Reda El Gammal and Orabi, J Geol Geophys 2019, 8:1
DOI: 10.4172/2381-8719.1000456

Journal of Geology & Geophysics

ISSN: 2381-8719

Research Article | Open Access

Coniacian-late Campanian Planktonic Events in the Duwi Formation, Red Sea Region, Egypt

Reda El Gammal MH[1] and Orabi H[2]*

[1]GeoMine Company, Egypt
[2]Geology Department, Faculty of Science, Menoufia University, Egypt

Abstract

This study is the first one on accurate identification of the planktonic foraminiferal biozones of the Duwi Formation and no similar study was carried out before. About 70 m thick complete succession of Coniacian-Late Campanian of the Duwi Formation at Gabal Duwi Range, Red Sea Coast, were completely examined and the following biozones have been recognized and correlated to the world bioevent; *Dicarinella concavata* Zone (Coniacian), *Dicarinella asymetrica* Zone (Santonian), where the two zones are correlated to the Coniacian–Santonian time interval of Oceanic Anoxic Event 3 (OAE 3), *Globotruncanita elevate* Zone (Early Campanian), which correlated to the Santonian/Campanian Boundary Event (SCBE), *Contusotruncana plummerae* Zone (Middle Campanian) which correlated with the Mid Campanian Event (MCE), *Radotruncana calcarata* Zone and *Globotruncana aegyptiaca* Zone (Late Campanian) of the Late Campanian Event (LCE).

Keywords: Gabal Duwi; Coniacian; Campanian; Planktonic foraminifera; Red Sea; Egypt

Introduction

During the last 10 years, the Late Cretaceous timescale has been improved by the integration of floating Astronomical Time Scales (ATS), higher resolution biostratigraphic frameworks and high-resolution carbon-isotope stratigraphy dates [1-5] of the Coniacian/Santonian and Santonian/Campanian Stage boundaries. The Coniacian/Santonian boundary has now been formalized, with a Global Boundary Stratotype Section and Point (GSSP) at Olazaguta, northern Spain [6] the defining marker is the lowest occurrence of the inoceramid bivalve *Cladoceramus undulatoplicatus* (Römer), which occurs widely in the northern hemisphere. However, detailed correlation with deep-water Tethyan successions (which generally lack inoceramid bivalves), where the base of the Santonian is traditionally taken at the lowest occurrence of the planktonic foraminifera *Dicarinella asymetrica* [7], is not currently possible. There is no formal agreement on either a marker or a type-locality for the Santonian/Campanian boundary GSSP, for which a series of potential markers have been discussed [8], including the Highest Occurrence (HO) of the crinoid Marsupites, the Lowest Occurrence (LO) of the planktonic foraminifera *Globotruncanita elevata* (Brotzen), and the LO of *Dicarinella asymetrica.*

Here, we present a correlation of new high-resolution planktonic datasets through the Coniacian and Upper Campanian of Gabal Duwi (Figure 1) and provide the first attempt to calibrate the new Regional sea level curve of Egypt [9] and Global Eustatic curves [10] to the Duwi Basin. This study brings new insights into the presence of condensed levels and potential hiatuses in existing Egyptian records and provides evidence of the position of the Santonian/Campanian boundary in the standard Tethyan biostratigraphic record of Gabal Duwi.

The Duwi Formation has not been formally defined, through various proposals have been made. For example, this formation has been assigned to Late Campanian age [11,12]. Meanwhile, it assigned to Maastrichtian age by Hegab [13], later on, assigned to Late Campanian-Maastrichtian as suggested by Schrank and Perch-Nielsen [14] Baiumy and Tada [15]. But up till now, there is no agreement on the age assignments of the fine subdivision of the Campanian rock units in the region.

In order to solve the problem of the age of the Duwi Formation, the planktonic foraminiferal assemblages have been analyzed and identified to allow us to recognize the faunal turnover, biostratigraphy and sea level changes during the Duwi Formation deposit.

The Coniacian-Santonian time interval is the inferred time of oceanic anoxic event 3 (OAE 3), the last of the Cretaceous OAEs. A detailed look on the temporal and spatial distribution of organic-rich deposits attributed to OAE 3 suggests that black shale occurrences are restricted to the equatorial to mid-latitudinal Atlantic and adjacent basins, shelves and epicontinental seas like parts of the Caribbean, the Maracaibo Basin and the Western Interior Basin, and are largely absent in the Tethys, the North Atlantic, the southern South Atlantic, and the Pacific. Here, oxic bottom waters prevailed as indicated by the widespread occurrence of red deep-marine CORBs (Cretaceous Oceanic Red Beds). Widespread CORB sedimentation started during the Turonian after Oceanic Anoxic Event 2 (OAE 2) except in the Atlantic realm where organic-rich strata continue up to the Santonian. The temporal distribution of black shales attributed to OAE 3 indicates that organic-rich strata do not define a single and distinct short-time event, but are distributed over a longer time span and occur in different basins during different times. This suggests intermittent and regional anoxic conditions from the Coniacian to the Santonian [16,17].

Palaeontological and geochemical evidence increasingly indicates that the Coniacian–Santonian interval represents the transition from the mid-Cretaceous extreme greenhouse characterized by elevated temperatures, increased volcanic activity, high sea-level and regional

***Corresponding author:** Orabi H, Geology Department, Faculty of Science, Menoufia University Egypt, Tel: +201061940775; E-mail: oraby1952@yahoo.com

Received December 12, 2018; **Accepted** January 12, 2019; **Published** January 18, 2019

Citation: Reda El Gammal MH, Orabi H (2019) Coniacian-late Campanian Planktonic Events in the Duwi Formation, Red Sea Region, Egypt. J Geol Geophys 8: 456. doi: 10.4172/2381-8719.1000456

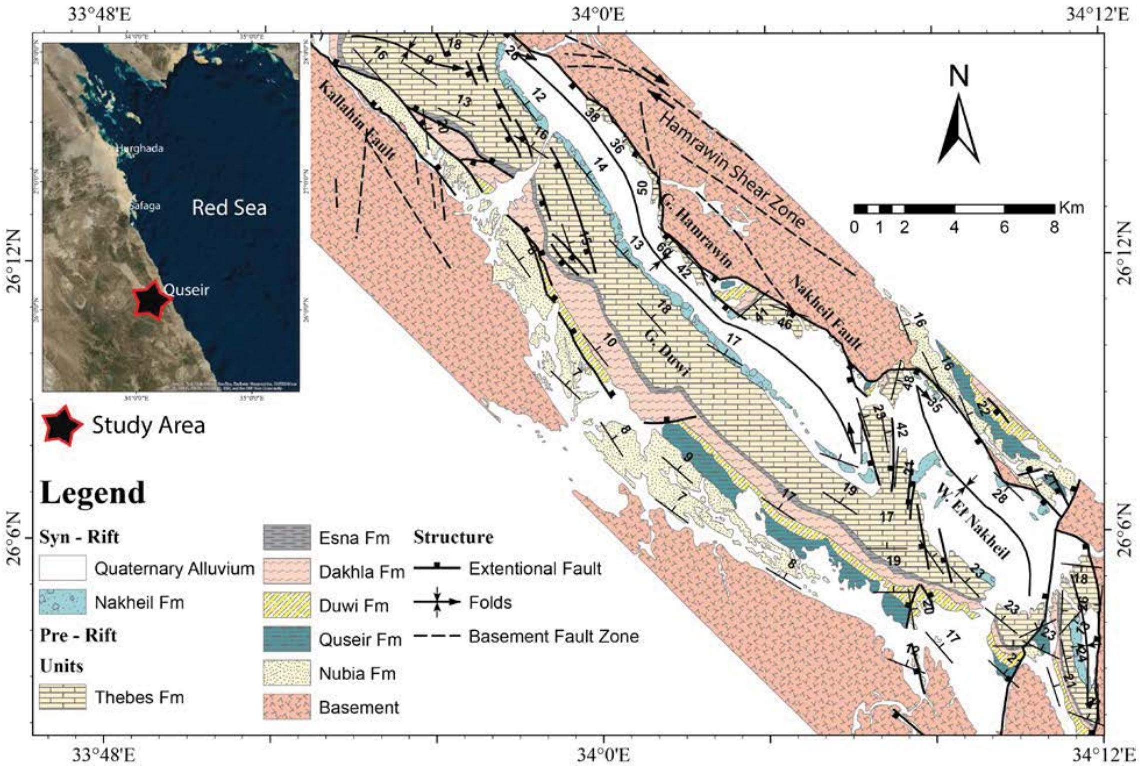

Figure 1: Geological map of Gabal Duwi range.

to global ocean anoxic events, to more temperate climatic conditions in the Campanian-Maastrichtian [18-20]. Coincident with this climatic transition, planktonic foraminiferal assemblages underwent major compositional changes due largely to a pulse of diversification.

Causes of the Santonian turnover, however, are still a matter of debate. Hypotheses proposed to suggest that it could be related to the following:

(i) tectonically forced changes in surface-water and deep-water circulation [7]; (ii) the onset of the Late Cretaceous cooling trend during the late Santonian Patrizzo, [21] combined with taxa competition within particular depth habitats [22]; and (iii) the development of minor and regional anoxic events (i.e. Oceanic Anoxic Event 3; e.g. [23-25] in the Atlantic and adjacent epicontinental sea, which resulted in the enlargement of the ecological niches in the more oxygenated Tethys, Pacific and Indian oceans [16].

Long-term cooling during the remainder of the Late Cretaceous Friedrich et al. [17]; Linnert et al. [20], where this cooling trend accelerated during the beginning of the Campanian [17,20], but its mechanisms and dynamics are not yet well understood.

The Campanian is also characterized by significant fluctuations of the sea level [10,26,27], a major shift in the δ15 N of marine organic matter [28], clay mineralogical changes, and the occurrence of positive and negative carbon isotope events: the Santonian/Campanian Boundary Event (SCBE) [1,27], the Mid Campanian Event (MCE) [1,3,27], the Conica Event [29], the Late Campanian Event (LCE) [1,3,4,5,27], the Epsilon Event (EE) (also called C1-Event) [3,30], and the Campanian/Maastrichtian Boundary Event (CMBE) [3,5,30].

The aim of the present study is to present an illustrated record for the planktonic foraminifera of the study area which has never been attempted before and to discuss the chronostratigraphy of the foraminiferal interval of the investigated locality in other Egyptian occurrences in the light of the application of up to date biostratigraphical standards. The study of the biostratigraphy of Campanian planktonic foraminifera in Egypt started in connection to the solution of the problem of tracing exactly the Santonian/Campanian boundary in the region.

Geological Setting

The northwestern Red Sea-Gulf of Suez rift consists of four distinct sub-basins as half graben separated by complex accommodation zones [31]. Each sub–basin is asymmetric and bounded on one side by a major NW trending border fault system, Duwi sub-basin is the largest one (Figure 1).

The pre-rift sediments of the northwestern margin of the Red Sea are filled these hanging wall synclines by Paleozoic, Cretaceous, Paleogene sediments [32,33]. The basement complex is unconformably overlain by the 500-700 m thick of Pri-Rift sediments. The lower part of these sediments is of siliciclastic Nubian facies, overlain by the 350-570 m thick of interbedded shales, sandstones and limestones of the Quseir, Duwi, Dakhla, Tarawan, Esna and Thebes formations [34,35]. Duwi area can be considered as tilted faulted blocks of the Red Sea tectonics dissected by major and minor faults (Figure 1). A major phosphogenic episode took place during Late Cretaceous of shallow to the epicontinental shelf of the southern Neo-Tethys Ocean [36]. The common association of phosphatic strata with chert and organic-rich sediments (black shales in Duwi) in both the Middle East and Egypt

Citation: Reda El Gammal MH, Orabi H (2019) Coniacian-late Campanian Planktonic Events in the Duwi Formation, Red Sea Region, Egypt. J Geol Geophys 7: 456. doi: 10.4172/2381-8719.1000456

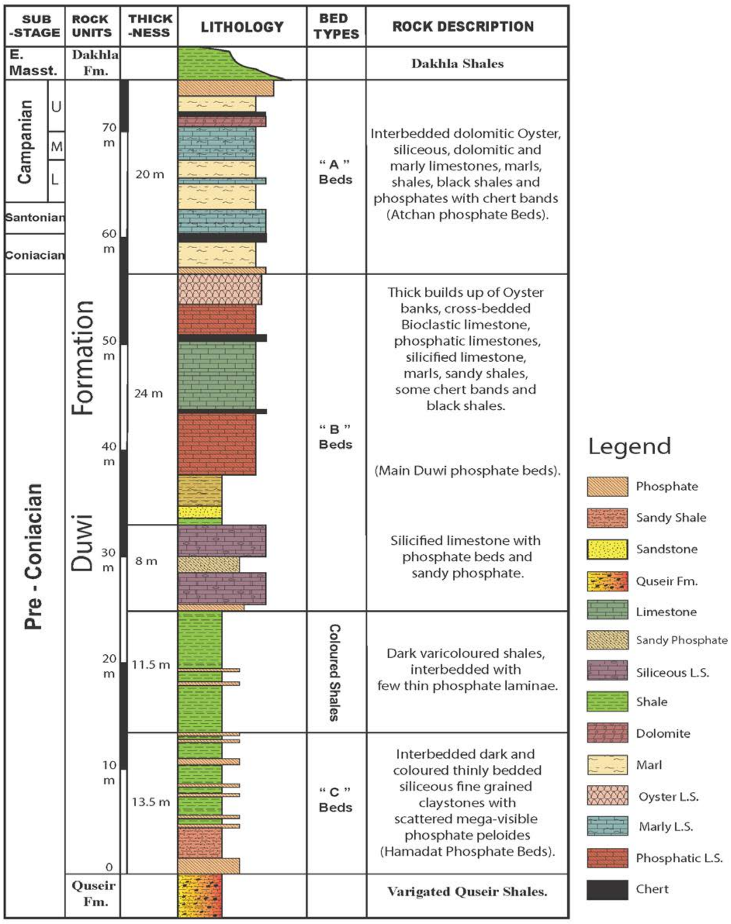

Figure 2: Lithostratigraphic subdivision of the Late Cretaceous. Duwi Formation, Red Sea, Egypt.

has been interpreted as an indication that phosphorite accumulation association with highly productive surface waters possibly caused by upwelling currents [37].

The Duwi (Phosphate) Formation (Figure 2) was laid down in extremely shallow Epicontinental-Neritic seas which flanked the southern margin of the Tethys sea in Egypt that extended across the northern margin of the Arabo-Nubian Craton and deeper towards the Tethyan seaway to the north.

Materials and Methods

The present study has focused on the 77 m thick of the total interval of the Duwi Formation at Gabal Duwi Range. This succession is well exposed at the type section which is easily accessible with its contact with Lower Quseir Formation and the Upper Dakhla Formation.

The entire interval includes the transition from the Coniacian stage to the Campanian stage of the Duwi Formation. Each marl, shale, silt and soft bioclastic carbonate beds are sampled for foraminiferal analyses, while the very hard rocks as bioclastic and silicified limestones, chert and phosphatic beds are not sampled. Throughout the Duwi Formation interval, a total of 80 samples are collected from Pre-Coniacian-Late Campanian. The sampling system is doomed to rock nature where the general hardness nature of the rocks throughout all intervals exacerbated the problem.

The samples were soaked in H_2O_2 and water with some heating for several days and then washed through a 63 µm sieve with tap water. Planktonic foraminiferal species were picked from the residues, identified, mounted on covered micro slides for a permanent record.

The present work (77 m thick) yielded more than 70 species belonging to 27 genera of planktonic foraminifera suitable for the establishment of chronozones for worldwide correlation. The identified species are listed in Figure 3. Only the very important, as well as the index species, are scanned and illustrated (Figures 4-6). The material has been deposited in the Museum of the Geology Department, Menoufia University.

Stratigraphy

In the Quseir-Safaga District, the Cretaceous–Eocene deposits occupy the troughs of synclinal–like folds within the crystalline basement hill ranges. The Cretaceous sections especially Duwi Formation are described by many authors focusing on the phosphate formations and black shales [33-35,37,38].

At its type locality at Quseir, Youssef [34] give to the Duwi Formation a Campanian age. In the south of the Western Desert, Luger [39], Farouk [40] consider that the Duwi Formation is of Late Campanian to early Maastrichtian age. Stratigraphically, the Duwi Formation is underlain by the Quseir Formation (Quseir Variegated Shale) and overlain by the Dakhla Formation.

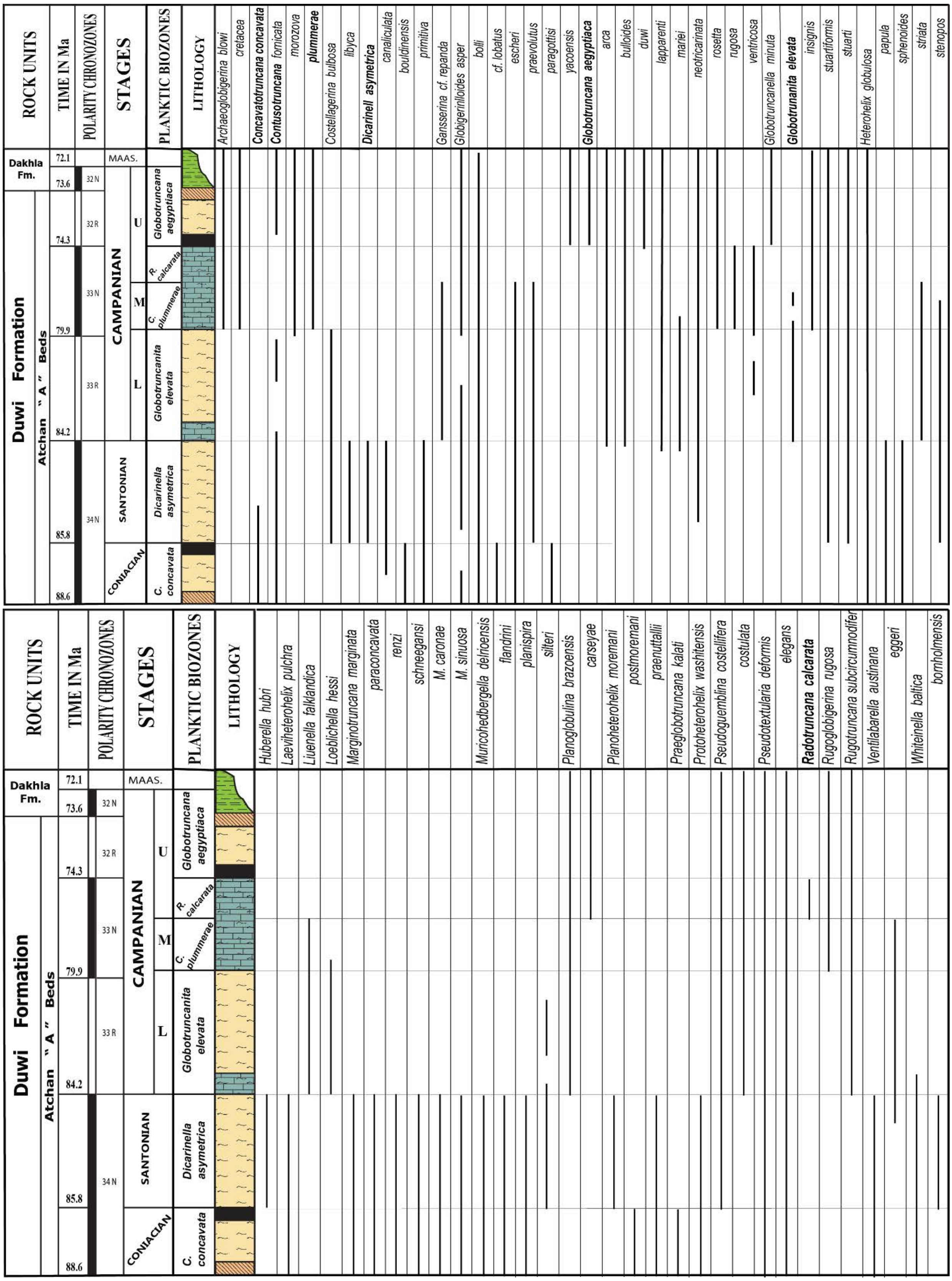

Figure 3: Coniacian-late Campanian planktonic foraminiferal species distribution chart, upper part of the Duwi Formation, Gabal Duwi, Egypt. Specimens in bold are index fossils for the identical biozones.

Citation: Reda El Gammal MH, Orabi H (2019) Coniacian-late Campanian Planktonic Events in the Duwi Formation, Red Sea Region, Egypt. J Geol Geophys 7: 456. doi: 10.4172/2381-8719.1000456

The rocks of the Duwi Formation overlie marginal marine to shallow marine shales of the Quseir Formation which overlie the Nubian Formation [33] and underlie deeper-marine marls and carbonates of the Maastrichtian Dakhla Formation [33]. Duwi Formation is considered the first of fully marine transgression of the Egyptian Late Cretaceous.

The lithostratigraphic subdivision of Gabal Duwi emerged after long series of investigations which started from the beginning of the century [33,34,37,41].

The following litho-stratigraphic subdivisions (Figure 2) for the Pre-Coniacian-Late Campanian in Gabal Duwi are proposed, from the base to the top as follow:

Lower phosphate unit (Pre-Coniacian): It attains 13.5 m thick named Abu Shigeila or Hamadat “C” horizon [33]. It is made of interbedded dark colored thinly bedded siliceous fine-grained claystones with scattered mega-visible phosphate peloids. It is underlain by Quseir Formation of Nubian facies which attains 11.5 m thick of dark varicolored shales with very thin two phosphate lamina.

Middle phosphate unit (Pre-Coniacian): It is about 32 m thick named Main Duwi Beds or “ B ” beds. The lower 8 m thick is made of unfossiliferous silicified limestone with phosphate beds and sandy phosphate. The upper 24 m is made of thick builds up of the oyster bank, cross-bedded bioclastic limestone covered phosphatic limestone, silicified limestone, marly, sandy shales, some chert bands, ochreous shale and black shale bands.

Upper phosphatic unit (Coniacian-Late Campanian): It is named locally Atchan phosphate or “A” Beds, its thickness 20 m of more marly limestone, marls and shale beds, the dolomitic oyster band near the top. It is capped by a conglomeratic phosphate bed (50 cm thick) marked the top of the Duwi Formation. The direct overlying shale beds are belonging to the Dakhla Formation.

In general, the Duwi Formation is mainly hard varieties of silicified phosphatic oyster beds builds up with chert bands and black shales, few and thin intervening throughout scattered marly, shale and marly limestone bands.

The Duwi Formation had a large number of vertebrate remains, coprolites, fish teeth and heteromorph ammonites [33]. Dominik and Schall [42] assigned these faunal assemblages to the Bostrychoceras polypolocum Zone.

Biostratigraphy

All of the samples collected from the informal member (A–Beds)

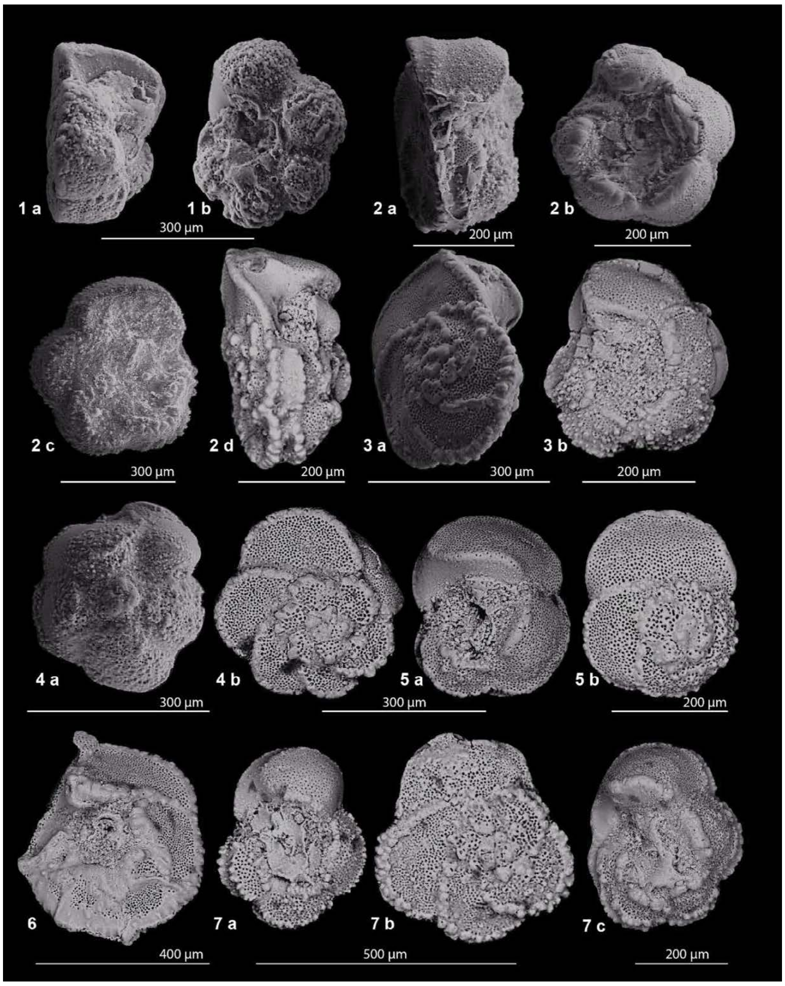

Figure 4: SEM Images for foraminiferal species; **1a, 1b**: *Concavatotruncana concavata*; **2a-2d**: *Dicarinella asymetrica*; **3a, 3b:** *Globotruncanita elevata;* **4a, 4b:** *Globotruncana ventricosa*; **5a, 5b:** *Contusotruncana plummerae;* **6:** *Radotruncana calcarata;* **7a-7c:** *Globotruncana aegyptiaca*.

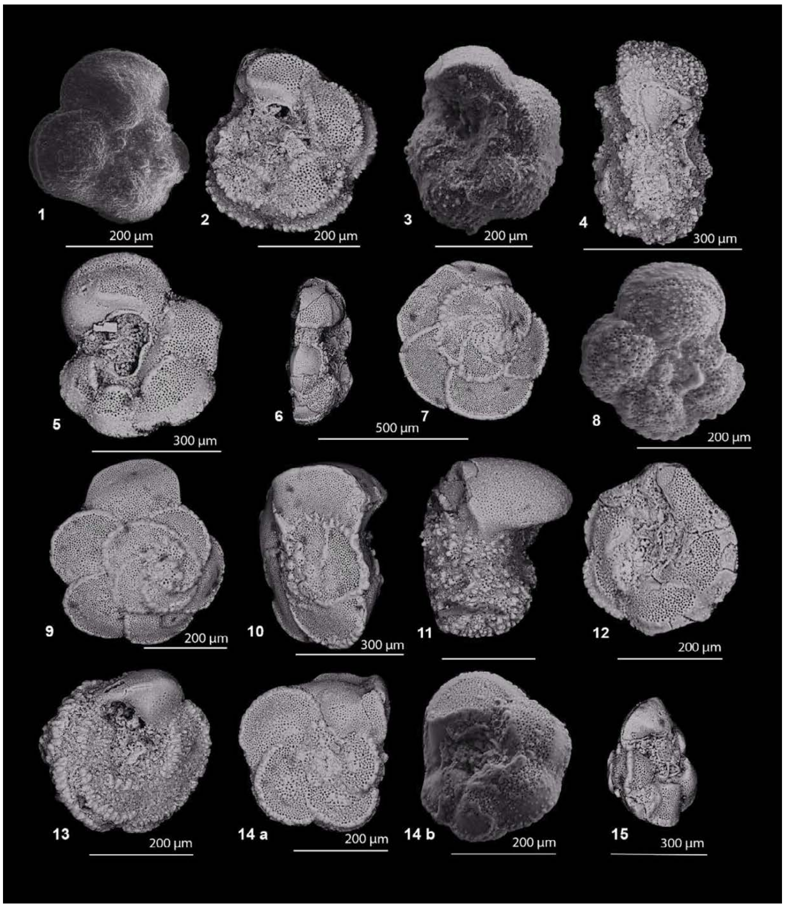

Figure 5: SEM Images for foraminiferal species; **1:** *Dicarinella* primitive; **2:** *D. canaliculata*; **3:** *Marginotruncana paraconcavata*; **4:** *Marginotruncana caronae*; **5:** *Marginotruncana sinuosa*; **6:** *Marginotruncana renzi*; **7:** *Globotruncanita stuarti;* **8:** *Praeglobotruncana kalaati*; **9:** *Globotruncanita stuartiformis*; **10:** *Globotruncanita insignis;* **11:** *Gansserina cf. repanda*; **12:** *Contosutruncana morozova*; **13:** *Contosutruncana fornicate*; **14a,14b:** *Globotruncana bulloides*; **15:** *Globotruncana arca.*

of about 20 m thick of the Duwi Formation section forming several assemblages of moderate to well preserved taxa including small biserial and large multiserial heterohelicids, trochospiral keeled, unkeeled and very rare planispiral *Globigerinelloides*.

The lower beds (about 57 m thick) of the Duwi Formation are barren interval including (B-beds and C-Beds) (Figures 2 and 7a), (Oxygen, Anoxic Environment) including rhythmically bedded dark shales and black shales facies with phosphatic beds and lamina [15,37,41]. The upper part which overlying the B-beds had black shale and marl interval (7.7 m thick) (Figures 7d), which may coincide with the Coniacian–Santonian time interval OAE 3 of [23-25,43].

The planktonic foraminiferal assemblages of studied upper 20 m thick of the Duwi Formation are diverse and from low to middle abundance (Figures 3-6). Several low latitude Tethyan zonal schemes for the Upper Cretaceous have been previously as [7,44-46]. The following 6 standard Tethyan biozones covering the sedimentary sequence of the upper 20 m thick of the interval (A-beds) are from older to younger as follows (The Upper Cretaceous (CF) numerical zonal scheme of Li and Keller [47,48] is used.

Dicarinella concavata zone (CF 14)

Definition: Interval zone from the FO of *Dicarinella concavata* (Brotzen) to the FO of *Dicarinella asymetrica* (Sigal).

Author: Sigal [49] as Globotruncana concavata Zone.

Age: Coniacian to early Santonian.

Thickness: 4.3 m equivalent to the lowermost marl beds of the A-Beds that overlies the barren interval of may propose Pre-Coniacian interval (Figures 3 and 8) of Coniacian–Santonian time interval of Oceanic Anoxic Event 3 (OAE 3).

Assemblage: *Dicarinella concavata* is not abundant in its individuals. Some individuals of *Whiteinell baltica* are present within this zone might probably of Turonian span. Species as *Marginotruncana rensi*, *M. marginata, M. schneegansi, M. paraconcavata, Globigerinelloides paragottisi, G. esceri, G. cf. lobatus, G. bolli, G. asper, Muricohedbergella fandrini, Murico. planispira, Murico delrioensis, Heterohelix globulosa, Hx. papula, Hx. sephenoides, Pseudotextularia deformis, P. elegans, Dicarinella canliculata, Contusotruncana fornicate, Ventilabarella austiniana* and others.

Remarks: According to the diverse of shell shapes as marginotruncanids, dicrinellids and concvatotruncanids groups evolved from Praeglobotrucana [50,51], cited concavata species within the genus *Dicarinella* where the species of *Dicarinella* have an umbilically convex shell with closely positioned keels near the dorsal side. The extinction of *C. concavata* within the nearly lowest appearance of *Dicarinella asymertrica* correlates with observations of Petrizzo [52]

Citation: Reda El Gammal MH, Orabi H (2019) Coniacian-late Campanian Planktonic Events in the Duwi Formation, Red Sea Region, Egypt. J Geol Geophys 7: 456. doi: 10.4172/2381-8719.1000456

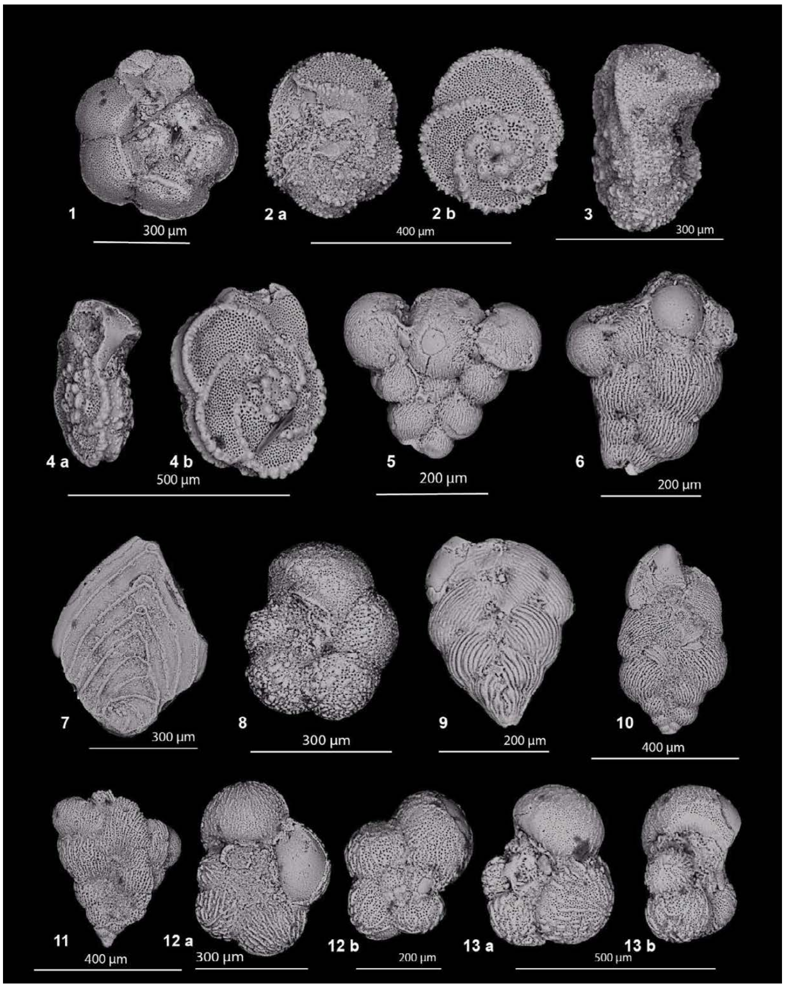

Figure 6: SEM Images for foraminiferal species; **1:** *Globotruncana rosetta*; **2a, 2b:** *Globotruncana duwi*; **3:** *Archaeglobigerina cretacea*; **4a, 4b:** *Globotruncana neotricarinata*; **5:** *Ventilabrella eggeri;* **6:** *Ventilabrella austinana*; **7:** *Neoflabellina suturalis.* **8:** *Costallegrina bulbosa*; **9:** *Pseudoguemblina cotulatal*; **10:** *Planoheterohelix moremani*; **11:** *Planogloulina carseyae*; **12a, 12b:** *Rugoglobigerina rugosa*; **13a, 13b:** *Rugoglobigerina macrocephala.*

in southern Tanzania, South Spain, Petrizzo [53] in Exmouth Plateau, Indian Ocean, in the Tethyan Tunisia [54] and in southern Tibet [55].

Dicarinella asymetrica zone (CF 13)

Definition: Total range of *Dicarinell asymetrica* (Sigal).

Author: Postoma [56] from Tethys region as *Globotruncana concavata* carinata Zone.

Age: Santonian

Thickness: 3.4 m equivalent to marly limestone beds that overlie the Coniacian marls (Figure 7e).

Assemblage: The nominate taxon is consistently present but not abundant. Most taxa of the *C. concavata* Zone are present in *D. asymetrica* Zone except *C. concavata* or any form of *Sigalia* spp.

Still, the occurrence of *Contusotruncana fornicate*, *Whiteinella baltica* and *W. bornholmensis* and all *Margiotruncana* spp. The first appearance of *Muricohedbergella silteri, Globotruncana bulloides, G. arca* and *Ventilabarella eggeri* are marked this zone. The total range of *Huberella hubris, laeviheterohelix pulchra* and *Planoheterohelix moremani, Heterohelix papula* and *Hx. sphenoides* are extinct in Santonian.

Remarks: El Gammal [57], Obidalla and Kassab [58], Farouk and Faris [59] and many others reported the occurrence of the Santonian *marker Dicarinella asymetrica* of the Matulla Formation in Southern and Northern Sinai [9]. Aref and Ramadan [60] recorded D. asymetrica also in the Matulla Formation of Esh El Mallaha Range, Gulf of Suez.

Globotruncanita elevate zone (CF 12)

Definition: Partial range zone, recognized from the last occurrence of *Dicarinella asymetrica* and the first occurrence of *Globotruncana ventricosa* White.

Author: Dalbiez [61].

Age: Early Campanian

Thickness: About 4.5 m thick equivalent to marl interval overlie the Santonian marly limestone (Figure 8e).

Assemblage: It is characterized by abundance and more species diversification of both globotruncanids and heterohelicids. Taxa

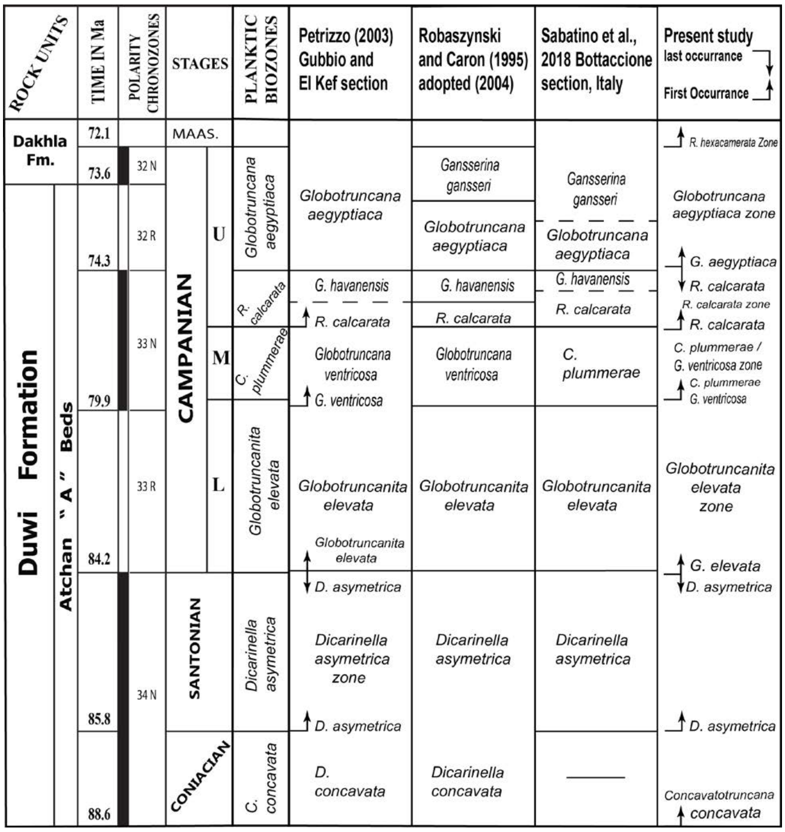

Figure 7: Comparison of chronostratigraphic assignment of the Coniacian–Late Campanian Tethyan biozones recorded in Tunisia, Europe–Mediterranean, Italy and present study.

as *Globotruncanita stuartiformis, Globotruncana bulloides, G. arca, G. neotricarinata, G. lapparenti, Contusotruncana fornicata* and *Marginotruncana marginata* are very rare. It is marked also by the extinction of both *Muricohedbergella flandrini* and *Dicarinella asymetrica*.

Remarks: Some authors as [62,63] have equated the first appearance of *Globotruncanita elevata* with different levels of the Santonian Stage in North Africa and Mediterranean regions. Recent works show the base of the Campanian [64-66] and present work is marked by first appearance of *Globotruncanita elevata* taxon. In Egypt, many authors reported previously that *Globotruncanita elevata* is marked the base of the Campanian Stage in most localities [9,58,60,67,68,].

The Santonian/Campanian Boundary Event (SCBE) of Jarvis [1,27] is well recognized in both *Dicarinella asymetrica* Zone and *Globotruncanita elevate* Zone by the extinction of *Heterohelix papula* and *Hx. sphenoides* of Santonian and by the extinction of *Muricohedbergella flandrini* and *Dicarinella asymetrica* of Early Campanian.

Contusotruncana plummerae zone (CF 11)

Definition: The interval from the first occurrence of the *Contusotruncana plummerae* to the first occurrence of *Radotruncana calcarata*. Duration of about 3.2 Ma is inferred for this zone [69].

Author: Petrizzo [69] in tropical and subtropical areas.

Age: Middle Campanian

Thickness: About 3 m thick (Figures 7c and 7e).

Assemblage: *C. plummerae, C. fornicata, Rugotruncana subcircumnodifer, Globotruncana ventricosa, G. arca, G. bulloides, Heterohelix striata, Liunella falklandica, Archaeoglobigerina blowi, A. cretacea, Rugoglobigerina rugosa* and very rare individuals of *Globotruncanita elevate*.

Remarks: [7,44,46,54,69] proposed the *Globotruncana ventricosa* Zone as interval between Gta. elevata and first appearance of *Radotruncana calcarata* in Tethyan Campanian foraminiferal zonation. Petrizzo [69] discussed unreliability of *Globotruncana ventricosa* for several reasons and introduced *Contusotruncana plummerae* as first appearance as zonal marker for tropical and subtropical regions instead of *G. ventricosa* to first occurrence of *Radotruncana calcarata*. Besides that, the exact position of *G. ventricosa* stratigraphically is imprecise [53,70], the Conica Event [29]. Farouk and Faris [59] recorded *G. ventricosa* over the *Globotruncanita elevata* in the shallow marine carbonate platform in the Mitla Pass, Northern Sinai.

Radotruncana calcarata zone (CF 9 and CF 10)

Definition: Total range zone of *Radotruncana calcarata* taxon.

Author: Dalbiez [61].

Age: Early Late Campanian

Thickness: About 2.4 m thick (Figure 8c).

Assemblage: In general, the Late Campanian was identified as a time of planktonic foraminiferal turnover [71]. No major change is observed in planktonic communities and no evolutionary trend of any

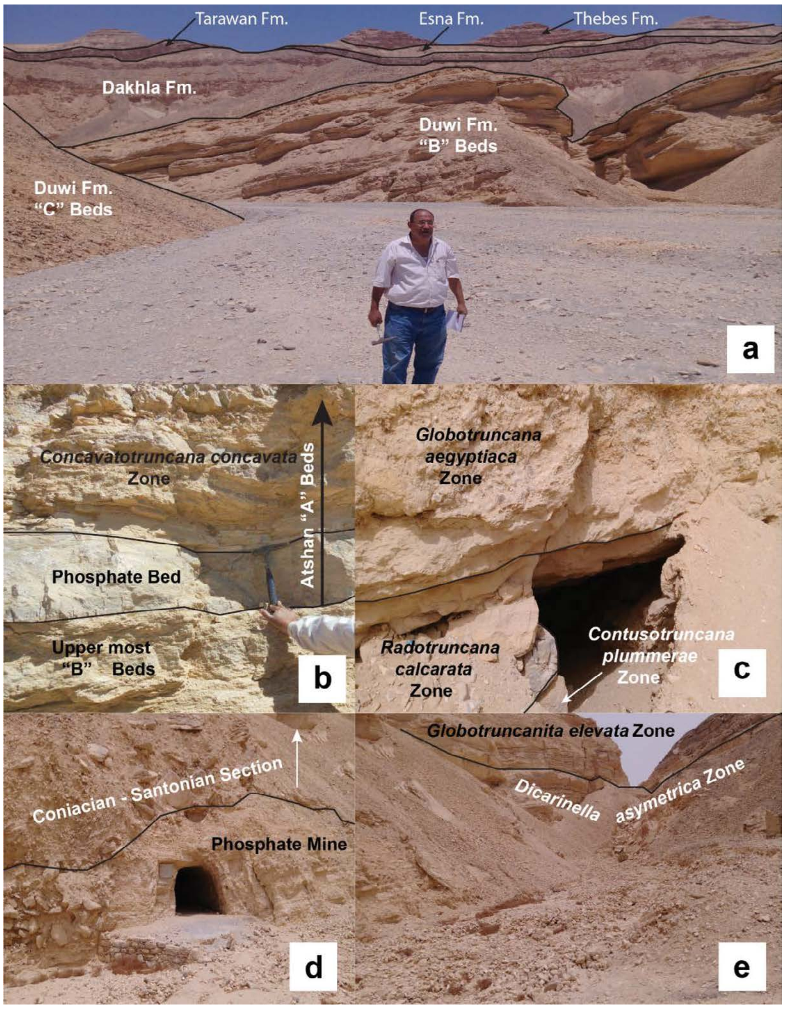

Figure 8: Field photographs; **(a)** General photo showing some of stratigraphic boundaries of Gabal Duwi; **(b)** Field zonal boundaries, *Contusotruncana plummearae, Radotruncana calcarata, Globotruncana aegyptiaca;* **(c)** Field boundary between B-Beds and A-Beds of the Duwi Fm. and Zone; **(d)** Contact between last mined phosphatic bed of B-Beds and the direct overlying Coniacian-Santonian sequence of A-Beds; **(e)** Relation between *Dicarinella asymetrica* Zone and the overlying *Globotruncanita elevata* Zone.

noticeable environmentally influenced change in the composition. It is considered a stable interval in terms of paleoecology despite the general trend of cooling of Cretaceous Greenhouse. The distinctive morphology of *R. calcarata* most certainly suggests an adaptation to environmental changes may relate to hydrodynamic regime [72].

Remarks: *Radotruncana calcarata* biozone is considered the narrowest biozone of the present study. Georgescu [73] suggested this case and support the hemipelagic to pelagic setting [74]. In the present material of the Duwi Formation, *Globotruncana aegyptiaca* Nakkady is recorded at the upper level of the *R. calcarata* as primitive forms where *Praeglobotruncana (Globotruncanella) havanensis* is not seen. Therefore, this biozone encompassed interval of both *R. calcarata* and *Praeglobotruncana havanensis* zones.

Globotruncana aegyptiaca zone (CF 8a)

Definition: Interval zone, from the last occurrence of *Radotruncana calcarata* at the base to the first appearance of *Rugoglobigerina hexacamerata Bronnimann* at the top.

Author: Caron [44].

Age: Late Campanian

Thickness: about the last 4.2 m thick of uppermost marl bed of the Duwi Formation (A-Beds) under the chert band which rests over the last conglomeratic siliceous phosphatic band (Figure 8c). The conglomerate bed shows the Late Campanian Event (LCE) [1,3-5,27].

Assemblage: Abundance of *Heterohelix reussi, Hx. globulosa, Globotruncanella minuta, Rugoglobigerina rugosa, Globotruncanita insignis, Rugotruncana subcircumnodifer, Globotruncana duwi*, flood of *Globotruncana aegyptiaca* and other *globotruncanids, Globgerinelloides yacoensis* and *Pseudoguemblina costulata* are present.

Remarks: The interval of the *Globotruncana aegyptiaca* Zone of the Upper Cretaceous of the Gabal Duwi Range is of two parts (Figure 9b). The marly lower one belonged to the Duwi Formation and the other upper part is of lower shale beds belongs the Lower Maastrichtian Dakhla Formation (Figures 9a and 9b).

Later works have documented G. *aegyptiaca* Zone as Late Campanian age [7,47,48,75,76]. The maximum abundance and diversification of globotruncanids of this biozone coincide with global abundance data presented by Premoli Silva and Sliter [77]. Smith and Pessagno [78], in their study on the Corisican Formation, Texas, USA, introduced *Globotruncana aegyptiaca* Zonule, as a lower biounit within their Gansserina (Globotruncana) gansseri Subzone of the *Globotruncana contuse-stuartiformis* Assemblage Zone. They differentiated the *G. aegyptiaca* Zonule from the underlying Early *Maastrichtian Rugotruncana subcircumnodifer* Subzone. Smith and Pessagno [78] defined the base of *G. aegyptiaca* Zonule by the first occurrence of the maker species with the limit distribution of *Gansserina gansseri (Bolli)* at the lower level. In Egypt, Aref [79] reported *G. aegyptiaca* Zone in the lower part of the Dakhla Formation of some Red Sea coastal areas, conformably overlain by *Gansserina gansseri* Zone (Duwi, Safaga and Esh El Mellaha sections) within the Matulla and partly Sudr Formation and unconformably underlain by the *Globotruncana ventricosa*. El-Deeb and El Gammal [80] recorded *G. aegyptiaca* Zone in southern Sinai and correlated its occurrence with Central Egypt and Red Sea regions. This zone is correlated to the Late Campanian Event (LCE) [1,3-5,27].

Results

Stage boundaries

The present taxonomic analysis has revealed several traceable levels that show changes in the composition of the planktonic foraminiferal assemblages belonged to the upper 20 m thick of the Duwi Formation through the Coniacian–Santonian and Campanian Stages with comparable Tethyan events (Figure 10).

Oceanic anoxic event 3: Oceanic Anoxic Event 3 (OAE 3) was designated by Arthur and Schlanger [24] and occupied the *Dicarinella concavata* Zone of the earliest event, the sharp contact between first lower marl bed of "Atchan Phosphate A-Beds" where the first appearance of *Dicarinella concavata* and its assemblage of the Coniacian markers. oyster phosphate thin bed (Figure 7b). The late Coniacian Event [25,81] has marked by the boundary between the Coniacian/Santonian Stages (Figures 9c and 9d). It is indicated by the first occurrence of both *Dicarinella asymetrica* and the first appearance of very rare benthic *Neoflabelin suturalis* [82]. Premoli Silva and Sliter [7,46] and others accepted the first appearance of *Dicarinella asymetrica* as a zonal marker of the Santonian Stage in low latitude globotruncanid zonal

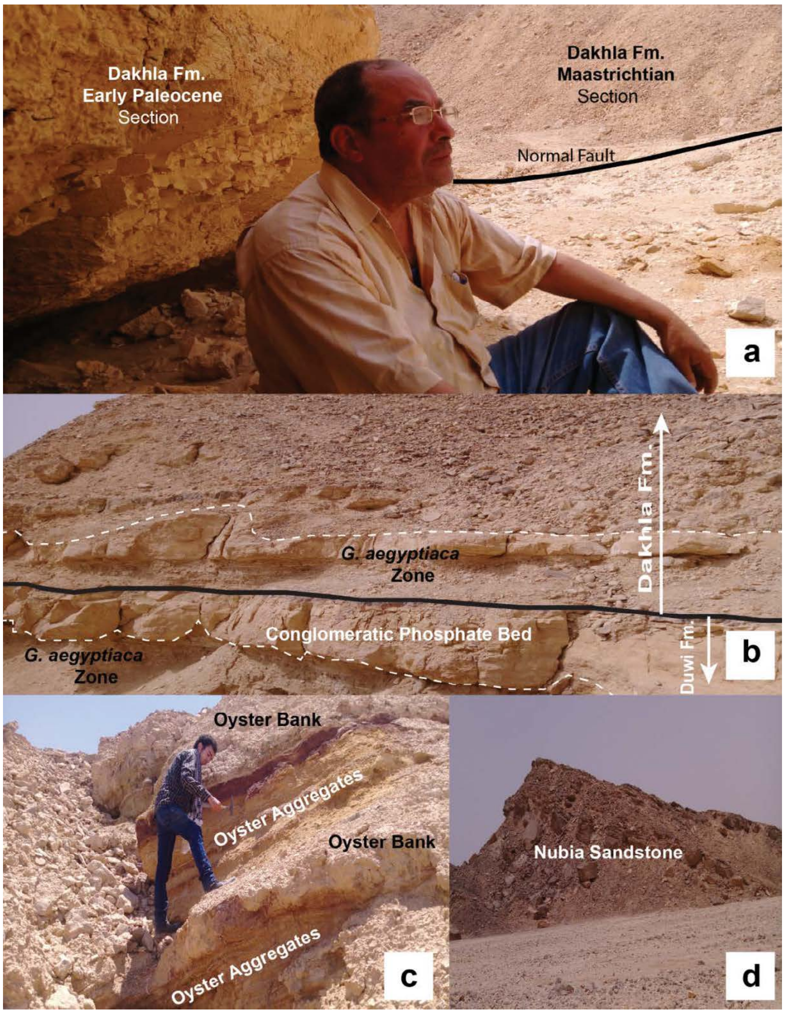

Figure 9: Field Photographs; **(a)** The Dakhla Formation, lower Maastrichtian and lower Paleocene; **(b)** Field relation between the Duwi and Dakhla Fm., two parts of the *Globotruncana aegyptiaca* separated by a last phosphate bed; **(c)** Oyster bank and oyster aggregates; **(d)** Part of the Nubian Sanstone sequence rounded the Duwi Basin.

Citation: Reda El Gammal MH, Orabi H (2019) Coniacian-late Campanian Planktonic Events in the Duwi Formation, Red Sea Region, Egypt. J Geol Geophys 7: 456. doi: 10.4172/2381-8719.1000456

schemes. The disappearance of *Concavotruncana concavata is* reported at the same level of the first appearance of *Dicarinella asymetrica* but the highest occurrence of *C. concavata* cannot be used as marker event for the Asymetrica event [52].

The Coniacian–Santonian phase is followed by the extinction of *Marginotruncana* and *Dicarinella* in the Latest Santonian-Earliest Campanian, the two evolutionary steps are traditionally regarded as due to a broader, major turnover (the so-called Santonian turnover) that affected all trophic groups within the planktonic foraminifera but was especially important in the history of more oligotrophic, keeled taxa [62,77].

Santonian/Campanian Boundary (SCBE): The Santonian/Campanian Boundary Event (SCBE) [1,27] is marked by major faunal turnover of *margintruncanids* were replaced by globotruncanids [46,83,84] confirm the cooling event of the lower Campanian. The first appearance of *Globotruncanita elevata* as indicated the marker of this boundary means prograding to oligotrophic episodes [85]. The lower Campanian shows the real diversity and radiation of globotruncanids, contusotruncanids and many foraminferal species in Northern Africa and Mediterranean regions [63,64], so at this boundary, many of mid Cretaceous planktonic foraminiferal taxa extincted globally [21,63]. The definition of Santonian/Campanian boundary is still mater of discussion [86-88].

Mid-Campanian Event (MCE): This event named the Mid Campanian Event (MCE) [1,3,27] and/or the Conica Event [29] at the base of the upper Campanian and recently as (Trunch event) by Thibault [86]. The Mid-Campanian Event at 78.7 Ma may equate to the high level of *Globotruncana ventricosa* Zone. The mid Campanian transgression therefore, corresponds to the Mid-Campanian Event [27] as leading to a Campanian sea level maximum to the top of *Contusotruncana plummerae* Zone of Middle Campanian. On another hand, some factors worked as a divide between the Boreal realm and the Tethyan basin [89] specifically the connection between the Boreal realm, the proto-Atlantic Ocean and the Tethyan basin became narrow gateways, progressively more restricted during the mid-Campanian possibly due to northward movement of Africa associated with the effects of the South Atlantic Ocean opening with consequent reduced intermediate/deep water mass exchange. This could cause a change in the Tethyan fauna as a result of changes the ecological supply of niche and large number of Campanian species of genera *Planoglobulina, Contusotruncana, Rugoglobigerina*, and *Pseudoguemblina* were adopted [90].

As in Duwi section, diversification of keeled species of genera like *Globotruncana, Radotruncana, Contusotruncana* show strong fluctuations and the assemblages became strong affinity with typical Tethyan assemblages [7,91]. The Late Campanian regression and subsequent sea level rise correspond to the present *Radotruncana calcarata* taxon of the Duwi section may suggest sediment starvation during sea level rise (note that the rate of sedimentation of *R. calcarata* Zone is calculated as 0.06 cm/kyr. (Figure 11), this supported by weakly phosphatized.

Late Campanian Event (LCE): During the *R. calcarata* Zone**,** the

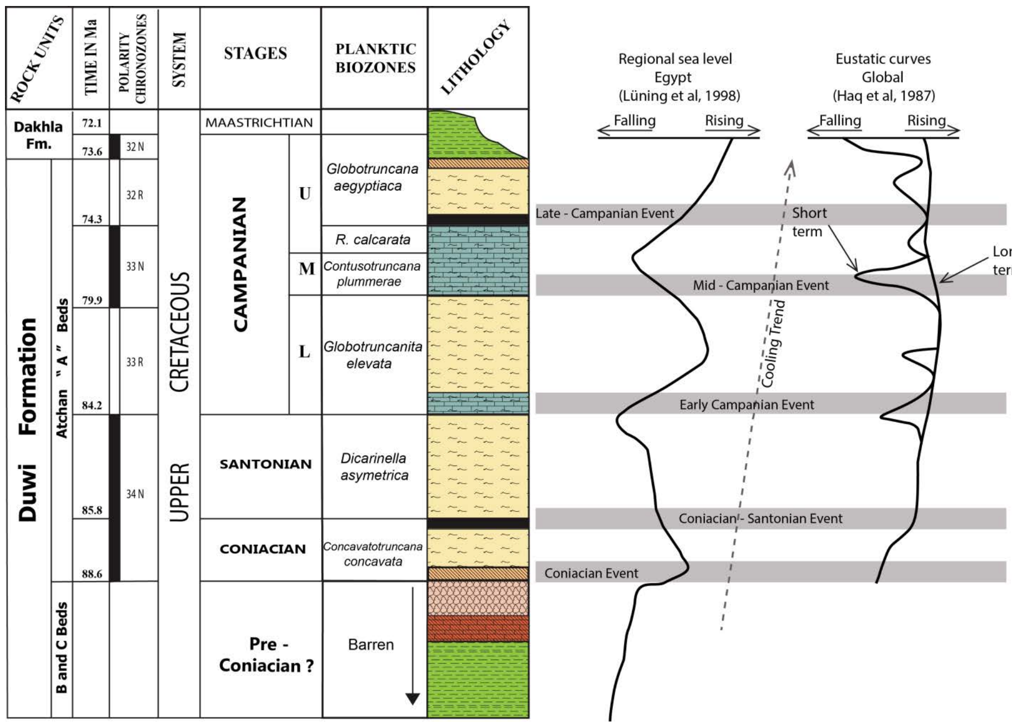

Figure 10: Sea–level changes *vs*. planktonic foraminiferal biozones, chronozones and biostratigraphic events, Coniacian-late Campanian succession of the Duwi Fm., Gabal Duwi, Egypt.
- Time scale adopted after.
- Assuming a constant sedimentation rate.
- Tethyan planktonic foraminiferal zones.
- Peaks and troughs on the regional sea level curves of Egypt.

Citation: Reda El Gammal MH, Orabi H (2019) Coniacian-late Campanian Planktonic Events in the Duwi Formation, Red Sea Region, Egypt. J Geol Geophys 7: 456. doi: 10.4172/2381-8719.1000456

Late Campanian Event (LCE) [1,3-5,27,] the Epsilon Event (EE) (also called C1-Event) [3,30] in many Tethyan cases, *Praeglobotrancanella (Globotruncanella) havanensis* as an index species is not present in Duwi section, it stratigraphically downs into the hardgrounds indicating to maximum sea level rise. *Radotruncana calcarata* as in Tunisia and SW France [92] and may correlate with the Duwi section where the *Globotruncana aegyptiaca* Zone rests over *R. calcarata* directly marking the LCE. During the early late Campanian, a great transgression event occurred indicated by very large double–keeled foraminifera as *Globotruncana aegyptiaca* indicative of oligotrophic conditions. Jarvis [27] concluded that a broad agreement between the NW Europe and Egypt curves of Lüning [9] with generally rising sea levels throughout the Campanian, and major earliest Campanian and mid-late Campanian transgressions.

Egypt (even Duwi section) and in Tunisia [27], the Late Campanian Event coincides with the final phase of marl sedimentation at the base of the Gansserina (Globotruncana) gansseri Subzone, which directly overlying the *Globotruncana aegyptiaca* Zone.

In conclusion, of all three Campanian events in Europe and North Africa (Tunisia and Egypt) are mainly sea level changes appear to be synchronous indicating a dominance of eustatic over regional tectonic forcing [27].

Discussion

Sedimentary basin environment

The Gabal Duwi Range is considered a sub-basin of Neo-Tethys started opening to the Tethys oceanic basin may since the Late Cretaceous starting from Coniacian Stage. The Coniacian deposition started with a vast shallowest platform receiving siliciclastic sediments of the Quseir Formation and lower parts of the Duwi Formation. With a tectonic normal subsidence of basin toward N-NW, a geodynamic evolution of the Duwi basinal area was related to the opening of gateways of Neo-Tethys depositing more basinal sediments from Coniacian up to Eocene Epochs, with significant sea level changes globally or periodic local regime.

Duwi is as asymmetric syncline system (Figure 12) of hanging a wall with extensional fault–related folding [31]. It is bounded by Precambrian basement rocks (Figure 1) where gentle, moderately east dipping Nubia Sandstones occur in the footwall (Figure 8d). These Pre-Coniacian (Hamadat C–Beds) accompanying eustatic sea level rise phosphorites are developed through intermittent winnowing of fine–grained phosphatic shales (OAE 3) (Oxygen Anoxic Environment) and biosiliceous sediments. Gradual sea level fall of the basin during Oyster banks of Duwi (B-Beds) (Figure 8c) with brackish back–reef sediments and deltaic sediments dominated the middle beds of the Duwi Formation all these sediments are barren of foraminifera.

The start of Coniacian Stage (Atchan A–Beds), the sea level rise cycle of inner to outer neritic marine conditions with strong waves and surface-bottom currents usually tend to distribute the local influx of terrigenous sediments over large areas [37], such terrigenous sediments rich in planktonic, benthonic foraminifera and ostracods. Such conditions often persist over long time intervals without being filled up to sea level [93]. This case is true for widely extended shallow marine basins as long as excess sediment volume in relation to space provided by gradual subsidence (averaged 0.96 cm/Kyr in case of upper beds of Duwi Formation). The margin of such basins is characterized by a kind of Ramp Morphology as Duwi basin case.

Sea level fluctuations

An erosional contact is present between the middle (B-beds) and upper (A–Beds) of the Duwi Formation. This contact is characterized by the long wall of bioclastic, silicified limestone of reefal facies (Oyster banks) or Lime–coquina and overlying silicified phosphatic bed of about 40 cm thick rich in glauconite (Figures 8a, 8b and 9c). The (A–Beds) sequence displays a transgressive nature with silty, marls

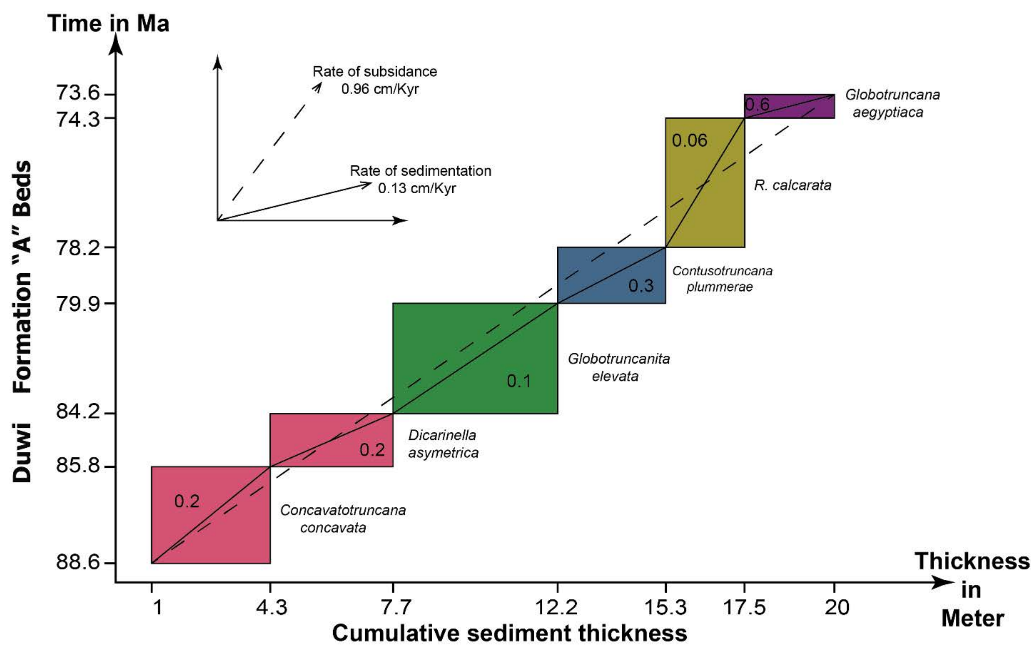

Figure 11: Age-thickness for the Coniacian–Late Campanian sedimentary sequence of the Duwi Fm., Gabal Duwi, Egypt; Sedimentation rates based on thickness of each biozone, its age (Ma) and corresponding chrones. Numbers of sedimentation rates correspond to each biozone.

Citation: Reda El Gammal MH, Orabi H (2019) Coniacian-late Campanian Planktonic Events in the Duwi Formation, Red Sea Region, Egypt. J Geol Geophys 7: 456. doi: 10.4172/2381-8719.1000456

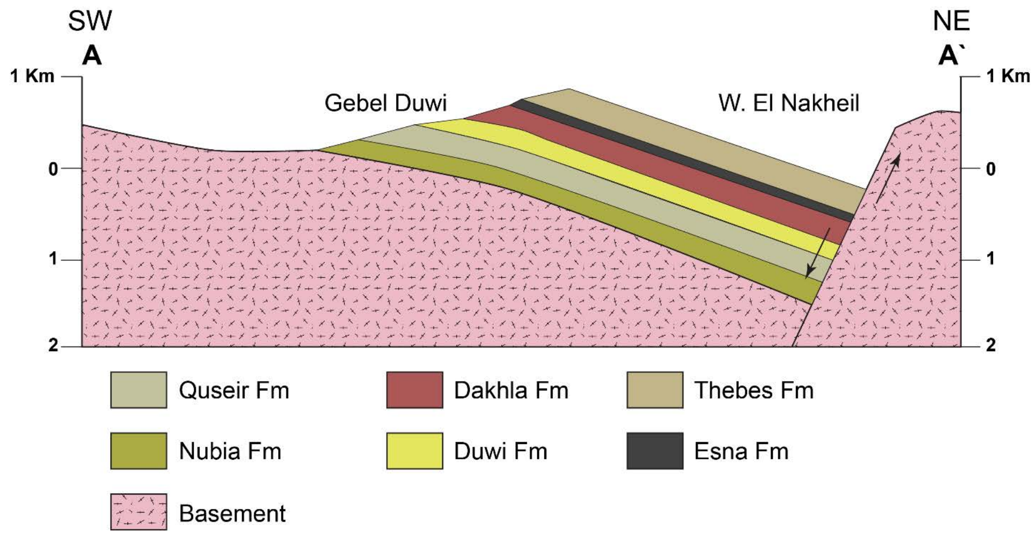

Figure 12: Structural cross section A-À (on Figure 1) across the Gabal Duwi area demonstrating the asymmetric syncline system of hanging wall with extensional fault related folding as Duwi Basin as continental or interior sag basins.

and argillaceous limestone containing the migrated foraminiferal fauna from Coniacian to Late Campanian of the Duwi Formation. Fluctuations in relative sea level occurred in form of two chert bands and topmost conglomeratic phosphatic bed which separating the Duwi Formation and the overlying the Upper Campanian–Maastrichtian Dakhla Formation (Figure 9b). Also, very short–lived regressive episodes and drops in sea level are observed before the accumulation of thin black shales through all interval of the Duwi Formation especially the Coniacian–Campanian interval. There is no unconformity surfaces existed through the Coniacian–Late Campanian. In general, the sea level changes were quite marked no turbidity through Coniacian–Late Campanian. The succession may be explained as tectonically quiet as well as eustatic changes of sea levels.

Rates of sedimentations

According to authors Jarvis, Hart and Thibault N [1,94,95] it can be calculate or estimate the rate of sedimentation for the studied foraminiferal 20 m thick of the (A-Beds) from Coniacian to Late Campanian about 17 Ma [95] according to the Age-Depth Models described by Thibault (Figure 12) [95] allows to estimating the variations of sedimentation rates and age assignments of calcareous microfossils (planktonic foraminifera). It is clear that the lowest rate of sedimentation is of the *Radotruncana calcarata* Zone (0.06 Cm/Kyr), the higher one is of the *Globotruncana aegyptiaca* (0.6 Cm/Kyr) and other biozones are from 0.2 to 0.3 Cm/Kyr (Figure 13). The overall average is about 0.13 Cm/Kyr, indicating very calm and quiet, no tectonics or abnormal cases except only eustatic sea level basinal conditions comparing with the Tethyan Bottaccione and Contessa High Way sections, 1.16 Cm/Kyr [96]. Also, the average of basin subsidence is 0.96 Cm/Kyr, it means very gentle subsidence with a long time. These data may explain the small thicknesses of biozones comparing with deeper oceanic basins. Comparing our data with most Tethyan Cretaceous sections, the differences do not exceed ± 0.01 Cm/Kyr. Through the Late Campanian, there is an increase in depth with an increase of flavoring and specification in planktonic components indicating a time of expansion of the Egyptian continental shelf and continued into the Maastrichtian.

Summary and Conclusions

1. The present study has focused on the Duwi Formation at Gabal Duwi Range. This succession is easily accessible with its contact with lower Quseir Formation and the upper Dakhla Formation. The entire interval includes the transition from the Coniacian stage to the Campanian stage.

2. The lithostratigraphic subdivisions for the Pre-Coniacian-Late Campanian in Gabal Duwi are proposed, from the base to the top as follow; Lower phosphate unit (Pre-Coniacian), which named Abu Shigeila or Hamadat “C” horizon; Middle phosphate unit (Pre-Coniacian), which named Main Duwi Beds or “ B ” beds and Upper phosphatic unit (Coniacian-Late Campanian), which named locally Atchan phosphate or “A” Beds.

3. About 6 standard Tethyan biozones covering the sedimentary sequence of the interval (A-beds) have been recognized and correlated to the world bioevent as follows; *Dicarinella concavata* Zone (CF 14) and *Dicarinella asymetrica* Zone (CF 13), where these two zones are correlated to the Coniacian–Santonian time interval of Oceanic Anoxic Event 3 (OAE 3). *Globotruncanita elevate* Zone (CF 12) is correlated with the Santonian–Campanian Boundary Event (SCBE). *Contusotruncana plummerae* Zone (CF 11) is correlated with the Mid Campanian Event (MCE). *Radotruncana calcarata* Zone (CF 9 and CF 10) and *Globotruncana aegyptiaca* Zone (CF 8a), these two zones are correlated to the Late Campanian Event (LCE). In conclusion, of all three Campanian events in Europe and North Africa (Tunisia and Egypt) are mainly sea level changes appear to be synchronous indicating a dominance of eustatic over regional tectonic forcing.

4. The Gabal Duwi Range is considered a sub-basin of Neo-Tethys started opening to the Tethys oceanic basin may since the Late Cretaceous starting from Coniacian Stage. Through the Late Campanian, there is an increase in depth with an increase of flavoring and specification in planktonic components indicating a time of expansion of the Egyptian continental shelf and continued into the Maastrichtian.

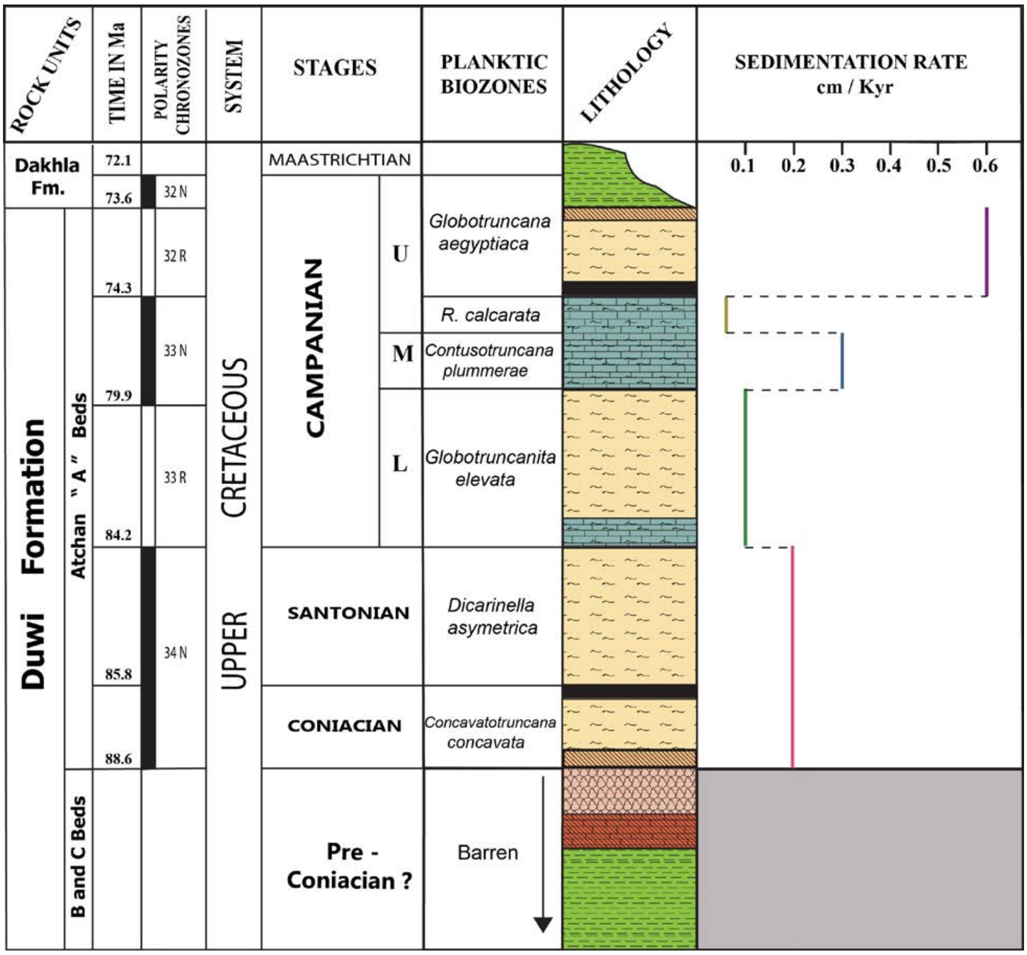

Figure 13: Changes in the sedimentation rates with the eustatic sea level changes of Coniacian- Late Campanian, Duwi Fm., Gabal Duwi section.

Acknowledgements

The authors are grateful to Dr. Friedrich van der Wart from the Netherlands for critically reading the manuscript and suggesting improvements. The authors thank the anonymous reviewers for their very helpful comments on the manuscript.

References

1. Jarvis I, Gale AS, Jenkyns HC, Pearce MA (2006) Secular variation in late Cretaceous carbon isotopes: a new δ13C carbonate reference curve for the Cenomanian–Campanian (99.6–70.6 Ma). Geol Mag 143: 561–608.
2. Gradstein FM, Ogg JG, Schmitz MD, Ogg GM (2012) The Geologic Time Scale 2012. 2. Elsevier BV, Amsterdam, The Netherlands 1176.
3. Thibault N, Harlou R, Schovsbo N, Schiøler P, Minoletti F, et al. (2012a) Upper Campanian–Maastrichtian nannofossil biostratigraphy and high-resolution carbon–isotope stratigraphy of the Danish Basin: Towards a standard δ13C curve for the Boreal Realm. Cretac Res 33: 72–90.
4. Thibault N, Husson D, Harlou R, Gardin S, Galbrun B, et al. (2012b) Astronomical calibration of upper Campanian–Maastrichtian carbon isotope events and calcareous plankton biostratigraphy in the Indian Ocean (ODP Hole 762C): Implication for the age of the Campanian–Maastrichtian boundary. Palaeogeogr Palaeoclimatol Palaeoecol 337-338: 52–71.
5. Voigt S, Gale AS, Jung C, Jenkyns HC (2012) Global correlation of upper Campanian–Maastrichtian successions using carbon–isotope stratigraphy: Development of a new Maastrichtian timescale Newsl Stratigr 45: 25–53.
6. Lamolda MA, Paul CRC, Peryt D, Pons JM (2014) The Global Boundary Stratotype and Section Point (GSSP) for the base of the Santonian stage, “Cantera de Margas“, Olazagutia, Northern Spain. Episodes 37: 2–13.
7. Premoli Silva I, Sliter WV (1995) Cretaceous planktonic foraminiferal biostratigraphy and evolutionary trends from the Bottacioni section, Gubbio, Italy. Paleontographica Italica 82: 1–89.
8. Gale AS, Hancock JM, Kennedy WJ, Petrizzo MR, Lees JA, et al. (2008) An integrated study (geochemistry, stable oxygen and carbon isotopes, nannofossils, planktonic foraminifera, inoceramid bivalves, ammonites and crinoids) of the Waxahachie Dam Spillway section, north Texas: A possible boundary stratotype for the base of the Campanian Stage. Cretaceous Res 29: 131–167.
9. Lüning S, Marzouk AM, Morsi AM, Kuss J (1998) Sequence stratigraphy of the Upper Cretaceous of south-east Sinai, Egypt: Cretaceous Res 19: 153-196.
10. Haq BU, Hardenbol J, Vail PR (1987) Chronology of fluctuating sea levels since the Triassic. Science 235: 1156–1167.
11. Tantawy AA, Keller G, Adatte T, Stinnesbeck W, Kassab A, et al. (2001) Maastrichtian to Paleocene (Dakhla Formation) depositional environment of the Western Desert in Egypt: sedimentology, mineralogy and integrated microand macrofossils biostratigraphies. Createaceous Res 22: 795-827.
12. Ismail AA (2012) Late Cretaceous-Early Eocene benthic foraminifera from Esh El-Mallaha area, Egypt. Rev. Paleobiol Gen 31: 15-50.
13. Hegab AA (1986) Contribution to the biostratigraphy of the Maestrichtian-Paleocene rocks, Gebel Duwi (Quseir area, Red Sea), Egypt. Bull Fac 15: 125-139.
14. Schrank E, Perch-Nielsen K (1985) Late Cretaceous Palynostratigraphy in Egypt with comments on Maastrichtian and early Tertiary calcareous nannofossils. Newsl Stratigraphy 15: 81-99.

15. Baioumy HM, Tada R (2005) Origin of Late Cretaceous phosphorites in Egypt. Cretaceous Res 26: 261-275.

16. Wagreich M (2009) Coniacian-Santonian oceanic red beds and their link to Oceanic Anoxic Event 3, in: Cretaceous Oceanic Red Beds: Stratigraphy, Composition, Origins, and Paleoceanographic and Paleoclimatic Significance. SEPM Spec Pub 91: 235–242.

17. Friedrich O, Norris RD, Erbacher J (2012) Evolution of middle to late cretaceous oceans a 55 m. y. record of Earth's temperature and carbon cycle. Geol 40: 107–110.

18. Huber BT, Norris RD, MacLoed KG (2002) Deep sea paleotemperature record of extreme warmth during the Cretaceous. Geol 30: 123-126.

19. Hay WW (2011) Can humans force a return to a Cretaceous climate? Sediment. Geol 235: 5-26.

20. Linnert C, Robinson SA, Lees JA, Bown PR, Pérez-Rodríguez I, et al. (2014) Evidence for global cooling in the Late Cretaceous. Nat Commun 5: 4194.

21. Petrizzo MR (2002) Paleoceanographic and paleoclimatic inferences from Late Cretaceous planktonic foraminiferal assemblages from the Exmouth Plateau (ODP Sites 762 and 763, Eastern Indian Ocean). Marine Micropaleon 45: 117-150.

22. Falzoni F, Petrizzo MR, Clarke LC, MacLeod KG, Jenkyns HJ, et al. (2016) Long-Term cretaceous carbon and oxygen isotope trends and planktonic foraminiferal turnover: A new record from the southern mid-latitudes. GSA Bull 128: 1725-1735.

23. Schlanger SO, Jenkyns HC (1976) Cretaceous anoxic events: Causes and consequences. Geol Mijnbow 55: 179–184.

24. Arthur MA, Schlanger SO (1979) Cretaceous "oceanic anoxic events" as causal factors in development of reef-reservoired giant oil fields. Am Ass Pet Geol Bull 63: 870-885.

25. Wagreich M (2012) "OAE 3" – regional Atlantic organic carbon burial during the Coniacian-Santonian. Clim Past 8: 1447–1455.

26. Barrera E, Savin SM, Thomas E, Jones CE (1997) Evidence for thermohaline-circulation reversals controlled by sea-level change in the latest Cretaceous. Geol 25: 715–718.

27. Jarvis I, Mabrouk A, Moody RT, de Cabrera S (2002) Late Cretaceous (Campanian) carbon isotope events, sea-level change and correlation of the Tethyan and Boreal realms. Palaeogeogr. Palaeoclimatol. Palaeoecol 188: 215–248.

28. Algeo TJ, Meyers PA, Robinson RS, Rowe H, Jiang GQ, et al. (2014) Icehouse-greenhouse variations in marine denitrification. Biogeosci 11: 1273–1295.

29. Perdiou A, Thibault N, Anderskouv K, van Buchem F, Buijs GJA, et al. (2015) Orbital calibration of the late Campanian carbon isotope event in the North Sea. J Geol Soc Lond 173: 504-517.

30. Thibault N, Jelby ME, Anderskouv K, Bjerager M, Surlyk F, et al. (2015) Upper Campanian–Maastrichtian chronostratigraphy of the Skælskør-1 core (eastern Sjælland, Danish Basin): Correlation at the basin and global scale and paleoclimatic changes. Lethaia 48: 549-560.

31. Khalil SM, McClay K (2002) Extentional fault related folding, northwestern Red Sea, Egypt. J Struc Geol 24: 743-762.

32. Akkad MK, Noweir AM (1980) Geology and lithostratigraphy of the Arabian desert orogenic belt between latitudes 25° 35ˮ and 26° 30ˮ. Permagon Press, New York 127-135pp.

33. Said R (1990) The geology of Egypt. Egyptian General Petro Corp 734 pp.

34. Youssef MI (1957) Upper cretaceous rocks in Kosseir area. Bulletin de l'Institute, 7: 35-53.

35. Isawi B, El-Hinnawi M, Francis M, Mehanna A (1969) Contribution to the structure and phosphate deposits of Qusseir area: Egypt. Organ Geol Res Min Geol Survey 50: 35.

36. Sheldon RP (1981) Ancient marine phosphorites. Annu Rev Earth Planet Sci 9: 251-284.

37. Glenn CR (2016) Depositional sequences of the Duwi, Sibaiya and phosphate formations, Egypt: phosphogenesis and glauconitization in a Late Cretaeous epeiric sea. Geol Soc Spec Pub 52: 205-222.

38. Soliman MF, Essa MA (2003) Upper Dakhla Formation (Beida Shale member) at G. Duwi, Red Sea, Egypt: Mineralogical and geochemical aspects . Third Intern Conf Geol Africa 2: 283-305.

39. Luger P (1985) Stratigraphy of the marine Cretaceous and Early Tertiary in the south-western Obernil Basin (SW-Aegypt), with particular reference to micropalaontology, palocology and palaeography. Berliner geowiss. Abh. 63: 11-151.

40. Farouk S, Ahmed F, Powell JH, Marzouk A (2016) Integrated microfossil biostratigraphy, facies distribution,and depositional sequences of the upper Turonian to Campanian succession in northeast Egypt and Jordan. Facies 62: 8.

41. Baioumy H, Lehmann B (2017) Anomalous enrichment of redox-sensitive trace elements in the marine black shales from the Duwi Formation, Egypt: Evidence for the late Cretaceous Tethys anoxia. J African Earth Sci 133: 7-14.

42. Dominik W, Schall S (1984) Notes on the stratigraphy of the Upper Cretaceous phosphates (Campanian) of the Western Desert, Egypt. A Geol & Paleontol 50: 153-1 75.

43. De Souza FL, Kral G, Fauth G (2018) Late Cretaceous (Cenomanian-Maastrichtian) planktonic forminifera from Goban Spur (DSDP Sites 549 and 550): Biostratigraphy inferences. Cretaceous Res 86: 238-250.

44. Caron M (1985) Cretaceous planktic foraminifera. Plankton Stratigraphy. Cambridge Univ. Press, 17-86.

45. Nederbragt AJ (1991) Late Cretaceous biostratigraphy and development of Heterohelicidea (planktic foraminifera). Micropaleontol 44: 385-412.

46. Robaszynski F, Caron M (1995) Cretaceous planktonic foraminifera: commentary on the Europe-Mediterranean zonation. Bull Soc Geol Fr 166: 681–692.

47. Li L, Keller G (1998a) Maastrichtian climate, productivity and faunal turnovers in planktic foraminifera in South Atlantic DSDP Sites 525A and 21. Mar Micropaleontol 33: 55-86.

48. Li L, Keller G (1998b) Abrupt deep-sea warming at the end of the Cretaceous. Geol 26: 995-998.

49. Sigal J (1955) Notes micropaleontologiques Nord-Africanines. 1Du Cenomanien au Santonien: zones et limites en facies pelagique. CR Somm Soc Geol.Fr.7/8, 157-160.

50. Kopaevich LF, Vishnevskaya VS (2015) Cenomanian-Campanian (Late Cretaceous) planktonic assemblages of the Crimea-Caucasus area: Paleoceanography, paleoclimate and sea level changes. Paleogeog paleoclim paleoecol 441: 493-515.

51. Kopaevich LF, Gorbachic TN (2017) Shell morphology of Cretaceous planktonic foraminifers as a means for paleoenvironment reconstructions. Paleontolo J 51: 1-12.

52. Petrizzo MR, Berrocoso AJ, Falzoni F, Huber BT, Macleod KG, et al. (2017) The Coniacian-Santonian Sedimentary recrd in southern Tanzania (Ruvuma Basin, East Africa): Planktonic foraminiferal evolutionary, geochemical and paleoceanographic patterns. Sedimentol 64: 252-285.

53. Petrizzo MR (2000) Upper Turonian-lower Campanian planktonic foraminifera from southern mid-high latitudes (Exmouth Plateau NW Australia): Biostratigraphy and taxonomic notes. Cretaceous Res 21: 479-505.

54. Robaszynski F, Gonzáles Donoso JM, Linares D, Amédro F, Caron M, et al. (2000) The Upper Cretaceous region of Kalaat Senan, central Tunisia. Integrated Litho-Biostratigraphy: Zone of ammonites, planktonic foraminifers and nannofossils from the Turonian Higher than the Maastrichtian. Bull Elf-Aquitaine Exploration-Prod Res Cen 22: 35–490.

55. Wendler I, Wendler J, Grafe KU, Lehmann J, Willems H, et al. (2009) Turonian to Santonian carbon isotope data from the Tethys Himalaya, southern Tibet, Cretaceous Res 30: 961–979.

56. Postuma J (1971) Manual of planktonic foraminifera. Elsevier Publishing Co, Amsterdam, 420pp.

57. El-Gammal RM (1985) Geological studies on the stratigraphic succession of Um-Bogma District, Gebel Musabaa Salama, west Central Sinai,Egypt. MSc Thesis, Cairo University, 158p.

58. Obaidalla NA, Kassab AS (2002) Integrated biostratigraphy of the Coniacian-Sntonian sequence, Southwestern Sinai, Egypt. Egypt. J Paleont 2: 85-104.

59. Farouk S, Faris M (2012) Late Cretaceous Calcareous nannofossil and

planktonic Foraminiferal bioevents of the shallow-marine Carbonate Platform in the Mitla Pass, west central Sinai, Egypt. Cretaceous Res 33: 50–65.

60. Aref M, Ramadan M (1990) New recorded planktonic foraminifera from the Upper Cretaceous rocks of Esh El-Mellaha range, Red Sea, Egypt, MERC. Ain Shams Sci Ser 4: 123-141.

61. Dalbiez F (1955) The Genus Globotruncana in Tunisia. Micropaleontol 1: 161-171.

62. Barr FT (1972) Cretaceous biostratigraphy and planktonic foraminifera of Libya. Micropaleontol 18: 1-46.

63. Wonders AAH (1980) Middle and late Cretaceous planktonic foraminifera of the western Mediterranean area. Utrecht Micropaleontol Bul 24: 1- 158.

64. Elamri Z, Farouk S, Zaghbib-Turki D (2014) Santonian planktonic foraminiferal biostratigraphy of the northern Tunisia. Geologia Croatica 67: 111–126.

65. Soycan H, Hakyemez A (2018) The first calibration of radiolarian biochronology with Late Cretaceous (latest Coniacian-Santonian to early Campanian) planktonic foraminifera in the volcano- sedimentary sequences of the Eastern Pontides, NE Turkey. Cretaceous Res 85: 319-348.

66. Wolfgring E, Wagreich M, Dinarès Turell J, Yilmaz I O, Böhm K, et al. (2018) Plankton biostratigraphy and magnetostratigraphy of the Santonian–Campanian boundary interval in the Mudurnu–Göynük Basin, northwestern Turkey. Cretaceous Res 87: 296-311.

67. Ayyad SN, Abed MM. Abu Zied RH (1997) Biostratigraphy of the Upper Cretaceous rocks in Gebel Arif El-Naga, northeastern Sinai, Egypt, based on benthonic foraminifera. Cretaceous Res 18: 141-159.

68. El-Gammal RMH, Sayah TA, El-Assy IE , Shahata RM (1996) The Campanian-Maastrichtian boundary at Southern Wadi Araba area, Southwestern Sinai, Egypt. Geol Soc Egypt Special Publ 2: 13-31.

69. Petrizzo MR, Falzoni F, Premoli Silva I (2011) Identification of the base of lower to middle Campanian Globotruncana ventricosa Zone: Comments on reliability and global correlations. Cretaceous Res 32: 387-405.

70. Mancini EA, Puckett MT (2005) Jurassic and Cretaceous Transgressive-Regressive (T-R) Cycles, Northern Gulf of Mexico, USA. Stratigraphy 2: 31-48.

71. Petrizzo MR (2003) Late Cretaceous plabktonic foraminiferal bioevents in the Tethys and in the southern ocean record: an overview. J Foraminiferal Res 33: 330-337.

72. Neuhuber S, Gier S, Hohenegger J, Wolfgring E, Spotl CH, et al. (2016) Paleoenvironmental changes in the northwestern Tethys during the Late Campanian *Radotruncana calcarata* Zone : Implications from stable isotopes and geochemistry. Chem Geol 420: 280-296.

73. Georgescu MD (2017) Upper Cretaceous planktic foraminiferal biostratigraphy. Studia UBB Geologia 61: 5-20.

74. Wolfgring E, Wagreich M (2016) A quantitative look on northwestern Tethyan foraminiferal assemplages. Campanian Nierental formation, Austria. Peer j 4: e1757.

75. Premoli Silva I, Verga D (2004) Practical manual of Cretaceous planktonic Foraminifera. Int School on Planktonic Foraminifera, 3rd Course: Cretaceous. University of Perugia and Milano, Tipografia Pontfelecino, Perugia, Italy, 283pp.

76. Jaff RBN, Wiliams M, Wilkinson IP, Lawa F, Zalasiewicz J (2014) A refind foraminiferal biostratigraphgy for the Late Campanian-Early Maastrichtian succession of northeast Iraq. GeoArabia 19: 161-180.

77. Premoli Silva I, Sliter WV (1999) Cretaceous paleoceanography: Evidence from planktonic foraminiferal evolution. Geol Soc Am Spec Pap 332: 301–328.

78. Smith CC, Pessagno EA JR (1973) Planktonic foraminifera and stratigraphy of the Corsicana formation (Maastrichtian) North Central Texas. Cushman Found. Foram Res Spec Publ No. 12.

79. Aref M, Philobbos ER, Ramadan M (1988) Upper Cretaceous-Lower Tertiary planktonic biostratigraphy along the Egyptian Red Sea region and its tectonic implication. Bull Fac Sci Assuit Univ 17: 171-201.

80. El-Deeb WZM., El-Gammal RMH (1994) Contributions to the biostratigraphy of Sudr and Esna formations in Southwestern Sinai, Egypt. Egypt J Petrol 3: 65- 84.

81. Lowery Ch M, Leckie RM, Sageman BB (2017) Micropaleontological evidence for redox changes in the OAE3 interval of the US Western Interior: Global vs. local processes. Cretaceous Res 69: 34-48.

82. Peryt D, Lamolda MA (2007) Neoflabellinids (benthic foraminifers) from the Upper Coniacian and Lower Santonian at Olazagutia, Navarra province, Spain; taxonomy and correlation potential. Cretaceous Res 28: 30-36.

83. Robaszynski F, Caron M, Gonzáles Donoso JM, Wonders AH and the European Working Group on Planktonic foraminifera, et al. (1984) Atlas of the Late Cretaceous globotruncanids. Rev Micropaleontol 26: 145-305.

84. Farouk S, Faris M, Elamri Z, Ahmed F, Wagreich M (2018) Tethyan plankton bioevents calibrated to stable isotopes across the upper Santonian-lower Campanian transition in north-western Tunisia. Cretaceous Res 85: 128-141.

85. Kopaevich LF (2009) Planktonic foraminiferal zonal schem for Late cretaceous of Crimea-Caucasus. Byull Mosk Obschh Inspyt prir Ser Geol 85: 40-52.

86. Thibault N, Jarvis I, Voigt S, Gale AS, Attree K, et al. (2016) Astronmical calibration and global correlation of the Santonian (Cretaceous) based on the marine carbon isotope record. Paleoceanography 31: 847-865.

87. Dubicka Z, Jurkowska A, Thibault N, Razmjooei MJ, Wojcik K, et al. (2017) An integrated stratigraphic study across the Santonian/Campanian boundary at Bocieniec, southern Poland: A new boundary stratotype candidate. Cretaceous Res 80: 61-85.

88. Razmjooei MR, Thibault N, Kani A, Dinares-Turell J, Puceat E, et al. (2018) Integrated bio- and carbon-isotope stratigraphy of the Upper Cretaceous Gurpi Formation (Iran): A new reference for the eastern Tethys and its implications for large-scale correlation of stage boundaries. Cretaceous Res 91: 312-340.

89. Sabatino N, Meyers SR, Voigt S, Coccioni R, Sprovieri M, et al. (2018) A new high-resolution carbon-isotope stratigraphy for the campanian (Bottaccioni section): Its implications for global correlation, ocean circulation, and astrochronology. Paleogeo Paleoclimatol Paleoecol 489: 29-39.

90. Abramovich S, Keller G, Stuben D, Berner Z (2003) Characterization of late Campanian and Maastrichtian planktonic foraminiferal depth habitats and vital activities A based on stable isotopes. Paleogeog Paleoclimatol Paleoecol 202: 1-29.

91. Falzoni F, Petrizzo MR, MacLeod KG, Huber BT (2013) Santonian-Campanian planktonic foraminifera from Tanzania, Shasky Rise and Exmouth Plateau: species depth ecology and paleoceanographic inferences. Marine Micropaleontol 103: 15-29.

92. Odin GS (2001) The Campanian-Maastrichtian stage boundary: Characterisation at Tercis les Bains (France): correlation with Europe and other continents. Elsevier Sci Pub 910p.

93. Einsele G (1992) Sedimentary basins: evolution, facies and sedimentary budget. Springer New York, printed in Germany, 628p.

94. Hart MB (1990) Cretaceous sea level changes and global eustatic curves; evidence from SW England. Proc Ussher Soc 7: 268-272.

95. Thibault N (2010) Calcareous nannofossils from the boreal upper Campanian-Maastrichtian chalk of Denmark. J Nannoplankton Res 31: 39-56.

96. Gardin S, Galbrun B, Thibault N, Coccioni R, Silva IP, et al. (2012) Bio-magnetochronology for the upper Campanian-Maastrichtian from the Gubbioarea, Italy: New results from the Contessa Highway and Bottaccione sections. Newalet Stratigr 45: 75-103.