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Global warming and equatorial Atlantic paleoceanographic changes during early Eocene carbon cycle perturbation V

Anne H. Kegel^{1#}, Chris D. Fokkema^{1#}, Henk Brinkhuis^{1,2}, Ursula Röhl³, Thomas Westerhold³, Claudia Agnini⁴, Peter K. Bijl¹, Francien Peterse¹ and Appy Sluijs¹

¹Department of Earth Sciences, Faculty of Geoscience, Utrecht University, 3584CB Utrecht, The Netherlands

²Department of Ocean Systems research (OCS), Royal Netherlands Institute for Sea Research (NIOZ), 1790 AB Den Burg, The Netherlands.

³MARUM - Center for Marine Environmental Sciences, University of Bremen, Leobener Straße, 28359 Bremen, Germany

⁴Dipartimento di Geoscienze, Università degli Studi di Padova, I-35131 Padova, Italy.

Corresponding author: Chris D. Fokkema (c.d.fokkema@uu.nl)

#Shared first authorship

Key Points:

- Equatorial sea surface and deep ocean warming marked early Eocene carbon isotope excursion V, implying it was a global warming event.
- A longer isotope recovery in organic compared to inorganic carbon suggests a contribution of older dissolved organic carbon at Site 959.
- Dinoflagellate cysts indicate a productive setting related to upwelling that commenced since the end of the Early Eocene Climatic Optimum.

Abstract

A series of transient global warming events ("hyperthermals") in the early Eocene is marked by massive environmental and carbon cycle change. Among these events, the impacts of the Paleocene Eocene Thermal Maximum (~56 Ma), Eocene Thermal Maximum 2 (~54 Ma) and Eocene Thermal Maximum 3 (~53 Ma) are relatively well documented, but much less is known on the many later hyperthermals that apparently occurred on orbital eccentricity maxima until at least the end of the Early Eocene Climatic Optimum (EECO; ~53–49 Ma). Here, at Ocean Drilling Program (ODP) Site 959 (Equatorial Atlantic Ocean), we report a large negative carbon isotope excursion (CIE) in both organic and carbonate substrates that we correlate to the "V" event *sensu* Lauretano et al. (2016) (or C22nH1 *sensu* Sexton et al. (2011)) at ~49.7 Ma, following combined bio- and chemostratigraphic

39 constraints. Through TEX₈₆ paleothermometry, we reconstruct a sea surface temperature rise of 1.1–
40 1.9 °C associated with this CIE, which, combined with evidence for warming from the deep sea,
41 implies that this event indeed represents a transient global-scale warming episode like the earlier
42 hyperthermals. Organic walled dinoflagellate cyst assemblages indicate a productive
43 paleoceanographic background setting, likely through regional upwelling, which alternated with
44 episodes of stratification. Warming reconstructed across V at Site 959 is relatively similar to the
45 higher-latitude-derived deep ocean reconstructions. However, the presence of upwelling and its
46 variable intensity across the event compromises the use of the reconstructed warming as an estimate
47 for the complete tropical band.

48 1. Introduction

49 Persistent deep ocean warming during the late Paleocene and early Eocene climaxed at the
50 end of the Early Eocene Climatic Optimum (EECO; ~53–49 Ma) (Shackleton, 1986; Westerhold et
51 al., 2020). This trend is mimicked by sea surface temperature (SST) reconstructions in both low and
52 middle, and high latitudes, indicating that they represent global mean surface temperature (GMST)
53 variability (e.g., Bijl et al., 2009; Hollis et al., 2012; Cramwinckel et al., 2018; Gaskell et al., 2022).
54 During the EECO, GMSTs were approximately 10–16 °C elevated relative to pre-industrial
55 temperatures (Inglis et al., 2020) and atmospheric *p*CO₂ exceeded 1000 ppmv (Anagnostou et al.,
56 2020).

57 Superimposed on this long-term trend occur approximately 20 transient negative stable
58 carbon ($\delta^{13}\text{C}$) isotope excursions (CIEs) in the ocean-atmosphere system (Kennett and Stott, 1991;
59 Cramer et al., 2003; Lourens et al., 2005; Agnini et al., 2009; Sexton et al., 2011; Littler et al., 2014;
60 Kirtland Turner et al., 2014; Frieling et al., 2018; Lauretano et al., 2018). These transient CIEs and
61 associated deep ocean carbonate dissolution horizons likely reflect the rapid and massive injection of
62 ¹³C-depleted carbon from outside the global exogenic carbon reservoir pool into the ocean-
63 atmosphere system (Dickens et al., 1995, 1997). It is demonstrated that most or all of these events
64 are paced by eccentricity cycles of ~100- and ~405-kyrs (Cramer et al., 2003; Lourens et al., 2005;
65 Galeotti et al., 2010; Zachos et al., 2010; Littler et al., 2014; Lauretano et al., 2016; Laurin et al.,
66 2016; Westerhold et al., 2017, 2018; Piedrahita et al., 2022). This view is recently strengthened by a
67 recorded loss in resilience in climate-carbon cycle dynamics leading up to the events (Setty et al.,
68 2023). The coeval negative excursions in benthic foraminiferal oxygen isotope ratios ($\delta^{18}\text{O}$) are
69 interpreted to reflect deep ocean warming (Kennett and Stott, 1991; Lourens et al., 2005; Lauretano
70 et al., 2018). SST proxy records indicate that at least a great part of these events indeed represent
71 global warming events (Lourens et al., 2005; Sluijs et al., 2009; Dunkley Jones et al., 2013; Frieling
72 et al., 2018; Fokkema et al., 2023) and can therefore be termed ‘hyperthermals’ (Thomas and
73 Zachos, 2000). However, the exact driving mechanisms and carbon sources are still under debate
74 (e.g., Frieling et al., 2019). Potential external carbon sources can include (a combination of) methane
75 hydrates and terrestrial organic carbon (Dickens et al., 1995; Kurtz et al., 2003; DeConto et al.,
76 2012). Volcanic sources may have contributed to carbon injection during the largest hyperthermal,
77 the Paleocene-Eocene Thermal Maximum (PETM; ~56 Ma; e.g., Svensen et al., 2004; Frieling et al.,
78 2019; Berndt et al., 2023).

79 The PETM is associated with massive environmental and concomitant biotic change on land
80 and in the ocean (Sluijs et al., 2007; McInerney and Wing, 2011; Hupp et al., 2022). Some of the
81 subsequent lesser hyperthermals are regionally associated with biotic changes as well, notably
82 Eocene Thermal Maximum 2, and Eocene Thermal Maximum 3, but others, where documented,
83 show little to no change (Agnini et al., 2009; Sluijs et al., 2009; Gibbs et al., 2012; Willard et al.,
84 2019; Rush et al., 2023). In contrast, it has been proposed that in some tropical regions during the
85 PETM, high-resolution fossil organic walled dinoflagellate cyst (dinocyst) and planktic foraminifer
86 assemblages suggest the disappearance of eukaryotic plankton because of heat stress, when warming

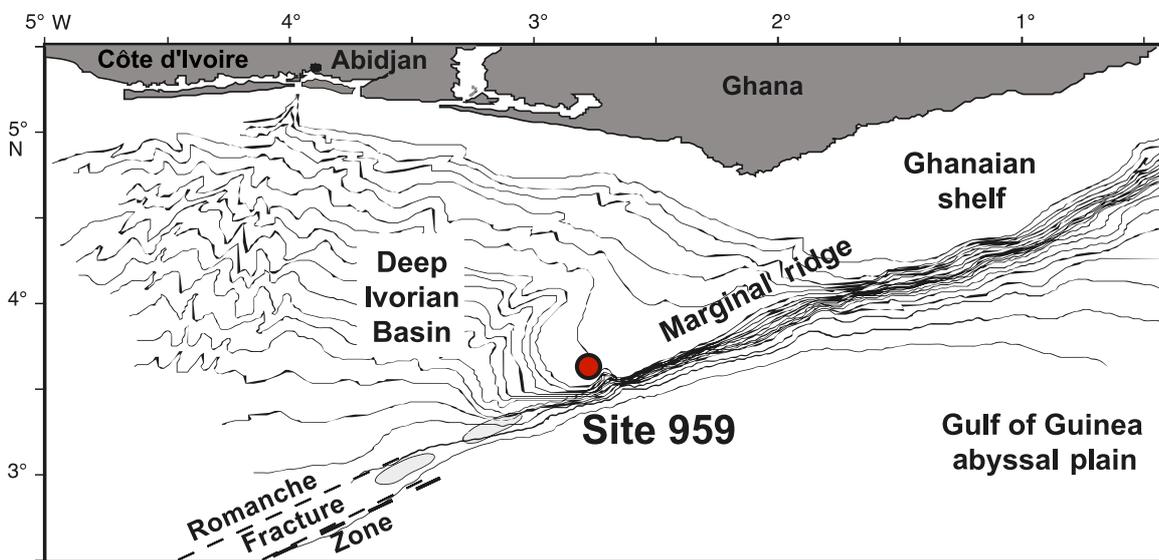
87 led to temperatures near thermo-physiological limits, or the rate of warming resulted in reduced
 88 prevalence and biodiversity (Aze et al., 2014; Frieling et al., 2017, 2018).

89 So far, most studies focused on the early set of hyperthermals, whereas the events that
 90 occurred during the later part of the EECO received much less attention (Sexton et al., 2011;
 91 Kirtland Turner et al., 2014; Westerhold et al., 2017, 2018). Consequently, while for a series of the
 92 earlier hyperthermals (PETM–L2, ~56–52.4 Ma) surface ocean warming has been reconstructed in
 93 multiple regions including the tropics (Sluijs et al., 2009; Frieling et al., 2017, 2018; Fokkema et al.,
 94 2023), evidence for warming of hyperthermals during the later part of the EECO (~52–49 Ma) is
 95 limited to negative excursions in benthic foraminiferal $\delta^{18}\text{O}$ (Westerhold et al., 2018), that
 96 presumably reflects Southern Ocean SST variability (Hollis et al., 2012; Zhang et al., 2022). In
 97 addition, to our knowledge, no studies of biotic change have been carried out for these later
 98 perturbations.

99 Therefore, here we aim to evaluate temperature and ecological variability across the later part
 100 of the EECO in sediments recovered at Ocean Drilling Program (ODP) Site 959, located in the
 101 Equatorial Atlantic (Fig. 1). Previous Eocene reconstructions on Site 959 using TEX₈₆ (TetraEther
 102 indeX of tetraethers consisting of 86 carbon atoms) paleothermometry have demonstrated that the
 103 PETM was associated with ~3 °C warming, and hyperthermals I1–L2 with ~1–1.5 °C from baseline
 104 temperatures exceeding 33 °C (Frieling et al., 2019; Fokkema et al., 2023). We follow up on this
 105 work by searching for CIEs using stable carbon isotope analyses on a late-EECO interval of Site 959.
 106 We will constrain the chronostratigraphy using the available biostratigraphic data and identify
 107 hyperthermal related variability in the record. Then we assess environmental and biotic changes over
 108 the interval using TEX₈₆ and palynological associations.

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111

112 **Figure 1.** Present day location of ODP Site 959 in the eastern equatorial Atlantic Ocean. Map
 113 adapted from Mascle et al. (1996).

114

115 2. Material and methods

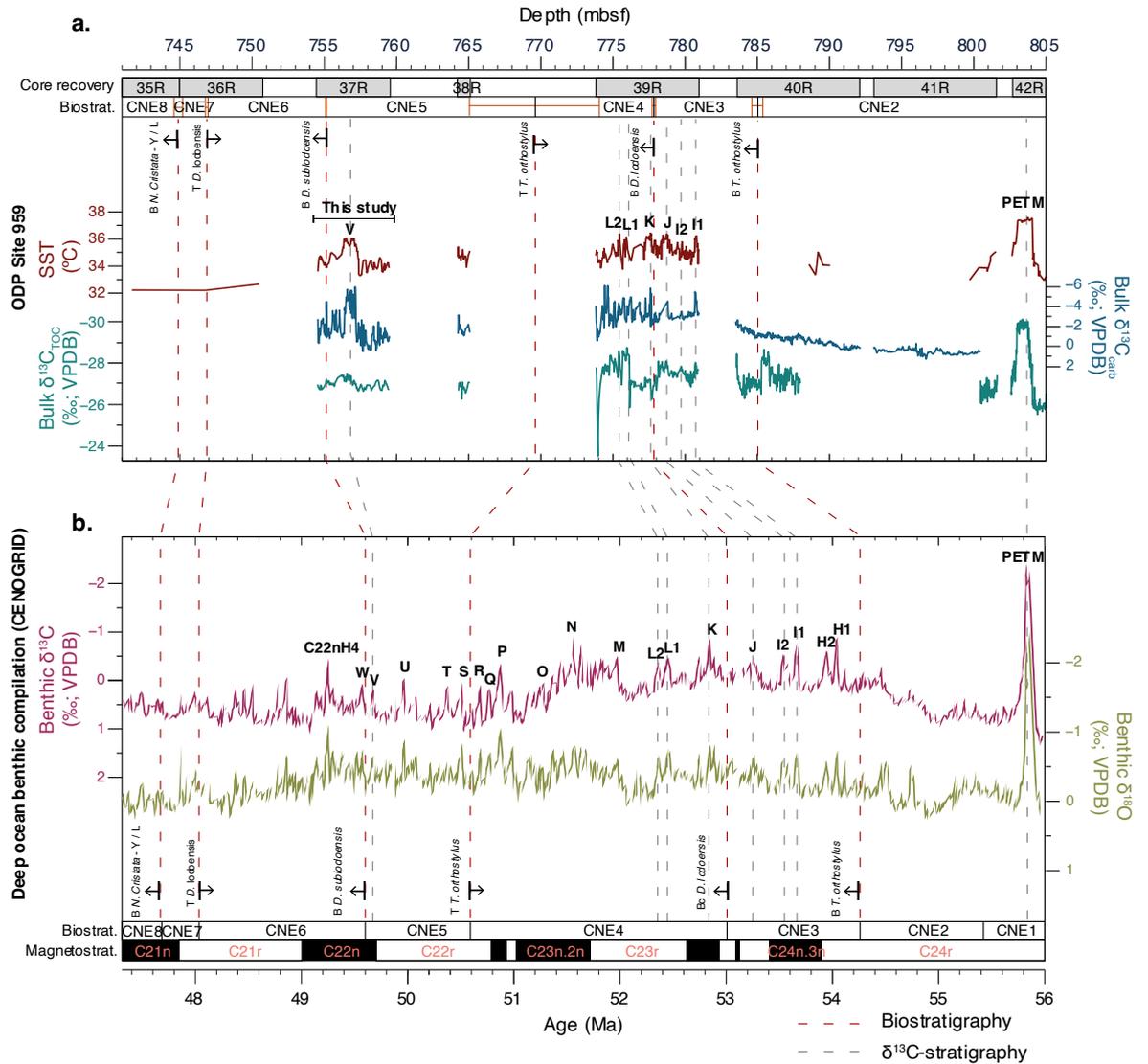
116 2.1 Ocean Drilling Program Site 959

117 ODP Leg 159 Site 959 (3.62760° N, 2.735817° W) is located ~140 km offshore of Ghana in
118 the eastern equatorial Atlantic Ocean at a depth of 2100 m on the northern slope of the Côte d'Ivoire-
119 Ghana Marginal Ridge, which divides the Deep Ivorian Basin from the abyssal plain of the Gulf of
120 Guinea (Mascle et al., 1996) (Fig. 1). Site 959 has accumulated sediments with ample organic matter
121 since the Late Cretaceous (Wagner, 2002), which has allowed for low latitude paleoclimate
122 reconstructions by lipid biomarkers and organic walled dinoflagellate cysts (dinocysts) throughout
123 the Cenozoic Era (e.g., Cramwinckel et al., 2018; van der Weijst et al., 2022). The lower Eocene
124 sediments, consisting of clay-bearing chalks and porcellanites (Mascle et al., 1996), were deposited
125 at a paleolatitude of ~9° S (<http://paleolatitude.org>, model version 2.1; (van Hinsbergen et al., 2015);
126 using the paleomagnetic reference frame of Torsvik et al. (2012)). An open marine depositional
127 environment during the early Eocene was inferred by previous work based on the (near) absence of
128 terrestrial palynomorphs and dominance of dinocyst taxa typical for outer shelf to open marine
129 environments (Fokkema et al., 2023). The appearance of heterotrophic dinocysts and increased total
130 organic carbon (TOC) content around the end of the early Eocene (~49 Ma) suggests the onset, or
131 intensification of upwelling around that time (Cramwinckel et al., 2018; Fokkema et al., 2023).

132 2.2 Stratigraphic framework

133 The PETM and CIEs I1 to L2 were previously identified in Hole D Cores 42R (Frieling et al.,
134 2018) and 39R (Fokkema et al., 2023; Fig. 2). Core 38R had very poor recovery (10 %, Mascle et al.,
135 1996). Therefore, our search for late EECO CIEs focuses on Core 37R, which has reasonable
136 recovery (51.8 %, Mascle et al., 1996) (Fig. 2). Incomplete recovery around the studied interval
137 complicates the construction of a reliable stratigraphic framework. However, the base of calcareous
138 nannofossil *Discoaster subloidoensis* in Core 37R at 755.11±0.04 mbsf (Fokkema et al., 2023)
139 provides important constraints. This event is generally considered a problematic biohorizon, due to
140 its relative proneness to diagenetic overgrowth, its sporadic occurrences in the lower part of its range
141 (Backman, 1986), its relatively recently improved taxonomy (Agnini et al., 2006) and its inconsistent
142 position at Walvis Ridge ODP Sites 1265 and 1263 (Westerhold et al., 2017). Nevertheless, the use
143 of a consistent taxonomy and the standardization of a semi quantitative counting result in a very
144 consistent position of this event between astronomically tuned ODP Sites 1265 (Walvis Ridge, South
145 Atlantic) and Site 1258 (Demerara Rise, equatorial Atlantic), where it is reported in the base of
146 magnetochron C22n, between the V (49.685 Ma; C22nH1) and W (49.585 Ma; C22nH2) CIEs
147 (Westerhold et al., 2017). The offset observed between Site 1263 and 1265 is potentially caused by
148 core disturbance of this interval at Site 1263 (Westerhold et al., 2017). Therefore, the available data
149 support the hypothesis that the base of *D. subloidoensis* is indeed positioned between the V and W
150 CIEs. The calculated age of this event at Sites 1265 and 1258 is also consistent with the tuned age of
151 this biohorizon at the Newfoundland Ridge International Ocean Discovery Program (IODP) Site
152 U1410, although that site lacks paleomagnetic data in that interval (Cappelli et al., 2019). Moreover,
153 the position between V and W agrees with unpublished observations of the base of *D. subloidoensis*
154 datum from the low-latitude Pacific ODP Site 1209 (C. Agnini, personal observation, 2023).
155 Collectively, we might expect to find CIEs V and/or W close to this bioevent.

156



157

158 **Figure 2.** Stratigraphic correlation of Site 959 to a global deep ocean benthic compilation. (a) Site
 159 959 data (TEX₈₆-based SST, bulk organic and carbonate $\delta^{13}C$, calcareous nannofossil
 160 biostratigraphy), including newly generated data for Core 37R and previously published data
 161 (Frieling et al., 2018, 2019; Cramwinckel et al., 2018; Fokkema et al., 2023). (b) benthic
 162 foraminiferal $\delta^{13}C$ and $\delta^{18}O$ from the CENOGRID compilation (Westerhold et al., 2020). The dashed
 163 lines mark the correlations based on calcareous nannofossils (red) and CIEs (grey). The calcareous
 164 nannofossil biostratigraphic zonation (CNE zones) follows definitions by Agnini et al. (2014), and
 165 the ages are the astronomically calibrated by Westerhold et al. (2017).

166

2.3 Methods

167

168 For this study, we continuously sampled the working halves of Hole 959D, Core 37R (759.5–
 169 754.5 mbsf) at 2 cm resolution at the Bremen Core Repository of the IODP. All samples were freeze-
 170 dried. Selections of samples were analyzed for magnetic susceptibility, bulk carbonate stable carbon
 171 and oxygen isotope ratios, bulk organic carbon isotope ratios, biomarkers, and palynology. Archive
 halves were subject to X-Ray Fluorescence (XRF) scanning.

172 2.3.1 Color Reflectance

173 Total color reflectance data was generated from the original sediment core images (Masclé et
174 al., 1996) following the methodology described by Zeeden et al. (2015). Images were prepared by
175 deleting cracks, drilling mud and other unconformities, and data was generated at 1-cm resolution
176 using the R-script by Kocken (2022).

177 2.3.2 Bulk magnetic susceptibility

178 For bulk magnetic susceptibility (MS) measurements, samples were crushed to ~0.5 cm large
179 pieces and weighed. MS was measured with a MFK1-FA on 212 samples at Utrecht University. Each
180 measurement was repeated at least thrice. Analytical uncertainty was smaller than $5.3 \times 10^{-10} \chi$.

181 2.3.3 XRF scanning

182 XRF Core Scanner data were collected in 2009 every 1 cm down-core over a 1.2 cm² area
183 with a down-core slit size of 10 mm using generator settings of 10 kV, a current of 0.2 mA, and a
184 sampling time of 30 seconds directly at the split core surface of the archive half with XRF Core
185 Scanner III (AVAATECH Serial No. 12) at the MARUM - University of Bremen. The split core
186 surface was covered with a 4 μm thin SPEXCerti Prep Ultralene1 foil to avoid contamination of the
187 XRF measurement unit and desiccation of the sediment. The here reported data have been acquired
188 by a Canberra X-PIPS Detector (Model SXP 5C–200–1500) with 200eV X-ray resolution, the
189 Canberra Digital Spectrum Analyzer DAS 1000 and an Oxford Instruments 100W Neptune X-ray
190 tube with rhodium (Rh) target material. Raw data spectra were processed by the analysis of X-ray
191 spectra by Iterative Least square software (WIN AXIL) package from Canberra Eurisys.

192 2.3.4 Isotope measurements

193 Stable carbon and oxygen isotope ratios of bulk carbonate ($\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$, respectively)
194 were measured on powdered samples using a Thermo Finnigan GasBench II system coupled to a
195 Thermo Delta-V isotope ratio mass spectrometer (IRMS) at Utrecht University. Results were
196 calibrated to delta values relative to the Vienna Pee Dee Belemnite (VPDB) using an in-house
197 carbonate standard “NAXOS” ($\delta^{13}\text{C} = 2.08 \text{‰}$; $\delta^{18}\text{O} = -6.83 \text{‰}$) and the international carbonate
198 standard IAEA-CO-1. Analytical precision, based on repeated measurements of NAXOS, were better
199 than 0.11 ‰ for $\delta^{18}\text{O}_{\text{carb}}$ and 0.05 ‰ for $\delta^{13}\text{C}_{\text{carb}}$ (1 σ). For 15 samples that required relatively large
200 amounts of sediment (>1000 μg) due to low carbonate content, analytical uncertainty increased up to
201 1 ‰ due to H₂S formation.

202 For bulk organic carbon isotope ratios ($\delta^{13}\text{C}_{\text{TOC}}$), approximately 0.3 gram per sample was
203 powdered, and treated with 25 mL HCl to dissolve the carbonates. Samples were crushed after
204 drying and analyzed on a Thermo Scientific Flash IRMS Elemental Analyzer coupled to a Thermo
205 Scientific Delta V Advantage isotope ratio mass spectrometer. Standard bracketing using the in-
206 house standard nicotinamide ($\delta^{13}\text{C} = -33.08 \text{‰}$) allowed for calibration of delta values relative to the
207 VPDB. Analytical precision, based on repeated measurements of the nicotinamide standard, were
208 smaller than 0.05 ‰ (1 σ).

209 2.3.5 Total organic carbon content and carbonate weight percentage

210 The CaCO₃ content (CaCO₃wt%) was approximated by two independent methods; the first
211 estimation was based on the weight loss after decalcification with HCl, here-after referred to as
212 "CaCO₃wt%_{wl}" (see section 2.3.4). This method presumably slightly overestimates the CaCO₃ content
213 due to removal of additional material (e.g., salts). In the second method, the CaCO₃wt% was
214 approximated based on the relation between the mass spectrometer m/z 44 signal intensity and mass
215 of the weighted sediment sample, in comparison to that of the pure carbonate standards (same

216 approach as described in Fokkema et al., (2022)) here-after referred to as " $\text{CaCO}_3\text{wt}\%_{\text{ms}}$ ". The
217 precision was better than 9.5% (1σ) based on the carbonate standards.

218 The weight percentage of total organic carbon content (TOCwt%) was measured on the EA
219 during $\delta^{13}\text{C}_{\text{TOC}}$ analysis. Analytical precision, based on repeated measurements of the nicotinamide
220 standard, was better than 0.05% (1σ).

221 2.3.6 GDGT analysis

222 For 92 samples, a subsample of ~4–16 g was powdered for the analysis of glycerol dialkyl
223 glycerol tetraethers (GDGTs). Lipids were extracted with a mixture of dichloromethane (DCM) and
224 MeOH (9:1, v/v) using a Milestone Ethos X Microwave extraction system (70 °C; 50 min). An
225 internal standard (99 ng of C_{46} GTGT) was added to allow for quantitative assessment of GDGT
226 concentrations. Subsequently, solvent mixtures of hexane:DCM (9:1), hexane:DCM (1:1) and
227 methanol:DCM (1:1) were used to respectively separate the total lipid extract into apolar, ketone and
228 polar fractions over an activated Al_2O_3 column. The polar fractions, containing the GDGTs, were
229 dried under a gentle N_2 stream, redissolved in hexane:isopropanol (99:1) and filtered through a 0.45
230 μm polytetrafluorethylene filter, after which they were measured on an Agilent 1290 ultra-High
231 Performance Liquid Chromatography (UHPLC) coupled to a Agilent 6135 Mass Spectrometer (MS),
232 following the methodology as described by Hopmans et al. (2016). GDGTs were identified by
233 detecting their $[\text{M}+\text{H}]^+$ ions in selected ion monitoring mode. Quantification was achieved by peak
234 area integration and comparing that with the area of the internal standard, assuming that the response
235 of the MS was similar for all compounds. A minimum signal-to-noise ratio of >3 was maintained as
236 detection limit.

237 GDGT distributions were first tested for non-thermal controls on the TEX_{86} using various
238 published ratios and indices by applying the R script of Bijl et al. (2021), resulting in exclusion of
239 two samples (Supplementary Fig. S1). TEX_{86} values were translated to temperatures using a SST
240 calibration ($\text{TEX}_{86}^{\text{H}}$; Kim et al., 2010) and, because sedimentary GDGTs most likely derive from just
241 below the mixed layer (Massana et al., 2000; Taylor et al., 2013; Rattanasriampaipong et al., 2022),
242 also with a subsurface temperature calibration (SubT) that targets the layer between 100 and 250 m
243 water depth (SubT_{100-250m}; Ho and Laepple, 2016). To properly assess SST variability, we use the
244 range of variability in SST and SubT-calibrated TEX_{86} records to cover the plausible range of depth-
245 depended TEX_{86} -temperature slopes in the modern system (see Fokkema et al. (2023)). Analytical
246 precision, based on repeated integrations of the standard ($n = 4$), were smaller than 0.0015 TEX_{86}
247 values, which translates to 0.06 °C following the $\text{TEX}_{86}^{\text{H}}$ calibration in the range of TEX_{86} values
248 encountered.

249 2.3.7 Palynology

250 Forty-one samples were crushed to ~5 mm size pieces and prepared in the Palynology Lab at
251 Utrecht University, using standard methods (e.g., Cramwinckel et al., (2018)). In short, samples were
252 treated with HCl and HF and residues were sieved over 250- and 15- μm sieves. The addition of a
253 known number of *Lycopodium clavatum* spores prior to processing provides for quantitative
254 assessment (Stockmarr, 1972). Microscope slides were prepared using glycerine jelly. Dinocysts
255 were classified at genus and if possible, at species level, using the taxonomy of that cited in Williams
256 et al. (2017). Taxa were grouped based on inferred ecological niches, following modern (Zonneveld
257 et al., 2013; Thöle et al., 2023) and Paleogene distributions and affinities (Brinkhuis, 1994; Pross and
258 Brinkhuis, 2005; Sluijs and Brinkhuis, 2009; Frieling and Sluijs, 2018).

259 3. Results

260 3.1 Bulk carbonate and bulk organic isotopic compositions

261 3.1.1 Carbon isotope excursion "V"

262 We record a significant CIE in the $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{TOC}}$ data between $\sim 757.2\text{--}756.3$ mbsf
263 (Figs. 2, 3), with magnitudes of 3.4 ‰ and 0.6 ‰, respectively. The timing of recovery to
264 background values differs between $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{TOC}}$, the rise in $\delta^{13}\text{C}_{\text{carb}}$ at 756.36 mbsf precedes
265 that of the $\delta^{13}\text{C}_{\text{TOC}}$ by ~ 55 cm. The $\delta^{18}\text{O}_{\text{carb}}$ data exhibits overall low values (mean = -4.7 ‰) that
266 drop to values as low as -7.1 ‰ during the CIE (Supplementary Fig. S1). Although this negative
267 oxygen isotope excursion might support the presence of a hyperthermal, such low values may also
268 point to secondary overprint of the $\delta^{18}\text{O}$ signals (Schrag et al., 1995). Therefore, we do not further
269 interpret $\delta^{18}\text{O}_{\text{carb}}$ as a paleoclimatic proxy indicator. Collectively, however, this CIE stands out
270 significantly relative to background values suggesting the presence of a globally recorded CIE. The
271 CIE occurs just below the base of *D. subloidoensis*, implying that it most likely represents the "V"
272 CIE at 49.7 Ma (Lauretano et al., 2016) (also termed "C22nH1"; Sexton et al., 2011) (see section 2.2;
273 Fig. 2).

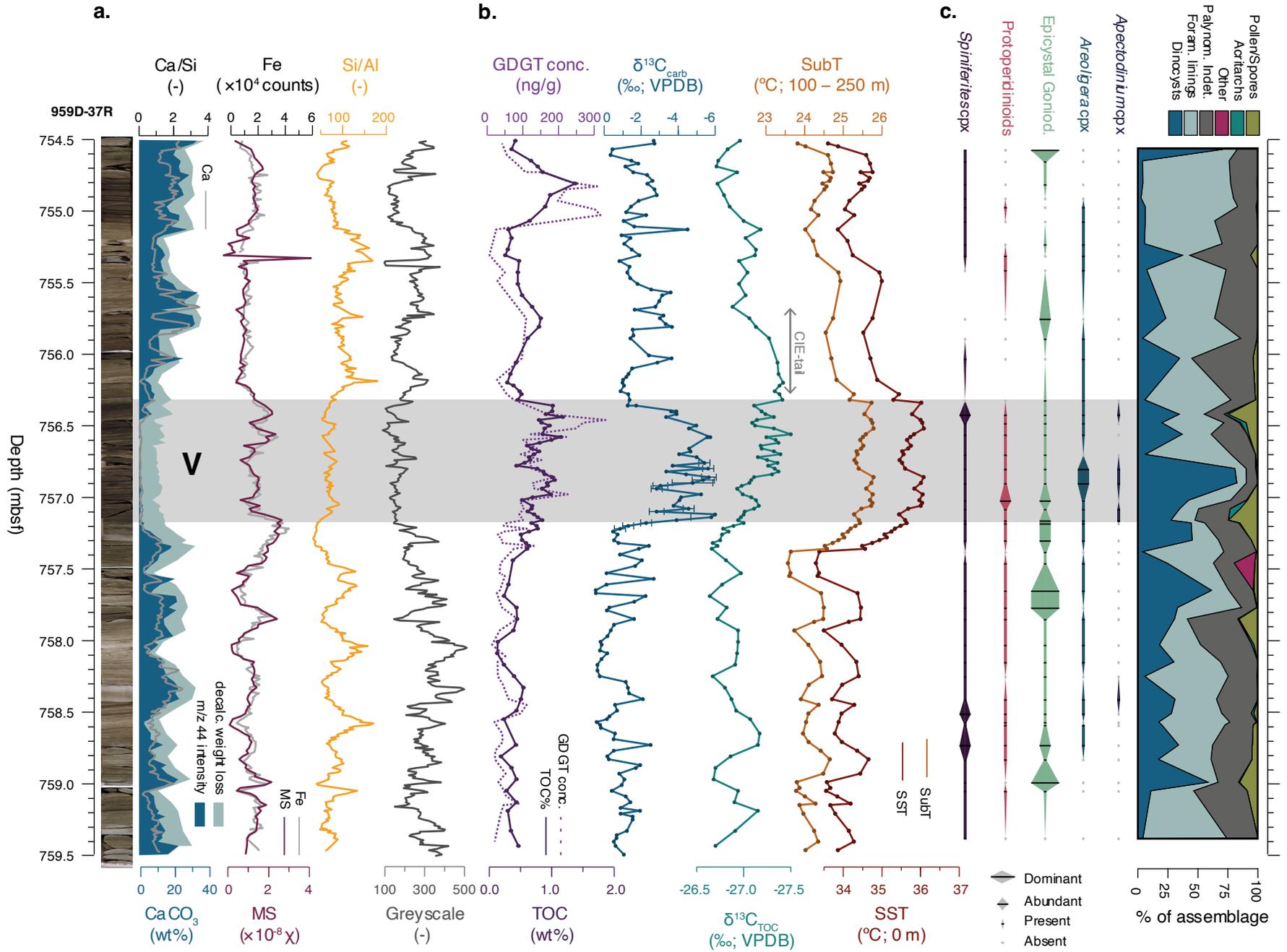
274 3.1.2 Background $\delta^{13}\text{C}$ variability

275 The $\delta^{13}\text{C}_{\text{carb}}$ ranges between ~ -2 and 0 ‰ from the base of Core 37R up to the V event
276 ($759.5\text{--}757.2$ mbsf). Above the event (<756.3 mbsf), $\delta^{13}\text{C}_{\text{carb}}$ values recover to average values
277 between ~ -3 and -1 ‰ (Fig. 3). The $\delta^{13}\text{C}_{\text{TOC}}$ data shows ~ 50 cm background variability between -
278 26.7 ‰ and -27.2 ‰ surrounding V. Within the CIE, values vary between -27.5 and -27.1 ‰.

279 3.2 Sediment composition

280 3.2.1 Core picture and sediment color

281 Site 959D Core 37R is characterized by prominent variations in sediment color, ranging from
282 dark brown to light grey (Fig. 3). Across the complete core, the darker and lighter intervals alternate
283 on submeter scale. An extended dark horizon marks the V event in the middle of the core between
284 757.2 to 756.3 mbsf, and relatively light-colored sediments mark the lower part of the section. One
285 occurrence of a black chert at ~ 755.4 mbsf stands out in the core picture.



287 **Figure 3.** Downcore analysis results for Site 959D Core 37R. **(a)** Calcareous nannofossil
 288 biostratigraphic zonation (data from Fokkema et al. (2023); zonation definitions follows Agnini
 289 et al. (2014)). Two methods for CaCO₃wt% (dark and light blue areas, see section 2.3.5) and
 290 Ca/Si (grey line), bulk magnetic susceptibility (MS) and XRF-based Fe counts, Si/Al, Greyscale.
 291 **(b)** TOCwt% and GDGT concentrations (dotted line), bulk carbonate δ¹³C, total organic δ¹³C,
 292 TEX₈₆-based SubT (orange, following the SubT_{100-250m} calibration; Ho and Laepple, 2016) and
 293 SST (red, following the TEX₈₆^H calibration; Kim et al. 2010). **(c)** Palynology results with semi-
 294 quantitative dinocyst assemblages and the relative abundances of constituents that make up the
 295 complete palynomorph assemblage. Dark grey band indicates the CIE horizon.

296 3.2.2 CaCO₃ and organic carbon content

297 The two independent CaCO₃wt% proxies (light and dark blue areas in Fig. 3a) correlate
 298 well ($R^2=0.87$; $p < 0.001$), but the decalcification-weight-loss method yields 5–10 wt% higher
 299 values (Supplementary Fig. S1), presumably due to the loss of non-CaCO₃ components during
 300 HCl treatment. The CaCO₃wt% (m/z 44 method) displays a continuous ~50 cm variability across
 301 the bottom half of the core, ranging between ~2–24 % and strongly declines to negligible
 302 amounts during the V event. Above this horizon, CaCO₃wt% rises to values of approximately 10
 303 % higher than the lower interval, varying between 10–34 % on meter scale.

304 The TOCwt% varies in-phase with CaCO₃wt% in the top and bottom part of the interval,
 305 ranging between 0.2–0.6% in ~50 cm intervals in the bottom half of the core. In the dark
 306 colored interval marking V, TOCwt% rises to higher values up to 1.2%. Above, the TOCwt%
 307 remains elevated compared to the base of the section and reaches two maxima at 755.75 and
 308 754.75 mbsf, with values of 0.85 and 1.4 %, respectively.

309 3.2.3 Magnetic Susceptibility

310 The MS values (Fig. 3a) range between -3.7×10^{-9} – 3.9×10^{-8} χ . Overall variability in the
 311 MS record corresponds to that of TOCwt%: showing a continuous ~50 cm variability from the
 312 base of the section to 756.36 mbsf, with relatively higher values during V. The ~1m-scale
 313 variability above the V event to the top of the section is relatively dampened. Interestingly, one
 314 high MS peak coincides with the bottom of the identified black chert, immediately followed by
 315 slightly negative MS.

316 3.2.4 X-Ray Fluorescence (XRF)

317 Ca and Si compose the bulk signal intensity of the XRF scan. Ca significantly covariates
 318 with the CaCO₃wt%_{ms} ($R^2 = 0.82$; $p < 0.001$) and CaCO₃wt%_{owl} ($R^2 = 0.73$, $p < 0.001$)
 319 (Supplementary Fig. S1). For most of the record, Si and Ca display a strong anticorrelation
 320 across regular ~50-cm-scale alternations, whereas both elements decline during V (757.4–756.4
 321 mbsf). Interestingly, the regular occurring maxima of Si and Ca are both in the lighter core
 322 intervals. Furthermore, Fe correlates strongly with the MS, Al and K (Supplementary Fig. S1),
 323 representatives of the terrestrially derived clay component.

324 3.3 TEX₈₆ paleothermometry

325 Concentrations of isoprenoid GDGTs range between 6–332 ng/g throughout the core. All
 326 isoprenoid GDGTs relevant for TEX₈₆ paleothermometry were above detection limit in all

327 samples. An average GDGT-2/GDGT-3 ratio of 3.7 indicates predominant GDGT export origin
 328 from the upper top 200 meters of the ocean water column (Taylor et al., 2013; Hurley et al.,
 329 2018; van der Weijst et al., 2022; Rattanasriampaipong et al., 2022). There is no clear correlation
 330 between the TEX₈₆ and the GDGT-2/GDGT-3 ratio ($R^2 = 0.05$; $p = 0.026$). Branched GDGTs are
 331 absent or only present in low concentrations, hence the branched and isoprenoid tetraether (BIT)
 332 index that can indicate potential GDGT inputs from the terrestrial realm (Hopmans et al., 2004)
 333 reflects consistently very low values (<0.01), characteristic for open marine settings. Two
 334 samples show large (>0.3) deviations from the modern Ring Index-TEX₈₆ relationship (Δ -RI;
 335 Zhang et al., 2016) indicating a possible non-thermal overprint on the temperature signal and
 336 were discarded (Supplementary Fig. S2).

337 Reconstructed temperatures change from background SSTs varying between 33.3–34.5
 338 °C (SubT = 23.6–24.5 °C) rising to values varying between 35.5–36.1 °C (SubT = 25.3–25.8 °C)
 339 during the V event, reflecting an averaged warming of 1.1 (SubT)–1.9 °C (SST) (Fig. 3b). The
 340 background SSTs of ~34 °C are close to two previous data points from Cramwinckel et al. (2018)
 341 for this interval at Site 959. After the CIE, background temperatures are on average ~0.5 °C
 342 higher than before. Interestingly, the onset of warming leads the onset of the CIE by ~10 cm. In
 343 addition, the warming includes two warm phases with temperatures up to ~36 °C, separated by a
 344 0.5 °C cooling at the level where carbon isotope ratios reach minimum levels. Finally,
 345 throughout the studied interval, TEX₈₆ shows a continuous ~50 cm variability, which is
 346 dampened during the hyperthermal.

347 3.4 Palynology

348 The palynological associations are mainly composed of dinocysts and remains of the
 349 organic inner-wall ‘linings’ of benthic foraminifera (Fig. 3c). The contribution of terrestrial
 350 elements, (notably various types of pollen and spores) is generally low, with larger numbers,
 351 occasionally up to ~25% in certain samples. These intervals of higher proportions of terrestrial
 352 palynomorphs coincide with higher MS and Fe values, and darker sediment color. Unfortunately,
 353 the overall poor preservation of dinocysts prohibits the confident and complete quantification of
 354 the dinocyst assemblages for most analyzed samples; because of large number of unidentifiable
 355 (fragments of) dinocysts, only 10 samples have more than 30 specimens that could be identified
 356 at family or genus level. Therefore, we discuss the dinocyst assemblages only in terms of broad
 357 qualitative and semi-quantitative results.

358 We do record the near-continuous occurrence of dinocysts that belong to the ecological
 359 groupings of *Areoligera*, *Spiniferites*, Protoperidinoids, Goniodomideae with an epicystal
 360 archeopyle (epicystal Goniodomids) and a rest group of Gonyaulacoids with a precingular
 361 archeopyle (cf. Frieling and Sluijs, 2018). In addition, many of the unidentified dinocysts
 362 represent skolochorate dinocysts probably of Gonyaulacoid affinity. The occurring
 363 Protoperidinoid cysts, notably *Selenopemphix* spp. and *Lejeunecysta* spp., represent obligate
 364 heterotrophic dinoflagellates today (e.g., Zonneveld et al., 2013). The identified epicystal
 365 Goniodomids, which in the modern ocean are typical of lagoonal, hypersaline settings
 366 (Zonneveld et al., 2013) are notably *Homotryblium* spp. and have been linked to strong
 367 stratification (and even harmful algal blooms in some cases) in the paleodomain (Brinkhuis,
 368 1994; Pross and Schmiedl, 2002; Reichart et al., 2004; Cramwinckel et al., 2019). In effect, a
 369 major signal in the dinocysts assemblages is the sporadic dominance of *Homotryblium* spp.,
 370 reaching above 90% of the identified component of the assemblages. In the middle part of the

371 core, we record a few more numerous and/or prolonged occurrences of representatives of
 372 *Apectodinium*, *Areoligera*, Protoperidinioids and a single occurrence of a representative of
 373 *Senegalinum*.

374 4. Discussion

375 4.1 Carbon isotope magnitude

376 The magnitude of the V-related CIE differs strongly between the bulk organic and
 377 carbonate records at Site 959 ($\Delta^{13}\text{C}_{\text{carb}} = 3.4 \text{ ‰}$, $\Delta^{13}\text{C}_{\text{TOC}} = 0.7 \text{ ‰}$). Importantly, the $\delta^{13}\text{C}_{\text{carb}}$
 378 excursion is much larger than what is expected from the global exogenic signal, approximately
 379 $\sim 0.6 \text{ ‰}$ (Fig. 4) (Sexton et al., 2011; Lauretano et al., 2016). We surmise that in the sedimentary
 380 environment of early Eocene Site 959, with relatively low biogenic CaCO_3 accumulation,
 381 authigenic carbonates, which are typically ^{13}C -depleted due to local organic carbon respiration
 382 (e.g., Botz et al., 1988), comprise a significant component of the bulk carbonate, similar to
 383 observations during earlier Eocene CIEs I1–L2 at Site 959 (Fokkema et al., 2023) and consistent
 384 with pore-water analysis (Masclé et al., 1996). Low $\delta^{13}\text{C}_{\text{carb}}$ values typically correspond with low
 385 $\text{CaCO}_3\text{wt}\%$ in lower Eocene deep ocean records (Zachos et al., 2010; Kirtland Turner et al.,
 386 2014). Consequently, this would exaggerate both background variability and the magnitude of
 387 the CIE in our record due to an increased fraction of authigenic carbonate in intervals with
 388 relatively low biogenic $\text{CaCO}_3\text{wt}\%$.

389 4.2 Paleoenvironment and water column structure

390 4.2.1 Background climate state and oceanography

391 An open marine early Eocene depositional setting at Site 959 is reflected by the very low
 392 BIT index values (< 0.01 ; Supplementary Fig. S2), generally low numbers of terrestrial
 393 palynomorphs (mean = 4.8 %) (Fig. 3a), and bulk $\delta^{13}\text{C}_{\text{TOC}}$ values around -27 ‰ , indicative of
 394 marine particulate organic carbon for this time interval (Hayes et al., 1999; Sluijs and Dickens,
 395 2012). Also the dinocyst assemblages are consistent with an open ocean setting with the
 396 continuous abundances of *Spiniferites* and particularly a few representatives of the open-ocean
 397 genus *Impagidinium* (Frieling and Sluijs, 2018). In contrast with these records, multiple samples
 398 are dominated by epicystal Goniodomid-cysts, specifically *Homotryblium* spp. ($> 75\%$ of the
 399 relative dinocyst assemblage), traditionally interpreted to reflect lagoonal, hypersaline conditions
 400 (Brinkhuis, 1994; Pross and Schmiedl, 2002). However, dominances of such types of dinocysts
 401 have also been recorded under unequivocal open-ocean conditions, including the Eocene Site
 402 959 (Frieling et al., 2018; Cramwinckel et al., 2019; Fokkema et al., 2023), attributed to so-
 403 called ‘hyper-stratified conditions’—potentially under high temperature (Cramwinckel et al.,
 404 2019), analogous to records from the ‘hyperstratified’ episodes during the Pleistocene of the
 405 Arabian Sea (cf. Reichart et al., 2004), which enables a lifecycle without a sea floor interface. It
 406 is possible that the acmes we record here indeed reflect actual ancient dinoflagellate blooms,
 407 which is common in modern representatives of this group, e.g., *Pyrodinium bahamense* and its
 408 cyst *Polysphaeridium zoharii* (Usup et al., 2012). In the paleo-domain, such a bloom-forming
 409 strategy of this group is supported by very high $\delta^{13}\text{C}$ values of *Eocladopyxis peniculata* cysts in
 410 an acme during the PETM on the New Jersey shelf (Sluijs et al., 2018). Hence, given the absence
 411 of evidence for large-scale transport of organic matter off the shelf, we suggest that the
 412 *Homotryblium* acmes/blooms at Site 959 occurred under warm and strongly stratified oceanic

413 conditions. We postulate that the few intervals of relatively high contribution of terrestrial
 414 palynomorphs (Fig. 3c) (partly) result from preservation bias, as terrestrial palynomorphs are
 415 more resistant to degradation than dinocysts.

416 The consistent presence of *Protoperidinium* cysts, derived from heterotrophic
 417 dinoflagellates, implies sufficient food supply (Zonneveld et al., 2013). Moreover, sufficient
 418 organic material must have reached the ocean floor to sustain the relatively high TOCwt% and
 419 abundance of benthic foraminifera linings. The required high primary productivity presumably
 420 originated from regional wind-driven upwelling, inferred to occur at least since the middle
 421 Eocene based on organic geochemical data and palynological data (Wagner, 2002; Cramwinckel
 422 et al., 2018). Importantly, this productive setting around 49.7 Ma contrasts the records from ~54–
 423 52 Ma, where rich dinocyst assemblages lack Protoperidinioids (Fokkema et al., 2023),
 424 suggesting that Eocene upwelling at Site 959 did not intensify until the end of the early Eocene
 425 (i.e., ~52–50 Ma). Neogene upwelling conditions at this site have previously been linked to the
 426 annual migration of the Atlantic Intertropical Convergence Zone (ITCZ) and the resulting West-
 427 African monsoonal response (Vallé et al., 2017; Wubben et al., in press). Simulations suggest
 428 that West African monsoons should already have established during the Cretaceous and early
 429 Paleogene, and were relatively strong in this time interval compared to the Miocene due to the
 430 narrower Atlantic basin (Acosta et al., 2022). Based on the dominant factors enhancing the
 431 strength of monsoons, e.g., geography and topography (Acosta et al., 2022), we infer that
 432 upwelling potentially started or intensified during the Eocene by progressive northward
 433 movement of the African plate, drifting Site 959 towards the equator.

434 4.2.2 Orbital variability

435 The regular alternations observed in multiple proxy records (e.g., CaCO₃ content, MS,
 436 Fe, TOC%, total GDGT concentration), specifically on ~50-cm scale below the CIE (759.5–
 437 757.4 mbsf), suggests an orbital influence on the local oceanographic conditions and/or monsoon
 438 strength, analogous to that recorded in the Cretaceous (Beckmann et al., 2005) and Miocene
 439 (Wubben et al., in press) at Site 959. As there is only one age-depth tiepoint in the studied core
 440 section, the responsible orbital cycle cannot be confidently identified. Although sedimentation
 441 rates of the studied section are uncertain, the long-term background sedimentation rates of 0.8–
 442 1.4 cm/kyr (Fokkema et al., 2023) imply that the persistent 50-cm variability would approximate
 443 the 41-kyr obliquity band. However, obliquity has neglectable direct influence on equatorial
 444 insolation, whereas variability in the strength of the monsoons is usually associated with Earth's
 445 ~20-kyr precession cycles, and only obliquity when remotely forced (e.g., Tuenter et al., 2003).
 446 On the other hand, dominant obliquity forcing has been observed in low latitudes (Weedon et al.,
 447 1997) and exerts influence on Neogene Mediterranean sapropel occurrences and monsoon
 448 dynamics (e.g., Lourens et al., 1996; Bosmans et al., 2015). In addition, obliquity is also possibly
 449 more pronounced under very low orbital eccentricity configurations (Westerhold et al., 2014), as
 450 also recorded in the early Miocene at Site 959 (Wubben et al., 2023). Indeed, a ~2.4-Myr orbital-
 451 eccentricity minimum just precedes the V event (Laskar et al., 2011; Lauretano et al., 2018) (Fig.
 452 4), supporting the possibility of obliquity forcing on monsoonal strength prior to V.
 453 Nevertheless, without additional age control, there remains a possibility that the cyclic variations
 454 prior to V represent precession.

455 The cycles below the V event coincide with temperature variability of 0.8–1 °C. This
 456 variability is mirrored in $\delta^{13}\text{C}_{\text{TOC}}$, with minimum TEX₈₆ values coinciding with $\delta^{13}\text{C}_{\text{TOC}}$ minima,

457 which suggests that during enhanced upwelling of ^{13}C depleted subsurface waters, the shallow
 458 waters cooled and vice versa (Fig. 3). No clear cyclic patterns are present in the interval above V.
 459 Importantly, the absence of the W CIE, which occurred ~ 100 kyr after V, implies at least a
 460 doubling of sedimentation rate if the pre-event variations represent obliquity. Accordingly, a
 461 relative increase in, CaCO_3 , Si and TOCwt% (Fig. 3a) suggest higher productivity and biogenic
 462 sediment accumulation rates compared to the lower interval, potentially facilitating the required
 463 increase if the supply of siliciclastic materials remained constant.

464 4.2.3 Environmental change during hyperthermal V

465 The peak SSTs reached during the V event of $36.1\text{ }^\circ\text{C}$ are consistent with the
 466 hyperthermals in the earlier interval of the Eocene (Fokkema et al., 2023). The magnitude of
 467 warming ($\Delta T = 1.1\text{--}1.9\text{ }^\circ\text{C}$) is, however, higher than recorded during the preceding early Eocene
 468 hyperthermals at Site 959 (averaged $\sim 0.7\text{--}1\text{ }^\circ\text{C}$; (Fokkema et al., 2023)). Peak warmth during
 469 hyperthermal V manifests in two peaks (757.0 & 756.5) separated by a plateau. We surmise that
 470 this reflects a continuation of the pre-event orbitally forced variability in monsoon strength.
 471 Accordingly, the two temperature peaks are marked by relatively high $\delta^{13}\text{C}_{\text{TOC}}$ values,
 472 suggesting they represent periods of minimal upwelling, similar to the intervals of higher
 473 $\delta^{13}\text{C}_{\text{TOC}}$ and SST surrounding the event.

474 In addition to warming, the V event is characterized by increased TOCwt%, Fe, MS and a
 475 decline in the CaCO_3 wt% and Si in the sediment. Frieling et al. (2018) discussed multiple
 476 mechanisms that might have induced low CaCO_3 wt% contents in PETM records at Site 959.
 477 Ocean acidification may well have impacted carbonate preservation, but this may have been
 478 limited, due to the relatively shallow paleodepth (~ 1000 m) of Site 959 during the early Eocene
 479 (Frieling et al., 2018). However, similar to the PETM interpretation by Frieling et al. (2018), we
 480 infer that the low CaCO_3 wt% during the CIE can be attributed to remineralization of organic
 481 material in the sediment and possibly also to attenuation of biogenic CaCO_3 formation—
 482 potentially by heat stress (Aze et al., 2014; Frieling et al., 2018).

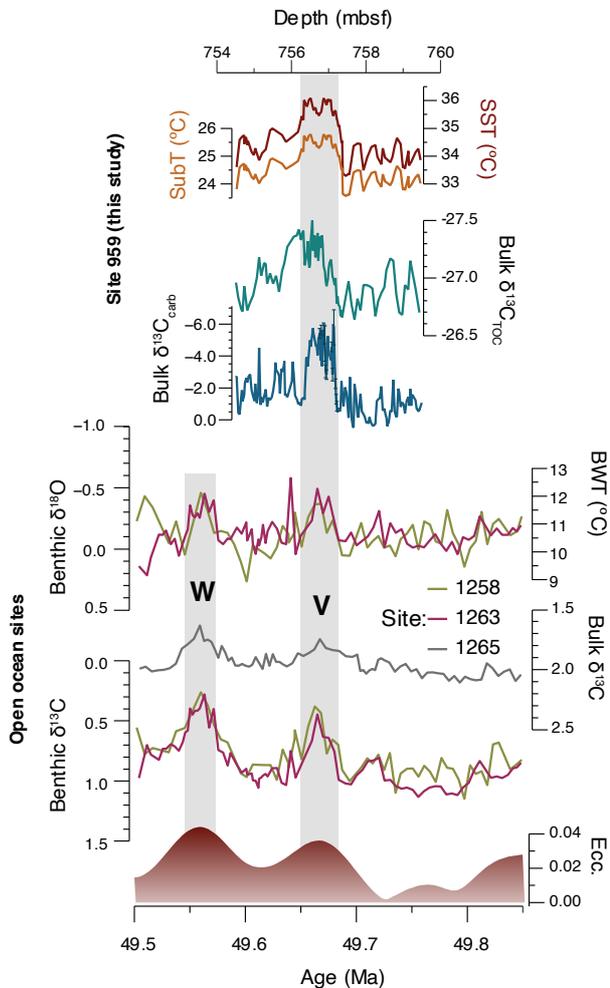
483 Although — as discussed above — dinocyst preservation being poor in general, we do
 484 record a subtle dinocyst response during hyperthermal V. Notably, few occurrences of the
 485 thermophilic dinocyst taxon *Apectodinium* are recorded during the hyperthermal (Fig. 3c).
 486 Representatives of this genus are known to occur in high numbers surrounding the PETM
 487 interval at Site 959 (Frieling et al., 2018) (and globally during the PETM, e.g., Crouch et al.,
 488 2001; Denison, 2021). Presumably other factors than solely temperature allowed for the
 489 occurrence of *Apectodinium* spp. to Site 959; perhaps the supply of nutrients. Another feature is
 490 abundant *Areoligera* spp. at 756.8 mbsf. This peak coincides with the highest relative abundance
 491 of dinocysts ($>75\%$), and elevated TOCwt% (Fig. 2). Although perhaps masked by poor
 492 preservation, the epicystal Goniodomid-cysts do not seem to significantly increase during the
 493 event, conflicting with the occurrence of (thermal) hyperstratification during the peak MECO
 494 (Cramwinckel et al., 2019).

495 4.3 Polar amplification of climate change

496 Previous work has compared the warming of early Eocene hyperthermals in the tropics to
 497 that of higher latitudes to assess ice-free polar amplification of climate change, finding a relative
 498 consistent amplification factor of approximately ~ 2 (Frieling et al., 2017; Tierney et al., 2022;
 499 Fokkema et al., 2023). The only temperature and ecological records from the high latitudes that

500 cover internal climate variability in the EECO (i.e. IODP Site 1356, Wilkes Land) are still not
 501 fully resolved on eccentricity timescales (Pross et al., 2012; Bijl et al., 2013). Therefore, to
 502 assess potential extratropical amplification of warming during hyperthermal V, we compare the
 503 warming at Site 959 with that in the deep ocean benthic foraminiferal $\delta^{18}\text{O}$ -based bottom water
 504 temperature (BWT) records, following the approach of Cramwinckel et al. (2018) and Fokkema
 505 et al. (2023). Importantly, the $\delta^{18}\text{O}$ -based BWTs match the pollen-derived coldest month
 506 temperatures at Wilkes Land (Pross et al., 2012), in line with the notion that early Eocene bottom
 507 water formation occurred dominantly in the Southern Ocean close to the Antarctic margin
 508 (Hollis et al., 2012; Zhang et al., 2022). In the deep ocean, hyperthermal V was marked by a
 509 warming of $\sim 1.5\text{ }^\circ\text{C}$ (Lauretano et al., 2018) (Fig. 4). This implies a near-absent amplification of
 510 warming between the tropics and the polar region of deep-water formation (0.8–1.4), which
 511 mismatches the polar amplification recorded during multiple early EECO hyperthermals (1.7–
 512 2.2) (Fokkema et al., 2023) and during the PETM $\sim 1.6\text{--}2.7$ (Frieling et al., 2017; Tierney et al.,
 513 2022). We surmise that the upwelling at Site 959 during this interval led to reduced polar
 514 amplification estimates, possibly due to a decrease in upwelling intensity during the V event,
 515 amplifying surface warming at Site 959.

516



517

518 **Figure 4.** Representation of the V event at Site 959 and other Atlantic Ocean sites. Top: Site 959
 519 data (against depth) with TEX₈₆-based SST and SubT, $\delta^{13}\text{C}_{\text{TOC}}$ and $\delta^{13}\text{C}_{\text{carb}}$. Bottom: benthic
 520 $\delta^{18}\text{O}$ from Site 1258 (Sexton et al., 2011) and Site 1263 (Lauretano et al., 2016, 2018) with $\delta^{18}\text{O}$ -
 521 based bottom water temperatures on the right (following (Kim and O'Neil, 1997), modified by
 522 (Bemis et al., 1998) and seawater $\delta^{18}\text{O}$ of -1‰ VPDB). Bulk $\delta^{13}\text{C}$ from Site 1265 (Westerhold et
 523 al., 2017), benthic $\delta^{13}\text{C}$ of Site 1258 and 1263 and orbital eccentricity (Laskar et al., 2011). Grey
 524 bands mark globally recognized CIEs V and W.

525

526 4.4 Prolonged organic carbon isotope excursion

527 The recovery of the CIEs in the $\delta^{13}\text{C}_{\text{TOC}}$ and $\delta^{13}\text{C}_{\text{carb}}$ are significantly offset; $\delta^{13}\text{C}_{\text{TOC}}$ lags
 528 $\delta^{13}\text{C}_{\text{carb}}$ by 55 cm, which represents ~40–70 kyr, following long-term background sedimentation
 529 rates of 0.8–1.4 cm/kyr (calculated between Top *Tribarchaitus orthostylus* (Core 36R) and Top
 530 *Discoaster lodoensis* (Core 39R–Core 38R); Fig. 2). This is inconsistent with the premise that
 531 carbon utilized in photosynthesis and calcification are both fixed from the same dissolved
 532 inorganic carbon (DIC) pool. It suggests that an organic matter component with low $\delta^{13}\text{C}$ values
 533 contributes to sedimentary $\delta^{13}\text{C}_{\text{TOC}}$ values in the direct aftermath of the CIE. Potentially, the
 534 longer tail in the $\delta^{13}\text{C}_{\text{TOC}}$ record represents terrestrial organic carbon produced during the CIE,
 535 transported to the marine realm after the event. However, this is inconsistent with the relatively
 536 short (~5 kyr) time between the production of terrestrial organic matter and its transport to the
 537 ocean (Cole and Caraco, 2001) and the consistently low relative amounts of terrestrial organic
 538 matter in our materials.

539 In the modern ocean, the oldest (most ^{14}C -depleted) phase of carbon resides as
 540 (refractory) dissolved organic carbon (DOC_{ref}) in the Pacific Ocean at 2 km water depth, with an
 541 estimated age of 12 kyrs (Follett et al., 2014). Potentially, the long bulk $\delta^{13}\text{C}_{\text{TOC}}$ tail thus reflects
 542 contributions from this older DOC_{ref} pool in the aftermath of V. This chemically inert phase of
 543 DOC is likely generated by the microbial carbon pump (Jiao et al., 2010). The bioavailability of
 544 DOC_{ref} increases by upwelling (Fang et al., 2020), which allows for a biotic conversion of ^{13}C -
 545 depleted DOC_{ref} to ^{13}C -depleted POC (Shen and Benner, 2018). Specifically, DOC_{ref} is suggested
 546 to physically disintegrate by radiation at the surface ocean and biologically degrade by the
 547 addition of more readily degradable organic matter (Shen and Benner, 2018; Fang et al., 2020).
 548 Furthermore, based on the presence of sponge spicules in younger intervals at Site 959 (Masclé
 549 et al., 1996), while they would not be well preserved in the Eocene porcellanites, this site might
 550 facilitate the DOC to POC conversion by benthic filter feeding sponges. We surmise that Site
 551 959 holds the potential to transform $\delta^{13}\text{C}$ depleted DOC_{ref} to POC, either at the surface and/or the
 552 deep ocean, which will eventually leave an imprint in the sediment record as a prolonged
 553 $\delta^{13}\text{C}_{\text{TOC}}$ tail.

554 Conclusions

555 Tropical warming and biotic change accompanied a significant CIE during the end of the
 556 EECO in the tropical Atlantic at Site 959. By previously published biostratigraphic constraints
 557 and new carbon isotope stratigraphy the CIE is identified as the "V" event at 49.7 Ma. TEX₈₆
 558 paleothermometry indicates a temperature increase of 1.1–1.9 °C relative to pre-event values,
 559 confirming that this event represents a transient global warming event, or hyperthermal. Peak
 560 temperatures during V (SSTs of ~36.1 °C) match that of earlier hyperthermals during the EECO.

561 Interestingly, the recovery of the CIE in $\delta^{13}\text{C}_{\text{TOC}}$ is delayed relative to that in $\delta^{13}\text{C}_{\text{carb}}$. We
 562 postulate that upwelling of old and deep (^{13}C -depleted) DOC_{ref} contributed to the long bulk
 563 $\delta^{13}\text{C}_{\text{TOC}}$ tail.

564 The recorded background paleoceanography during this time differs from earlier Eocene
 565 records from the same site: wind-driven upwelling facilitated a more productive depositional
 566 setting reflected in relatively organic rich sediments and presence of heterotrophic dinoflagellate
 567 cyst remains. Variations in upwelling intensity, presumably orbitally driven by monsoons, are
 568 reflected in multiple proxy indicators: notably affecting carbon isotopes and temperatures in the
 569 (shallow sub-) surface ocean. In strong contrast to the recorded upwelling is the sporadic
 570 dinocyst dominance by hyper-stratification-related genera, which seems a common feature in
 571 Eocene Site 959 records (Frieling et al., 2018; Cramwinckel et al., 2019; Fokkema et al., 2023).

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582 **Competing interests**

583 One of the co-authors is a member of the editorial board of *Paleoceanography and*
 584 *Paleoclimatology*.

585 **Open Research**

586 All new data presented in this work will be made publicly available on Zenodo upon
 587 publication.

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