Calcareous nannofossil biostratigraphy of Eocene oil shales from central Jordan

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ABSTRACT

Cretaceous and Paleogene marls, rich in total organic carbon, are widespread throughout Jordan and adjoining areas. Based on planktonic foraminifera these oil shales have been assigned a late Campanian–Paleocene age in previous studies. For the current analysis a total of 283 smear slides from five wells in central Jordan have been investigated for calcareous nannofossil biostratigraphy. Findings suggest a much more differentiated age model of the oil shales than previously proposed. The oil shales studied contain abundant calcareous nannofossil taxa of Eocene age along with varying abundances of Maastrichtian and Paleocene taxa. The encountered marker species Rhomboaster cuspis, Tribrachiatus bramlettei, Tribrachiatus orthostylus, Discoaster lodoensis, Coccolithus crassus, Discoaster sublodoensis, Nannotetrina quadrata, Reticulofenestra umbilicus and Chiasmolithus solitus, indicate an Early to Middle Eocene age, while the presence of Maastrichtian and Paleocene forms suggests major reworking. The presence of Cretaceous taxa reflects either subaerial erosive input from the hinterland or submarine reworking of Cretaceous strata within the basin. The highly variable amount of reworked material and associated deposition rates in the basin may represent changes in the tectonic setting during the Eocene. We propose that the high abundances of Cretaceous and Paleocene taxa reflect an increase in accommodation space by active graben flank movements. A dominance of Eocene taxa, on the other hand, indicates either periods of little accommodation space due to graben infill or inversion-type movements of the graben itself. In any case, the youngest Eocene and autochthonous taxa represent shallower or low topography graben phases.

INTRODUCTION

Oil shales as an energy resource have recently experienced a revival and become an important target for oil exploration. This renewed attention and the unique high hydrocarbon yields, when compared to generally accepted, world-class source rock systems, highlight the need for a better understanding of their depositional environment. As a result of the rising demand for unconventional resources, oil shale occurrences and their distribution have been studied and reported globally in recent years (e.g. Dyni, 2003). The oil shales of central Jordan studied here have been described by different lithological terms such as “bituminous limestone” (Bender, 1975; Abed and Amireh, 1983) and “chalk marl or shale that has solid immature organic content” (Hamarneh, 2006). The occurrence of oil shales has been reported for large areas of Jordan, with thicknesses up to 200 m (Hamarneh, 2006) and high but variable amounts of total organic carbon (TOC) of more than 20 percent.

Little is known about the age and depositional paleoenvironment of these oil shales. Previous studies have commonly related them to the Muwaqqar Chalk Marl Formation (MCM) (Bender, 1975; Powell, 1989; Mihdawi and Mustafa, 2007) with ages ranging from Campanian to Paleocene (Bender, 1975; Yassini, 1979; Powell, 1989). These age assignments are based on material from surface occurrences along, or influenced by, the Jordan Valley shoulder uplift such as Yarmouk, Muwaqqar, Lajjun and Jurf ed Darawish (from north to south, respectively, Figure 1). In most of these publications the dating and stratigraphy are based either on macrofossils and/or on planktonic and benthonic foraminifera (e.g. Futyan, 1976; Yassini, 1979). Basha (1982) used benthonic and planktonic foraminifera for correlating wells from the Azraq Sirhan Basin (Saudi Arabian-Jordanian-Syrian Basin) of central east Jordan, assigning the bituminous marl layers an Eocene age.
Figure 1: (a) Paleofacies map of the Arabian Peninsula during the Middle Eocene (after Ziegler, 2001). Jordan is indicated by the red outline in the northwest with its relative position towards the epicontinental sea in the north and northeast. (b) Geographic map of Jordan showing the locations of wells (blue star) used in this study and those of other authors (red pentagon; Yarmouk, Muwaqqar and Lajjun, Jurf ed Darawish, see references in the text); (1) Maastrichtian oil shales, (2) Maastrichtian–Paleocene oil shales, (3) Maastrichtian–Eocene oil shales, and highlighted areas with different sub-basins throughout Jordan (proposed by Basha, 1982; Abed, 2013).
Calcareous nannofossils are a reliable tool for biostratigraphical dating and age assignment for Mesozoic and Cenozoic sediments. Their rapid evolution also allows them to be used for describing and quantifying geological phenomena such as in-situ deposition versus reworking. Due to their widespread geographic distribution, they have been employed in numerous biostratigraphic and palaeoenvironmental analyses (e.g. Martini, 1971; Sissingh, 1977; Okada and Bukry, 1980; Perch-Nielsen, 1985; Haq, 1998; Fioroni et al., 2012) (Figure 2). On the global scale, the first subdivisions of Paleogene strata go back to Bramlette and Sullivan (1961), Hay and Mohler (1967), Martini (1970), Bukry (1973) and Okada and Bukry (1980). Bown (2005) summarized numerous papers performed on calcareous nannofossils of Paleogene sediments.

The current study of oil shales from central east Jordan is a complementary and more detailed supplement of the previously published biostratigraphic framework of Jordan (Yassini, 1979; Basha, 1982). It aims to determine the relative age and the establishment of a stratigraphic scheme of the oil shale occurrences in Jordan by using calcareous nannofossils from five wells. Our biostratigraphic scheme is adopted in a companion paper that interprets depositional paleoenvironments from two of these wells (Ali Hussein et al., 2014). Furthermore, our study correlates phases of reworking with tectonic events and the related basin formation in the Paleogene.

**BIOSTRATIGRAPHY USED IN THIS WORK**

The data of previous biostratigraphic studies (Edwards, 1971; Martini, 1971; Okada and Bukry, 1980; Naji, 1983; Perch-Nielsen, 1985; Toker, 1989; Wie and Thierstein, 1991; Yzbek, 1999) are shown in Figure 2. For Jordan only Naji (1983) described calcareous nannofossils from Coniacian to Middle Eocene sections, whereas many biostratigraphic investigations based on calcareous nannofossils have been carried out in areas adjacent to Jordan (e.g. Sadek, 1972: Egypt; Toker, 1989: Turkey; Eshet and Almogi-Labin, 1996: Levant; Yzbek, 1999: Syria). A major difficulty for dating the Jordanian oil shales is the inconsistent abundance patterns of index species typically used for defining the base or top of biozones due to rare abundance and poor preservation of calcareous nannofossils.

The current study uses the Martini (1971) biozonation (Figure 2), with few exceptions: (1) The *Rhomboaster cuspis* Zone defines for the lowermost Eocene. (2) The base of the *Discocysta binodosus* Zone is identified by the first occurrence (FO) of *Tribrachiatus orthostylus* as suggested by Kerdany (1970) and Perch-Nielsen (1985). (3) The *T. orthostylus* Zone was introduced by Brönnimann and Stradner (1960) and correlates with the CP10 Zone of Okada and Bukry (1980), the FO of *Discocysta lodoensis* defines the base of this zone. (4) The FO of the *Coccolithus crassus* marks the base of the *D. lodoensis* Zone which was introduced by Brönnimann and Stradner (1960) and which correlates with CP11 of Okada and Bukry (1980). (5) *Nannotetrina fulgens* and other species of the genus *Nannotetrina* are used to define the base of the *N. fulgens* Zone following the suggestions of Perch-Nielsen (1985). This zone was introduced by Hay et al. (1967). Wie and Thierstein (1991) used *Nannotetrina cristata* instead of *N. fulgens* to define the base of this zone. (7) The base of the *Discocysta tanii nodifer* Zone is identified by the FO of *Reticulofenestra umbilicus* (Okada and Bukry, 1980) or by the FO of *Discocysta bifax* (Martini, 1970).

**GEOLOGICAL BACKGROUND**

During the Maastrichtian, central Jordan was positioned paleogeographically on a broad shallow-marine shelf along the southern rim of the Neo-Tethys Ocean (Figure 1). The Syrian Arc Fold System initiated in the Late Cretaceous coincided with the closing of the Neo-Tethys Ocean and the convergence of the Arabian Plate with the Eurasian Plate (Abd El-Motaal and Kusky, 2003). This system extends in an S-shaped belt from northern Egypt through Palestine, Lebanon and Syria, and influenced the regional paleotopography. This resulted in anticlinal crests forming clusters of local highs, potentially in the form of subaerially exposed islands. The synclines or depressions, on the other hand, formed local basins that represent a suitable environment for oil shale deposition (e.g. Abu-Jaber et al., 1989; Ziegler, 2001). As part of the Alpine Orogeny and the associated closure of the Neo-Tethys Ocean, major shortening occurred in the Arabian Plate, which affected central east Jordan and the Mediterranean region. The synchronous subsidence of the Azraq Sirhan Basin was...
Figure 2: Correlation of various calcareous nannofossil zonation schemes.
described as a regional effect of this plate tectonic induced process. The development of the Azraq Sirhan Basin appears to have started in the Early Cretaceous based on the paleogeographic analysis of the sedimentary rocks. As a consequence, the Mesozoic strata underwent intense deformation, regional uplift, erosion and volcanic activity was observed (Abu-Jaber et al., 1989).

The last phase of the Arabian-Eurasian plate convergence is of Eocene age, when the island arcs of the Arabian and Eurasian plates collided. With the ongoing closure of the Neo-Tethys Ocean in the Middle Eocene, the Arabian Plate fully collided with Eurasia along its northeastern margin (Butterlin et al., 1993). The “deformed package” that was located north of the Arabian Peninsula represents the ongoing collision (Alsharhan and Nairn, 1997). For the entire Maastrichtian to Late Eocene, the plate-tectonic movements played a major role in regional and local basin formation, with resulting topography controlling the sediment facies types and transport mechanisms.

In conjunction with plate tectonic movement, relative sea-level changes from the late Middle Eocene in Jordan show a regressive trend following a Paleocene to Middle Eocene highstand (Haq and Al-Qahtani, 2005). Paleogeographical and sedimentological data point to the creation of subaerial topography as a result of these plate tectonic movements and sea-level changes (Alsharhan and Nairn, 1995).

**LITHOLOGY, LOCATIONS AND SAMPLES**

The Muwaqqar Chalk Marl unit of Jordan, which contains the oil shales, has been described by numerous authors as being composed of soft, thick-bedded, highly bituminous chalky marls to chalky limestones (Figure 3). These grey marls or grey argillaceous limestones contain limestone concretions and chert nodules (e.g. Bender, 1975; Powell, 1989; Andrews, 1992). Basha (1982) described the sediments as brownish, phosphatic-silty and bituminous marls interbedded with black cherts. According to the latter author, the unit changes laterally into limestones or dolomitic limestones interbedded with marl.

Core and cutting samples were obtained and studied from five wells (OS-14, OS-22, OS-23, OS-25 and OS-28) located in central east Jordan (Figure 3). These wells were drilled by the Jordan Oil Shale Company (JOSCO) in 2010–2012. Wells OS-22 and OS-23 are located southeast of Amman, wells OS-14 and OS-28 were drilled southeast of Azraq, and well OS-25 is positioned northwest of Safawi (Figure 1). Wells OS-22, OS-23 and OS-28 were cored, with OS-22 and OS-23 having 5-inch diameter cores for research purposes, and OS-28 having a 3.5-inch core. Wells OS-14 and OS-25 were not cored, and therefore cutting samples were used for this analysis.

Wells OS-14, OS-22, OS-23 and OS-25 exhibit similar lithological units that consist of phosphatic limestones alternating with chert bands at the bottom, which are overlain by ca. 120 m of organic-rich marls. Chalky limestones with chert bands re-occur towards the top. Core OS-28 contains a succession of 150 m of organic-rich marls with a 30 m section of intercalated white limestones at the top from 44.90–13.75 m. The biostratigraphic zones and their respective lithologies are illustrated in Figure 3. A total of 283 samples have been taken from the five wells (OS-22, n = 111; OS-23, n = 101; OS-28, n = 34; OS-14, n = 20 and OS-25, n = 17). Relative depths are given for the samples starting from the bottom.

**METHODS**

Simple smear slides were prepared following the procedure described by Roth (1984). Calcareous nannofossils were taxonomically examined using an Olympus BH2 light microscope with a magnification of 1,250 x. Calcareous nannofossils from 100 fields of view were counted in each simple smear slide. Preservation of species was determined using the procedure of Roth (1973). In order to estimate the abundance of autochthonous and allochthonous species, counting was performed on smear slides following procedures described by Backman and Shackleton (1983) and Wie and Wise (1989). The abundance pattern for each species was defined as follows: Abundant (A) = more than 10 specimens per 1 FOV (field of view); Common (C) = 1–10 specimens per 1 FOV;
Few (F) = 1 specimen per 2–50 FOVs; Rare (R) = 1 specimen per 51–100 FOVs; Absent (NO) = no specimens in more than 100 FOVs.

The abundance proportion of Cretaceous specimens is the total of these specimens divided by the total amount of specimens encountered in a given sample. Paleocene and Eocene proportions were calculated using the same method. Total abundance of calcareous nannofossils in each sample therefore add to one hundred percent. TOC and CaCO₃ content data were measured using a calibrated LECO-CS-244 elemental analyzer at Newcastle University (Aqleh et al., 2013).

The biostratigraphic zonation used in this study is based on Martini (1971), Edwards (1971), Okada and Bukry (1980) and Perch-Nielsen (1985). The boundaries between different biozones were established by the first occurrence (FO) and last occurrence (LO) of the following marker species: R. cuspis, Tribrachiatus cramletae, T. orthostylus, D. lodoensis, C. crassus, Discoaster sublodoensis, Nannotetrina quadrata, R. umbilicus and Chiasmolithus solitus.

### RESULTS

Calcareous nannofossils species, which were used as index fossils are illustrated in Figure 4 (a–e) and Plates 1 and 2.

### Core OS-28

In the lower part of this well, at sample 48.20 m, the Paleocene/Eocene boundary is characterized by a distinctive gradual darkening of the section, containing a high amount of organic matter. Rare calcareous nannofossils (Figure 4a) have been found in sample 48.20 m taken directly at the boundary itself.
Plate 1: Transmitting plane polarized and cross-polarized light micrograph of Early to Middle Eocene calcareous nanofossils in the wells of this study. Scale bar on upper left applies to all.
Plate 2: Transmitting plane polarized and cross-polarized light micrograph of the Middle Eocene and an example of Cretaceous calcareous nannofossils in the wells of this study. Scale bar on upper left applies to all.
Figure 4: (a and b) Marker species used in this study, and their distribution in the wells.
NANNOFOSSIL DISTRIBUTION IN EOCENE

### (c) Core OS-22

| Depth (m) | Lithology | Nannofossil Sample |
|-----------|-----------|--------------------|
| 0         | Marl      | R. umbilicus       |
| 20        | Marl      | D. bifax           |
| 40        | Phosphate | N. quadrata        |
| 60        | Limestone | D. sublodoensis    |

### (d) Well OS-25

| Depth (m) | Lithology | Nannofossil Sample |
|-----------|-----------|--------------------|
| 0         | Marl      | D. scrippaeae      |
| 20        | Phosphate | N. quadrata        |
| 40        | Limestone | B. stilus          |

Figure 4: (c and d) Marker species used in this study, and their distribution in the wells.
NANNOFOSIL DISTRIBUTION IN EOCENE

Figure 4: (e) Marker species used in this study, and their distribution in the wells.
Samples 48.00 m and 47.75 m were assigned to the Early Eocene R. cuspis Zone based on the presence of R. cuspis. The top of the R. cuspis Zone is determined by the FO of T. branlettei. The FO of T. branlettei was observed in sample 45.99 m, which is therefore taken as the base of Tribrachiatus contortus Zone. The assemblage composition in the interval 47.45–45.99 m suggests mixing of nannofossils from the R. cuspis and T. contortus zones. Due to the absence of T. contortus, the FO of T. orthostylus can be used as an approximation of the T. contortus Zone/D. binodosus Zone boundary at sample 45.99 m. Few calcareous nannofossils were observed in samples from 45.99–44.00 m, a 2 m-thick interval. Paleocene species, like Towieus pertusus and Fasciculithus tympaniformis disappear at the top of the T. contortus Zone in samples 45.00 m and 44.00 m respectively.

Samples 43.30–41.83 m were assigned to the D. binodosus Zone. The top of the zone is marked by the FO of Discoaster lodoensis. Chiasmolithus californicus, Discoaster binodosus, Sphenolithus conspicuus, Pontosphaera pulchra and Imperiaster obscurus have their FOs throughout this zone. The LOs of Discoaster diastypus, T. branlettei and Tribrachiatus digitalis were observed within this zone (Figure 4a). Poor to moderately preserved calcareous nannofossils were encountered just above the Paleocene/Eocene boundary in samples 41.83–17.70 m an increase in calcareous nannofossils with moderately preserved specimens has been observed. Towards the top of the core, well-preserved species occur with abundances ranging from rare to abundant.

The occurrences of T. orthostylus and D. lodoensis overlap in samples 40.23 m and 39.00 m. The FO of C. crassus is more reliable than the FO of D. lodoensis. This zone is considered to be relatively condensed with poor calcareous nannofossil preservation. The FO of C. crassus was observed in sample 39.00 m, which coincides with the base of the Discoaster lodoensis Zone. Reticulofenestra dicoxyda, Sphenolithus radians and Girgisia gammanum have their LOs throughout the D. lodoensis Zone. In the literature (e.g. Perch-Nielsen, 1985), the top of the D. lodoensis Zone is defined by the FO of D. sublodoensis. Common to rare T. orthostylus and D. lodoensis have been observed in the interval 38.00 m to the top of the core. In samples 41.83–17.70 m an increase in calcareous nannofossils with moderately preserved specimens has been observed. Towards the top of the core, well-preserved species occur with abundances ranging from rare to abundant.

Wells OS-14, 22, 23 and 25

The lowermost part of the cores OS-22 (218–222 m) and OS-23 (254–213 m) were assigned to a composite zone as suggested by Haq and Aubry (1981), which includes the T. orthostylus, D. lodoensis and D. sublodoensis zones. Rare calcareous nannofossils were observed in core OS-23 (254–206 m) and core OS-22 (222–215 m). The Early/Middle Eocene boundary is based on the FO of D. sublodoensis, with rare calcareous nannofossils occurring in this zone.

In well OS-25, D. sublodoensis is absent. The FO of Blackites stilus which approximates the base of the D. sublodoensis Zone (Bown, 2005) was observed in sample 169 m. In core OS-23, the thickness of the D. sublodoensis Zone is condensed and represented by a 3 m interval; however, it is expanded in wells OS-25 and OS-22 with 12 m and 45 m thickness, respectively.

The base of the N. fulgens Zone is defined by the FO of N. fulgens in wells OS-14, OS-22, OS-23 and OS-25 (320 m, 171 m, 209 m and 157 m, respectively). Samples 209–80 m (core OS-23), 171–22 m (core OS-22), 157–56 m (well OS-25) and 320–277 m (well OS-14) were assigned to this zone, which yields common N. quadrata. The genus Nannotetrina, typical for the N. fulgens Zone of the Middle Eocene (Perch-Nielsen, 1985), is common in these samples. Sedimentation rates range from 2.8 cm per 1,000 year to 4.1 cm per 1,000 year. The D. sublodoensis and N. fulgens zones yield poorly to moderately preserved calcareous nannofossils. These two nannofossil biozones are described by a high amount of organic matter and high abundance of allochthonous specimens. The abundance of Cretaceous species significantly increases upwards, becoming dominant around sample 75 m (core OS-23).

The base of the D. tanii nodifer Zone is identified by the FO of R. umbilicus or by the FO of D. bifax. The FO of R. umbilicus was observed in samples 80 m (core OS-23) and 22 m (core OS-22). R. umbilicus is a useful marker in wells OS-14, OS-25 and OS-23, but absent at sample 22 m (core OS-22). D. bifax was absent in wells OS-23 and OS-25. The FO of R. umbilicus was observed in sample 277 m (well OS-14) accompanied by D. nodifer. The top of the D. tanii nodifer Zone is marked by the
LO of *D. bifax* or the LO of *C. solitus*. The *D. tanii nodifer* Zone has a thickness of 20 m (core OS-22), 80 m (core OS-23), and 160 m (well OS-14). Poor to moderately preserved specimens were also observed in the *D. tanii nodifer* Zone in well OS-14 between samples 277–120 m. The *D. tanii nodifer* Zone in samples 22–2 m (core OS-22) and in samples 23–6 m (core OS-23) yields abundant and well preserved species.

The LO of *C. solitus*, which was observed in sample 116 m (well OS-14), indicates the base of the *D. saipanensis* Zone. The FO of *C. oamaruensis* defines the top of the *D. saipanensis* Zone. The thickness of the *D. saipanensis* Zone is 300 m.

**Correlation of Wells**

When correlating wells OS-14, OS-22, OS-23 and OS-25 biostratigraphically, problems arose over the top of core OS-28 and the base of the other wells (Figure 5). The *D. lodoensis* Zone, which is the top horizon of core OS-28, does not match biostratigraphically with the composite zone at the bottom of cores OS-22 and OS-23. This was ascribed to the scarcity of calcareous nannofossils in cores OS-22 and OS-23.

Both the *D. sublodoensis* Zone and the overlying *N. fulgens* Zone are expanded in core OS-22. The *D. sublodoensis* Zone is missing in well OS-14 since the marker species of this zone are absent. The *N. fulgens* Zone is thick in wells OS-14, OS-22, OS-23 and OS-25. The base of the *D. tanii nodifer* Zone can be correlated throughout the wells OS-14, OS-22 and OS-23. It is difficult to use the *D. tanii nodifer* Zone for correlating well OS-25 with other wells due to the absence of *R. umbilicus* in OS-25. The lower part of the *D. tanii nodifer* Zone is not recorded in wells OS-14 and OS-23, due to the absence of the indicative nannofossils. The *D. tanii nodifer* Zone, which represents the Bartonian (late Middle Eocene) was found at 277 m (well OS-14) based on *D. nodifer* co-occurring with *R. umbilicus* (Figure 6). The *D. saipanensis* Zone only occurs in well OS-14.

**Reworking**

Allochthonous calcareous nannofossils have varying abundances in the studied material. Rare to abundant allochthonous calcareous nannofossils species were detected throughout the studied wells with Maastrichtian species co-occurring with Eocene taxa.

The percentages of Cretaceous, Paleocene and Eocene species in core OS-23 are illustrated in Figure 7. Five intervals (A–E) can be identified based on the occurrence of allochthonous species of different ages (Maastrichtian–Eocene). These five intervals are from 254–206 m (interval A), from 205–177 m (interval B), from 172–119 m (interval C), from 112–45 m (interval D), and from 44–0 m (interval E). Intervals A and E contain dominant autochthonous species and minor Paleocene allochthonous species. In both intervals calcareous nannofossils are well preserved with differences in abundances. Calcareous nannofossils are rare in interval A, but common in interval E. Abundance of allochthonous species fluctuate from rare to abundant on a centimeter scale in interval B. Allochthonous Paleocene species are absent at the top of this interval and throughout interval C. In contrast, the abundance of Cretaceous species increases upwards and exceeds the abundance of autochthonous Eocene species in specific samples.

Allochthonous *Micula murus* was found from 203–48 m. This suggests that the Cretaceous source is of late Maastrichtian age. The presence of the taxa *Cruciplacolithus primus*, *Cruciplacolithus tenuis* and *T. pertusus* reflects erosion, transportation and re-deposition of Paleocene strata. A typical example for the complexity of the multiple sources of reworking is sample 74 m: in addition to the autochthonous Middle Eocene species, the presence of *Micula murus*, *Chiasmolithus danicus*, *T. pertusus* and *F. tympaniformis*, *T. orthostylus*, *D. lodoensis* and *Helicosphaera seminulum* indicates reworking of Maastrichtian, Paleocene and Early Eocene strata into the basin during the Middle Eocene.
DISCUSSION

Biostratigraphy

Our stratigraphic findings correspond to the Martini (1971) biozonation. A major difficulty for dating the wells in this study is the inconsistent abundance patterns of index species typically used for defining the base or top of biozones. The inconsistent abundance patterns may reflect: (1) a regional hiatus; (2) specific climatic and/or environmental conditions unsuitable for calcareous nannofossils; and/or (3) poor preservation.

Figure 5: Stratigraphic correlation between well sites using marker calcareous nannofossils.
A hiatus has been detected in wells OS-14 and OS-23 based on the FO of *D. bifax*, *R. umbilicus* and *D. nodifer*. Taking into account that there is a chronological difference in the FOs of these species, the FOs of *D. bifax* and *R. umbilicus* equate the *N. fulgens/D. tanii nodifer* Zone boundary. The FO of *D. nodifer* is associated with the Lutetian (Middle Eocene)/Bartonian (late Middle Eocene) boundary within the *D. tanii nodifer* Zone (Perch-Nielsen, 1985; Aubry, 1988; Berggren et al., 1985). The co-occurrence of *D. nodifer* and *R. umbilicus* in well OS-14 may be explained by the poor preservation of these index species in the sediments predating the *D. tanii nodifer* Zone (Figure 6).

The absence or scarcity of calcareous nannofossils in the Middle Eocene of the OS-22 and OS-23 cores may be explained by a depositional environment that was unsuitable for calcareous nannofossils. The accumulation of high amounts of organic matter in the marly sediments suggests a restricted basin environment. The source of the organic matter remains unknown, but is most likely of bacterial and not of phytoplanktonic origin.

**Age Differentiation**

One of the main findings of this study is the complex age model for the oil shales in central east Jordan. Previous studies concentrated on the Lajjun area (Bender, 1975; Yassini, 1979) in central west Jordan, which experienced erosion due to topographic uplift in post-Miocene times (Bender, 1975). Other authors focused on the oil shales of north or central east Jordan (Yassini, 1979; Basha, 1982; Powell, 1989; Andrews, 1992; Nazzal and Mustafa, 1993). The current study assigns the oil shales in central east Jordan as outlined by the OS wells (Figure 1) an Early to Middle Eocene age. This age is confirmed by the findings of Basha (1982) and Andrews (1992) who studied marls with bituminous sections from the Azraq Sirhan Turayf Basin in central east Jordan and northern Saudi Arabia.

The different ages suggested for the oil shales in Jordan (central west Jordan: Maastrichtian; north Jordan: Maastrichtian–Paleocene; central east Jordan: Maastrichtian–Eocene, Figure 1) can be explained by the progressing differentiation.
of the geological setting from a broad shelf environment in the Maastrichtian to localized tectonically controlled depositional settings in the Eocene.

In the early Maastrichtian a rapid eustatic sea-level rise led to the formation of an epicontinental sea with limited connection to the Neo-Tethys Ocean (Powell and Muh'd, 2011). In this epicontinental sea sub-basins developed (Jafir Sub-basin, Azraq-Hamza Sub-basin, Yarmouk Sub-basin and Lajjun Sub-basin, Figure 1) caused by syn-sedimentary subsidence (Abed et al., 2005; Abed, 2013). The genesis of these sub-basins can be ascribed to mainly NW-trending and sometimes EW-trending normal faults, which created horst and graben structures. Alternatively, the sub-basins were proposed to have formed from gentle, synformal folding related to the Syrian Arc or Alpine Orogeny (Abu-Jaber et al., 1989). In both cases, the increase in topographic relief may have created sills between the sub-basins that became increasingly restricted.

Irrespective of the alternatives mentioned for structural genesis of the sub-basins, there are also different scenarios that need to be taken into account for the origin of the water mass anoxia.
A reduction in water circulation leading to periods with a positive water balance caused an estuarine dominated circulation pattern (Hay, 1995). Stagnation of the water column and internal nutrient recycling took place, which is generally associated with the development of more reducing (i.e. anoxic) conditions of the bottom waters such as is observed in, for example, the present-day Black Sea. Alternatively, the described formation of anticlines or horsts obstructed the east-west-flowing Tethyan Circumglobal Current in central Jordan, i.e., shielding off the synclines or grabens from full marine water exchange, and resulting in the trapping of nutrients thereby creating stagnant localized waterbodies (e.g. Abed and Amireh, 1983; Almogi-Labin et al., 1993; Abed, 2013). For the former concept (Black Sea model) an isthmus-like single connection to the Neo-Tethys Ocean in northwest Jordan would have been needed. For the second model central Jordan would have been connected to the Neo-Tethys Ocean both via the west (northern Levant and southern Lebanon), and the east (east Jordan-Iraq). As no material was available in this study to prove either scenario, the two concepts remain potentially valid as both in turn would have favored the deposition of oil shales (e.g. Demaison and Moore, 1980; Hay, 1995).

Such oil shales continued to be deposited in central east Jordan, i.e., the Azraq-Hamza Sub-basin until the end of the Middle Eocene. A major regression leading to a depositional environment with shallower water depths was induced by tectonic activity. As an effect of this regression and filling of the tectonically related sub-basins, the dense anoxic bottom-water masses were replaced by more oxygenated waters. The depositional environment favorable for oil shale formation retreated to the basin center (well OS-14) during the late Middle Eocene. This lateral shift is reflected by the shift in the occurrence of the index species of the *D. tanii nodifer* Zone.

In central west Jordan (e.g. Lajjun), the Early to Middle Eocene represents a period during which the subsiding sub-basins were already filled with sediments and bituminous intervals are mostly absent (Powell, 1989; Andrews, 1992; Powell and Moh'd, 2011).

**Reworking**

Allochthonous species in the samples are helpful for the interpretation of the otherwise complex depositional setting. There are significant changes in the source and amount of terrigeneous input throughout the interval studied (Maastrichtian–Eocene). The change in the source and amount of reworking reflects rapid reversal movements of the graben structures with a resulting complex paleotopographic relief of juxtaposed structural units (horsts, grabens, tilted blocks, half grabens). Horsts or crests induced by these movements reflect the erosional source areas and therefore directly influence the volume of reworked material. For subaqueous crests or horsts, i.e. those remaining submerged under water but above wave-base depth, reworking by wave action is a feasible hypothesis. The nature of the reworked material, sub- to unconsolidated carbonate dominated muds and marls, favors easy erosion and long-distance transport of nano-scale matter.

The studied oil shales have abundant Cretaceous, Paleocene and intra-Eocene allochthonous species. Throughout the cores, five intervals were found, each one having a different age of reworking and, therefore, a different source of erosion. During the Early Eocene no reworking took place (interval A). The first allochthonous species appeared in the early Middle Eocene documented by Paleocene and Cretaceous species (interval B). This first reworking phase is here linked to the initial graben formation and increased topographic relief of the graben shoulders (Figure 8). An accentuation of the graben during the Middle Eocene is reflected by a decline of allochthonous Paleocene species and simultaneously an increase in allochthonous Cretaceous specimens (interval C and D). Whether this truly reflects a reversed stratigraphy remains inconclusive as no intra-Paleocene or intra-Maastrichtian biozonation was possible. This increase in Cretaceous specimens corresponds with an increase of the TOC content in the late Middle Eocene (Figure 8). Finally, the re-occurrence of allochthonous Paleocene species during the latest Middle Eocene indicates that another phase of graben movement and the renewed juxtaposition of Paleocene strata along the shoulders of the graben must have occurred in an area not eroded during the first cycle (interval E).
An Early to Middle Eocene age was assigned to oil shales in central east Jordan based on autochthonous species of calcareous nannofossils. Rare to few calcareous nannofossils occur in the Early and early Middle Eocene, few to common taxa are present in the sediments of late Middle Eocene age. Typical Eocene index species were not observed in the studied wells. Variable amounts of Cretaceous and Paleocene species are mixed in with the autochthonous Eocene species. The former represent fluctuations in reworking of a nearby sedimentary source most likely caused by a very complex sub-regional tectonic setting. Material from the Cretaceous was eroded from either horst structures or crests of folds, which themselves were induced by the Syrian Arc or Alpine Orogeny. The eroded material was transported into the sub-basins and re-deposited. Observation of intra-Eocene allochthonous species suggests other types of reworking such as intra-basinal transportation by sub-tidal wave activity. Changes in the source and transportation rate reflect syn-depositional Eocene tectonic movements resulting in shallowing or deepening of the grabens.

CONCLUSIONS

An Early to Middle Eocene age was assigned to oil shales in central east Jordan based on autochthonous species of calcareous nannofossils. Rare to few calcareous nannofossils occur in the Early and early Middle Eocene, few to common taxa are present in the sediments of late Middle Eocene age. Typical Eocene index species were not observed in the studied wells. Variable amounts of Cretaceous and Paleocene species are mixed in with the autochthonous Eocene species. The former represent fluctuations in reworking of a nearby sedimentary source most likely caused by a very complex sub-regional tectonic setting. Material from the Cretaceous was eroded from either horst structures or crests of folds, which themselves were induced by the Syrian Arc or Alpine Orogeny. The eroded material was transported into the sub-basins and re-deposited. Observation of intra-Eocene allochthonous species suggests other types of reworking such as intra-basinal transportation by sub-tidal wave activity. Changes in the source and transportation rate reflect syn-depositional Eocene tectonic movements resulting in shallowing or deepening of the grabens.
A tectonically induced closure of the basinal structures during the late Middle Eocene led to the restriction that allowed oil shales to form in the sub-basin centers.

**APPENDIX: TAXONOMIC LIST OF SPECIES USED IN THIS STUDY**

| Cyclagelosphaera | C. tenuis | C. primus | C. latipons | C. frequens |
|------------------|----------|----------|-------------|-------------|
|                  |          |          |             |             |
| Biscutum         |          |          |             |             |
| Blackites        |          |          |             |             |
| B. spinosus      |          |          |             |             |
| Braneriozonin    |          |          |             |             |
| B. stellus       |          |          |             |             |
| Brahmutella      |          |          |             |             |
| C. deta          |          |          |             |             |
| C. eudela        |          |          |             |             |
| Chiasmolithus    |          |          |             |             |
| C. dela          |          |          |             |             |
| C. gigas         |          |          |             |             |
| C. granis        |          |          |             |             |
| C. californicus  |          |          |             |             |
| C. constrictus   |          |          |             |             |
| C. denticus      |          |          |             |             |
| C. gigas         |          |          |             |             |
| C. gravis        |          |          |             |             |
| C. mahmoudi      |          |          |             |             |
| C. mediusus      |          |          |             |             |
| C. multiradiatus |          |          |             |             |
| C. petrevinitii   |          |          |             |             |
| C. pharensis     |          |          |             |             |
| C. pelagicus     |          |          |             |             |
| C. pseudopatula  |          |          |             |             |
| C. primus        |          |          |             |             |
| C. rectus        |          |          |             |             |
| C. rupétus       |          |          |             |             |
| C. saipanensis   |          |          |             |             |
| C. saipanensis   |          |          |             |             |
| C. seminulum     |          |          |             |             |
| C. pseudopatula  |          |          |             |             |
| C. seminulum     |          |          |             |             |
| C. seminulum     |          |          |             |             |
| C. seminulum     |          |          |             |             |
| C. seminulum     |          |          |             |             |
| C. seminulum     |          |          |             |             |

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Lithraphidites Deflandre 1963
L. carniolensis Deflandre 1963
L. quadratus Bramlette and Martini 1964
L. laxithecus Noël 1965
L. armilla (Black in Black and Barnes 1959) Noël 1965
Markallites Bramlette and Martini 1964
M. apertura Perch-Nielsen 1979
M. inversus (Deflandre in Deflandre and Fert 1954)
Bramlette and Martini 1964
Micrantholithus Deflandre in Deflandre and Fert 1954
M. astrum Bown 2005
M. varius Perch-Nielsen 1979
M. breviflora Deflandre in Deflandre and Fert 1954
Microrhabdulus Deflandre 1959
M. decoratus Deflandre 1959
Micula Yekshina 1959
M. concava (Stradner in Martin and Stradner 1960) Verbeek 1976
M. accrescita Yekshina 1959
M. prinsi Perch-Nielsen 1979
M. morus (Martini 1961) Bukry 1973
Nannotetra Acuthan and Stradner 1969
N. alata (Martini 1960) Haq and Lohmann 1976 (= C. alatus)
N. fulgens (Stradner 1960) Acuthan and Stradner 1969
N. quadrata (Bramlette and Sullivan 1961) Bukry 1973
N. crista (Martini 1958) Perch-Nielsen 1971 (= C. cristaus)
Neocheiostegyus Perch-Nielsen 1971
N. circatus (Bramlette and Sullivan 1961) Perch-Nielsen 1971
N. imbriici (Haq and Lohmann 1976)
N. juris (Bramlette and Sullivan 1961) Perch-Nielsen 1971
N. perfectus Perch-Nielsen 1971
N. primitives Perch-Nielsen 1981
N. saepes Perch-Nielsen 1971
Neoconocollithus Sujkowski 1931
N. dubius (Deflandre in Deflandre and Fert 1954) Black 1967
N. protensis (Bramlette and Sullivan 1961) Black 1967
Neocollithus Black 1967
N. dirimulous (Perch-Nielsen 1979) Perch-Nielsen 1981
N. fossus (Romein 1977) Romein 1979
Penna Klump 1953
P. basquesensis (Martini 1959) Báldi-Beke 1971
Placogygus Hoffman 1970
P. fibulariformis (Reinhardt 1964) Hoffman 1970
P. squamata (Bramlette and Sullivan 1961) Romein 1979
Postopora Deflandre 1902
P. exilis (Bramlette and Sullivan 1961) Romein 1979
P. multiloba (Kamptner 1984) Roth 1970
P. obliqua (Deflandre in Deflandre and Fert 1954)
Romein 1979
P. pectinata (Bramlette and Sullivan 1961) Sherwood 1974
P. plana (Bramlette and Sullivan 1961) Haq 1971
P. pulchra (Deflandre in Deflandre and Fert 1954) Romein 1979
P. punctata (Bramlette and Sullivan 1961) Perch-Nielsen 1984
Prediscosphaera Tekshina 1959
P. cretacea (Arkhangelsky 1912) Gartner 1968
P. grandis Perch-Nielsen 1973
P. pinus (Bramlette and Martini 1964) Gartner 1968
Prissius Hay and Mohler 1967
P. bisculus (Stradner 1963) Hay and Mohler 1967
P. dimorphous (Perch-Nielsen 1969) Perch-Nielsen 1977
P. tenificulum (Okada and Thierstein 1979) Perch-Nielsen 1984
Reinhardtites Perch-Nielsen 1968
R. levis Prins and Sissingh in Sissingh 1977
Retractus Black 1971
R. crenulata (Bramlette and Martini 1964) Grün in Grün and Allemann 1975
R. ficala (Stover 1966) Burnett 1998
Reticulofenestra Hay, Mohler and Wade 1966
R. dictyoda (Deflandre in Deflandre and Fert 1954) Stradner in Stradner and Edwards 1968
R. hampdenensis Edwards 1973
R. hillae Bukry and Percival 1971
R. minuta Roth 1970
R. reticula (Gartner and Smith 1967) Roth and Thierstein 1972
R. umbilicus (Levin 1965) Martini and Ritzkowski 1968
Rhabdolithus Kamptner ex Deflandre in Grasse 1952
Rhabdosphaera Haeckel 1894
R. gladius Locker 1967
R. influenza Bramlette and Sullivan 1961
R. tenus Bramlette and Sullivan 1961
Rhagodiscus Reinhardt 1967
R. angustus (Stradner 1963) Reinhardt 1971
R. reniformis Perch-Nielsen 1973
R. splendidus (Deflandre 1953) Verbeek 1977
Rhombaster Bramlette and Sullivan 1961
R. braamletti (Bönnimann and Stradner 1960) Bybell and Self-Traill 1995
R. cuspid Bramlette and Sullivan 1961
Sphenolithus Deflandre in Grasse 1952
S. anarrhopus Bukry and Bramlette 1969
S. conspicuus Martini 1976
S. furcatolithoides Locker 1967
S. moriformis (Bönnimann and Stradner 1960) Bramlette and Wilcoxen 1967
S. orphans (Perch-Nielsen 1971)
S. primus Perch-Nielsen 1971
S. pseudoradians Bramlette and Wilcoxen 1967
S. radians Deflandre in Grasse 1952
S. spiniger Bukry 1971
Staurolithus Caratini 1963
S. f. caratini Bukry 1998
S. integer (Bukry 1969) Burnett 1998
Tegumentum Thierstein in Roth and Thierstein 1972
T. stradneri Thierstein in Roth and Thierstein 1972
Thoracosphaera Kamptner 1927
T. operculata Bramlette and Martini 1964
Touvier Hay and Mohler 1967
T. africanaus (Perch-Nielsen 1980) Perch-Nielsen 1984
T. callosus Perch-Nielsen 1971
T. eminens (Bramlette and Sullivan 1961) Perch-Nielsen 1971
T. occultus (Locker 1967) Perch-Nielsen 1971
T. pertusus (Sullivan 1965) Romein 1979
Tranolithus Stover 1966
T. minimus (Bukry 1969) Perch-Nielsen 1984
T. orinatus (Reinhardt 1966) Perch-Nielsen 1968
T. transversapontis Hay, Mohler and Wade 1966
T. umbilicus (Bramlette and Sullivan 1961) Locker 1968
Tribrachiatus Shamrai 1963
T. bramlettei Deflandre 1963
T. orionatus (Stradner 1958) Bukry 1972 (= M. contortus)
T. digitalis Bukry 1996
T. orthostylus Sharmai 1963 (= M. tribrachiatus)
Umbilicosphaera Lohmann 1902
U. jordani Bown 2005
Watznaueria Reinhardt 1964
W. barnesi (Black 1971) Perch-Nielsen 1968
W. biporta Bukry 1969
W. fossacincta (Black 1967) Bown in Bown and Cooper 1989
W. ovata Bukry 1969
Zeugrhabdotus Reinhardt 1965
Z. bicrescenticus (Stover 1966) Burnett in Gale et al. 1996
Z. noeliae Rood et al. 1971
Z. sigmoides (Bramlette and Sullivan 1961) Bown and Young 1997
Z. spiralis (Bramlette and Martini 1964) Burnett 1998
Zeynerhabdotus Reinhardt 1965
Z. bicrescenticus (Stover 1966) Burnett in Gale et al. 1996
Z. noeliae Rood et al. 1971
Z. sigmoides (Bramlette and Sullivan 1961) Bown and Young 1997
Z. spiralis (Bramlette and Martini 1964) Burnett 1998
Zygodsiculites Bramlette and Sullivan 1961
Z. adamas Bramlette and Sullivan 1961
Z. braamletti Perch-Nielsen 1968
Zygolithites Deflandre 1959
Z. bijugatus (Deflandre in Deflandre and Fert 1954)
Deflandre 1959
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