

# 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 localizedregional effects, such as freshwater influx and basin restriction,  
22 specific 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<sub>2</sub> 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, EOOGM of Hutchinson et al., 2020) and  
43 (2) Oi-1a, corresponding to early part of [magnetic Chronpaleomagnetic Subchron](#) C12r, took place during the early Rupelian  
44 (Pekar et al., 2002 and references therein). The EOOGM, spanning 490 kyr from ~33.65 to 33.16 Ma, marks a sustained period  
45 of cold climate and glaciation during [magnetic Chronpaleomagnetic Subchron](#) C13n (Hutchinson et al., 2020). The Oi-1a  
46 glaciation event corresponds to the appearance of the cold-water dinocyst taxa *Svalbardella cooksoniae* and/or *Svalbardella*  
47 spp. in the North Sea Basin (Śliwińska, 2019a and references therein). Episodes of southward migration of *Svalbardella*  
48 *cooksoniae* have consequently been interpreted as evidence of cooling events (e.g., Van Simaeys et al., 2005). *Svalbardella*  
49 *cooksoniae* has also been documented in a brief interval during the earliest Oligocene at numerous sites across the Northern  
50 Atlantic and Western 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 Bati,](#)  
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 [markerindex](#) species. Furthermore, it sheds light on how climatic

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

## 75 2 Geological Setting

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

83

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

95 **3 Regional Stratigraphy**

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

109 **4 Material & Methods**

110 **4.1 Lithostratigraphy**

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

117

118 The studied sediments predominantly consist of hemipelagic light gray marls, dark brownish clays, thin- to medium-bedded  
119 light gray, whitish, and beige carbonates, calcareous siltstones, and sandstones, which occasionally display planar lamination.  
120 The sequence also includes submarine fan (turbiditic) conglomerates, as well as debris flow and slump deposits toward the  
121 top. The hemipelagic fine-grained deposits contain rich microfossil assemblages of planktic and benthic foraminifera,  
122 calcareous nannofossils, and dinocysts, indicating a latest Eocene–early Oligocene age (this study). The submarine fan deposits  
123 are characterized by thin to medium thick, erosive-based conglomeratic beds with mainly carbonate pebbles, organic matter,  
124 and microfossil fragments (e.g., foraminifera and shell debris) and intercalated with thin silty layers. They often grade vertically  
125 into sandstone layers. Although these submarine fan deposits exhibit variable lateral thicknesses, they provide solid key

126 horizons for correlation of the subsections (e.g. correlation of the KR1 and KR3 subsections). Brownish organic-rich clay  
127 layers occasionally contain red-yellow nodules. Pyrite is commonly found in these organic-rich layers. Additionally, pyritized  
128 coral fossils are rarely observed in these organic-rich clay deposits. A distinctive white tuff layer at approximately 71.5 m  
129 within the composite section was sampled for U-Pb zircon dating (see section 4.5) (Figure 2). Debris flow horizons, with a  
130 maximum thickness of 5 meters, exhibit channel geometries and primarily consist of carbonate pebbles. These horizons  
131 increase in frequency toward the topmost 20 meters of the succession. This study focuses on the lower and middle portions of  
132 the section including the EOT, ending around the tuff layer (Figure 2), and does not include analysis of the uppermost part of  
133 the succession including the debris flow deposits.

#### 134 4.2 Sequence Stratigraphy

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

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

#### 147 4.3 Biostratigraphy

##### 148 4.3.1 Calcareous Nannofossilnannofossil

149 The study on Calcareous Nannofossilcalcareous nannofossil assemblages was carried out on 84 samples, prepared at the  
150 Department of Earth Sciences of the University of Milan (Italy), following the smear-slide standard technique described by  
151 Bown & Young (1998). Calcareous nannofossils were analyzed using an Axioscop Zeiss light microscope (LM) at 1250X  
152 magnification. Preservation of the specimens was generally good, as indicated by the presence of holococcoliths, coccospores  
153 and small coccoliths. Quantitative analysis was performed by counting 300 specimens per sample, in a variable number of  
154 fields of view, depending on the nannofossils total abundance. Nannofossil frequency data were converted into the number of  
155 specimens per square millimeter for the evaluation of the biostratigraphic signal, and into percentages to estimate the  
156 paleoecological significance of the assemblage variations. The position of biohorizons recognized in this study is based on

157 abundance patterns of index species, according to Agnini et al. (2014) and Viganò et al. (2023a) and are labelled as follows:  
158 Top (T): the highest occurrence of a taxon, Base common and continuous (Bc) and Top common and continuous (Tc): the  
159 lowest and highest common and continuous occurrence of a taxon. For calcareous nannofossils taxonomy, we refer to Perch-  
160 Nielsen (1985), Agnini et al. (2014), and the Nannotax web library (<https://www.mikrotax.org/Nannotax3>). The  
161 biostratigraphic schemes adopted here are those of Martini (1971) and Agnini et al. (2014).

#### 162 **4.3.2 Marine Palynology - Dinoflagellate cystsdinocysts**

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

173

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

#### 177 **4.4 Geochronology**

##### 178 **4.4.1 U-Pb dating**

179 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  
180 constraining the age of the deposits. Three kilograms of tuff material were crushed and zircon crystals were separated by  
181 standard heavy liquid techniques and mounted in epoxy resin. Thirty five zircon crystals were dated via U-Pb at the Envitop  
182 analytical facility at CEREGE using an Element XR ICP-MS connected to a NWR193 laser (ArF 193 nm) ablation system.  
183 Zircon crystals were ablated with a 25-micron spot diameter, a 15 Hz pulse repetition rate, an energy fluence of 1.5 J/cm<sup>2</sup>, and  
184 a carrier gas flow of 0.975 L/min. Data reduction, date, and date uncertainty calculations were conducted with an in-house  
185 MATLAB script. We applied rigorous filtering based on zircon morphology, U-Pb concordance, and common Pb content to  
186 ensure the reliability of the final age determination. The dataset was filtered for concordant grains using the Concordia distance  
187 method of Vermeesch (2021), which applies isometric logratios with a discordance threshold of 10 (SI units) and a reverse

188 [discordance threshold of 5 \(SI units\)](#). Details about our U-Pb dating workflow, and data reduction steps, and discordance filter  
189 are given in Licht et al. (2024). The three zircon validation reference materials used during these sessions yielded offsets  
190 around TIMS ages < 1% in most cases, and < 2% otherwise. Out of the 35 analyzed zircon crystals, 12 yield concordant U-Pb  
191 ages (see Supplementary Table S3). The final Concordia age was calculated with concordant ages only using IsoPlotR  
192 (Vermeesch, 2018).

## 193 4.5 Geochemistry

### 194 4.5.1 $\delta^{18}\text{O}$ & $\delta^{13}\text{C}$ analyses

195 Measurements of stable oxygen ( $\delta^{18}\text{O}$ ) and carbon ( $\delta^{13}\text{C}$ ) isotopes on benthic foraminiferal (*Cibicidoides* spp.) and planktic  
196 foraminifera (*Turborotalia ampliapertura*) test fragments were performed at GEOMAR, Kiel on a Thermo Scientific MAT  
197 253 mass spectrometer with an automated Kiel IV carbonate preparation device. The isotope values were calibrated versus the  
198 NBS 19 (National Bureau of Standards) carbonate standard and the in-house carbonate standard (“Standard Bremen”,  
199 Solnhofen limestone). Isotope values in delta-notation ( $\delta$ ) are reported in ‰ relative to the VPDB (Vienna PeeDee Belemnite)  
200 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  
201 were not done due to the low numbers of specimens found.

## 202 5 Results

### 203 5.1 Biostratigraphy

#### 204 5.1.1 Calcareous [Nannofossils](#)

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

211

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

215 The Base [of](#) common (Bc)-[of](#) *Clasicoccus subdistichus* group defines the onset of Zone CNO1 of Agnini et al. (2014), which  
216 corresponds to the upper part of Zone NP21 (Martini, 1971). The increase in abundance of this informal taxonomic group  
217 represents the most reliable nannofossil bioevent to approximate the EOB across different regions (e.g. Marino and Flores,

218 2002; Toffanin et al., 2013; Fioroni et al. 2015, Viganò et al., 2023a). In the studied area, this event is well-delineated, showing  
219 an abrupt increase in abundance from approximately 17 to over 100 specimens/mm<sup>2</sup> in sample N18, ca. 13 m from the base of  
220 the studied section (Figure 2).

221  
222 The top (T) of *Ericsonia formosa* marks the base of Zone NP22 of Martini (1971) and the base of Zone CNO2 (Agnini et al.,  
223 2014). However, in the studied composite section, the precise position of this bioevent remains uncertain due to the rarity and  
224 scattered occurrence of the taxon in its final range, nevertheless it is likely located before the end of the *C. subdistichus* acme  
225 (Figure 2). This Top of common (Tc) and continuous occurrence of the *C. subdistichus* group, was positioned before the top  
226 (T) of *E. formosa* by Agnini et al. (2014) in the biozonation adopted here. However, recent studies challenge this interpretation  
227 (Viganò et al., 2023a, 2024a). In fact, recent findings suggest that the top common and continuous (Tc) of this group occurs  
228 above the top (T) of *E. formosa*, as previously reported by Backman (1987) and Catanzariti et al. (1997). The top common and  
229 continuous (Tc) of *C. subdistichus* gr., occurring early in Subchron C12r, has also been documented above the top (T) of *E.*  
230 *formosa* in the Pacific (Toffanin et al., 2013; Viganò et al., 2023b), the Atlantic (Bordiga et al., 2015, Viganò 2023b) and the  
231 Indian Ocean (Fioroni et al., 2015; Villa et al., 2021, Viganò 2023a). These studies indicate that *C. subdistichus* persists and  
232 remains common even after the T of *E. formosa*. In our dataset, the abundance of *C. subdistichus* exhibits a marked decrease  
233 (Tc) in the upper 5 meters of the studied section (Figure 2). Consequently, the T of *E. formosa* (i.e. the boundary between  
234 NP21 and NP22) should be positioned at some point prior to this bioevent. This interpretation is supported by data from  
235 dinoflagellatesdinocysts (section 5.1.2), which provide a better and more precise constraint for the stratigraphic position of the  
236 upper part of the investigated section.

237 **5.1.2 Dinoflagellate cystsDinocysts**

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

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

247  
248 The dominant palynological groups throughout the succession are organic linings of benthic foraminifera, and notably their  
249 fragments, besides dinocysts, and pollen and spores of terrestrial higher plants (Table S1). We also recovered several other  
250 aquatic algal taxa and acritarchs. These include representatives of e.g., fresh to brackish water elements like *Cyclopsiella*,

251 *Pterospermella* and, *Tasmanites* spp., besides very small (<~20 µm) psilate and skolochorate cysts (viz, ‘acritarchs’, and other  
252 ‘small skolochorate cysts’) of unknown ecology. Fungal spores and fruitbodies, as well as scolecodonts are occasionally  
253 encountered as well. In terms of palynofacies (palynodebris) composition and trends, the samples are all very similar in  
254 displaying a rich mix in mainly terrestrial plant-derived elements of varying sizes, combined with various amorphous materials.  
255 Truly opaque material is conspicuously absent. No trends or breaks are apparent from this visual, non-quantitative assessment  
256 (Table S1). Overall, these results match the findings by Sancay and Bati (2020).

257

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

277

278 These correlations are here bolstered by the spot-occurrence of *Svalbardella cooksoniae* in sample Y41 at 71 m, assigned to  
279 the Cin Zone (Figure 2, Figure S5a, b, and c). This event was previously described from the central Italian EOT section, within  
280 the same Cin Zone (Brinkhuis and Biffi, 1993). At high northern latitudes, this species ranges from the late Eocene way into  
281 the Oligocene (e.g., Eldrett et al., 2004; Eldrett and Harding, 2009). Subsequent work noted that colder episodes during the  
282 Oligocene likely induced equatorward migration of such typical high-latitude taxa (e.g., van Simaeys et al., 2005). In effect,  
283 our finding reflects the earliest one of such migration pulses, an event well documented by Śliwińska and Heilmann-Clausen  
284 (2011). These authors showed that *Svalbardella cooksoniae* is consistently present in the same narrow interval calibrated to

285 the basal Subchron C12r, close to the NP21/NP22 boundary, in many high- and mid-latitude Northern Hemisphere sections,  
286 ranging from the Greenland Sea in the north to Italy in the south. Moreover, they correlated this event to the Oi-1a oxygen  
287 isotope maximum of Pekar and Miller (1996) and Pekar et al. (2002). Another interesting finding is specimens of the ~~bizarre~~  
288 acritarch *Ascostomocystis potane* in samples from the sub-section KR2. Documented from the basal Rupelian type section in  
289 Belgium (Stover and Hardenbol, 1993), this further confirms assignment to the basal Oligocene.

290 **5.2 U-Pb dating of the tuff**

291 Concordia plot of the dated tuff is available in Supplementary Figure S10. The tuff sample yields an early Rupelian  $32.55 \pm$   
292  $0.38$  Ma ( $2\sigma$ ) age, based on 11 out of the 12 dated zircons. The age of the tuff layer aligns with the biostratigraphic dating,  
293 which places the top of the KR composite section within the Cin dinoflagellate cysts zone (Figure 2).

294 **5.3 Sequence Stratigraphy**

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

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

309

310 The upper part of the section, including the Early Oligocene Cooling (see Section 6.3), consists of four sequences (S7, S8, S9,  
311 and S10). A notable difference in the thickness of the depositional sequences is observed between the lower section (up to ~11  
312 m) and the overlying sequences (from ~11 m to the top) (Figure 3). This variation in depositional thickness may could be  
313 linked related to orbital obliquity forcing across during the latest Eocene to early Oligocene, a process that has been identified  
314 in several terrestrial and marine records (e.g., Westerhold Abels et al., 20202011; Jovane et al., 2006; Boulila et al., 2021).  
315

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

323 **5.4 Geochemistry**

324 **5.4.1  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  isotope analyses**

325 The  $\delta^{18}\text{O}$  values of benthic foraminifera (*Cibicidoides* spp.) in the Karaburun composite section range from -9.5‰ to 1.0‰,  
326 displaying distinct temporal variations throughout the sequence. At the base of the section, at 11.40 m, just before the EOB,  
327 a small positive peak (1.0‰) is noticeable (Figure 4). Following this, a pronounced and abrupt negative shift from -0.1‰ to -  
328 9.5‰ is evident just after the Eocene-Oligocene Boundary (EOB). Following the abrupt negative shift, the  $\delta^{18}\text{O}$  values increase  
329 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  
330 as in other EOT records (e.g., Katz et al., 2008), however, with a negative shift in between. Following the second increase, a  
331 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  
332 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‰  
333 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  
334 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  
335 modest increase toward the top of the section (Figure 4).

336

337 The  $\delta^{13}\text{C}$  values of benthic foraminifera (*Cibicidoides* spp.) in the Karaburun composite section range from -0.7‰ to 2.2‰,  
338 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  
339 prominent shift towards more depleted levels (from 2.1‰ to 0‰) just after the EOB (Figure 5). The two-phase increase in  
340  $\delta^{13}\text{C}$  (from the EOB up to ca 24 m) is again evident, interrupted by a sharp decline (ca. 20 m) that corresponds precisely with  
341 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  
342 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‰  
343 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  
344 EOB, two more major positive carbon isotope excursions are observed at the base of the section: one between 1 and 7.5 m  
345 (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  
346 11.20 m), exhibit a significant positive shift of  $\geq 1.0\text{‰}$ . After the excursion between 19.60 m and 23 m, the  $\delta^{13}\text{C}$  values drop  
347 sharply (around 23–27 m) before exhibiting another series of positive carbon isotope excursions in the middle and upper parts

348 of the section (Figure 5). Like the excursions at the base of the section, the excursions in the middle and upper parts also  
349 display positive shifts of approximately 1‰, with the most pronounced reaching 1.5‰. This pattern reflects a dynamic carbon  
350 cycle with notable variations throughout the sequence.

351

352 Due to the intermittent presence of planktic foraminifera *Turborotalia ampliapertura* along the composite section, the  $\delta^{18}\text{O}$   
353 and  $\delta^{13}\text{C}$  records of planktic foraminifera exhibit some gaps (Figures 6 and S1). Nevertheless, the overall trends and shifts  
354 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  
355 shift at the base of the section at 11.40 m just before the EOB, although slightly smaller (1.2‰). Following a slight positive  
356 shift (0.4‰) just after the EOB at 14.45 m, the most prominent and significant positive shift recorded by the benthic  
357 foraminifera is not fully captured in the planktic foraminifera record due to the absence of *T. ampliapertura* in the 4 m interval  
358 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  
359 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.*  
360 *ampliapertura* follows the pronounced positive shift between 19.90 m and 26.5 m, obscuring the  $\delta^{18}\text{O}$  record for that interval.  
361 Following the barren interval, a gradual increase of 0.9‰ is clearly observed, extending up to 36.75 m. In the interval between  
362 36.75 m and 61 m, the  $\delta^{18}\text{O}$  record of planktic foraminifera fluctuates, exhibiting two significant alternating trends of decrease  
363 and increase. Following this fluctuation, a gradual decrease is observed at the top of the section, followed by a sharp decline.  
364

365 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  
366 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  
367 the infrequent presence of *T. ampliapertura*, the interval between 15.90 m and 26.5 m is not fully represented; however, this  
368 interval includes the highest  $\delta^{13}\text{C}$  values observed. After these peak values, a gradual decrease is noted towards the top of the  
369 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  
370 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  
371 noticeable.

## 372 5.5 Paleoenvironment & Paleoecology

### 373 5.5.1 Calcareous ~~Nannofossils~~nannofossils

374 Calcareous nannoplankton are highly sensitive to environmental changes in their surface water habitats, and fluctuations in  
375 nannofossil assemblages are interpreted as responses to shifts in sea surface temperature (SST), nutrient concentrations,  
376 salinity, and other environmental factors (e.g., Aubry, 1992; Winter et al., 1994) thereby reflecting palaeoceanographic  
377 perturbations. Numerous studies have explored the ecological tolerance of extinct taxa, establishing paleoecological  
378 preferences through biogeographic studies (e.g., Wei and Wise, 1990) and comparison with diverse environmental proxies

379 (e.g., Villa et al., 2014). We discuss the behavior of several taxa within the Karaburun assemblage, based on paleoecological  
380 affinities outlined in previous works.

381

382 *Reticulofenestra daviesii* and *Chiasmolithus* spp. are considered cool-water taxa (Wei et al., 1992; Villa et al., 2008) with  
383 preference for eutrophic conditions (Villa et al., 2014, Viganò et al., 2024b). This paleoecologic group is recorded with very  
384 low relative abundances, with few positive peaks in the middle part of the studied section. *Cyclicargolithus floridanus*, a typical  
385 eutrophic open-ocean species (Auer et al., 2014), occurs with abundances reaching up to 50% in the lower and upper part of  
386 the studied section, suggesting high productivity conditions (Aubry, 1992, Dunkley Jones et al., 2008, Villa et al., 2021) (Figure  
387 S3). Small reticulofenestrids constitute a significant component of the assemblages, reaching over 60% in the middle part of  
388 the section. They have been reported as dominant components of the nannoflora along continental margins (Haq, 1980). These  
389 settings are typically characterized by eutrophic conditions, driven by continental runoff and/or riverine input. Consequently,  
390 these small coccoliths are regarded as opportunistic taxa with broad ecological tolerance, yet particularly well-adapted to  
391 nutrient-rich environments (Aubry, 1992) and indicative of increased availability of terrigenous nutrient (Wade and Bown,  
392 2006).

393

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

399

400 Evidence of nutrient availability is further supported by the presence of braarudosphaerids, which are associated with coastal,  
401 low-salinity waters (Peleo-Alampay et al., 1999; Thierstein et al., 2004; Konno et al., 2007), eutrophic conditions (Cunha and  
402 Shimabukuro, 1997, Bartol et al., 2008) and the influx of terrigenous material (Švábenická, 1999). Braarudosphaerids are  
403 rarely found in the open ocean and thrive under unusual marine conditions, demonstrating a tolerance for environmentally  
404 stressed settings. Similar conditions are indicated by the presence of *Micrantholitus*, a taxon typically associated with shallow  
405 marine environments (Bown, 2005), reduced salinity, and eutrophic conditions (Street and Bown, 2005; Bown and Pearson  
406 2009). These pentoliths occur from the base of the investigated section, albeit at low percentages and with a discontinuous  
407 distribution, further suggesting eutrophication and reduced salinity (Figure S3).

408

409 The presence of Ascidian spicules, with their highest and continuous occurrence in the middle-upper part of the section, also  
410 points to a shallow-marine depositional setting (e.g. Varol, 2006; Ferreira et al., 2019) and high surface-water productivity  
411 (Toledo et al., 2007). Furthermore, the relatively common occurrence of holococcoliths (mainly *Lanternithus minutus* and

412 *Zigrablithus bijugatus*), *Pontosphaera* spp. and *Helicosphaera* spp., taxa prone to dissolution (Bown, 2005; Monechi et al.,  
413 2000) reinforces the interpretation of a shallow-water environment.

#### 414 **5.5.2 Marine Palynology – Dinoflagellate cystspalynology - dinocysts**

415 For the analysis of the marine palynological assemblages, emphasizing dinocysts, we rely on the taxonomical and ecological  
416 dinocyst groups derived from modern distributions (e.g., Zonneveld et al., 2013; Marret and De Vernal 2024) and empirically  
417 based paleoecological information on the Paleogene dinocysts following previous works (e.g., Brinkhuis, 1994; Pross and  
418 Brinkhuis, 2005; Sluijs et al., 2005; Frieling and Sluijs, 2018). However, as mentioned above, the assemblages are generally  
419 too poorly preserved to allow for detailed quantitative considerations. Yet, given the overall quantitative characteristics of the  
420 studied samples, viz, substantial terrestrial input, and consistent dominances of taxa empirically known from restricted marine  
421 to inner neritic (incl. lagoonal) like the goniomedoid-group of dinoflagellate cysts (in this case e.g., *Homotryblium*,  
422 *Polysphaeridium*, *Heteraulacocysta*, *Eocladopyxis* spp.), and the peridinioids (*Lentinia*, *Phthanoperidinium*, *Senegalinium*,  
423 and *Deflandrea* spp.), neritic to outer neritic (e.g., *Areoligera*, *Glaphyrocysta*, *Enneadocysta*, *Spiniferites* and *Operculodinium*  
424 spp.) combined with a small, but consistent contribution from offshore, oceanic taxa like *Impagidinium* and  
425 *Nematosphaeropsis* spp. points to an essentially open marine, offshore, hemipelagic setting, comparable to e.g., the central  
426 Italian sections (cf. Brinkhuis and Biffi, 1993).

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

## 434 **6 Discussion**

### 435 **6.1 Age control overview**

436 An initial age model for the KR composite section was constructed using tie points derived from nannofossil and dinocyst  
437 biozonations, combined with U-Pb dating of a tuff layer at 71.5 m (Figure 2). Additional age constraints were obtained by  
438 aligning the Karaburun benthic foraminifera  $\delta^{18}\text{O}$  data with the high-resolution benthic foraminifera  $\delta^{18}\text{O}$  record from the  
439 Atlantic sites 522 and 1263 (Figure 4). This alignment was achieved by identifying corresponding features in the isotope  
440 records, such as positive and negative shifts and their amplitudes. The carbon isotope record of benthic foraminifera provided  
441 independent validation of this tuning (Figure 5). The  $\delta^{13}\text{C}$  benthic foraminifera data from the Karaburun area was correlated  
442 and aligned with global high-resolution benthic foraminifera  $\delta^{13}\text{C}$  records from deep-sea sites, including 1218 (equatorial

443 Pacific), 689 (sub-Antarctic Atlantic), 522 (South Atlantic), and 744 (southern Indian Ocean). Similarly, the planktic  
444 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  
445 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  
446 hemipelagic sediment cores retrieved from the African margin of the Indian Ocean (Tanzania Drilling Project sites 12 and 17,  
447 Pearson et al., 2008). All ages were assigned following the integrated magneto-biostratigraphic GTS2012 timescale. Our  
448 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  
449 South Atlantic, Southern Ocean and Pacific are also present in the Paratethys, verifying that these signals are genuinely global  
450 and valuable for stratigraphic correlation.

451 According to the constructed age model, the base of the section dates to the latestlate Eocene (late Priabonian). The EOB is  
452 identified at 12.75 m. The middle and upper parts of the composite KR section correspond to the early Oligocene (early  
453 Rupelian) (Figure 2). The revised age constraints established in this study offer a robust chronostratigraphic framework for the  
454 latest Eocene to early Oligocene interval of the Karaburun section, surpassing the accuracy of prior studies (e.g., Less et al.,  
455 2011; Okay et al., 2019; Simmons et al., 2020). However, we have to note that the absence of the typically abundant index  
456 taxon *Areosphaeridium diktyoplakum* in all studied sub-sections currently hampers precise recognition of the Adi/Rac zonal  
457 boundary and may reflect local paleoenvironmental or taphonomic conditions, or alternatively, point to a slight stratigraphic  
458 gap or condensed interval, a possibility that merits further investigation.

## 459 6.2 The EOT

460 At the deep Atlantic Site 522 and Pacific Site 1218, the Latelate Eocene Event is marked by a transient interval of positive  
461  $\delta^{18}\text{O}$  values, reflecting a short-lived cooling or glacial episode (Hutchinson et al., 2021) (Figure 4). This isotopic shift measures  
462 approximately 0.6‰ and 0.4‰ at Site 522 and Site 1218, respectively. Similarly, the base of the KR composite section exhibits  
463 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  
464 Eocene Event (Figure 4). The onset of this event coincides with the extinction of *Discoaster saipanensis* at 34.44 Ma at Site  
465 1218. Based on calcareous nannofossil data (i.e., the absence of *D. saipanensis*), the base of the KR section is inferred to be  
466 younger than 34.44 Ma, supporting this correlation. The Late Eocene Event represented by this 0.7‰ positive shift in  $\delta^{18}\text{O}$   
467 values at approximately 5.5 m marks the onset of the EOT in the KR composite section (e.g., Hutchinson et al., 2020) (Figure  
468 4).

469 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  
470 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  
471 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  
472 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  
473 al., 2008). The onset of Step 1 is dated to 34.15 Ma, with an estimated duration of approximately 40 kyr (Hutchinson et al.,  
474 2020).

476

477 The Earliest Oligocene Oxygen Isotope Step (EOIS) represents a rapid  $\delta^{18}\text{O}$  increase (0.7‰ or more) occurring well after the  
 478 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,  
 479 with the entire EOIS lasting around 40 kyr. In the KR composite section, the positive shift associated with EOIS is  
 480 approximately 2‰, peaking at 1‰ at 19.60 m, marking the end of the EOT (Hutchinson et al., 2020) (Figure 4).

481

482 The Early Oligocene Glacial Maximum (EOGM) is characterized as a prolonged period of cold climate and glaciation during  
 483 the early Oligocene, corresponding to the most of ~~magnetic~~ Chron ~~paleomagnetic~~ Subchron C13n (Hutchinson et al., 2020)  
 484 (Figure 4). It spans from approximately 33.65 Ma to 33.16 Ma, lasting about 490 kyr. Correlation between the Karaburun data  
 485 and global deep-sea records was achieved by aligning the peak-to-peak  $\delta^{18}\text{O}$  stratigraphic intervals, starting at the top of the  
 486 EOIS at 19.60 m and extending to another peak at 36.75 m (0.6‰) corresponding to the top of Chron Subchron C13n (Figure  
 487 4).

488

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

### 493 6.3 The Early Oligocene cooling

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

508

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

513

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

#### 518 6.4 The ~~local~~regional & global effects in the Paratethys

519 The benthic foraminiferal oxygen and carbon isotope records from the Karaburun area closely resemble deep-sea records from  
520 Atlantic sites 522 and 1263 during the latest Eocene and early Oligocene (Figures 4, 5). However, a notable distinction is the  
521 pronounced negative  $\delta^{18}\text{O}$  shift just after the EOB, a feature characteristic of the Paratethys region (Figures 4 and 7) which  
522 will be discussed in the following. Firstly, it is noticeable that the overall benthic and planktic foraminifera  $\delta^{18}\text{O}$  values are  
523 more depleted than global records from the EOT. These depleted values likely reflect a ~~local~~regional effect rather than  
524 diagenetic alteration, as the exceptional preservation and glassy appearance of the foraminiferal shells from the KR composite  
525 section and other Paratethys sites (e.g., Ozsvárt et al., 2016) suggest minimal recrystallization. A major diagenetic overprint  
526 affecting the entire basin is also improbable given the differing tectonic and depositional histories across the Paratethys sub-  
527 basins. Additionally, the observed timescale (<100 ka) and the significant magnitude of changes in proxy records from the  
528 Paratethys Basin are unlikely to be explained by regional tectonic processes. Instead, ~~local~~ factors such as basin restriction,  
529 enhanced precipitation and/or freshwater input due to increased runoff or changing hydrological conditions during the EOT  
530 seem more plausible explanations. Similar  $\delta^{18}\text{O}$  depletion has been documented in other marginal basins during the EOT (e.g.,  
531 Pearson et al., 2008; De Lira Mota et al., 2023), further supporting a localized effect in semi-restricted environments due to  
532 local hydrology and climate. Indeed, these values are consistent with the isotopic composition of meteoric waters at mid-  
533 latitude coastal regions ( $\delta^{18}\text{O} \sim -5\text{\textperthousand}$  to  $-10\text{\textperthousand}$ ; Dansgaard, 1964; Gat, 1996). During the Rupelian (35–31 Ma), the dominant  
534 influence of Atlantic-derived westerlies likely brought increased precipitation with a depleted  $\delta^{18}\text{O}$  signature to the western  
535 and central Paratethys. This interpretation was supported by  $\delta^{18}\text{O}_{\text{PO}_4}$  values from herbivore tooth enamel, which reflect the  
536 depleted isotopic composition of drinking water (Kocsis et al., 2014). The enhanced precipitation was likely due to the  
537 intensification of the westerlies and the reorganization of oceanic and atmospheric circulation (e.g., Hou et al., 2022). As global  
538 cooling progressed and the meridional temperature gradient steepened, the westerlies strengthened, enhancing vapor transport  
539 from the Atlantic into Eurasia, contributing to higher precipitation in the western and central Paratethys (e.g., Kocsis et al.,  
540 2014; Li et al., 2018). Modeling studies further suggest prevailing westerly winds during winter at mid-high latitudes in the  
541 Rupelian (Li et al., 2018). The "continental effect," where  $\delta^{18}\text{O}$  in meteoric water becomes progressively fractionated with

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542 increasing transport distance from the Atlantic, likely contributed to more negative  $\delta^{18}\text{O}$  values in the Karaburun area (e.g.,  
543 Kocsis et al., 2014). Further east, evaporation over the Basin geometry, particularly water depth, plays a crucial role in  
544 evaporation dynamics within a semi-isolated system like the Paratethys. The Western Paratethys, characterized by deeper  
545 basins, had a higher heat capacity and greater thermal inertia, leading to moderated surface temperatures and lower evaporation  
546 rates. In contrast, the shallower Eastern Paratethys had lower heat capacity, allowing for rapid surface warming, especially in  
547 warmer seasons, which enhanced evaporation. Additionally, wind-driven effects likely played a significant role. In deeper  
548 basins, wind-driven mixing redistributed heat within the water column, reducing extreme surface warming and stabilizing  
549 evaporation rates. In shallower areas, reduced mixing allowed surface waters to remain warmer, leading to increased  
550 evaporation, particularly under strong seasonal winds. Hence, further east, evaporation over the shallower Eastern Paratethys  
551 may have added moisture to westerly air trajectories, resulting in relatively less negative  $\delta^{18}\text{O}$  values in the Northern Caucasus  
552 (Karaburun, Belaya and Chirkei sections in Figure 7) and increased inland precipitation (Figure 7). A similar precipitation  
553 gradient, with wetter conditions in the western-central Paratethys and drier conditions in the east, is also evident in an  
554 Oligocene climate reconstruction based on plant macrofossil data (Li et al., 2018). Additionally, the increased fresh  
555 waterfreshwater input in the Paratethys at the EOB could be plausibly explained by the major sea level fall and falling base  
556 level, driven by glacio-eustasy associated with the growth of Antarctic ice sheets during the EOT. The reorganization of rivers  
557 due to the falling base level would have introduced fresh water into the depositional epicenters of the Paratethys. Combined  
558 together, the effects of localregional hydrological change and the base level fall due to the global major sea level fall at the EOB  
559 might have resulted in the depleted  $\delta^{18}\text{O}$  values observed in the Paratethys. Despite the localized variations in the depletion of  
560  $\delta^{18}\text{O}$  values, the relatively consistent  $\delta^{18}\text{O}$  depletion observed across Paratethys sections suggests a uniform basin-wide isotopic  
561 background. This consistency allows for reliable identification of major trends and isotopic excursions in the Paratethys Basin  
562 during the EOT.

563  
564 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  
565  $\delta^{13}\text{C}$  values (Figure 6). Particularly during the Step1 same trends in isotopic shifts could be clearly recognized. Most  
566 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  
567 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  
568 significant stratification and a reduction in vertical mixing. The pronounced negative  $\delta^{18}\text{O}$  shift in benthic foraminifera likely  
569 reflects a significant influx of isotopically light cold freshwater into the bottom waters. The slight increase in planktic  
570 foraminifera  $\delta^{18}\text{O}$  at the same time suggest that the surface waters might have become relatively saline due to evaporation  
571 exceeding freshwater input in the surface layer, which would increase  $\delta^{18}\text{O}$  values. Cold, freshwater inflow might have been  
572 funneled into deeper areas of the basin, displacing or mixing with bottom waters. This might have been happened through the  
573 submarine channels providing sediment-laden freshwater as underflows into the deep-marine turbiditic systems. However, this  
574 would have required an anti-estuarine circulation model where marine saltwater flows upstream and overrides the freshwater  
575 inflow. Indeed, an anti-estuarine circulation model for the early Oligocene Paratethys was proposed earlier (Dohmann, 1991)

576 which was later supported by Schulz et al. (2005) showing also increasing surface salinities due to evaporation of marine water  
577 based on increasing di-/tri-MTTC ratios. An early anti-estuarine circulation model for the Paratethys aligns perfectly with our  
578 abovementioned stable isotope data. The deep ~~fresh water~~<sup>freshwater</sup> input could be explained by the early evolution of the  
579 Paratethys (e.g., Schulz et al., 2005). Early evolution of the Paratethys was mainly controlled by narrowing seaways connecting  
580 it to the Tethys Ocean which led to ingressions of cold boreal water from the north (through Polish straits), initially as  
581 undercurrents (generating an anti-estuarine circulation) into the Eastern Paratethys first and then to Central and Western  
582 Paratethys (e. g., Schulz et al., 2005; Soták, 2010). In parallel to the  $\delta^{18}\text{O}$  values, a divergent trend could be observed between  
583 the benthic and 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  
584 an increase (Figure 6). In addition to the boreal fresh-water ingressions, enhanced organic matter production due to increased  
585 nutrient input and then the subsequent decomposition in the isolated, stratified Paratethys waters might have released light  
586 carbon into the bottom waters. Increased  $\delta^{13}\text{C}$  in planktic foraminifera could have resulted from elevated primary productivity  
587 driven by increased nutrient input, which preferentially removes isotopically light carbon from surface waters during  
588 photosynthesis, leaving the remaining carbon pool enriched in heavier carbon.

589

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

604

605 The decrease in  $\delta^{18}\text{O}$  of benthic foraminifera at ca. 22 m is ~~likelypossibly~~ related to the slight warming ~~observedrecorded~~ in  
606 the North Sea (Śliwińska et al., 2019b) in ~~magnetic chron~~<sup>paleomagnetic Subchron</sup> C13n above the EOIS (Figure 6). The  
607 increase in  $\delta^{18}\text{O}$  planktic foraminifera after ca. 25 m up to the ca. 40-41 m is likely related with further cooling during the  
608 EOOG. A sharp declining trend could be noticed for benthic foraminifera  $\delta^{13}\text{C}$  during the onset of this interval (at ca. 25 m)

609 suggesting a relatively less organic carbon burial. This was due to a decrease in primary productivity at the surface represented  
610 by lowering  $\delta^{13}\text{C}$  of planktic foraminifera and lowered terrestrial input.

611  
612 The relative sea-level in the Paratethys starts to lower after the ca. 40-41 m level (Figure 3) which is followed by another fresh  
613 waterfreshwater input likely due to enhanced precipitation at level ca. 45 m. This is evidenced by the depleted  $\delta^{18}\text{O}$  planktic  
614 foraminifera values and an increase in terrestrial palynomorphs (Figures 3, 6). Once again this was followed by an increase in  
615 organic carbon burial represented by a peak in benthic  $\delta^{13}\text{C}$  values at ca. 47 m.

616  
617 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  
618 and suggests a decrease in organic carbon burial in the Karaburun area (Figure 6). Decreasing relative sea-level and related  
619 increase in bottom current velocities and wave action combined with a decrease in fresh water input should have likely  
620 decreased the organic carbon burial suggested by decreasing benthic  $\delta^{13}\text{C}$  at ca. 52 -53 m. Another fresh waterfreshwater  
621 ingress as 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  
622 that the global Oi-1a cooling signal dominates the upper part of the section between levels ca. 60 m and 75 m. At ca. 63 ma-m  
623 a decrease in planktic  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values corresponds to surface freshwater input accompanied by increased organic carbon  
624 burial (peak 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  
625 fresh waterfreshwater influx at level ca. 65 m. The uppermost peak in benthic  $\delta^{13}\text{C}$  values at ca. 70 m represents and enhanced  
626 organic carbon burial in the Karaburun area due to more favorable conditions for organic carbon sequestration provided by a  
627 relative sea level rise (e.g., lower bottom current velocities and lower wave action).

628  
629 The divergent trend between benthic and planktic  $\delta^{13}\text{C}$  values indicates a highly stratified Paratethys Sea from time to time  
630 and different surface and bottom carbon cycling processes after a change from anti-estuarine to estuarine circulation during  
631 the EOT. These changes reflect both regional hydrological and basin reconfiguration (restriction) controls and global climatic  
632 and eustatic shifts within the EOT. The global cooling during the EOT and Oi-1 must have amplified the stratification, reduced  
633 ventilation and triggered local environmental shifts in the semi-enclosed Paratethys Basin. These environmental shifts provided  
634 favorable conditions for the deposition of organic-rich fine-grained sediments with high Total Organic Carbon (TOC) values.  
635

636 Overall, this observed contrasting pattern between the isotopic values of benthic and planktic foraminifera highlights the  
637 complex interplay of global climate trends (Antarctic glaciation and global cooling) and regional factors (basin isolation and  
638 hydrological changes) during the EOT. Our findings align with the proposed isolation of the Paratethys, driven by the  
639 prolonged African-Arabian-Eurasian collision coupled with eustatic sea-level decline at the EOB and the cooling during the  
640 EOT. In turn this led to the development of a distinct Paratethyan domain marked by mesophilic, humid climatic conditions  
641 and intensified runoff (Popov et al., 2002). Moreover, the observedreconstructed relative sea-level fluctuationevolution  
642 closely correspond withto other reconstructions of relative sea-level changes from the latestlate Eocene to early Oligocene

643 period (e.g., from the [Neo-Tethys, Brinkhuis, 1994](#); North Sea, Jarvse et al., 2015), further reinforcing the presence of a global  
644 climatic signal, as well.

#### 645 **6.5 The Boreal water in the Paratethys during the [Early](#)[early](#) Oligocene and its paleoceanographic significance**

646 The prominent negative shift in  $\delta^{18}\text{O}$  of benthic foraminifera [likely](#) representing the boreal water [ingress](#)[observed](#)[ingress](#)  
647 [recorded](#) in the Karaburun composite section shortly after the EOB (ca. 33.7 Ma) appears to be widespread across the entire  
648 Paratethys Basin (Figure 7; Soták, 2010; Ozsvárt et al., 2016; Gavrilov et al., 2017; van der Boon et al., 2019). In the Eastern  
649 Paratethys, this shift appears to have occurred abruptly. A similar negative shift in  $\delta^{18}\text{O}$  of bulk carbonates [was](#)[is](#) also recorded  
650 further west in the West Alpine Foreland Basin (Chalufy section, Soutter et al., 2022), which connected the Western Paratethys  
651 to the Mediterranean Tethys Ocean (Figures 7, 8). Hence, [the](#)-boreal water [ingress](#)[ingress](#) into the Paratethys, beginning in  
652 the Eastern Paratethys and progressively reaching the Central and Western regions, is clearly represented by this negative shift  
653 across the entire Paratethys Basin just after [the](#) EOB. This process was likely linked to the restriction or closure of the Arctic-  
654 Atlantic gateway and the onset of anti-estuarine circulation between the Atlantic and the Nordic Seas during the EOT.  
655 [Together, the Atlantic-Arctic closure and onset of anti-estuarine circulation events could have triggered or intensified the](#)  
656 [Atlantic Meridional Overturning Circulation \(AMOC\) \(e.g., Abelson and Erez, 2017; Coxall et al., 2018; Hutchinson et al.,](#)  
657 [2019\).](#)

658  
659 Proxy records indicate that the Eocene Arctic Ocean was significantly fresher than today, with salinities ranging from 20 to 25  
660 psu and occasional drops below 10 psu (Brinkhuis et al., 2006; Kim et al., 2014; Waddell and Moore, 2008). The Arctic  
661 freshwater outflow into the North Atlantic may have inhibited deep-water formation during the Eocene (Baatsen et al., 2020;  
662 Hutchinson et al., 2018). [Sea-level and paleo-shoreline reconstructions in the Nordic Seas support the hypothesis that the](#)  
663 [Arctic became isolated during the latest Eocene to early Oligocene due to the closure of the Arctic-Atlantic gateway \(Hegewald](#)  
664 [and Jokat, 2013; O'Regan et al., 2011; Hutchinson et al., 2019\).](#) Recent evidence suggests that the deepening of the Greenland-  
665 Scotland Ridge (GSR) around the EOT (just before the EOIS, ca. 33.7 Ma) facilitated increased exchange between the Atlantic  
666 and the Nordic Seas, enabling the formation of anti-estuarine circulation and the salinization of North Atlantic surface waters  
667 (Abelson and Erez, 2017; Stářík et al., 2017). [Sea-level and paleo-shoreline reconstructions in the Nordic Seas support the](#)  
668 [hypothesis that the Arctic became isolated during the latest Eocene to early Oligocene due to the closure of the Arctic-Atlantic](#)  
669 [gateway \(Hegewald and Jokat, 2013; O'Regan et al., 2011; Hutchinson et al., 2019\).](#) Consequently, the gradual restriction of  
670 Arctic-Atlantic connectivity, followed by the onset of anti-estuarine circulation driven by the deepening of the GSR, may have  
671 played a critical role in developing a robust AMOC (e.g., Coxall et al., 2018; Hutchinson et al., 2019).

672  
673 [The- Together, the Atlantic-Arctic closure of the Arctic gateway and the initiationonset of Nordic anti-estuarine circulation](#)  
674 [around the EOT likely enhanced deep water formation, strengthening the Atlantic overturning circulation and establishing the](#)  
675 [interhemispheric northern sourced circulation cellevents could have triggered or intensified the Atlantic Meridional](#)

676 OVERTURNING CIRCULATION (AMOC) (e.g., Abelson and Erez, 2017; Coxall et al., 2018; Hutchinson et al., 2019). The onset of  
677 Nordic anti-estuarine circulation around the EOT might have likely influenced salinity gradients and circulation in connected  
678 basins like the North Sea and Paratethys, contributing to the isolation and freshening of the latter (Figure 8). Our data indicate  
679 that the onset of anti-estuarine circulation in the Paratethys (ea. around 33.7 Ma) coincided, coinciding with its the development  
680 of similar circulation patterns between the Nordic Seas and the North Atlantic (Abelson and Erez, 2017). This suggests that at  
681 ea. 33.7 Ma We propose a hypothetical pathway for this circulation: warm surface water entering from the North Sea would  
682 have flowed into the eastern Nordic Seas, joining the warm surface waters from the North Atlantic. After losing heat, this  
683 water likely sank in the northern Nordic basin, flowed southward as deep water, and eventually reached the North Atlantic,  
684 the North Sea, the Paratethys ~~with its anti-estuarine circulation was part of the, and the Mediterranean Tethys, forming an~~  
685 interhemispheric northern-sourced circulation cell (Figure 8a).

686  
687 The onset of anti-estuarine circulation in the Paratethys was likely related to the deepening of the Greenland-Scotland Ridge  
688 (GSR), similar to the development of anti-estuarine circulation in the Nordic Seas. During most of the Paleogene, the GSR  
689 was shallower, restricting deep-water exchange between the North Atlantic and the Nordic Seas. A key feature of the modern  
690 AMOC is the formation of North Atlantic Deep Water, which flows over the GSR from the Nordic Seas. The initiation of  
691 Nordic anti-estuarine circulation events around the EOT likely enhanced deep-water formation, strengthening the Atlantic  
692 Meridional Overturning Circulation (AMOC) and establishing an interhemispheric northern-sourced circulation cell (e.g.,  
693 Abelson and Erez, 2017; Coxall et al., 2018; Hutchinson et al., 2019). This suggests that by ~33.7 Ma, the Paratethys, with its  
694 anti-estuarine circulation, was integrated into this larger circulation system, contributing to global ocean circulation (Figure  
695 8). Shortly after, subsequent geographic restrictions and hydrological changes during the EOT changed this anti-estuarine  
696 circulation to an estuarine circulation (e.g. Figure 8b). The closure of the Mediterranean Seaway—along with the narrowing of  
697 other seaways likely caused by the major sea-level fall during the EOIS and EOGM—and increased freshwater influxes from  
698 the continent diluted the surface Tethyan waters, disrupted the anti-estuarine circulation pattern, and led to brackish surface  
699 salinities in the Paratethys (e.g., Schulz et al., 2005; Soták, 2010).

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701 Further evidence supporting boreal water ingress and circulation through the Nordic Seas, North Sea and Paratethys  
702 comes from the distribution of *Svalbardella cooksoniae* in the Greenland, Norwegian, and North Seas, as well as in the Eastern  
703 Paratethys during the Oi-1a cooling event (Śliwińska and Heilmann-Clausen, 2011). This suggests a possible migration  
704 pathway for *Svalbardella* spp. The presence of *Svalbardella* in the Massicore and Monte Cagnero sections of central Italy  
705 (Brinkhuis and Biffi, 1993; van Mourik and Brinkhuis, 2005) aligns with our interpretation of boreal water circulation  
706 extending to the Mediterranean Tethys through the Paratethys (Figure 8). Sinking cold boreal freshwater likely propagated  
707 through interconnected Nordic marine basins, reaching the Paratethys and eventually the Mediterranean Tethys (Figure 8).  
708 The invasion of the Mediterranean Tethys by higher-latitude taxa around the EOB (Brinkhuis and Biffi, 1993) further supports  
709 this circulation pathway. The incursion of old boreal bottom waters into the Mediterranean Tethys during the EOT is

710 corroborated by a marked transient increase in deep-water ostracod *Krithe* and a decline in deep-water ostracod diversity in  
711 the Massignano composite section (Slotnick and Schellenberg, 2013). The *Krithe* pulse and the subtle change in the deep-  
712 water ostracod fauna reflects intensified thermohaline flow of cooler deep waters, likely linked to boreal freshwater circulation  
713 through the Paratethys (e.g., Dall'Antonia et al., 2003; Slotnick and Schellenberg, 2013). Variations in seafloor ventilation and  
714 productivity due to changes in paleoceanographic conditions of the Tethys during the late Eocene - early Oligocene described  
715 in previous studies (Jovane et al., 2007 and references therein) are also likely related to the circulation of the boreal water.

716

717 The influence of boreal waters on the Mediterranean Tethys might also ~~explain~~explain the absence of well-defined positive  
718  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  shifts typically characterizing the EOT in Italian sections, including the GSSP for the EOB (e.g., Houben et al.,  
719 2012). 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.,  
720 Jovane et al., 2007; Brown et al., 2009; Jaramillo-Vogel et al., 2013), resembling the Paratethys records and likely reflecting  
721 deep boreal freshwater incursion through the Paratethys.

## 722 7 Conclusion

723 This study provides a comprehensive examination of the EOT and Earlyearly Oligocene cooling in the eastern Paratethys, with  
724 a focus on the Karaburun composite section. By integrating high-resolution biostratigraphy, geochemistry, sequence  
725 stratigraphy, and precise geochronology, we constructed a robust chronostratigraphic framework spanning the latest Eocene  
726 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  
727 records, underscoring the influence of global climatic drivers during this critical transition. However, significant localregional  
728 deviations, such as depleted  $\delta^{18}\text{O}$  values and pronounced stratification, highlight the impact of regional hydrological changes  
729 and basin restriction of the Paratethys. These results emphasize the dual influence of global icehouse dynamics and regional  
730 hydrological processes on the eastern-Paratethys during the EOT. The identification of key cooling events, such as the  
731 *Glyphyrocysta semitecta* influx for the EOB, and the *Svalbardella*-spp. event and its alignment with the Oi-1a glaciation,  
732 further enhances the utility of the Karaburun section for refining regional and global stratigraphic correlations. Additionally,  
733 the observed sequence stratigraphic patterns illustrate the interplay between eustatic sea-level changes and localregional  
734 depositional dynamics during this interval. The abrupt depleted values in  $\delta^{18}\text{O}$  values in benthic foraminifera just after the  
735 EOB (ca. 33.7 Ma) in the Paratethys Basin is attributed to boreal water ingressions, driven by the closure of the Arctic-Atlantic  
736 gateway and the onset of anti-estuarine circulation between the Nordic seas and Atlantic during the EOT. This event, which  
737 funneled low-salinity boreal water through the Nordic and North Seas into the Paratethys, aligns with the distribution of boreal  
738 taxa like *Svalbardella cooksoniae* and extends as far as the Mediterranean Tethys, explaining isotopic anomalies in these  
739 regions. We have to note that the large-scale oceanic circulation scenario proposed here requires the movement of deep waters  
740 through narrow and shallow seaways surrounding the Paratethys, such as the Polish Strait, the Alpine Seaway, and smaller  
741 Neo-Tethys-Paratethys gateways. However, the timing of their closure and their paleobathymetry during the early Oligocene

742 remain uncertain, as existing paleogeographic maps for this period are inconsistent (e.g. Barrier et al., 2018; Palcu &  
743 Krijgsman, 2023; Straume et al., 2024). Therefore, high-resolution paleogeographic reconstructions, integrated with ocean  
744 circulation modelling (e.g., Vahlenkamp et al., 2018), are essential for testing and validating this hypothesis. Overall, the  
745 Karaburun section emerges as a critical archive for studying the EOT in epicontinental seas. Our findings contribute to a deeper  
746 understanding of how global climatic transitions manifest in marginal marine settings and highlight the potential for further  
747 high-resolution studies to refine our knowledge of early icehouse climate evolution in the Paratethys region and beyond.

748

#### 749 **Author contribution**

750 MYK designed the study. MYK and SGA conducted fieldwork. TV prepared samples for geochemical analyses. DN  
751 performed geochemical analyses. AL conducted geochronological analysis. HB ~~analyzed dinocyst stratigraphy, performed the~~  
752 ~~marine palynological analyses.~~ CF analyzed ~~nannofossil stratigraphy, the calcareous nannofossils.~~ MYK wrote the paper with  
753 input from all authors. All authors analyzed and discussed the data.

754

#### 755 **Competing interests**

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

757

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