsch; P1-2_L34. **h** *Corsinipollenites* sp.; Am-09. **i** *Echiperiporites estelae* Germeraad, Hopping & Muller; P1-2_K38. **j** *Graminidites* sp.; P2-1_H35. **k** *Monoporopollenites annulatus* Jaramillo & Dilcher; P1-1_T37. **l** *Mauritiidites crassiexinus* van Hoeken-Klinkenberg; P2-1. **m** *Auriculopollenites echinatus* Salard-Cheboldaeff; Am-15_K21. **n** *Retibrevitricolporites ibadanensis* Jan du Chêne et al.; Am-07. **o** *Psilatricolporites crassus* van der Hammen & Wymstra; Am-10. **p** *Zonocostites ramonae* Germeraad, Hopping & Muller; Am-13. **q** *Perfotricolpites digitatus* González; P1-2_N38. **r** *Peregrinipollis nigericus* Clarke; Am-07_P22. **s** *Striatopollis catatumbus* (González) Ward; P1-1_E26. **t** *Pachydermites diederixi* Germeraad, Hopping & Muller; Am-01_J22. Scale bar = 20 μm

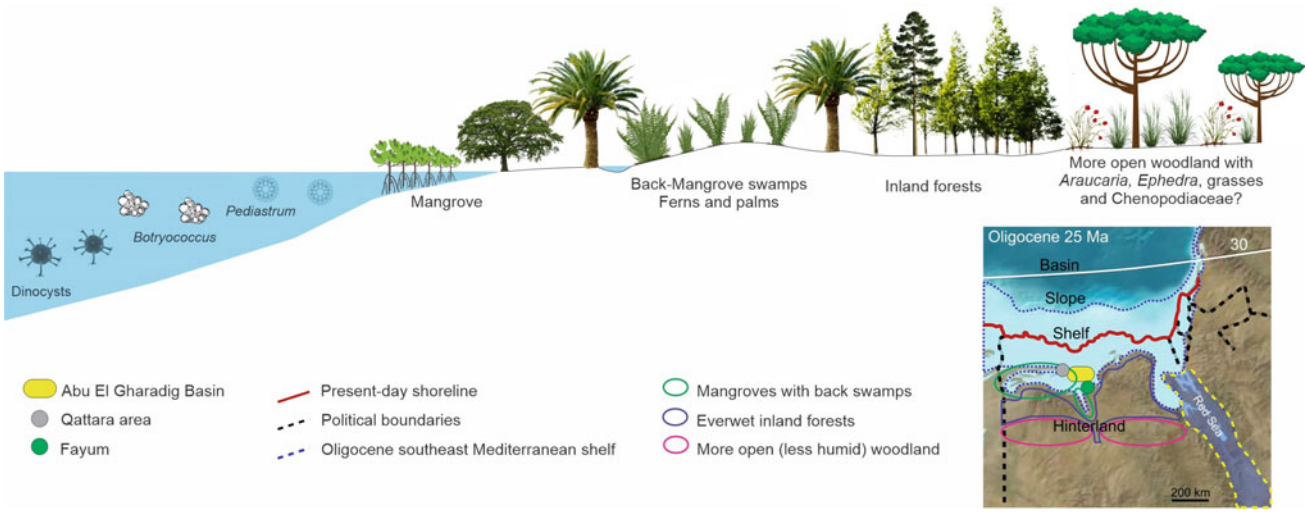


Fig. 4 Tentative reconstruction of the probable vegetation belts during the deposition of the Dabaa Formation, Amana-1X well, north–Western Desert, Egypt (modified after El Atfy et al., 2021). The palaeogeographic base map is after Ron Blakey, Colorado Plateau Geosystems, Arizona, USA (http://cpgeosystems.com)

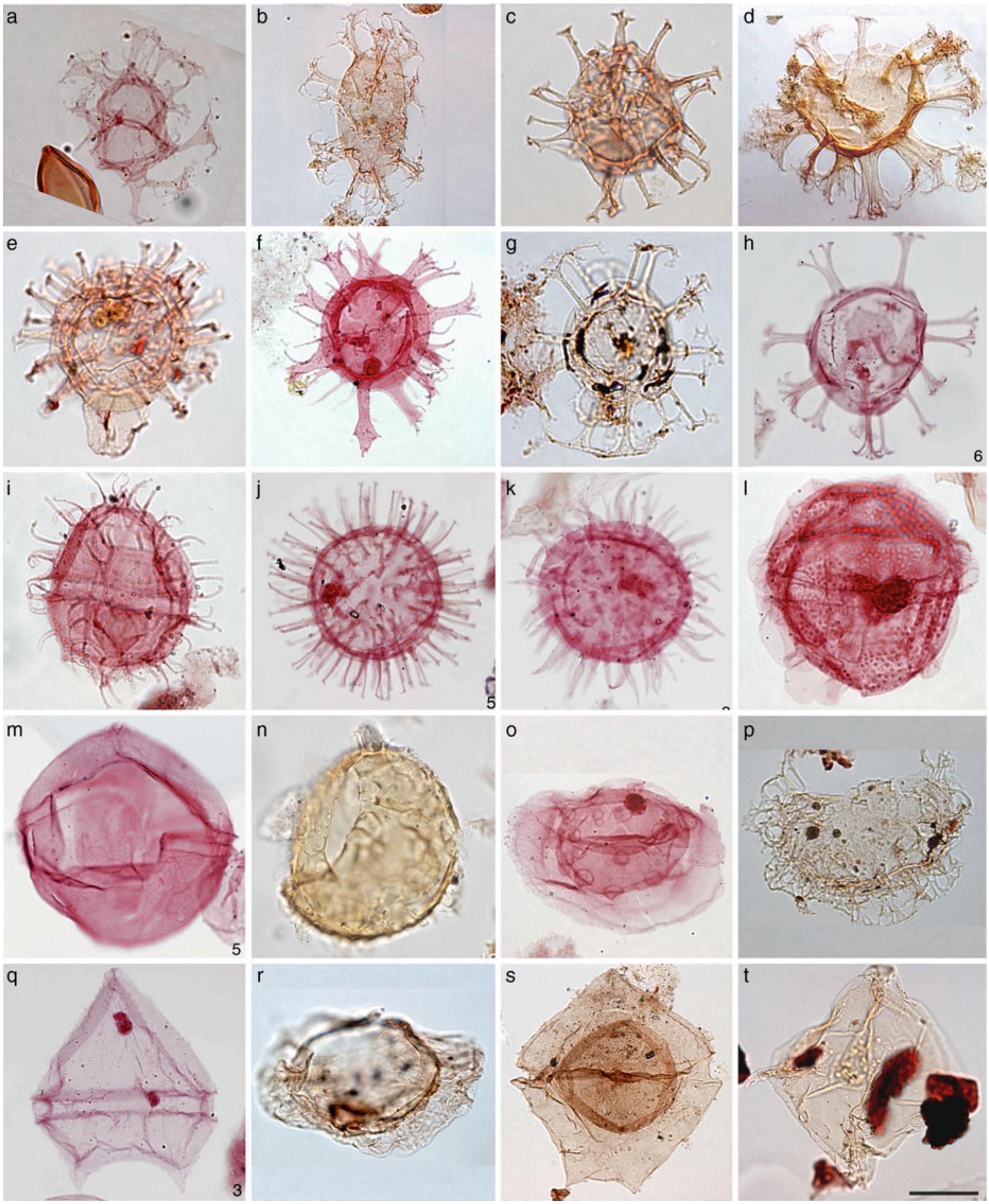
well as a few numbers of acritarchs, dispersed fungal remains, algae, and foraminiferal test linings.

The Dabaa Formation contains diverse assemblages of dinoflagellate cysts, some of which are sufficiently restricted as to be useful in age determination. The dinocyst *Thalassiphora pelagica* (Fig. 5r) is not usually recorded above the lowermost Miocene (e.g. Fensome et al., 2009 for offshore Canada). However, it has a known range of Late Cretaceous to late Oligocene in Italy, Turkey and Tunisia (Wilson, 1971; Powell, 1986; Erkmen & Sadek, 1981; Torricelli & Biffi, 2001). In Egypt, *T. pelagica* was previously recorded from the late Oligocene of the Nile Delta, (El-Beialy, 1990b), the early Miocene of North Sinai (El-Beialy & Gheith, 1989), the Gulf of Suez (El Atfy et al., 2017), and the early Oligocene of the north Western Desert (El Beialy et al., 2019).

Diphyes colligerum (Fig. 5e) shows a reliable end Lutetian (Chron C19n) highest occurrence (HO) in the North Sea and the Norwegian–Greenland Sea (Eldrett et al., 2004), but exhibits a range of mid-Lutetian to lower Oligocene highest occurrences in places like Italy, Denmark, Germany, and the Labrador Sea (Brinkhuis & Biffi, 1993; Heilmann-Clausen & van Simaeys, 2005; Köthe & Piesker, 2008; Firth et al., 2012). In Egypt, this species was recorded from the upper Eocene to lower Oligocene of the north Western Desert and the Nile Delta (El-Bassiouni et al., 1988; El Beialy, 1988b; El-Beialy, 1990a; El Beialy et al., 2019).

Dinopterygium cladoides (Fig. 5l) has an early Oligocene to middle Miocene HO in Africa. It was previously reported from the early Oligocene of Tunisia (Torricelli & Biffi, 2001), the late Oligocene of Morocco (Chekar et al., 2018), the Oligocene of the north Western Desert, Egypt (El Beialy et al., 2019), and the middle Miocene of the Gulf of Suez, Egypt (El Beialy & Ali, 2002; Soliman et al., 2012).

Fig. 5 Most significant Oligocene dinocyst assemblage from the Dabaa Formation, Qattara area, north-Western Desert, Egypt. All ▶ photomicrographs are in bright field illumination. An England Finder reference follows the sample number for each specimen. **a** *Distatodinium craterum* Eaton; P1-2_N36. **b** *Distatodinium paradoxum* Brosius; Am23B_E24.3.4_5.6 m.; central body length 43 µm. **c** *Melitasphaeridium pseudorecurvatum* (Morgenroth) Bujak et al.; Am24B_D36_515.1 m.; central body maximum diameter 30 µm. **d** *Cordosphaeridium inodes* (Klumpp) Eisenack; Am22B_T14_496.8 m; central body maximum diameter 45 µm. **e** *Diphyes colligerum* (Deflandre & Cookson) Cookson, emend. Goodman & Witmer; Am5A_O17.1.3_341.4 m. **f** *Hystrichokolpoma rigaudiae* Deflandre & Cookson; P1-2_W36; central body length 41 µm. **g** *Enneadocysta multicornuta* (Eaton) Stover & Williams, emend. Stover & Williams; Am1A_K13.1.3_304.8 m; central body maximum diameter 36 µm. **h** *Homotryblium floripes* subsp. *floripes* (Deflandre & Cookson) Stover; P1-1_U40; central body maximum diameter 45 µm. **i** *Phthanoperidinium comatum* (Morgenroth) Eisenack & Kjellström; P1-2_K41; central body length 44 µm. **j** *Polysphaeridium*? sp. 2; P1-2_C31, central body maximum diameter 39 µm. **k** *Lingulodinium machaerophorum* (Deflandre & Cookson) Wall; P1-2_C37; central body maximum diameter 40 µm. **l** *Dinopterygium cladoides* sensu Morgenroth; P1-2_C44; central body length 49 µm. **m** *Cribroperidinium tenuitabulatum* (Gerlach) Helenes; P1-2_B43; cyst length 67 µm. **n** *Samlandia*? sp.; Q1_Q15; central body length 52 µm. **o** *Tuberculodinium vancampoae* (Rossignol) Wall 1967; P1-2_W38; central body maximum diameter 50 µm. *Polysphaeridium zoharyi* (Rossignol) Bujak et al.; P1-2_K42; central body maximum diameter 54 µm. **p** *Glaphyrocysta intricata* (Eaton) Stover & Evitt; Am2A_G21. **q** *Lentinia serrata* Bujak in Bujak et al.; P1-1_W39; cyst length 56 µm. **r** *Thalassiphora pelagica* (Eisenack) Eisenack & Gocht, emend. Benedek & Gocht; Am2A_O17. **s** *Deflandrea phosphoritica* Eisenack; Am11A_U44; central body maximum length 40 µm. **t** *Rhombodinium draco* Gocht; Am16A_T26; central body length 54 µm. Scale bar = 20 µm



Phthanoperidinium comatum (Fig. 5i) has a consistent record of lower Rupelian highest occurrences from the western North Atlantic (Stover, 1977, the Labrador Sea Head & Norris 1989, the North Sea (van Simaeys et al., 2005), Italy (van Mourik & Brinkhuis, 2005), Germany (Köthe & Piesker 2008), Denmark (Śliwińska et al., 2012). In North Africa, *P. comatum* has been documented from the lower Oligocene of the Nile Delta and Qattara Depression of Egypt as well as the Lutetian–Chattian of Morocco (El Beialy, 1988b; El Beialy et al., 2019; Chekar et al., 2018).

Glaphyrocysta intricata (Fig. 5p) has its HO in the lower Miocene of western Tasmania in the Indian Ocean (Brinkhuis et al., 2003) and Germany (Strauss, 1993). The HO of this taxon is also documented from the upper Oligocene of west

Germany (Weiler, 1982; Sonne & Weiler, 1984) and the Nile Delta, Egypt (El-Beialy, 1990b), lower Oligocene of Italy and Tunisia (Brinkhuis & Biffi, 1993; Brinkhuis, 1994; van Mourik & Brinkhuis, 2005; Torricelli & Biffi, 2001). The lowest occurrence of this species has been recorded in the upper Paleocene (Thanetian) of England (Powell et al., 1996).

The genus *Rhombodinium* has a Bartonian–Rupelian stratigraphic range (Fensome et al., 2009). Its top range, Rupelian, is based on the HO *R. draco* (Fig. 5t) by Powell and Brinkhuis in Gradstein et al. (2004) as well as Williams et al. (2004) from the mid-latitudes of the Northern Hemisphere. In Egypt, *R. draco* was recorded from the early Oligocene of the Nile Delta by El Beialy (1988b). Reported anomalous Miocene records of *R. draco* from Russia and East Germany by Zosimovich (1991) and Strauss (1993) are assumed to be reworked.

Enneadocysta multicornuta (Fig. 5g) was recorded in deposits ranging in age from the late Paleocene of India (Khanna, 1979) to early Miocene of Europe (Hochuli, 1978). Elsewhere, this species has the HO in the early Oligocene of Italy (Gruas-Cavagnetto & Barbin, 1988), Mississippi, Alabama (Jaramillo & Oboh-Ikuenobe, 1999, 2001), Poland (Gedl & Leszczynski, 2005), and Azerbaijan (Bati, 2015).

Williams et al. (2004) recorded the first occurrence of *Melitasphaeridium pseudorecurvatum* (Fig. 5c) from the early Eocene (Ypresian) in northern Hemisphere mid-latitudes, southern Hemisphere high latitudes and equator. The high occurrence of this species is recorded from the early Eocene in the southern Hemisphere high latitude (51.4 Ma), the late Eocene (Priabonian) in the equator (35 Ma), and the early Oligocene (Rupelian) in northern Hemisphere mid-latitude (33.07 Ma) (Williams et al., 2004). Elsewhere, the highest occurrence of this species was recorded from the lower Oligocene of Belgium (De Coninck, 1999), Poland (Gedl & Leszczynski, 2005), Italy (van Mourik & Brinkhuis, 2005), and Azerbaijan (Bati, 2015). In Egypt, *M. pseudorecurvatum* was firstly recorded from the lower Oligocene Dabaa Formation in the Qattara Depression, Western Desert (El Beialy et al., 2019).

Lentinia serrata (Fig. 5q) has been previously recorded from the Lutetian to lower Rupelian strata of multiple European countries including England, Belgium, and Italy (Aubry, 1986; De Coninck, 2001; Pross et al., 2010) as well as from the offshore of eastern Canada (Fensome et al., 2008). In North Africa, it was recorded from the middle Eocene of Morocco (Chekar et al., 2018) and from the lower Oligocene of the Qattara Depression, Egypt (El Beialy et al., 2019).

According to Brinkhuis et al. (2009), *Deflandrea phosphoritica* (Fig. 5s) has a mid-Aquitanian highest occurrence in low- and mid-latitudes of the Northern Hemisphere. In North Africa, *D. phosphoritica* was reported from the Eocene of Morocco (Chekar et al., 2018), the early Miocene of Tunisia (Torricelli & Biffi, 2001) and Libya (El-Mehdawi & El Beialy, 2008), and the late Oligocene of the Nile Delta, Egypt (El Beialy, 1990b).

Distatodinium paradoxum (Fig. 5b) has middle Miocene highest occurrences in many parts of the world including the USA (de Verteuil & Norris, 1996), the Canadian Scotian shelf (Fensome et al., 2008), Germany (Köthe & Piesker, 2007), Belgium (Louwye et al., 2000), Norwegian-Greenland Sea (Poulsen et al., 1996), and Italy (Zevenboom, 1995). In Egypt, *D. paradoxum* has its HO in the early Miocene (e.g., El Beialy & Ali, 2002; El Atfy et al., 2017).

Distatodinium craterum (Fig. 5a) has relatively a long stratigraphic range (lower Eocene to middle Miocene, collectively) in Egypt as reported from multiple localities including the Gulf of Suez and the north Western Desert (Ahmed & Pocknall, 1994; El Beialy et al., 2019).

7 Oligocene Palynostratigraphical Framework in Egypt

The excellent state of preservation of dinoflagellate cysts, as well as sporomorphs encountered in the studied materials of the Dabaa Formation by El Beialy et al. (2019) and El Atfy et al. (2021, 2022) promises considerable future advances to the Oligocene palynology of Egypt. This will provide a sound basis for surface and subsurface correlations in the region. This palynostratigraphical correlation (Figs. 6 and 7) is still seriously hampered by the limited extent of published data. This chapter is intended in part to highlight such a gap, as further sample coverage is needed to allow for a more complete regional picture to emerge.

The biostratigraphy of the late Paleogene subsurface sequences in the Western Desert, Egypt, based initially on calcareous microfossils mainly foraminifera and nannoplankton, was discussed by several authors (e.g., Ouda, 1998). So far, there is no standard zonation established for dinocysts or sporomorphs of the Paleogene rocks in Egypt, where published records are existing only from the Western Desert (El Beialy & Kora, 1987; Mahmoud, 1998; El Beialy et al., 2019; El Atfy et al., 2021) and Nile Delta (El Beialy, 1988a, b; El Beialy, 1990a, b).

From a palynostratigraphical perspective, it is still not tenable to provide accurate Paleogene dating and biozonation based on terrestrial evidence in Egypt. However, some sporomorph taxa retrieved in the Egyptian strata (especially from the Dabaa Formation) display biostratigraphic potential, especially when they are correlated with regional counterparts in Africa and the Tethyan realm, as follows:

Cicatricosisporites dorogensis has been recorded from Eocene to early Miocene strata of Nigeria (Legoux, 1978), Tunisia (Torricelli & Biffi, 2001), Cameroon (Salard-Cheboldaeff, 1979), and Sudan (Stead & Awad, 2005;

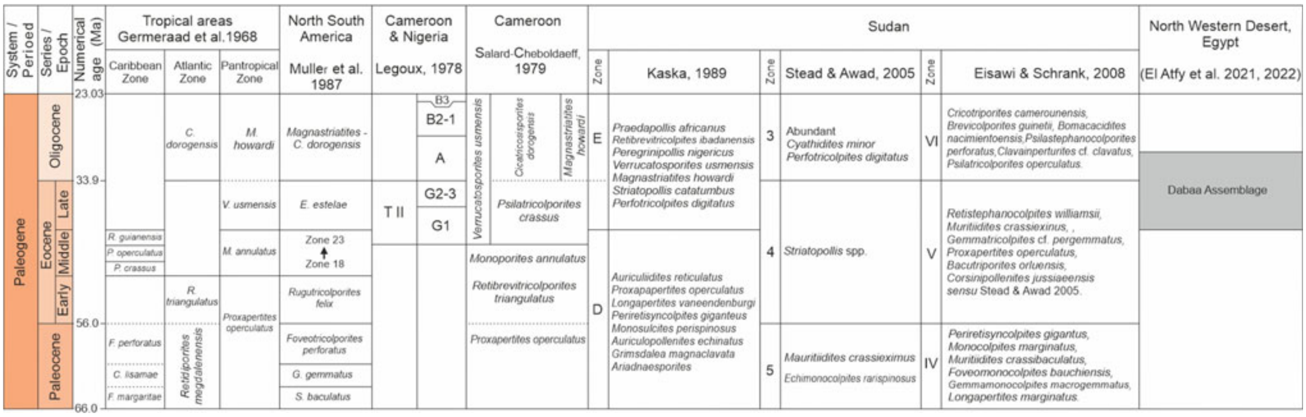


Fig. 6 Correlation of the sporomorph assemblage identified in this study with other zonations elsewhere, mainly from Sudan, West Africa, and North–South America

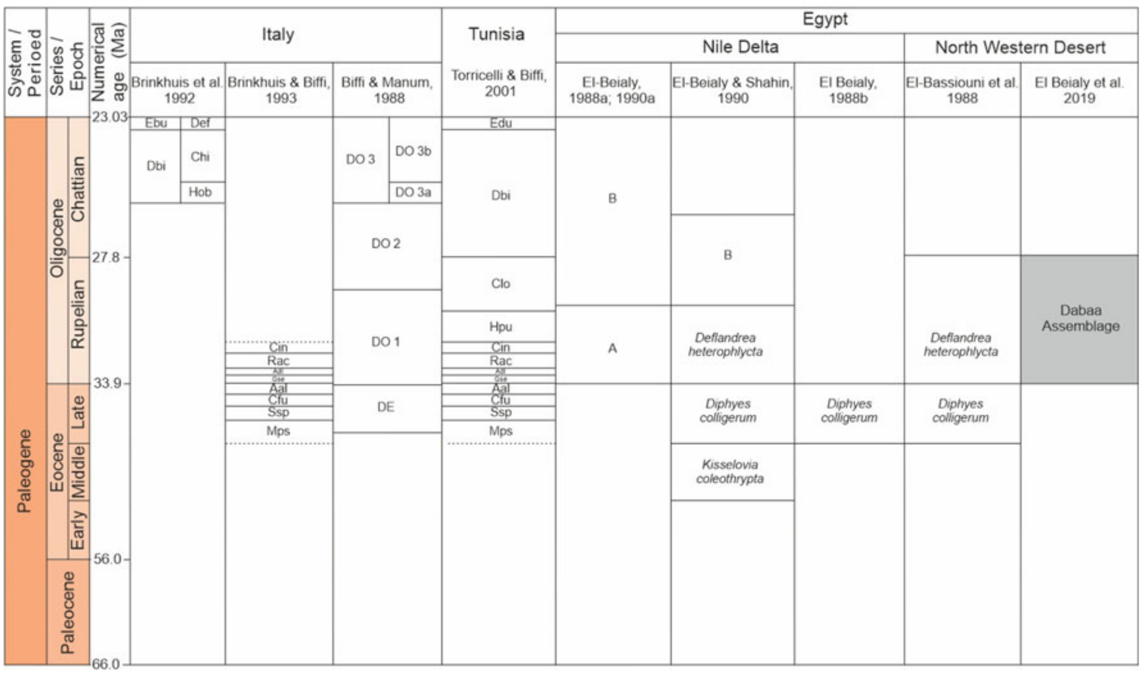


Fig. 7 Dinocyst zonation scheme for the studied Oligocene succession, Western Desert, Egypt. Correlation with previously published local and regional dinocyst zonations is shown

Eisawi & Schrank, 2008). In Egypt, *C. dorogensis* has been reported from the Rupelian of the Western Desert (El Atfy et al., 2022) and the early Miocene of the Gulf of Suez (El Atfy et al., 2017).

Magnastriatites howardi (Fig. 3b) is generally found in Eocene to Oligocene rocks in many African countries including Tunisia (Torricelli & Biffi, 2001), Sudan (Kaska, 1989; Stead & Awad, 2005), and Cameroon (Salard-Cheboldaeff, 1979), despite the early Miocene record by Eisawi and Schrank (2008) from Sudan. In Egypt, *M. howardi* has been reported from the late Eocene of the Saccara area (Kedves, 1986; as *Cicatricosisporites grandiosus*), the Rupelian of the Western Desert (El Atfy et al., 2021, 2022) as well as the early Miocene of the Gulf of Suez (Ahmed & Pocknall, 1994; El Atfy et al., 2013) and Sinai (Wescott et al., 2000).

Verrucatosporites usmensis (Fig. 3c) is generally a long ranging (Eocene–Pliocene) species in Africa (e.g., Salard-Cheboldaeff, 1979, 1990; Kaska, 1989; Stead & Awad, 2005; Eisawi & Schrank, 2008). In Egypt, however, it was recorded from the Burdigalian of Sinai (Wescott et al., 2000), early Miocene of the Gulf of Suez (El Atfy et al., 2013), and the Rupelian of the Western Desert (El Atfy et al., 2021, 2022).

Mauritiidites crassiexinus (Fig. 3l) was previously recorded from Paleocene to Eocene strata of Nigeria and

Sudan (Jan du Chêne et al., 1978; Adegoke et al., 1978; Stead & Awad, 2005; Eisawi & Schrank, 2008). In Egypt, it was reported by El Atfy et al. (2021) from the Rupelian of the Qattara area.

Echiperiporites estelae (Fig. 3i) was recovered from Oligocene to Miocene rocks from Tunisia (Torricelli & Biffi, 2001), Sudan (Eisawi & Schrank, 2008), and Cameroon (Salard-Cheboldaeff, 1979). In Egypt, *E. estelae* has a rather longer age range (Eocene–Miocene) as reported by El-Beialy and Shahin (1990), El Beialy et al. (2005) and El Atfy et al. (2021).

The first record of *Magnaperiporites spinosus* from Egypt was reported by El Atfy et al. (2022) from the Rupelian of the Dabaa Formation. Its actual age range in Egypt is yet to be determined through further investigation.

Perfotricolpites digitatus (Fig. 3q) was previously reported from the upper Eocene to Miocene strata from Nigeria and Sudan (Germaraad et al., 1968; Kaska, 1989; Stead & Awad, 2005). In Egypt, it was recorded from the Burdigalian of Sinai, (Wescott et al., 2000), early Miocene of the Gulf of Suez (El Atfy et al., 2017), as well as the Rupelian of the Western Desert (El Atfy et al., 2021, 2022).

Praedapollis africanus has been repeatedly documented from the late Oligocene to early Miocene of Cameroon, (Salard-Cheboldaeff, 1978), Tunisia, (Torricelli & Biffi, 2001) and the Sudan (Eisawi & Schrank, 2008). In Egypt, it was recorded from the early Miocene of the Gulf of Suez (El Atfy et al., 2013) and the Rupelian of the Western Desert (El Atfy et al., 2021, 2022).

Striatopollis catatumbus (Fig. 3s) was previously recorded from the upper Eocene to Oligocene of Sudan (Kaska, 1989; Stead & Awad, 2005) as well as the lower Eocene of Nigeria (Takahashi & Jux, 1989b). In Egypt, it was recorded from the Rupelian of the Western Desert as well as the Burdigalian of the Gulf of Suez (El Atfy et al., 2021, 2022; El Beialy et al., 2005).

Zonocostites ramonae (Fig. 3p) was recorded in Egypt by El Atfy et al. (2022) from the Rupelian of the north Western Desert.

It is worth noting that *Cicatricosisporites dorogensis*, *Magnaperiporites spinosus*, *Magnastriatites howardi*, *Peregrinipollis nigericus* (Fig. 3r), *Perfotricolpites digitatus*, *Praedapollis africanus*, *Retibrevitricolporites ibadanensis* (Fig. 3n), *Striatopollis catatumbus,* and *Verrucatosporites usmensis* have an Eocene first occurrence that extended to the Oligocene in West Africa (Salard-Cheboldaeff, 1979, 1990). These taxa correlate (Fig. 6) well with the late Eocene to Oligocene Zone E of Kaska (1989) and the Oligocene-early Miocene Zone VI-VII of Eisawi and Schrank (2008) from Sudan. The stratigraphic ranges of the concurrent species *Cicatricosisporites dorogensis* (Zones 22–25, middle Eocene to Oligocene)**,** indicate that the spore/pollen assemblage from the Dabaa Formation within the Western Desert can be allocated to the *Magnastriatites-Cicatricosisporites dorogensis* Zone (Zone 25, Oligocene) of Muller et al. (1987). Alike postulation was established for material from Venezuela (Helenes & Cabrera, 2003) and Zone F (Oligocene-early Miocene) *Verrucatosporites usmensis-Magnastriatites howardi* (Ikegwuonu et al., 2020) from Nigeria. On the other hand, the dinocyst evidence from the exposed Dabaa Formation in the Qattara Depression, north Western Desert (El Beialy et al., 2019) assigns it to Rupelian.

From the above discussion, one can conclude that the age of the Dabaa Formation is postulated to be late Eocene to early Oligocene based on combined dinoflagellate and miospore evidence from different locations within the Qattara Depression (El Beialy et al., 2019; El Atfy et al., 2021).

8 Conclusions

A review of the palynology of the Oligocene Dabaa Formation in the north Western Desert, based on data gathered from the Qattara surface sections and the Amana-1X well (Abu El Gharadig Basin), shows the existence of diverse palynomorph assemblages including dinoflagellate cysts, spores, pollen, acritarchs, fungal spores, and foraminiferal test linings. The palynostratigraphic and palaeofloristic significance of these assemblages have been discussed highlighting the following points as especially important:

1. The early Oligocene vegetation was composed of several zones that extended approximately parallel to the Tethys' coastline (Fig. 4). These include the mangroves which expanded greatly, in addition to marshes, lakes, swamps, streams, and rivers. Tropical forests probably developed in mountainous areas.
2. The Oligocene was marked by predominantly wet climatic conditions with fluctuating moisture regimes. A tropical coastal plain with wet soil and seasonal rainfall developed in the northern Western Desert which led to considerable floral abundance and diversity.
3. The dating of the Dabaa Formation is based mainly on diagnostic dinoflagellate cysts including *Phthanoperidinium comatum*, *Rhombodinium draco*, *Lentinia serrata*, *Diphyes colligerum*, *Deflandrea phosphoritica*, *Distatodinium paradoxum*, *D. craterum,* and *Melitasphaeridium pseudorecurvatum*, which is also supplemented with some age-diagnostic sporomorphs.
4. Dinoflagellate cysts and sporomorphs are numerous, morphologically diverse, and well preserved in most of the preparations retrieved from the Oligocene Dabaa Formation in the north Western Desert. The general aspect of the preparations is consistent with late Eocene to early Oligocene age. Good analogies exist with associations described from the Oligocene of the Sudan,

Tunisia, and Libya demonstrating common sedimentation environments on the northern margin of Gondwana. These conclusions need further verification based on materials from other localities.

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