Planktonic foraminiferal turnover across the Cenomanian–Turonian boundary (OAE2) in the northeast of the Tethys realm, Kopet-Dagh Basin

BEHNAZ KALANAT[1], MOHAMMAD VAHIDINIA[1✉], HOSSEIN VAZIRI-MOGHADDAM[2] and MOHAMAD HOSSEIN MAHMUDY-GHARAIE[1]

[1]Department of Geology, Faculty of Science, Ferdowsi University of Mashhad, Mashhad, Iran;
Be.kalanat@stu.um.ac.ir, ✉vahidinia@ferdowsi.um.ac.ir, mahmudygharaie@gmail.com
[2]Department of Geology, Faculty of Science, University of Isfahan, Isfahan, Iran; avaziri7304@gmail.com

Abstract: Two Late Cenomanian–Early Turonian (C–T) intervals of the eastern part of the Kopet-Dagh basin, NE Iran have been investigated to evaluate the response of planktonic foraminifera to the geological event OAE2. The Gharesu and Taherabad sections with the thicknesses of 30 m and 22.5 m are composed of shale and marl interbedded with glauconitic sandstone. Three biozones *Rotalipora cushmani*, *Whiteinella archaeocretacea* and *Helvetoglobotruncana helvetica* were recognized based on study of planktonic foraminifera, in these sections. We observed the patterns of planktonic foraminiferal assemblage changes around the C–T boundary and divided this succession into several successive intervals. This study confirms that OAE2 was a long term event. A gradual perturbation in the study successions starts in the interval 1 with low abundance and diversity of planktonic foraminifera. An enhanced oxygen minimum zone (OMZ) occurs in the interval 3 which coincides with a temporary absence of planktonic foraminifera and sedimentation of framboidal pyrite. High diversity of planktonic foraminifera and appearance of new genera in the interval 5 indicate return of normal conditions to the basin. A significant short-term sea surface temperature cooling is also indicated by planktonic foraminiferal turnover and carbonate contents in the interval 2 which is comparable with other parts of the Tethys Ocean, Boreal sea and Atlantic region.

Keywords: biostratigraphy, palaeoecology, planktonic foraminifera, Cenomanian–Turonian boundary, OAE2, Tethyan realm, Kopet-Dagh basin.

Introduction

Oceanic anoxic events (OAEs) were episodes of widespread marine anoxia. There were arguably between two and seven OAEs during the Mid-Cretaceous (Leckie et al. 2002). The most prominent and widespread of these OAEs spans the Cenomanian–Turonian boundary interval and is called Cenomanian–Turonian Boundary Event (CTBE) or OAE2 (Schlanger & Jenkyns 1976; Leckie et al. 2002).

The Cenomanian–Turonian boundary (~93.5 Ma) was a time of transition in the nature of the ocean-climate system. This interval is known as a typical greenhouse period caused largely by increased CO_2 from elevated global igneous activity (Jones & Jenkyns 2001; Poulsen et al. 2001). It is marked by a major warming peak and globally averaged surface temperatures more than 14 °C higher than those of today (Tarduno et al. 1998), ~100–200 m higher sea level than that of today (Haq et al. 1987; Miller et al. 2005), a lack of permanent ice sheets and oceanic circulations (Frakes et al. 1992).

These conditions could have increased the rate of chemical weathering, flux of nutrient to the ocean and biological productivity (Pedersen & Calvert 1990). Production of a lot of organic matter during the time of decreasing oceanic circulations leads to an anoxic condition and superimposition of three crises during the event: wide scale occurrence of black shales (Arthur & Premoli Silva 1982; Luciani & Cobianchi 1999; Coccioni & Luciani 2005), faunal turnovers (Leckie et al. 1998; Keller et al. 2001, 2008; Keller & Pardo 2004; Caron et al. 2006; Soua et al. 2009, 2011) and a positive $\delta^{13}C$ excursion (Paul et al. 1999; Keller et al. 2004).

The C–T boundary has been studied worldwide (Luderer & Kuhnt 1997; Leckie et al. 1998; Luciani & Cobianchi 1999; Keller et al. 2001, 2004, 2008; Coccioni & Luciani 2005; Mort et al. 2008 among others) but this boundary was little known in Iran. Only a few studies have been conducted in the Kopet-Dagh basin. Previous reports from the Shurab and Hmam-Ghale sections in the east of the Kopet-Dagh basin have suggested the presence of oxygen-restricted deposition in the region, based on the presence of dark shale and marl sediments and foraminiferal turnovers (Abdoshahi et al. 2010; Ghoorchaei et al. 2011). Other studies described this boundary as a disconformity (Afshar-Harb 1994; Sadeghi & Foroughi 2004) or paraconformity (Vahidinia et al. 1999) in the Kopet-Dagh basin.

The investigated sections in the Gharesu and Taherabad localities are the only well described Cenomanian–Turonian boundary intervals in the northeastern part of Tethys. It is therefore a unique opportunity to study the changes in fossil

content across the Cenomanian–Turonian boundary and the palaeoceanographic development in this part of the Mid-Cretaceous world.

The turnover of the planktonic foraminifera, is a particular focus of this study because their perturbation has been a sensitive monitor of events during the Cenomanian–Turonian boundary in the previous studies.

Geographical and geological setting of the studied areas

The Kopet-Dagh basin, which stretches over nearly 700 kilometres from the east of the Caspian Sea to NE Iran, Turkmenistan and north Afghanistan (Fig. 1C) is composed of a 5000–7000 m thick sequence of gently folded rocks of Middle Jurassic to Eocene age (Afshar-Harb 1994). The Cretaceous sediments were deposited on the northern shelf of a deeper marine basin separating the Iran Plate from Eurasia (Turan) Plate in the northeast of the Tethys realm (Fig. 1A, B). This sequence reaches more than 3000 m in thickness and seems to represent all stages of the Cretaceous (Stocklin 1968). The Aitamir and Abderaz formations are two major lithostratigraphic units in the Kopet-Dagh basin with Lower to Upper Cretaceous age (Afshar-Harb 1994).

The Late Cenomanian to Early Turonian interval in the Gharesu and Taherabad sections spans the upper part of the

Aitamir Formation and lower part of the Abderaz Formation. The Gharesu section is located a few kilometres west of the city of Kalat (Fig. 1D). This section is composed of 30 m of shale and marl interbedded with glauconitic sandstone. The Taherabad section is 22.5 m thick and is located along the Mashhad-Kalat main road (Fig. 1D). The lithology of this section is similar to the Gharasu section except that the Gharesu section has more sandstone beds than the Taherabad section.

Material and methods

After preliminary sampling to locate the C–T interval, 22 samples spanning 30 m of the Gharesu section and 49 samples spanning 22.5 m of the Taherabad section were finally collected. The samples were taken at 20–50 cm intervals close to the C–T boundary and at 1–2 m intervals farther away from the boundary.

In the laboratory, samples were crushed and put into a mixture of water and a small amount of H_2O_2, then washed through a 53 μm sieve. For planktonic foraminiferal studies, about 250–300 specimens were counted and identified except in the samples around the C–T boundary where foraminifera are rare or absent. Species identification follows that of Caron (1985) and Premoli-Silva & Verga (2004) (Figs. 2, 3). The carbonate contents were measured using the calcimeter method in the laboratory.
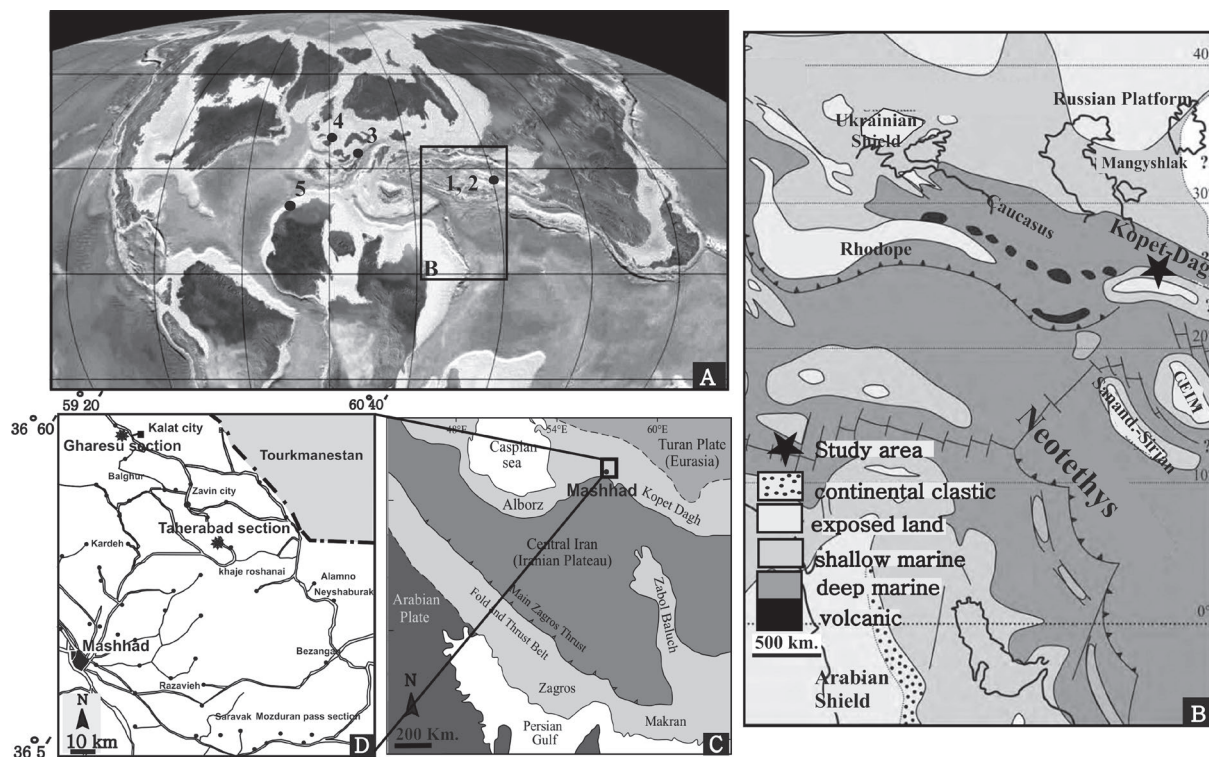

Fig. 1. A — Palaeogeographic map of the Late Cretaceous (Late Cretaceous; © Blakey R.; http://jan.ucc.nau.edu/~rcb7/RCB.html) showing the location of investigated sections in the Kopet-Dagh basin (1, 2) and important sections referred in the text (3 — Pont d'Issole, 4 — Eastbourn, 5 — Tarfaya). The area of map B is outlined; **B** — Cenomanian palaeogeographic map of the Middle East (modified after Philip & Floquet 2000); **C** — structural units of Iran and location of the Kopet-Dagh basin in northeast Iran, southwestern Turkmenistan and north Afghanistan (after Berberian & King 1981); **D** — the Gharesu and Taherabad sections along the Mashhad-Kalat main road.
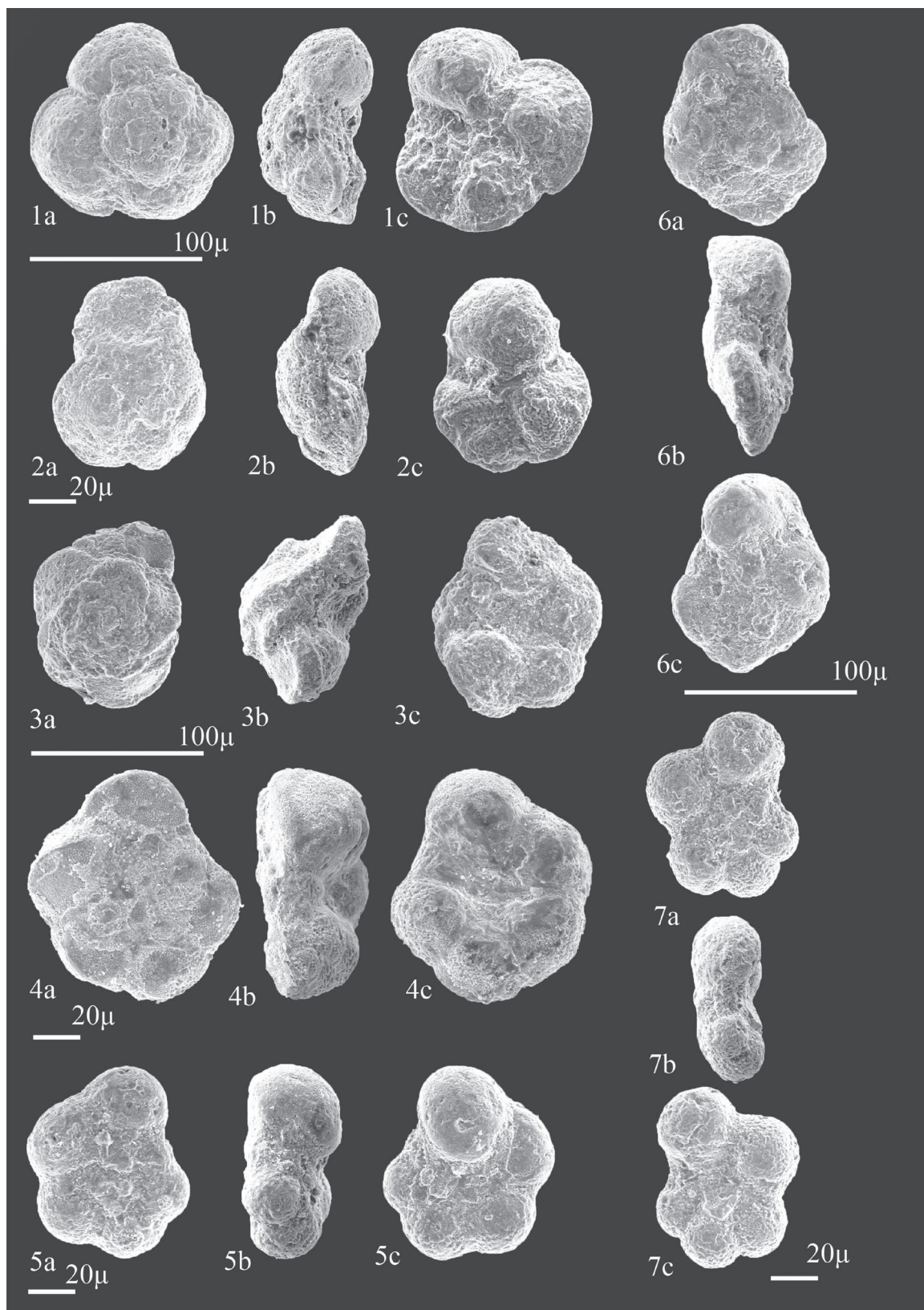
Fig. 2. SEM illustrations of planktonic foraminifera from the Taherabad section ('a' for spiral side, 'b' for lateral side, 'c' for umbilical side). **1a,b,c** — *Rotalipora cushmani* (Morrow), sample T4. **2a,b,c** — *Rotalipora appenninica* (Renz) sample T14. **3a,b,c** — *Praeglobotruncana gibba* Klaus, sample T48. **4a,b,c** — *Helvetoglobotruncana helvetica* (Bolli), sample T46. **5a,b,c** — *Whiteinella praehelvetica* (Trujillo), sample T46. **6a,b,c** — *Dicarinella hagni* (Scheibnerova), sample T46. **7a,b,c** — *Whiteinella archaeocretacea* Pessagno, sample T47.
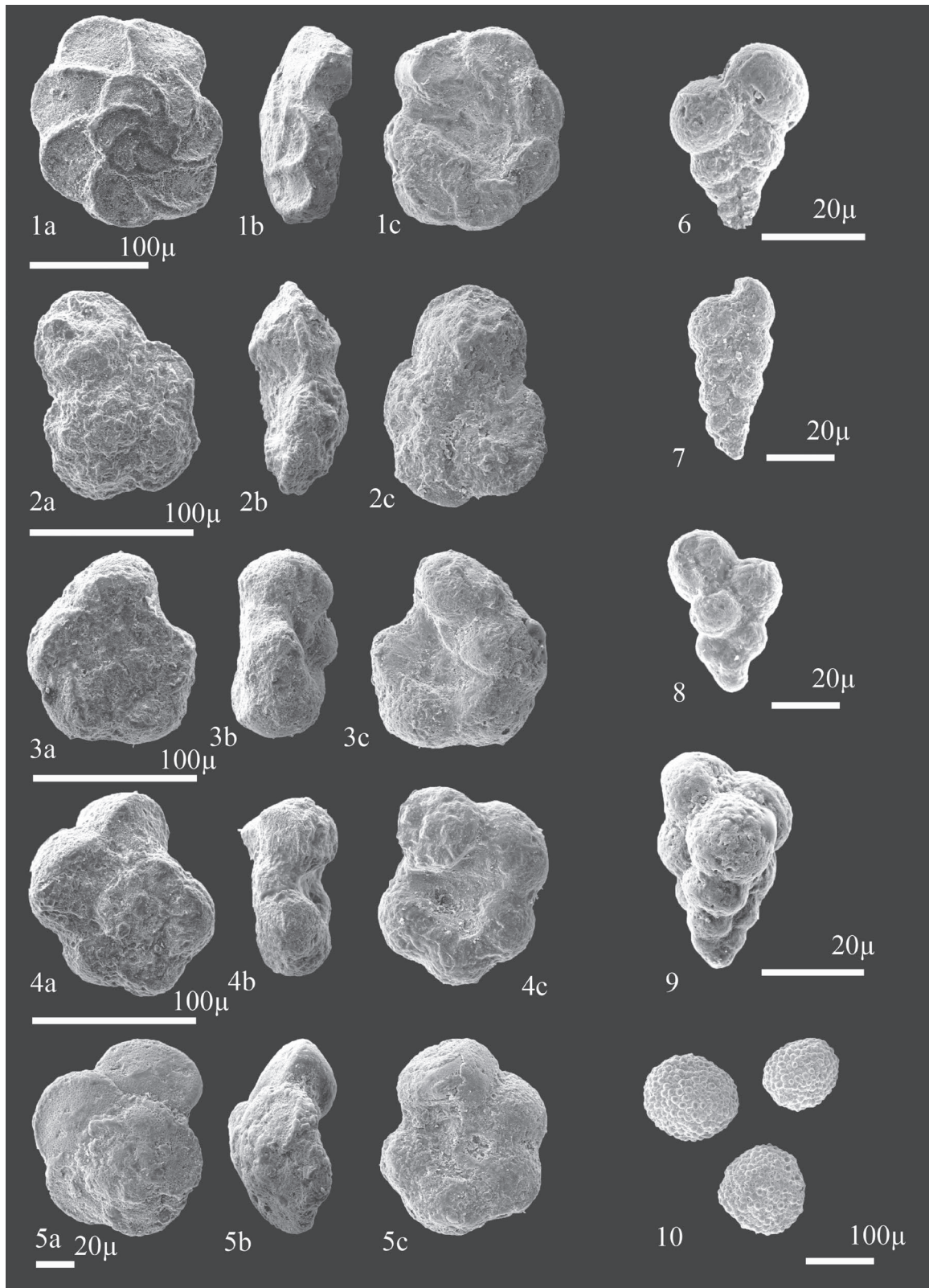
Fig. 3. SEM illustrations of planktonic foraminifera from the Taherabad and Gharesu sections ('a' for spiral side, 'b' for lateral side, 'c' for umbilical side). **1a,b,c** — *Marginotruncana pseudolinneiana* Pessagno, sample gh21. **2a,b,c** — *Rotalipora appenninica* (Renz), sample gh5. **3a,b,c** — *Helvetoglobotruncana helvetica* (Bolli), sample gh21. **4a,b,c** — *Helvetoglobotruncana helvetica* (Bolli), sample gh22. **5a,b,c** — *Praeglobotruncana stephani* (Gandolfi), sample gh6. **6** — *Heterohelix globulosa* (Ehrenberg), sample T47. **7** — *Heterohelix moremani* (Cushman), sample T48. **8** — *Guembelitria cenomana* Keller, sample T46. **9** — *Guembelitria cenomana* Keller, sample gh21. **10** — framboidal pyrite, sample T2.

Biostratigraphy

Despite ammonites, changes in the planktonic foraminiferal assemblages are less indicative and seem to occur over a broad interval of time coeval with the contemporary oceanic environmental perturbation across the Cenomanian–Turonian boundary.

The foraminiferal biostratigraphy of the Cretaceous has been studied in great detail by many authors. Wonders (1980) proposed that C–T boundary corresponds to the base of the *Whiteinella archeocretacea* zone. But subsequent studies (Caron 1985; Robaszynski et al. 1993; Robaszynski & Caron 1995; Hardenbol et al. 1998; Premoli-Silva & Verga 2004; Zaghbib-Turki & Soua 2013) indicated that this boundary is placed in the middle of the *W. archeocrheetacea* zone (Fig. 4).

28 species belonging to 11 genera and 22 species belonging to 10 genera of planktonic foraminifera were recognized respectively in the Taherabad and Gharesu sections. The stratigraphic distributions of these planktonic foraminifera are plotted in Figs. 5 and 6.

Based on the planktonic foraminiferal zonal scheme of Premoli-Silva & Verga (2004), the following planktonic foraminiferal biozones were identified in the sections:

***Rotalipora cushmani* Total Range Zone:** This zone is defined by the first and last occurrences of *Rotalipora cushmani* and corresponds to the Middle–Late Cenomanian. Premoli-Silva & Verga (2004) divided this zone into two subzones (*Rotalipora greenhornensis* and *Dicarinella algeriana*) which are differentiated by the first appearance of *D. algeriana*.

Rotaliporids are not present from the base of this zone in our sections, but this group appears from a height of 2.5 m in the Taherabad section and 10.5 m in the Gharesu section. Rotaliporids (*R. cushmani, R. appenninica*) are scarce in this biozone and are associated with whiteinellids, heterohelicids, muricohedbergellids and dicarinellids.

In the Gharesu section the first appearance of rotaliporids coincides with the first appearance of dicarinellids, but in the Taherabad section *D. algeriana* appears before rotaliporids. The absence of rotaliporids and dicarinellids in the base of sections resulted from eutrophic condition and/or low oxygen content or by shallower marine environmental conditions. So we suggest the interval from the base of sections to the last occurrence of *R. cushmani* belongs to the *D. algeriana* subzone.

***Whiteinella archaeocretacea* Partial Range Zone:** This zone spans the interval from the last occurrence of *R. cushmani* to the first occurrence of *H. helvetica* and extends from the latest Cenomanian to earliest Turonian.

This biozone appears from a height of 6.5 m in the Taherabad section and 14 m in the Gharesu section. Assemblages in this interval, if present, consist of whiteinellids, muricohedbergellids, heterohelicids and guembelitriids.

***Helvetoglobotruncana helvetica* Total Range Zone:** The base and top of the zone coincide respectively with the first and last occurrences of the index-species *H. helvetica*. This zone ranges from Early to Middle Turonian.

We recognized the species about 19.5 m and 29 m above the base of the Taherabad and Gharesu sections, respectively. The most important and common foraminifera in this zone are: dicarinellids, whiteinellids, muricohedbergelids, heterohelicids and praeglobotruncanids. Also Marginotrucana species are present in the Gharesu section.

Palaeo-ecology and planktonic foraminiferal turnover

In the past 20 years, the palaeo-ecology of foraminifera became a dominant field of micropalaeontology. Significant changes in foraminiferal assemblages can be interpreted as reflecting an ecological response to palaeo-oceanographic variations. Among other foraminifera the ecology of planktonic foraminifera plays an important role in the study of oceanic systems (Nebrigic 2006).

Recent planktonic foraminifera reach their highest diversity within a stratified water column with normal salinity,

Age (Ma)	Stage	Tethys realm biozonation					
		Hardenbol et al., 1998		Wonders, 1980	Caron, 1985	Premoli-Silva & Verga, 2004		*This study*
92	Turonian	*H. helvetica*	*H. helvetica* 91.31	*H. helvetica*	*H. helvetica*	*H. helvetica*		*H. helvetica*
	93.5	*W. archaeocretacea*	*H. helvetica* 93.29	*W. archaeocretacea*	*W. archaeocretacea*	*W. archaeocretacea*		*W. archaeocretacea*
94	Cenomanian	*R. cushmani*	*R. cushmani* 93.9	*R. cushmani*	*R. cushmani*	*R. cushmani*	*D. algeriana*	*R. cushmani*
			W. archaeocretacea 94.5				*R. greenhornensis*	
			R. cushmani 95.39					

D: *Dicarinella*   H: *Helvetoglobotruncana*   R: *Rotalipora*   W: *Whiteinella*

Fig. 4. Comparison of standard zonations of the Tethyan Realm with study sections. Time scales of Gradstein et al. (1995) and Hardenbol et al. (1998) are adopted.
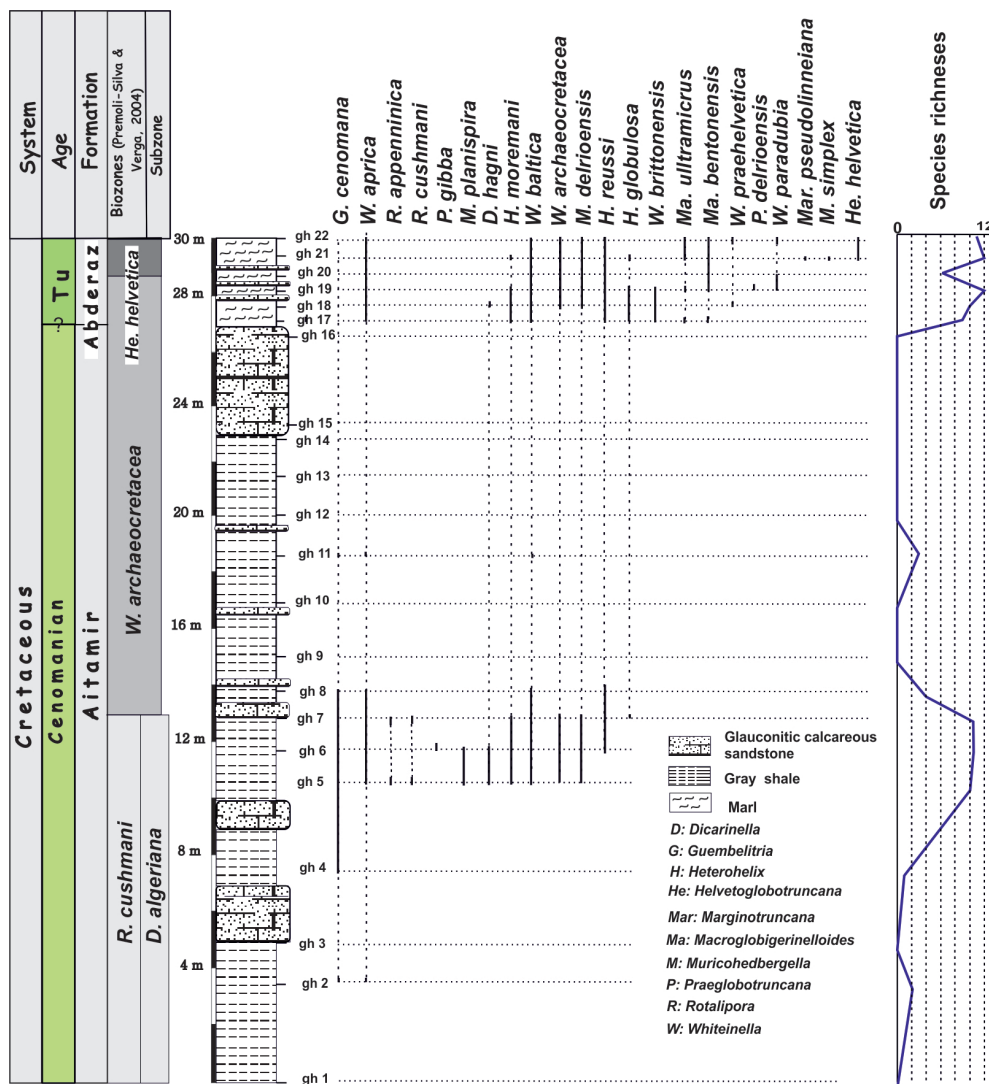
Fig. 5. Distribution and species richness of planktonic foraminifera in the Gharesu section.

nutrient and oxygen content (Hart 1980a,b; Leckie et al. 1998; Hallock et al. 1991; Keller et al. 2001; Keller & Pardo 2004). In these conditions large, keeled, deep-water dwelling forms (K strategists, such as: *Rotalipora* and *Helvetoglobotruncana*) increase. But when environmental conditions are more extreme, diversity is low and opportunistic taxa (r strategists, such as: *Heterohelix* and *Muricohedbergella*) are common (Keller et al. 2001; Coccioni and Luciani 2005). These simpler morphotypes with little surface ornamentation and thin test walls generally occupy the surface mixed layers (above thermocline), or shallow waters with generally unstable and/or eutrophic conditions (Hart 1980a,b; Caron & Homewood 1983; Leckie et al. 1998; Keller 1988; Li & Keller 1998; Keller et al. 2001).

In order to interpret foraminiferal distribution across the C–T transition, Coccioni and Luciani (2005) summarized published data on stable isotopic analysis together with depth ranking based on environmental inferences (morphology and palaeobiogeographic distribution) in their article (Fig. 7).

Based on changes in planktonic foraminiferal distribution and diversity, we divided the Gharesu and Taherabad sections into 5 discrete intervals:

Interval 1: This interval (samples gh1–gh4 in Gharesu section and T0–T3 in the Taherabad section, the lower part of *R. cushmani* zone) is characterized by low species diversity and abundance and low carbonate content (Fig. 8). In the Gharesu section planktonic foraminifera are not present or too scarce to be counted. In the Taherabad section simple morphotype of planktonic foraminifera with globular chambers and trochospiral, biserial and triserial test (*Whiteinella*, *Muricohedbergella*, *Heterohelix* and *Guembelitria*) are present. Keeled forms (*Dicarinella algeriana*) are present in very low abundance in sample T1 (Fig. 9).

We propose this interval was deposited in a marine environment with high terrigenous influx and eutrophic condition. Framboidal pyrite (Fig. 3-*10*) is also present in all samples of this interval. These framboids are 50–150 µm in diameter and are composed of tiny crystallites of 2–5 µm which are spheroidal, cubic, octahedra or irregular in shape. The formation of framboidal pyrite in sediments or during sedimentation requires an anaerobic environment (Schallreuter 1982). Under low oxygen content conditions, sulphate-reducing bacteria produce hydrogen sulphide which reacts with available iron to form microconcretions of pyrite, possibly by the mechanisms suggested by Berner (1969).

Interval 2: At the onset of this interval (samples gh5–gh7 in the Gharesu section and T4–T14 in the Taherabad section, the upper part of the *R. cushmani* zone), rotaliporids appear in both sections. A marked increase in diversity and abundance of planktonic foraminifera occurs in this interval, as the number of species in the Gharesu section increases from 1 to 10 and in the Taherabad section from 6 to 15 (Fig. 9), this is accompanied by a relative increase of CaCO₃ content (Figs. 8, 9) which suggests, terrigenous input decreases in
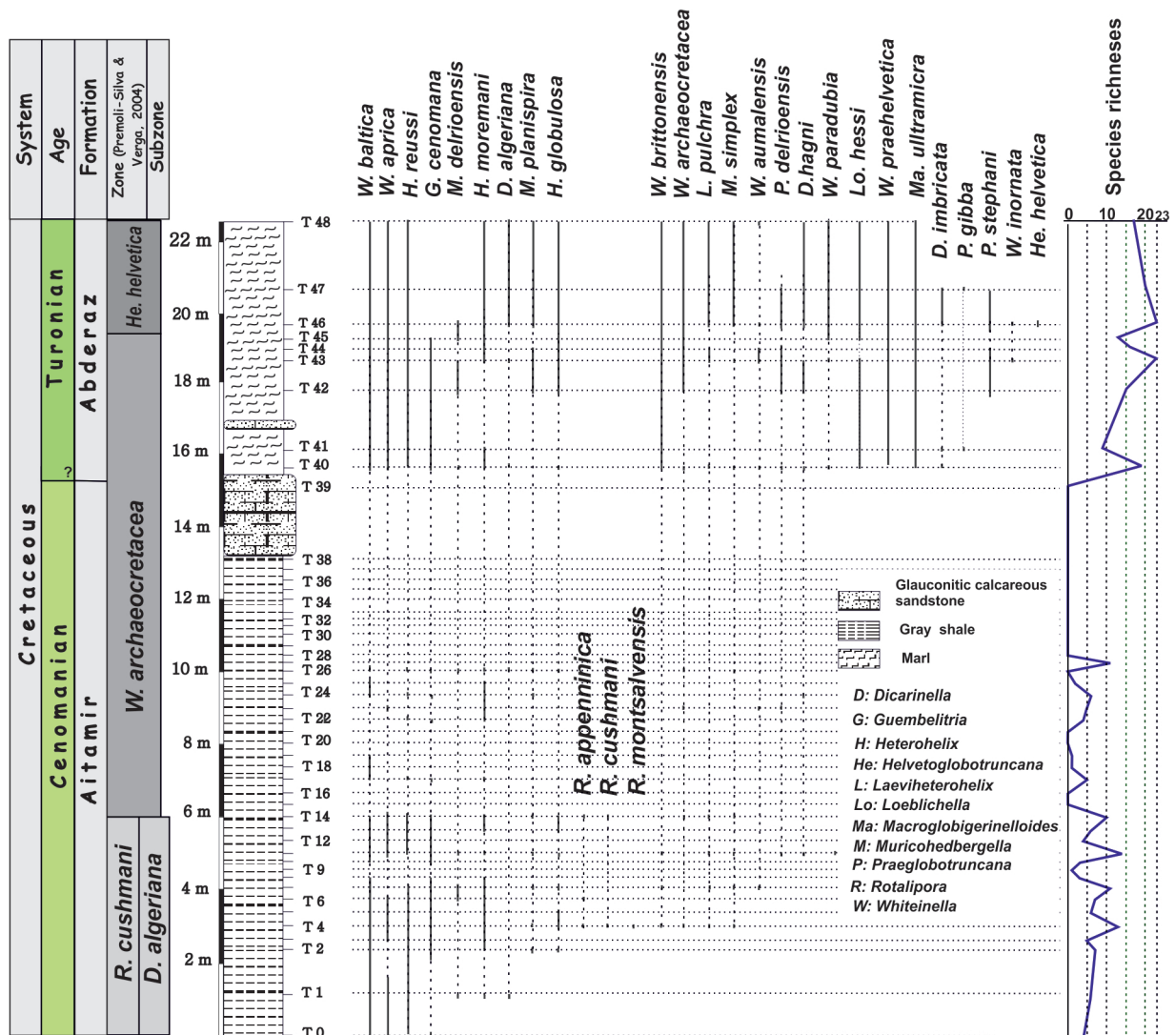
Fig. 6. Distribution and species richness of planktonic foraminifera in the Taherabad section.

this interval. However, the condition in this interval is more stable, a relatively high percentage of *Whiteinella*, *Muricohedbergella* and *Guembelitria* and low abundance of keeled forms (2–5 % in samples which are present) indicate a general mesotrophic regime in interval 2.

Interval 3: This interval (samples gh8–gh16 in the Gharesu section and T15–T39 in the Taherabad section, the lower part of the *W. archaeocretacea* zone) starts with the disappearance of rotaliporids and ends with a thick glauconitic sandstone. At the base of this phase, a dramatic
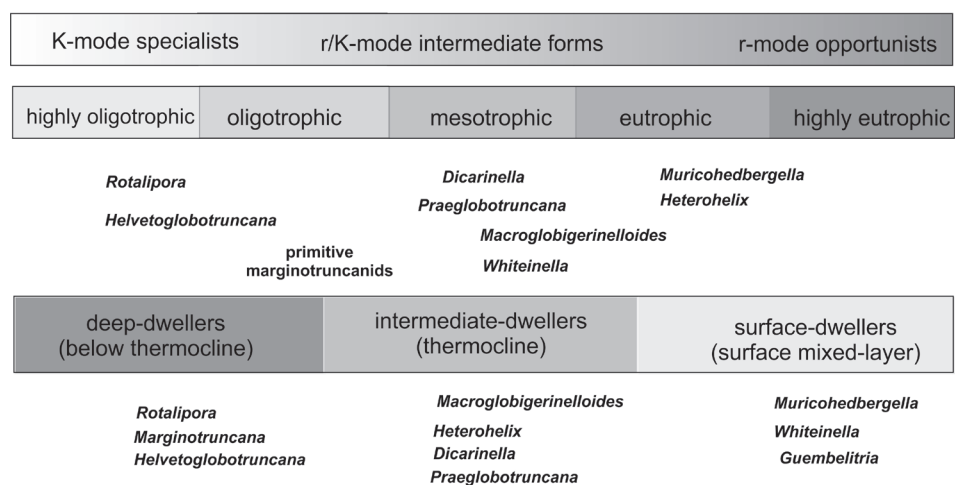

Fig. 7. Inferred life strategy of upper Cenomanian and lower Turonian planktonic foraminifera derived mainly from latitudinal distribution, abundance and depth ranking based on environmental inferences (morphology and biogeographic distribution), and stable isotopic data, plotted against the oceanic surface trophic resources continuum (modified after Coccioni & Luciani 2005).
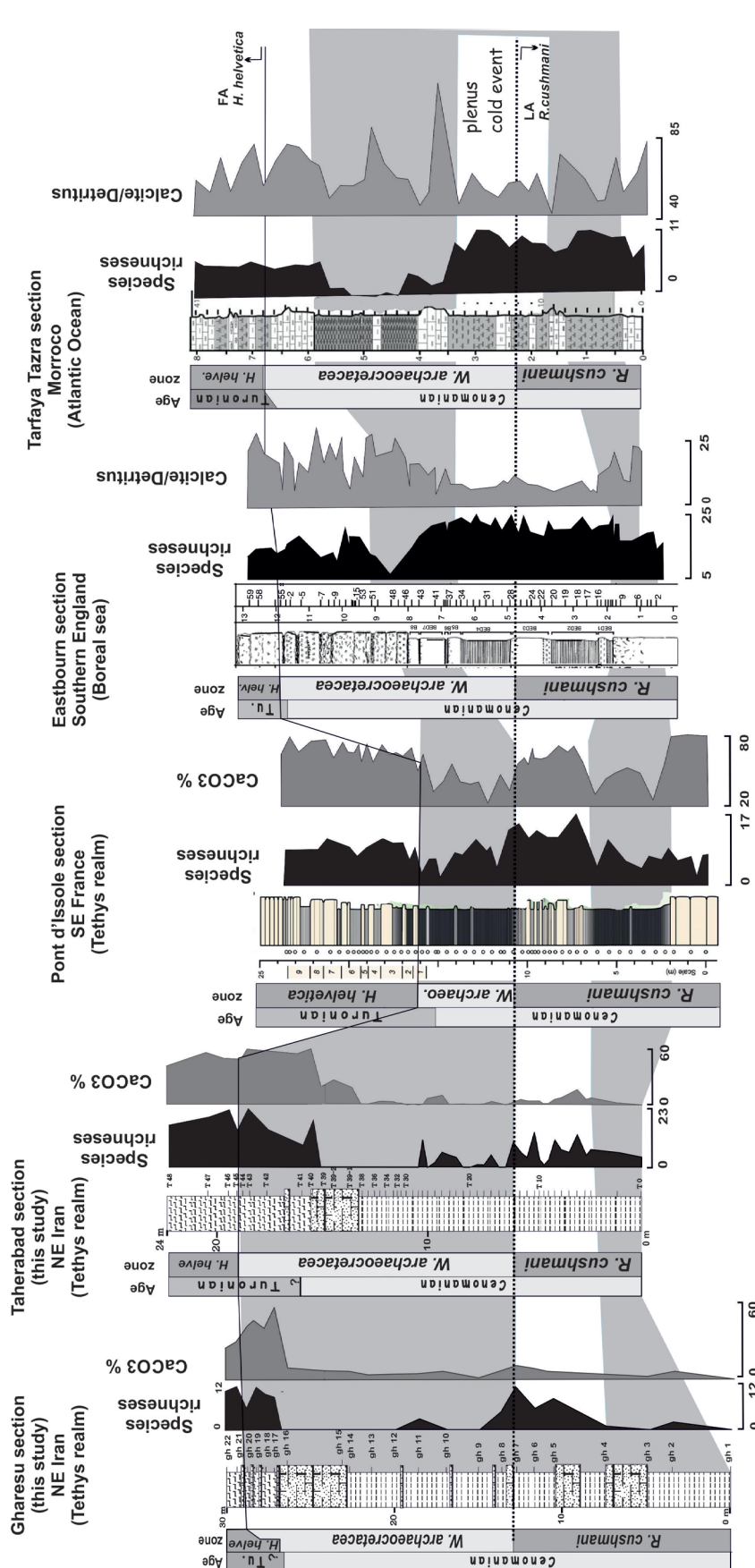
Fig. 8. Species richness and carbonate content profiles for Tethys, Boreal and Atlantic sites. Tie lines are based on biostratigraphic correlation. Horizontal white band indicates the extent of the sea surface temperature cooling. Horizontal grey band shows expansion of OAE2 in the sections. Pont d'Issole data are from Grosheny et al. (2006) and Jarvis et al. (2011), Eastbourne data are from Keller et al. (2001) and Tarfaya data are from Keller et al. (2008).

decrease in diversity and abundance of planktonic foraminifera occurs. Planktonic foraminifera are absent except for sample gh11 in the Gharesu section and samples 17–19, 22–26 in the Taherabad section. The assemblage contains dwarfed and scattered *Guembelitria* and *Whiteinella* in the Gharesu section and *Guembelitria*, *Heterohelix*, *Whiteinella* and *Muricohedbergella* in the Taherabad section. Keeled morphotypes are present in very low abundance (almost 2 %) in samples 23, 24, 26 in the Taherabad section (Fig. 9).

All the above, low percentage of carbonate, presence of sandstone beds and framboidal pyrite in the sediments, indicate sedimentation occurs in a marine environment with high productivity and low oxygen content.

The presence of benthic foraminifera in samples where planktonic foraminifera disappear temporarily indicates that the water column and bottom water is not extremely depleted in oxygen content and absence of planktonic foraminifera is a result of a high detrital input and productivity beside a developed oxygen minimum zone (OMZ).

Interval 4: At the onset of this interval (samples gh17–gh20 in the Gharesu section and samples T40–T45 in the Taherabad section, the upper part of the *W. archaeocretacea* zone) planktonic foraminifera reappear with higher abundance and diversity. Heterohelicids dominate the assemblage. *Whiteinella* and *Muricohedbergella* are present with high abundance. *Guembelitria* are very rare in this interval. Keeled forms are present in all samples from the Taherabad section and in
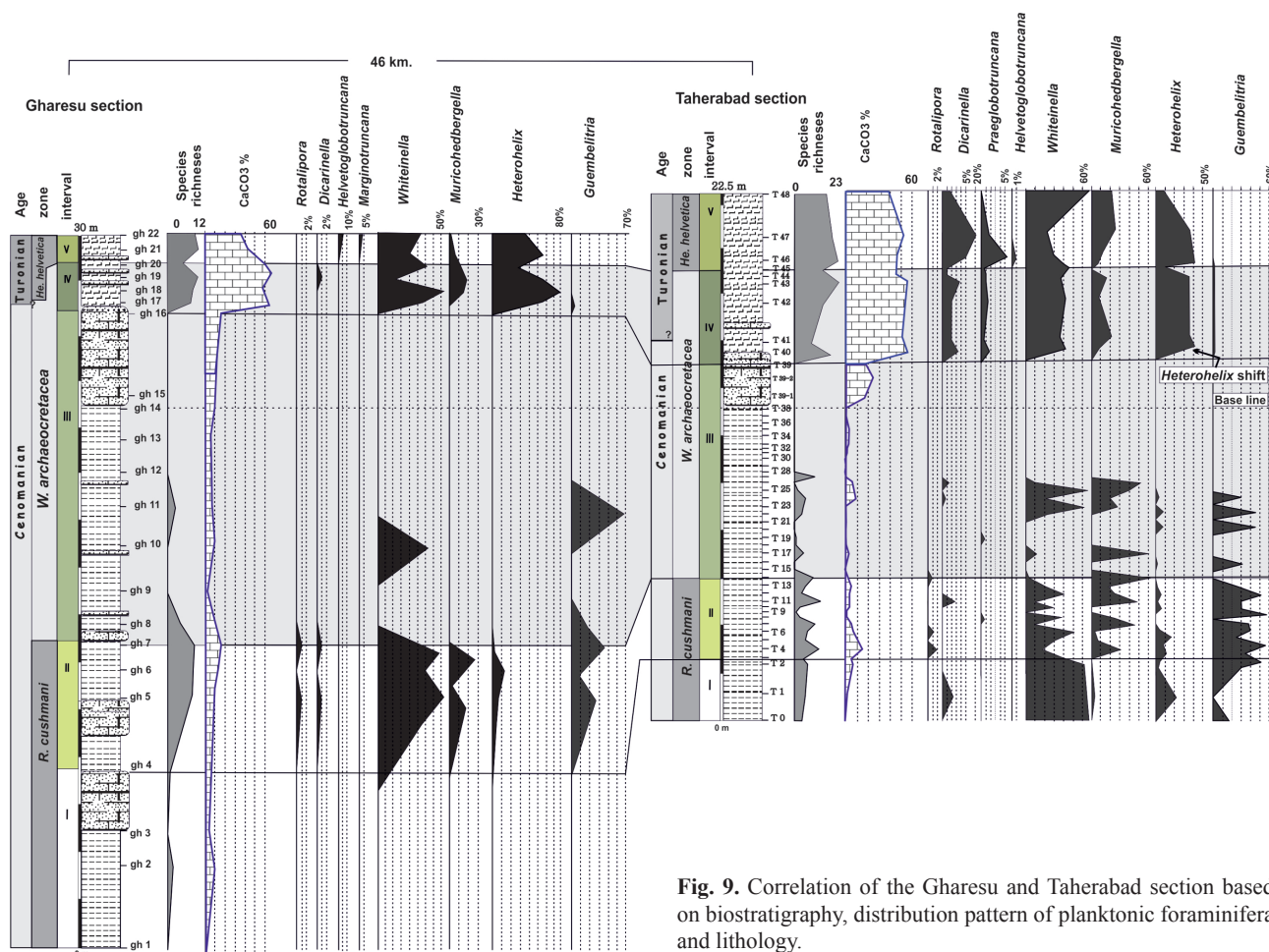
Fig. 9. Correlation of the Gharesu and Taherabad section based on biostratigraphy, distribution pattern of planktonic foraminifera and lithology.

sample 18 from the Gharesu section (although with low abundance ~4–5 %) (Fig. 9).

This interval coincident with maximum *Heterohelix* abundance. This event (the so called *Heterohelix* shift) was also observed at the same stratigraphic interval in the world (Leckie et al. 1998; Keller et al. 2001, 2008; Keller & Pardo 2004; Zaghbib-Turki & Soua 2013) and is an excellent biomarker for global correlations. *Heterohelix* shift appears to be associated with a global expansion of the oxygen minimum zone (Keller & Pardo 2004, Zaghbib-Turki & Soua 2013).

Increasing of planktonic foraminiferal abundance and carbonate content, show that terrigenous input is low in this interval.

Interval 5: This interval (samples gh21–gh22 in the Gharesu section, samples T46–T49 in the Taherabad section, *H. helvetica* zone) starts with the appearance of *Helvetoglobotruncana helvetica*. The diversity and abundance of planktonic foraminifera are higher than in other intervals (Fig. 9).

The appearance of new genera (*Helvetoglobotruncana* and *Marginotruncana* in the Gharesu section and *Helvetoglobotruncana* in the Taherabad section) and abundance of keeled forms indicate this interval is deposited under a relatively normal salinity, oxygen content and oligotrophic condition.

Guembelitria disappear in this interval because this genus thrived in eutrophic surface waters of shallow marginal marine environments with variable salinities at times of severe ecological stress (Keller 2002; Keller et al. 2002).

Correlation and discussion

Lower abundance and number of species of planktonic foraminifera, lower keeled forms and more sandstone intercalation in the Gharesu section suggest that this section was deposited in a shallower water environment than the Taherabad section but all other changes in the sea level and appearance and disappearance of species are comparable in these sections (Fig. 9).

In the Gharesu and Taherabad sections, decrease of oxygen content starts in interval 1 and continues in intervals 2, 3 and 4, as is revealed by the presence of framboidal pyrite in the deposits and planktonic foraminiferal content. High numbers of the opportunists and *Heterohelix* shift suggest that the environmental perturbation related to the CTBE did not end during interval 3 but continued into interval 4.

It seems that, recovery phase occurs in the *Helvetoglobotruncana helvetica* zone (interval 5) and is marked by an

increase of both abundance and diversity of planktonic foraminifera.

Good correlation between diversity of planktonic foraminifera and $CaCO_3$ content suggests that detrital input and productivity beside oxygen content are effective factors in patterns of change in planktonic foraminiferal assemblages in the Gharesu and Taherabad sections, as in the intervals 1 and 3, high detrital input leads to a eutrophic condition and decreases the diversity of planktonic foraminifera (Fig. 9).

The Late Cenomanian sea surface temperature cooling

On a global scale, the C–T boundary is characterized by its warm climate interval accompanied by a major sea level transgression (Hancock & Kauffman 1979; Haq et al. 1987; Uličný et al. 1993; Hardenbol et al. 1998). However, many authors have described an ephemeral but significant decrease in sea surface temperatures at the onset of OAE2 in some CTB sections (Lamolda et al. 1994; Forster et al. 2007; Jarvis et al. 2011; Kaiho et al. 2014 among others). This cooling has been attributed to a drop in atmospheric pCO_2 (Symbol for the negative decadic logarithm of the CO_2 concentration) levels which in turn was caused by enhanced carbon sequestration by burial of organic matter (Hasegawa et al. 2003; Forster et al. 2007). This event was first recognized as an incursion of boreal fauna in the shelf seas of NW Europe (e.g., Jefferies 1962; Gale & Christensen 1996; Voigt et al. 2004) and called the Plenus Cold Event. In the Western Interior Seaway of North America and in the equatorial proto-Atlantic, this interval was characterized by repopulation of the seafloor by benthic foraminifera (the Benthic Oxic Event), indicating regional reoxygenation of bottom waters (Lekie et al. 1998; Keller et al. 2001, 2008; Keller & Pardo 2004).

This short-lived cooling event in the Gahresu and Taherabad sections is characterized by a decreased detrital input and increased carbonate production and diversity of planktonic foraminifera in interval 2, at the end of the *R. cushmani* zone (Fig. 9).

This event is compared with those at Pont d'Issole (France, Tethys Ocean), Eastbourne (England, Boreal sea) and Tazra (Morocco, Atlantic region) in Figure 8.

The record of Pont d'Issole in France (Tethys Ocean) (Grosheny et al. 2006; Jarvis et al. 2011) displays two prominent black shale intervals within an interbedded limestone-grey marl succession. This succession is similar to Gharesu and Taherabad in many ways (increase of carbonate and planktonic foraminiferal diversity in cooling interval) but differs from Gharesu and Taherabad by its further depth and sedimentation of black shale before and after the cold event.

The CTB interval at Eastbourne section in England (Boreal sea) consists of chalk and rhythmically bedded marls (Plenus Marls Formation) (Gale et al. 1993). In the Plenus marls, the relative abundance increase of keeled taxa and increase of planktonic foraminiferal diversity, coincident with decreased *Heterohelix* abundances, suggest a well-stratified water mass and reduced oxygen minimum zone (Keller et al. 2001).

In contrast to Gharesu and Taherabad (Tethys realm), marls and marly chalk deposits at Eastbourne (Boreal Sea), with decreased calcite, increased phyllo-silicates, quartz and feldspar were deposited in the Plenus Cold Event period (Keller et al. 2001). This difference in carbonate content of the Plenus Cold Event interval may be explained by the different palaeo-latitude of sections. The Gharesu and Taherabad sections with their lower palaeo-latitude are less effected by sea-level lowstand periods, increased erosion and accelerated detrital input.

As before mentioned, synchronous with the Plenus Cold Event, a period of extensive bottom water reoxidation occurred throughout the Atlantic region. In this interval at the Tazra section (Morocco, Atlantic Ocean), the low oxygen tolerant *Heterohelix* populations dramatically decreased in response to more oxygenated waters throughout the water column, *Guembelitira* thrived and the surviving rotaliporids species rapidly disappeared (Keller et al. 2008). Low Calcite/Detritus ratios in the cooling event indicate increased detrital input, linked to intensified upwelling in the western African margin (Keller et al. 2008).

In fact, this cooling event is an unusual interlude in the long-term trend of oceanic anoxia during the Cenomanian–Turonian interval. At the start of OAE2, higher sea surface temperature together with enhanced runoff and nutrient input resulted in burial of organic matter. Burial of organic matter (OM) lowered pCO_2, cooling the greenhouse climate. At the end of the OAE2, when carbon burial rates were reduced, pCO_2 increased again (van Bentum et al. 2012). In all described sections, faunal turnover suggests the expansion of the oxygen minimum zone before and after the cold event (Fig. 8)

Conclusion

The two Cenomanian–Turonian sections at Gharesu and Taherabad in the Kopet-Dagh basin coincide with the upper part of the Aitamir Formation and lower part of the Abderaz Formation. The C–T boundary in these two sections is conformable based on presence of the three following zones: *Rotalipora cushmani* Total Range zone, *Whiteinella archaeocretacea* Partial Range Zone and *Helvetoglobotruncana helvetica* Total Range Zone. Planktonic foraminiferal assemblage changes show five intervals with different degrees of environmental perturbation associated with the OAE2. In the two sections, the most dramatic changes took place during interval 3. Temporary disappearance of all planktonic foraminifera in some samples, low carbonate content and presence of framboidal pyrite indicate an increased detrital input, surface productivity and an enhanced OMZ in interval 3. High abundance of simple planktonic morphotypes especially heterohelicids shows that low oxygen

content continues into interval 4. Appearance of the new genera *Helvetoglobotruncana* and *Marginotruncana* in the upper part of sections indicates that this part of the sediments was deposited under more normal and stable condition. A decreased run off also coincides with increased carbonate contents and diversity of planktonic foraminifera in interval 2 (at the end of the *R. cushmani* zone) can be interpreted as evidence of a short-term cooling in the studied succession, which is comparable to the Plenus Cold Event in the Tethys realm and benthic oxic zone in the Atlantic Ocean.

References

Abdoshahi M., Vahidinia M., Ashuri A. & Rahimi B. 2010: Micro-biostratigraphy of Cenomanian–Turonian boundary in the Shurab section, east of Kopet-Dagh basin. *Sed. Facies* 1, 61–70 (in Persian).

Afshar-Harb A. 1994: Geology of Kopet Dagh. In: Hushmandzadeh A. (Ed.): Treatise on the Geology of Iran. *Geological Survey of Iran*, Tehran, 1–275 (in Persian).

Arthur M.A. & Premoli-Silva I. 1982: Development of widespread organic carbon rich strata in the Mediterranean Tethys. In: Schlanger S.O. & Cita M.B. (Eds.): Nature and Origin of Cretaceous rich Facies. *Academic Press*, London, 7–54.

Berberian M. & King G.C.P. 1981: Toward a paleogeography and tectonic evolution of Iran. *Canad. J. Earth Sci.* 18, 210–265.

Berner R.A. 1969: The synthesis of framboidal pyrite. *Econ. Geol.* 64, 383–384.

Caron M. & Homewood P. 1983: Evolution of early planktonic foraminifers. *Mar. Micropaleont.* 7, 453–462.

Caron M. 1985: Cretaceous planktonic foraminifera. In: Bolli H.M., Saunders J.B. & Perch-Nielsen K. (Eds.): Plankton Stratigraphy. *Cambridge Univesity*, Cambridge, 17–86.

Caron M., Dall' Agnolo S., Accarie H., Barrera E., Kauffman E.G. Amedro F. & Robaszynski F. 2006: High-resolution stratigraphy of the Cenomanian/Turonian boundary interval at Pueblo (USA) and Wadi Bahloul (Tunisia): stable isotope and bio-events correlation. *Geobios* 39, 171–200.

Coccioni R. & Luciani V. 2005: Planktonic foraminifers across the Bonarelli Event (OAE2, latest Cenomanian): The Italian record. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 224, 167–185.

Frakes L.A., Francis J.E. & Syktus J.I. 1992: Climate modes of the Phanerozoic. *Cambridge University Press*, Cambridge, 1–274.

Forster A., Schouten S., Moriya K., Wilson P. & Sinninghe Damsté J.S. 2007: Tropical warming and intermittent cooling during the Cenomanian/Turonian oceanic anoxic event 2: Seasurface temperature records from the equatorial Atlantic. *Paleoceanography* 22, 1219.

Gale A.S., Jenkyns H.C., Kennedy W.J. & Corfield R.M. 1993: Chemostratigraphy versus biostratigraphy: data from around the Cenomanian–Turonian boundary. *J. Geol. Soc. London*, 150, 29–32.

Gale A.S. & Christensen W.G. 1996: Occurrence of the belemnite Actinocamax plenus in the Cenomanian of SE France and its significance. *Bull. Geol. Soc. Den.* 43, 68–77.

Ghoorchaei Sh., Vahidinia M., Ghasemi-Nejad E. & Mahmudy-Gharaie M.H. 2011: Microbiostratigraphy of Cenomanian–Turonian boundary in the Hamam-Ghale section, east of Kopet-Dagh basin. *Journal of Geology of Iran* 19, 31–44 (in Persian).

Gradstein F.M., Agterberg F.P., Ogg J.G., Hardenbol J., Veen P.V. &

Huang Z. 1995: A Triassic, Jurassic and Cretaceous time scale. *SEPM Spec. Publ.* 54, 95–126.

Grosheny D., Beaudoin B.B., Morel L. & Esmares D.D. 2006: High resolution biostratigraphy and chemostratigraphy of the Cenomanian/Turonian boundary event in the Vocontian Basin, southeast France. *Cretaceous Res.* 27, 629–640.

Hallock P., Premoli Silva I. & Boersma A. 1991: Similarities between planktonic and larger foraminiferal evolutionary trends through Paleogene paleoceanographic changes. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 83, 49–64.

Hancock J.M. & Kauffman E.G. 1979: The great transgressions of the Late Cretaceous. *J. Geol. Soc, London* 136, 175–186.

Haq B.U., Hardenbol J. & Vail P.R. 1987: Chronology of Fluctuating Sea Levels since the Triassic. *Science* 235, 156–167.

Hardenbol J., Thierry J., Farley M.B., Jacquin T., Graciansky P.C. & Vail P.R. 1998: Mesozoic and Cenozoic sequence chronostratigraphic framework of European basins. *SEPM Spec. Publ.* 60, 3–13.

Hart M.B. 1980a: The recognition of mid-Cretaceous sea-level changes by means of foraminifera. *Cretaceous Res.* 1, 289–297.

Hart M.B. 1980b: A water-depth model for the evolution of the planktic Foraminiferida. *Nature* 286, 252–254.

Hasegawa T., Pratt L.M., Maeda H., Shigeta Y., Okamoto T., Kase T. & Uemura K. 2003: Upper Cretaceous stable carbon isotope stratigraphy of terrestrial organic matter from Sakhalin, Russian Far East: A proxy for the isotopic composition of paleoatmospheric CO_2. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 189, 97–115.

Jarvis I., Lignum J.S., Gröcke D.R., Jenkyns H.C. & Pearce M.A. 2011: Black shale deposition, atmospheric CO_2 drawdown, and cooling during the Cenomanian–Turonian Oceanic Anoxic Event. *Paleoceanography* 26, 3201.

Jefferies R.P.S. 1962: The palaeoecology of the Actinocamax plenus subzone (lowest Turonian) in the Anglo–Paris Basin. *Palaeontology* 4, 609–647.

Jones Ch.E. and Jenkyns H.C. 2001: Seawater strontium isotopes, oceanic anoxic events, and seafloor hydrothermal activity in the Jurassic and Cretaceous. *Amer. J. Sci.* 301, 112–149.

Kaiho K., Katabuchi M., Oba M. & Lamolda M. 2014: Repeated anoxia-extinction episodes progressing from slope to shelf during the latest Cenomanian. *Gondwana Res.* 25, 1357–1368.

Keller G. 1988: Extinction, survivorship and evolution of planktic foraminifera across the Cretaceous–Tertiary boundary at El Kef, Tunisia. *Mar. Micropaleont.* 13, 239–263.

Keller G. 2002: Guembelitria dominated late Maastrichtian planktic foraminiferal assemblages mimic early Danian in the Eastern Desert of Egypt. *Mar. Micropaleont.* 47, 71–99.

Keller G. & Pardo A. 2004: Age and paleoenvironment of the Cenomanian–Turonian global stratotype section and point at Pueblo, Colorado. *Mar. Micropaleont.* 51, 95–128.

Keller G., Han Q., Adatte T. & Burns S.J. 2001: Palaeoenvironment of the Cenomanian–Turonian transition at Eastbourne, England. *Cretaceous Res.* 22, 391–422.

Keller G., Adatte T., Stinnesbeck W., Luciani V., Karoui N. & Zaghbib-Turki D. 2002: Paleoecology of the Cretaceous–Tertiary mass extinction in planktic foraminifera. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 178, 257–298.

Keller G., Berner Z., Adatte T. & Stueben D. 2004: Cenomanian–Turonian and delta C^{13}, and delta O^{18}, sea level and salinity variations at Pueblo, Colorado. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 211, 19–43.

Keller G., Tantawy A.A., Berner Z., Adatte T., Chellai E.H. & Stueben D. 2008: Oceanic events and biotic effects of the Cenomanian–Turonian anoxic event, Tarfaya Basin, Morocco.

Cretaceous Res. 29, 976–994.

Lamolda M.A., Gorostidi A. & Paul C.R.C. 1994: Quantitative estimates of calcareous nannofossil changes across the Plenus Marls (latest Cenomanian), Dover, England; implication for the generation of the Cenomanian–Turonian boundary event. *Cretaceous Res.* 15, 143–164.

Leckie R.M., Uretich R.F., West O.L.O., Finkelstein D. & Schmidt M.G. 1998: Paleoceanography of the southwestern Western Interior Sea during the time of the Cenomaniane–Turonian boundary (Late Cretaceous). In: Arthur M.A. & Dean W.E. (Eds.): Stratigraphy and Paleoenvironments of the Cretaceous Western Interior Seaway. *SEPM Concepts in Sedimentology and Paleontology* 101–126.

Leckie R.M., Bralower T.J. & Cashman R. 2002: Oceanic anoxic events and plankton evolution: biotic response to tectonic forcing during the mid-Cretaceous. *Paleoceanography* 17, 13.1–13.29.

Li L. & Keller G. 1998: Diversification and extinction in Campanian–Maastrichtian planktic foraminifera of northwestern Tunisia. *Eclogae Geologicae Helvetiae* 91, 75–102.

Luciani V. & Cobianchi M. 1999: The Bonarelli level and other black shales in the Cenomanian–Turonian of the northeastern Dolomites (Italy): calcareous nannofossil and foraminiferal data. *Cretaceaous Res.* 20, 135–167.

Luderer F. & Kuhnt W. 1997: A high resolution record of the Rotalipora extinction in laminated organic-carbon rich limestone of the Tarfaya Atlantic Coastal Basin (Morocco). *Annales de Société Géologique du Nord* 5, 199–205.

Miller K.G., Kominz M.A., Browning J.V., Wright J.D., Mountain G.S., Katz M.E., Sugarman P.J., Cramer B.S., Christie-Blick N. & Pekar S.F. 2005: The Phanerozoic record of global sea-level change. *Science* 310, 1293–1298.

Mort H.P., Adatte T., Keller G., Bartels D., Follmi K.B., Steinmann P., Berner Z. & Chellai E.H. 2008: Carbon deposition and phosphorus accumulation during Oceanic Anoxic Event 2 in Tarfaya, Morocco. *Cretaceous Res.* 29, 1008–1023.

Nebrigic D.D. 2006: Cenomanian–Turonian planktic foraminiferal bioevents during the time of global oceanic anoxic event 2 and maximum sea level rise, Western Interior Basin and Texas Gulf Coast. *Ph.D. Thesis, University of Texas at Dallas*, 1–465.

Paul C.R.C., Lamolda M.A., Mitchell S.F., Vaziri M.R., Gorostidi A. & Marshall J.D. 1999: The Cenomanian–Turonian Boundary at Eastbourne (Sussex, UK): a proposed european reference section. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 150, 83–121.

Pedersen T.F. & Calvert S.E. 1990: Anoxia vs. productivity: what controls the formation of organic-carbon rich sediments and sedimentary rocks? *American Association of Petroleum Geologists Bulletin* 74, 454–466.

Philip J. & Floquet M. 2000: Late Cenomanian. In: Dercourt J., Gaetani M. Vrielynck B., Barrier E., Biju-Duval B., Brunet M.F., Cadet J.P., Crasquin S. & Sandulescu M. (Eds.): Atlas Peri-Tethys, Palaeogeographical maps. *CCGM/CGMW,* Paris, 129–136.

Premoli Silva I. & Verga D. 2004: Practical manual of Cretaceous planktonic foraminifera. In: Verga D. & Rettori R. (Eds.): International School on Planktonic Foraminifera. *University of Perugia and Milan,* 283.

Poulsen C.J., Barron E.J., Arthur M.A. & Peterson W.H. 2001: Response of the mid-Cretaceous global ocean circulation to tectonic and CO_2 forcings. *Paleoceanography* 16, 576–592.

Robaszynski F., Caron M., Amédro F., Dupuis C., Hardenbol J., Gonzalez-Donoso J.M., Linares D. & Gartner S. 1993: Le Cénomanien de la region de Kalaat Senan (Tunisie Centrale): Litho-biostratigraphie et interpretation séquentielle. *Revue de Paléobiologie* 12, 351–505.

Robaszynski F. & Caron M. 1995: Foraminifères planctoniques du Crétacé: commentaire de la zonation Europe-Méditerranée. *Bulletin de la Société Géologique de France* 166, 681–692.

Sadeghi A. & Forughi F. 2004: The effect of Sub-Hercynian phase movements in the east of Kopet-Dagh basin (east and northeast of Mashhad). *Journal of Earth Science Facility of Shahid Beheshti University* 10, 53–68 (in Persian).

Soua M., Echihi O., Herkat M., Zaghbib-Turki D., Smaoui J., Fakhfakh-Ben Jemia H. & Belghaji H. 2009: Structural context of the paleogeography of the Cenomanian–Turonian anoxic event in the eastern Atlas basins of the Maghreb. *CR Geosci.* 341, 1029–1037.

Soua M., Zaghbib-Turki D., Ben Jemia H., Smaoui J. & Boukadi A. 2011: Geochimical record of the Cenomanian–Turonian anoxic event in Tunisia: Is it correlative and isochronous to the biotic signal? *Acta Geol. Sin.* 85, 1310–1335.

Schallreuter R. 1982: Framboidal pyrite in deep-sea sediments. *DSDP* 75, 875–891.

Schlanger S.O., Jenkyns H.C. 1976: Cretaceous oceanic anoxic events: Causes and consequences. *Geologie en Mijnbouw* 55, 179–184.

Stocklin J. 1968: Structural history and tectonics of Iran: a review. *Bull. Amer. Assoc. Petrol. Geol.* 52, 58–129.

Tarduno J.A., Brinkman D.B., Renne P.R., Cottrell R.D., Scher H. & Castillo P. 1998: Evidence for extreme climatic warmth from the Late Cretaceous arctic vertebrates. *Science* 282, 2241–2244.

Uličný D., Hladíková J. & Hradecká L. 1993: Record of sea-level changes, oxygen depletion and the $\delta^{13}C$ anomaly across the Cenomanian–Turonian boundary, Bohemian Cretaceous Basin. *Cretaceous Res.* 14, 211–234.

Vahidinia M., Ariai A. & Sobhani E. 1999: A new approach to Abderaz Formation in the east of Kopet-Dagh basin. *Science Journal of Islamic Azad University* 31, 2325–2385 (in Persian).

van Bentum E.C., Reichart G.J., Forster A. & Sinninghe Damsté J.S. 2012: Latitudinal differences in the amplitude of the OAE-2 carbon isotopic excursion: pCO_2 and paleoproductivity. *Biogeosciences* 9, 717–731.

Voigt S., Gale A.S. & Flogel S. 2004: Midlatitude shelf sea in the Cenomanian-Turonian greenhouse world: Temperature evolution and North Atlantic circulation, *Paleoceanography* 19, 4020.

Wonders A.A. 1980: Middle and late Cretaceous planktonic Foraminifera of the western Mediterranean area. *Utrecht Micropaleontology Bulletin* 24, 1–158.

Zaghbib-Turki D. & Soua M. 2013: High resolution biostratigraphy of the Cenomanian–Turonian interval (OAE2) based on planktonic foraminiferal bioevents in North-Central Tunisia. *J. African Earth Sci.* 78, 97–108.