Biostratigraphic analysis across the Eocene–Oligocene boundary in the southern Hellenic Thrace basin (Lemnos Island, north Aegean Sea)

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Received: 28.03.2017  •  Accepted/Published Online: 28.12.2017  •  Final Version: 17.05.2018

Abstract: The time interval of the sedimentation in the southern Hellenic Thrace Basin (Lemnos Island, northeast Aegean Sea) was studied, based on the qualitative analysis of calcareous nannoplankton and quantitative analysis of planktonic foraminifera. The samples were collected from 4 different stratigraphic sections (Ifestia, Panagia, Kaspaka, and Kaminia). The biostratigraphic analyses on both calcareous nannoplankton and planktonic foraminifera reveal an age of Late Eocene–Early Oligocene (Priabonian–Rupelian). In particular, the Eocene–Oligocene boundary is recorded at the lower part of the turbiditic system, in the Ifestia section. The present study offers the first taxonomic analysis of planktonic foraminifera of this age in the Hellenic southern Thrace Basin (Lemnos Island). The species Catapsydrax unicavus, Chilognemella cubensis, Dentoglobigerina sp., Dipsidripella liqianyui, Globigerina officinalis, Paragloborotalia nana, Turborotalia cerroazulensis, and Turborotalia sp. were identified. Furthermore, D. liqianyui adds new data worldwide, as it is identified in the Mediterranean Sea (Lemnos Island) for the time interval of the O1–O4 zones (Early Oligocene).

Key words: Oligocene, planktonic foraminifera, D. liqianyui, calcareous nannofossils, Lemnos Island, eastern Mediterranean Sea

1. Introduction
The Eocene–Oligocene (E–O) transition corresponds to a critical interval in Earth history since it is associated with very significant climatic changes in the Cenozoic that caused the last major extinction of marine fauna (e.g., Miller et al., 1991; Prothero, 1994; Lear et al., 2000; Coxall and Pearson, 2007; Maravelis and Zelilidis, 2012). The climatic conditions on Earth changed from the warm Early Eocene conditions to the glaciated Early Oligocene conditions associated with the rapid expansion of continental-scale ice sheets on Antarctica (Zachos et al., 2001; Pagani et al., 2005; Pälike et al., 2006).

Calcareous nannofossils are abundant in deep-sea sediments and are considered to be very sensitive to fluctuations in sea-surface temperature and nutrient concentration (e.g., McIntyre and Bé, 1967; Winter and Siesser, 1994). They have been utilized in paleoecological and paleoceanographical studies (e.g., Haq and Lohmann, 1976; Winter et al., 1994; Wade and Bown, 2006; Villa et al., 2008). In addition, planktonic foraminifera are regarded as a valuable tool for both biostratigraphic correlations (e.g., Casford et al., 2007; Budillon et al., 2009; Triantaphyllou et al., 2009; Siani et al., 2010) and paleoenvironmental reconstructions (e.g., Geraga et al., 2010; Kontakiotis, 2016; Kontakiotis et al., 2016; Drinia et al., 2016).

Thick successions of near-continuous clastic sedimentation are well preserved on Lemnos Island; they are a new opportunity to reconstruct the Eocene–Oligocene depositional history in the southern Hellenic Thrace Basin (HTB). The recent study of Maravelis and Zelilidis (2012) documented an Upper Eocene–Lower Oligocene succession. This succession is composed of basin-floor fan deposits that underlie slope and shallow-marine sediments, and gives the opportunity for a high-resolution biostratigraphic analysis across the Eocene–Oligocene Boundary (EOB). This research presents for the first time an integrated study of calcareous nannofossils and planktonic foraminifera in the Hellenic Thrace Basin (Lemnos Island), and aims to refine the biostratigraphic dating of this succession.

2. Geological setting
The Aegean region corresponds to a first-class example of a retreating plate-boundary zone (Ring and Layer, 2003).
It is subdivided into the Rhodope and Serbo-Macedonian massif (RSM), Circum-Rhodope Belt (CRB), Pelagonian zone, Vardar–İzmir–Ankara oceans, and Pindos ocean (Figure 1A) (Robertson et al., 1991). The northern HTB (Hellenic Rhodope region) overlies the metamorphic rocks of the RSM (Bonev et al., 2006). The basement is not exposed in the southern HTB (Lemnos Island), but the provenance analysis, including conglomerate clast composition and geochemical data, suggests that the sub-Pelagonian zone is one of the most likely candidates (Maravelis et al., 2015, 2016a). The HTB was formed as an elongate trough between the Rhodopian magmatic arc to the north and the accretory prism (Pindic Cordillera in Greece and Biga Peninsula in Turkey) to the south (Tranos, 2009; Maravelis and Zelilidis, 2010, 2013). This prism is considered to represent the westward continuation of the Intra-Pontide or the more southern İzmir–Ankara oceans, which are related to the closure of the Neo-Tethyan oceanic basin (Okay et al., 2008). The HTB is deformed by the North Anatolian Fault (NAF), which postdates the development of the HTB (active since Late Miocene–Early Pliocene times) (Dewey and Şengör, 1979; Armijo et al., 1999; Tranos, 2009).

The Hellenic subduction zone in the Aegean has controlled magmatism in the Balkan Peninsula since the Late Cretaceous (Fytikas et al., 1984; Marchev et al., 2004; Bonev and Beccaleto, 2007). The Upper Cretaceous magmatism in southern Bulgaria has been ascribed to a subduction mechanism, with the Rhodope massif being considered the frontal part of the Srednogorie zone (Marchev et al., 2004). A southward magmatic migration from 92 to 78 Ma is evidenced by U–Pb zircon dating (Peytcheva et al., 1998). The occurrence of the Srednogorie zone into the southern Rhodope region is documented by the 70–42 Ma granitoid intrusions. Lower Miocene volcanic rocks are common in the northern Aegean Sea, indicating further southward migration (Innocenti et al., 1994; Yilmaz et al., 2001; Maravelis et al., 2007). An extended subduction event has been invoked to explain the continuous regional magmatism (Marchev et al., 2004). The magmatic activity started in Rhodope during the Late Eocene, and its migration suggests that the present-day northward subduction in the Aegean region began during the Eocene (Fytikas et al., 1984; Spakman et al., 1988).

At the subduction zone, the accretion of oceanic crust and sediment to the upper plate caused the increase of the accretionary prism (Maravelis and Zelilidis, 2010, 2013; Görür and Elbek, 2013). Between the magmatic arc and the accretionary prism, sediment filling the Thrace forearc basin is composed of Eocene–Oligocene clastic strata (Figure 1B) (Maravelis et al., 2007; Maravelis and Zelilidis, 2011).

Figure 1. A) Map depicting the plate tectonic configuration of the eastern Balkan Peninsula and the position of the study area in relation to the main Alpine orogenic elements of the region (Dimitriadis et al., 1998; Maravelis and Zelilidis, 2013; Tranos and Lacombe, 2014). CRB: Circum-Rhodope belt; RMA: Rhodopian magmatic arc; RMMCC: Rhodope massif metamorphic core complex; NSZ: Nestos shear zone. B) Schematic diagram illustrating the depositional setting of the Hellenic Thrace Basin during the Late Eocene to Oligocene (Maravelis et al., 2016a).
Provenance analyses utilizing sandstone petrography, conglomerate clast composition, and bulk-rock geochemistry have revealed 2 sediment sources (Maravelis et al., 2007; Maravelis and Zelilidis, 2010; D’Atri et al., 2012). The northern HTB is influenced by the RSM and the CRB, and fed abundant coarse-grained sediment from the slope up to the deep parts of the forearc basin. This sediment formed extensive submarine fan deposits that crop out along the northern HTB (Maravelis and Zelilidis, 2010, 2013). The sedimentation in the southern HTB was principally controlled by the accretionary prism (Pindic Cordillera) and is dominated by a sand-rich submarine fan system that underlies slopes and shallow-marine deposits (Maravelis et al., 2007). The source area was eroded, leading to the deep incision of the bedrock and enabling a significant amount of coarse-grained material to be deposited in the forearc region (Maravelis and Zelilidis, 2010; Maravelis et al., 2016b).

3. Materials and methods
Biostratigraphic analyses were performed on samples collected from 4 stratigraphic sections (Ifestia, Panagia, Kaspakas and, and Kaminia), based on the sedimentary facies observed in the region (Figure 2). One hundred twenty-seven samples were analyzed for planktonic foraminiferal analysis. For each sample, 150–200 g of dry sediment was disaggregated using hydrogen peroxide (H2O2). Samples were gently ultrasonicated to enhance the disaggregation. Because of the strong cementation, several samples were soaked in acetic acid (80% CH3COOH and 20% H2O) for 2 h maximum, and then washed through a 63-µm sieve. Residues were dried at 50 °C. It is important to note that the lithified nature of the samples made disaggregation very difficult; therefore, the washing process was repeated multiple times for each sample, with some samples requiring up to a week of processing. The dry material was split into aliquots using an Otto

![Figure 2](image.png)

Figure 2. Detailed facies map of the southern HTB (Lemnos) depicting the lateral distribution of the deep and shallow-marine deposits (modified from Maravelis and Zelilidis (2011)). Note the different depositional environments and subenvironments. Black dots correspond to the studied sections. Palaeocurrent analysis is from Maravelis et al. (2007, 2015).
microsplitter. In each sample, at least 300 specimens were selected and identified under a Leica S4E stereoscope at 80× magnification. The systematic taxonomy of planktonic foraminifera was based on the atlas by Pearson et al. (2006); the biostratigraphy was based on the biozonal definitions given by Berggren et al. (1995), Berggren and Pearson (2005), and Wade et al. (2011). Planktonic foraminiferal shells were also investigated using a scanning electron microscope (SEM), type Jeol JSM 6360, in the laboratory of Historical Geology and Paleontology of the Faculty of Geology and Geoenvironment, National and Kapodistrian University of Athens.

For calcareous nanofossil analysis, 161 samples were collected from the sand-rich turbiditic sequence and the overlying slope and shallow-marine deposits. The preparation of these samples was made using standard smear slide techniques, as described in Bown and Young (1998) and Giunta et al. (2003). Smear slides were analyzed with an optical Leica DMLSP microscope at 1250× magnification. Calcareous nanofossils were generally abundant and well preserved. The biostratigraphy was based on the biozonal definitions of Martini (1971) and Agnini et al. (2014).

Chronostratigraphic consideration refers to Luterbacher et al. (2004), Pearson et al. (2006) and Ogget al. (2016).

4. Studied sections

The studied sections (Ifestia, Panagia, Kaspakas, Kaminia) were selected according to the facies analysis by Maravelis et al. (2007) and Maravelis and Zelilidis (2011). Therefore, the Ifestia section consists of a lobe (outer fan) and channel-fill deposits (inner fan), the Panagia section is composed of channel-fill deposits (inner fan), the Kaspakas section constitutes slope sediments, and finally the Kaminia section represents shallow-marine deposits. According to Maravelis et al. (2007), the stratigraphic order from the base is Ifestia, Panagia, Kaspakas, and Kaminia.

4.1. Ifestia section

The lower part is ~50-m thick and is composed of light brown to light green very thin- to very thick-bedded sandstone, with interbedded mudstone. Sandstone displays parallel, cross-, and/or climbing ripple-laminations (Bouma Tb and Tc sandstone), but structureless sandstone beds have been also observed (Bouma Ta). Mudstone is brown and typically lacks internal structures, although occasionally layers include silt laminae. The middle part is ~35-m thick, and consists of very thin- to very thick-bedded sandstone with mudstone. The upper part is about 10-m thick, and is composed of very thin- to very thick-bedded sandstone and conglomerate that is interbedded with mudstone. The color of these intervals is the same as that of the overlying parts. The sandstone beds mostly exhibit parallel-, cross-, and/or climbing ripple-laminations (Bouma Tb and Tc), and the thicker parts of them are usually structureless, although beds with normal grading are also present. Conglomeratic beds are polymictic, disorganized, and rarely exhibit inverse to normal grading. Mudstone beds typically lack internal structure (Figure 3A).

4.2. Panagia section

The lower part of the section is about 10-m thick and consists of very thin- to very thick-bedded sandstone with mudstone. The color of the sandstone beds is light brown to light green; the thinner beds mostly display parallel-, cross-, and/or climbing ripple-laminations (Bouma Tb and Tc), and the thicker ones are usually structureless. Beds with normal grading have also been observed. Mudstone beds are brownish and typically lack internal structure. Mudstone beds with parallel lamination are also present. The upper part of the section is about 20-m thick and is composed of 5 sedimentary cycles. These cycles evolve into very thin- to very thick-bedded sandstone and conglomerate interbedded with mudstone. Conglomeratic beds are polymictic, disorganized, and rarely exhibit inverse to normal grading. The sandstone and mudstone beds have the same characteristics as the lower part of the section (Figure 3B).

4.3. Kaspakas section

The lower part of the section is about 10-m thick and consists of very thin- to very thick-bedded sandstone with mudstone couplets. Mudstone beds are brownish and typically lack internal structure, although parallel-laminated mudstone occurs rarely. Sandstone units are light brown to light green, with the thinner beds mostly displaying parallel-, cross-, and/or climbing ripple-laminations (Bouma Tb and Tc). The thicker parts are usually structureless; beds with normal grading are not uncommon. The upper part of the section is about 40-m thick and is characterized by mudstone with rare thick sandstone interbeds. Sandstone beds are light brown to light green with ripple cross-lamination (Bouma Tc), while mudstones are similar to those in the lower part of the section (Figure 3C).

4.4. Kaminia section

The Kaminia section is about 100-m thick. The lower part consists of thin to thick-bedded sandstone interbedded with mudstone. Sandstone beds are light brown and sharp-based. Most sandstone beds are structureless, although parallel and cross-lamination occasionally occurs. The middle part is composed of thin- medium bedded sandstone and mudstone forming sedimentary cycles with a trend of becoming finer further up. Sandstone units have flat bases including rare trace fossils, while interbedded mudstone is mostly structureless and/or with rare parallel lamination and locally contains a high amount of coalfield debris. The uppermost part of the section is composed of parallel-laminated mudstone, ~10 m in thickness (Figure 3D).
Figure 3. Stratigraphic columns on corresponding photographs from the outcropped studied sections in Lemnos Island. Red dots correspond to the samples’ positions. A) Ifestia section (39°57′37.11″N, 25°18′57.64″E); B) Panagia section (39°59′12.08″N, 25°23′58.15″E); C) Kaspakas section (39°52′49.24″N, 25°4′50.23″E); and D) Kaminia section (39°51′36.34″N, 25°19′48.78″E).
Panagia section

Figure 3. (Continued).
Figure 3. (Continued).
Kaminia section

Figure 3. (Continued.)
5. Discussion
The analysis of both, the planktonic foraminifera and calcareous nannofossils adds new constraints to the age of sedimentation in the Lemnos Island. More specifically, the biostratigraphic analysis has demonstrated that sedimentation occurred during the Late Eocene (Priabonian, 38.0–33.9 Ma, e.g., Ogg et al., 2016) to Early Oligocene (Rupelian 28.1–33.9 Ma, e.g., Ogg et al., 2016).

The Late Eocene (Priabonian; Figure 4) is identified by the presence of the planktonic foraminifera *Globigerinatheka index* (the last occurrence [LO] of the taxon is the marker event of the upper boundary of the E15 biozone), *Turborotalia cerroazulensis* (LO of the taxon is the marker of E16/O1 boundary), *Hantkenina alabamensis* (taxon LO is the upper boundary of the O1 biozone) (e.g., Wade et al., 2011), and calcareous nannofossils *Ismolithus recurvus* (first occurrence [FO] to the top of the CNE18 biozone), *Cribrocentrum reticulatum* (last occurrence to the top of the CNE19 biozone), *Discoaster barbadiensis* (last occurrence in the CNE21 biozone), *Discoaster saipanensis* (last occurrence at the top of the CNE21 biozone) (e.g., Agnini et al., 2014).

The Eocene–Oligocene boundary is mainly characterized by the disappearance of planktonic foraminifera *T. cerroazulensis* and *H. alabamensis* (e.g., Wade et al., 2011), and the appearance of calcareous nannofossils *Ilselithina fusa* and *Clausicoccus subdistichus* (e.g., Agnini et al., 2014). The Early Oligocene is recognized by the presence of calcareous nannofossils *C. subdistichus* (last occurrence in the CNO1 biozone), *Reticulofenestra umbilicus* (last occurrence at the top of the CNO2 biozone), *Cyclicargolithus abisectus* (first occurrence within the CNO3 biozone), *Sphenolithus distentus* (CNO4 biozone), *Sphenolithus predistentus* (last occurrence at the top of the CNO4 biozone), *Sphenolithus ciperoensis* (within CNO4 to CNO5 biozones) (e.g., Agnini et al., 2014). According to the planktonic foraminifera, the *T. ampliapertura* LO confirms the O2–O3 boundary; the *Globigerina angulisuturalis* FO confirms the O3–O4 boundary; the *Chiloguembelina cubensis* LO suggests the Early Oligocene–Late Oligocene boundary (Roupelian–Chattian; O4–O5).

5.1. Ifestia section
The lower part of the Ifestia section (L100–L55; Figure 5A) is characterized by the planktonic foraminifera...
Catapsydrax unicavus, C. cubensis, Dentoglobigerina sp., Globigerina officinalis, Turborotalia sp. (Plates 1 and 2, see Appendix), and calcareous nannofossils D. barbadiensis, Reticulofenestra bisecta, Reticulofenestra floridana, and Reticulofenestra hillae, which are typical in E14–E15 planktonic foraminifera biozones and CNE18–CNE20 nannofossil biozones, respectively. Globigerinatheka semiinvoluta is absent within the planktonic foraminifera of the section, and this prevents recognition of the E14/E15 boundary. The midpart of that section (samples L54–L32) contains C. unicavus, C. cubensis, Dentoglobigerina sp., Dipsidripella liqianyui, G. officinalis, T. cerroazulensis, and Turborotalia sp. among the planktonic foraminifera and Cyclicargolithus floridanus, Ericsonia formosa, R. bisecta, R. floridana, and S. predistentus among the calcareous nannofossils. Considering also the absence of the nannofossil D. barbadiensis (Plate 3), it is possible to refer this part of the section to the CNE21–CNO2 interval biozones, while in combination with the presence of the above planktonic foraminifera, it confirms the E16–O1 time interval (EOB). The presence of planktonic foraminifer T. cerroazulensis on sample L50 only is particularly interesting if we consider it as the LO of the taxon. In fact, this event closely approximates the Eocene–Oligocene boundary (upper boundary of E16, 33.9 Ma). The presence of planktonic foraminifera C. unicavus, C. cubensis, Dentoglobigerina sp., D. liqianyui, G. officinalis, Paragloborotalia nana, Turborotalia sp. (Plates 1 and 2, see Appendix) and calcareous nannofossils C. floridanus, R. bisecta, R. floridana, S. distentus, S. predistentus at the top of the Iفestia section (samples L30–L3) confirms the O2–O4 and CNO3–CNO4 biozones, respectively. There is no biostratigraphic index of planktonic foraminifera that
indicate the O2/O3/O4 boundary. Finally, the sample L1 contains *S. ciperoensis*, which suggests the CNO4 biozone (Early Oligocene).

### 5.2. Panagia section

In the Panagia section (Figure 5B), the combination of planktonic foraminifera *C. unicavus*, *C. cubensis*, *Dentoglobigerina* sp., *D. liqianyui*, *G. officinalis*, *P. nana*, and *Turborotalia* sp. and calcareous nannofossils *S. ciperoensis*, *S. distentus*, and *S. predistentus* (Plate 3) indicates the O4 planktonic foraminifera biozone or CNO4 nannofossil biozone.

### 5.3. Kaspakas section

The Kaspakas section (Figure 5C) is characterized by the total absence of planktonic foraminifera. However, biostratigraphic reference was possible due to the presence of the calcareous nannofossils *S. ciperoensis*, *S. distentus*, and *S. predistentus*, which indicates the CNO4 biozone.

### 5.4. Kaminia section

The Kaminia section (Figure 5D) consists of planktonic foraminifera *C. unicavus*, *C. cubensis*, *Dentoglobigerina* sp., *D. liqianyui*, *G. officinalis*, *P. nana*, and *Turborotalia* sp. and calcareous nannofossils *S. ciperoensis*, *S. distentus*, and *S. predistentus* (Plate 3) indicates the O4 planktonic foraminifera biozone or CNO4 nannofossil biozone.

![Figure 5. (Continued).](image)
Figure 6. Biochronostratigraphic correlation of the studied sections in Lemnos Island.
and \textit{S. predistentus}, which confirms that these sediments were deposited during the O4 and CNO4 biozones of planktonic foraminifera and calcareous nannofossils zonation, respectively.

The fact that \textit{C. cubensis} was identified in all sections confirms that there are no sediments of Chattian age. Therefore, the Rupelian–Chattian (O4–O5) boundary was not detected.

According to the new biostratigraphic data, the Upper Eocene–Lower Oligocene (CNE18 to CNO4 interval zone in terms of nannofossil zonation) sediments represent lobe facies in the Ifestia section. Additionally, the channel-fill facies is observed in both the Ifestia and Panagia sections, but the sedimentation in the Ifestia section (CNO3–CNO4 biozones) is older than in the Panagia section (CNO4 biozone). Finally, the upper sediments (CNO4 biozone) in Lemnos Island are identified in the Kaspakas and Kaminia sections, which constitute the slope and shelf facies, respectively (Figures 6 and 7).
6. Conclusion
The detailed and integrated calcareous nannofossil and planktonic foraminiferal biostratigraphic analysis in the southern HTB (Lemnos Island) indicates the following:

1. Sediments were deposited during the biostratigraphic intervals E14–O4 (planktonic foraminifera zonation of Wade et al. (2011)) and CNE18–CNO4 (calcareous nannofossils zonation of Agnini et al. (2014)). Therefore, the age of the succession is Late Eocene–Early Oligocene (Priabonian–Roupelian). This succession encompasses submarine fan sediments that are sitting below slope and shallow-marine deposits (Figure 7).

2. The Eocene–Oligocene boundary is recognized with high precision within the lower parts of the submarine fan system (lobe facies: outer fan) 33 m from the base of the Ifstia section.

3. The detailed analysis of planktonic foraminifera in Lemnos Island defines a new geographic and biostratigraphic distribution of the planktonic foraminifer Dipsidripella liqianyui. In fact, this species was recovered for the first time in the Mediterranean area. Previously, it was recognized only in the southern Indian Ocean (ODP Site 738: Kerguelen Plateau, and ODP Site 761: Wombat Plateau). Moreover, according to our data, D. liqianyui has a distribution range in the O1–O4 interval zone, which is different from that given in previous literature, in which it is considered to be in the O1 biozone only.

Further investigations need to be conducted on Upper Eocene–Lower Oligocene sediments of this region to better characterize paleoenvironmental, paleoclimatological, and paleoecological changes, as they have been highlighted in other areas by a decrease in temperature, salinity, productivity, and oxygenation in surface waters (e.g., Berggren and Prothero, 1992; Zachos et al., 2001; Dall’Antonia et al., 2003).

Acknowledgments
This work is part of the PhD thesis by Sofia Kostopoulou (first author), who conducted her studies in the Department of Geology and Geo-Environment, National and Kapodistrian University of Athens, Greece. The authors would like to thank Asst Prof MD Dimiza and Prof MV Triantafyllou for their contribution to this research. We would like also to thank Prof WA Berggren and Prof D Olsson, Rutgers University, USA, for their suggestions on the taxonomy and biostratigraphy of planktonic foraminifera in the Late Eocene–Oligocene time interval. Moreover, we would like to acknowledge the journal editor (Prof Dr Emin Çiftçi) and the 3 anonymous reviewers for their helpful comments, which have improved the final version of the manuscript.

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Appendix

The systematic taxonomy of the planktonic foraminifera species identified in the study region was based on the specificities in morphological characteristics and the new data recorded for the Eocene–Oligocene planktonic foraminifera of the eastern Mediterranean Sea. The identification and systematic taxonomy of the species is based on Loeblich and Tappan (1987), the subsequent Loeblich and Tappan (1992), and Pearson et al. (2006).

Order FORAMINIFERIDA Eichwald, 1830
Superfamily GLOBIGERINACEAE Carpenter, Parker and Jones, 1862
Family GLOBIGERINIDAE Carpenter, Parker and Jones, 1862
Genus GLOBIGERINA d’Orbigny, 1826
Globigerina officinalis Subbotina, 1953
Plate 1, Figures 1–4
1953 Globigerina officinalis Subbotina, 105, pl. 11: figs. 1–7, figs. 1 (holotype), 3, 4, 6, 7, fig. 2; Blow and Banner, 1962, 88, pl. 9: figs A–C; Blow, 1969, 78, pl. 1, figs 1–7 (reillustration of Blow and Banner, 1962, pl. 9, figs A–C); Blow, 1979, 798, pl. 1, figs 1–7 (reillustration of Blow and Banner, 1962, pl. 9, figs A–C); Poag and Commeau, 1995, Pl. 7, figs 8–9; Olsson et al., 2006, 114–115, pl. 6.1, figs 6–16.
1962 Globigerina praebulloides leroyi Blow and Banner, 93, pl. 9, figs R–T.
Description. Perforate, spinose, bulloides-type wall structure. Test is low trochospiral with globular chambers; in spiral view, there are 4 chambers, while in umbilical view there are 3½ chambers. The suture lines are straight and slightly depressed, while umbilical area is small and open. The aperture is umbilical, low to high arch bordered by an imperforate rim. Biostratigraphic range. E10 to Oligocene.
Geographic distribution. Low to mid-latitudes (Pearson et al., 2000).
Paleoecological distribution. Living in surface waters (Spezzaferri et al., 2002; Pearson et al., 2006).
Genus CATAPSYDRAX Bolli, 1957
Catapsydrax unicusavis Bolli, 1957
Plate 1, Figures 5–10
1939 ?Globigerina linaperta Finlay var. turigida Finlay, 125; Bronnimann, 1952, 19–21, pl. 3, figs 1–3.
1952 ?Globigerina tarouibaensis Brönnimann, 18, pl. 2, figs 16–18.
1957 Catapsydrax unicusavis Bolli, Loeblich and Tappan, 37, pl. 7, fig. 9A–C, pl. 37, fig. 7A–B; Stainforth et al., 1975, 328, fig. 149–1, 2, 4, fig. 149 5A–C, fig. 149–3A–B; Kennett and Srivinhasan, 1983, 26, pl. 3, figs 1–3; Huber, 1991, 439, fig. 5, fig. 18; Spezzaferri, 1994, 48, pl. 33, fig. 5A–C; Olsson et al., 2006, 75–76, 79, pl. 3.3, figs 1–7.
1962 Globigerina ishayayensis Asano, 55, pl. 21, fig. 9A–C.
1962 Globigerinita unicusavis unicusavis (Bolli); Blow and Banner, 109, pl. 14, figs M–N.
1964 ?Globorotaloides turigida (Finlay); Jenkins, pl. 7, figs 1–10, pl. 8, figs 1–12, 13 (lectotype selected); Fleisher, 1974, 1029, pl. 13, fig. 8.
1979 ?Globigerinita turigida (Finlay); Blow, pl. 179, figs 6–7.
1983 ?Globorotaloides turigidus (Finlay); Krasheninnikov and Basov, 840, pl. 7, figs 7–9.
Description. Alveolated, spinose, rubber/sacculifer-type wall structure. Low trochospiral test, with 4 chambers in the umbilical and spiral side. Chambers are globular, embracing, increasing rapidly in size. The suture lines are straight to slightly curved, moderately depressed on umbilical side. The main diagnostic feature of this species is a small, slightly inflated bulla which extends above the umbilicus and has 1 infralaminar aperture with a continuous thickened imperforate rim. Biostratigraphic range. E2–N6.
Paleoecological distribution. Living in midwater (Sexton et al., 2000) and deep-sea waters (e.g., Poore and Matthews, 1984; Spezzaferri, 1995; Wade et al., 2000, 2007; Spezzaferri et al., 2002).
Genus PARAGLOBOROTALIA Cifelli, 1982
Paragloborotalia nana Bolli, 1957
Plate 1, Figures 11–12
1957b Globorotalia opima nana Bolli, 118, pl. 28, fig. 3A–C; Bolli and Saunders, 1985, 202, fig. 26: 16A–C (holotype reillustrated), fig. 26: 15, 17–20 (paratypes); Toumarkine, 1978, 714, pl. 8, figs 3–4.
1969 Globorotalia (Turborotalia) opima nana Bolli; Blow, 154, pl. 39, fig. 1.
1977 Globorotalia nana Bolli; Krasheninnikov and Pflaumann, 606, pl. 6, figs. 10–11C.
1995 Jenkinsella opima nana (Bolli); Poag and Commeau, 149, pl. 6, figs 21, 22.
2004 Paragloborotalia nana (Bolli); Pearson et al., 36, pl. 1, fig. 21. Olsson et al., 2006: 95–96: pl. 5.8, figs 1–16.
Description. Perforate, coarsely cancellate, rubber/sacculifer-type wall structure. The test is very small and consists of very low trochospiral, globular much inflated chambers. In both sides of the test are 4, occasionally 4½, globular chambers. In the ultimate whorl, chambers increase slowly in size, while the last chamber is almost equal in size with the penultimate. As a result, the test has a square shape, which is also a diagnostic feature of these foraminifera. Sutures are straight and slightly depressed, while the umbilical view is a very small, deep opening, sometimes closed off by surrounding chambers. The aperture is umbilical–extraumbilical, bordered by a low arch with a thickened lip. Biostratigraphic range. The first appearance of Paragloborotalia nana recorded in Late Eocene (E13 biozone; Blow, 1979 and Toumarkine and Luterbacher, 1985). Nevertheless, transitional forms of Paragloborotalia griffinioides (from which P. nana is evolutionarily derived) have been revealed from the Ε7 biozone in Tanzania (Pearson et al., 2004). P. nana identified until Early Miocene.
Geographic distribution. Low to mid-latitudes (Pearson et al., 2000).
Paleoecological distribution. Deep-water dwellers (Sexton et al., 2000).
Family HEDBERGELLIDAE Loeblich & Tappan, 1961
Genus TURBOROTALIA Cushman & Bermúdez, 1949
Turborotalia sp.
Plate 2, Figures 1–4
Description. Smooth, pustulose, sometimes weakly cancellate wall texture. The test is trochospiral, while the chambers are globular and inflated. There are 3 chambers in umbilical view. The aperture is umbilical–extraumbilical, usually including a high arch with bordering lip.
The poor preservation of those foraminifera did not allow their further classification at species level.

Superfamily HETEROHELICOIDEA Cushman, 1927
Family CHILOGUEMBELINIDAE Reiss, 1963
Genus CHILOGUEMBELINA Loeblich and Tappan, 1956
Chiloguembelina cubensis Palmer, 1934
Plate 2, Figures 5–6
1934 Guembelina cubensis Palmer, 74, figs 1–6.
1939 Chiloguembelina garretti Howe, 61, pl. 8, fig. 14.
1955 Chiloguembelina barnardi Ansary, 77, pl. 2, fig. 26.
1957 Chiloguembelina cubensis (Palmer); Beckmann, 89, pl. 21, fig. 21, text–fig. 14 (5–8); Hornibrook, 1990, 368, pl. 1, figs 1–11; Resig 1993, 241, pl. 1, figs 1–8, 10, 13, 17; Huber et al., 2006, 473–474, pl. 16.3, figs 23–24.
1985 Streptochilus cubensis (Palmer); Poore and Gosnell, 5, pl. 2, figs 1–13.
Description. Microperforate, pustular wall texture. The test is biserial, elongate, with subtriangular form. Sutures are depressed, while the aperture is low, including a symmetrical arch with bordering lip at the base of the final chamber, bordering on the attachment point of penultimate chamber. Main diagnostic feature of Chiloguembelina cubensis is the costate vertical wall texture on sutures.
Biostratigraphic range. Middle Eocene (E11)–Early Oligocene (O1) (Huber et al., 2006). The present study has redefined the biostratigraphic range: Middle Eocene (E11)–Early Oligocene (O4).
Geographic distribution. According to Huber et al. (2006), Dipsidripella liqianyui has been identified in the southern Indian Ocean at ODP Site 738 (Kerguelen Plateau) and Site 761 (Wombat Plateau). This study has certified the presence of D. liqianyui in the Mediterranean Sea (Lemnos Island), adding new data to the geographic distribution of the species on a global basis.
Comments. According to Huber et al. (2006), the small number of D. liqianyui that have been studied and described worldwide and the poor preservation status of these specimens make the systematic taxonomy of this species uncertain. In order to correct the systematic classification of the species, it is necessary for better preserved specimens to be studied, and for more information to be acquired about the microstructure wall of tests and the paleoecological preferences of the species.

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