

1 The Eocene-Oligocene Transition in the Paratethys: Boreal Water 2 Ingression and its Paleoceanographic Implications

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15 **Abstract.** The Eocene-Oligocene Transition (EOT) represents a pivotal period in Earth's climatic history, marked by the onset
16 of Antarctic glaciation and global cooling. While deep-sea records have extensively documented this transition, its impacts on
17 marginal and epicontinental seas remain less understood. This study investigates the impacts of the EOT in the Karaburun
18 composite section, located in the eastern Paratethys. Using a multidisciplinary approach that integrates biostratigraphy,
19 geochemistry, geochronology and sequence stratigraphy, a robust chronostratigraphic framework for the latest Eocene to early
20 Oligocene was established. The stable isotopic shifts observed in benthic and planktic foraminifera $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records at
21 Karaburun align with global patterns but also reveal regional effects, such as freshwater influx and basin restriction, specific
22 to the semi-restricted Paratethys. The abrupt negative $\delta^{18}\text{O}$ shift across the EOB in the Paratethys reflects boreal water
23 ingressions driven by the onset of anti-estuarine circulation between the Nordic seas and Atlantic and the closure of the Arctic-
24 Atlantic gateway, which redirected cold, low-salinity boreal waters through interconnected basins towards the Paratethys.
25 These findings highlight the interplay between global climate drivers and regional hydrological dynamics, providing critical
26 insights into the evolution of marginal marine environments during the EOT. Our results underscore the significance of the
27 Paratethys as a unique archive for studying the onset of global icehouse climate conditions and regional responses.

28 1 Introduction

29 The Earth's geological history has witnessed several significant long-term climate transitions, along with short-term disruptions
30 to the carbon cycle. The most recent of these transitions occurred over the last 50 million years during the Cenozoic and is
31 characterized by a long-term cooling trend and a decline in atmospheric CO_2 levels, culminating in the onset of Antarctic
32 glaciation during the Eocene-Oligocene Transition (EOT) (e.g., Zachos et al., 2001; Caves et al., 2016). The EOT marks the

33 end of an extended period of predominantly greenhouse conditions and represents a phase of accelerated biotic change lasting
34 approximately 500–800 kyr, bracketing the Eocene-Oligocene Boundary (EOB) (Coxall & Pearson, 2007; Hutchinson et al.,
35 2021). This transition is also associated with a deepening of the ocean's calcite compensation depth (CCD) (Coxall et al.,
36 2005), a northward migration of the Intertropical Convergence Zone (ITCZ) (Hyeong et al., 2016), and increased seasonality
37 in northern high latitudes (Eldrett et al., 2009).

38

39 The Antarctic glaciation events during the Oligocene are referred to as Oi events (Pälike et al., 2006; Pekar and Miller, 1996).
40 At deep-ocean sites Oi glaciation events are characterized by positive excursions in the oxygen isotope records of benthic
41 foraminifera (Miller et al., 1991; Pälike et al., 2006; Wade and Pälike, 2004). During the earliest Oligocene there are two
42 significant cooling/glaciation events: (1) Oi-1 (Early Oligocene Glacial Maximum, EOGM of Hutchinson et al., 2020) and
43 (2) Oi-1a, corresponding to early part of paleomagnetic Subchron C12r, took place during the early Rupelian (Pekar et al.,
44 2002 and references therein). The EOGM, spanning 490 kyr from ~33.65 to 33.16 Ma, marks a sustained period of cold climate
45 and glaciation during paleomagnetic Subchron C13n (Hutchinson et al., 2020). The Oi-1a glaciation event corresponds to the
46 appearance of the cold-water dinocyst taxa *Svalbardella cooksoniae* and/or *Svalbardella* spp. in the North Sea Basin
47 (Śliwińska, 2019a and references therein). Episodes of southward migration of *Svalbardella cooksoniae* have consequently
48 been interpreted as evidence of cooling events (e.g., Van Simaey et al., 2005). *Svalbardella cooksoniae* has also been
49 documented in a brief interval during the earliest Oligocene at numerous sites across the Northern Atlantic and Western
50 Tethyan regions (Śliwińska and Heilmann-Clausen, 2011).

51

52 The most comprehensive insights into the EOT come from deep-sea marine records of benthic foraminiferal oxygen and carbon
53 isotopes ($\delta^{18}\text{O}$ and $\delta^{13}\text{C}$), extensively analyzed using cores from the Deep Sea Drilling Project (DSDP), Ocean Drilling Program
54 (ODP), and Integrated Ocean Discovery Program (IODP) (e.g., Zachos et al., 1996; Pekar and Miller, 1996; Salamy and
55 Zachos, 1999; Coxall et al., 2005; Bordiga, et al., 2015; Hutchinson et al., 2021 and references therein). In contrast, changes
56 associated with the EOT in marginal and epicontinental seas have been the focus of relatively few studies (e.g., Pearson et al.,
57 2008; Ozsvárt et al., 2016; van Der Boon et al., 2019; Dickson et al., 2021). Nevertheless, geochemical and sedimentary data
58 from these shallow regions can offer valuable insights into the impacts of the EOT, including Antarctic glaciation and cooling,
59 within restricted marine environments influenced by local factors such as freshwater influx, salinity variations, weathering,
60 erosion and terrestrial (sediment and carbon) fluxes.

61

62 This study addresses the gap in understanding the EOT conditions in epicontinental seas by revisiting and reanalyzing an
63 Eocene-Oligocene Boundary (EOB) section in the Karaburun area of northern Türkiye (Figure 1a). The Karaburun section
64 with its exceptionally well-preserved and diverse assemblages of microfossils (e.g. Simmons et al., 2020; Sancay and Batı,
65 2020) provides an excellent opportunity to investigate the distribution of key latest Eocene - earliest Oligocene organic walled
66 dinoflagellate cyst (dinocyst) and calcareous nannofossil index species. Furthermore, it sheds light on how climatic changes

67 influenced stratigraphic sequences in the eastern Paratethys Sea—the largest Cenozoic epicontinental sea, with no modern
68 analogue. To investigate these processes, the EOB section at Karaburun is analyzed using marine palynology emphasizing
69 dinocysts and calcareous nannofossil biostratigraphy, as well as high-resolution stable oxygen and carbon isotope analyses of
70 benthic (*Cibicidoides* spp.) and planktic (*Turborotalia ampliapertura*) foraminifera. U-Pb dating of a tuff layer within the
71 section further constrained the age model, complementing the biostratigraphic and chemostratigraphic data, also allowing
72 sequence stratigraphic interpretations. The findings are compared with the EOT records from other Paratethys sites and global
73 oceans to discern the regional and global climatic and oceanographic effects.

74 **2 Geological Setting**

75 The interplay of paleoclimatic and tectonic processes fragmented the largely enclosed Paratethys water body into numerous
76 sub-basins, separated by narrow, shallow gateways and land bridges (Palcu et al. 2022). Extending from southern Germany to
77 China (Figure 1b), Paratethys encompassed three distinct regions. The Western and Central Paratethys are characterized by
78 active tectonics and comprised smaller, short-lived basins. The Western Paratethys included the western Alpine foreland basin,
79 while the Central Paratethys included sub-basins spanning from Austria to Romania (e.g., Popov et al., 2004). In contrast, the
80 Eastern Paratethys, centered around the Black Sea and Caspian Sea basins, evolved within a tectonically stable region (Popov
81 et al., 2004).

82

83 The Karaburun area is situated along the southern margin of the Western Black Sea Basin, a back-arc basin formed during the
84 late Cretaceous as a sub-basin of the Paratethys Sea (Okay and Nikishin, 2015) (Figure 1c). Since its formation, the Western
85 Black Sea Basin has experienced continuous subsidence, resulting in a sedimentary thickness exceeding 14 km (Okay et al.,
86 2019). To the south of the Karaburun area lies the Thrace Basin (Figure 1a), which is younger and characterized by Eocene-
87 Oligocene clastic fill deposits, reaching a maximum thickness of approximately 9 km in its central region (Turgut, 1991).
88 During the Eocene and Oligocene, the Strandja Massif—a polydeformed, deeply eroded orogenic belt composed of
89 metamorphic and magmatic rocks—formed a paleo-high that separated the Western Black Sea Basin from the Thrace Basin
90 to the south (Cattò et al., 2018) (Figure 1a). The only marine connection between these basins was through the Çatalca Gap
91 (Figure 1a), located west of İstanbul, where sedimentation abruptly ceased during the early Oligocene due to an uplift event
92 (Okay et al., 2019). This region represents the sole contact point between the Eocene-Oligocene sequences of the Black Sea
93 and Thrace Basins (Okay et al., 2019).

94 **3 Regional Stratigraphy**

95 The Soğucak Formation, characterized by shallow marine, massive reefal limestone, underlies the uppermost Eocene–
96 lowermost Oligocene hemipelagic deposits in both the Thrace Basin and the Karaburun area (Figure 1b). In the Karaburun

97 area, the Soğucak Formation has been dated to the Priabonian based on benthic foraminiferal biozonation (Yücel et al., 2020).
98 Overlying this formation is a 120-meter-thick sedimentary succession from the latest Eocene to early Oligocene, predominantly
99 composed of hemipelagic marls and carbonates (Figure 1c). This sequence also includes intermittent submarine fan deposits,
100 debris flows, slumps, pebbly sandstones, and conglomerates near the top. A prominent tuff layer within this succession is
101 linked to a significant Rupelian volcanic event originating from the Rhodope Massif (Marchev et al., 2024). Although the
102 hemipelagic succession in the Karaburun area has often been referred to as the Ceylan Formation—following the terminology
103 used in the Thrace Basin (e.g., Natal'in and Say, 2015)—we adopt the designation "İhsaniye Formation," as recommended by
104 Okay et al. (2019) and Simmons et al. (2020). While prior studies assigned an early Oligocene age to the İhsaniye Formation
105 in the Karaburun area (e.g., Less et al., 2011; Okay et al., 2019; Simmons et al., 2020), our findings refine its age to the latest
106 Eocene–early Oligocene. This revision is based on the calcareous nannofossil and dinocyst biostratigraphy, stable oxygen and
107 carbon isotope analyses, and U-Pb dating of the tuff layer.

108 **4 Material & Methods**

109 **4.1 Lithostratigraphy**

110 Eocene and Oligocene deposits are well-exposed in 50-meter-high cliffs along the Black Sea coast in the Karaburun area
111 (Figure 1d). These deposits have been the focus of recent studies (e.g., Okay et al., 2019; Sancay and Bati, 2020; Simmons et
112 al., 2020; Tulan et al., 2020), which documented their paleoenvironment, biostratigraphy, and source rock potential. To build
113 upon these studies, we revisited the area and measured three adjacent stratigraphic sections—designated as KR1, KR2, and
114 KR3 (Figure 1d). By integrating these sections, we constructed a composite Karaburun section comprising hemipelagic
115 deposits of the İhsaniye Formation (Figure 2).

116

117 The studied sediments predominantly consist of hemipelagic light gray marls, dark brownish clays, thin- to medium-bedded
118 light gray, whitish, and beige carbonates, calcareous siltstones, and sandstones, which occasionally display planar lamination.
119 The sequence also includes submarine fan (turbiditic) conglomerates, as well as debris flow and slump deposits toward the
120 top. The hemipelagic fine-grained deposits contain rich microfossil assemblages of planktic and benthic foraminifera,
121 calcareous nannofossils, and dinocysts, indicating a latest Eocene–early Oligocene age (this study). The submarine fan deposits
122 are characterized by thin to medium thick, erosive-based conglomeratic beds with mainly carbonate pebbles, organic matter,
123 and microfossil fragments (e.g., foraminifera and shell debris) and intercalated with thin silty layers. They often grade vertically
124 into sandstone layers. Although these submarine fan deposits exhibit variable lateral thicknesses, they provide solid key
125 horizons for correlation of the subsections (e.g. correlation of the KR1 and KR3 subsections). Brownish organic-rich clay
126 layers occasionally contain red-yellow nodules. Pyrite is commonly found in these organic-rich layers. Additionally, pyritized
127 coral fossils are rarely observed in these organic-rich clay deposits. A distinctive white tuff layer at approximately 71.5 m
128 within the composite section was sampled for U-Pb zircon dating (see section 4.5) (Figure 2). Debris flow horizons, with a

129 maximum thickness of 5 meters, exhibit channel geometries and primarily consist of carbonate pebbles. These horizons
130 increase in frequency toward the topmost 20 meters of the succession. This study focuses on the lower and middle portions of
131 the section including the EOT, ending around the tuff layer (Figure 2), and does not include analysis of the uppermost part of
132 the succession including the debris flow deposits.

133 4.2 Sequence Stratigraphy

134 We analyzed various surfaces that indicate either a seaward or landward shift of successive facies belts, including erosive
135 surfaces which could be equivalent to subaerial unconformities on land, transgressive surfaces, and maximum flooding
136 surfaces. These surfaces demarcate the boundaries of different systems tracts—lowstand, transgressive, and highstand—which
137 together form the depositional sequences (Catuneanu, 2006 and references therein).

138

139 In addition to the identification of the systems tracts based on the recognition of key surfaces, sedimentary facies and
140 microfossil assemblages have been utilized to reconstruct past water depths and identify sea-level evolution, typically indicated
141 by shifts towards offshore (or onshore) characteristics. The distribution and relative abundance of planktic and benthic
142 foraminifera have further been employed to discern variations in relative sea level. Additionally, the relative abundance of
143 lagoonal and inner neritic dinocysts, combined with the distribution patterns of associated brackish and terrestrial
144 palynomorphs as well as the grain size of the deposited sediments, were examined to assess proximity to the coast.

145 4.3 Biostratigraphy

146 4.3.1 Calcareous nannofossils

147 The study on calcareous nannofossil assemblages was carried out on 84 samples, prepared at the Department of Earth Sciences
148 of the University of Milan (Italy), following the smear-slide standard technique described by Bown & Young (1998).
149 Calcareous nannofossils were analyzed using an Axioscop Zeiss light microscope (LM) at 1250X magnification. Preservation
150 of the specimens was generally good, as indicated by the presence of holococcoliths, coccospheres and small coccoliths.
151 Quantitative analysis was performed by counting 300 specimens per sample, in a variable number of fields of view, depending
152 on the nannofossils total abundance. Nannofossil frequency data were converted into the number of specimens per square
153 millimeter for the evaluation of the biostratigraphic signal, and into percentages to estimate the paleoecological significance
154 of the assemblage variations. The position of biohorizons recognized in this study is based on abundance patterns of index
155 species, according to Agnini et al. (2014) and Viganò et al. (2023a) and are labelled as follows: Top (T): the highest occurrence
156 of a taxon, Base common and continuous (Bc) and Top common and continuous (Tc): the lowest and highest common and
157 continuous occurrence of a taxon. For calcareous nannofossils taxonomy, we refer to Perch-Nielsen (1985), Agnini et al.
158 (2014), and the Nannotax web library (<https://www.mikrotax.org/Nannotax3>). The biostratigraphic schemes adopted here are
159 those of Martini (1971) and Agnini et al. (2014).

160 4.3.2 Marine Palynology - dinocysts

161 For marine palynological analysis, emphasizing dinocysts 42 samples were prepared at Petrostrat laboratories (Conwy, Wales,
162 UK; sections KR1 and KR2), and another 10 (from the sub-section KR3) at Utrecht University laboratories, according to
163 typical palynological processing techniques (see e.g., Cramwinckel et al., 2020). This involves freeze-drying and precision
164 weighing, subsequent HCl and HF treatments, followed by sieving residues over a 15 µm mesh sieve, before slides were
165 produced for light microscopy from the residues. Samples are spiked with a known amount of *Lycopodium clavatum* spores to
166 allow for absolute quantitative analysis (Stockmarr, 1972). After a broad palynofacies characterization (non-quantitative), light
167 microscopical analysis included counting of broad categories of aquatic and terrestrial palynomorphs up to a minimum of 100
168 identifiable dinocysts per sample. Fragments of palynomorphs identifiable or not (viz, fragments of indeterminable
169 palynomorphs to e.g., fragments of – therefore - indeterminable dinocysts, and including fragments of inner linings of benthic
170 foraminifera), were quantified as well (see Table S1).

171

172 For dinocyst taxonomy, we refer to that cited in Williams et al. (2019), except for taxa belonging to the Wetzelielloideae (see
173 discussion in Bijl et al., 2017). All materials are stored in the collection of the Marine Paleooceanography and Palynology group,
174 at the Laboratory of Palaeobotany and Palynology (Utrecht University, Faculty of Geosciences).

175 4.4 Geochronology

176 4.4.1 U-Pb dating

177 A 30 cm-thick volcanic tuff layer was identified at the 71.5 m level of the KR composite section, serving as a key marker for
178 constraining the age of the deposits. Three kilograms of tuff material were crushed and zircon crystals were separated by
179 standard heavy liquid techniques and mounted in epoxy resin. Thirty five zircon crystals were dated via U-Pb at the Envitop
180 analytical facility at CEREGE using an Element XR ICP-MS connected to a NWR193 laser (ArF 193 nm) ablation system.
181 Zircon crystals were ablated with a 25-micron spot diameter, a 15 Hz pulse repetition rate, an energy fluence of 1.5 J/cm², and
182 a carrier gas flow of 0.975 L/min. Data reduction, date, and date uncertainty calculations were conducted with an in-house
183 MATLAB script. We applied rigorous filtering based on zircon morphology, U-Pb concordance, and common Pb content to
184 ensure the reliability of the final age determination. The dataset was filtered for concordant grains using the Concordia distance
185 method of Vermeesch (2021), which applies isometric logratios with a discordance threshold of 10 (SI units) and a reverse
186 discordance threshold of 5 (SI units). Details about our U-Pb dating workflow and data reduction steps are given in Licht et
187 al. (2024). The three zircon validation reference materials used during these sessions yielded offsets around TIMS ages < 1%
188 in most cases, and < 2% otherwise. Out of the 35 analyzed zircon crystals, 12 yield concordant U-Pb ages (see Supplementary
189 Table S3). The final Concordia age was calculated with concordant ages only using IsoPlotR (Vermeesch, 2018).

190 4.5 Geochemistry

191 4.5.1 $\delta^{18}\text{O}$ & $\delta^{13}\text{C}$ analyses

192 Measurements of stable oxygen ($\delta^{18}\text{O}$) and carbon ($\delta^{13}\text{C}$) isotopes on benthic foraminiferal (*Cibicidoides* spp.) and planktic
193 foraminifera (*Turborotalia ampliapertura*) test fragments were performed at GEOMAR, Kiel on a Thermo Scientific MAT
194 253 mass spectrometer with an automated Kiel IV carbonate preparation device. The isotope values were calibrated versus the
195 NBS 19 (National Bureau of Standards) carbonate standard and the in-house carbonate standard (“Standard Bremen”,
196 Solnhofen limestone). Isotope values in delta-notation (δ) are reported in ‰ relative to the VPDB (Vienna Pee Dee Belemnite)
197 scale. The long-term analytical precision is 0.06 ‰ for $\delta^{18}\text{O}$ and 0.05 ‰ for $\delta^{13}\text{C}$ (1-sigma value). Replicate measurements
198 were not done due to the low numbers of specimens found.

199 5 Results

200 5.1 Biostratigraphy

201 5.1.1 Calcareous nannofossils

202 So far a number of high-resolution studies on calcareous nannofossils during the EOT have been published, focusing on
203 specific regions, including high latitudes (Villa et al., 2014) and mid- to low-latitudes (Bordiga et al., 2015; Fioroni et al.,
204 2015; Villa et al., 2021; Jones et al., 2019). Recently, new studies on IODP sediments have further enhanced our understanding
205 of this critical interval in the Paleogene paleoclimate history (Raffi et al., 2024, Viganò et al., 2023a, b, 2024a). Our study
206 contributes to the knowledge of nannofossil biostratigraphy in this interval, providing detailed documentation of the EOT
207 under marine conditions within the eastern Paratethyan realm (see Table S2 and Figure S9).

208

209 The extinction of *D. saipanensis* defines the base of Zone NP21 (Martini, 1971), which corresponds to the base of Zone CNE21
210 as described by Agnini et al. (2014). This species, as well as the “rosette shaped” *Discoaster*, is absent in the lowermost
211 samples analyzed, indicating that its last occurrence (at 34.4 Ma) predates the studied interval.

212 The Base of common (Bc) *Clausicoccus subdistichus* group defines the onset of Zone CNO1 of Agnini et al. (2014), which
213 corresponds to the upper part of Zone NP21 (Martini, 1971). The increase in abundance of this informal taxonomic group
214 represents the most reliable nannofossil bioevent to approximate the EOB across different regions (e.g. Marino and Flores,
215 2002; Toffanin et al., 2013; Fioroni et al. 2015, Viganò et al., 2023a). In the studied area, this event is well-delineated, showing
216 an abrupt increase in abundance from approximately 17 to over 100 specimens/mm² in sample N18, ca. 13 m from the base of
217 the studied section (Figure 2).

218

219 The top (T) of *Ericsonia formosa* marks the base of Zone NP22 of Martini (1971) and the base of Zone CNO2 (Agnini et al.,
220 2014). However, in the studied composite section, the precise position of this bioevent remains uncertain due to the rarity and

221 scattered occurrence of the taxon in its final range, nevertheless it is likely located before the end of the *C. subdistichus* acme
222 (Figure 2). This Top of common (Tc) and continuous occurrence of the *C. subdistichus* group, was positioned before the top
223 (T) of *E. formosa* by Agnini et al. (2014) in the biozonation adopted here. However, recent studies challenge this interpretation
224 (Viganò et al., 2023a, 2024a). In fact, recent findings suggest that the top common and continuous (Tc) of this group occurs
225 above the top (T) of *E. formosa*, as previously reported by Backman (1987) and Catanzariti et al. (1997). The top common and
226 continuous (Tc) of *C. subdistichus* gr., occurring early in Subchron C12r, has also been documented above the top (T) of *E.*
227 *formosa* in the Pacific (Toffanin et al., 2013; Viganò et al., 2023b), the Atlantic (Bordiga et al., 2015, Viganò 2023b) and the
228 Indian Ocean (Fioroni et al., 2015; Villa et al., 2021, Viganò 2023a). These studies indicate that *C. subdistichus* persists and
229 remains common even after the T of *E. formosa*. In our dataset, the abundance of *C. subdistichus* exhibits a marked decrease
230 (Tc) in the upper 5 meters of the studied section (Figure 2). Consequently, the T of *E. formosa* (i.e. the boundary between
231 NP21 and NP22) should be positioned at some point prior to this bioevent. This interpretation is supported by data from
232 dinocysts (section 5.1.2), which provide a better and more precise constraint for the stratigraphic position of the upper part of
233 the investigated section.

234 5.1.2 Dinocysts

235 Our work builds on the recent integrated study by Simmons et al. (2020) and notably that by Sancay and Bati (2020), targeting
236 the Karaburun area and outcrops using palynological approaches, emphasizing dinocysts. Their pioneering effort, using more
237 locations and sections, but with much lower sample resolution, now show the need for a higher-resolution approach,
238 particularly while considering the potential recognition of a continuous EOT interval. Therefore, here, we focus on the lower
239 parts of the section, with our higher-resolution sampling.

240

241 Unfortunately, palynomorph preservation and fragmentation varies significantly across the section, ranging from (most often)
242 very poor to only occasionally reasonable, and always typically heavily fragmented (Table S1; note the high number of
243 undeterminable - fragments of - specimens).

244

245 The dominant palynological groups throughout the succession are organic linings of benthic foraminifera, and notably their
246 fragments, besides dinocysts, and pollen and spores of terrestrial higher plants (Table S1). We also recovered several other
247 aquatic algal taxa and acritarchs. These include representatives of e.g., fresh to brackish water elements like *Cyclopsiella*,
248 *Pterospermella* and, *Tasmanites* spp., besides very small (<~20 µm) psilate and skolochorate cysts (viz, ‘acritarchs’, and other
249 ‘small skolochorate cysts’) of unknown ecology. Fungal spores and fruitbodies, as well as scolecodonts are occasionally
250 encountered as well. In terms of palynofacies (palynodebris) composition and trends, the samples are all very similar in
251 displaying a rich mix in mainly terrestrial plant-derived elements of varying sizes, combined with various amorphous materials.
252 Truly opaque material is conspicuously absent. No trends or breaks are apparent from this visual, non-quantitative assessment
253 (Table S1). Overall, these results match the findings by Sancay and Bati (2020).

255 Although the dinocyst assemblages are difficult to quantify because of preservation and fragmentation issues, taken together,
256 assemblages are highly diversified throughout, and essentially composed of well-known late Eocene to early Oligocene taxa
257 (See Figure S4, S5, S6, S7 and S8). The assemblages are quite comparable to those known from other EOT sections in the
258 larger Tethyan region, e.g., from central and northeast Italy (Brinkhuis and Biffi, 1993; Brinkhuis, 1994; Van Mourik and
259 Brinkhuis, 2005; Houben et al., 2012; Iakovleva, 2025), to North Africa (Egypt, El Beialy et al., 2019, Tunisia, Toricelli and
260 Biffi, 2001, and Morocco, Chekar et al., 2018, Mahboub et al., 2019; Slimani et al., 2019; Slimani and Chekar, 2023), and
261 further to the East, e.g., the Caspian Sea region (Bati, 2015), and Armenia (Iakovleva et al., 2024). In terms of robust dinocyst-
262 based age-assessment, best calibrated information is available from central and northeast Italy (e.g., Brinkhuis and Biffi, 1993
263 and follow-up studies), including a detailed zonal scheme for the EOT matched with magneto- and calcareous microfossil
264 stratigraphies. Based on the first, and last regional occurrence of *Glaphyrocysta semitecta* the base of the Gse and the Rac/Cin
265 zonal boundary of Brinkhuis and Biffi (1993) can be recognized in the Karaburun sections (Figure 2, Table S1). Furthermore,
266 based on the last occurrence of *Hemiplacophora semilunifera*, the Gse/Adi zonal boundary can be recognized as well (Figure
267 2). Remarkably, the important index species *Areosphaeridium diktyoplokum* is so far not recorded in any of the sub-sections
268 with confidence, hampering the recognition of the Adi/Rac zonal boundary. This is noteworthy, as elsewhere in the region, the
269 species is typically quite abundantly present in the deposits assigned to the absolute earliest Oligocene (as defined by the
270 extinction of the Hantkeninids, planktonic foraminifera; see e.g., Brinkhuis and Visscher, 1995; Van Mourik and Brinkhuis,
271 2005; Houben et al., 2012). Yet, the recognition of the other zonal boundaries allows confident correlation to the EOT interval,
272 throughout matching the assignments by calcareous nannofossils, and the typical benthic foraminifer $\delta^{18}\text{O}$ -EOT-profile
273 (including the Oi-1a event – cooling during the early part of Subchron C12r) discussed further below.

274

275 These correlations are here bolstered by the spot-occurrence of *Svalbardella cooksoniae* in sample Y41 at 71 m, assigned to
276 the Cin Zone (Figure 2, Figure S5a, b, and c). This event was previously described from the central Italian EOT section, within
277 the same Cin Zone (Brinkhuis and Biffi, 1993). At high northern latitudes, this species ranges from the late Eocene way into
278 the Oligocene (e.g., Eldrett et al., 2004; Eldrett and Harding, 2009). Subsequent work noted that colder episodes during the
279 Oligocene likely induced equatorward migration of such typical high-latitude taxa (e.g., van Simaey et al., 2005). In effect,
280 our finding reflects the earliest one of such migration pulses, an event well documented by Śliwińska and Heilmann-Clausen
281 (2011). These authors showed that *Svalbardella cooksoniae* is consistently present in the same narrow interval calibrated to
282 the basal Subchron C12r, close to the NP21/NP22 boundary, in many high- and mid-latitude Northern Hemisphere sections,
283 ranging from the Greenland Sea in the north to Italy in the south. Moreover, they correlated this event to the Oi-1a oxygen
284 isotope maximum of Pekar and Miller (1996) and Pekar et al. (2002). Another interesting finding is specimens of the acritarch
285 *Ascotomocystis potane* in samples from the sub-section KR2. Documented from the basal Rupelian type section in Belgium
286 (Stover and Hardenbol, 1993), this further confirms assignment to the basal Oligocene.

287 **5.2 U-Pb dating of the tuff**

288 Concordia plot of the dated tuff is available in Supplementary Figure S10. The tuff sample yields an early Rupelian $32.55 \pm$
289 0.38 Ma (2σ) age, based on 11 out of the 12 dated zircons. The age of the tuff layer aligns with the biostratigraphic dating,
290 which places the top of the KR composite section within the Cin dinoflagellate cysts zone (Figure 2).

291 **5.3 Sequence Stratigraphy**

292 Based on lithological, sedimentological, and paleontological characteristics, we interpret the depositional environment as a
293 shelf setting exhibiting littoral and neritic characteristics, depending on fluctuating sea levels. Our sequence stratigraphic
294 analysis identifies ten distinct depositional sequences within this setting (Figure 3). The lower part of the section, up to
295 approximately 11 m, comprises three depositional sequences (S1, S2, and S3), characterized by intercalated
296 pebbly/conglomeratic layers, marls, and claystones (Figure 3). The Lowstand Systems Tracts (LSTs) and Highstand Systems
297 Tracts (HSTs) contain both pebbly/conglomeratic layers and fine-grained deposits, whereas the Transgressive Systems Tracts
298 (TSTs) are represented exclusively by fine-grained marls and claystones. The relatively low abundance of lagoonal
299 dinoflagellate cysts (20–40%) in this interval suggests a distal position far from the coastline. Between approximately 20 m
300 and 23 m, within Sequence 5, the depositional setting represents the deepest marine conditions, marking a more distal position
301 relative to the coastline. This interval corresponds to the TST within Sequence 5.

302 At around 23 m, just below the last conglomeratic layer, a maximum flooding surface marks the base of the HST within
303 Sequence 5, coinciding with the highest percentage of lagoonal dinoflagellate cysts. This increase in lagoonal dinoflagellates
304 (up to 60–70%) continues into the LSTs of Sequences 6 and 7, indicating a more proximal position near the coastline during
305 the Eocene-Oligocene Glacial Maximum (EOGM).

306

307 The upper part of the section, including the Early Oligocene Cooling (see Section 6.3), consists of four sequences (S7, S8, S9,
308 and S10). A notable difference in the thickness of the depositional sequences is observed between the lower section (up to ~11
309 m) and the overlying sequences (from ~11 m to the top) (Figure 3). This variation in depositional thickness could be related to
310 obliquity forcing during the latest Eocene, a process that has been identified in several terrestrial and marine records (e.g.,
311 Abels et al., 2011; Jovane et al., 2006; Boulila et al., 2021).

312

313 Finally, we compared our reconstructed relative sea-level variations, based on the depositional sequences and systems tracts
314 described above, with the global sea-level reconstruction of Miller et al. (2020) (Figure 3). In the Eastern Paratethys region,
315 relative sea-level evolution during the latest Eocene and early Oligocene appear to follow a pattern parallel to global sea-level
316 changes. (Note that the base of the Karaburun composite section is younger than 34.4 Ma, see Section 5.1.1.) These high-
317 resolution sea-level evolution offer a refined reconstruction of Eastern Paratethys sea-level changes (e.g., Popov et al., 2010)
318 and provide a valuable framework for future studies to further reveal the history of Paratethys sea-level evolution..

319 5.4 Geochemistry

320 5.4.1 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ isotope analyses

321 The $\delta^{18}\text{O}$ values of benthic foraminifera (*Cibicidoides* spp.) in the Karaburun composite section range from -9.5‰ to 1.0‰,
322 displaying distinct temporal variations throughout the sequence. At the base of the section, at 11.40 m, just before the EOB,
323 a small positive peak (1.0‰) is noticeable (Figure 4). Following this, a pronounced and abrupt negative shift from -0.1‰ to -
324 9.5‰ is evident just after the Eocene-Oligocene Boundary (EOB). Following the abrupt negative shift, the $\delta^{18}\text{O}$ values increase
325 gradually and then sharply, reaching a peak of 1.0‰ at 19.60 m. This represents a two-phase increase in $\delta^{18}\text{O}$ during the EOT
326 as in other EOT records (e.g., Katz et al., 2008), however, with a negative shift in between. Following the second increase, a
327 negative shift to -2.3‰ occurs at 22 m, followed by a renewed increase in $\delta^{18}\text{O}$ values, forming a plateau that culminates at
328 0.6‰ by 36.75 m. The $\delta^{18}\text{O}$ values then decline, reaching a negative peak of -1.6‰ at 40.25 m, before gradually rising to 0.4‰
329 at 60 m, marking a second, shorter plateau. This plateau is interrupted by a decrease to -1.6‰ at 65.75 m, followed by a slight
330 recovery to 0.5‰ at 69 m. In the uppermost portion of the section, $\delta^{18}\text{O}$ values drop to -2.4‰ at 74.50 m and then exhibit a
331 modest increase toward the top of the section (Figure 4).

332
333 The $\delta^{13}\text{C}$ values of benthic foraminifera (*Cibicidoides* spp.) in the Karaburun composite section range from -0.7‰ to 2.2‰,
334 exhibiting greater variability compared to the $\delta^{18}\text{O}$ values. However, similar to the $\delta^{18}\text{O}$ values, the $\delta^{13}\text{C}$ values also display a
335 prominent shift towards more depleted levels (from 2.1‰ to 0‰) just after the EOB (Figure 5). The two-phase increase in
336 $\delta^{13}\text{C}$ (from the EOB up to ca 24 m) is again evident, interrupted by a sharp decline (ca. 20 m) that corresponds precisely with
337 changes in $\delta^{18}\text{O}$ (Figure 4 & 5). The $\delta^{13}\text{C}$ appears to lag behind by several tens of thousands years compared to the two-phase
338 increase in $\delta^{18}\text{O}$ (e.g., Coxall et al., 2011). The increase just before the EOB between 11.20 m and 12 m represents a 1.3‰
339 shift, followed by a 1.5‰ increase between 13.20 m and 15 m and a 1.0‰ increase between 19.60 m and 23 m. Before the
340 EOB, two more major positive carbon isotope excursions are observed at the base of the section: one between 1 and 7.5 m
341 (1.0‰) and the second between 7.5 m and 11.20 m (0.1‰). All excursions, except the second at the base (between 7.5 m and
342 11.20 m), exhibit a significant positive shift of ≥ 1.0 ‰. After the excursion between 19.60 m and 23 m, the $\delta^{13}\text{C}$ values drop
343 sharply (around 23–27 m) before exhibiting another series of positive carbon isotope excursions in the middle and upper parts
344 of the section (Figure 5). Like the excursions at the base of the section, the excursions in the middle and upper parts also
345 display positive shifts of approximately 1‰, with the most pronounced reaching 1.5‰. This pattern reflects a dynamic carbon
346 cycle with notable variations throughout the sequence.

347
348 Due to the intermittent presence of planktic foraminifera *Turborotalia ampliapertura* along the composite section, the $\delta^{18}\text{O}$
349 and $\delta^{13}\text{C}$ records of planktic foraminifera exhibit some gaps (Figures 6 and S1). Nevertheless, the overall trends and shifts
350 remain discernible. Similar to the $\delta^{18}\text{O}$ record of benthic foraminifera, the $\delta^{18}\text{O}$ of planktic foraminifera exhibits a clear positive
351 shift at the base of the section at 11.40 m just before the EOB, although slightly smaller (1.2‰). Following a slight positive

352 shift (0.4‰) just after the EOB at 14.45 m, the most prominent and significant positive shift recorded by the benthic
353 foraminifera is not fully captured in the planktic foraminifera record due to the absence of *T. ampliapertura* in the 4 m interval
354 between 15.90 m and 19.90 m. However, the positive 2.3‰ shift observed from 15.90 to 19.90 m represents partly the major
355 positive shift, albeit smaller than the shift recorded in the benthic foraminifera $\delta^{18}\text{O}$. An interval of ca. 7 m without any *T.*
356 *ampliapertura* follows the pronounced positive shift between 19.90 m and 26.5 m, obscuring the $\delta^{18}\text{O}$ record for that interval.
357 Following the barren interval, a gradual increase of 0.9‰ is clearly observed, extending up to 36.75 m. In the interval between
358 36.75 m and 61 m, the $\delta^{18}\text{O}$ record of planktic foraminifera fluctuates, exhibiting two significant alternating trends of decrease
359 and increase. Following this fluctuation, a gradual decrease is observed at the top of the section, followed by a sharp decline.
360
361 A distinct decline in the $\delta^{13}\text{C}$ record of planktic foraminifera (1.5‰) is observed just before the EOB at 11.60 m (Figures 6
362 and S2). Following this decline, $\delta^{13}\text{C}$ values increase at the EOB, reaching to 1.5‰ at 15.90 m with a shift of 1.8‰. Due to
363 the infrequent presence of *T. ampliapertura*, the interval between 15.90 m and 26.5 m is not fully represented; however, this
364 interval includes the highest $\delta^{13}\text{C}$ values observed. After these peak values, a gradual decrease is noted towards the top of the
365 section, with six positive shifts interrupting this trend: between 34.50 m and 36.75 m, 45.20 m and 47 m, 49 m and 51 m, 56.50
366 m and 61 m, 62.25 m and 65.75 m, and 65.75 m and 70.10 m. At the top of the section, a sharp decline in $\delta^{13}\text{C}$ values is
367 noticeable.

368 **5.5 Paleoenvironment & Paleoecology**

369 **5.5.1 Calcareous nannofossils**

370 Calcareous nannoplankton are highly sensitive to environmental changes in their surface water habitats, and fluctuations in
371 nannofossil assemblages are interpreted as responses to shifts in sea surface temperature (SST), nutrient concentrations,
372 salinity, and other environmental factors (*e.g.*, Aubry, 1992; Winter et al., 1994) thereby reflecting palaeoceanographic
373 perturbations. Numerous studies have explored the ecological tolerance of extinct taxa, establishing paleoecological
374 preferences through biogeographic studies (*e.g.*, Wei and Wise, 1990) and comparison with diverse environmental proxies
375 (*e.g.*, Villa et al., 2014). We discuss the behavior of several taxa within the Karaburun assemblage, based on paleoecological
376 affinities outlined in previous works.

377

378 *Reticulofenestra daviesii* and *Chiasmolithus* spp. are considered cool-water taxa (Wei et al., 1992; Villa et al., 2008) with
379 preference for eutrophic conditions (Villa et al., 2014, Viganò et al., 2024b). This paleoecologic group is recorded with very
380 low relative abundances, with few positive peaks in the middle part of the studied section. *Cyclicargolithus floridanus*, a typical
381 eutrophic open-ocean species (Auer et al., 2014), occurs with abundances reaching up to 50% in the lower and upper part of
382 the studied section, suggesting high productivity conditions (Aubry, 1992, Dunkley Jones et al., 2008, Villa et al., 2021) (Figure
383 S3). Small reticulofenestrads constitute a significant component of the assemblages, reaching over 60% in the middle part of

the section. They have been reported as dominant components of the nannoflora along continental margins (Haq, 1980). These settings are typically characterized by eutrophic conditions, driven by continental runoff and/or riverine input. Consequently, these small coccoliths are regarded as opportunistic taxa with broad ecological tolerance, yet particularly well-adapted to nutrient-rich environments (Aubry, 1992) and indicative of increased availability of terrigenous nutrient (Wade and Bown, 2006).

The genus *Helicosphaera* has been linked to increased nutrient availability (De Kaenel and Villa, 1996; Ziveri et al., 2004). Studies on extant coccolithophorids confirm the relationship of helicosphaerids with high primary productivity rates (Haidar and Thierstein, 2001; Toledo et al., 2007), and their preference for near-shore environments (Ziveri et al., 2004; Guerreiro et al., 2005). At the Karaburun section, this genus is recorded at low abundances but occurs consistently throughout the section (Figure S3).

Evidence of nutrient availability is further supported by the presence of braarudosphaerids, which are associated with coastal, low-salinity waters (Peleo-Alampay et al., 1999; Thierstein et al., 2004; Konno et al., 2007), eutrophic conditions (Cunha and Shimabukuro, 1997, Bartol et al., 2008) and the influx of terrigenous material (Švábenická, 1999). Braarudosphaerids are rarely found in the open ocean and thrive under unusual marine conditions, demonstrating a tolerance for environmentally stressed settings. Similar conditions are indicated by the presence of *Micrantholitus*, a taxon typically associated with shallow marine environments (Bown, 2005), reduced salinity, and eutrophic conditions (Street and Bown, 2005; Bown and Pearson 2009). These penthaliths occur from the base of the investigated section, albeit at low percentages and with a discontinuous distribution, further suggesting eutrophication and reduced salinity (Figure S3).

The presence of Ascidian spicules, with their highest and continuous occurrence in the middle-upper part of the section, also points to a shallow-marine depositional setting (e.g. Varol, 2006; Ferreira et al., 2019) and high surface-water productivity (Toledo et al., 2007). Furthermore, the relatively common occurrence of holococcoliths (mainly *Lanternithus minutus* and *Zigrablithus bijugatus*), *Pontosphaera* spp. and *Helicosphaera* spp., taxa prone to dissolution (Bown, 2005; Monechi et al., 2000) reinforces the interpretation of a shallow-water environment.

5.5.2 Marine palynology - dinocysts

For the analysis of the marine palynological assemblages, emphasizing dinocysts, we rely on the taxonomical and ecological dinocyst groups derived from modern distributions (e.g., Zonneveld et al., 2013; Marret and De Vernal 2024) and empirically based paleoecological information or the Paleogene dinocysts following previous works (e.g., Brinkhuis, 1994; Pross and Brinkhuis, 2005; Sluijs et al., 2005; Frieling and Sluijs, 2018). However, as mentioned above, the assemblages are generally too poorly preserved to allow for detailed quantitative considerations. Yet, given the overall quantitative characteristics of the studied samples, viz, substantial terrestrial input, and consistent dominances of taxa empirically known from restricted marine

417 to inner neritic (incl. lagoonal) like the goniodomid-group of dinoflagellate cysts (in this case e.g., *Homotryblum*,
418 *Polysphaeridium*, *Heteraulacacysta*, *Eocladopyxis* spp.), and the peridinioids (*Lentinia*, *Phthanoperidinium*, *Senegalinium*,
419 and *Deflandrea* spp.), neritic to outer neritic (e.g., *Areoligera*, *Glaphyrocysta*, *Enneadocysta*, *Spiniferites* and *Operculodinium*
420 spp.) combined with a small, but consistent contribution from offshore, oceanic taxa like *Impagidinium* and
421 *Nematosphaeropsis* spp. points to an essentially open marine, offshore, hemipelagic setting, comparable to e.g., the central
422 Italian sections (cf. Brinkhuis and Biffi, 1993).

423 Despite the issues with preservation throughout, a percentage-plot of fresh water tolerant, and restricted to inner neritic marine
424 taxa vs more offshore taxa still reveals stronger influxes in the latter part of the Gse, and within the Adi Zone (Table S1; see
425 above, and compare e.g., Frieling and Sluijs, 2018, and Sluijs and Brinkhuis, 2024). In effect, this aspect matches the records
426 from elsewhere (e.g., the Italian sections), and was earlier interpreted to reflect general eustatic sea level lowering associated
427 with the Oi-1 stable isotope event reflecting the earliest glaciation of Antarctica (e.g., Brinkhuis, 1994). In terms of
428 temperature/climatic changes, the conspicuous increase in Gymnospermous (conifer) bisaccate pollen input may be significant
429 as well. Again, a similar trend was noted in the Italian sections across the EOT (Brinkhuis and Biffi, 1993; Brinkhuis, 1994).

430 **6 Discussion**

431 **6.1 Age control overview**

432 An initial age model for the KR composite section was constructed using tie points derived from nannofossil and dinocyst
433 biozonations, combined with U-Pb dating of a tuff layer at 71.5 m (Figure 2). Additional age constraints were obtained by
434 aligning the Karaburun benthic foraminifera $\delta^{18}\text{O}$ data with the high-resolution benthic foraminifera $\delta^{18}\text{O}$ record from the
435 Atlantic sites 522 and 1263 (Figure 4). This alignment was achieved by identifying corresponding features in the isotope
436 records, such as positive and negative shifts and their amplitudes. The carbon isotope record of benthic foraminifera provided
437 independent validation of this tuning (Figure 5). The $\delta^{13}\text{C}$ benthic foraminifera data from the Karaburun area was correlated
438 and aligned with global high-resolution benthic foraminifera $\delta^{13}\text{C}$ records from deep-sea sites, including 1218 (equatorial
439 Pacific), 689 (sub-Antarctic Atlantic), 522 (South Atlantic), and 744 (southern Indian Ocean). Similarly, the planktic
440 foraminifera provided a further confirmation for our age model (Figures S1, S2). The planktic foraminifera $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data
441 from the Karaburun area were correlated and aligned with high-resolution $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records of planktic foraminifera in
442 hemipelagic sediment cores retrieved from the African margin of the Indian Ocean (Tanzania Drilling Project sites 12 and 17,
443 Pearson et al., 2008). All ages were assigned following the integrated magneto-biostratigraphic GTS2012 timescale. Our
444 geochemical results indicate that the increases in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ observed during the EOT at mid- and high-latitude sites in the
445 South Atlantic, Southern Ocean and Pacific are also present in the Paratethys, verifying that these signals are genuinely global
446 and valuable for stratigraphic correlation.

447 According to the constructed age model, the base of the section dates to the late Eocene (late Priabonian). The EOB is identified
448 at 12.75 m. The middle and upper parts of the composite KR section correspond to the early Oligocene (early Rupelian) (Figure

2). The revised age constraints established in this study offer a robust chronostratigraphic framework for the latest Eocene to early Oligocene interval of the Karaburun section, surpassing the accuracy of prior studies (e.g., Less et al., 2011; Okay et al., 2019; Simmons et al., 2020). However, we have to note that the absence of the typically abundant index taxon *Areosphaeridium diktyoplokum* in all studied sub-sections currently hampers precise recognition of the Adi/Rac zonal boundary and may reflect local paleoenvironmental or taphonomic conditions, or alternatively, point to a slight stratigraphic gap or condensed interval, a possibility that merits further investigation.

6.2 The EOT

At the deep Atlantic Site 522 and Pacific Site 1218, the late Eocene Event is marked by a transient interval of positive $\delta^{18}\text{O}$ values, reflecting a short-lived cooling or glacial episode (Hutchinson et al., 2021) (Figure 4). This isotopic shift measures approximately 0.6‰ and 0.4‰ at Site 522 and Site 1218, respectively. Similarly, the base of the KR composite section exhibits an increase of 0.7‰ in benthic foraminifera $\delta^{18}\text{O}$ values at approximately 5.5 m, which we interpret as evidence of the Late Eocene Event (Figure 4). The onset of this event coincides with the extinction of *Discoaster saipanensis* at 34.44 Ma at Site 1218. Based on calcareous nannofossil data (i.e., the absence of *D. saipanensis*), the base of the KR section is inferred to be younger than 34.44 Ma, supporting this correlation. The Late Eocene Event represented by this 0.7‰ positive shift in $\delta^{18}\text{O}$ values at approximately 5.5 m marks the onset of the EOT in the KR composite section (e.g., Hutchinson et al., 2020) (Figure 4).

The initial $\delta^{18}\text{O}$ step increase, occurring just before the EOB, has been identified as Step 1 in some records (e.g., EOT-1 in Katz et al., 2008; Precursor Glaciation in Scher et al., 2011). The first 1.0‰ $\delta^{18}\text{O}$ increase observed in the KR composite section at 11.40 m is interpreted as Step 1 as in the previous records (e.g., EOT-1 in Katz et al., 2008; Precursor Glaciation in Scher et al., 2011) (Figure 4). A similar $\delta^{18}\text{O}$ increase of 0.9‰ is recorded in the Alabama St. Stephens Quarry core (Miller et al., 2008). The onset of Step 1 is dated to 34.15 Ma, with an estimated duration of approximately 40 kyr (Hutchinson et al., 2020).

The Earliest Oligocene Oxygen Isotope Step (EOIS) represents a rapid $\delta^{18}\text{O}$ increase (0.7‰ or more) occurring well after the EOB, within the lower part of Chron C13n (Hutchinson et al., 2020). The peak $\delta^{18}\text{O}$ is recorded at approximately 33.65 Ma, with the entire EOIS lasting around 40 kyr. In the KR composite section, the positive shift associated with EOIS is approximately 2‰, peaking at 1‰ at 19.60 m, marking the end of the EOT (Hutchinson et al., 2020) (Figure 4).

The Early Oligocene Glacial Maximum (EOGM) is characterized as a prolonged period of cold climate and glaciation during the early Oligocene, corresponding to the most of paleomagnetic Subchron C13n (Hutchinson et al., 2020) (Figure 4). It spans from approximately 33.65 Ma to 33.16 Ma, lasting about 490 kyr. Correlation between the Karaburun data and global deep-

481 sea records was achieved by aligning the peak-to-peak $\delta^{18}\text{O}$ stratigraphic intervals, starting at the top of the EOIS at 19.60 m
482 and extending to another peak at 36.75 m (0.6‰) corresponding to the top of Subchron C13n (Figure 4).

483

484 Overall, the $\delta^{18}\text{O}$ record of benthic foraminifera from the Karaburun composite section closely mirrors global $\delta^{18}\text{O}$ trends from
485 deep-sea sites, except for a sharp decrease observed just after the EOB (Figure 4). The EOT signal is clearly recorded in the
486 $\delta^{18}\text{O}$ benthic foraminifera data from the Karaburun composite section. However, the relatively lower $\delta^{18}\text{O}$ values and sharp
487 decrease just after the EOB are attributed to regional conditions in the Paratethys Sea, as discussed in Section 6.4.

488 6.3 The early Oligocene cooling

489 The presence of the cold-water dinoflagellate *Svalbardella cooksoniae* within a brief interval of the early Oligocene in the
490 North Atlantic and Western Neo-Tethyan realms has been previously documented and linked to the Oi-1a oxygen isotope
491 maximum (Śliwińska and Heilmann-Clausen, 2011). This oxygen isotope maximum representing a cooling event occurs during
492 the early part of Subchron C12r, near the NP21/NP22 boundary. In the KR composite section, the *Svalbardella cooksoniae*-
493 bearing sample aligns with an oxygen isotope maximum at approximately 71 m, occurring during the early phase of Subchron
494 C12r (Figure 4). Consequently, the boundary between NP21 and NP22 is likely located near this level. In the North Sea, the
495 *S. cooksoniae* event was identified at the top of the regressive systems tract (OSS-21 RST) in the 11/10-1 well (Śliwińska,
496 2019a). Similarly, in the KR composite section, the *S. cooksoniae* event is positioned at the top of a Lowstand Systems Tract,
497 in agreement with the North Sea data (Figure 3). Strontium isotope analyses by Jarsve et al. (2015) suggest an age of 32.66
498 Ma for this event. Our U-Pb dating of the tuff layer located just above the *Svalbardella* spp.-bearing interval yields an age of
499 32.55 ± 0.38 Ma, aligning closely with the strontium-based age reported by Jarsve et al. (2015). These findings further support
500 the interpretation of Śliwińska and Heilmann-Clausen (2011) that the earliest Rupelian *S. cooksoniae* interval across the
501 Tethys, Central Europe, the North Sea Basin, the Norwegian-Greenland Sea, and the Eastern Paratethys is coeval with the Oi-
502 1a event and corresponds to a significant sea-level fall (Figure 3).

503

504 In support of the geochemical evidence provided by $\delta^{18}\text{O}$ values in benthic foraminifera, a notable increase in gymnospermous
505 (conifer) bisaccate pollen is clearly observed at the KR composite section during the EOT, EOGM, and Oi-1a events (Figure
506 3). This increase is likely associated with cooling and glaciation events occurring during these intervals, as previously
507 suggested by Brinkhuis and Biffi (1993) and Brinkhuis (1994).

508

509 The early Oligocene cooling event (Oi-1a) was previously dated to 32.8 Ma by Pekar et al. (2002). At the KR composite
510 section, the peak $\delta^{18}\text{O}$ values in benthic foraminifera ($\sim 0.4\text{‰}$) observed around 58–60 m are interpreted as representing the
511 Oi-1a event. Our age model corroborates the age proposed by Pekar et al. (2002), further supporting a timing of 32.8 Ma for
512 these peak values.

513 6.4 The regional & global effects in the Paratethys

514 The benthic foraminiferal oxygen and carbon isotope records from the Karaburun area closely resemble deep-sea records from
515 Atlantic sites 522 and 1263 during the latest Eocene and early Oligocene (Figures 4, 5). However, a notable distinction is the
516 pronounced negative $\delta^{18}\text{O}$ shift just after the EOB, a feature characteristic of the Paratethys region (Figures 4 and 7) which
517 will be discussed in the following. Firstly, it is noticeable that the overall benthic and planktic foraminifera $\delta^{18}\text{O}$ values are
518 more depleted than global records from the EOT. These depleted values likely reflect a regional effect rather than diagenetic
519 alteration, as the exceptional preservation and glassy appearance of the foraminiferal shells from the KR composite section
520 and other Paratethys sites (e.g., Ozsvárt et al., 2016) suggest minimal recrystallization. A major diagenetic overprint affecting
521 the entire basin is also improbable given the differing tectonic and depositional histories across the Paratethys sub-basins.
522 Additionally, the observed timescale (<100 ka) and the significant magnitude of changes in proxy records from the Paratethys
523 Basin are unlikely to be explained by regional tectonic processes. Instead, factors such as basin restriction, enhanced
524 precipitation and/or freshwater input due to increased runoff or changing hydrological conditions during the EOT seem more
525 plausible explanations. Similar $\delta^{18}\text{O}$ depletion has been documented in other marginal basins during the EOT (e.g., Pearson et
526 al., 2008; De Lira Mota et al., 2023), further supporting a localized effect in semi-restricted environments due to local
527 hydrology and climate. Indeed, these values are consistent with the isotopic composition of meteoric waters at mid-latitude
528 coastal regions ($\delta^{18}\text{O} \sim -5\text{‰}$ to -10‰ ; Dansgaard, 1964; Gat, 1996). During the Rupelian (35–31 Ma), the dominant influence
529 of Atlantic derived westerlies likely brought increased precipitation with a depleted $\delta^{18}\text{O}$ signature to the western and central
530 Paratethys. This interpretation was supported by $\delta^{18}\text{OPO}_4$ values from herbivore tooth enamel, which reflect the depleted
531 isotopic composition of drinking water (Kocsis et al., 2014). The enhanced precipitation was likely due to the intensification
532 of the westerlies and the reorganization of oceanic and atmospheric circulation (e.g., Hou et al., 2022). As global cooling
533 progressed and the meridional temperature gradient steepened, the westerlies strengthened, enhancing vapor transport from
534 the Atlantic into Eurasia, contributing to higher precipitation in the western and central Paratethys (e.g., Kocsis et al., 2014;
535 Li et al., 2018). Modeling studies further suggest prevailing westerly winds during winter at mid-high latitudes in the Rupelian
536 (Li et al., 2018). The "continental effect," where $\delta^{18}\text{O}$ in meteoric water becomes progressively fractionated with increasing
537 transport distance from the Atlantic, likely contributed to more negative $\delta^{18}\text{O}$ values in the Karaburun area (e.g., Kocsis et al.,
538 2014). Basin geometry, particularly water depth, plays a crucial role in evaporation dynamics within a semi-isolated system
539 like the Paratethys. The Western Paratethys, characterized by deeper basins, had a higher heat capacity and greater thermal
540 inertia, leading to moderated surface temperatures and lower evaporation rates. In contrast, the shallower Eastern Paratethys
541 had lower heat capacity, allowing for rapid surface warming, especially in warmer seasons, which enhanced evaporation.
542 Additionally, wind-driven effects likely played a significant role. In deeper basins, wind-driven mixing redistributed heat
543 within the water column, reducing extreme surface warming and stabilizing evaporation rates. In shallower areas, reduced
544 mixing allowed surface waters to remain warmer, leading to increased evaporation, particularly under strong seasonal winds.
545 Hence, further east, evaporation over the shallower Eastern Paratethys may have added moisture to westerly air trajectories,

546 resulting in relatively less negative $\delta^{18}\text{O}$ values in the Northern Caucasus (Karaburun, Belaya and Chirkei sections in Figure
547 7) and increased inland precipitation (Figure 7). A similar precipitation gradient, with wetter conditions in the western-central
548 Paratethys and drier conditions in the east, is also evident in an Oligocene climate reconstruction based on plant macrofossil
549 data (Li et al., 2018). Additionally, the increased freshwater input in the Paratethys at the EOB could be plausibly explained
550 by the major sea level fall and falling base level, driven by glacio-eustasy associated with the growth of Antarctic ice sheets
551 during the EOT. The reorganization of rivers due to the falling base level would have introduced fresh water into the
552 depositional epicenters of the Paratethys. Combined, the effects of regional hydrological change and the base level fall due
553 the global major sea level fall at the EOB might have resulted in the depleted $\delta^{18}\text{O}$ values observed in the Paratethys. Despite
554 the localized variations in the depletion of $\delta^{18}\text{O}$ values, the relatively consistent $\delta^{18}\text{O}$ depletion observed across Paratethys
555 sections suggests a uniform basin-wide isotopic background. This consistency allows for reliable identification of major trends
556 and isotopic excursions in the Paratethys Basin during the EOT.

557

558 Second to notice is that before the EOB a parallel trend could be observed for the benthic and planktic foraminifera $\delta^{18}\text{O}$ and
559 $\delta^{13}\text{C}$ values (Figure 6). Particularly during the Step1 same trends in isotopic shifts could be clearly recognized. Most
560 significantly, just after the EOB a distinct contrasting trend between benthic and planktic foraminifera $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ could be
561 noticed (Figure 6). These contrasting trends between benthic and planktic foraminifera $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ just after the EOB suggest
562 significant stratification and a reduction in vertical mixing. The pronounced negative $\delta^{18}\text{O}$ shift in benthic foraminifera likely
563 reflects a significant influx of isotopically light cold freshwater into the bottom waters. The slight increase in planktic
564 foraminifera $\delta^{18}\text{O}$ at the same time suggest that the surface waters might have become relatively saline due to evaporation
565 exceeding freshwater input in the surface layer, which would increase $\delta^{18}\text{O}$ values. Cold, freshwater inflow might have been
566 funneled into deeper areas of the basin, displacing or mixing with bottom waters. This might have been happened through the
567 submarine channels providing sediment-laden freshwater as underflows into the deep-marine turbiditic systems. However, this
568 would have required an anti-estuarine circulation model where marine saltwater flows upstream and overrides the freshwater
569 inflow. Indeed, an anti-estuarine circulation model for the early Oligocene Paratethys was proposed earlier (Dohmann, 1991)
570 which was later supported by Schulz et al. (2005) showing also increasing surface salinities due to evaporation of marine water
571 based on increasing di-/tri-MTTC ratios. An early anti-estuarine circulation model for the Paratethys aligns perfectly with our
572 abovementioned stable isotope data. The deep freshwater input could be explained by the early evolution of the Paratethys
573 (e.g., Schulz et al., 2005). Early evolution of the Paratethys was mainly controlled by narrowing seaways connecting it to the
574 Tethys Ocean which led to ingressions of cold boreal water from the north (through Polish straits), initially as undercurrents
575 (generating an anti-estuarine circulation) into the Eastern Paratethys first and then to Central and Western Paratethys (e. g.,
576 Schulz et al., 2005; Soták, 2010). In parallel to the $\delta^{18}\text{O}$ values, a divergent trend could be observed between the benthic and
577 planktic $\delta^{13}\text{C}$ values just after the EOB where benthic $\delta^{13}\text{C}$ declines largely whereas the planktic $\delta^{13}\text{C}$ shows an increase (Figure
578 6). In addition to the boreal fresh-water ingression, enhanced organic matter production due to increased nutrient input and
579 then the subsequent decomposition in the isolated, stratified Paratethys waters might have released light carbon into the bottom

580 waters. Increased $\delta^{13}\text{C}$ in planktic foraminifera could have resulted from elevated primary productivity driven by increased
581 nutrient input, which preferentially removes isotopically light carbon from surface waters during photosynthesis, leaving the
582 remaining carbon pool enriched in heavier carbon.

583

584 Initial boreal freshwater input by undercurrents then changes into ingressions of freshwater runoff as overflowing currents and
585 diluting the former Paratethyan sea water (i.e., a change into estuarine circulation) (e.g., Schulz et al., 2005; Soták, 2010). This
586 later ingression of overflowing freshwater runoff is likely due to an enhanced precipitation and represented as declining
587 planktic foraminifera $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values at ca. 15m (Figure 6). The $\delta^{18}\text{O}$ benthic foraminifera shows an increase during this
588 time which is likely related to the cooling of the bottom waters where as $\delta^{13}\text{C}$ of benthic foraminifera shows a positive peak
589 suggesting an enhanced organic carbon burial. The ingressions of freshwater runoff as overflowing currents likely formed a
590 freshwater surface layer reducing ventilation of bottom waters. The formation of the freshwater surface layer and subsequent
591 restricted mixing could have led pronounced stratification in the water column with isotopically lighter freshwater dominating
592 the surface waters which is evidenced by more depleted $\delta^{18}\text{O}$ planktic foraminifera values (between 15 m and the top of the
593 section) (Figure 6). The subsequent stratification in the water column would have exacerbated the buildup of oxygen-depleted
594 conditions and the isotopically depleted carbon pool at the bottom. This would also favor sulfate reduction by microbial
595 processes which produce further isotopically light carbon and reduce $\delta^{13}\text{C}$ values in an euxinic benthic environment.
596 Consequently, stratification and reduced oxygenation must have enhanced the preservation of organic matter in bottom
597 sediments.

598

599 The decrease in $\delta^{18}\text{O}$ of benthic foraminifera at ca. 22 m is possibly related to the slight warming recorded in the North Sea
600 (Śliwińska et al., 2019b) in paleomagnetic Subchron C13n above the EOIS (Figure 6). The increase in $\delta^{18}\text{O}$ planktic
601 foraminifera after ca. 25 m up to the ca. 40-41 m is likely related with further cooling during the EOGM. A sharp declining
602 trend could be noticed for benthic foraminifera $\delta^{13}\text{C}$ during the onset of this interval (at ca. 25 m) suggesting a relatively less
603 organic carbon burial. This was due to a decrease in primary productivity at the surface represented by lowering $\delta^{13}\text{C}$ of
604 planktic foraminifera and lowered terrestrial input.

605

606 The relative sea-level in the Paratethys starts to lower after the ca. 40-41 m level (Figure 3) which is followed by another
607 freshwater input likely due to enhanced precipitation at level ca. 45 m. This is evidenced by the depleted $\delta^{18}\text{O}$ planktic
608 foraminifera values and an increase in terrestrial palynomorphs (Figures 3, 6). Once again this was followed by an increase in
609 organic carbon burial represented by a peak in benthic $\delta^{13}\text{C}$ values at ca. 47 m.

610

611 During the Oi-1a cooling (between section levels ca. 45 m and 75 m) a long-term declining trend in benthic $\delta^{13}\text{C}$ is distinctive
612 and suggests a decrease in organic carbon burial in the Karaburun area (Figure 6). Decreasing relative sea-level and related
613 increase in bottom current velocities and wave action combined with a decrease in fresh water input should have likely

614 decreased the organic carbon burial suggested by decreasing benthic $\delta^{13}\text{C}$ at ca. 52 -53 m. Another freshwater ingress as
615 surface runoff could be seen at level ca. 56 m represented by a sharp decrease in planktic $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$. It appears that the
616 global Oi-1a cooling signal dominates the upper part of the section between levels ca. 60 m and 75 m. At ca. 63 m a decrease
617 in planktic $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values corresponds to surface freshwater input accompanied by increased organic carbon burial (peak
618 in benthic $\delta^{13}\text{C}$ values). A sharp decline in both benthic and planktic $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values suggest a significant freshwater
619 influx at level ca. 65 m. The uppermost peak in benthic $\delta^{13}\text{C}$ values at ca. 70 m represents an enhanced organic carbon burial
620 in the Karaburun area due to more favorable conditions for organic carbon sequestration provided by a relative sea level rise
621 (e.g., lower bottom current velocities and lower wave action).

622

623 The divergent trend between benthic and planktic $\delta^{13}\text{C}$ values indicates a highly stratified Paratethys Sea from time to time
624 and different surface and bottom carbon cycling processes after a change from anti-estuarine to estuarine circulation during
625 the EOT. These changes reflect both regional hydrological and basin reconfiguration (restriction) controls and global climatic
626 and eustatic shifts within the EOT. The global cooling during the EOT and Oi-1 must have amplified the stratification, reduced
627 ventilation and triggered local environmental shifts in the semi-enclosed Paratethys Basin. These environmental shifts provided
628 favorable conditions for the deposition of organic-rich fine-grained sediments with high Total Organic Carbon (TOC) values.

629

630 Overall, this contrasting pattern between the isotopic values of benthic and planktic foraminifera highlights the complex
631 interplay of global climate trends (Antarctic glaciation and global cooling) and regional factors (basin isolation and
632 hydrological changes) during the EOT. Our findings align with the proposed isolation of the Paratethys, driven by the
633 prolonged African-Arabian–Eurasian collision coupled with eustatic sea-level decline at the EOB and the cooling during the
634 EOT. In turn this led to the development of a distinct Paratethyan domain marked by mesophilic, humid climatic conditions
635 and intensified runoff (Popov et al., 2002). Moreover, the reconstructed relative sea-level evolution closely corresponds to other
636 reconstructions of relative sea-level changes from the late Eocene to early Oligocene period (e.g., from the Neo-Tethys,
637 Brinkhuis, 1994; North Sea, Jarve et al., 2015), further reinforcing the presence of a global climatic signal, as well.

638 **6.5 Boreal water in the Paratethys during the early Oligocene and its paleoceanographic significance**

639 The prominent negative shift in $\delta^{18}\text{O}$ of benthic foraminifera likely representing the boreal water ingress recorded in the
640 Karaburun composite section shortly after the EOB (ca. 33.7 Ma) appears to be widespread across the entire Paratethys Basin
641 (Figure 7; Soták, 2010; Ozsvárt et al., 2016; Gavrillov et al., 2017; van der Boon et al., 2019). In the Eastern Paratethys, this
642 shift appears to have occurred abruptly. A similar negative shift in $\delta^{18}\text{O}$ of bulk carbonates is also recorded further west in the
643 West Alpine Foreland Basin (Chalufy section, Soutter et al., 2022), which connected the Western Paratethys to the
644 Mediterranean Tethys Ocean (Figures 7, 8). Hence, boreal water ingress into the Paratethys, beginning in the Eastern
645 Paratethys and progressively reaching the Central and Western regions, is clearly represented by this negative shift across the
646 entire Paratethys Basin just after the EOB. This process was likely linked to the restriction or closure of the Arctic-Atlantic

647 gateway and the onset of anti-estuarine circulation between the Atlantic and the Nordic Seas during the EOT. Proxy records
648 indicate that the Eocene Arctic Ocean was significantly fresher than today, with salinities ranging from 20 to 25 psu and
649 occasional drops below 10 psu (Brinkhuis et al., 2006; Kim et al., 2014; Waddell and Moore, 2008). The Arctic freshwater
650 outflow into the North Atlantic may have inhibited deep-water formation during the Eocene (Baatsen et al., 2020; Hutchinson
651 et al., 2018). Sea-level and paleo-shoreline reconstructions in the Nordic Seas support the hypothesis that the Arctic became
652 isolated during the latest Eocene to early Oligocene due to the closure of the Arctic-Atlantic gateway (Hegewald and Jokat,
653 2013; O'Regan et al., 2011; Hutchinson et al., 2019). Recent evidence suggests that the deepening of the Greenland–Scotland
654 Ridge (GSR) around the EOT (just before the EOIS, ca. 33.7 Ma) facilitated increased exchange between the Atlantic and the
655 Nordic Seas, enabling the formation of anti-estuarine circulation and the salinization of North Atlantic surface waters (Abelson
656 and Erez, 2017; Stärz et al., 2017). Consequently, the gradual restriction of Arctic-Atlantic connectivity, followed by the onset
657 of anti-estuarine circulation driven by the deepening of the GSR, may have played a critical role in developing a robust AMOC
658 (e.g., Coxall et al., 2018; Hutchinson et al., 2019). Together, the Atlantic-Arctic closure and onset of anti-estuarine circulation
659 events could have triggered or intensified the Atlantic Meridional Overturning Circulation (AMOC) (e.g., Abelson and Erez,
660 2017; Coxall et al., 2018; Hutchinson et al., 2019). The onset of Nordic anti-estuarine circulation around the EOT might have
661 likely influenced salinity gradients and circulation in connected basins like the North Sea and Paratethys, contributing to the
662 freshening of the latter (Figure 8). Our data indicate the onset of anti-estuarine circulation in the Paratethys around 33.7 Ma,
663 coinciding with the development of similar circulation patterns between the Nordic Seas and the North Atlantic (Abelson and
664 Erez, 2017). We propose a hypothetical pathway for this circulation: warm surface water entering from the North Sea would
665 have flowed into the eastern Nordic Seas, joining the warm surface waters from the North Atlantic. After losing heat, this
666 water likely sank in the northern Nordic basin, flowed southward as deep water, and eventually reached the North Atlantic,
667 the North Sea, the Paratethys, and the Mediterranean Tethys, forming an interhemispheric northern-sourced circulation cell
668 (Figure 8a).

669

670 The onset of anti-estuarine circulation in the Paratethys was likely related to the deepening of the Greenland-Scotland Ridge
671 (GSR), similar to the development of anti-estuarine circulation in the Nordic Seas. During most of the Paleogene, the GSR
672 was shallower, restricting deep-water exchange between the North Atlantic and the Nordic Seas. A key feature of the modern
673 AMOC is the formation of North Atlantic Deep Water, which flows over the GSR from the Nordic Seas. The initiation of
674 Nordic anti-estuarine circulation events around the EOT likely enhanced deep-water formation, strengthening the Atlantic
675 Meridional Overturning Circulation (AMOC) and establishing an interhemispheric northern-sourced circulation cell (e.g.,
676 Abelson and Erez, 2017; Coxall et al., 2018; Hutchinson et al., 2019). This suggests that by ~33.7 Ma, the Paratethys, with its
677 anti-estuarine circulation, was integrated into this larger circulation system, contributing to global ocean circulation. Shortly
678 after, subsequent geographic restrictions and hydrological changes during the EOT changed this anti-estuarine circulation to
679 an estuarine circulation (Figure 8b). The closure of the Mediterranean Seaway—along with the narrowing of other seaways
680 likely caused by the major sea-level fall during the EOIS and EOGM—and increased freshwater influxes from the continent

681 diluted the surface Tethyan waters, disrupted the anti-estuarine circulation pattern, and led to brackish surface salinities in the
682 Paratethys (e.g., Schulz et al., 2005; Soták, 2010).

683

684 Further evidence supporting boreal water ingress and circulation through the Nordic Seas, North Sea and Paratethys comes
685 from the distribution of *Svalbardella cooksoniae* in the Greenland, Norwegian, and North Seas, as well as in the Eastern
686 Paratethys during the Oi-1a cooling event (Śliwińska and Heilmann-Clausen, 2011). This suggests a possible migration
687 pathway for *Svalbardella* spp. The presence of *Svalbardella* in the Massicore and Monte Cagnero sections of central Italy
688 (Brinkhuis and Biffi, 1993; van Mourik and Brinkhuis, 2005) aligns with our interpretation of boreal water circulation
689 extending to the Mediterranean Tethys through the Paratethys (Figure 8). Sinking cold boreal freshwater likely propagated
690 through interconnected Nordic marine basins, reaching the Paratethys and eventually the Mediterranean Tethys (Figure 8).
691 The invasion of the Mediterranean Tethys by higher-latitude taxa around the EOB (Brinkhuis and Biffi, 1993) further supports
692 this circulation pathway. The incursion of boreal bottom waters into the Mediterranean Tethys during the EOT is corroborated
693 by a marked transient increase in deep-water ostracod *Krithe* and a decline in deep-water ostracod diversity in the Massignano
694 composite section (Slotnick and Schellenberg, 2013). The *Krithe* pulse and the subtle change in the deep-water ostracod fauna
695 reflects intensified thermohaline flow of cooler deep waters, likely linked to boreal freshwater circulation through the
696 Paratethys (e.g., Dall’Antonia et al., 2003; Slotnick and Schellenberg, 2013). Variations in seafloor ventilation and productivity
697 due to changes in paleoceanographic conditions of the Tethys during the late Eocene - early Oligocene described in previous
698 studies (Jovane et al., 2007 and references therein) are also likely related to the circulation of the boreal water.

699

700 The influence of boreal waters on the Mediterranean Tethys might also explain the absence of well-defined positive $\delta^{18}\text{O}$ and
701 $\delta^{13}\text{C}$ shifts typically characterizing the EOT in Italian sections, including the GSSP for the EOB (e.g., Houben et al., 2012).
702 Instead, these sections display distinct negative shifts in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ bulk carbonate values around the EOB (e.g., Jovane et
703 al., 2007; Brown et al., 2009; Jaramillo-Vogel et al., 2013), resembling the Paratethys records and likely reflecting deep boreal
704 freshwater incursion through the Paratethys.

705 **7 Conclusion**

706 This study provides a comprehensive examination of the EOT and early Oligocene cooling in the eastern Paratethys, with a
707 focus on the Karaburun composite section. By integrating high-resolution biostratigraphy, geochemistry, sequence
708 stratigraphy, and precise geochronology, we constructed a robust chronostratigraphic framework spanning the latest Eocene
709 to early Oligocene. Our findings reveal that the isotopic shifts in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ at Karaburun site align closely with global
710 records, underscoring the influence of global climatic drivers during this critical transition. However, significant regional
711 deviations, such as depleted $\delta^{18}\text{O}$ values and pronounced stratification, highlight the impact of regional hydrological changes
712 and basin restriction of the Paratethys. These results emphasize the dual influence of global icehouse dynamics and regional

hydrological processes on the Paratethys during the EOT. The identification of key cooling events, such as the *Glaphyrocysta semitecta* influx for the EOB, and the *Svalbardella* event and its alignment with the Oi-1a glaciation, further enhances the utility of the Karaburun section for refining regional and global stratigraphic correlations. Additionally, the observed sequence stratigraphic patterns illustrate the interplay between eustatic sea-level changes and regional depositional dynamics during this interval. The abrupt depleted values in $\delta^{18}\text{O}$ values in benthic foraminifera just after the EOB (ca. 33.7 Ma) in the Paratethys Basin is attributed to boreal water ingression, driven by the closure of the Arctic-Atlantic gateway and the onset of anti-estuarine circulation between the Nordic seas and Atlantic during the EOT. This event, which funneled low-salinity boreal water through the Nordic and North Seas into the Paratethys, aligns with the distribution of boreal taxa like *Svalbardella cooksoniae* and extends as far as the Mediterranean Tethys, explaining isotopic anomalies in these regions. We have to note that the large-scale oceanic circulation scenario proposed here requires the movement of deep waters through narrow and shallow seaways surrounding the Paratethys, such as the Polish Strait, the Alpine Seaway, and smaller Neo-Tethys–Paratethys gateways. However, the timing of their closure and their paleobathymetry during the early Oligocene remain uncertain, as existing paleogeographic maps for this period are inconsistent (e.g. Barrier et al., 2018; Palcu & Krijgsman, 2023; Straume et al., 2024). Therefore, high-resolution paleogeographic reconstructions, integrated with ocean circulation modelling (e.g., Vahlenkamp et al., 2018), are essential for testing and validating this hypothesis. Overall, the Karaburun section emerges as a critical archive for studying the EOT in epicontinental seas. Our findings contribute to a deeper understanding of how global climatic transitions manifest in marginal marine settings and highlight the potential for further high-resolution studies to refine our knowledge of early icehouse climate evolution in the Paratethys region and beyond.

731

732 **Author contribution**

733 MYK designed the study. MYK and SGA conducted fieldwork. TV prepared samples for geochemical analyses. DN
734 performed geochemical analyses. AL conducted geochronological analysis. HB performed the marine palynological analyses.
735 CF analyzed the calcareous nannofossils. MYK wrote the paper with input from all authors. All authors analyzed and discussed
736 the data.

737

738 **Competing interests**

739 The authors declare that they have no conflict of interest.

740

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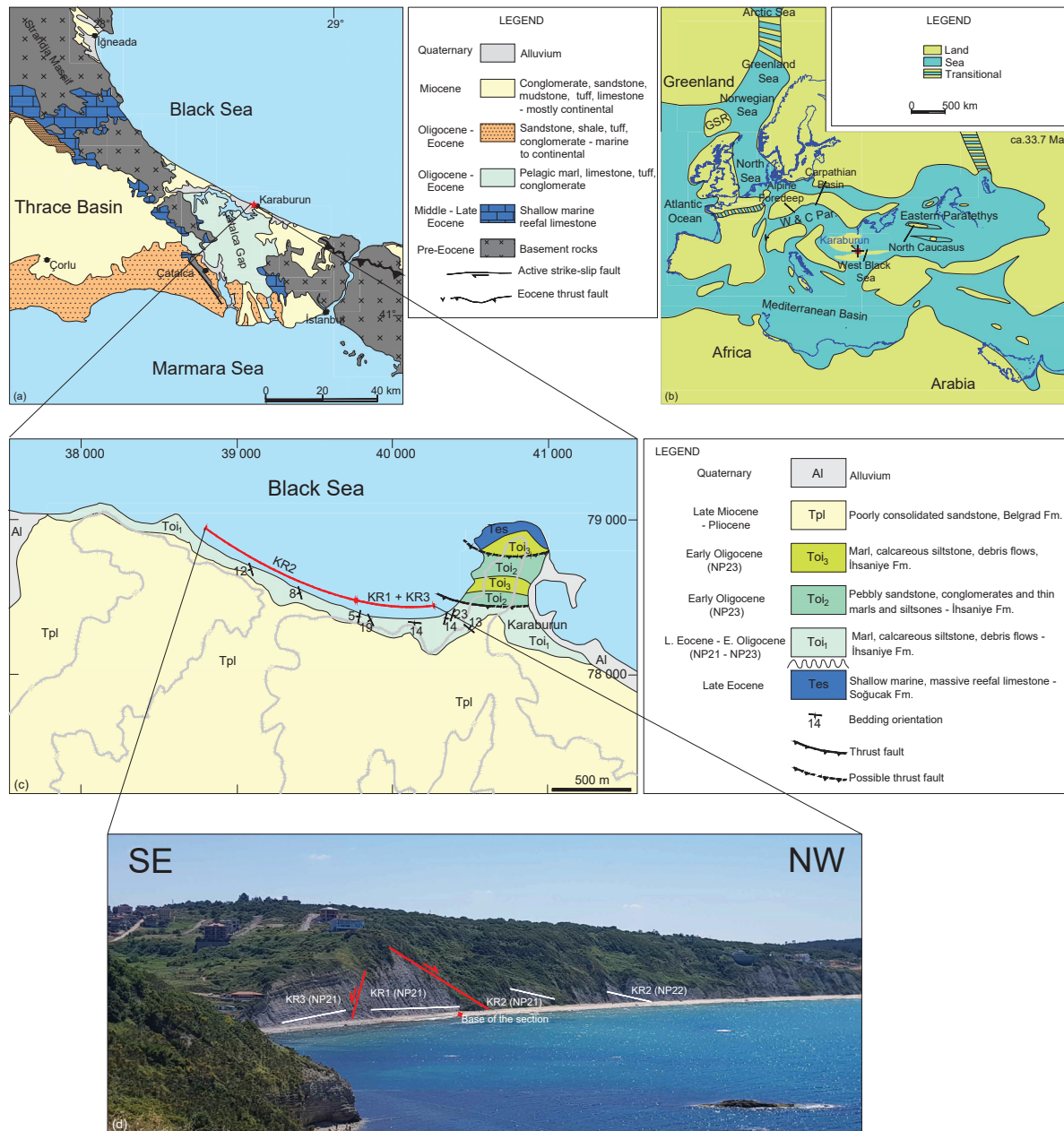


Figure 1 a) Geological map of the Thrace region (Türkiye) showing the location of the Karaburun area in relation to Thrace Basin, Strandja Massif and Çatalca Gap (modified from Okay et al., 2020). b) Paleogeography of the Paratethys during the early Oligocene (Rupelian, 33.7 Ma) (modified from Sachsenhofer et al., 2018) GSR: Greenland Scotland Ridge. W & C Par.: Western & Central Paratethys. Red dot marks the Karaburun area. c) Geological map of the Karaburun area showing the locations of the sub-sections KR1, KR2 and KR3 (revised from Okay et al., 2019). d) View of the studied sub-sections facing south from Cape Karaburun.

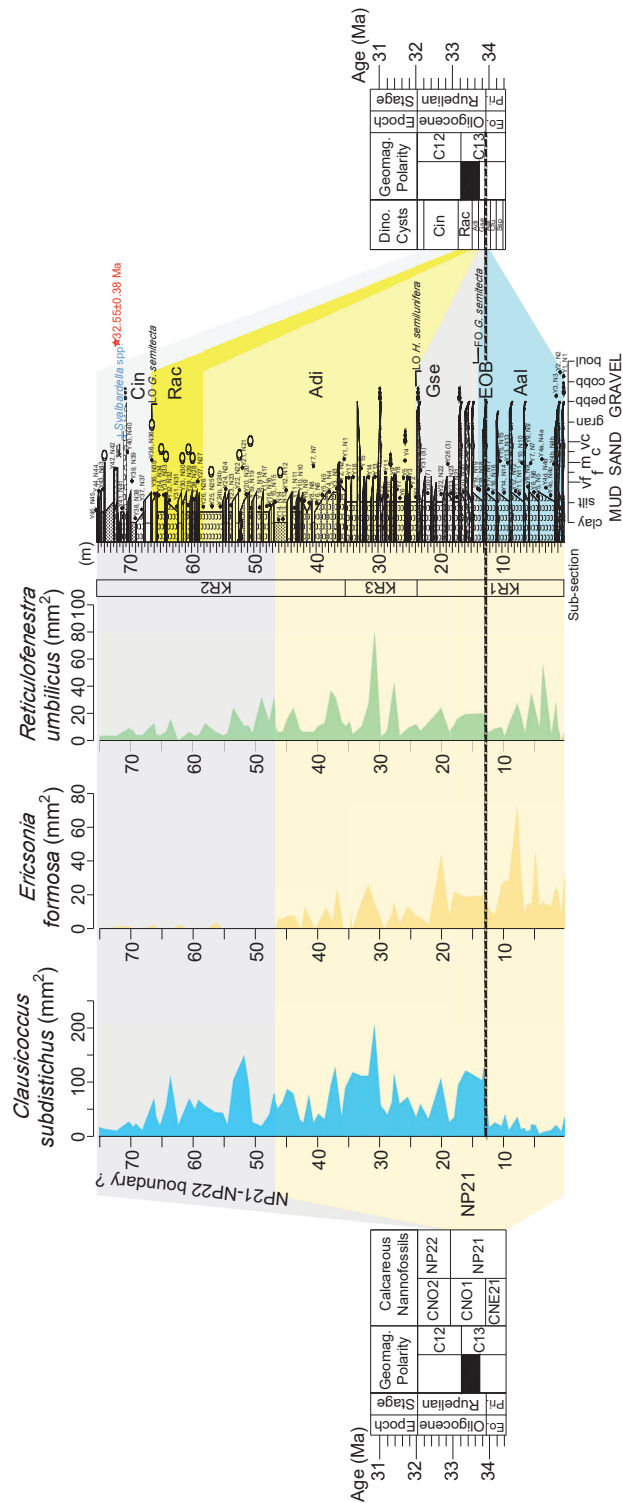


Figure 2 Stratigraphic log (in meters) of the Karaburun composite section with relative abundances of marker calcareous nannofossils. KR1, KR2, and KR3 are the abbreviations for the studied sub-sections from the Karaburun area. The biostratigraphic (calcareous nannofossil and dinoflagellate cyst) correlations to the geological time scale (Gradstein et al., 2012) are indicated by colored shading. The red star on the log shows the level of the tuff layer (32.65±0.38 Ma) while the blue cross indicates the level of sample with cold-water dinoflagellate *Svalbardella cooksoniae*, indicating a cooling event occurred during the early part of Chron C12r, near the NP21/NP22 boundary.

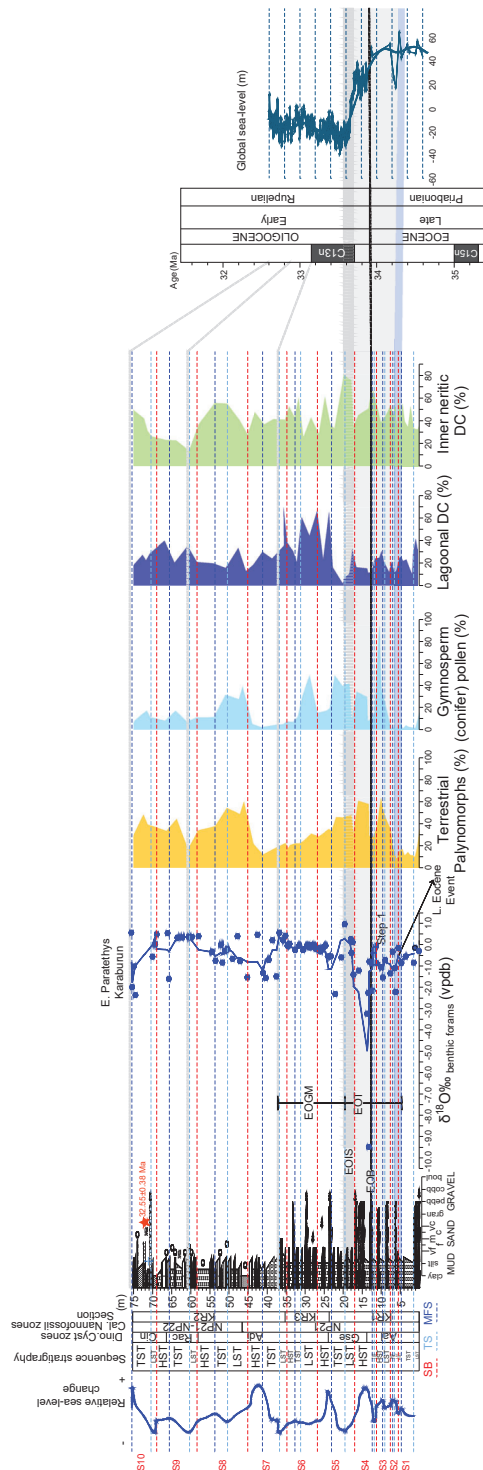


Figure 3 Stratigraphic log (in meters) of the Karaburun composite section showing sequence stratigraphic interpretations and reconstructed relative sea-level. The sequence stratigraphic interpretation and the reconstructed relative sea-level changes are based on the analysis of benthic foraminifera $\delta^{18}\text{O}$ values, along with the abundance of terrestrial palynomorphs and lagoonal and inner neritic dinocysts. Red star shows the level of the tuff layer on the log while blue cross indicates the level of sample with cold-water dinocyst *Svalbardella cooksoniae*. Correlation to the reconstructed global sea-level curve (Miller et al., 2020) and to the geological time scale (Gradstein et al., 2012) could be seen on the right side. S: Sequence, SB: Sequence boundary, TS: Transgressive surface, MFS: Maximum flooding surface, LST: Lowstand systems tract, TST: Transgressive Systems Tract. HTS: Highstand Systems Tract, EOB: Eocene Oligocene Boundary, EOT: Eocene Oligocene Transition, EOIS: Earliest Oligocene oxygen isotope step, EOGM: Early Oligocene glacial maximum.

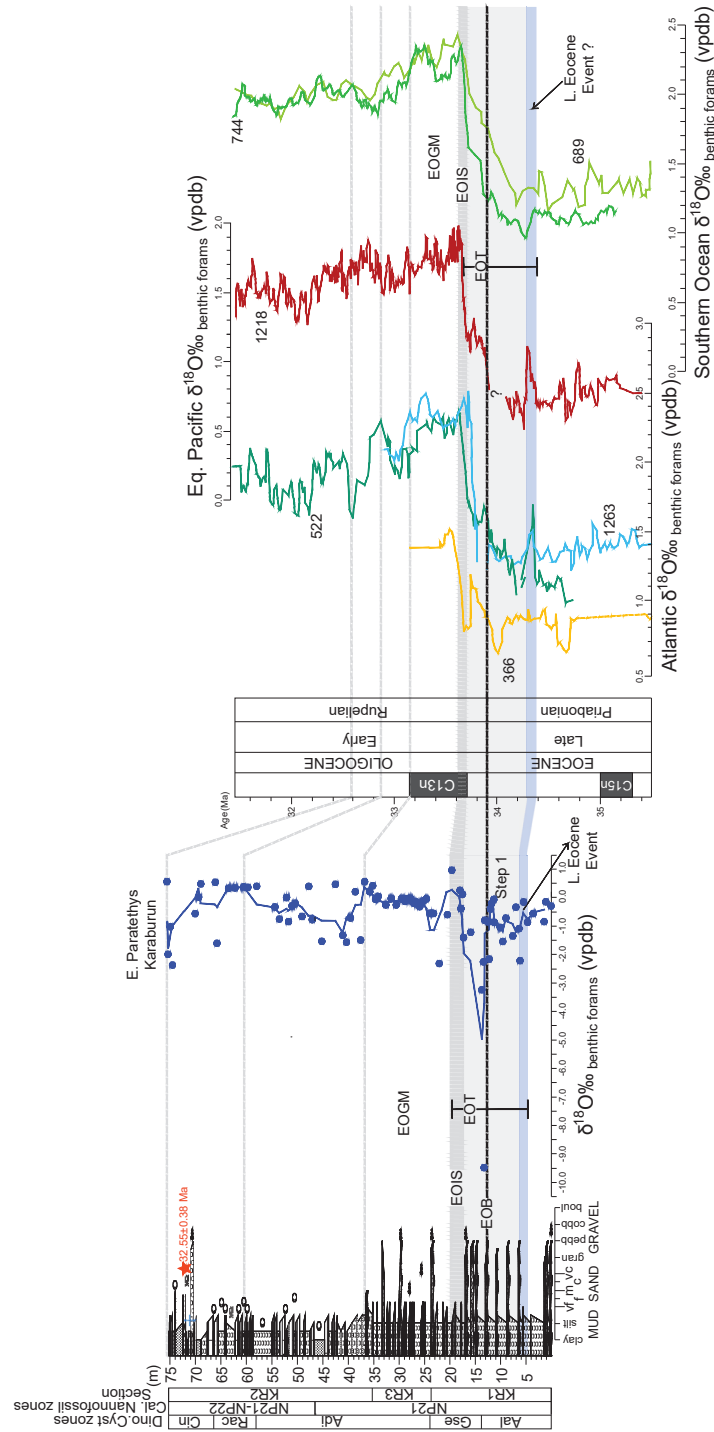


Figure 4 The stratigraphic log (in meters) of the Karaburun composite section including the results of benthic foraminifera $\delta^{18}\text{O}$ record (blue dots and line showing three-point running mean), highlighting the chronostratigraphic characteristics of the Eocene-Oligocene Transition (EOT). Apparent correlations were established by aligning the Karaburun data with high-resolution deep sea benthic foraminifera $\delta^{18}\text{O}$ records from the South Atlantic sites 522 (Zachos et al., 1996) and 1263 (Langton et al., 2016), and compared to the tropical Atlantic site 366 (Langton et al., 2016), the Southern Ocean sites 744 and 689 (Zachos et al., 1996; Diester-Haass and Zahn, 1996), and the equatorial Pacific site 1218 (Coxall and Wilson, 2011). Key features in the $\delta^{18}\text{O}$ records, such as positive and negative shifts and their amplitudes, were used to define EOT characteristics, including the Late Eocene Event, the Earliest Oligocene Oxygen Isotope Step (EOIS) and the Early Oligocene Glacial Maximum (EOGM). The Eocene-Oligocene Boundary (EOB) was identified through biostratigraphic analyses (see section 5.1). The red star on the log marks the tuff layer, while the blue cross indicates the occurrence of the cold-water dinocyst *Svalbardella cooksoniae*.

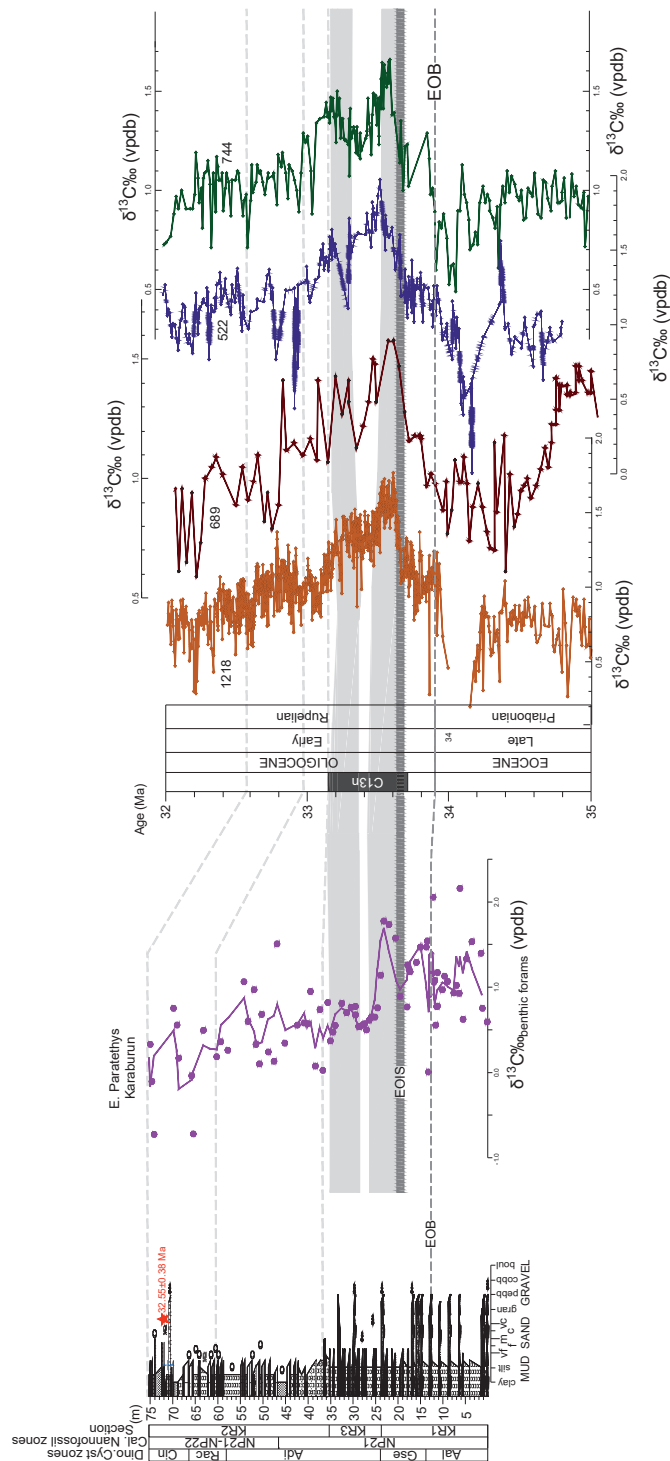


Figure 5 Benthic foraminifera $\delta^{13}\text{C}$ record (pink dots and line showing three-point running mean) of the Karaburun composite section along the stratigraphic log (in meters) highlighting the chronostratigraphic features of the Eocene-Oligocene Transition (EOT). Correlations were made with deep marine records from Site 689 in the sub-Antarctic Atlantic (Diester-Haass and Zahn, 1996), Site 1218 in the equatorial Pacific (Coxall and Wilson, 2011), Site 522 in the South Atlantic (Zachos et al., 1996), and Site 744 in the Southern Ocean (Zachos et al., 1996). Alignment of high-resolution benthic foraminifera $\delta^{13}\text{C}$ records from these sites with the Karaburun data revealed corresponding double peak isotope feature (gray shading) after the Earliest Oligocene Oxygen Isotope Step (EOIS) within the Early Oligocene Glacial Maximum (EOGM). The Eocene-Oligocene Boundary (EOB) was determined through biostratigraphic analysis (see section 5.1). The red star marks the tuff layer, while the blue cross indicates the level of sample containing the cold-water dinocyst *Svalbardella cooksoniae*.

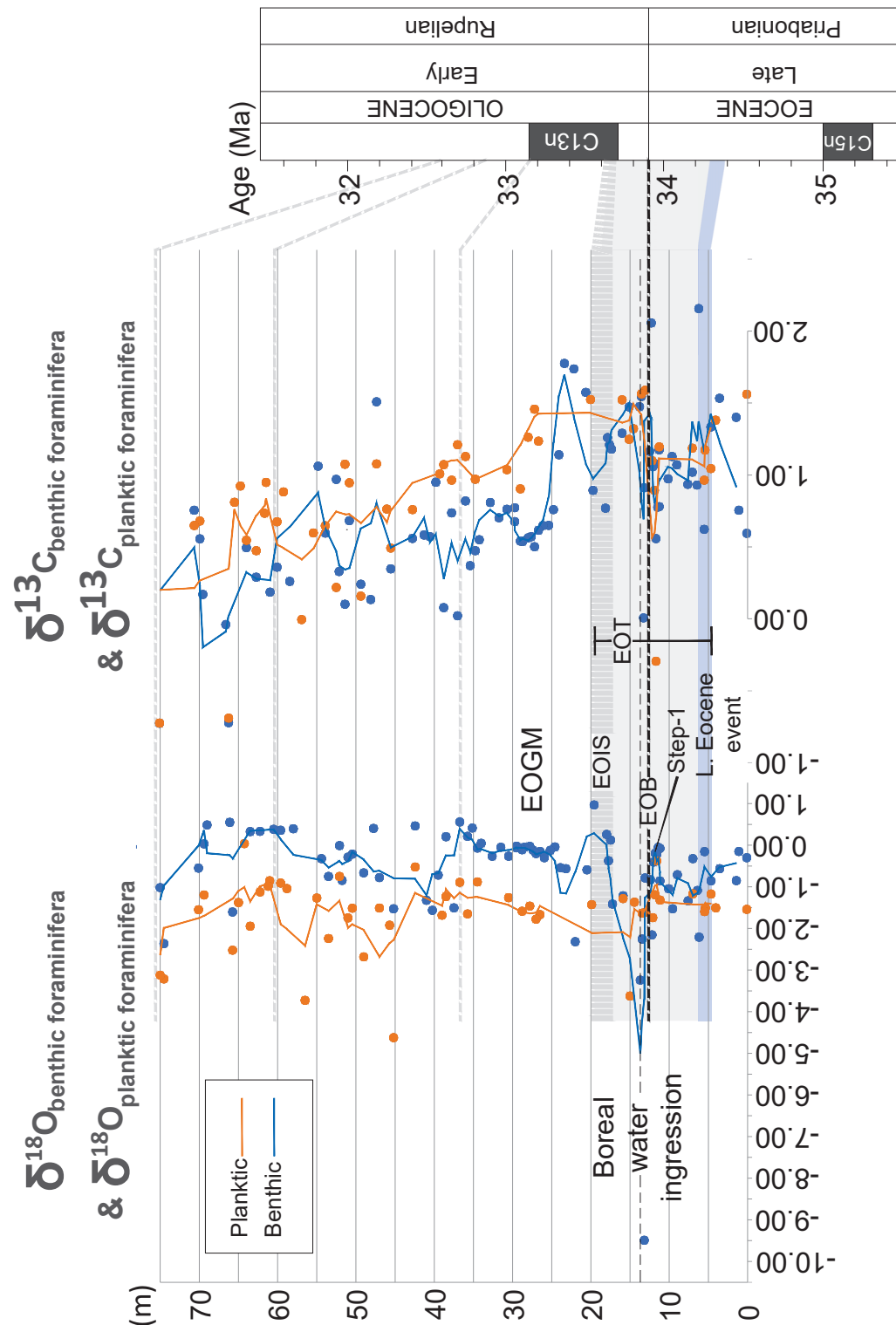


Figure 6 Oxygen ($\delta^{18}\text{O}$) and Carbon ($\delta^{13}\text{C}$) stable isotope values for the benthic and planktic foraminifera from the Karaburun composite section. The black dashed line represents the level for the boreal water ingress just after the Eocene-Oligocene Boundary (EOB). EOT: Eocene-Oligocene Transition. EOIS: Earliest Oligocene Oxygen Isotope Step. EOGM: Early Oligocene Glacial Maximum.

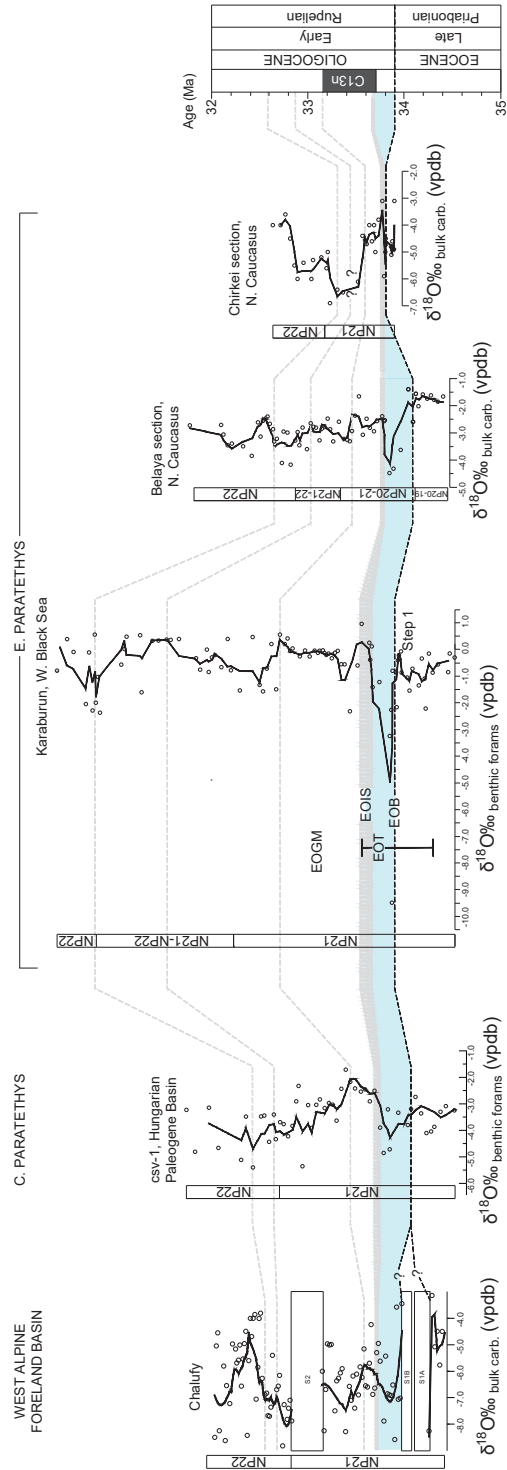


Figure 7 Oxygen stable isotope ($\delta^{18}\text{O}$) values and the characteristic negative $\delta^{18}\text{O}$ shift during the EOT observed in various Paratethys marine records. Correlations highlight the interval (light blue shading) between the Eocene-Oligocene Boundary (EOB) and the Earliest Oligocene Oxygen Isotope Step (EOIS, gray dashed line) across various sections, including the Chalufy Section in the West Alpine Foreland Basin, which links the Western Paratethys to the Mediterranean Tethys (Soutter et al., 2022); the csv-1 core from the Hungarian Paleogene Basin (Ozsvárt et al., 2016); the Karaburun Section (this study); and the Belaya (van der Boon et al., 2019) and Chirkei (Gavrilov et al., 2017) sections from the Northern Caucasus. To account for varying sample resolutions, a three-point running mean filter was applied uniformly to all sites.

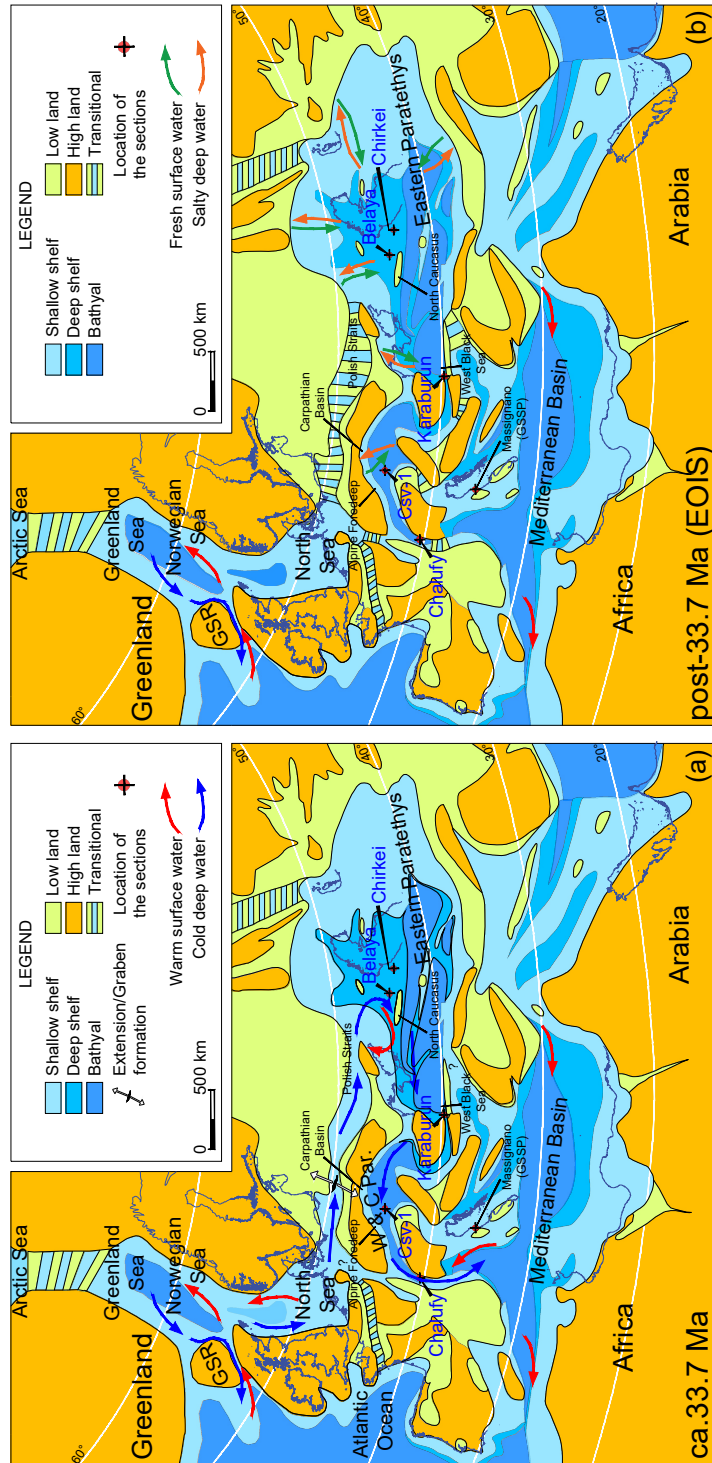


Figure 8 Paleogeographic maps at 33.7 Ma and post-33.7 Ma during the EOIS (The Earliest Oligocene Oxygen Isotope Step) (modified from Sachsenhofer et al., 2018). (a) boreal water anti-estuarine circulation at ca. 33.7 Ma through the Paratethys, extending into the Mediterranean Tethys. GSR: Greenland Scotland Ridge. (b) estuarine-circulation in the Paratethys during the EOIS due to the geographic restrictions (closure of the Mediterranean and other seaways due to the major sea-level fall) and hydrological changes (increased fresh water influx from land).