A new stratigraphic correlation for the upper Campanian phosphorites and associated rocks in Egypt and Jordan

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ABSTRACT

Facies analyses and a sequence stratigraphical framework with regional correlation of the upper Campanian phosphate province are described and interpreted based upon three main sections located in Egypt (Gebel Duwi and Abu Tartur sections) and north Jordan (Umm Qais section).

Fifteen facies types were grouped into: phosphate (FT1–5), carbonate (FT6–11) and siliciclastic (FT12–15) facies associations. The main component of phosphate rocks are pellets *in situ* and common reworked biogenic debris especially in the upper phosphate beds (e.g., fish teeth and bones) with abundant *Thalassinoides* burrows suggests that the skeletal materials are the main source for phosphatized inputs in Egypt, while the common authigenically phosphatic grains (pristine) in Jordan reflects upwelling regime in oxic to suboxic zones.

Based on age assignment as well as stratigraphical position, the phosphorite beds show great similarity that may suggest the similar origin and adjacency during the period of deposition of the Duwi Formation in Red Sea coast of Egypt and its

equivalent the Al-Hisa Phosphorite Formation in Jordan that represents the early transgressive system tract.

On the Abu Tartur Plateau, the presence of sandy pyritic phosphatic grainstone (FT1) and glauconitic quartz arenite (FT12) in the middle part of the studied section, along with the absence of the chert facies (FT14), reflects shallower marine depositional environment with increased fluvial sediment-supply than in those along the Red Sea coast and north Jordan.

Keywords: Egypt, Jordan, phosphorites, Late Campanian, glauconite, black shale

Introduction

The Campanian Duwi and Amman phosphate-rich formations, deposited within shallow marine environments in Egypt and Jordan, have been intensely studied due to their economic interest. These deposits belonged to the giant Tethyan phosphorite belt extending from the Caribbean in the west, through North Africa to the Middle East in the east (Notholt 1980). This province accounts for the greatest known accumulation of marine phosphorites, possibly in excess of 70 billion metric tons of phosphate rocks (Glenn and Arthur 1990).

An increased attention to the Mesozoic and Cenozoic phosphorite deposits of the Middle East coincides with the discoveries of oil shale and rare earth elements associated with the phosphate deposits (El-Kammar et al. 1979; Muhammad et al. 2011; Abd 2011)**.** The Campanian deposits were studied by several authors and described their petrography, geochemistry and phosphogenesis (e.g., Hassan and El Kammar 1975; Glenn and Arthur 1990; Baioumy and Tada 2005) in Egypt, while in Jordan (e.g., Bandel and Mikbel 1985; Abed and Ashour 1987; Abed and Kraishan

1991; Abed and Fakhouri 1996; Abed and Amireh 1999; Abed et al. 2005; Abed et al. 2007; Abed 2011 and 2013, Powell and Moh'd 2011, 2012).

Sedimentation of the phosphorite, chert, oil shale, and associated sediments was a function of upwelling currents influence bioproductivity, as well as the light, relative sea level, and paleobathymetry of the epicontinental shelf floor (Kolodny and Garrison 1994; Abed et al. 2005; Powell and Moh'd 2011). The depositional setting caused rapid lateral variations in thickness and facies. The Egyptian and Jordanian phosphates are shallow marine deposits of Late Cretaceous (Campanian to Maastrichtian) age. The maximum phase of phosphorite sedimentation was associated with a transgressive shoreline of the Neo-Tethys Ocean that encroached from north to south over the northern slope of Africa in the Coniacian to Campanian times. The precise correlation of the major upper Cretaceous phosphate provinces in Egypt and Jordan, as well as the comparison between the local and global sequences is still uncertain due to the controversy in age assignments and strong lateral and vertical variation of lithofacies.

The aim of this paper is to present results of investigation on age assignment, lithofacies, biofacies, and the depositional environments to determine the relative sea-level curve of the Campanian deposits in Egypt and Jordan. It provides interesting correlation of phosphorite sequences in the inner-ramp setting from south Egypt and north Jordan, and the timing of causal factors such as global/regional sea-level changes.

Geological setting

In Mesozoic times, Egypt and Jordan were situated at the southern margin of the Neo-Tethys Ocean (Stampfli and Borel 2002). Sedimentation was controlled by the configuration of the Tethys to the north and north-west and the intra-plate tectonics of the Arabo-Nubian shield (Powell 1989; Powell and Moh'd 2011). The closure of Neo-Tethys during the convergence of the African-Arabian Craton caused basins and swells that are mainly related to the major tectonic pulse of the 'Syrian Arc' fold belt during the Late Cretaceous (Krenkel 1924; Bowen and Jux 1987). This is reflected in a series of syn-depositional anticlines and synclines in north Sinai and the Negev; west of the Dead Sea transform (Said 1962; Bartov et al. 1972; Bartov and Steinitz 1977; Garfunkel 1978; Bartov et al. 1980) and in the Palmyrides to the southeast of Syria (Al-Saad et al. 1992; Chaimov et al. 1992; Sawaf et al. 1993).

Upper Campanian phosphate provinces in Egypt and Jordan occur in basins that remained relatively undeformed with a less deformed stable shelf to the south (Fig. 1). Otherwise, the Duwi area, on the northwest margin of the Red Sea, can be considered as tilted faulted blocks that are generally dissected by minor faults, especially those of the northern and southern parts (Akkad and Dardir 1966). A major phosphogenic episode took place during the late Campanian in the Levant and North Africa. The Levant and North Africa was part of the broad, shallow, southern epicontinental shelf of the Tethys Ocean, situated between latitude 10-20°N (Sheldon 1981) and prevailing winds were blowing to the west and southwest onto the southern Neo-Tethys epicontinental shelf. This created an upwelling regime from the deeper Neo-Tethys Ocean onto this shelf (Cook and Mc Elhinny 1979; Sheldon 1987; Almogi-Labin et al. 1993; Kolodny and Garrison 1994; Powell and Moh'd 2011). Since the late Turonian, the Levant epicontinental shelf was shallow marine, with water depth not more than 50 meters, in these basins and swells, Phosphorites and oil shales were deposited across the inner-shelf (Abed and Sadaqah 1998; Abed et al. 2005).

The common association of phosphatic strata with chert and organic-rich sediments in both Middle East and Egypt has been interpreted as an indication that phosphorite accumulation was associated with highly productive surface waters possibly caused by upwelling (Reiss 1988; Shemesh et al. 1988; Almogi- Labin et al. 1990; Almogi-Labin et al. 1993; Kolodny and Garrison 1994; Nathan et al. 1997).

Material and Methods

Three exposed sections were measured and sampled bed by bed and investigated sedimentologically in detail. These are the Abu Tartur (25º25’34”N and 30º05’08”E) and Gebel Duwi sections (26º06’2”N and 34º05’10”E) in Egypt, and the Umm Qais section (32º38’50”N and 35º41’13”E) in Jordan. Carbonate and siliciclastic rocks were analyzed in the field using a hand lens and classified according to their depositional fabrics as well as grain-size and composition. Ninety samples were collected from characteristic microfacies recognized in the field in order to fully characterize these by thin section analyses. In addition, the most important calcareous nannofossil and planktonic foraminiferal taxa are discussed to give an improved biostratigraphical correlation between Egypt and Jordan.

Lithostratigraphy

The Duwi Formation (Youssef 1957) is represented by an economically-valuable phosphate belt concentrated mainly in southern Egypt. This phosphate belt occurs outside north Latitude 26º00´´due to the facies changes and productivity of carbonate, which are represented by Gebel Thelmet Formation or the lower part of the Sudr Formation. It represents high rates of carbonate sedimentation with an observed Late Cretaceous sea-level rise reported across many areas of the Arabian and African plates such as northern Egypt, Israel and Jordan. It unconformably overlies the fluvial shale sequence of the lower-middle? Campanian Quseir Varigated Shales and underlies unconformably the lower Maastrichtian - Upper Paleocene Dakhla Formation.

A marked regional and local variation in the lithofacies of the Duwi Formation between the Red Sea, Nile Valley and Western Desert areas is noticed.

In Kharga Oasis as well as in Nile Valley, the grater thickness of the economic phosphorites occurred in the lower part of the formation, reaching about 6 m at the Abu Tartur mine and decreasing outwards to the El-Kharga and north El-Dakhla oases. Large-scale cross bedding is recorded in siliciclastics phosphate beds on Abu Tartur plateau reflecting the deposition in intertidal marine conditions. This phosphate bed is followed by sub-oxic black shale, and glauconitic sandstones overlain by shale intercalated with less economic phosphate beds (Fig. 2A). The base of these phosphate beds is marked by an intensive burrow system of *Thalassinoides* with siliceous granules and pebbles.

The Gebel Duwi section consists of shale with phosphate, followed by oyster limestone and marl intercalated with (0.5-1.0m thick) phosphatic beds with a concentrated *Thalassinoides* burrow system at the Duwi / Dakhla formational boundary (Fig. 2B).

Age assignments: The precise age of the Duwi Formation is poorly known (Baioumy and Tada 2005). Previous studies reached two different age attributions: late Campanian (Issawi 1972; Abd El-Razik 1979; Kassab and Hamma 1991; Cherif and Ismail 1991; Tantawy et al. 2001; Ismail 2012; El-Azabi and Farouk 2012) and late Campanian –Maastrichtian (Glenn and Arthur 1990; Schrank and Perch-Nielsen 1985; El Beialy 1995).

The ammonites *Bostrychoceras polyplocum*, recorded from the upper phosphate beds directly below the Dakhla Formation, and *Libycoceras ismaeli* from the lowermost part of the Duwi Formation by Kassab and Hamma (1991) in Gebel Abu Had, confirm a late Campanian age for the Duwi Formation. This was also observed by Reiss (1955) and El Naggar (1966) who correlated the ammonite *Bostrychoceras polyplocum* with the late Campanian planktonic foraminiferal zone *Globotruncanita calcarata*, which confirms that the Duwi Formation is late Campanian. Cherif and Ismail (1991) and Ismail (2012) mentioned that in the Esh El Mallaha area the Duwi Formation is devoid of index planktonic foraminifera and its age is uncertain, but it could be Campanian as it is overlain by chalk yielding *Globotruncanita calcarata* of Late Campanian age.

In the present study, the flooding of planktonic foraminifera and calcareous nannofossilis from the lowermost beds of the Dakhla Formation indicates an early Maastrichtian age. According to Li et al. (1999), the first occurrence of *Rugoglobigerina hexacamerata* is considered a marker bioevent for the Campanian/Maastrichtian boundary. Well-preserved, abundant and diverse assemblages include *Rugoglobigerina rugosa* and *R. hexacamerata*, *Globotruncana bulloides*, *G. fornicata*, *G. linneiana*, *G. aegyptica* and *G. hevanesis*. These assemblages indicate the earliest Maastrichtian zone *Rugoglobigerina hexacamerata* CF8b, as suggested by the absence of *Gansserina gansseri* and presence of *R. hexacamerata*. The calcareous nannofossil assemblages are represented by *Aspidolithus parcus*, *Quadrum gothicus*, *Q. sissinghii*, *Q. trifidum*, *Tranolithus phacelosus*, *Reinhardtites levis*, *Eiffelithus turriseiffelii*, *Rhagodiscus angustus*, *Arkhangelskiella Cymbiformis*, *Micula decussata* and *Prediscosphaera cretacea*. These assemblages indicate zone CC23b, which spans the earliest Maastrichtian.

Amman Formation (Parker 1970)

The Amman Formation in Jordan is divided into two units, the Amman Silicified Limestone Formation, and the overlying Al-Hisa Phosphorite Formation (El-Hiyari 1985; Powell 1989; Powell and Moh'd 2011) (Fig. 2C).

Amman Silicified Limestone Formation (Masri 1963)

Description: This formation is well-exposed in the study area within deep wadis where the lower part forms steep slopes, occasionally with undulating beds.

The **Amman Silicified Limestone Formation** is characterized by a major unconformity at the base of the underlying Wadi Umm Ghudran Formation of Late Coniacian and the absence of Santonian and early Campanian deposits (Al-Rifaiy et al. 1993); the upper boundary is marked by the increase of thick-bedded chert that unconformably underlies the Al-Hisa Phosphorite Formation. The formation has a thickness of about 94 m at the Umm Qais section, and can be divided into the following lithologic units; from base to top:

Unit 1 consists of 20 m of thick greyish limestone, with bivalves and gastropods intercalated with thinly discontinuous chert interbedded with chert concretions. A few beds of phosphatic limestone can be seen, especially at base.

Unit 2 is ca. 22 m thick and composed of phosphatic argillaceous limestone intercalated with thick bedded dark brown pen contemporaneous chert ranging in thickness from one to two meters. The phosphate particles are derived mainly from bone fragments and larger intraclasts. This interval is marked by *Baculites* ammonites and the large size of *Pycnodonte* (*Ph.*) *vesicularis* (Fig. 2D).

Age assignments: Larger-sized *Pycnodonte* (*Ph.*) *vesicularis* and *Baculites* cf. *ovatus* are considered the most characteristic megafossils of the Campanian in Jordan.

Haggart (2000) assigned Amman Silicified Limestone FormationFormation to the upper Campanian based on the occurrence of *Baculites* cf. *ovatus*, which supports Wetzel and Morton's (1959) assignment of the Amman Silicified Limestone Formation to the Campanian. Using calcareous nannofossils, all the chalk bearing *P.* (*Ph.*) *vesicularis* is preferably placed in the late Campanian Zones CC21 and CC22 (Faris and Abu Shama 2006). On the other hand, Ismail (2012) reported the disappearance of the *Pycnodonte vesicularis* below the marker index planktonic foraminifera of *Globotruncanita calcarata* Zone. The Amman Silicified Limestone Formation yielded rare and sporadic planktonic foraminifera such as: *Rugoglobigerina rugosa*, *Globotruncana arca*, *Globotruncanita calcarata*, *G. fornicata*, *Heterohelix globulosa*. The calcareous nannofossil assemblages are represented by *Watznaueria barnesae, Micula decussate Micula concava Reinhardtites anthophorus, Arkhangelskiella cymbiformis, Eiffellithus gorkae, E. turriseiffelii, Gartnerago obliquum, Prediscosphaera cretacea, Quadrum sissinghii*, *Ahmuellerella octoradiata, Calculites obscurus* (Plate 1). These assemblages are assigned to the CC222 zone of Perch-Nielsen (1985). Al-Rifaiy et al. (1993), using the presence of *Rugoglobigerina rugosa,* suggests an age not older than the middle Campanian for the base of the Amman Silicified Limestone Formation.

Al-Hisa Phosphorite Formation (El-Hiyari 1985)

Description: This formation crops out as discontinuous thin strips along some wadis in the study area. The Al-Hisa Phosphorite Formation consists of three units: the 25 m thick main phosphorite, and 2 m massive limestone and the upper phosphorite units. The phosphate consists of massive granular, sand-grade phosphate embedded in calcareous cement characterized by a lack of any sedimentary structures. The Al-Hisa

Phosphorite Formation is overlain conformably the Muwaqqar Chalk Marl Formation (Masri 1963), which is composed mainly of chalky limestone and marl (Fig. 2C).

Age assignments: Previous studies of the Al-Hisa Phosphorite Formation reached two significant differences in age assignments due to sparse biostratigraphic data. Previous authors assigned late Campanian age, such as (Powell 1989; Powell and Moh'd 2011); Early Middle Maastrichtian age (Abed and Ashour 1987; Abed 2011) and Campanian –Maastrichtian age (Abed et al. 2005 and Abed et al. 2007). Abed and Ashour (1987) gave a Maastrichtian age for the Al-Hisa Phosphorite Formation based on the presence of the planktonic foraminifera *Gansserina gansseri* in the overlying sediment above the phosphate deposits. Different age determinations are obtained in this current study, notably the first appearance of *Gansserina gansseri* is higher in the Muwaqqar Chalk Marl Formation. In addition, the presence of calcareous nannofossil reliable bioevents of *Reinhardtites anthophorus* and *Eiffellithus eximius* indicates a Late Campanian for the Al-Hisa Phosphorite Formation in Jordan.

Lithofacies and microfacies

The dominance of a particular or group of lithofacies, in different locations allows the definition of a phosphate association (FT1–5)**,** carbonate association (FT6-11)**,** and siliciclastic association (FT12-15) as outlined below.

Phosphate associations (FT1–5)

Sandy pyritic phosphatic grainstone (FT1): This lithofacies is recorded only in the main phosphatic rocks at Abu Tartur section. It is composed of large-scale cross bedding in two opposite directions, massive, sandy, moderately sorted, coarse-grained, ferruginous, granular phosphorite intercalated with thin shale laminas. In thin section, phosphatic pellets and bioclasts compose about 20-35% of the rocks and sized from 0.2-2mm, while the lithoclasts up to 3mm and their size is decreased upward. These phosphatic grains are of subrounded to well-rounded and oval to spherical shape. Phosphatic bioclastics are composed of vertebrate fragments (bones and teeth). The carbonate hydroxyapatite mineral (dahlite) of these biogenic fragments is partially transformed into silica minerals (especially chalcedony). Non-phosphate grains consist essentially of pyrite, detrital quartz, gypsum and hematite. Quartz grains are characterized by rounded to angular, elongated and spherical shape and sized in 0.1-0.2mm is well detected. The percent of quartz grains is decreased upward. Pyrite crystals are dispersed within the cement in dark gray to black color in the fresh phosphatic samples. The yellowish brown weathered phosphatic samples are cemented with iron oxide and gypsum which formed by the oxidation of pyrite (Fig 2E).

Interpretation: The abundant detrital quartz, the presence of marine vertebrates, the large-scale cross bedding suggest near-shore depositional environments with high a gaited and storm sediment deposited as lag deposits. The large amounts of pyrite in the Abu Tartur phosphorites suggests an increasingly abundant source of iron-bound phosphate associated with terrigenous sediment input to the continental margin. Phosphatic *Thalassinoides* reflect storm, high energy conditions associated by reworking and upwelling of phosphatic deposits. Avery common depositional environment of *Thalassinoides* include not only shelves and epeiric embayments but also littoral to sub-littoral parts of certain estuaries, bays, lagoons, and tidal flats (Miller 2007).

Pufahl et al. (2003) mentioned that *Thalassinoides* burrow networks are classic examples of condensed beds that can also be interpreted as firmgrounds associated with increasing accommodation during transgression, extensive high energy conditions developed with elevated reworking intensity and upwelling, these phosphate deposits were transported and accumulated in the intercontinental shallow basins. The association of sands and phosphorite reflects the fluvial supply that may have been, according to Glenn and Arthur (1990), the major source of phosphorus for the formation of these phosphorites. This phosphorite facies interpretated by Abed and Amireh (1999) as near-shore deposits and the lack of clay matrix are taken to indicate reworking and winnowing of the phosphorite deposits and redeposition with the sand.

Calcareous bioclastic phosphatic grainstone (FT2): This lithofacies have a wide distribution along the lowermost part of the economic phosphatic beds in and Umm Qais section, while it occurs in the upper part of Gebel Duwi section ranging in thickness from 2 to 5 m. It is composed of yellowish brown to light grey, massive, moderately sorted, coarse-grained, granular phosphorite beds characterized by lack of any sedimentary structures.

Microscopically, it is made up of bioclasts, phosphatic pellets as well as lithoclasts. The bioclastic grains are represented by vertebrate fragments (bones and teeth, up to 25% of the rock) of different sizes (0.2-2mm) with well-preserved of their original structure. The replacement of phosphatic mineral (dahlite) by calcite can be noted in bone fragments. Partial decomposition of bone-forming minerals by bacteria is well developed. Phosphatic pellets and lithoclasts compose about (40-60% of the rock) with different size and type. Granular phosphatic pellets are sized from 0.3-1mm with rounded to subrounded, spherical and oval shape. In situ phosphatic pellets (pristine) are characterized by their irregular outer relief. The lithoclasts (e.g. composite phosphatic pellets) can be recorded with their large grain size (up to 3mm) and shape and they mostly contain quartz and shell fragments of preexisting rocks (Fig. 2F). In general, the cement material in these phosphorites is macrocrystalline spary calcite.

Interpretation: The main characteristic feature of this microfacies is the accumulations of small, grain-supported, subrounded and subangular peloids. This microfacies is represented by peloidal grains of different size and type (pristine and reworked), forming irregularly distributed grainstone fabrics. These sediments are deposited in suboxic, protected-shallow basins with moderate water circulation. The peloidal phosphates were formed authigenically in oxic to suboxic zones, from phosphate-rich sediments, followed by storm wave winnowing and storm-generated, shelf-parallel geostrophic currents and minor compaction (Glenn et al. 1994; Pufahl *et al*. 2003; Brookfield *et al*. 2009). The lack of tidally generated sedimentary structures in other sections indicates a low tidal range, suggesting that tidal currents played a minimal role in transporting and sediments redistribution (Pufahl et al. 2003).

Biopeloidal dolomitic grainstone (FT3): This lithofacies exist commonly as intercalated, relatively thin (20-30 cm), beds within the black shale beds. This facies occurs in the lower part of both Gebel Duwi and Abu Tartur sections with sharp contacts with their surrounding strata. It is consists of light cream to yellowish brown, medium to coarse sand–size and well to moderately sorted phosphatic grainstone.

Petrographically, it is composed of bioclasts (bone and teeth, algae and oyster shell fragments) and phosphatic pellets. Trace algal fragments are sized from 0.3 to 6mm of different forms and they act as traps of other phosphatic grains. Dolomite cement is dominated and generally of medium to coarse, polymodal, euhedral planar crystalline nature. Pyrite can be noted as fine-grained dark cubes within and/or around the phosphatic grains. In addition to dolomite, carbonate and gypsum cement have been observed. Well-developed dolomite crystals indicating partial dolomitzation of calcite mineral (Fig. 3A).

Interpretation: These sediments were deposited in protected, intertidal to shallow subtidal, slightly reducing environments with moderate water circulation. Also as indicated by the fossil assemblage of algae and bivalves this microfacies was deposited under moderate-energy conditions in lagoonal or protected shelf environments. During storms, the epeiric platform interior tidal flats would be flooded and much shallow subtidal sediment deposited upon them. In the subtidal, skeletal debris would be transported and sorted during storms and post-storm surges, and deposited to give grainstone beds (Tuker et al. 1990).

The enrichment of dolomite reflects relative saline conditions. As a diagenetic product, dolomite formed from high concentrated Mg^{+2} solutions and also from solutions with less salinity such as that associated with mixing between a brackish fluid and seawater. The Mg^{+2} may be leached from the overlying shales by the meteoric water and penetrated into the phosphatic deposits where the dolomite is crystallized. Rifai and Shaaban (2007) mentioned that this dolomite cement was formed from mixed hypo-saline fluids within a mixing marine-meteoric zone in association with organic matter degradation.

Biopeloidal dolomitic rudstone (FT4): It is non-economic dolomitic phosphatic beds and made up of coarse lithoclasts of gravely-size flat pebble conglomerate (>3mm) consisting of partially dolomitized, laminated and exhibiting desiccation cracks. This lithofacies has been recorded in the uppermost phosphatic bed of Umm Qais and Abu Tartur sections, and also in the intercalated phosphatic beds in the middle part of Gebel Duwi section.

It has an average thickness of about 10cm and made up of massive, gypsiferous, ill-sorted phosphatic grains with brownish red color due to the presence of iron oxides.

The grain size is decreased vertically. Few phosphorite beds with a thickness less than 20 cm were recorded in the Amman Silicified Limestone Formation in Jordan; they are composed mainly of reworked phosphatic bones and lithoclasts derived from pre-exit phosphate, and differ completely from the overlying phosphates of the Al-Hisa Phosphorite Formation.

Microscopically, the phosphatic pellets are made up of subrounded to subangular, oval and lath shape and they are mainly of composite type and contain small phosphatic pellets. Few of these lithoclasts phosphatic pellets are fractured that may due to the compaction diagenetic process. Phosphatic grains are cemented together by well crystalline fibrous gypsum and dolomite (Fig. 3B).

Interpretation: Enrichment of ill-sorted gravely-sized phosphatic lithoclasts reflects high a gaited and storm sedimentation environment. These grains were deposited as lag deposits in tidal channels and on tidal flats. The corresponding carbonate-rudstone facies with coarse lithoclasts and bioclasts are considered as lag deposit in tidal channels (Tuker et al. 1990). Dolomite cement was formed from mixed hypo-saline fluids within a mixing marine-meteoric zone.

Glauconitic phosphatic grainstone/rudstone (FT5): This lithofacies occurs as intercalated thin beds (10-20 cm) within the well-developed glauconitic sandstones in the upper part of Abu Tartur section. It made up of pale brown, moderately sorted, coarse to very coarse sand-size, semi-hard to friable glauconitic phosphorite. Very thin gypsum vienlets and hematite patches have been noted.

Microscopically, this microfacies consists of well-developed phosphatic grains up to 80% and sized from 0.4 up to 3mm, with the dominance of the 0.6 mm grain size. Phosphatized organic remains are common as well as glauconite pyrite microcrystals and minor amount of quartz grains (Fig. 3C). Authigenic glauconite as well as the dispersed glauconitic pellets is well developed. All grains are cemented by gypsum. The original structure of bones is well observed.

Interpretation: The combination of grainstone / rudstone facies can reflect storm, lag / intertidal to shallow subtidal depositional environment with Fe-enrichment. These conditions are followed by post-depositional oxidation stage forming the glauconite cement. The oxidation of Fe-rich minerals (especially pyrite) played a considerable role in the formation of authigenic glauconite, hematite and gypsum. The formation of authigenic glauconite occurs mainly during diagenesis of sediments by synthesis from interstitial solutions and/or alteration of clay minerals (Logvinenko 1982). Glauconite cement develops in Fe-rich, oxidized shallow marine environments (Odin and Matter 1981). The unusual glauconite phosphate assemblage is typical of suboxic zones and may be related to sulphide oxidation (Coleman 1985). Removal of fine-grained sediment by strong winnowing occurs in high-energy near-coast environments (Flügel, 2004).

Carbonate associations (FT6-11)

The carbonate associations are recorded mainly in the Umm Qais section, Gebel Duwi section, and completely absent in Abu Tartur plateau (Fig. 4).The observed facies types suggest a depositional environment ranging from deep-subtidal to intertidal, reflecting minor fluctuations in relative sea-level from tidal flat to deep subtidal environments.

Lime-mudstone (FT6): The lime mudstone lithofacies is well represented in the Duwi Formation, at Gebel Duwi and Amman Silicified Limestone Formation. It is brown, well indurated, thick-bedded and argillaceous intercalated with bedded chert (Fig. 3D). Microscopically, it consists of a dense lime mud matrix which may exhibit a clotted texture due to the patchy occurrence of microsparite. Few foraminiferal tests and corroded fine sands may occur. In thin section it consists of fine dense and dark grey microcrystalline calcite. Aggrading neomorphism into microsparry calcite is observed.

Interpretation: The lime-mudstone was deposited in a restricted lower intertidal regime shoreward of a quiet water lagoon on the marine shelf due to the absence of variable amount of skeletal particles. Most of the lime mud may have been derived from the abrasion and micritization of skeletal particles, especially in the restricted areas.

Sandy foraminiferal packstone (FT7): This lithofacies is recorded at the lower part of the Amman Silicified Limestone Formation and middle part of the Al-Hisa Phosphorite Formation with a thickness of about 3.0 m and 5.0 m respectively. It is composed of yellowish gray argillaceous limestone, moderately hard, massive with some chert nodules and bands containing highly fragments of bivalves. In thin section, the lithofacies is made essentially of low diversity and badly preserved planktonic foraminiferal tests of *Globotruncana* sp. and *Heterohelix* sp., well sorted thin-shelled pelagic bivalve and few corroded sub-angular quartz grain. The quartz grains are fine to medium size, moderately sorted display a floating grain texture. All components are closely packed in a dark dense lime–mud matrix. Some of these tests are susceptible to change to microspare by aggrading neomorphism while their internal structure is still well preserved.

Interpretation: The sandy foraminiferal packstone (FT13) suggests a restricted lower intertidal regime shoreward of a quiet water lagoon below normal wave base. The presences of low–diversity planktonic foraminiferal content with broken bivalves and sand grains may be attributed to wave transportation, while the sand content may be attributed to episodes of high wind blown sands.

Ammonite/oyster bearing planktonic foraminiferal packstone (FT8): This lithofacies is recorded in the middle part of the Amman Silicified Limestone Formationwith thickness of about 20 m. The rock is yellowish brown, well indurated, thick-bedded and fossiliferous with *Baculites* ammonites and the large size of *Pycnodonte vesicularis*. Microscopically, it is composed of foraminifera (40-60%), yellowish pellets of collophne (10-20%), lithoclastic, large teeth and bone fragments (0.4-2mm) that scattered randomly. Few *in-situ* phosphatic pellets can be recorded of up to 3mm. All components are embedded in cryptocrystalline calcitic matrix (Fig.3E).

Interpretation: The occurrence of baculitid ammonite bearing planktonic foraminiferal packestone interbedded with cherts and micritic limestones suggest open-marine, deep subtidal environments (i.e., distal outer ramp) with low energy, below storm wave-base.

Foraminiferal wackestone (FT9): This facies marks the lower part of the Muwaqqar Chalk Marl Formation. It is composed of massive marly chalk. Well preserved foraminiferal assemblages with medium-diversity are the characteristic fauna; forming about 30% of the rock. They are loosely packed in a partially recrystallized lime mud with rare phosphate pellets.

Interpretation: The increase of foraminiferal assemblages, total absence of shallow-water indicators and absence of bioturbation all point to deposition in a low-energy, shallow middle shelf regime under rapid sediment accumulation. Similar deposits in north Egypt, Libya, Negev, and Syria indicated on major transgressive phase with increase of carbonate productivity during the Masstrichtian age (e.g. Soudry et al. 2006; El-Azabi and Farouk 2011).

Foraminiferal phosphatic packstone (FT10): This lithofacies is recorded from the upper part of the Duwi Formation at Gebel Duwi with a thickness of about 5m. It is yellowish gray, massive, argillaceous limestone with worm tubes. Low diversity of biserial and non-keeled, biserial and trochospiral planktonic foraminifera, with the presence of some benthonic foraminifera and bone fragments are important grains in this microfacies. All components are embedded in cryptocrystalline calcitic matrix

Interpretation: The assemblage of fossil types that comprise this facies is suggestive of a shallow-water marine environment. The low diversity assemblages of planktonic and benthic foraminiferal assemblages with bone fragments and shark teeth suggest that this facies was derived from more proximal shelf settings and transported offshore during storms.

Oyster pelloidal rudstone/framestone (FT11): This lithofacies is recorded in the Duwi Formation at Gebel Duwi and Amman Silicified Limestone Formation. It has yellowish gray to creamy color, hard and bedded with varies in thickness from 0.5 to 20 m. The maximum thickness (20m) of this oyster facies is recorded at Gebel Hamadat, south of Gebel Duwi). The thickness of this facies is decreased northward (Safaga area) and changed laterally into calcareous mudstones. This limestone facies is characterized by large sized bioclasts in essentially micritic matrix. The bioclasts include large sized shell fragments especially pelecypods debris, which most probably were leached by solutions and the cavities had filled by well-formed mosaic of drusy spary calcite. Low diversity echinoids and gastropods fauna are recorded in this facies. Oyster shells (0.1 to >3 mm) partially showing their internal foliated microstructure packed in coarse drusy and granular sparite (Fig. 3F). The rock is highly phosphatic with many reworked bone fragments and subangular to rounded phosphatic pellets.

These facies are recorded in Upper Cretaceous stratigraphic sections (Ruseifa and northern portion of the Al Abiad/Alhisa district) and reorganized into banks and isolated bioherms (Pufahl et al. 2003). Oyster banks in Jordan consist of a basal bed of fragmented oyster rudstone overlain by a set of megascrossbedded oyster rudstone that is truncated at its top by a bed of chalk-rich fragmented oyster rudstone.

Interpretation: The oyster buildups are similar to other Upper Cretaceous and Cenozoic oyster buildups that dominate (brackish/hypersaline) highly productive tropical marine environments (Pufahl et al. 2003). The presence of oysters shows shallow-marine influences represent a submarine carbonate skeletal shoal developed in a restricted shallow subtidal regime. Deposition took place behind barriers of carbonate skeletal shoals. Sea surface waters were generally enriched with respect to $CaCO_3$. The low faunal diversity of bioclastic facies, the highly broken shell fragments, and their parallel-to-bedding geometry, may indicate reworking of adjacent oyster buildups and deposition in shallow subtidal environments (Abed and Amireh 1999; Pwell and Moh'd 2011). The enormous size, limited species diversity, and rapid community growth observed within banks and bioherms is attributable to reduced competition for space and nutrients (Glenn and Arthur, 1990). Pufahl et al (2003) suggested that these oyster banks developed in more distal environments and prograded landward during continued sea level rise through the attack of onshore-directed storm and fair-weather waves. The progradation of oyster buildups developed behind the distal banks in shallower environments.

Siliciclastic associations (FT12-15)

This facies association is characterized by glauconite, quartz, chert and shale as described below.

Glauconitic quartz arenite (FT12): This lithofacies is recorded only at the Abu Tartur mine and is missing in other sections with thickness 8 m (Fig. 2A). It is light yellowish-greenish, medium-grained glauconitic sandstone contains gypsum-filled fractures. Terrigenous-allogenic and authigenic glauconite are the main component in this microfacies. Glauconitic pellets are well developed forming about 40% of the rock. The main glauconite morphologies are ovoidal, spheroidal or lobate. They are sized from 0.1 to 1mm with green to brown color. Fine to very fine sand are abundant. The constituents are embedded in clay matrix.

Interpretation: The majority of grains in this facies types are predominantly glauconite grains with terrigenous supply represented by fine grained quartz. The morphology and the mineralogy and sedimentary environment of the glauconites can be interpreted as autochthonous. Glauconite is one of the most sensitive indicators of low sedimentation rates in marine environments within comparatively thin (less than 1- 3m) deposits and was deposited during the remaining high stand systems tract and a lowstand prograding wedge (Glenn and Arthur 1990; Amorosi1997; El-Hassan and Tichy 2000). The abundant detrital quartz, the presence of glauconite, the absences of calcareous fossils, and the association with the carbonate facies suggest near-shore depositional environments with source of rivers sediment-supply input (Glenn and Arthur 1990).

Bioclastic phosphatic quartz arenite (FT13): This lithofacies is noted in the middle part of Gebel Duwi and Abu Tartur sections. It is made up of yellow to pale brown, coarse to very coarse sand-size, moderately to well sorted phosphatic sandstones. The thicknesses of this unite reaches about 1.7 m at Gebel Duwi section and 0.3 m at Abu Tartur section as thin beds intercalated. The phosphatic sandstone facies are friable to massive with no internal structures or bioturbation. Minor glauconite is recorded with oyster shell fragments. Microscopically, bioclastic grains are represented by oyster shell fragments, phosphatic particles and rarely algal debris. Phosphatic particles, up to 8% of the rock, are invoked by reworked bone fragments, pellets and shark teeth (Fig. 3H). Simple phosphatic pellets are subangular to rounded, elongated, oval and spherical shape and sized from 0.2 to 0.7mm and rarely reach more than 3mm. Quartz grains are well sorted and sized from 0.1-0.3mm, monocrystalline, angular to sub rounded, elongated to spherical shape. Oyster fragments are well developed and compose mostly of fibrous calcite in their internal structure. Bone fragments can be recognized with longitudinal shape and sized from 0.1 to 1mm. They are grey to black and compose of apatite minerals, which sometimes replaced by silica minerals mostly chalcedony. All components are embedded in clay and sometimes cryptocrystalline calcitic matrix.

Interpretation: The intensive siliciclastic supply as well as large size oyster shell fragments may reflect the deposition under intertidal conditions. The comparable size of quartz grains and phosphatic particles and minor mudstone matrix can indicate reworking, winnowing of near-shore sediments and redeposition of these facies.

Chert facies (FT14): is recorded in Jordan, Gebel Duwi section, and Nile Valley in Egypt with a complete absence of chert in the Abu Tartur section. Three types of chert are recorded: 1) massive bedded chert; 2), discontinuous chert in the lower Amman Silicified Limestone Formationand 3) nodular chert.

Interpretation: Senonian chert is biogenic and is derived mainly from a low diversity assemblage of diatoms during early diagenesis and proceeded through various stages to produce the distinctive chert textures. These indicate a shallow near-shore environment for chert formation and are associated with upwelling on the epeiric shelf of the Tethys from the north (Abed and Amireh 1999; Pufahl *et al*. 2003; Powell and Moh'd 2012).

Shale FT15: Different types of shale are recorded in Egypt only. Non-marine shales (FT15a) recorded from the main part of the Qussier Formation underling the Duwi Formation. Black shales (FT15b) are recorded in large scale on the Abu Tartur plateau and at Gebel Duwi. No foraminifera have been detected in this rock unit. Sediek and Amer (2001) interpreted the Abu Tartur black shales to have been deposited under reducing conditions in a quiet (low energy), sub-oxic marine setting. According to Glenn and Arthur (1990), the Western Desert shales exhibit lower Corg contents and HI values than the Eastern Desert shales, thus implying stronger terrigenous influence, sediment dilution and perhaps more oxidizing bottom conditions. The foraminiferal shale (FT15c) is recorded in the Dakhla Formation overlying the Duwi Formation of middle shelf environment as indicated by the high diversity of foraminifera (Fig. 4).

Discussion

The petrographical studies revealed that biogenic phosphatic debris (e.g., faecal pellets, fish teeth and bones) are the main source of phosphate minerals in Egypt. Thin sections of the pellets show two types: reworked granular and *in situ* phosphatic grains (pristine). The alternation of intervals of pristine phosphate and granular phosphorite is also recognized in correlative deposits in south-central Jordan (Abed and Sadaqah 1998; Pufahl et al. 2003), Iraq (Al-Bassam et al. 1983), Negev (Nathan et al. 1979; Soudry and Champetier 1983), and Egypt (El-Kammar et al. 1979; Glenn 1990).

Reworked granular phosphatic pellets have been recorded in all the samples of the studied localities and reflect locally mechanical accretion and rearrangement of peloids in a relatively high-energy environment. They are structureless, rounded to subangular, spherical to ovular and nearly homogeneous in their internal structure. In all studied phosphatic lithofacies, peloidal grain-size decreases upwards. In the studied phosphate rocks the size of the pellets is larger in both Umm Quais (Jordan) and Abu Tartur (Egypt) than those along the Red Sea coast. In general, the grain size of phosphatic rocks ranges from fine sand grade to coarse gravel with an average of medium sand-grade.

Authigenic phosphatic grains (pristine) are very common in the Jordanian phosphorites and have undifferentiated form (FT1; Fig.2F). The majority are made up of intensive micropellets that accumulated into larger phosphatic pellets. In general, the phosphatic pellets are more common in the Jordanian phosphorites than in the Egyptian examples.

In contrast to phosphatic pellets, the phosphatic lithoclasts are much more abundant in the Egyptian phosphorites than in the phosphatic rocks of Jordan. The phosphatic lithoclasts are formed by multiple event concentration and associated with sediment transportation. These grains contain detrital and carbonate grains, showing that they are reworked of pre-existing accumulations.

Bone and teeth fragments are dominant in the different investigated phosphorites. Their abundance and size increase in the Abu Tartur phosphate rocks and the majority of these grains are fractured, reflecting a high degree of load-compaction on these sediments. The replacement of apatite minerals in vertebrate skeletal grains is well developed in all the studied phosphorites. The main difference is that the phosphatic minerals in the phosphorites of Abu Tartur are mostly transformed into silica minerals (chalcedony), while in other phosphorites they are often replaced by calcite.

Phosphatized algal clasts are better developed in the Egyptian phosphorites than in Jordanian ones. Some of these algal fragments acted as traps for small phosphatic grains and the majority are stained by iron oxide pigments. Algae played a greater role in the formation of the Egyptian phosphorites and are recorded in different forms. Microbial action is well developed in phosphatic grains (pellets and bones), causing the dissolution and concentration of apatite minerals that is associated with an increase of some elements such as phosphorous and fluorine in Egypt.

The cement material in phosphorites of both the Red Sea coast (Egypt) and Jordan is mainly calcite while, in the Abu Tartur area, it is represented mostly by gypsum and hematite. This may reflect the difference in the chemical characteristics of intragranular fluids during the sedimentation and/or early diagenetic processes.

The main phosphatic beds of the Campanian successions in central and south Jordan lie in the Al-Hisa Phosphorite Formation, intercalated mainly with phosphatic chert, and limestone; the latter including bedded coquina limestone and oyster build-ups (Abed et al, 2007; Powell and Moh'd 2011). Towards south and south-west Jordan there is abundant siliciclastics sedimentation on the inner shelf in the Alia and Batn El Ghul areas, where they are massive with no internal sedimentary structures or bioturbation except at the base of the phosphorite (Abed and Amireh 1999; Powell 1989; Powell and Moh'd 2011).

Along the Red Sea coast of Egypt as well as in Nile Valley, the economically exploited beds occur in the upper part of the formation, the middle part and occasionally in the lower part of the formation. This may indicate that the depositional setting in some basins was more suitable to a rapid accumulation rate of phosphates rather than in other localities. In these two areas the phosphate beds are associated by chert, marls, shales, silicified phosphate and oyster limestone. Baioumy and Tada (2005) showed that the Egyptian phosphorites in the upper horizon in Gebel Duwi are similar in composition to the Jordanian phosphorites. Furthermore, the Late Campanian stratigraphical units in Jordan; Red Sea and Nile Valley are approximately similar to each other. These facies accumulated in protected inner shelf environments where the phosphorite beds were deposited during storm events (Powell and Moh's 2011). Glenn and Arthur (1990) considered that the phosphate lithofacies on the Red Sea coast were deposited inshallow hemipelagic environments.

On the Abu Tartur plateau the main phosphorite bed occurs in the lower part of the formation followed by black shales and glauconitic sandstones (Fig. 2A). The presence of quartz grains in the lower part of phosphatic bed at Abu Tartur reflects fluvial sediment-supply and shallower depositional environment in these basins than in those along Red Sea coast and Umm Qais (Jordan), although siliciclastic lithofacies have been reported close to the palaeo-shoreline south-west Jordan (Powell, 1989). Soudry et al. (2006) mentioned that the Abu Tartur phosphorites associated by terrigenous input (and perhaps some continental P supply) in their formation. On the other hand, the Middle East (e.g., Jordan and Syria) phosphorites with a little or no terrigenous influence were felt in the course of their deposition. This reflects different water masses and palaeogeographical settings at the edges of the southeastern Neo-Tethys during the Campanian. The large amounts of pyrite and glauconite in the Abu Tartur phosphorites suggests an increasingly abundant source of iron-bound phosphate associated with terrigenous sediment input to the continental margin. The Late Campanian Egyptian phosphorites deposits of Abu Tartur may have formed largely from remobilized iron-bound phosphate derived from local rivers (Glenn and Arthur 1990; Glenn et al. 1994).

Enhanced fluvial solution supply of phosphorus sporadically initiates phosphogenesis associated with intensive glauconitization in the presence of Fe, Al, and Mg. These grains are commonly reworked and represent the top lap condensation (Trappe 1998).

Pufahl et al. (2003) mentioned that the bulk sediment composition and absence of Fe-bearing authigenic phases such as glauconite, pyrite (including pyrite molds), siderite, and goethite within Upper Cretaceous (Campanian) pristine phosphorites in Jordan phosphates suggests that deposition and authigenesis occurred under conditions of detrital starvation and that ‘‘iron-pumping’’ played a minimal role in phosphogenesis.

Based on the relative abundance of REE in the Egyptian phosphorites, Hassan and El Kammar (1975) concluded that the phosphorite in both the Red Sea and the Nile Valley are enriched in U content more than the phosphate rocks on the Abu Tartur plateau. On the other hand, the phosphate rocks in Abu Tartur are characterized by a relative abundance of Y, Yb and La more than phosphorite in other localities. It can concluded that the relatively deeper-water phosphate beds in the Red Sea and the Nile Valley are characterized by their low REE content due to their precipitation from sea water. The high content of Y, Yb and La reflects the relatively shallower phosphate deposits of Abu Tartur Plateau than those in the other areas.

SEQUENCE DEVELOPMENT

Facies development of the Late Campanian deposits is very important for understanding the environmental changes and (relative) sea-level variations across the shallow marine shelf in Egypt and Jordan. Systems tracts are packages of strata composed of age-equivalent depositional systems, and correspond to specific stages of shoreline shifts (Brown and Fisher 1977; Posamentier and Vail 1988).

The Campanian phosphate province in Egypt and Jordan can be subdivided into two 3rd-order depositional sequences.

Depositional sequence 1. It is comprises the lower beds of the Duwi Formation and Al-Hisa Phosphorite Formation. The base of the late Campanian phosphatic beds in Egypt and Jordan is marked by an well-known unconformity surface (sequence boundary 1), separating the non-marine Quseir Variegated Shale (FT15a) from the main Phosphate facies (FT1 and FT2) of the Duwi Formation in Egypt and the Amman Silicified Limestone Formation (FT6 and FT14) from the phosphate facies (FT2) of Al-Hisa Phosphorite Formation (Fig. 4). The contact between this phosphorite facies and the underlying rocks is bioturbated and characterized by *Thalassinoides* burrow networks of irregularly inclined to horizontal components and less scattered vertical cylindrical burrows. These *Thalassinoide* burrows are infilled with granular phosphatic grainstone.

A well-marked global marine transgression caused frequent reworking and re-deposition, explaining the huge phosphorite deposits followed by organic carbon-rich shales observed in the eastern Mediterranean and North Africa compared with southern Europe (Glenn *et al*., 1994).

At Eshidiyya area (south of Jordan), the Al-Hisa Phosphorite Formation overlies the Paleozoic strata with a pronounced regional erosional unconformity (Abed et al. 2007). This sequence boundary is primarily related to regional tectonism of the Arabian Plate (Powell, 1989). The studied section at Umm Qais in north Jordan lay in a more basinal setting and seems less affected by tectonism; the equivalent sequence boundary is recorded between strong vertical facies changes at the formational boundary and is coincident with a eustatic sea-level fall between the middle/late Campanian (Haq *et al*. 1987).

Sedimentation of the phosphorite, chert, oil shale, and associated sediments was a function of upwelling currents (bio-productivity and photic zone position in the water column), relative sea level, and paleobathymetry of the epicontinental shelf floor (Kolodny and Garrison 1994; Powell and Moh'd 2011).

The MFS is observed in whole studied sections characterised by a shift vertical facies changes overlain by a pro-gradational highstand facies are observed from the deposition of glauconite intercalated in the higher part with coarse-grained dolomitic phosphate at Abu Tartur plateau, while in Gebel Duwi the equivalent is composed of bioclastic phosphatic quartz arenite (FT13) followed upward by oyster debris coquinoidal limestone intercalated with chert (FT11 and FT14) represents a sea-level highstand (HST) period.

In south Jordan, the highstand (HST) sea-level period of latest Campanian age represents by oyster debris coquinoidal limestone at Eshidiyya platform (Abd et al. 2007). This oyster debris coquinoidal limestone laterally changed in a more basinal setting at Umm Qais section, north Jordan to sandy foraminiferal packstone (FT7) represents a correlative highstand system tract.

Depositional sequence 2. The second depositional sequence in the Egypt and Jordan spanned the time of deposition of the upper phosphate beds which is placed in the latest Campanian age. The basal sequence boundary of depositional sequence 2, marked by siliciclastic gravel and *Thalassinoides* burrows at the base, and coarse-grained upper phosphate beds, indicates aggradation, higher energy shelf environment with increasing accommodation during transgression (Figs. 2B & 4). The granular phosphorites with

Thalassinoides burrow networks are classic examples of condensed beds that can also be interpreted as firmgrounds associated with increasing accommodation during transgression (Pufahl et al. 2003). A transgressive systems tract overlies these firmgrounds, marking a rise in relative sea level that resulted in the deposition of inner neritic hemipelagic chalk of the Muwaqqar Chalk Marl Formation in Jordan or foraminiferal shale of the Dakhla Formation in Egypt. These hemipelagic facies are characterized by variable abundances of calcareous nannofossils and planktonic foraminifera, and the open marine conditions likely resulted in a low sedimentation rate. This major transgressive event is observed in many parts of Arabian and African plates during the early Maastrichtian (Soudry et al. 2006; El-Azabi and Farouk 2011).

Conclusions

A detailed facies analysis of the Late Campanian succession of Egypt and Jordan, including litho-, bio- and microfacies analyses, resulted in the recognition of 15 characteristic lithofacies types grouped into phosphate (FT1-5), carbonate (FT6-11) and siliciclastic (FT12-15a-c) associations that have been used to characterize the depositional environments.

The phosphate province in Egypt (Duwi Formation) and Jordan (Al-Hisa Phosphorite Formation) is represented by five lithofacies types. The basal phosphate beds in Egypt and Jordan show that major transgressive facies development occurred during the Late Campanian above strong facies changes from the non-marine Qusseir Variegated Shale in Egypt, or, equivalent carbonate facies of the Amman Silicified Limestone Formation. The upper phosphate beds represent another transgressive facies characterized by coarse-grained phosphate with siliciclastic gravel and *Thalassinoides* burrows at the base.

Phosphatic pellets are represented by reworked granular and in situ phosphatic grains (pristine). Granular phosphatic pellets have been recorded from all studied localities. Authigenic phosphatic grains (pristine) are more common in the Jordanian phosphorites. Phosphatic lithoclasts are much more common in the Egyptian phosphorites than in phosphatic rocks in Jordan. Bones and teeth fragments are dominant in the different investigated phosphorites. Their abundance and size increase in the Abu Tartur phosphate rocks.

Bioactive is well developed in the phosphatic grains (pellets and bones) of the Egyptian phosphorites. These phosphatic grains are microbially tunneled (by bacteria), commonly with a micritic carbonate fluorapatite (francolite) and phase filling the tunnels and gradually replacing the bone matrix. The redeposition of francolite in the bored bone fragments is associated by dissolving hydroxyapatite mineral (dahlite).

The main phosphatic beds of the Campanian successions in Jordan lie in the Al-Hisa Phosphorite Formation. Along the Red Sea coast of Egypt as well as in Nile Valley, the thickest phosphatic beds may be in the upper part of the formation or in the middle part and sometimes in the lower part of the formation, that may indicate that some depositional basins were more suitable and/or affected by rapid accumulation of phosphates more than other localities. The Campanian rock units in Jordan, Red Sea and Nile Valley are approximately similar to each other and suggest similar lithofacies development in response to global/regional relative sea-level changes across the Nubo-Arabian Shield on the southern margin of the Neo-Tethys Ocean.

These facies accumulated under protected inner shelf environments where the phosphorite beds were deposited during slight storm include events. On the Abu Tartur plateau, the main phosphorite bed occurs in the lower part of the formation followed by black shales and glauconitic sandstones. The presence of quartz grains in the lower part of the phosphatic bed in Abu Tartur reflects fluvial sediment-supply and shallower depositional environment in these basins than in those along Red Sea coast and in north Jordan. The large amounts of pyrite and glauconite in the Abu Tartur phosphorites suggest an increasingly abundant source of iron-bound phosphate associated with terrigenous sediment input to the continental margin.

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Figures captions

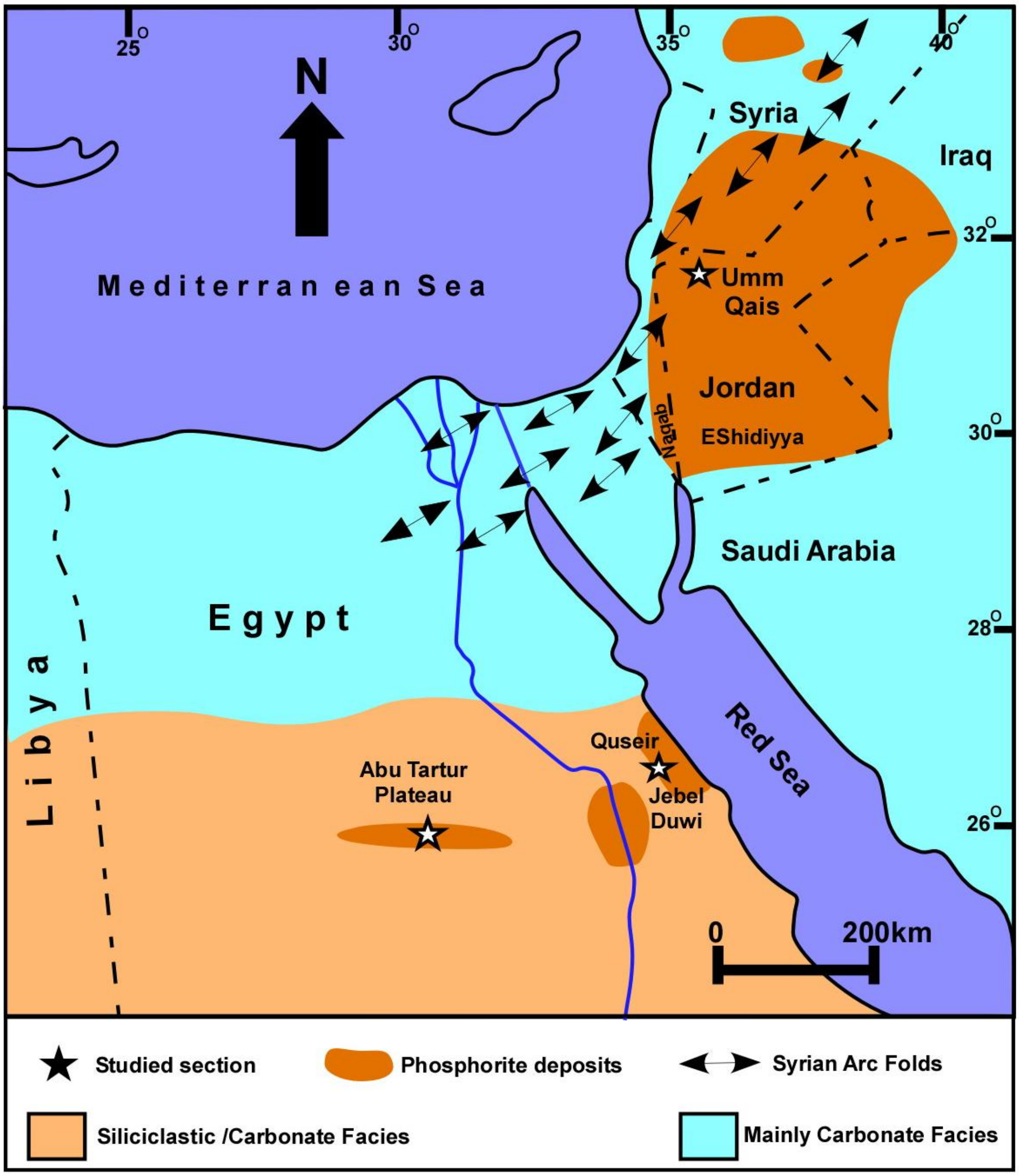

Fig. 1. Location map of the study area showing the phosphorite deposits and their facies associations.

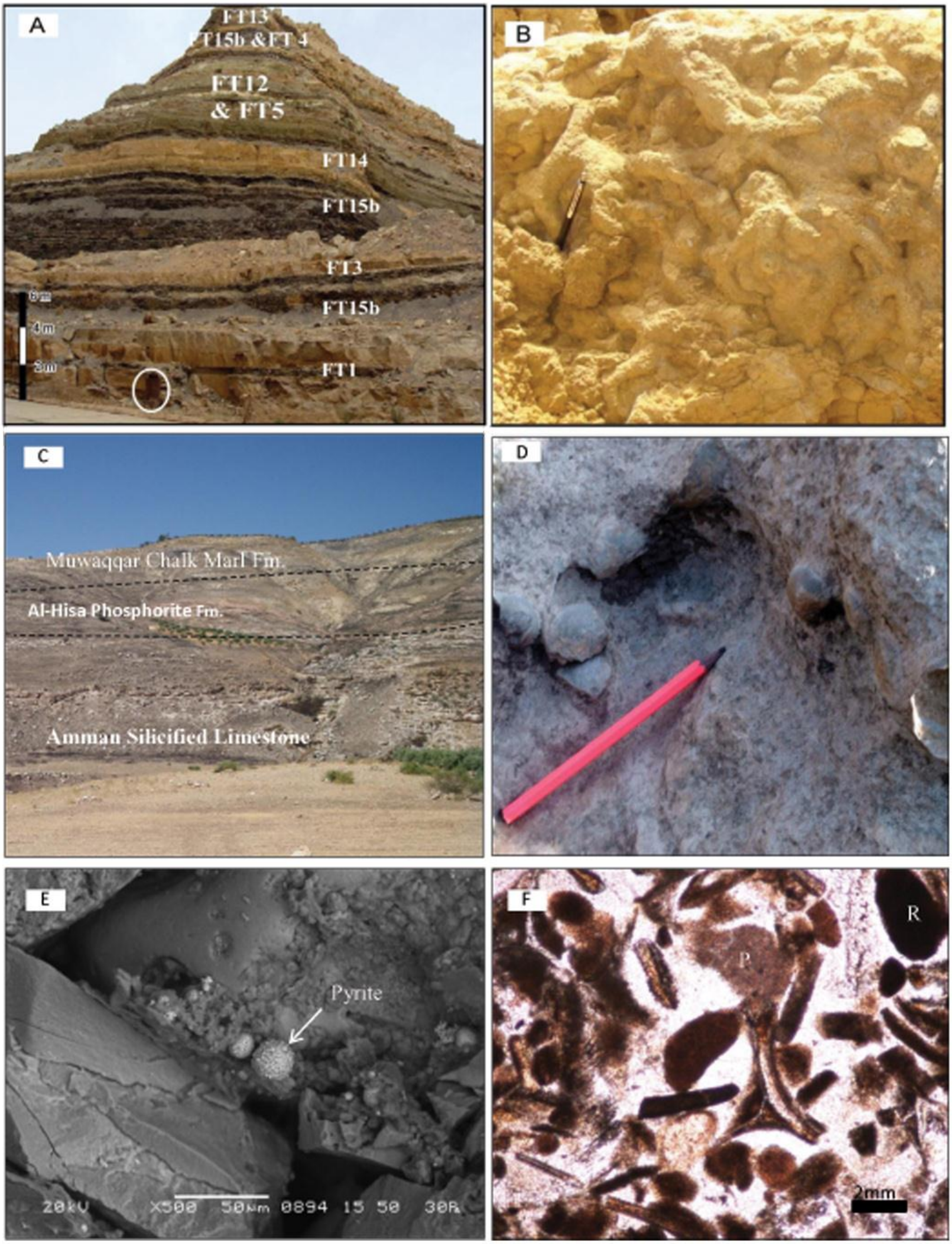
A
FT13
FT15b &FT 4
FT12
& FT5
FT14
FT15b
FT3
FT15b
FT1
6m
4m
2m
B
C
Muwaqqar Chalk Marl Fm.
Al-Hisa Phosphorite Fm.
Amman Silicified Limestone
D
E
Pyrite
20kV X500 50µm 0894 15 50 30R
F
P
R
2mm

Fig. 2A. Photograph showing the Abu Tartur section and its related microfacies types (Circle=1.5m).

Fig. 2B.A higher phosphate bed is marked by intensive burrow system of bifarcuted *Thalassinoides* with siliceous gravel and pebbles at Gebel Duwi section.

Fig. 2C. Filed photograph showing the Amman Silicified Limestone, Al-Hisa Phosphorite Formation and Muwaqqar Chalk Marl Formations exposed at Umm Qais section.

Fig. 2D.Phosphatic limestone rich in *Pycnodonte vesicularis* near the topmost part of the Amman Silicified Limestone.

Fig. 2C. Scan Electron Microscope (SEM) image showing the phosphatic bioclastics are composed of vertebrate fragments with pyrite.

Fig. 2D.Bioclastic phosphatic grainstone (FT1) showing phosphatic pellets of different types: reworked and pristine at Umm Qais section.

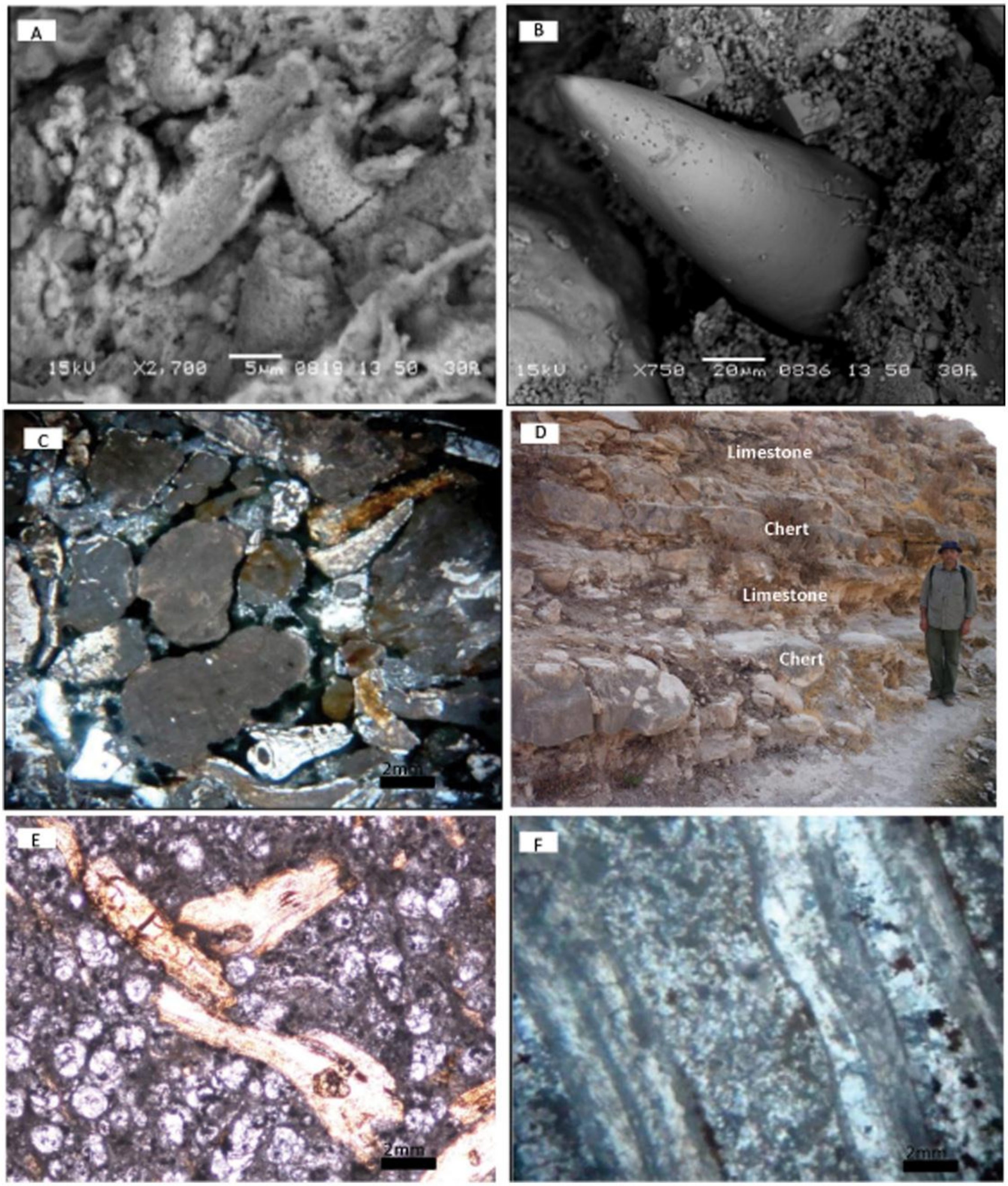
A
15kV X2,700 5μm 0819 13 50 30R
B
15kV X750 20μm 0836 13 50 30R
C
2mm
D
Limestone
Chert
Limestone
Chert
E
2mm
F
2mm

Fig. 3A. SEM image showing the reworked phosphatic pellets and fragments of bone with their cellular structure showing bacterial action (FT3).

Fig. 3B. SEM image showing teeth fragment and dolomite crystal developing in clay matrix characterized the higher phosphate beds at Gebel El- Duwi. Algal bio-pelloidal dolomitic gainstone (FT4).

Fig. 3C.Glauconitic phosphatic grainstone/rudstone (FT5): Authigenic glauconite is well developed in between the phosphatic grain as diagenetic mineral.

Fig. 3D. Filed photograph showing the intercalated of limestone with chert near the upper part of the Amman Silicified Limestone.

Fig. 3E. Microscopic image showing the bioclastic foraminiferal phosphatic wackestone (FT8) showing reworked bone fragments and iron oxide patches in Amman Silicified Limestone.

Fig. 3F. Microscopic image showing oyster bio-pelloidal rudstone/floatstone (FT11) with fibrous crystalline calcite in the oyster shell fragments.

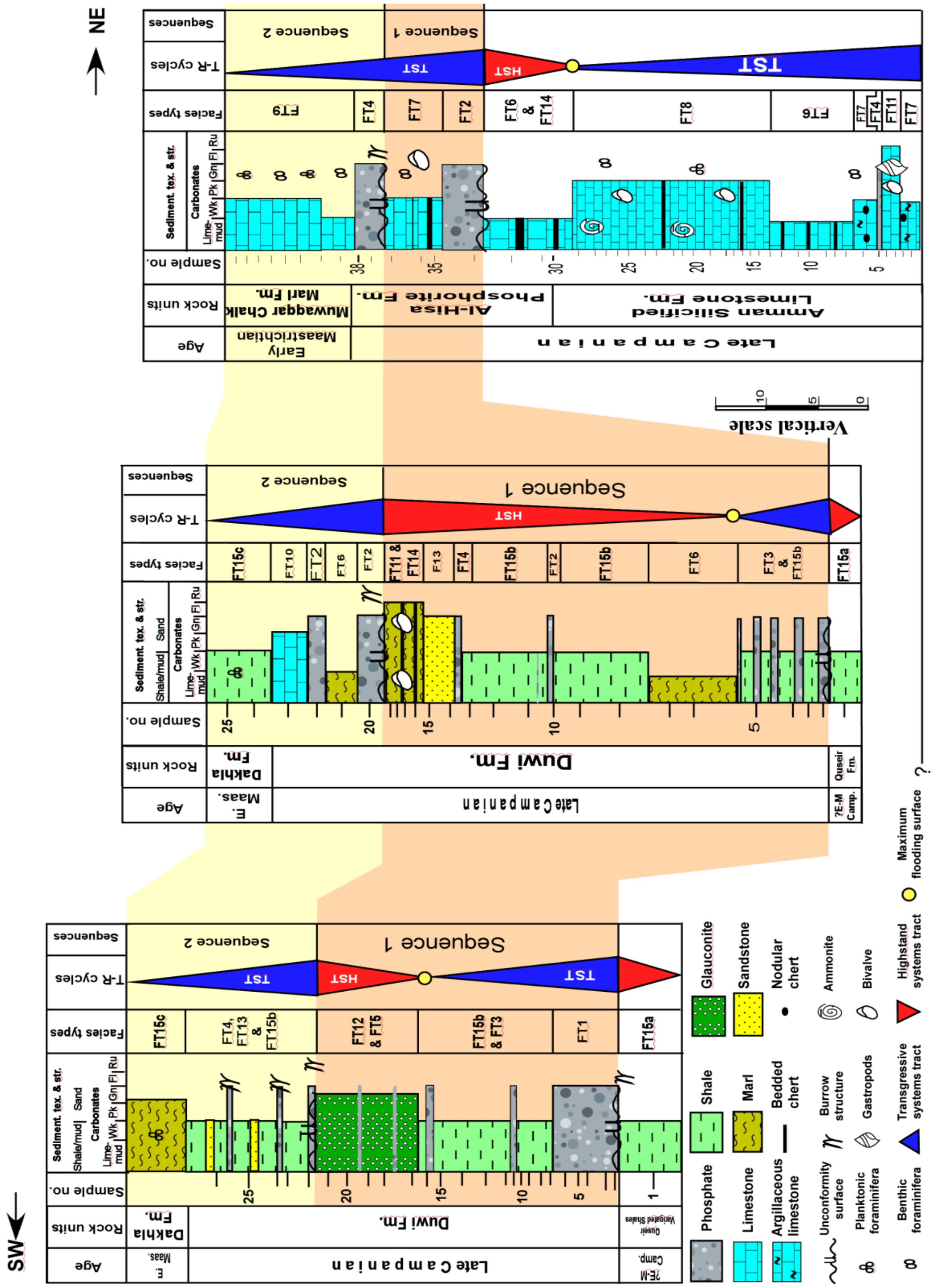

Fig. 4. Sequence stratigraphic correlation and facies associations of the three measured sections from Egypt and Jordan; for location see Fig.1.

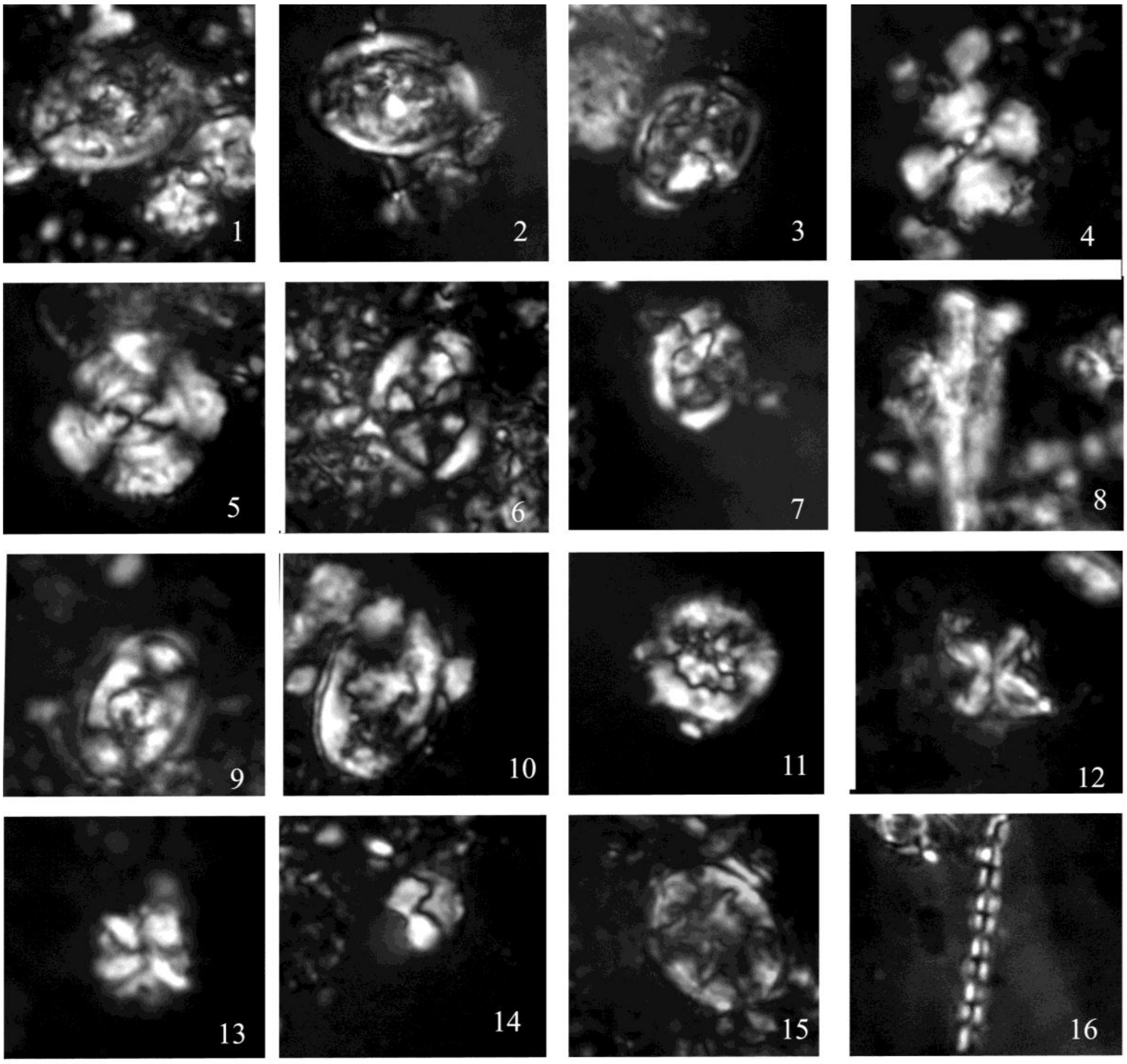

Plate 1 (All photograph taken from late Campanian Amman Silicified Limestone Formation and Al-Hisa Phosphorite Formation at Umm Qais section): 1-3*Reinhardtitesanthophorus* (Deflandre, 1959), sample 37; 4-5 *Watznaueria barnesae* (Black *in* Black & Barnes, 1959), sample 20; 6-7 *Arkhangelskiella cymbiformis* Vekshina 1959, sample 2; 8- *Lucianorhabdus cayeuxii* Dellandre 1959, sample 20;9-10 *Zeugrhabdotus pseudanthophorus* (Bramlette & Martini, 1964), sample15; *Stradneria crenulata* (Bramlette & Martini, 1964), sample ; 12-13 *Micula decussata* Vekshina 1959, sample 37; *Calculites obscurus* (Deflandre, 1959), sample; 15 *Eiffellithus turriseiffelii* (Deflandre *in* Deflandre & Fert, 1954) sample 37*; Microrhabdulus decoratus* Deflandre (1959) sample 37.