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Biozonation, Paleobathymetry and paleoenvironmental study of the Gurpi Formation in southwestern Iran

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Abstract

Planktonic foraminifera and major palynomorph groups (i.e. terrestrial, marine, and amorphous organic matter) of the Campanian-Selandian strata of the Gurpi Formation were studied to evaluate the changes in the Zagros paleobasin in southwestern Iran. Planktonic foraminifera were categorized into four morphotypes according to their paleodepth, extending from Morphotype 1, comprising surface generalists, to Morphotype 4, comprising deep-water aphotic heterotrophic species. The relative amount of species belonging to Morphotype 4 is noticeable in the Campanian-Maastrichtian. It shows that the depth of the basin was generally greater than 200 m in the Late Cretaceous. Two major sea-level regressional phases could be recognized during the formation deposition. The first one occurred near the middle of the Maastrichtian in the *Contusotruncana contusa* Biozone, during which the relative abundance of terrestrial palynomorphs increased prominently, and the species belonging to Morphotype 4 became notably rare. This regression correlates well with the eustatic curves, suggesting that it was controlled by global factors. Also, the species belonging to Morphotype 4 disappeared in the Danian, the relative abundance of terrestrial palynomorphs increased, and the relative abundance of planktonic foraminifera (P%) decreased prominently. However, this regression does not correlate with the eustatic curves and indicates that local tectonic events should be considered. The palynofacies study of the strata shows that during the Campanian-Maastrichtian, the strata were deposited in a deep suboxic-anoxic setting (Palynofacies IX), whereas the Danian strata were deposited in a shallow-shelf setting (Palynofacies VI). The studies show that during the Selandian, the depth of the basin increased again.

Keywords: *Planktonic Foraminifera, Palynomorph, Palynofacies, Tethys, Paleobasin*

1. Introduction

The Zagros Mountains, also called the Zagros fold-thrust belt, extends from the Ararat Mountains in Turkey to Iranian Fars geological province in southwestern Iran. The formation of the Zagros Basin (Neotethys Ocean) is the result of rifting and separation of Central Iran Block from northern Gondwana (Agard et al. 2005). In fact, there were three major events affecting the Neotethys Ocean basin and the creation of its rock units; subduction of the Neotethys oceanic plate beneath Iranian plates during the Early to Late Cretaceous, obduction of the Neotethys ophiolites over the Arabian plate in the Late Cretaceous and the collision of the Arabian plate with the Iranian plate in the Late Cretaceous and, finally, creation of the Zagros Mountains (Alavi 2004), which is an Alpine-Himalayan orogeny type. However, the vanishing of the Neotethys Ocean basin did not occur until the middle Miocene (Ajirlu et al. 2016). Due to the tectonic events and the continental collision during the late Cretaceous-middle Miocene, the Zagros trough was a foreland basin developing in a NW-SE trend parallel with the collision line, causing sea-level fluctuations in different times and localities within the Neotethys Ocean basin. Even though different localities are studied by different authors, there is not a large-scale reconstruction of the environmental change during the Late Cretaceous-Paleogene. The strata of the Gurpi Formation, extending throughout the whole Zagros mountain range, are an excellent rock unit to analyze to help elucidate the most important changes in the sedimentary basin properties.

The foraminiferal and palynomorph content of the Campanian-Selandian strata of the Gurpi Formation have been widely studied by many authors (e.g. Vaziri Moghaddam 2002; Esmaeilbeig 2012; Fereydounpour et al. 2014; Esfandyari-Bayat and Rameh 2016). Considering all of these reports can lead to a large-scale reconstruction of the Neotethys Ocean basin for this period of time. The aim of this study is to understand and illustrate the changes in the Zagros paleobasin during the late Campanian-Selandian period, using planktonic foraminifera as depth-indicating groups and palynomorph groups as environmental proxies. This study shows that microfossils are useful tools for paleobasin reconstruction, when coupled with tectonic data.

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2. Geological setting

The Gurpi Formation outcrops well in many areas of the Zagros Mountains, from Fars to Lurestan geological provinces. In the studied area, the Cretaceous part of the formation is composed of marly limestone, while in Paleocene-aged units, mudstone intervals with clastic origin are dominant due to tectonic events and the erosion of the continental lithosphere (Alavi 2004).

The studied section of the Gurpi Formation is located at Chenareh Mountain in the north of Khuzestan geological province, Iran (Fig 1). The formation lies on massive limestone of the Sarvak Formation with an unconformity and is, in turn, overlain by purple shales of the Pabdeh Formation with a gradual boundary (Fig 2). The total thickness of the studied formation at this section from the base of marly limestones to the beginning of the purple shale of the Pabdeh Formation is measured to be 125 m.

3. Materials and methods

Fifty-six rock samples were collected from marly limestones and mudstones from the study section. A total of 65 thin sections were prepared to study foraminiferal content. Moreover, 190 palynological slides were prepared from the 56 rock samples, using HCl and HF acids to dissolve the rock, and extract particulate organic matter (Traverse 2007). Due to the poor preservation of organic matter in carbonate rocks, good and consistent data could not be gathered from Maastrichtian strata. Both thin sections and palynological slides were studied using an optical microscope with the maximum magnification of 1000X, following Riding and Head (2018). The sharpness and contrast of photos taken from samples were minimally enhanced (Appendix 1).

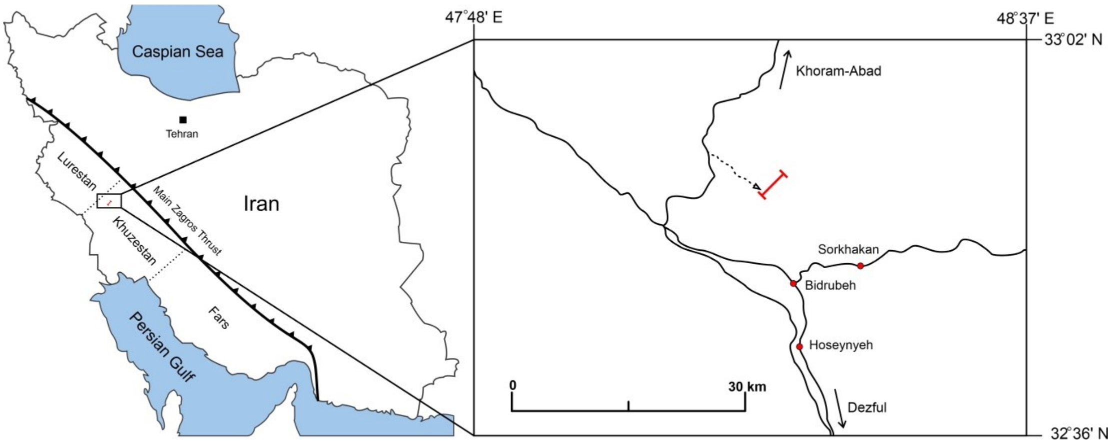


Fig 1. Location of the studied section at Chenareh Mountain in Khuzestan geological province, southwestern Iran. The dashed line shows the trail to the study section.

4. Biostratigraphy

To establish a biostratigraphic framework and to date the formation precisely, the foraminiferal contents of the thin sections were studied. However, the main purpose of foraminiferal biozonation is to facilitate the explanation of the changes in paleobathymetry and paleoenvironment during selected period of time. Thin sections revealed a relatively rich assemblage of well-preserved planktonic foraminifera, based on which the following biozones were differentiated consecutively (Fig 2, Appendix 1).

The *Globotruncana aegyptiaca* Interval Zone is the first biozone and occupies the base of the formation up to the first appearance datum (FAD) of *Gansserina gansseri*. Even though the zonal nominal species was not recorded at this interval, the accompanying species including *Globotruncana havanensis*, *Globotruncana linneiana*, *Globotruncana hilli*, *Globotruncana bulloides*, *Macroglobigerinelloides ultramicrus* and *Macroglobigerinelloides alvarezi* indicate that the interval belongs to the *G. aegyptiaca* zone. Due to the coverage of the base of the formation, it was not possible to sample and identify the lowermost part of this biozone. This biozone shows an age of late to latest Campanian (Premoli Silva and Verga 2004).

The *Gansserina gansseri* Interval Zone overlies the *Globotruncana aegyptiaca* Interval Zone beginning with the FAD of the nominal species and ending with the FAD of *Contusotruncana contusa*. Accompanying species are *Globotruncana linneiana*, *Globotruncana falsostuarti*, *Globotruncana arca*, *Macroglobigerinelloides alvarezi* and *Rugoglobigerina rugosa*. This biozone shows an age of latest Campanian to early Maastrichtian (Premoli Silva and Verga 2004).

The *Contusotruncana contusa* Interval Zone is the last biozone distinguishable in Upper Cretaceous strata. This zone begins with the FAD of the nominal species to the FAD of *Abathomphalus mayaroensis*. Accompanying species are *Globotruncana conica*, *Contusotruncana patelliformis*, *Macroglobigerinelloides subcarinatus*, *Macroglobigerinelloides alvarezi* and *Rugoglobigerina rugosa*.

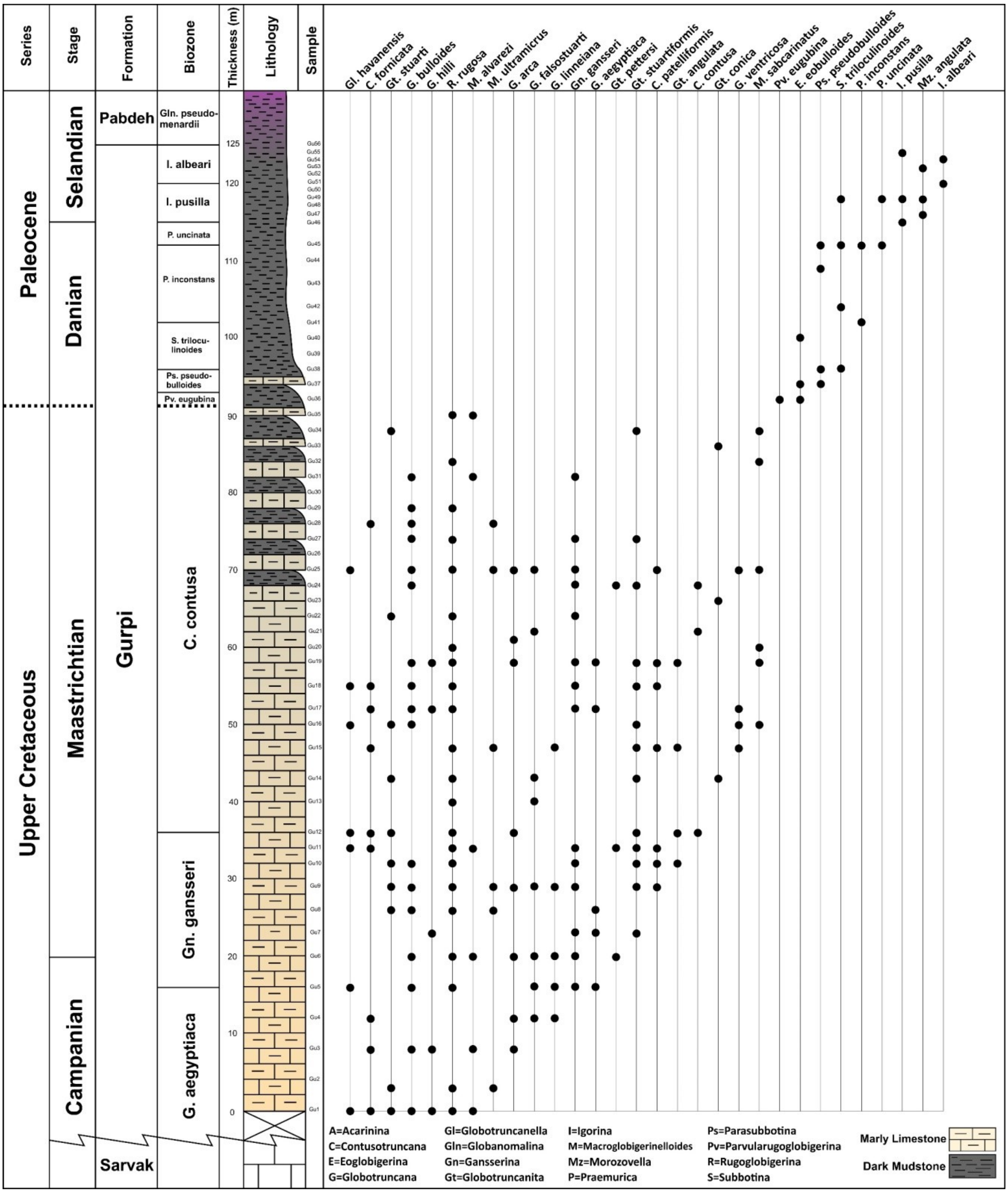


Fig 2. Lithostratigraphy, range chart and biozonation of the important foraminifera species recorded.

This biozone shows an age of early to late Maastrichtian (Premoli Silva and Verga 2004).

The *Parvularugoglobigerina eugubina* Total Range Zone overlies the *Contusotruncana contusa* Interval Zone and indicates the beginning of the Paleocene strata. This biozone begins from the FAD to the last appearance datum (LAD) of the nominal species indicating an age of earliest Danian (Berggren and Pearson 2005). Therefore, a hiatus and sediment starvation from late Maastrichtian to earliest Danian is recorded. Even though this hiatus cannot be identified lithologically, it encompasses the *Abathomphalus mayaroensis* and *Guembelitria cretacea* Biozones (about three million years). Accompanying species are *Eoglobigerina eobulloides* and *Muricohedbergella monmouthensis*. *Muricohedbergella monmouthensis* is the only species, which survived the mass extinction in the studied area.

The *Parasubbotina pseudobulloides* Partial Range Zone starts with the LAD of *Pv. Eugubina* to the FAD of *Subbotina triloculinoides*. This biozone indicates an age of middle Danian (Berggren and Pearson 2005). As a result of the mass extinction at the Cretaceous-Paleogene boundary, the species diversity of planktonic foraminifera is very low in the early and middle Danian. For this reason, only *Eoglobigerina eobulloides* could be observe with the nominal species.

The *Subbotina triloculinoides* Interval Zone starts with the FAD of the nominal species and ends with the FAD of *Praemurica inconstanse*. It is indicative of the early to middle Danian (Berggren and Pearson 2005). In this biozone, the species diversity is still very low and accompanying species are only *E. edita*, *Ps. Pseudobulloides* and biserial species (i.e. *Woodringina* sp.).

The *Praemurica inconstans* Interval Zone occupies the interval between the FAD of the nominal species and the FAD of *Praemurica uncinata* and is indicative of the middle to late Danian (Berggren and Pearson 2005). Accompanying species are *Ps. Pseudobulloides*, *S. triloculinoides* and *Eoglobigerina edita*.

The *Praemurica uncinata* Interval Zone starts with the FAD of the nominal species and ends with the FAD of *Igorina pusilla*. Accompanying species are *S. cancellata*, *Ps. Pseudobulloides* and *S. triloculinoides*. This biozone indicates an age of late Danian (Berggren and Pearson 2005).

The *Igorina pusilla* Interval Zone and the *Igorina albeari* Interval Zone overlie Danian strata respectively. The *Igorina pusilla* Interval Zone starts with the FAD of the nominal species to the FAD of *Igorina albeari*. The *Igorina albeari* Interval Zone starts with the FAD of the nominal species to the FAD of *Globanomalina pseudomenardii*. These two biozones indicate an age of early to middle Selandian (Berggren and Pearson 2005). In the early Selandian, accompanying species are *Morozovella angulata*, *Acarinina strabocella* and, rarely, *Globanomalina compressa*, while accompanying species in the *Igorina albeari* Interval Zone are *Igorina pusilla*, *Morozovella angulata*, *Morozovella velascoensis* and *Acarinina strabocella*. In the middle Selandian, Morozovella species became abundant and by the beginning of the Pabdeh Formation these species became abundant.

According to the identified biozones, an age of late Campanian–middle Selandian is defined for the formation at this section. As neither index species *A. mayaroensis* and *G. cretacea* were recorded in thin sections nor in the washed samples, the zones that they represent are missing. Therefore, a gap encompassing about three million years in the upper Maastrichtian strata is confirmed within the Gurpi Formation at this section. However, more sampling and detailed studies may be needed for more clarification. A sampling method with shorter interval and finer resolution may reveal a sharp boundary between the Cretaceous and Paleogene at this location, but such a detailed study requires different approaches, which are beyond the goals of this research.

A hiatus at the end of the Cretaceous or the beginning of the Paleocene has been reported by many authors on different sections of the formation and other formations of the same age deposited in different basins of the Neotethys (e.g. Ghasemi-Nejad et al. 2006; Vaziri Moghadam et al. 2006; Zarei and Ghasemi-Nejad 2006; Hemati-Nasab et al. 2008; El-Ayyat and Obaidallah 2013; Farouk et al. 2014). However, a recent detailed biostratigraphy of the Gurpi Formation conducted in Khuzestan geological province shows that a consistent biozonation at the Cretaceous-Paleocene boundary may be available at some localities (Beiranvand et al. 2014). The correlation of the biozonation from this study with some other papers is illustrated at Fig 3.

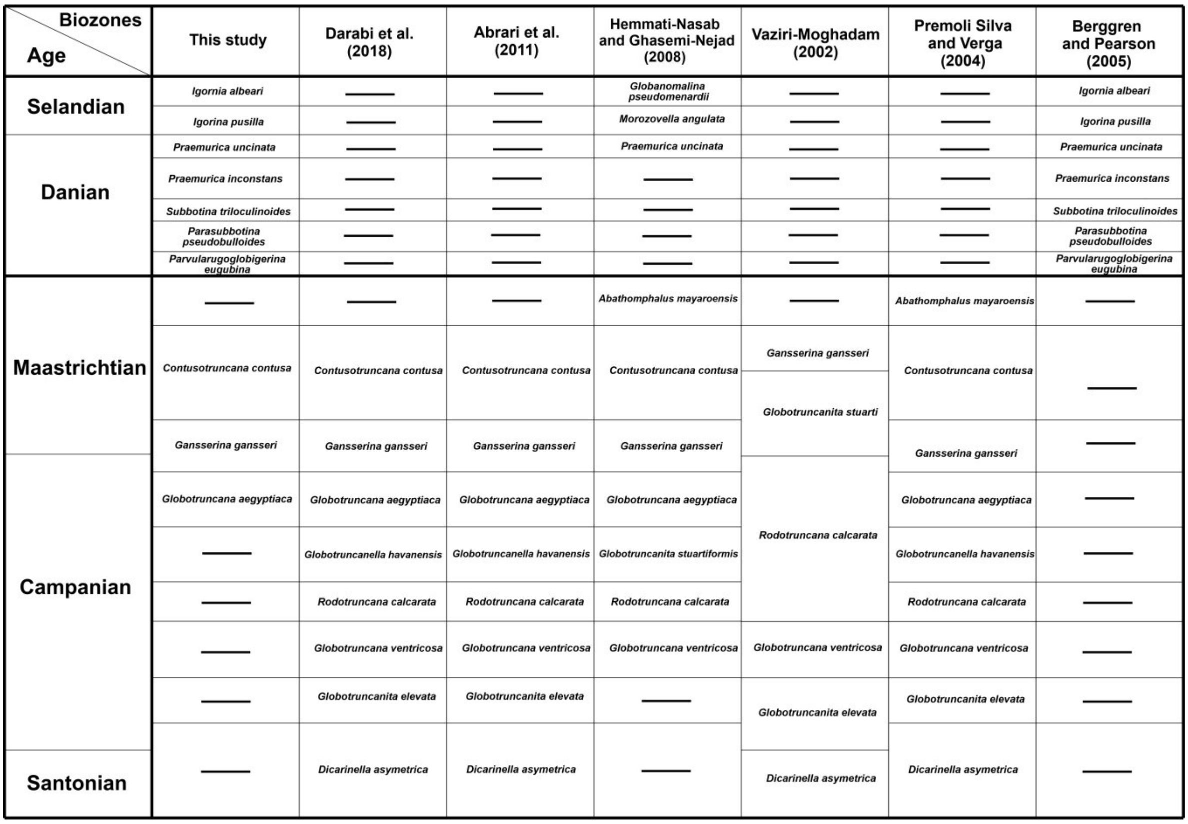
5. Paleobathymetry and paleoenvironmental changes

5.1. Paleobathymetry

One of the useful proxies for evaluating paleobathymetry and sea-level changes is the relative abundance of planktonic foraminifera (P%). The planktonic/benthic ratio has been widely used by researchers as a proxy to determine the proximity to the shoreline and changes in paleobathymetry (e.g. Orabi and Zahran 2014; Orabi and Hassan 2015; Darabi et al. 2018). Moreover, some authors managed to present advanced formulas for using P% to determine paleobathymetry, studying the distribution pattern of both benthic and planktonic foraminifera (Hallam 1967; Berger 1971; Van der Zwaan et al. 1990; Wilson 2003). Murray (1976) studied Persian Gulf and Celtic Sea foraminifers and realized that in the upper part of the slope, the planktonic/benthic ratio is more than 70%, while in shallower areas, it is less. In addition, Olsson and Nyang (1984) argued that the slope is characterized by a high abundance of P% (more than 70%).

The population of planktonic and benthic foraminifera were studied within the prepared thin sections, and the P% was calculated (P% = planktonic / planktonic + benthic × 100). The data show that the fluctuations in the relative abundance of planktonic foraminifera (P%) throughout the formation are minimal and always above 70%, indicating a deep shelf to slope sedimentary basin (Appendix 2). However, the mean change of P% shows that there are two noticeable decreases in P%; the first occurring near the base of the *Gansserina gansseri* Biozone and the second in Danian strata (Fig 4a).

In addition to the relative abundance of planktonic foraminifera, the relative abundance of their morphotypes is another useful proxy to study paleobathymetry and evaluate changes in sea level.



Biozones / Age	This study	Darabi et al. (2018)	Abrari et al. (2011)	Hemmati-Nasab and Ghasemi-Nejad (2008)	Vaziri-Moghadam (2002)	Premoli Silva and Verga (2004)	Berggren and Pearson (2005)
Selandian	*Igornia albeari*	—	—	*Globanomalina pseudomenardii*	—	—	*Igornia albeari*
	Igorina pusilla	—	—	*Morozovella angulata*	—	—	*Igorina pusilla*
Danian	*Praemurica uncinata*	—	—	*Praemurica uncinata*	—	—	*Praemurica uncinata*
	Praemurica inconstans	—	—	—	—	—	*Praemurica inconstans*
	Subbotina triloculinoides	—	—	—	—	—	*Subbotina triloculinoides*
	Parasubbotina pseudobulloides	—	—	—	—	—	*Parasubbotina pseudobulloides*
	Parvularugoglobigerina eugubina	—	—	—	—	—	*Parvularugoglobigerina eugubina*
Maastrichtian	—	—	—	*Abathomphalus mayaroensis*	—	*Abathomphalus mayaroensis*	—
	Contusotruncana contusa	*Contusotruncana contusa*	*Contusotruncana contusa*	*Contusotruncana contusa*	*Gansserina gansseri*	*Contusotruncana contusa*	—
					Globotruncanita stuarti		
	Gansserina gansseri	*Gansserina gansseri*	*Gansserina gansseri*	*Gansserina gansseri*		*Gansserina gansseri*	—
Campanian	*Globotruncana aegyptiaca*	*Globotruncana aegyptiaca*	*Globotruncana aegyptiaca*	*Globotruncana aegyptiaca*	*Rodotruncana calcarata*	*Globotruncana aegyptiaca*	—
	—	*Globotruncanella havanensis*	*Globotruncanella havanensis*	*Globotruncanita stuartiformis*		*Globotruncanella havanensis*	—
	—	*Rodotruncana calcarata*	*Rodotruncana calcarata*	*Rodotruncana calcarata*		*Rodotruncana calcarata*	—
	—	*Globotruncana ventricosa*	*Globotruncana ventricosa*	*Globotruncana ventricosa*	*Globotruncana ventricosa*	*Globotruncana ventricosa*	—
	—	*Globotruncanita elevata*	*Globotruncanita elevata*	—	*Globotruncanita elevata*	*Globotruncanita elevata*	—
Santonian	—	*Dicarinella asymetrica*	*Dicarinella asymetrica*	—	*Dicarinella asymetrica*	*Dicarinella asymetrica*	—

Fig 3. The correlation of biozonation from this study with some other papers.

Studying foraminifera according to taxonomic data lies within the principle that species with similar morphological specification live in similar environments and these specifications do not change in time (Sliter and Baker 1972; Lipps 1979). Most of the recent foraminiferal genera developed from the Late Cretaceous to the early Paleogene and most of the recent species developed in the middle Miocene. Thus, studying foraminifera regarding their similarities, and more specifically paleobathymetry and paleoecology, is only possible for genera developed since the Cretaceous system (Gibson 1988). However, several authors believe that some genera developed in the Jurassic and extend the possibility to use them in Jurassic deposits as well (Sliter and Baker 1972; Lipps 1966, 1979). In addition, these researchers used depth index genera (based on similarities to modern genera) to find a pattern in species diversity and distribution in continental shelf and slope during the Cretaceous. In addition to the authors mentioned above, many other scientists studied planktonic foraminifera regarding their paleoenvironment and used them as a sea-level proxy (e.g. Douglas 1979; Boersma and Premoli Silva 1991; Keller 1996; Nederbragt et al. 1998; Keller et al. 2002; Schmidt et al. 2004; Punekar et al. 2016; Keller et al. 2016; Mateo et al. 2017; Gold et al. 2017) all concluding the possibility to classify all genera from the Late Cretaceous to the Paleogene into four main categories regarding their living depth.

All planktonic foraminifera deposited in the study section were counted and categorized and the relative percentages of the different morphotypes were calculated and graphed considering the pycnocline zone to be at a 200 m depth (Fig 4b). These morphological groups from shallow to deep areas are as follows (Fig 5):

The first group (Morphotype 1) is comprised of generalist and opportunist foraminifera and widely show depths of approximately 50 m, which include A. biserial and triserial genera, such as *Heterohelix* and *Guembelitria*, and B. planspiral and trochospiral genera, such as *Macroglobigerinelloides* and *Hedbergella*. The second group (Morphotype 2) is comprised of relatively specialist genera, which generally occupy depths from 50 to 100 m below water surface. These genera are divided into two subcategories: A. trochospiral genera with relatively well developed processes such as *Rugoglobigerina* and *Acarinina* and B. trochospiral genera with compressed chambers or initial keel such as *Praemurica* and *Morozovella*. The third group (Morphotype 3) is comprised of more developed and more specialized genera.

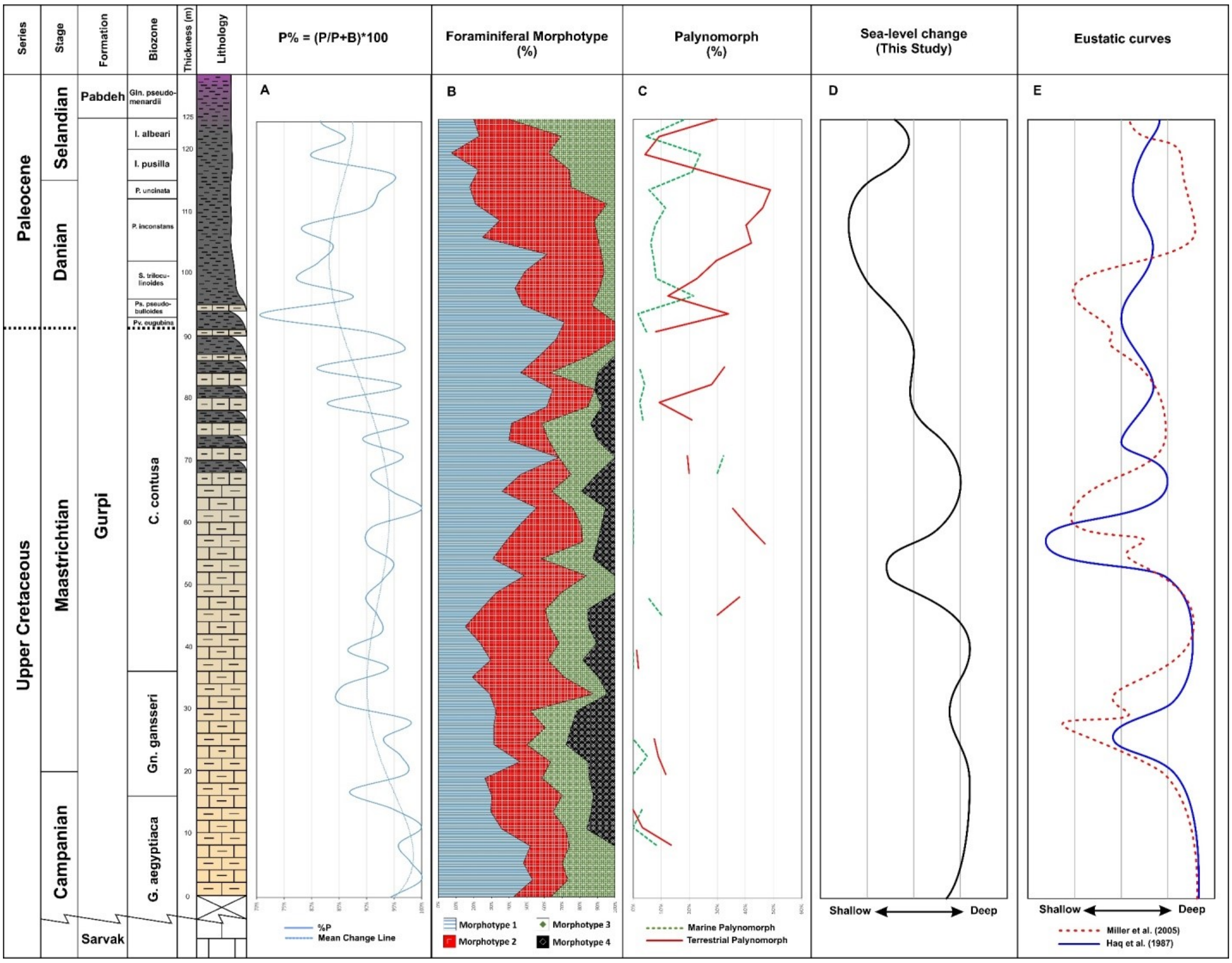


Fig 4. Proxies used for the estimation of sea-level changes in the study section. a. Planktonic foraminiferal percentage. Note that the strongest decrease in mean value occurs in the Danian. b. Relative abundance of foraminiferal morphotypes. c. Changes in the relative amount of terrestrial and marine palynomorphs. Particulate organic matter is not well-preserved in limestone, causing gaps in the chart. d. Estimated sea-level curve for the Gurpi Formation according to charts A, B and C. e. Eustatic curves according to Haq et al. (1987) and Miller et al. (2005).

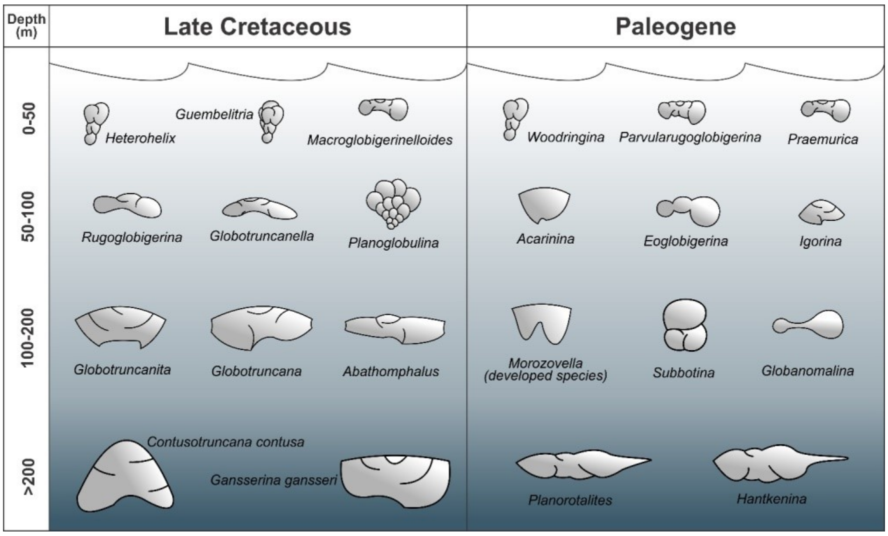


Fig 5. Depth distribution of some planktonic foraminifera during the Late Cretaceous-Paleogene, modified after BouDagher-Fadel (2015). For detailed information refer to the text.

They are mostly indicators of depths from approximately 100 to 200 m (in upper bathyal) above the aphotic zone. These genera are relatively large and are divided into three subcategories: A. genera with one or two keels such as *Globotruncana* and developed species of *Morozovella* with relatively bigger tests compared to Morphotype 2, B. biconvex genera with mild convexity such as *Globotruncanita* and *Marginotruncana* and C. genera with relatively big cancellate tests in the Paleogene such as *Subbotina*. The fourth group (Morphotype 4) involves particular species of Morphotype 3 groups, which are large forms, as well as the ones with more convexity. The species of this group are believed to be able to live in the environments with the least amount of light. Thus, this group represents depths of approximately 200 m and below the photic zone of the ocean (middle to lower bathyal), which include A. planoconvex species, such as *Gansserina gansseri* and *Dicarinella concavata*, B. species with high convexity in the dorsal side, such as *Globotruncanita conica* and *Contusotruncana contusa* and C. long and flat or semi-flat genera, such as *Globanomalina planocompressa* and *Planorotalites*.

The data gathered from planktonic morphotypes throughout the formation show that two major regressions occurred in sea level. The first occurred near the base of the *Contusotruncana contusa* Biozone, when species belonging to Morphotype 4 disappeared for a short period of time, suggesting that sea level fell greatly and, after a short period, rose again. This regression could not be identified in lithological units. The second great regression occurred in the Danian. At this time, Morphotypes 1 and 2 were dominant, suggesting a great shallowing trend, which lasted until the beginning of the Selandian.

The second proposed regression (in the Danian) cannot be determined with great confidence due to the evolutionary trends of planktonic foraminifera after the Cretaceous-Paleogene boundary extinction. Even though the species diversity of planktonic foraminifera recovered after approximately 3 million years (Coxall et al. 2006), deep water forms did not evolve until the late Danian (BouDagher-Fadel 2015). However, even in the late Danian, deep forms such as *Globanomalina planocompressa* or *G. compressa* are rare. These species lived from the middle Danian to Selandian and their rarity can indicate a shallow basin. Due to the constraints in the results gathered from Danian strata, palynomorph data were also considered in order to understand the changes in the basin at this time (Section 5.3).

5.2. Changes in palynomorph assemblages in response to sea level changes

Beside foraminiferal data, amounts and relative portions of palynological groups can be used to determine sea-level changes and the paleobathymetry. Compared to proximal environments, where the relative abundance of phytoclasts and spores are higher, in distal environments, the only terrestrial palynomorphs able to migrate long distances are saccate pollen (i.e. bisaccate pollen). However, the absolute abundance of pollen grains is still very low in deep environments (e.g. Habib 1982; Tyson 1984; Traverse 2007). Higher amounts of wood debris and other phytoclasts indicate a higher proximity to the shoreline. They are mostly transported to the sea by rivers and seasonal streams. The higher amounts of terrestrial palynomorphs show the effect of deltaic systems and shallow shelf or turbidity currents (Tyson 1995).

Relative proportions of palynomorphs calculated for the Gurpi Formation in the study area show three relatively acme values. The first rise in the amount of terrestrial palynomorphs is recorded in the middle Maastrichtian in the *Contusatruncana contusa* Biozone and the second in the Danian from the *Parasubbotina pseudobulloides* Biozone to the *Praemurica uncinata* Biozone, which coincides with the second decrease in mean value of planktonic foraminifera (P%). The third rise in terrestrial palynomorphs occured in the middle Selandian in the *Igorina pusilla* Biozone.

Fluctuation in the relative abundance of terrestrial palynomorphs near the Cretaceous-Paleogene boundary is believed to be due to turbidity currents (Fig 4c, Appendix 3). These fluctuations correlate well with P% fluctuations at the same time (Fig 4a, Appendix 2) and show a shallowing trend in sea level, but cannot be used to properly track detailed sea-level changes.

5.3. Palynofacies and paleoenvironmental changes

According to Tyson (1993), palynofacies analysis is a proven tool to determine changes in depositional environment, as well as hydrocarbon evaluations. The use of geochemical methods based on organic matter has led to the concept of palynofacies and its broad use in hydrocarbon evaluations and basin analyses (e.g. Cross 1964; Rodgers 1979; Tyson 1984; Whitaker 1984; Hart 1986; Jones 1987; Van der Zwaan 1990; Ghasemi-Nejad et al. 2009; Schiøler et al. 2010; Juliao-Lemus et al. 2016; Adams et al. 2017; El Atfy et al. 2017; Garnit et al. 2017).

Three major palynomorph groups can be retrieved from the rocks and studied in palynological slides: terrestrial palynomorphs (i.e. phytoclasts and sporomorphs), marine palynomorphs (i.e. dinoflagellate cysts and acritarchs) and amorphous organic matter (AOM). Despite the fact that the relative amount of palynomorph groups can be used to track sea-level changes (Section 5.3), they can be used to understand some of the properties of the marine sedimentary basins as well. One of the proven methods to study the paleoenvironment of marine basins is the ternary diagram of Tyson (1993), using the three major palynomorph groups.

Plotting the relative contents of mentioned palynomorph groups recovered from the Gurpi Formation on this diagram revealed that the dominant palynofacies of the formation strata are Palynofacies IX and VI. The Upper Campanian and Maastrichtian strata dominantly show Palynofacies IX. This palynofacies is associated with a distal suboxic-anoxic basin, where sediment starvation is common, amorphous organic matter is dominant, and the only terrestrial palynomorphs available are bisaccate pollen grains (very low absolute amounts of pollen grains). The sediments of this facies are believed to be highly oil prone and could be potential hydrocarbon source rock (Tyson 1993). On the other hand, the Danian strata show Palynofacies VI. This facies is indicative of a shallow suboxic-anoxic shelf, where phytoclasts and AOM are dominant (Tyson 1993). In the Selandian, Palynofacies VIII and IX are dominant indicating a dysoxic-anoxic distal shelf. These changes in the palynofacies of strata show that the basin underwent a shallowing trend in the Danian (Fig 6). A plate of some of some of the important dinocysts can be found in Appendix 4.

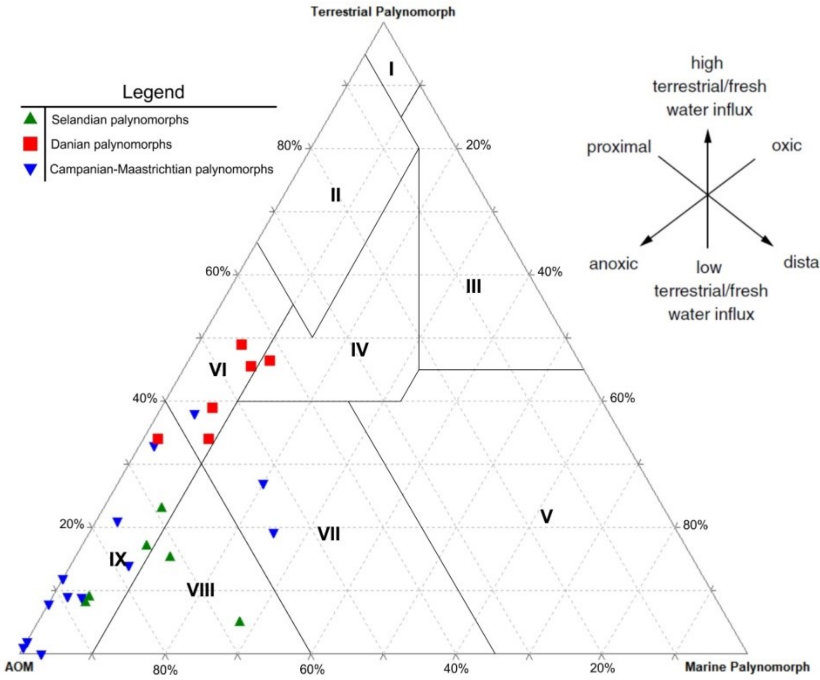


Fig 6. Palynofacies of the Gurpi Formation according to the three major palynomorph groups. The data plotted show that the late Campanian-Maastrichtian strata deposited in a deep suboxic-anoxic basin (Palynofacies IX). The palynofacies for Danian strata is VI showing a shallow shelf, while it changes to VIII and IX expressing a deep dysoxic-anoxic shelf.

6. Discussion

Two thirds of planktonic foraminifera species went extinct during the Cretaceous-Paleogene extinction event, and only generalist species managed to survive this mass extinction (Keller et al. 2002). The evolutionary trend of foraminifera shows that during mass extinctions, bigger species will vanish and smaller ones (generalists) will survive or evolve (Gould 1988). Studying the size trend in the studied section shows that after the Cretaceous-Paleogene boundary, foraminiferal species decreased. The size of the species before the boundary varied between 100 µm and 500 µm, while after the extinction until the end of the Danian, when species diversity recovered, all species were smaller than 200 µm.

When studying species richness of foraminiferal assemblages, it is also worth mentioning that the changes in relative abundance of foraminifera is not solely related to the number of living forms present at the time of sedimentation, but also to the productivity of living foraminifera (e.g. Murray 1976; Diester-Haass 1978; Lipps 1979; Lutze 1980; Berger and Diester-Haass 1988). Thus, the depth of water is not the only controlling factor in the composition of assemblages. However, the study of paleoproductivity is beyond the goals of this study and requires different approaches.

Due to the reasons mentioned above, studying foraminiferal morphotypes in times of extinction can be difficult. High abundance of Morphotype 1 near the Cretaceous-Paleogene boundary should not be interpreted as a falling sea level, because at this time, mainly surface generalists survived (Morphotype 1). In addition, the extinction of Globotruncanidae at the end of the Cretaceous (Morphotypes 3 and 4) can also lead to paleobathymetry misunderstandings. To minimize the errors in paleobathymetric evaluations via planktonic foraminifera, the relative amount of terrestrial and marine palynomorphs were taken into account. Two noticeable sea-level changes can be seen when studying different paleobathymetric proxies, showing that the depth of the sedimentary basin experienced two great shallowing regressions. At the middle of the *Contusatruncana contusa* Biozone, near the middle Maastrichtian, when the species belonging to Morphotype 4 disappeared, this indicates that the sea level fell for a short period of time. During this time, the mean P% value slightly decreased and the relative abundance of terrestrial palynomorphs increased prominently, suggesting a seaward transgression of the shoreline. This great shallowing also correlates well with the eustatic curves presented by Haq et al. (1987) and Miller et al. (2005), suggesting affection by global factors and not local tectonic events. In addition, the palynofacies of the Upper Cretaceous strata show that during the Campanian and Maastrichtian, the sedimentary basin of the Gurpi Formation was a deep suboxic-anoxic basin (Palynofacies IX, Fig 6) confirming the depth of deposition being greater than 200 m (Fig 7a).

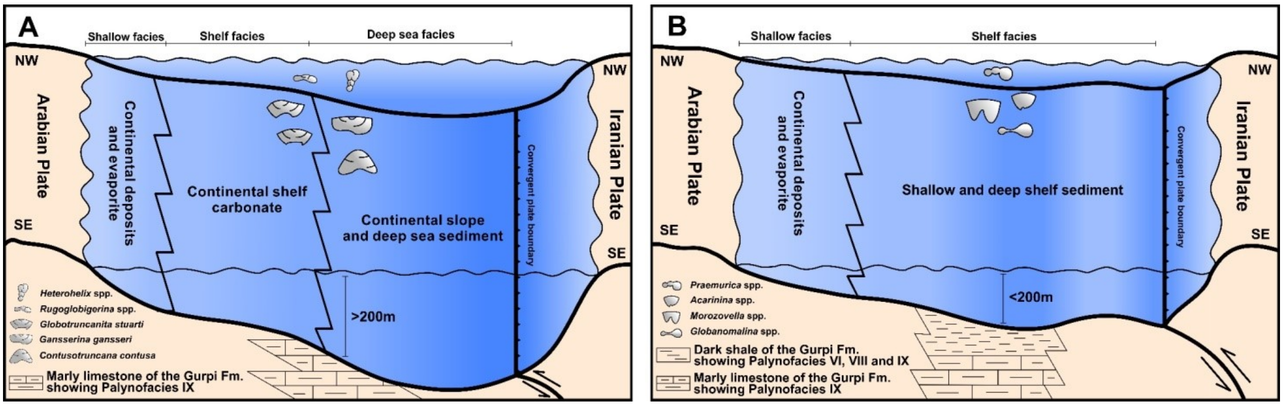


Fig 7. Schematic paleobasins for the Neotethyan Zagros in which the Gurpi Formation deposited. a. During the Campanian-Maastrichtian, paloebathymetry and palynofacies data show that the depth of the deposition was more than 200 m in a suboxic-anoxic environment (Palynofacies IX). b. The basin transformed into a shallow shelf with a relatively low oxygen content during the Danian (Palynofacies VI) as a result of continental collision of Iranian and Arabian plates. In the Selandian, the depth increased again due to lithosphere folding, but it was not greater than 200 m (Palynofacies VIII and IX).

The second noticeable sea-level regression is recorded in the Danian. During this time, the mean P% value decreased and the relative amount of terrestrial palynomorphs increased prominently, indicating an extended sea-level regression until the end of the *Praemurica uncinata* Biozone. During this time, deep-water forms became very rare. This decrease in the depth of the basin does not correlate with the eustatic curves (Fig 4e). Thus, it is believed that this change in the depth of the basin is not the result of global factors controlling the sea level, but the result of local tectonic events (Fig 7b). The palynofacies data show that during the Danian, the sedimentary basin of the Gurpi Formation experienced the first outcome of continental closure transforming the sedimentary basin into a shallow suboxic-anoxic shelf (Palynofacies VI). During this time, the lithosphere thickened and folded in a NW-SE direction forming a foreland basin (Alavi 2004).

The first species of *Morozovella* and *Acarinina* appear in the Selandian. These species rely on photo symbiosis of their algal partner and, thus, cannot live below 200 m in the aphotic zone (Kelly et al. 1996). The Selandian strata fall within the Palynofacies VIII and IX showing a relatively deep dysoxic-anoxic sedimentary basin (Fig 6). However, deep genera such as *Globanomalina* are rare, suggesting that even though the depth of the basin increased, it was not more than 200 m (Fig 4d).

The results of this study correlate well with reports on other sections of the formation in different geological provinces (e.g. Hemmati-Nasab and Ghasemi-Nejad 2008; Abrari et al. 2011; Beiranvand et al. 2013; Darabi et al. 2018). However, it seems that the effects of continental closure were stronger in Khuzestan geological province during the Danian. The data gathered from the Chenareh section suggests that the basin transformation was more conspicuous compared to other parts of the basin during this time.

7. Conclusions

Studying paleobathymetric and paleoenvironmental proxies from the strata of the Gurpi Formation reveals that the Zagros paleobasin was a deep suboxic-anoxic basin during the Campanian and Maastrichtian stages. The facies and lithological studies show that the formation extended from deep continental shelf to lower parts of continental slope, where the depth was more than 200 m. A great sea-level regression occurred near the middle of the Maastrichtian for a short period of time. This regression cannot be identified in lithological units and correlates well the eustatic curves. The basin subsequently transformed into a suboxic-anoxic shallow shelf in the Danian. Both paleobathymetric proxies and palynofacies data show that during this time, due to the continental collision of the Iranian and Arabian plates, the lithosphere folded and thickened, causing an extended shallowing trend. This transformation in the sedimentary basin was more conspicuous in Khuzestan geological province. During the Selandian, the depth of the basin increased again indicating deep dysoxic-anoxic shelf facies. Even though the depth increased again, paleobathymetric data show that it was not greater than 200 m.

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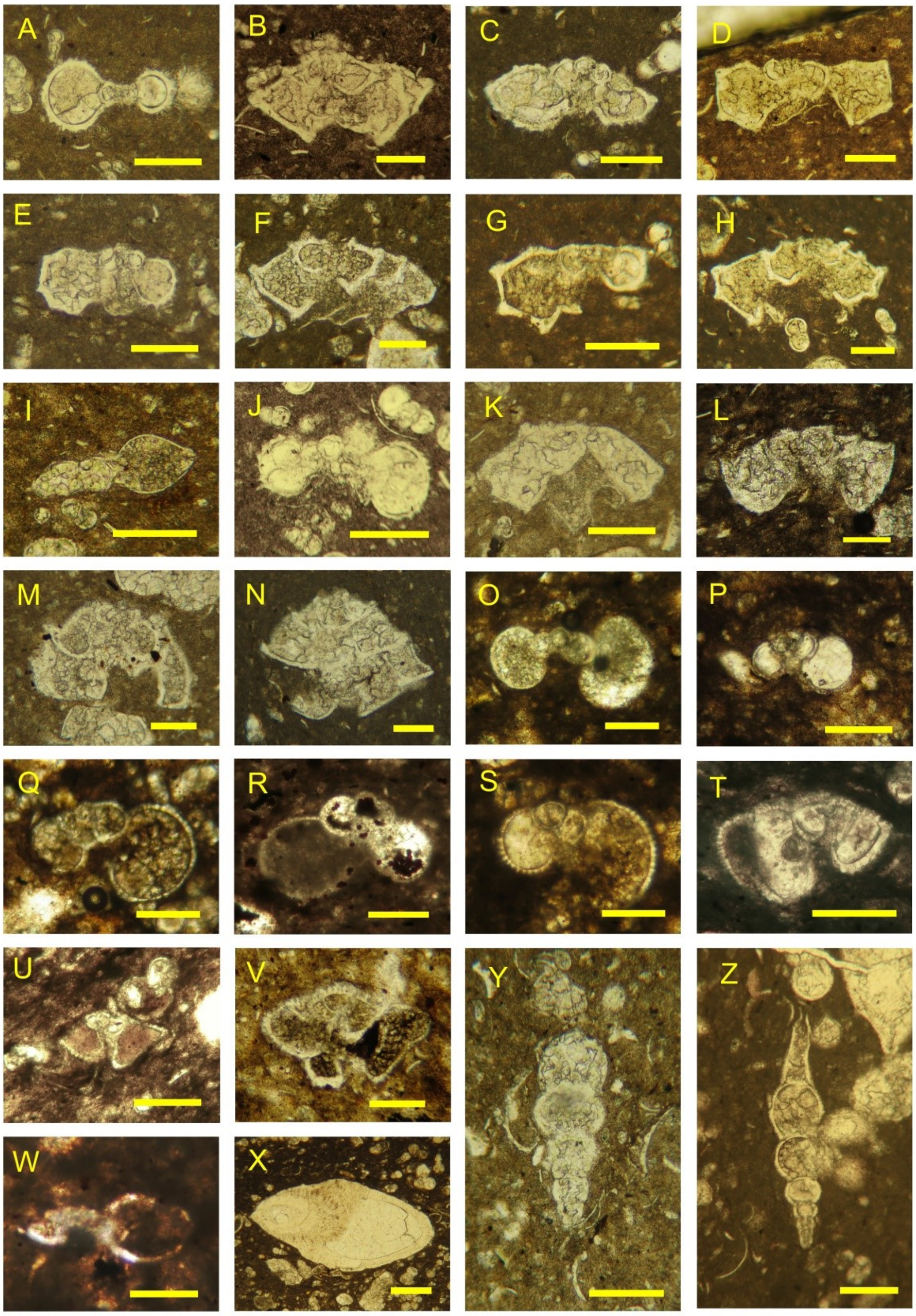
Supplementary Material



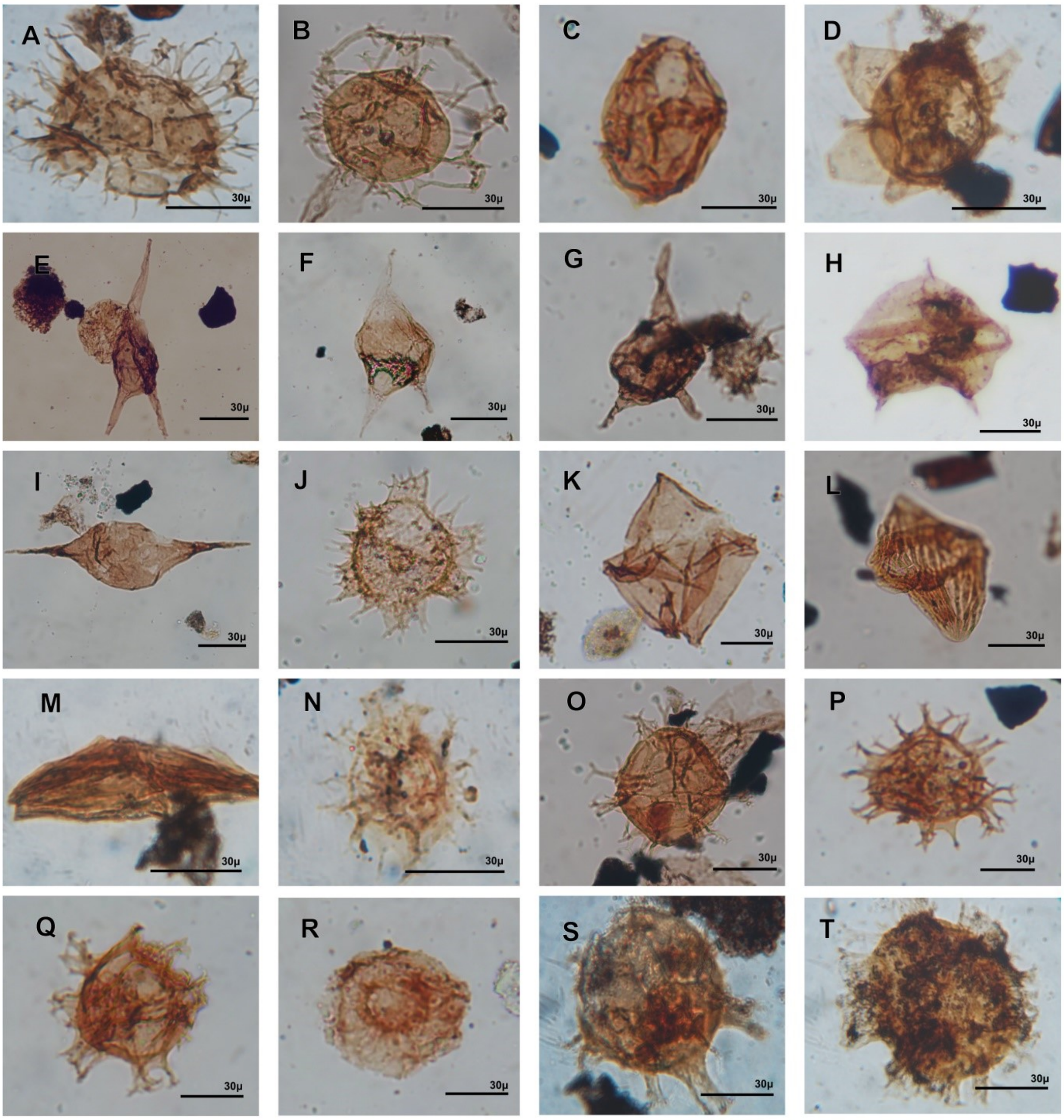
Appendix 1. The plate of biostratigraphically important foraminifera. A. *Macroglobigerinelloides alvarezi* (Eternod-Olvera 1959) B. *Globotruncanita stuarti* (de Lapparent 1918) C. *Globotruncanella havanensis* (Voorwijk 1937) D. *Globotruncana linneiana* (d'Orbigny 1839) E. *Globotruncana bulloides* (Vogler 1941) F. *Globotruncana arca* (Cushman 1926) G. *Globotruncana hilli* (Pessagno 1967) H. *Globotruncana lapparenti* (Brotzen 1936) I. *Macroglobigerinelloides subcarinatus* (Broennimann 1952) J. *Rugoglobigerina rugosa* (Plummer 1926) K. *Contusotruncana fornicata* (Plummer 1931) L. *Gansserina gansseri* (Bolli 1951) M. *Contusotruncana contusa* (Cushman 1926) N. *Globotruncanita conica* (White 1928) O. *Muricohedbergella monmoutensis* (Olsson 1960) P. *Parvularugoglobigerina eugubina* (Luterbacher and Premoli-Silva, 1964) Q. *Parasubbotina pseudobulloides* (Plummer 1926) R. *Eoglobigerina eobulloides* (Morozova 1959) S. *Subbotina triloculinoides* (Plummer 1926) T. *Acarinina strabocella* (Loeblich 1957) U. *Igorina pusilla* (Bolli 1957) V. *Morozovella angulata* (White 1928) W. *Globanomalina compressa* (Plummer 1926) X. *Lenticulina* sp. Y. *Heterohelix* sp. Z. *Gublerina cf. cuvillieri* (Kikoine 1948). The scale bar represents 100 µm.

Appendix 2. The table of relative abundance of particulate organic matter.

Sample	Percentage of Morphotype 1	Percentage of Morphotype 2	Percentage of Morphotype 3	Percentage of Morphotype 4	P%
Gu1	42%	21%	36%	0%	94%
Gu2	53%	20%	27%	0%	100%
Gu3	48%	22%	30%	0%	98%
Gu4	52%	22%	26%	0%	96%
Gu5	36%	36%	12%	16%	100%
Gu6	30%	35%	22%	14%	95%
Gu7	30%	40%	18%	13%	87%
Gu8	26%	32%	26%	15%	97%
Gu9	46%	17%	21%	15%	96%
Gu10	31%	19%	22%	28%	93%
Gu11	31%	29%	15%	25%	98%
Gu12	33%	20%	26%	22%	85%
Gu13	29%	58%	9%	4%	86%
Gu14	19%	52%	19%	10%	94%
Gu15	30%	32%	20%	18%	87%
Gu16	24%	45%	21%	11%	93%
Gu17	15%	48%	22%	15%	92%
Gu18	24%	37%	24%	15%	92%
Gu19	33%	37%	30%	0%	93%
Gu20	49%	35%	16%	0%	95%
Gu21	31%	27%	29%	13%	90%
Gu22	38%	44%	8%	10%	91%
Gu23	46%	35%	12%	8%	100%
Gu24	56%	20%	19%	6%	95%
Gu25	36%	28%	17%	19%	91%
Gu27	47%	28%	15%	10%	97%
Gu28	68%	0%	32%	0%	89%
Gu29	40%	23%	28%	10%	98%
Gu31	41%	17%	28%	14%	83%
Gu32	62%	23%	8%	8%	96%
Gu33	65%	24%	0%	12%	81%
Gu34	47%	17%	27%	10%	97%
Gu35	57%	29%	14%	0%	91%
Gu36	67%	33%	0%	0%	71%
Gu37	71%	29%	0%	0%	88%
Gu38	46%	38%	16%	0%	77%
Gu39	44%	47%	9%	0%	81%
Gu41	46%	42%	12%	0%	84%
Gu43	45%	22%	33%	0%	78%
Gu46	19%	48%	33%	0%	90%
Gu47	31%	48%	21%	0%	92%
Gu48	17%	59%	24%	0%	95%
Gu50	29%	67%	5%	0%	80%
Gu51	23%	52%	25%	0%	86%
Gu53	8%	55%	37%	0%	82%
Gu54	23%	46%	31%	0%	87%
Gu56	11%	22%	67%	0%	87%

Appendix 3. The table of P% and the relative abundance of foraminiferal morphotypes.

Sample	AOM%	Marine Palynomorph%	Terrestrial Palynomorph%
Gu2	78%	8%	14%
Gu3	97%	0%	3%
Gu4	97%	3%	0%
Gu6	88%	0%	12%
Gu7	86%	5%	9%
Gu8	93%	0%	8%
Gu12	98%	0%	2%
Gu13	99%	0%	1%
Gu15	60%	10%	30%
Gu16	57%	5%	38%
Gu19	53%	0%	47%
Gu20	58%	0%	42%
Gu21	65%	0%	35%
Gu25	50%	30%	20%
Gu26	48%	32%	19%
Gu30	76%	3%	21%
Gu31	88%	2%	10%
Gu32	68%	4%	28%
Gu33	65%	2%	33%
Gu36	87%	5%	8%
Gu38	64%	2%	34%
Gu39	66%	22%	12%
Gu41	69%	8%	22%
Gu42	63%	7%	30%
Gu43	51%	6%	42%
Gu44	52%	8%	40%
Gu46	42%	11%	46%
Gu48	45%	6%	49%
Gu50	53%	21%	26%
Gu51	72%	24%	4%
Gu54	86%	5%	9%
Gu56	52%	19%	30%



Appendix 4. A. *Areoligera senonensis* (Lejeune-Carpentier 1938) B. *Cannosphaeropsis utinensis* (O. Wetzel 1933) C. *Carpodinium sp.* D. *Hystrichokolpoma bulbosum* (Ehrenberg 1838) E. *Cerodinium diebelli* (Alberti 1959) F and G: *Cerodinium sp.* H. *Deflandrea cf. phosphoritica* (Eisenack 1938) I. *Palaeocystodinium bulliform* (Ioannides 1986) J. *Xenascus cf. ceratoides* (Deflandre 1937) K. *Phelodinium magnificum* (Stanley 1965) L and M: *Dinogymnium spp.* N-Q. *Spiniferites spp.* R. *Thalassiphora sp.* S. *Florenitinia sp.* T. *Damassidinium californicum* (Drugg 1967)