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Calcareous nannoplankton proxies for palaeoenvironmental reconstruction of the Albian–Cenomanian succession in North-western Israel (Mount Carmel Region)

Maria N. Ovechkina^{a,*} mariaio@gsi.gov.il, Elisabetta Erba^b, Cinzia Bottini^b

^aGeological Survey of Israel, 32 Yesha'yahu Leibowitz St., Jerusalem 95501, Israel; School of Agricultural, Earth and Environmental Sciences, University of KwaZulu-Natal, Durban, RSA

^bDepartment of Earth Sciences “Ardito Desio”, Università di Milano, Via Mangiagalli 34, Milan 20133, Italy

*Corresponding author.

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ABSTRACT

We present data on the calcareous nannofossil distribution in the Albian–Cenomanian borehole CT8 obtained from the Carmel area, NW Israel. The nannofloral assemblage consists of 113 species, with *Watznaueria* spp. dominating the entire succession. The detailed calcareous nannofossil biostratigraphy of this section is established, and seven nannofossil (sub)zones are recognised. For the first time the age of the Tavasim Tuff (V₂) is biostratigraphically identified as Late Albian. The quantitative analysis of the calcareous nannofossils suggests warm, open marine or coastal, mainly oligotrophic conditions throughout the entire succession of borehole CT8, with evident fluctuations of temperatures and fertility and a general trend towards cooling in the uppermost part of the sequence (Middle–Late Cenomanian). Temperature and nutrient variations correspond to those in the western Tethys, thus suggesting at least a supra-regional extent of processes detected in the Carmel Region. The biostratigraphy shows no evidence of hiatuses in the studied section, but the absence of positive carbon isotope anomalies associated with the Oceanic Anoxic Event (OAE) 1d and Mid-Cenomanian Event elsewhere suggests that the OAE 1d is possibly missing in the studied section due to a hiatus around the Isfiye Fm. / Arqan Fm. boundary.

Keywords: Calcareous nannoplankton; planktic foraminifera; biostratigraphy; stable isotopes; palaeoenvironmental reconstructions; Albian; Cenomanian; Mount Carmel; Levant Basin.

1. Introduction

As one of the most abundant marine phytoplankton groups, calcareous nannoplankton is an important component of the marine ecosystem and it has been one of the primary oceanic producers since the Triassic. The distribution of calcareous nannoplankton is strongly influenced by climatic and oceanographic conditions (e.g., Mutterlose et al., 2005). Quantitative studies of Cretaceous calcareous nannofossil assemblages have resulted in the identification of the palaeoecological affinities of some taxa and the establishment of palaeotemperature and palaeofertility indices (Roth and Krumbach, 1986; Mutterlose and Wise, 1990; Mutterlose, 1991, 1996; Erba et al., 1992; Burnett et al., 2000; Mutterlose and Kessels, 2000; Street and Bown, 2000; Herrle, 2003; Herrle and Mutterlose, 2003; Lees et al., 2005; Tiraboschi et al., 2009; Bottini et al., 2015). Due to their sensitivity to temperature, nutrients, salinity and $p\text{CO}_2$ calcareous nannoplankton can be an excellent proxy of surface water conditions (Erba and Tremolada, 2004; Mutterlose et al., 2005; Tiraboschi et al., 2009; Erba et al., 2010).

The Early to Late Cretaceous transition was a time of global warming, when the Earth's climate was in an extreme greenhouse mode (e.g., Huber et al., 2002; Jenkyns, 2003, 2010; Hay, 2008; Ando et al., 2015; O'Brien et al., 2017). The northern and southern continents were separated by a wide eastern Tethys and narrow western Tethys and the Atlantic oceans. Starting from the late Albian, the Tethys and Atlantic oceans provided a circum-global connection at low latitudes probably with a stable, westward-flowing current throughout the Tethys (Roth, 1986; Barron, 1987; Hay et al., 1999; Giorgioni et al., 2015). Reconstruction of the surface-water palaeotemperature, fertility, and chemistry dynamics in mid-Cretaceous oceans attract broad attention (Watkins et al., 2005; Forster et al., 2007; Ando et al., 2010; Gambacorta et al., 2016).

Despite prolific literature on the sedimentology, micropalaeontology, tectonics, and geological mapping of NW Israel (Kafri, 1972; Rosenfeld and Raab, 1974; Bein and Weiler, 1976; Lewy and Raab, 1978; Sass and Bein, 1982; Lipson-Benitah et al., 1995, 1997; Segev, 2009; Karcz and Sneh, 2011), the palaeoecology of the region is still poorly investigated. Calcareous nannoplankton of Israel, both fossil and recent, has been studied for almost half a century (Moshkovitz, 1967, 1984, 1995; Reiss et al., 1971, 1985; Moshkovitz and Ehrlich, 1976; Ehrlich and Moshkovitz, 1978, 1982; Winter et al., 1979; Flexer and Honigstein, 1984; Gvirtzman et al., 1985, 1989, 2004; Almogi-Labin et al., 1991; Eshet et al., 1992; Eshet and Moshkovitz, 1995; Fiorentino, 1995; Eshet and Almogi-Labin, 1996; Weinbaum-Hefetz et al., 2000; Weinbaum-

Hefetz and Benjamini, 2011), although the coverage of stratigraphic intervals and the selection of research aims have been uneven. In particular, calcareous nannofossil assemblages of the mid-Cretaceous remain unstudied, despite broad development of mid-Cretaceous carbonate platforms in the country (Braun and Hirsch, 1994).

The southern Tethyan margin experienced a large transgression during the Cenomanian, initiating the deposition of calcareous sediments of up to 400 m thickness in Israel (Bein and Weiler, 1976; Sass and Bein, 1982). These rocks are widespread in Israel including the Carmel Region (Fig. 1) in the north-western part of the country (Bartov, 1990), as well as below the coastal plain and offshore (Lipson-Benitah et al., 1995). The Albian–Cenomanian succession in the Carmel Region was deposited during four sedimentary cycles separated by phases of maximal regression or transgression (Segev and Sass, 2014). The end of the second cycle is recorded by the Isfye Fm. recovered in borehole CT8 (Fig. 1). The miliolid facies documented by Lipson-Benitah et al. (1995, 1997) in the upper part of dolomites in the nearby borehole CT2 of the Yagur Fm. point to a shallow and warm-water environment on the inner platform (Sliter and Baker, 1972). The third cycle, which corresponds to the Arqan Fm., was deposited during a general deepening of the basin, as inferred from the faunistic and lithologic changes (Lipson-Benitah et al., 1995, 1997). The Lower–Middle Cenomanian chalks from the northwest Carmel Region were deposited in the outer shelf environment (Lipson-Benitah et al., 1995, 1997).

The Cenomanian and Turonian chalks and limestones of Carmel contain a rich macrofauna composed of cephalopods, gastropods, bivalves echinoids and ostracods (Rosenfeld and Raab, 1974; Lipson-Benitah et al., 1997) used for biostratigraphic applications (Avnimelech, 1965; Lewy and Raab, 1978). The local biostratigraphy of the Cenomanian outcrops has been based mainly on ammonites (Lewy and Raab, 1978). Planktic foraminifers have been found sporadically in these outcrops, and the derived planktic foraminiferal biostratigraphy of the Cenomanian has been established on the coastal plain and offshore, where foraminifera occur more frequently (Lipson-Benitah, 1980, 1994; Martinotti, 1993; Lipson-Benitah et al., 1995, 1997).

The first palaeoenvironmental reconstruction of the Carmel area during the Cenomanian was made by Lipson-Benitah et al. (1995, 1997), based on quantitative analyses of planktic foraminifera from boreholes CT2 and CT0 at 1 km distance from borehole CT8 (Fig. 1). According to Lipson-Benitah et al. (1995, 1997) the Cenomanian deposits were accumulated in

an outer shelf environment, and five ecostratigraphic intervals of alternating low and high relative abundances of planktic foraminifera were recognised, reflecting sea-level changes that correspond to the third order cycles 2.2 to 2.4 of the supercycle UZA-2 of Haq et al. (1988) and to sequences 2–5 of Robaszynski et al. (1993).

The present endeavour offers a new regional biostratigraphy based on calcareous nannofossils and supplemented by carbon and oxygen stable isotope data. The main aim of our research is to document the quantitative fluctuations of nannofossils during the Albian–Cenomanian to improve our understanding of the depositional environment in the Levant Basin in this interval.

2. Material

2.1. Geological setting

Borehole CT8 (32.79°N 34.98°E) (Fig. 1) drilled in the north-western part of the Carmel area was chosen for analysis due to the completeness of the obtained core that represents nearly 100% of the drilled interval (Lipson-Benitah et al., 1995, 1997). Samples for calcareous nannofossil investigation were taken at approximately 2 m intervals, at the same levels as for foraminiferal samples. A total of 108 samples were collected along the 225 m section of the borehole.

The Carmel area is an isolated hilly belt composed mainly of Cretaceous (mostly Albian–Turonian) sequences (Sass and Bein, 1978). Mount Carmel is known for its complex stratigraphy and structure, which have been extensively studied over half a century (Picard and Kashai, 1958; Arkin and Hamaoui, 1967; Bein, 1976; Sass and Bein, 1978, 1982; Sass, 1980; Segev and Sass, 2006; Segev, 2009). In the southern part of this area, an anticlinal structure trending north-east is clearly recognised, other parts of the Carmel area are described as block structures (Sass, 1980). Local occurrences of volcanic material with varying thickness and lithology add to this complexity (Sass, 1980; Segev, 2009).

The sequence forming Mount Carmel consists of dolomites, reefoidal limestones, chalks with some cherts and marlstones. Facies changes that mark many stratigraphic units were attributed to the position of the area close to the edge of a large carbonate platform during the Cretaceous (Sass and Bein, 1978). Sediment supply and distribution was influenced by a rudist reef along the outer margin of the platform (Bein, 1976). Volcanics are represented by 16 separate volcanic events, some being good time markers (Sass, 1980). Segev (2009) recognised five tectonomagmatic events in the Mount Carmel sequence and four Cenomanian volcanic phases.

In the Mount Carmel sequence four depositional cycles are recognised, i.e. Albian, Lower Cenomanian, Lower–Middle Cenomanian, and Middle Cenomanian–Turonian (Segev and Sass,

2014). Each cycle, composed of several lithofacies, accumulated under a specific regime of transgression and regression. The northern Carmel is characterised by only one volcanic horizon, but the cycles established in the southern areas can still be recognised, with the exception of the fourth cycle that is missing (Segev and Sass, 2014).

2.2. Lithology

Several geological maps have been produced for the Carmel area (Segev and Sass, 2006, 2014; Karcz and Sneh, 2011). The Yagur Formation, and the Isfiye and Arqan formations separated by the Tavasim Tuff (V_2) are recognised in the northwest Carmel borehole according to the newest map of Segev and Sass (2014). The Cenomanian has not been subdivided into substages in available maps.

A detailed lithological description of the CT8 core was given by Lipson-Benitah et al. (1995, 1997). Borehole CT8 penetrates the Yagur, Isfiye and Arqan formations. The Yagur Fm. (10 m) is composed of dolomite. The Isfiye Fm. (18 m) is represented by dolomitised chalk. The Arqan Fm. starts with the Tavasim Tuff (V_2) layer (16 m), higher the sequence is composed of dolomitised chalk and micritic carbonates (57 m). There is a 1-m layer (between 126.75–124.5 m) of indurated chalk inside the Arqan Fm. The upper part of the Arqan Fm. (127 m) is represented by chalk with an increasing amount of chert nodules.

2.3. Calcareous nannofossils

Calcareous nannofossils have been found in all samples except for the lowermost part of the section (Yagur Fm., samples 1–3) and sample 15 (Tavasim Tuff). However, nannofossils are quite rare at some levels. Their preservation ranges from poor to good (see Methods for definitions), but on average is moderate throughout the sequence: taxa have been identified under a polarising light microscope since etching and overgrowth are minimal. All identified species are listed with their authorships in Appendix A.

3. Methods

3.1 Planktonic foraminifera and calcareous nannoplankton

Following the suggestion of the Cenomanian Working Group of the Subcommittee on Cretaceous Stratigraphy (Kennedy et al., 2004) to define the base of the Cenomanian as the level of the first occurrence (FO) of *Thalmaninella globotruncanoides*, foraminifera from borehole CT8 were re-examined. All foraminiferal material is stored on microslides at the Geological

Survey of Israel. A total of 60 microslides were re-checked and levels of foraminiferal FOs were confirmed.

For updating the foraminifera biostratigraphy of borehole CT8, the most recent revised Upper Albian – Maastrichtian planktic foraminiferal biostratigraphical zonation of Coccioni and Premoli Silva (2015) for the Tethyan region was used.

The study of the calcareous nannofossils was performed on permanent smear slides prepared using the standard techniques of Bown and Young (1998) and examined under an Olympus BX53 polarising light microscope at 1250× magnification using phase contrast and cross-polarized light. Photographs were taken with an Olympus SC100 digital camera.

For stratigraphical subdivision of borehole CT8 by calcareous nannofossils, the integrated stratigraphical zonation of Bralower et al. (1995) for the Tethyan Region (NC Zones for the Albian) and the zonation by Burnett (1998) for the Tethyan–Intermediate Province (UC Zones for the Cenomanian) were applied.

Quantitative analysis of nannofloral assemblages was performed counting at least 300 nannofossil specimens in random traverses in each smear slide. In samples with scarce assemblages, counting was done in as many as 350 fields of view. Counts were applied to 97 samples for palaeoecological reconstructions. To assess the general abundance of calcareous nannoplankton, the number of coccoliths per field of view was calculated.

Nannofossil preservation was determined based on criteria described in Watkins (1992):

G – good, most specimens exhibit little or no secondary alteration;

M – moderate, all specimens are easily identifiable, but exhibit the effects of some secondary alteration from calcite dissolution and/or overgrowth;

P – poor, identification of species hampered but possible in some cases; specimens exhibit profound effects of secondary alteration from etching and/or overgrowth.

The relation of diversity and abundance of each species is demonstrated by the Shannon index, Evenness and Species Richness, which was calculated using the software PAST, ver. 3.18 (Hammer et al., 2001). The Species Richness is often used as a measure of a relative stability of ecological conditions (Sanders, 1968; Dodd and Stanton, 1981; Watkins, 1989).

A correlation matrix and a multivariate factor analysis (R-mode) varimax rotation with principal component extraction were run using Statistica 10 (<https://www.tibco.com>). The factor analysis was performed on 18 taxa/groups including species with mean abundances $\geq 1\%$ (11 taxa, Table

1) and taxa with mean abundances <1%. The latter (*Z. embergeri*, *P. columnata*, *E. floralis*, *P. cretacea*, *L. carniolensis*, *Lithraphidites* spp. (*L. acutus*, *L. alatus*, *L. eccentricus*, *L. houghtonii*) and *Rh. asper* + *Rh. splendens*) were selected because of their palaeoecological significance. Factor loadings >±0.5 were assigned to “dominant taxa”, factor loadings ±0.3–0.5 were assigned to “associated taxa” and loadings ±0.2–0.3 were assigned to “weakly associated taxa”. The nannofossil Nutrient Index (NI) was calculated from relative abundances of selected nannofossil taxa following Bottini et al. (2015), with a minor amendment by adding *Zeugrhabdotus* sp. (small) and introducing *Zeugrhabdotus moulladei* Bergen, 1998, which is here assumed to have the same palaeoecological affinity as *Z. erectus* (Watkins et al., 2005):

$$\text{NI} = [(Z. \text{moulladei} + \text{Zeugrhabdotus sp. (small)} + \text{Biscutum constans} + \text{Discorhabdus ignotus}) / (Z. \text{moulladei} + Z. \text{sp. (small)} + \text{Biscutum constans} + \text{Discorhabdus ignotus} + \text{Watznaueria barnesiae})] \times 100,$$

where, *Z. moulladei* + *Z. sp. small* (species with the major axis <3.5 µm), *B. constans* and *D. ignotus* are high productivity indicators, and *W. barnesiae* is a low productivity indicator. High values of the NI specify high productivity and vice versa.

The calculation of the nannofossil Temperature Index (TI) also follows Bottini et al. (2015):

$$\text{TI} = [(E. \text{floralis} + R. \text{parvidentatum} + \text{Staurolithites sp.}) / (E. \text{floralis} + R. \text{parvidentatum} + \text{Staurolithites sp.} + \text{Zeugrhabdotus diplogrammus} + \text{Rh. asper})] \times 100,$$

where *E. floralis* + *R. parvidentatum* + *Staurolithites* sp. are cold-water taxa, and *Zeugrhabdotus diplogrammus* + *Rh. asper* are warm water taxa. Low values of the TI specify high temperatures and vice versa.

3.2 Carbon and oxygen stable isotope analyses

For the oxygen and carbon stable isotopes, 100 bulk sediment samples from borehole CT8 were analysed. Measurements of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ were taken according to the standard technique for bulk sediment samples. Samples were taken from freshly cut rock fragments, crushed and homogenised in an agate mortar. The isotopes were analysed at the Geological Survey of Israel using a Delta Plus mass spectrometer equipped with the gas bench system. All $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values were calibrated against the international standard NBS-19, and are reported in ‰ relative to the VPDB standard. The system has an accuracy of ±0.03‰ for oxygen and ±0.02‰ for carbon isotopes.

4. Results

4.1. Biostratigraphy and biochronology

4.1.1. Planktic foraminifera biostratigraphy

The taxonomy and stratigraphical distribution of the planktic foraminifera from the Carmel area (CT8 and two nearby boreholes) were studied by Lipson-Benitah et al. (1995, 1997). The following planktic foraminiferal zones were established by Lipson-Benitah et al. (1995, 1997) for borehole CT8: the *Rotalipora brotzeni* (Lower Cenomanian), the *Rotalipora reicheli* (lower Middle Cenomanian), and the *Rotalipora cushmani* (Middle Cenomanian) with the *Rotalipora greenhornensis* and the *Pseudotruncana algeriana* subzones (Fig. 2).

According to the revised planktic foraminiferal biostratigraphical zonation for the Upper Albian – Maastrichtian (Coccioni and Premoli Silva, 2015), the four planktic foraminifera zones are recognised (Fig. 2).

The planktic foraminifera are first recorded in sample 20 (depth 183.35 m). It is impossible to trace the base of the *Parathalmanninella appenninica* Zone unambiguously. The FO of *P. appenninica* was observed in sample 26 (depth 169.90 m) by Lipson-Benitah et al. (1995, 1997, as *Rotalipora appenninica*), but the presence of this species at this level has not been verified during the re-examination of the microslides. In the lower part of the Arqan Fm. (samples 32–45) we recognised the Upper Albian *P. appenninica* Zone as the interval from the lowest occurrence of *P. appenninica* to the lowest rare occurrence of *Thalmanninella globotruncanoides*.

The confirmed occurrence of *P. appenninica* has been recorded together with the FO of *Thalmanninella brotzeni* in sample 32 (depth 158.70 m). Thus, the lower boundary of this zone can be drawn at the level of sample 32, as was established by Lipson-Benitah et al. (1995, 1997) for the *R. brotzeni* Zone. The upper boundary of the *P. appenninica* Zone is unambiguously fixed by the FO of *Th. globotruncanoides* in sample 45 (depth 132.50 m). The *P. appenninica* Zone is correlated with the lower part of the *R. brotzeni* Zone of Lipson-Benitah et al. (1997), which is considered to be Lower Cenomanian. However, Coccioni and Premoli Silva (2015) accept the Late Albian age of the *P. appenninica* Zone. Higher, in the interval of samples 45–62, the *Th. globotruncanoides* Zone of the Lower Cenomanian fixed confidently as the interval from the lowest rare occurrence of *Th. globotruncanoides* to the lowest occurrence of *Thalmanninella reicheli*. The lower boundary of the *Th. globotruncanoides* Zone marks the Albian/Cenomanian boundary (sample 45, depth 132.50 m). The upper boundary of the zone is defined by the last occurrence (LO) of *Th. reicheli* in sample 63 (depth 101.60 m). The interval of this zone is correlated with the upper part of the *R. brotzeni* Zone of Lipson-Benitah et al. (1997).

In the middle part of the Arqan Fm. (samples 63–78) the Middle Cenomanian *Thalmaninella reicheli* Zone was established as the interval between the first appearance (FA) of *Th. reicheli* to the FA of *Thalmaninella deeckeri*. The zone is equivalent to the *Th. reicheli* Zone of Coccioni and Premoli Silva (2015), who used the total range of *Th. reicheli*. Because of extremely scarce *Th. reicheli* in our material (the species has been detected in sample 63 only), the definition of Lipson-Benitah et al. (1997) is retained. The lower boundary of the zone is established at the depth of 101.6 m (sample 63) and the upper boundary is fixed at the depth of 74 m (sample 78), with the FA of *Th. deeckeri* in sample 79 (depth 72.00 m).

The upper part of the Arqan Fm. can be assigned to the *Rotalipora cushmani* Zone of the upper Middle – uppermost Cenomanian as the interval from the FA of *Th. deeckeri* to the LO of *R. cushmani*. The zone is equivalent to the *R. cushmani* Zone of Coccioni and Premoli Silva (2015), who defined this zone as the interval from the highest occurrence of *Th. reicheli* to the highest occurrence of *R. cushmani*. Because *Th. reicheli* is extremely scarce in the studied section, the definition of the *R. cushmani* Zone by Lipson-Benitah et al. (1997) is accepted. The zone is confidently established for the interval of samples 79–108 (depths 72.00–5.00 m).

The *R. cushmani* Zone is subdivided into two subzones: the *Thalmaninella greenhornensis* Subzone (upper Middle – lower Upper Cenomanian) and the *Dicarinella algeriana* Subzone (Upper Cenomanian). The *Th. greenhornensis* Subzone is used as the interval between the FA of *Th. deeckeri* and the FA of *D. algeriana*. The subzone is equivalent to the *Th. greenhornensis* Subzone of Coccioni and Premoli Silva (2015), who defined this subzone as the interval from the highest occurrence of *Th. reicheli* to the lowest occurrence of *D. algeriana*. This subzone is confidently established for the interval of samples 79–106 (depths 72.00–9.00 m). Because of extremely scarce *Th. reicheli* in our material, the definition of this subzone by Lipson-Benitah et al. (1997) is followed here. Lipson-Benitah et al. (1997) accepted the Middle Cenomanian age of this subzone, whereas Coccioni and Premoli Silva (2015) suggested a late Middle to early Late Cenomanian age.

The *D. algeriana* Subzone is used as the interval from the lowest occurrence of *D. algeriana* to the highest occurrence of *R. cushmani*. The subzone is equivalent to the *Praeglobotruncana algeriana* Subzone of Lipson-Benitah et al. (1997) and is detected in the uppermost part of the borehole (samples 107–108, depths 7.00–5.00 m). The age of this subzone is accepted Middle – Late Cenomanian (Robaszynsky et al., 1993), but Lipson-Benitah et al. (1995, 1997) suggested a

Middle Cenomanian age for this interval, due to the continuous presence of *R. brotzeni* (cf. Grosheny et al., 1992) and *Orbitolina* spp. (Schroeder and Neumann, 1985) in the Junediya Fm. (Lipson-Benitah et al., 1995, 1997).

4.1.2. The Global Boundary of the Cenomanian Stage, the Stratotype Section

Following recommendations of the Cenomanian Working Group of the Subcommittee on Cretaceous Stratigraphy, a Global boundary Stratotype Section and Point was selected on the western flanks of Mont Risou, Alpes, southeastern France (Kennedy et al., 2004). The base of the Cenomanian stage is defined at the level of 36 m below the top of Marnes Bleues, which corresponds to the FO of the planktic foraminifera *Rotalipora globotruncanoides* (Kennedy et al., 2004), now classified as *Thalmanninella globotruncanoides* (Caron and Premoli Silva, 2007; Coccioni and Premoli Silva, 2015). Accordingly, in this study the Albian/Cenomanian boundary is placed at the FO of *Th. globotruncanoides* (depth 132.50 m).

4.1.3. Calcareous nannofossil biostratigraphy

Calcareous nannofossil assemblages in the borehole are fairly rich, consist of 113 taxa (Appendix A; Plates 1–3), and allow for recognition of zones and subzones of the NC (Bralower et al., 1995) and UC (Burnett, 1998) schemes, as follows (Fig. 2).

The lowest part of the Isfiye Fm. (samples 4–5) is referred to the Upper Albian Subzone NC9b, the lower boundary of which is defined by the FO of *Eiffellithus monechiae* (sample 5, depth 218.70 m) and upper boundary by the FO of *Eiffellithus turriseiffelii* s.l. (sample 6, depth 216.90 m). *Eiffellithus monechiae* is absent from the lowest sample 4 of the Isfiye Fm., which, however, may be explained by the extreme rarity of the species in the lower part of the section in general. This subzone is characterised by relatively scarce nannofossils that include rare *Axopodorhabdus biramiculatus* (= *A. albianus*), *Cyclagelosphaera margerelii*, *Eprolithus floralis*, *E. monechiae* and more abundant *Prediscosphaera columnata*, *Manivitella pemmatoidea*, *Retecapsa angustiforata*, *Tranolithus orionatus*, *Watznaueria barnesiae* and *W. fossacincta*.

The main part of the Isfiye Fm., the Tavasim Tuff and the succession of the lower part of the Arqan Fm. (samples 6–47, depths 216.90–128.50 m) belong to UC0 Zone of the Upper Albian – lowermost Cenomanian, a special transitional zone in Burnett's (1998) Upper Cretaceous scheme. The lower boundary of UC0 Zone is defined by the FO of *Eiffellithus turriseiffelii* s.l. (sample 6, depth 216.90 m) *sensu* Burnett (1998) [see Discussion for distinguishing between *E. turriseiffelii* s.str. and *Eiffellithus casulus* Shamrock in Shamrock & Watkins 2009]. The upper boundary is marked by the FO of *Corollithion kennedyi* (sample 48, depth 127.15 m). Zone UC0

is equivalent to subzone NC10a of Bralower et al. (1993, 1995). This zone straddles the Lower/Upper Cretaceous boundary and is shared by two discrete biozonations.

Subzones UC0a and UC0b cannot be recognised in the present section because of the absence of the stratigraphically important marker *Hayesites albiensis*, the LO of which defines the top of the UC0a subzone.

Subzone UC0c is defined in the interval of samples 39–47 (depths 144.45–128.50 m) of the studied section. The lower boundary of this subzone is fixed by the FO of *Calculites anfractus* (sample 39, depth 144.45 m) and the upper boundary is marked by the FO of *C. kennedyi*.

The interval of samples 48–54 (depths 127.15–118.55 m) is assigned to UC1 Zone of the Lower Cenomanian due to the presence of *C. kennedyi*, the FO of which was observed in sample 48. The upper boundary of this Zone is fixed by the FO of *Gartnerago segmentatum* in sample 55 (depth 116.50 m). Subzone UC1a and undivided subzones UC1b–d are recognised. The FO of *C. kennedyi* is generally used for demarcating the basal Cenomanian (Ando et al., 2015). This level is slightly above the FA of *Th. globotruncanoides* (depth 132.50 m); this pattern of successive FOs of *Th. globotruncanoides* and *C. kennedyi* has been documented elsewhere (e.g., Robaszynski et al., 1993; Watkins et al., 2005; Ando et al., 2015).

The interval of samples 48–52 (depths 127.15–122.50 m) belongs to Subzone UC1a due to the presence of *Watznaueria britannica*, which LO marks the upper boundary of UC1a and is recorded in sample 52 (depth 122.50 m).

Subzones UC1b–d cannot be established because of the absence of *Gartnerago chiasta*, which LO defines the top of UC1b. Subzone UC1c is described as the interval from the LO of *G. chiasta* to the FO of rare and sporadic *Kamptnerius magnificus* or the FO of *Helicolithus anceps*. In our material *H. anceps* is absent, whereas rare and sporadic *K. magnificus* has been recorded higher from sample 66 (depth 97.50 m).

The interval of samples 55–60 (depths 116.50–106.45 m) belongs to Zone UC2 of the Lower – basal Middle Cenomanian due to the presence of *Gartnerago segmentatum*, which FO is recorded in sample 55 (depth 116.50 m). Subzone UC2a and undivided subzones UC2b–c are recognised. The interval of samples 55–59 (depths 116.50–108.35 m) belongs to UC2a subzone due to the presence of *Zeugrhabdotus xenotus*, which LO is recorded in sample 59 (depth 108.35 m). The interval between samples 59–61 (depths 108.35–105.15 m, sample 60 at 106.45 m)

corresponds to undivided UC2b–c subzones; the marker species *Cylindralithus sculptus* has not been found.

The interval of UC1 and UC2 Zones correlates to the Lower Cenomanian NC10b Subzone of the Bralower et al. (1995) scheme.

The upper part of the borehole (samples 61–108, depths 105.15–5.00 m) is referred to the transitional Zone UC3 of the basal Middle–Upper Cenomanian due to the presence of *Lithraphidites acutus*, which FO is recorded in sample 61 (depth 105.15 m). *Kamptnerius magnificus* (sample 66, depth 97.50 m) and *Lithraphidites eccentricus* (sample 72, depth 86.05 m) are found in this interval. Another important species is *Microrhabdulus decoratus*, which according to Perch-Nielsen (1985) first appears together with *L. acutus* and has been reported as the index-species for the Upper Cenomanian Zone CC10. In our material the FO of *M. decoratus* was recorded slightly below the FO of *L. acutus* in sample 58 (depth 110.55 m).

Subzones UC3a–d cannot be subdivided because of the absence of *Gartnerago theta*, *Staurolithites gausorhethium* and *Gartnerago nanum*. The top of subzone UC3d is defined by the LO of *C. kennedyi*. This datum is not reached in the studied section since *C. kennedyi* is still present in the uppermost sample 108 (depth 5.00 m). Burnett (1998) recorded the LO of *C. anfractus* in the upper part of subzone UC3b in the Boreal Realm. In our material, the presence of very rare *C. anfractus* has been fixed in the penultimate sample 107 (depth 7.00 m).

The interval of samples 61–108 (depths 105.15–5.00 m) correlates to NC11* Zone of the Middle to lower Upper Cenomanian in the Bralower et al. (1995) scheme as the interval from the FO of *Lithraphidites acutus* to the LO of *Axopodorhabdus albianus* (*A. biramiculatus*), which presents in the uppermost sample 108 (depth 5.00 m).

4.2. Calcareous nannoplankton preservation and abundance

In our material, calcareous nannofossils are moderately preserved and quite abundant in general, with only a few samples containing poor assemblages or being barren.

Watznaueria is a dominant genus and one of the most resistant to dissolution, among Mesozoic calcareous nannofossils (e.g., Thierstein, 1980; Lees et al., 2005). Cretaceous nannofloras with a low species diversity and abundant *W. barnesiae* are considered as poorly-preserved (Lees et al., 2005), and assemblages with *W. barnesiae* exceeding 40% are thought to be diagenetically affected to such an extent that they no longer bear an original palaeoecological signal (Thierstein, 1980; Roth and Bowdler, 1981; Roth, 1984; Roth and Krumbach, 1986; Herrle et al.,

2003). However, Williams and Bralower (1995) proposed that a high ratio of *W. barnesiae* is indicative of the assemblage alteration only when it reaches greater values (70%). Erba (1992) documented original features preserved in nannofossil assemblages with relative abundance of *W. barnesiae* >40% and suggested that this taxon is typically oceanic and dominant under oligotrophic conditions.

In the studied samples, despite the predominance of *W. barnesiae*, dissolution sensitive taxa (e.g., *B. constans*, *D. ignotus*, *Z. moulladei* and small *Zeugrhabdotus*) are present and generally common (Figs. 3 and 6). In addition, the abundance or the dissolution-prone taxa is not controlled by the preservation index (Fig. 5). Also, the stable Shannon index (H) (1.8–2.85) indicates a low assemblage alteration. Therefore, we conclude that the abundance of *W. barnesiae* preserves the original palaeoenvironmental signal related to oligotrophy (Mutterlose et al., 2005).

Nannofossil abundance shows relatively high fluctuations, with an average of six specimens per one field of view (Fig. 5). The highest values (up to 27 nannofossils/field of view) are recorded in the middle part of the Arqan Fm., which corresponds to the Lower – lower Middle Cenomanian.

In all analysed samples (Figs. 3 and 4; Appendixes B, C), the dominant genus is *Watznaueria* (average 64.2%) while the genera *Zeugrhabdotus* (avg. 8.5%) and *Biscutum* (avg. 8.0%) exceed 5% of the assemblage (Fig. 3), followed by *Retecapsa/Cretarhabdus* (avg. 2.8%), *Eiffelithus* (avg. 2.5%) and *Tranolithus* (avg. 2.0%), *Prediscosphaera* spp. (1.8%), *Discorhabdus ignotus* (1.4%), *Lithraphidites* spp. (1.3%), *Rhagodiscus* spp. (1.0%), *Manivitella pemmatoidea* (0.9%), *Axopodorhabdus biramiculatus* (0.8%) and *Eprolithus floralis* (0.8%). Taxa with an average relative abundance <0.5% are grouped as “Rare taxa” (Fig. 4, Appendix D); these include *Cribrosphaerella ehrenbergii* (0.4%), *Radiolithus* spp. (0.3%), *Nannoconus* spp. (0.2%), *Broinsonia/Gartnerago* (0.2%), *Cyclagelosphaera* spp. (0.2%) and *Crucibiscutum* sp./*Seribiscutum* sp./*Repagulum parvidentatum* (0.2%). Very rare species (average relative abundance 0.1–0.5%) are treated as “Other taxa” (1.6%) (Appendix D).

Watznaueria is by far the commonest genus (min. 27.1%, max. 87.0%) throughout the CT8 core. The genus is dominated by *W. barnesiae* (avg. 60.6%) (Fig. 3); other six species (*W. britannica*, *W. biporta*, *W. ovata*, *W. fossacineta*, *W. cf. W. bayackii*, and *W. manivittiae*) occur rarely throughout the entire succession.

The genus *Zeugrhabdotus* (min. 1.7%, max. 21.7%) (Fig. 3) is represented by *Z. diplogrammus*, *Z. bicrescenticus* (common), *Z. embergeri*, *Z. scutula*, *Z. moulladei* (frequent), *Z. howei*, *Z. noeliae*, *Z. clarus*, *Z. xenotus*, and small *Zeugrhabdotus* sp. (rare).

The genus *Biscutum* (avg. 8.0%) (Fig. 3) shows sharp fluctuations from 1.3–35.4% and is represented by abundant *B. constans* (avg. 7.7%) (Fig. 3) and rare *B. thurwii* and *Biscutum* sp.

The *Retecapsa/Cretarhabdus* group (avg. 2.8%) (Fig. 3) includes *C. conicus*, *C. striatus*, *Cretarhabdus* sp., *R. angustiforata*, *R. sp.*, *R. ficula*, and *R. crenulata*.

The genus *Eiffellithus* (avg. 2.5%) (Fig. 3) is represented by *E. casulus*, *E. gorkae*, *E. hancockii*, *E. monechiae*, and *E. turriseiffelii*.

The genus *Tranolithus* was observed throughout the entire succession (avg. 2.0%) (Fig. 3) and is represented by *T. orionatus* (most abundant, avg. 1.7%) and *T. gabalus*.

The genus *Prediscosphaera* (avg. 1.8%) (Fig. 4) is represented by *P. columnata*, *P. cretacea* (common), *P. grandis* (rare), and *P. spinosa* (very rare).

The genus *Discorhabdus* (Fig. 6) is represented by a single species, *D. ignotus* (0.3–4.2%, avg. 1.4%), which occurs throughout the section.

The genus *Lithraphidites* (0–6.5%, avg. 1.3%) (Fig. 4) is represented by *L. carniolensis*, *L. alatus*, *L. acutus*, *L. eccentricus*, *L. houghtonii*, and *Lithraphidites* sp.

The genus *Rhagodiscus* (0–3%, avg. 1.0%) (Fig. 4) is represented by *Rh. asper*, *Rh. splendens*, *Rh. achlyostaurion*, *Rh. amphus*, *Rh. angustus*, *Rh. sageri*, and *Rhagodiscus* sp.

The fluctuations of the Species Richness, Shannon Index, and Evenness throughout the studied interval are summarised in Fig. 5. The Species Richness (S) is low/moderate (mean 37) and varies from 16 to 53 species per sample. The Shannon diversity Index (H) is moderate (mean 1.8) and varies from 1.8–2.85. Evenness (E) varies from 0.09–0.39 with an average of 0.19.

Pearson's correlation coefficients calculated for the selected nannofossil taxa are summarized in Table 2. *Watznaueria barnesiae* shows the highest negative correlation with *B. constans*, *Z. bicrescenticus*, *T. orionatus* and *E. turriseiffelii*. There is a strong negative correlation between *W. barnesiae* and the group of *Cretarhabdus* spp./*Retecapsa* spp., *Z. diplogrammus*, *Z. embergeri*, *L. carniolensis* and *Rh. asper*+*Rh. splendens*. *Biscutum constans* has a positive correlation with *T. orionatus* and *Rh. asper*+*Rh. splendens*. There are positive correlations between *Z. bicrescenticus* and *Z. diplogrammus*, *Z. embergeri*, *T. orionatus* and *E. turriseiffelii*. Positively correlated are the following pairs: *Z. moulladei*+*Zeugrhabdotus* sp. (small) and *E.*

turriseiffelii; *D. ignotus* and *Lithraphidites* spp.; *P. columnata* and *E. turriseiffelii*; and *Lithraphidites* spp.+*P. cretacea* and *L. carniolensis*.

Three significant factors were extracted from the varimax rotation factor analysis, representing 47.24% of the total variance (Table 3).

The PC1 (21.22% of total variance) shows the highest positive loadings for *Z. bicuspidatus*, *T. orionatus*, *E. turriseiffelii*, *Z. embergeri*, *Z. diplogrammus* and *Cretarhabdus* spp./*Retecapsa* spp., and a negative loading for *W. barnesiae*. The associated species *P. cretacea* shows a positive loading, while *Rh. asper*+*Rh. splendens* and *Lithraphidites* spp. (weakly associated species) present negative loadings. *Zeugrhabdotus moulladei*+*Zeugrhabdotus* sp. (small) show low positive loadings (associated species).

The PC2 (15.16% of total variance) shows high positive loadings for *D. ignotus*, *E. floralis* and *P. cretacea*, *L. carniolensis*, *Lithraphidites* spp. and, as associated taxa, *Z. moulladei*+*Zeugrhabdotus* sp. (small). Weakly associated taxa *Zeugrhabdotus* spp. and *Watznaueria* spp. demonstrate low negative loadings.

The PC3 (10.86% of total variance) shows high positive loadings for *B. constans*, *D. ignotus* (associated taxa), and a negative loading for *W. barnesiae*. Associated taxa *T. orionatus*, *Rh. asper*+*Rh. splendens* and *E. turriseiffelii* demonstrate high positive loadings.

4.3. Calcareous nannoplankton Temperature (TI) and Nutrient (NI) indices

Of the three factors extracted from the factor analysis, PC3 represents surface water fertility, since oligotrophic species *W. barnesiae* shows opposite loadings with respect to mesotrophic taxa *B. constans* and *D. ignotus* (Table 3, Fig. 6). On the contrary, the positive and negative loadings of the assemblages, given by PC1 and PC2, cannot be related to temperature. This probably depends on the relatively low abundances of species indicative of temperatures (*Z. diplogrammus*, *Rh. asper*, *E. floralis*, *R. parvidentatum*, and *Staurolithites* spp.), which together constitute ca. 3.6% of the total assemblage being a negligible part of the variability in the whole dataset (Fig. 7). However, although the 'temperature' species are scarce in the section, they show minor fluctuations that are independent from the total abundance variations, and may reflect a primary signal (see Discussion).

The NI is generally low (mean 15.52) and varies from 3.8–57.57 with rapid fluctuations throughout the succession (Figs. 5–7), therefore a relatively suppressed fertility is reconstructed for the studied interval with minor occasional increases as evidenced by NI values above 20

(detected around 180–170, 140–130, 110–90 and 70–60 m). The TI (mean 29.75) shows rapid fluctuations in the interval of 0–83.33 m (Figs. 5–7) pointing to generally warm temperatures, with lowest values (warmest conditions) displayed between 120–110 m (Lower Cenomanian). Some decrease in palaeotemperatures is estimated from ca. 100 m upwards, where the TI is shifted towards higher values. This phase corresponds to a minor decrease in abundance of *Z. diplogrammus* and a relative increase in *E. floralis*. Cooler interludes (highest TI values) are detected at ca. 215 m and 170 m (Upper Albian).

4.4 Stable isotope analyses

The oxygen isotope profile shows remarkable positive and negative fluctuations throughout the succession (Fig. 6), with two trends: one (below the tuff deposits) reflecting a gradual decrease of the oxygen values, and the other (above the tuff deposits) showing increasing values.

In the Isfiye Fm., a positive shift of ca. +1.06‰ [-2.97 to -1.91‰] is observed below the tuffaceous deposits. Directly above the tuffaceous deposits (samples 20 to 30), a negative shift of ca. -0.68‰ [-1.58 to -2.26‰] is detected. In the interval of samples 31–44, the values of oxygen are practically constant [-1.93 to -1.46‰]. A negative $\delta^{18}\text{O}$ shift of ca. -1.4‰ [-1.85 to -3.25‰] is identified in the interval of samples 45–75. The interval corresponding to samples 76–96 is characterised by constant values of $\delta^{18}\text{O}$. A positive $\delta^{18}\text{O}$ anomaly of ca. +0.81‰ [-3.11 to -2.30‰] marks depths 62–54.7 m. In the upper part of the section (samples 98–108), a positive shift of ca. +0.94‰ [-3.70 to -2.76‰] is identified.

The $\delta^{13}\text{C}$ value changes from 0.47–2.00‰ in the Isfiye Fm. below the tuff (Fig. 5). Above the tuffaceous layer, the lower and middle parts of the Arqan Fm. (samples 20–68) are characterised by almost constant $\delta^{13}\text{C}$ values (1.88–1.64‰) with minor fluctuations that include a relative decrease from ca. 2‰ at 180–170 m to ca. 1.5‰ at 170–160 m. The interval of samples 69–87 is characterised by very rapid fluctuations of $\delta^{13}\text{C}$ values from 0.88–2.07‰. Above this interval, a rapid two-step decrease of the $\delta^{13}\text{C}$ values is detected. The first step (samples 89–98) shows an abrupt decrease of the $\delta^{13}\text{C}$ value from 0.43 to -2.41‰. The second one (samples 99–108) displays a gentler decline of the $\delta^{13}\text{C}$ value from -1.39 to -1.66‰.

The $\delta^{13}\text{C}$ curve of the studied section differs significantly from $\delta^{13}\text{C}$ records in other localities (e.g., Jarvis et al., 2006; Gambacorta et al., 2015; Giorgioni et al., 2015; Bornemann et al., 2017). The $\delta^{13}\text{C}$ values detected throughout the Upper Albian to Lower Cenomanian of the CT8 section are relatively uniform, contained within ca. 1–2‰. The Middle Cenomanian is characterised by

transitional values of $\delta^{13}\text{C}$ within 1–2 ‰ in the lower part and sharp changes in the upper part. The Upper Cenomanian is marked instead by rather negative carbon isotopic values that do not match any record worldwide, except, possibly, the Shilaif and Natih basins (United Arab Emirates) where a prominent decrease in $\delta^{13}\text{C}$ values is detected in the Late Cenomanian prior to OAE 2 (Vahrenkamp, 2013; Wohlfend et al., 2016; Hennhofer et al., 2019). Such differences may reflect either diagenetic alteration or/and the presence of hiatuses. Noteworthy, the $\delta^{18}\text{O}$ values become more positive with depth (and not negative as expected from a strong impact of the diagenesis), and there is no correlation ($R^2=0.05$) between *W. barnesiae* abundance and the $\delta^{18}\text{O}$ record. The latter may imply the absence of a diagenetic imprint on both chemical and micropaleontological proxies, so the oxygen isotope data from borehole CT8 possibly preserve a primary isotopic signal.

5. Discussion

5.1. Biostratigraphy

The base of UC0 / NC10 was previously defined by Burnett (1998) and Bralower et al. (1993, 1995) by the FO of *Eiffellithus turriseiffelii*. Shamrock and Watkins (2009) demonstrated that this species is composite and established *E. casulus*, which differs from *E. turriseiffelii* s.str. in having smaller coccoliths and narrower central opening. Furthermore, they showed that their new species *E. casulus* appeared slightly earlier (~600 kyr) than the amended *E. turriseiffelii* s.str. In borehole CT8, the FO of *E. turriseiffelii* s.str. was fixed in sample 17 (depth 190.70 m), whereas *E. casulus* was recorded in sample 6 (depth 216.90 m). However, the definition of the zonal boundary has never been formally amended and it would be premature to do so on the basis of a single section. Therefore, at present we draw the base of Zone UC0 (NC10) in our succession by the FO of *E. turriseiffelii* s.l., i.e. *E. turriseiffelii* s.str. + *E. casulus*, bearing in mind that the actual boundary may lie higher. The latter is also supported by reconstructed palaeoenvironmental differences below and above the level of the FO of *E. turriseiffelii* s.str. Investigations of other continuous sections in the Levant Basin and elsewhere will undoubtedly help testing this hypothesis.

5.2. Age inferences for the Albian–Cenomanian interval of borehole CT8

The identified calcareous nannofossil and planktic foraminiferal datums provide a basis for the NW Carmel Region reference age–depth model, i.e. a linear extrapolation of the age vs. depth

(Fig. 8). The Albian–Cenomanian chronostratigraphy of borehole CT8 is based on the Geologic Time Scale 2012 (Ogg and Hinnov, 2012) and constricted by a simple linear relationship of two nearby nanofossil datums. This age model includes the following important bioevents: FO of *E. turriseiffelli* s.str. (large morphotype of *E. turriseiffelii* sensu Ogg and Hinnov, 2012: 1122) (103.13 Ma) at 190.70 m, FO of *C. kennedyi* (100.45 Ma) at 127.15 m, LO of *W. britannica* (100.03 Ma) at 122.50 m, FO of *G. segmentatum* (98.26 Ma) at 116.50 m, LO of *Z. xenotus* (97.73 Ma) at 108.35 m, and FO of *L. acutus* (96.16 Ma) at 105.15 m. Planktic foraminifera datums comprise the FO of *P. appenninica* (101.92 Ma) at 158.7 m, FO of *Th. globotruncanoides* (100.5 Ma) at 132.5 m, and FO of *Th. reicheli* (96.08 Ma) at 101.6 m. The model allows estimation of variations in sedimentation rates according to changes in slope of the correlation line. The sedimentation rate for the Albian–Cenomanian interval in borehole CT8 is expected to range from 0.20–2.37 cm/kyr, with most significant changes in the middle part of the Arqan Fm. within the *Th. globotruncanoides* Zone. The nanofossil biochronology supports the planktic foraminifera datums.

5.3. Integrated stratigraphy (Figs. 2 and 8)

The upper part of the interval of subzones UC0a–b corresponds to the Upper Albian foraminiferal *P. appenninica* Zone (from the level of sample 32). Subzone UC0c correlates with the upper part of *P. appenninica* Zone and lowermost part of *Th. globotruncanoides* Zone of the Lower Cenomanian. The interval of samples 48–54 (127.15–118.55 m) (foraminiferal *Th. globotruncanoides* Zone) is assigned to UC1 Zone of the Lower Cenomanian. The interval of samples 48–52 (127.15–122.50 m) (foraminiferal *Th. globotruncanoides* Zone).

The interval of samples 55–60 (depths 116.50–106.45 m) (foraminiferal *Th. globotruncanoides* Zone) belongs to Zone UC2.

The upper part of the borehole (samples 61–108, depths 105.15–5.00 m) (uppermost part of *Th. globotruncanoides*, *Th. reicheli* and *R. cushmani* foraminiferal zones; Lower–Upper Cenomanian) is referred to the transitional basal Middle–Upper Cenomanian Zone UC3.

5.4. Palaeoenvironmental reconstructions

Over the last couple of decades, Cretaceous climate has been in the focus of several studies, which demonstrated the early Late Cretaceous – Early Turonian to be one of the warmest periods during the past 140 million years (e.g., Huber et al., 2002; Forster et al., 2007; Linnert et al., 2014; Ando et al., 2015). Tropical sea surface temperatures (>35°C) were estimated for the Late

Cenomanian–Early Turonian with a strongly reduced latitudinal gradient (Linnert et al., 2014; O’Brien et al., 2017).

In this regard, the improved knowledge about the palaeoecological affinity of some mid-Cretaceous nannofossil species (e.g., Roth and Krumbach, 1986; Watkins, 1989; Erba et al., 1992; Herrle and Mutterlose, 2003; Herrle et al., 2003; Erba, 2004; Tiraboschi et al., 2009) allows better understanding of the evolution of palaeoenvironments during the mid-Cretaceous. Further on, we discuss temperature and nutrient variations reconstructed in the Carmel area on the basis of calcareous nannofossils and compare them with the results with the Late Albian–Late Cenomanian record of the western Tethys (Bottini and Erba, 2018).

Lipson-Benitah et al. (1995, 1997) recognises five ecostratigraphic intervals coinciding with sea-level changes (Fig. 9). Phase 1 corresponds to shallow marine conditions reflected by low relative abundance of planktic foraminifera. Phase 2 (183.35–166.00 m) in the lowermost part of the Arqan Fm. is associated with a sea-level rise, which commenced earlier. A peak of planktic foraminifera composed of rotaliporids, and occasionally abundant *Hedbergella*, points to the open marine depositional environment during this phase. Phase 3 (163.70–128.50 m) is recorded in the lower part of the Arqan Fm. and is characterised by a conspicuous sea-level drop inferred from an extremely low relative abundance of planktic foraminifera. In the western part of the Carmel Region biomicrites with *Pycnodonte vesiculosa* (Sowerby) in the growth position suggest a low-energy and more deep-water environment (Lipson-Benitah et al., 1995, 1997).

Phase 4 (127.15–97.50 m) characterises the middle part of the Arqan Fm. marked by a rich planktic foraminiferal assemblage, dominated by *Rotalipora brotzeni*, *R. globotruncanoides* and *R. greenhornensis*, and reflects a deeper oligotrophic environment (Lipson-Benitah et al., 1995, 1997). Species of *Hedbergella* or *Gubkinella* are occasionally numerous, and their abundance peaks may be associated with increased nutrient levels and/or a change in the stratification of the water column. Phase 5 (95.50–32.65 m) is recognised in the upper part of the Arqan Fm. and is characterised by a decrease in the relative abundance of planktic foraminifera, which may be connected to periodic sea-level drops (Lipson-Benitah et al., 1995, 1997).

The calcareous nannoplankton Temperature (TI) and Nutrient (NI) indices show some correspondence with quantitative data by planktic foraminifera and with cycles and sequences of Haq et al. (1988) and Robaszynski et al. (1993) (Fig. 9). The TI and NI are indicative of variations in surface water temperature and fertility throughout the succession, but temperature

shows irregular fluctuations, which are not necessarily associated with changes in fertility, suggesting that temperature and fertility varied independently.

In the Late Albian (nannofossil Subzone NC9b and lower part of Subzone NC10a) temperatures were relatively cooler and surface water fertility was low. Fertility increased in the middle part of Subzone NC10a under intermediate temperatures. During the *P. appenninica* foraminiferal Zone, surface water conditions shifted towards warmer temperatures and lower fertility. A minor increase in fertility and a relative decrease in temperature marked the end of the *P. appenninica* Zone. The Lower Cenomanian (Subzone NC10b) was characterised by the warmest temperature of the studied interval and a lower NI. In the Middle and Late Cenomanian (Zone NC11*), fertility was generally lower except for a peak at the base of Zone NC11*. The average temperature was cooler although the TI shows alternating cooler and warmer phases.

In relation to the planktic foraminifera ecostratigraphic intervals in borehole CT2 (Lipson-Benitah et al., 1995, 1997), relatively higher total nannofossil abundance and peaks in the NI characterise the equivalent intervals of phases 2 and 4 of Lipson-Benitah et al. (1995, 1997). The correspondence between higher surface water fertility conditions and transgressive phases is somehow contradictory. A possible explanation is that during a transgression phase, surface water fertility is reduced since phosphorous is extracted by sedimentation in the newly flooded shelf. Under such condition, the nitrogen-limited primary producers are favoured. Since coccolithophores are directly depending on nitrogen availability, the NI is shifted towards higher values.

Regarding the paleotemperature conditions of the Carmel area during the Late Albian–Cenomanian interval, there is a noticeable discrepancy between the two temperature proxies for the TI and the $\delta^{18}\text{O}$ curves show opposite trends (Figs. 6 and 7). The oxygen isotopes suggests temperate conditions during the Late Albian–Early Cenomanian followed by a warming period in the late Early Cenomanian, with temperatures gradually increasing toward the Middle Cenomanian. Evidence for a gradual increase from $\sim 30^\circ\text{C}$ in the Albian to $\sim 31^\circ\text{C}$ during the Cenomanian were provided by oxygen isotope data on phosphatic fish remains from the Cretaceous to Eocene of Israel (Kolodny and Raab, 1988), but the low resolution of that dataset does not allow for a reliable comparison. The oxygen isotope record does not show any evidence for a diagenetic alteration of the primary signal, thus the differences between the TI and the oxygen isotopes may depend on the relatively low abundances of temperature indicative species

used in the calculation of the TI. Compared to coeval sections in the Tethys (Bottini & Erba 2018), *E. floralis* and *Z. diplogrammus* have similar average abundances and trends, while *Rh. asper* is less abundant. Despite this difference, the palaeotemperature fluctuations indicated by the two TI records are comparable and, similarly to the Carmel area, the Cenomanian TI of the Tethys suggests cooler conditions with respect to the $\delta^{18}\text{O}$. In the following section we discuss two possible correlations between the Tethys and the Carmel area records.

5.5. Correlation with the western Tethys record

The studied section is correlated with the western Tethys (Bottini and Erba, 2018), which yields the only continuous record of the TI and NI through the Late Albian–Late Cenomanian interval (Fig. 10). The nannofossil data take into account SSTs derived from TEX_{86} analyses (Forster et al., 2007; O'Brien et al., 2017) and were interpreted to trace supra-regional palaeoenvironmental conditions being similar to nannofossil datasets from other basins (e.g., Hardas et al., 2012; Kanungo et al., 2018). Two correlations are proposed and discussed herein.

Correlation 1 is based on nannofossil and foraminiferal biozones, and shows no evidence of missing biostratigraphic intervals, thus suggesting that the Late Albian–Late Cenomanian interval is complete in the studied section. However, the carbon signal in borehole CT8 does not reflect the globally recorded carbon isotopic anomalies (e.g., Jarvis et al., 2006; Gambacorta et al., 2015; Giorgioni et al., 2015; Bornemann et al., 2017) that mark the Oceanic Anoxic Event (OAE) 1d and Mid Cenomanian Event (MCE) intervals. It is therefore plausible that the isotopic signal is diagenetically altered. It means that the $\delta^{18}\text{O}$ signal is affected too and, consequently, it cannot be used for palaeotemperature reconstructions. This scenario implies that there is no similar pattern of changes between the TI and NI compared to the western Tethys record (Bottini and Erba, 2018), and that the NI and TI of the Carmel area reflect very local fluctuations in this part of the Levant Basin.

Correlation 2 takes into account the biostratigraphic data as well as the $\delta^{13}\text{C}$ curve. The latter suggests that the OAE 1d interval is missing due to a hiatus coinciding with the lithostratigraphic boundary between the Tavasim Tuff (V_2) of the Isfiye Fm. and the Arqan Fm. The $\delta^{13}\text{C}$ curve does not display any positive anomaly that can be correlated with the OAE 1d positive carbon isotopic shift elsewhere. The $\delta^{13}\text{C}$ curve shows instead a decreasing trend, which may represent the post-OAE 1d positive carbon isotopic anomaly starting around 101.5 Ma (Fig. 10). The absence of a positive carbon isotope anomaly in the Middle Cenomanian of the studied section,

does not allow identification of the MCE. However, neither the biostratigraphic data nor chemostratigraphy permit to state whether the MCE interval is recorded or elided. The carbon isotope curve across the Late Cenomanian is characterised by negative values which do not find correspondence with other isotopic records worldwide, and are interpreted to be diagenetically altered. Contrary to Correlation 1, in the second scenario the TI and NI of the CT8 section show main trends comparable to those in the western Tethys. In particular, the cooling spike identified in Subzone NC9b and the lower part of Subzone NC10a, at the base of the section, may correspond to similar lower temperatures registered in the Tethys within the same subzones. The fertility spike detected in the middle of Subzone NC10a possibly corresponds to the fertility peak marking the end of the OAE 1d. In both records a warm interlude is paralleled by low fertility around ca. 98 Ma, followed by a minor cooling and a fertility spike. Another similarity is identified in the progressive cooling trend of the Middle–Late Cenomanian, which is marked by alternating phases of warming and cooling. The presence of relatively similar variations in the TI and NI suggest that Correlation 2 is more plausible compared to Correlation 1. On the basis of Correlation 2, the Carmel area appears to be characterised by palaeoclimatic and palaeoenvironmental conditions occurring during the Late Albian–Late Cenomanian at least at the supra-regional scale.

6. Conclusions

For the first time, we document the taxonomic composition and stratigraphic distribution of calcareous nannofossils from the Albian and Cenomanian of the Carmel area, NW Israel. The identified assemblage consists of 113 taxa.

The detailed calcareous nannofossil biostratigraphy of this area has been established and compared with planktic foraminiferal zones (Figs. 2 and 8).

The new nannofossil data indicate that the Isfiye Fm. and the Tavasim Tuff (V₂) are Late Albian.

The quantitative analysis of the calcareous nannoplankton suggests that the general dominance of *Watznaueria* spp. throughout the whole succession of borehole CT8 points to quite warm, open marine or coastal, generally oligotrophic conditions. Due to poor nutrient supply, the productivity of the calcareous nannoplankton was quite low except for two phases of higher fertility within the middle part of Subzone NC10a and Zone NC11*. Low values of the Shannon index, Evenness and Species Richness can be interpreted as reflecting unstable environment. The

Isfiye Fm. (Late Albian) accumulated in temperate, oligotrophic conditions. The lower part of the Arqan Fm. was deposited under temperate climate but mesotrophic conditions. The higher part of the Arqan Fm. (Early Cenomanian) was deposited in oligotrophic waters and relatively warm climate. The uppermost part of the Arqan Fm. (Middle–Late Cenomanian) was accumulated during a progressively cooling period, although characterised by alternating warming and cooling phases and oligotrophic conditions.

The calcareous nannoplankton quantitative data correlate very well with those obtained from planktic foraminifera (Lipson-Benitah et al., 1997) from the nearby area, and with cycles and sequences postulated by Haq et al. (1988), and suggest nutrification during phases of a risen sea level in the Late Albian and Early Cenomanian.

The biostratigraphy suggests no hiatuses in the studied section, but carbon isotope data seem to be affected by diagenetic alteration showing no evidence of the positive isotopic anomalies, which identify the OAE 1d and MCE worldwide. The correlation with the western Tethys record suggests that the OAE 1d is possibly missing in the studied section due to a hiatus at the Isfiye Fm. / Arqan Fm. boundary. Temperature and nutrient variations in the studied section correspond to the western Tethys record, thus implying that the detected palaeoclimatic variations occurred at a supra-regional scale.

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Appendix A. Taxonomic appendix

Acuturris Wind & Wise in Wise & Wind, 1977

A. scotus (Risatti, 1973) Wind & Wise in Wise & Wind, 1977

Amphizygus Bukry, 1969
A. brooksii Bukry, 1969
Assipetra Roth, 1973
A. infracretacea (Thierstein, 1973) Roth, 1973
Axopodorhabdus Wind & Wise in Wise, 1977
A. biramiculatus (Stover, 1966) Corbett & Watkins, 2014
Biscutum Black in Black & Barnes, 1959
B. constans (Górka, 1957) Black in Black & Barnes, 1959
B. thurrowii Burnett, 1997
Biscutum sp.
Braarudosphaera Deflandre, 1947
B. africana Stradner, 1961
B. bigelowii (Gran & Braarud, 1935) Deflandre, 1947
Broinsonia Bukry, 1969
B. cenomanica (Black, 1973) Bown, 2001
B. enormis (Shumenko, 1968) Manivit, 1971
B. matalosa (Stover, 1966) Burnett in Gale et al., 1996
Broinsonia sp.
Bukrylithus Black, 1971
B. ambiguus Black, 1971
Calcicalathina Thierstein, 1971
Calcicalathina sp.
Calciosolenia Gran, 1912
C. fossilis (Deflandre in Deflandre & Fert, 1954) Bown in Kennedy et al., 2000
Calculithus Prins & Sissingh in Sissingh, 1977
C. anfractus (Jakubowski, 1986) Varol & Jakubowski, 1989
Calculites sp.
Chiastozygus Gartner, 1968
Ch. amphipons (Bramlette & Martini, 1964) Gartner, 1968
Ch. litterarius (Górka, 1957) Manivit, 1971
Chiastozygus sp.
Corollithion Stradner, 1961
C. kennedyi Crux, 1981
Conusphaera Trejo, 1969
C. mexicana Trejo, 1969
Cretharhabdus Bramlette & Martini, 1964
C. conicus Bramlette & Martini, 1964
C. striatus (Stradner, 1963) Black, 1973
Cretarhabdus sp.
Cribrosphaerella Deflandre in Piveteau, 1952
C. ehrenbergii (Arkhangelsky, 1912) Deflandre in Piveteau, 1952
Crucibiscutum Jakubowski, 1986
Crucibiscutum sp.
Cyclagelosphaera Noël, 1965
C. margerelii Noël, 1965
C. rotaclypeata Bukry, 1969
Cylindralithus Bramlette & Martini, 1964
Cylindralithus sp.
Discorhabdus Noël, 1965
D. ignotus (Górka, 1957) Perch-Nielsen, 1968
Eiffellithus Reinhardt, 1965
E. casulus Shamrock & Watkins, 2009
E. gorkae Reinhardt, 1965
E. hancockii Burnett, 1997

E. monechiae Crux, 1991
E. turriseiffelli (Deflandre in Deflandre & Fert, 1954) Reinhardt, 1965
Eprolithus Stover, 1966
E. floralis (Stradner, 1962) Stover, 1966
Flabellites Thierstein, 1973
Fl. oblongus (Bukry, 1969) Crux in Crux et al., 1982
Gartnerago Bukry, 1969
G. segmentatum (Stover, 1966) Thierstein, 1974
Gartnerago sp.
Gorkaea Varol & Girgis, 1994
G. operio Varol & Girgis, 1994
G. pseudanthophorus (Bramlette & Martini, 1964) Varol & Girgis, 1994
Haqius Roth, 1978
H. circumradiatus (Stover, 1966) Roth, 1978
Helenea Worsley, 1971
H. chiasia Worsley, 1971
Helicolithus Noël, 1970
Hc. compactus (Bukry, 1969) Varol & Girgis, 1994
Hc. trabeculatus (Górka, 1957) Verbeek, 1977
Isocrystallithus Verbeek 1976
Isocrystallithus sp.
Kamptnerius Deflandre, 1959
K. magnificus Deflandre, 1959
Lithraphidites Deflandre, 1963
L. acutus Verbeek & Manivit in Manivit et al., 1977
L. alatus Thierstein in Roth & Thierstein, 1972
L. carniolensis Deflandre, 1963
L. eccentricus Watkins in Watkins & Bowdler, 1984
L. houghtonii Jeremiah, 2001
Lithraphidites sp.
Loxolithus Noël, 1965
L. armilla (Black in Black & Barnes, 1959) Noël, 1965
Loxolithus sp.
Manivitella Thierstein, 1971
M. pemmatoidea (Deflandre in Manivit, 1965) Thierstein, 1971
Mennerius Luljeva, 1967
Mennerius sp.
Micrantholithus Deflandre in Deflandre & Fert, 1954
Micrantolithus sp.
Microrhabdulus Deflandre, 1959
M. decoratus Deflandre, 1959
Nannoconus Kamptner, 1931
N. bucheri Brönnimann, 1955
N. cornuta Deres & Acheritéguy (1980)
N. elongatus Brönnimann, 1955
N. fragilis Deres & Acheritéguy, 1980
N. scyphoides Kamptner, 1938
N. truittii frequens Deres & Acheritéguy, 1980
N. truittii truittii Brönnimann, 1955
Nannoconus sp. (view from above)
Nannoconus sp. A
Nannoconus sp. B
Placozygus Hoffman, 1970
P. fibuliformis (Reinhardt, 1964) Hoffmann, 1970
Prediscosphaera Vekshina, 1959
P. columnata (Stover, 1966) Perch-Nielsen, 1984

P. cretacea (Arkhangelsky, 1912) Gartner, 1968
P. grandis Perch-Nielsen, 1979
P. spinosa (Bramlette & Martini, 1964) Gartner, 1968
Radiolithus Stover, 1966
R. orbiculatus (Forchheimer, 1972) Varol, 1992
R. planus Stover, 1966
Repagulum Forchheimer, 1972
R. parvidentatum (Deflandre & Fert, 1954) Forchheimer, 1972
Retecapsa Black, 1971
R. angustiforata Black, 1971
R. crenulata (Bramlette & Martini, 1964) Grün in Grün & Allemann, 1975
R. ficula (Stover, 1966) Burnett, 1997
Retecapsa sp.
Rhagodiscus Reinhardt, 1967
Rh. achlyostaurion (Hill, 1976) Doeven, 1983
Rh. amplus Bown, 2005
Rh. angustus (Stradner, 1963) Reinhardt, 1971
Rh. asper (Stradner, 1963) Reinhardt, 1967
Rh. sageri Bown, 2005
Rh. splendens (Deflandre, 1953) Verbeek, 1977
Rhagodiscus sp.
Seribiscutum Filewicz et al. in Wise & Wind, 1977
Seribiscutum sp.
Sollasites Black, 1967
S. horticus (Stradner et al. in Stradner & Adamiker, 1966) Cepek & Hay, 1969
Staurolithites Caratini, 1963
Staurolithites sp.
Tetrapodorhabdus Black, 1971
T. decorus (Deflandre in Deflandre & Fert, 1954) Wind & Wise, 1983
Tranolithus Stover, 1966
T. gabalus Stover, 1966
T. orionatus (Reinhardt, 1966) Reinhardt, 1966
Watznaueria Reinhardt, 1964
W. barnesiae (Black, 1959) Perch-Nielsen, 1968
W. biporta Bukry, 1969
W. britannica (Stradner, 1963) Reinhardt, 1964
W. cf. W. bayackii Worsley, 1971
W. fossacincta (Black, 1971) Bown in Bown & Cooper, 1989
W. manivittiae Bukry, 1973
W. ovata Bukry, 1969
Zeugrhabdotus Reinhardt, 1965
Z. bicrescenticus (Stover, 1966) Burnett in Gale et al., 1996
Z. clarus Bown, 2005
Z. diplogrammus (Deflandre in Deflandre & Fert, 1954) Burnett in Gale et al., 1996
Z. embergeri (Noël, 1958) Perch-Nielsen, 1984
Z. howei Bown in Kennedy et al., 2000
Z. moulladei Bergen, 1998
Z. noeliae Rood et al., 1971
Z. scutula (Bergen, 1994) Rutledge & Bown, 1996
Z. xenotus (Stover, 1966) Burnett in Gale et al., 1996
Zeugrhabdotus sp. (small)

Appendix B. Raw data of calcareous nannofossil counting.

Appendix C. Percentage of species in calcareous nannofossil assemblages.

Appendix D. Nannofossils groups.

Supplementary data (**Appendixes B-D**) associated with this article can be found in the online version.

ACCEPTED MANUSCRIPT

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Explanation to Tables, Figures and Plates.

Figure 1. The location of the studied area in the Carmel Region. Solid circle – studied borehole CT8, open circle – previously studied borehole CT2.

Figure 2. Lithology, vertical distribution of biostratigraphically important nannofossils, and planktic foraminifera events.

Figure 3. Vertical distribution of relative abundances of dominant and subdominant nannofossil taxa from the Carmel Region (NW Israel).

Figure 4. Vertical distribution of relative abundances (<2%) of selected nannofossil taxa from the Carmel Region (NW Israel).

Figure 5. Preservation, vertical distribution of relative nannofossil abundance, Species Richness, Shannon Diversity Index, Evenness, calcareous nannofossil Nutrient and Temperature Indices, 3pt. moving average of the TI (blue-red line) and NI (yellow-green), carbon isotope data from the Carmel Region (NW Israel).

Figure 6. Vertical distribution of relative abundances of calcareous nannofossil productivity indicators, calcareous nannofossil Nutrient and Temperature Indices, oxygen and carbon isotopes data in the Carmel Region (NW Israel).

Figure 7. Vertical distribution of relative abundances of calcareous nannofossil temperature indicators, calcareous nannofossil Nutrient and Temperature Indices, oxygen and carbon isotopes data in the Carmel region (NW Israel).

Figure 8. Age-depth model and derived sediment accumulation rates for the Albian – Cenomanian of the Carmel CT8 borehole, including events both nannofossil and planktonic foraminifera. Circles: nannofossil events, solid black – first occurrence, open – last occurrence; grey rhombs: first occurrences of planktonic foraminifera; MC – Middle Cenomanian.

Figure 9. Comparison of quantitative data of planktic foraminifera (Lipson-Benitah et al., 1997) from the nearby borehole CT2 with total calcareous nannofossil abundance, calcareous nannofossil Nutrient and Temperature Indices from borehole CT8.

Figure 10. Proposed alternative correlations between borehole CT8 (Carmel Region, NW Israel) and the western Tethys record (Bottini and Erba 2018): Correlation 1 is based on calcareous nannofossil and foraminiferal biostratigraphy which indicate that the record is continuous in CT8 but marked by anomalous carbon isotope record and different Nannofossil Nutrient (NI) and Temperature (TI) indices trends compared to the Tethys; Correlation 2 takes into account calcareous nannofossil and foraminiferal biostratigraphy as well as chemostratigraphy and infers the presence of a hiatus eliding OAE 1d. The TI and NI trends are relatively similar to those traced in the Western Tethys but the carbon isotope record largely

differs from the Tethys. Point by point TI and NI are reported together with the 3pt. moving average of the TI (blue-red line) and NI (yellow-green). Time scale is after Ogg and Hinnov (2012).

Plate 1

Calcareous nannofossils from the Albian–Cenomanian from the Mount Carmel Region.

Plate 2

Calcareous nannofossils from the Albian–Cenomanian from the Mount Carmel Region.

Plate 3

Calcareous nannofossils from the Albian–Cenomanian from the Mount Carmel Region.

Table 1. Descriptive statistics of selected nannofossil taxa from 97 samples of borehole CT8.

Species	Mean (%)	Median (%)	Min (%)	Max (%)	SD ¹	Std. error	Skewness	Kurtosis
<i>Watznaueria barnesiae</i>	60.59	61	23.9	84.7	10.34	1.05	-0.66	1.90
<i>Biscutum constans</i>	7.66	6	1.3	35.1	5.46	0.55	2.41	8.05
<i>Watznaueria</i> spp.	3.62	3.3	0.7	9.2	1.58	0.16	1.26	2.35
<i>Zeugrhabdotus bicrescenticus</i>	3.06	2.6	0	8.2	1.63	0.17	0.78	0.20
<i>Cretarhabdus</i> spp/ <i>Retecapsa</i> spp.	2.79	2.6	0.3	6.7	1.28	0.13	0.83	0.87
<i>Zeugrhabdotus diplogrammus</i>	2.10	1.7	0.3	7.4	1.20	0.12	1.53	3.62
<i>Zeugrhabdotus moulladei</i> + <i>Zeugrhabdotus</i> sp. (small)	1.81	1.8	0	3.9	0.74	0.08	0.32	0.76
<i>Tranolithus orionatus</i>	1.65	1	0.3	7.8	1.35	0.14	1.81	4.48
<i>Zeugrhabdotus</i> spp.	1.52	1.3	0	9	1.31	0.13	2.29	10.04
<i>Discorhabdus ignotus</i>	1.37	1.3	0	4.2	0.94	0.10	0.71	0.45
<i>Eiffellithus turriseiffelii</i>	1.24	1	0	4.6	0.94	0.10	1.16	1.61
<i>Zeugrhabdotus embergeri</i>	0.99	0.7	0	4.1	0.86	0.09	1.16	1.31
<i>Prediscosphaera columnata</i>	0.96	1	0	4.6	0.78	0.08	1.70	4.88
<i>Eprolithus floralis</i>	0.83	0.7	0	2.6	0.60	0.06	0.85	0.74
<i>Lithraphidites</i> spp.	0.70	0.3	0	5.4	1.04	0.11	2.45	6.52
<i>Prediscosphaera cretacea</i>	0.67	0.6	0	2.6	0.60	0.06	1.01	0.69
<i>Lithraphidites carniolensis</i>	0.63	0.6	0	2.3	0.59	0.06	1.04	0.73
<i>Rhagodiscus asper</i> / <i>splendens</i>	0.54	0.3	0	2.9	0.52	0.05	1.33	3.31

SD¹ - standart deviation

Table 2. Pearson's correlation matrix of selected calcareous nannofossil taxa. Significant coefficients ($p < 0.01$) are in bold.

	<i>W. barnesiae</i>	<i>B. constans</i>	<i>Watznaueria</i> spp.	<i>Z. bicrescenticus</i>	<i>Cretarhabdus</i> spp./ <i>Retacapsa</i> spp.	<i>Z. diplogrammus</i>	<i>Z. moulladei</i> + <i>Zeugrhabdodus</i> sp. (small)	<i>T. orionatus</i>	<i>Zeugrhabdodus</i> spp.	<i>D. ignotus</i>	<i>E. turriseiffelii</i>	<i>Z. embergeri</i>	<i>P. columnata</i>	<i>E. floralis</i>
<i>W. barnesiae</i>		-0.660	-0.063	-0.501	-0.306	-0.354	-0.113	-0.689	-0.302	-0.195	-0.510	-0.361	-0.260	-0.149
<i>B. constans</i>	-0.660		-0.070	-0.027	-0.087	-0.093	-0.094	0.469	0.033	0.173	0.202	-0.144	0.198	-0.086
<i>Watznaueria</i> spp.	-0.063	-0.070		-0.124	0.113	-0.056	-0.115	-0.109	0.100	-0.224	-0.085	-0.083	0.235	0.110
<i>Z. bicrescenticus</i>	-0.501	-0.027	-0.124		0.213	0.581	0.123	0.416	0.224	0.022	0.321	0.500	-0.037	0.142
<i>Cretarhabdus</i> spp./ <i>Retacapsa</i> spp.	-0.306	-0.087	0.113	0.213		0.240	-0.029	0.141	0.213	-0.153	0.170	0.354	0.256	0.078
<i>Z. diplogrammus</i>	-0.354	-0.093	-0.056	0.581	0.240		0.193	0.319	0.153	-0.259	0.247	0.527	0.071	0.134
<i>Z. moulladei</i> + <i>Zeugrhabdodus</i> sp. (small)	-0.113	-0.094	-0.115	0.123	-0.029	0.193		-0.053	-0.047	0.137	0.302	0.220	-0.016	0.170
<i>Tr. orionatus</i>	-0.689	0.469	-0.109	0.416	0.141	0.319	-0.053		0.146	0.053	0.353	0.418	0.130	-0.122
<i>Zeugrhabdodus</i> spp.	-0.302	0.033	0.100	0.224	0.213	0.153	-0.047	0.146		-0.185	-0.094	0.182	-0.134	-0.068
<i>D. ignotus</i>	-0.195	0.173	-0.224	0.022	-0.153	-0.259	0.137	0.053	-0.185		0.051	-0.143	-0.083	0.182
<i>E. turriseiffelii</i>	-0.510	0.202	-0.085	0.321	0.170	0.247	0.302	0.353	-0.094	0.051		0.459	0.336	0.152
<i>Z. embergeri</i>	-0.361	-0.144	-0.083	0.500	0.354	0.527	0.220	0.418	0.182	-0.143	0.459		0.202	-0.007
<i>P. columnata</i>	-0.260	0.198	0.235	-0.037	0.256	0.071	-0.016	0.130	-0.134	-0.083	0.336	0.202		-0.116
<i>E. floralis</i>	-0.149	-0.086	0.110	0.142	0.078	0.134	0.170	-0.122	-0.068	0.182	0.152	-0.007	-0.116	
<i>Lithraphidites</i> spp.	-0.063	-0.036	-0.018	-0.018	-0.100	-0.195	-0.034	-0.157	-0.110	0.430	-0.107	-0.131	-0.051	0.165
<i>P. cretacea</i>	-0.292	0.046	-0.181	0.264	0.107	0.193	0.022	0.088	-0.093	0.214	0.173	0.044	0.138	0.223
<i>L. carniolensis</i>	-0.315	0.146	-0.111	0.123	-0.020	-0.022	0.125	0.087	-0.078	0.405	0.093	0.000	0.058	0.238
<i>R. asper</i> / <i>R. splendens</i>	-0.307	0.328	-0.203	0.032	-0.186	-0.222	0.058	0.317	0.037	0.362	0.225	-0.103	-0.161	0.002
significant coefficients ($P < 0.01$) are in bold														

Table 3. Factor analysis Varimax normalized Rotation with principal component Extraction based on 18 taxa from borehole CT8, Carmel region. Rotation is R-mode.

	Factor 1	Factor 2	Factor 3
<i>Watznaueria barnesiae</i>	-0.60	-0.19	-0.69
<i>Biscutum constans</i>	-0.02	-0.05	0.85
<i>Watznaueria</i> spp.	0.02	-0.24	-0.16
<i>Zeughrabdotus bicrescenticus</i>	0.72	0.21	0.08
<i>Cretarhabdus</i> spp./ <i>Retecapsa</i> spp.	0.54	-0.07	-0.12
<i>Zeughrabdotus diplogrammus</i>	0.78	0.02	-0.16
<i>Zeughrabdotus moulladei</i> + <i>Zeughrabdotus</i> sp. (small)	0.26	0.29	-0.10
<i>Tranolithus orionatus</i>	0.50	-0.09	0.68
<i>Zeughrabdotus</i> spp.	0.29	-0.26	0.10
<i>Discorhabdus ignotus</i>	-0.26	0.62	0.35
<i>Eiffellithus turriseiffelii</i>	0.56	0.15	0.33
<i>Zeughrabdotus embergeri</i>	0.80	-0.03	-0.03
<i>Prediscosphaera columnata</i>	0.30	-0.08	0.14
<i>Eprolithus floralis</i>	0.15	0.51	-0.19
<i>Lithraphidites</i> spp.	-0.25	0.69	0.01
<i>Prediscosphaera cretacea</i>	0.24	0.66	0.04
<i>Lithraphidites carniolensis</i>	0.02	0.79	0.16
<i>Rhagodiscus asper</i> / <i>Rhagodiscus splendens</i>	-0.20	0.18	0.67
Total Variance percent	21.22	15.16	10.86
Factor Loadings (Varimax normalized) (18 taxa)			

dominant taxa associated and

Highlights

- The nanofloral assemblage the Albian–Cenomanian of the Carmel area, NW Israel, includes 113 species.
- The detailed calcareous nannoplankton biostratigraphy of this area is established for the first time.
- For the first time the age of the Tavasim Tuff (V₂) is biostratigraphically identified as Late Albian.
- The palaeoenvironment of the basin is reconstructed as warm, open marine or coastal, mainly oligotrophic.