Dear Climate of the Past Editor,

First of all, we warmly thank you for the observations that further improve our manuscript. Please find the uploaded manuscript and figures corrected with suggestions.

Specifically:
- We changed throughout the text *Morozovella* and *Acarinina* instead of morozovellids and acarininids. Micropaloentologists dealing with planktic foraminifera are used to discussing genera with 'ids'. (We could cite lots of papers by Wade, Premoli Silva, Boersma, Pearson, etc.). In fact, the two referees B. Wade and P. Pearson did not object to this issue. The 'ids', termination (not Latin) is not included within the Zoological code, and we suspect an origin to avoid the frequent use of italics. However, we realize that the double terminology may be confusing for non-foraminiferal specialists and we have amended the text and Figure 1 accordingly.

- While no referees pointe the problem, we have added in Figure 1 a line in the foraminiferal zonation column since we omitted a zonal boundary (between P4c and P5).

- We added a sentence about the complexity to identify the EECO termination.

-We changed figure 2 updating plaeolatitudes for Sites 577, 1258 and 1051 (Van Hinsbergen et al., 2015). The Possagno paleolatitude is still according to

http://www.odsn/de/services/paleomap.html since it is located in an unconstrained plate and the updated paleolatitude it is not yet available at www.paleolatitude.org.

- We changed ELMO with Elmo (as it is usually cited in the literature).
Going back to the history of this curious name, we find that the original given name is the acronym of Eocene Layer of Mysterious Origin (Leg 208, Preliminary report, 2003; p. 35).
In truth, we believe that good colleagues firstly thought of the red doll Sesame street character Elmo and then they found an appropriate acronym. Nonetheless, we modified.

Finally, we followed all remaining minor corrections suggested. Changes/corrections are in red color.

Sincerely,

Valeria Luciani on behalf of co-authors

1 **Major perturbations in the global carbon cycle and photosymbiont-bearing**

2 **planktic foraminifera during the early Eocene**

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19   **Abstract.** A marked switch in the abundance of the planktic foraminiferal genera

20   *Morozovella* and *Acarinina* occurred at low-latitude sites near the start of the Early Eocene

21   Climatic Optimum (EECO), a multi-million-year interval when Earth surface temperatures

22   reached their Cenozoic maximum. Stable carbon and oxygen isotope data of bulk sediment

23   are presented from across the EECO at two locations: Possagno in northeast Italy, and DSDP

24   Site 577 in the northwest Pacific. Relative abundances of planktic foraminifera are presented

25   from these two locations, as well as from ODP Site 1051 in the northwest Atlantic. All three

26   sections have good stratigraphic markers, and the δ^{13}C records at each section can be

27   correlated amongst each other and to δ^{13}C records at other locations across the globe. These

28   records show that a series of negative carbon isotope excursions (CIEs) occurred before,

29   during and across the EECO, which is defined here as the interval between the J event and the

30   base of *Discoaster sublodoensis*. Significant though ephemeral modifications in planktic

31   foraminiferal assemblages coincide with some of the short-term CIEs, which were marked by

32   increases in the relative abundance of *Acarinina* ~~acarininids~~, similar to what happened across

33   established hyperthermal events in Tethyan settings prior to the EECO. Most crucially, a

34   temporal link exists between the onset of the EECO, carbon cycle changes during this time,

35   and the decline of ~~morozovellids~~ *Morozovella*. Possible causes are multiple, and may include

36   temperature effects on photosymbiont-bearing planktic foraminifera and changes in ocean

37   chemistry.

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1 Introduction

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48   Cenozoic Earth surface temperatures attained their warmest long-term state during the Early

49   Eocene Climatic Optimum (EECO). This was a 2-4 Myr time interval (discussed below)

50   centered at ca. 51 Ma (**Figure 1**), when average high latitude temperatures exceeded those at

51   present-day by at least 10°C (Zachos et al., 2008; Bijl et al., 2009; Huber and Caballero,

52   2011; Hollis et al., 2012; Pross et al., 2012; Inglis et al., 2015). Several short-term (<200 kyr)

53   global warming events (**Figure 1**) occurred before the EECO. The Paleocene Eocene

54   Thermal Maximum (PETM) provides the archetypical example: about 55.9 Ma

55   (Vandenberghe et al., 2012; Hilgen et al., 2015) temperatures soared an additional 5-6°C

56   relative to background conditions (Sluijs et al., 2006, 2007; Dunkley Jones et al., 2013).

57   Evidence exists for at least two other significant Eocene warming events (Cramer et al., 2003;

58   Lourens et al., 2005; Röhl et al., 2005; Thomas et al., 2006; Nicolo et al., 2007; Agnini et al.,

59   2009; Coccioni et al., 2012; Lauretano et al., 2015; Westerhold et al., 2015): one ca. 54.1 Ma

60   and named H-1 or Eocene Thermal Maximum 2 (ETM-2, also referred as the ~~ELMO~~ Elmo

61   event), and one at 52.8 Ma and variously named K, X, or ETM-3 (hereafter called K/X).

62   However, additional brief warming events may have spanned the early Eocene (above

63   references; Kirtland-Turner et al., 2014), and the EECO may comprise a series of successive

64   events (Slotnick et al., 2012). Both long-term and short-term intervals of warming

65   corresponded to major changes in global carbon cycling, although the precise timing between

66   these parameters remains insufficiently resolved.

67   In benthic foraminiferal stable isotope records for the early Paleogene (**Figure 1**), $\delta^{18}O$

68   serves as a proxy for deep-water temperature, while $\delta^{13}C$ relates to the composition of deep-

69   water dissolved inorganic carbon (DIC). The highest $\delta^{13}C$ values of the Cenozoic occurred at

70   ca. 58 Ma. From this Paleocene Carbon Isotope Maximum (PCIM), benthic foraminiferal

71 $\delta^{13}C$ values plunge by approximately 2.5 ‰ to reach a near Cenozoic minimum at or near the

72 start of the EECO, and subsequently rise by approximately 1.5 ‰ across this interval

73 (Shackleton and Hall, 1984; Shackleton, 1986; Zachos et al., 2001, 2008; Cramer et al.,

74 2009). Benthic foraminiferal $\delta^{13}C$ records also exhibit prominent negative carbon isotope

75 excursions (CIEs) across the three hyperthermals mentioned above (Kennett and Stott, 1991;

76 Littler et al., 2014; Lauretano et al., 2015). Crucially, at least from the late Paleocene to the

77 start of the EECO, similar $\delta^{13}C$ records occur in other carbon-bearing phases, such as bulk

78 marine carbonate, planktic foraminifera, and various marine and terrestrial organic carbon

79 compounds (Shackleton, 1986; Schmitz et al., 1996; Lourens et al., 2005; Nicolo et al., 2007;

80 Agnini et al., 2009, submitted; Leon-Rodriguez and Dickens, 2010; Abels et al., 2012;

81 Coccioni et al., 2012; Sluijs and Dickens, 2012; Slotnick et al. 2012, 2015a; Clyde et al.,

82 2013). This strongly suggests that observed changes in $\delta^{13}C$, both long-term trends as well as

83 short-term perturbations, represent variations in the input and output of ^{13}C-depleted carbon

84 to the exogenic carbon cycle (Shackleton, 1986; Dickens et al., 1995; Dickens, 2000; Kurtz et

85 al., 2003; Komar et al., 2013).

86   Significant biotic changes occur in terrestrial and marine environments during times

87 when the early Paleogene $\delta^{18}O$ and $\delta^{13}C$ records show major variations. This has been

88 recognized for the PETM, where land sections exhibit a prominent mammal turnover

89 (Gingerich 2001, 2003; McInerney and Wing, 2011; Clyde et al., 2013), and where marine

90 sections reveal a profound benthic foraminiferal extinction (Thomas, 1998), turnovers in

91 calcareous nannoplankton, ostracods, corals and larger benthic foraminifera (Raffi and De

92 Bernardi, 2008; Scheibner and Speijer, 2008; Yamaguchi and Norris, 2012; Agnini et al.,

93 2014), and appearances of excursion taxa in calcareous nannoplankton, dinoflagellates and

94 planktic foraminifera (Kelly et al., 1996, 1998; Crouch et al., 2001; Sluijs et al., 2006; Self-

95 Trail et al., 2012). Major plant and mammal turnovers also occurred on land during the longer

4

96   EECO (Wing et al., 1991; Zonneveld et al., 2000; Wilf et al., 2003; Falkowski et al., 2005;

97   Woodbourne et al., 2009; Figueirido et al., 2012). In the marine realm, evolutionary trends

98   across the EECO have been noted, in particular the inception of modern calcareous

99   nannofossil community structure (Agnini et al., 2006, 2014; Schneider et al., 2011; Shamrock

100 et al., 2012) and possibly the same for diatoms (Sims et al., 2006; Oreshkina, 2012). These

101 observations, both from continents and the oceans, support an overarching hypothesis that

102 climate change drives biotic evolution, at least in part (Ezard et al., 2011).

103   Planktic foraminiferal assemblages are abundant in carbonate bearing marine sediments

104 and display distinct evolutionary trends that often can be correlated to climate variability

105 (Schmidt et al., 2004; Ezard et al., 2011; Fraass et al., 2015). This is especially true in the

106 early Paleogene, even though the relationship between climate variability and planktic

107 foraminiferal evolution remains insufficiently known. At the beginning of the Eocene,

108 planktic foraminifera had evolved over ca. 10 Myr following the Cretaceous-Paleogene mass

109 extinction event. Several early Paleogene phylogenetic lines evolved, occupying different

110 ecological niches in the upper water column. Subsequently, a major diversification occurred

111 during the early Eocene, which resulted in a peak of planktic foraminiferal diversity during

112 the middle Eocene (Norris, 1991; Schmidt et al., 2004; Pearson et al., 2006; Aze et al., 2011;

113 Ezard et al., 2011; Fraass et al., 2015).

114   In this study, we focus on the evolution of two planktic foraminiferal genera:

115 ~~morozovellids~~ *Morozovella* and *Acarinina* ~~acarininids~~ (**Figure 1**). These two genera belong

116 to the "muricate group", a term derived from the muricae that form layered pustules on the

117 test wall. These two genera are of particular interest because of their dominance among

118 tropical and subtropical assemblages of the early Paleogene oceans, and because these genera

119 show a major turnover in taxonomic diversity close to the beginning of the EECO, one that

120 comprises species reduction among ~~morozovellids~~ *Morozovella* and species diversification

121   among *Acarinina* ~~acarininids~~ (Lu and Keller, 1995; Lu et al., 1998; Pearson et al., 2006; Aze

122   et al., 2011).

123   Numerous lower Eocene sedimentary sections from lower latitudes contain well-

124   recognizable ~~preserved~~ (albeit often recrystallized) planktic foraminiferal tests. Changes in

125   foraminiferal assemblages presumably reflect relationships between climate and carbon

126   cycling across the EECO. The present problem is that no section examined to date provides

127   counts of foraminiferal assemblages, detailed stable isotope records and robust planktic

128   foraminiferal biostratigraphies across the entire EECO. Indeed, at present, only a few sites

129   have detailed and interpretable stable isotope records across much of the EECO (Slotnick et

130   al., 2012, 2015a; Kirtland-Turner et al., 2014). Furthermore, the EECO lacks formal

131   definition. As a consequence, any relationship between climatic perturbations during the

132   EECO and the evolution of planktic foraminifera remains speculative. Here, we add new data

133   from three locations: the Possagno section from the western Tethys, DSDP Site 577 from the

134   tropical Pacific Ocean, and ODP Site 1051 from the subtropical Atlantic Ocean **(Figure 2)**.

135   These sections hence represent a wide longitudinal span of low latitude locations during the

136   early Paleogene. By comparing stable isotope and planktic foraminiferal records at these

137   three locations, we provide a new foundation for understanding why the abundances of

138   *Acarinina* ~~acarininids~~ and *Morozovella* ~~morozovellids~~ changed during the EECO.

139

140   **2 The Early Eocene Climatic Optimum**

141

142   Evidence for extreme Earth surface warmth during a multi-million year time interval of the

143   early Eocene is overwhelming, and comes from many studies, utilizing both marine and

144   terrestrial sequences, and both fossil and geochemical proxies (Huber and Caballero, 2011;

145   Hollis et al., 2012; Pross et al., 2012). However, a definition for the EECO, including the

146 usage of "optimum", endures as a perplexing problem. This is for several reasons, including

147 the basic facts that: (i) proxies for temperature should not be used to define a time increment,

148 (ii) clearly correlative records across the middle of the early Eocene with temporal resolution

149 less than 50 kyr remain scarce, and (iii) absolute ages across the early Eocene have changed

150 significantly (Berggren et al., 1995; Vandenberghe et al., 2102). As a consequence, various

151 papers discussing the EECO give different ages and durations spanning from 2 to 4 Myr long

152 sometime between circa 49 and 54 Ma (e.g., Yapp, 2004; Lowenstein and Demicco, 2006;

153 Zachos et al., 2008; Woodburne et al., 2009; Bijl et al., 2009; Smith et al., 2010; Hollis et al.,

154 2012; Slotnick et al., 2012; Puljalte et al., 2015).

155   The EECO, at least as presented in many papers, refers to the time of minimum $\delta^{18}O$

156 values in "stacked" benthic foraminifera stable isotope curves (**Figure 1**). These curves were

157 constructed by splicing together multiple $\delta^{18}O$ records generated at individual locations onto

158 a common age model (originally Berggren et al., 1995). However, the stacked curves (Zachos

159 et al., 2001, 2008; Cramer et al., 2009), while they can be adjusted to different time scales,

160 show significant variance in $\delta^{18}O$ across the middle to late early Eocene. Some of this

161 variance belies imprecisely calibrated records at individual sites, where cores do not align

162 properly in the depth domain (Dickens and Backman, 2013). Some of this variance probably

163 reflects a dynamic early Eocene climate regime, where average temperatures and atmospheric

164 pCO$_2$ across Earth changed significantly, perhaps on orbital time scales (Smith et al., 2010;

165 Slotnick et al., 2012, 2015a; Kirtland-Turner et al., 2014).

166   There is also the root problem as to where EECO starts and ends. At a basic level, the

167 interval characterized by the lowest Cenozoic benthic foraminiferal $\delta^{18}O$ values begins at a

168 time that closely corresponds with a long-term minimum in $\delta^{13}C$ values (**Figure 1**). This is

169 important for stratigraphic reasons because the two stable isotope curves were generated

170 using the same benthic foraminiferal samples, but $\delta^{13}C$ records at different locations should

171 necessarily correlate in the time domain (unlike $\delta^{18}O$ and temperature). The rationale for such

172 carbon isotope stratigraphy lies in the rapid cycling of carbon across Earth's surface

173 (Shackleton, 1986; Dickens, 2000).

174   The Eocene minimum in $\delta^{13}C$ corresponds to the K/X event (**Figure 1**), which happened

175 in polarity chron C24n.1n and approximately 3 Myr after the PETM (Agnini et al., 2009;

176 Leon-Rodriguez and Dickens, 2010; Slotnick et al., 2012; Dallanave et al., 2015; Lauretano

177 et al., 2015; Westerhold et al., 2015). However, in several detailed studies spanning the early

178 Eocene, changes in long-term trends appear to have occurred about 400 kyr before the K/X

179 event, and at an event called "J" (after Cramer et al., 2003), which happened near the

180 boundary of polarity chrons C24n.2r and C24n.3n (Slotnick et al., 2015a; Lauretano et al.,

181 2015). Notably, the long-term late Paleocene-early Eocene decrease in detailed benthic

182 foraminiferal $\delta^{18}O$ records at Site 1262 on Walvis Ridge ceases at the J event (Lauretano et

183 al., 2015).

184   The end of the EECO has received limited attention from a stratigraphic perspective.

185 Indeed, the termination of the EECO may not be a recognizable global "event", because it

186 might relate to ocean circulation and gateways and expressed mostly in Southern Ocean and

187 deep ocean records (Pearson et al., 2007; Bijl et al. 2013). In Paleogene continental slope

188 sections now uplifted and exposed in the Clarence River Valley, New Zealand, a major

189 lithologic change from limestone to marl coincides with the J event (Slotnick et al., 2012,

190 2015a; Dallanave et al., 2015). The marl-rich unit, referred to as "Lower Marl", has been

191 interpreted to reflect enhanced terrigenous supply to a continental margin because of greater

192 temperature and enhanced seasonal precipitation. It has been suggested further that Lower

193 Marl expresses the EECO (Slotnick et al., 2012; Dallanave et al., 2015). The top of Lower

194 Marl, and a return to limestone deposition, lies within the upper part of polarity chron C22n

195 (Dallanave et al., 2015). This is interesting because it approximates the time when general

196   long-term Cenozoic cooling initiates at several locations that have records of polarity chrons

197   and proxies for temperature (Bijl et al., 2009; Hollis et al., 2012; Pross et al., 2012). It is also

198   useful from a stratigraphic perspective because the end of the EECO thus lies close to a well

199   documented and widespread calcareous nannofossil biohorizon, the base of *Discoaster*

200   *sublodoensis*. This marks the base of CP10, NP12 or CNE4, depending on the chosen

201   calcareous nannofossil zonal scheme (Okada and Bukry, 1080; Martini, 1971; Agnini et al.,

202   2014).

203   Without an accepted definition in the literature, we tentatively present the EECO as the

204   duration of time between the J event and the base of *D. sublodoensis*. This interval thus

205   begins at about 53 Ma and ends at about 49 Ma on the 2012 Time Scale (GTS; Vandenberghe

206   et al., 2012). However, while the EECO was characterized by generally warm conditions,

207   numerous fluctuations in average temperature likely occurred during the 4 Myr interval.

208

209   **3 Sites and stratigraphy**

210

211   **3.1 Possagno, Venetian Prealps, Tethys**

212

213   An Upper Cretaceous through Miocene succession crops out at the bottom of the Monte

214   Grappa Massif in the Possagno area, about 60 km northwest of Venice. The lower to middle

215   Eocene, of primary focus to this study, is represented by the Scaglia beds. These

216   sedimentary rocks represent pelagic and hemipelagic sediment that accumulated at middle to

217   lower bathyal depths (Cita, 1975; Thomas, 1998) in the western part of the Belluno Basin, a

218   Mesozoic–Cenozoic paleogeographic unit of the Southern Alps (Bosellini, 1989). The basin

219   very likely was an embayment connected to the western Tethys, with a paleolatitude of ca.

220   42° during the early Eocene (**Figure 2**).

221   A quarry at 45°51.0' N and 11°51.6' E exposed in 2002-2003 a 66 m thick section of

222   the Scaglia beds (Figure 3), although it is at present largely covered and inaccessible. This

223   section was examined for its stratigraphy (Agnini et al., 2006; Luciani and Giusberti, 2014),

224   and shown to extend from just below the PETM to within lower Chron C20r in the lower

225   middle Eocene. Like other lower Paleogene sections of the Venetian Pre-alps (Giusberti et

226   al., 2007; Agnini et al., submitted), a Clay Marl Unit (CMU) with a prominent negative CIE

227   marks the PETM.

228   The Possagno section appears to be continuous, but with an important decrease in

229   sedimentation rate (to below 1.4 m/Myr) between 14.66 m and 15.51 m (Agnini et al., 2006).

230   This interval lies within Chron C23r and near the start of the EECO, and predates the onset

231   of a major increase in ~~discoaster~~ *Discoaster* abundance (Agnini et al., 2006).

232

233   **3.2 Site 577, Shatsky Rise, Western Pacific**

234

235   Deep Sea Drilling Project (DSDP) Leg 86 drilled Site 577 at 32°26.5' N, 157°43.4' E, and

236   2680 m water depth, on Shatsky Rise, a large igneous plateau in the NW Pacific with a

237   relatively thin veneer of sediment (Shipboard Scientific Party 1985). During the early

238   Eocene, this site was located closer to 15° N (**Figure 2**), and probably at a slightly shallower

239   water depth (Ito and Clift, 1998).

240   Two primary holes were drilled at Site 577. Both Hole 577* and Hole 577A recovered

241   portions of a nominally 65 m thick section of Upper Cretaceous through lower Eocene

242   nannofossil ooze. Similar to the Possagno section, the lower Paleogene interval has

243   biomagnetostratigraphic information (Bleil, 1985; Monechi et al., 1985; Backman, 1986; Lu

244   and Keller, 1995; Dickens and Backman, 2013). Stable isotope records of bulk carbonate

245   have been generated for sediment from several cores at low sample resolution (Shackleton,

246 1986), and for much of Cores 577*-9H and 577*-10H at fairly high sample resolution

247 (Cramer et al. 2003).

248   The composition and relative abundances of planktic foraminifera were nicely

249 documented at Site 577 (Lu, 1995; Lu and Keller, 1995), and show a marked turnover

250 between *Morozovella* ~~morozovellids~~ and *Acarinina* ~~acaraninids~~ during the early Eocene.

251 These data, however, have remained on an out-dated view for the stratigraphy at this

252 location, where cores were not originally aligned to account for gaps and overlaps (Dickens

253 and Backman, 2013). As will become obvious later, the main phase of the EECO spans

254 Cores 577*-8H and 577A-8H, where detailed stable isotope records have not been generated

255 previously.

256

257 **3.3 Site 1051, Blake Nose, Western Atlantic**

258

259 The Blake Nose is a gentle ramp extending from 1000 m to 2700 m water depth east of

260 Florida (Norris et al, 1998). The feature is known for a relatively thick sequence of middle

261 Cretaceous through middle Eocene sediment with minimal overburden. Ocean Drilling

262 Program (ODP) Leg 171B drilled and cored this sequence at several locations, including Site

263 1051 at 30°03.2' N, 76°21.5' W, and 1994 m water depth (Shipboard Scientific Party 1998).

264 The site was located slightly to the south during the early Eocene (**Figure 2**). Benthic

265 foraminiferal assemblages indicate a lower bathyal depth (1000-2000 m) during the late

266 Paleocene and middle Eocene (Norris et al., 1998), although Bohaty et al. (2009) estimated a

267 paleodepth of about 2200 m for sedimentation ca. 50 Ma.

268   Sediments from 452.24 to 353.10 meters below sea floor (mbsf) at Site 1051 consist of

269 lower to middle Eocene carbonate ooze and chalk (Shipboard Scientific Party, 1998). The

270 site comprises two holes (1051A and 1051B), with core gaps and core overlaps existing at

271 both (Shipboard Scientific Party, 1998). However, the impact of these depth offsets upon

272 age is less than at Site 577, because of higher overall sedimentation rates.

273 The Eocene section at Site 1051 has good sediment recovery, except an interval between

274 382 mbsf and 390 mbsf, which contains significant chert. Stratigraphic markers across the

275 Eocene interval include polarity chrons (Ogg and Bardot, 2001), calcareous nannofossil

276 biohorizons (Mita, 2001), and planktic foraminiferal biohorizons (Norris et al., 1998; Luciani

277 and Giusberti, 2014). As first noted by Cramer et al. (2003), though, there is a basic

278 stratigraphic problem with the labelling of the polarity chrons. The intervals of normal

279 polarity between approximately 388 and 395 mbsf, and between approximately 412 and 420

280 mbsf were tentatively assigned to C22n and C23n, respectively (Ogg and Bardot, 2001). This

281 age assignment was assumed to be correct by Luciani and Giusberti (2014), who therefore

282 considered the last occurrence of *Morozovella subbotinae* as happening near the top of C23n,

283 an assumption that was also made for the revision of Eocene foraminiferal biozones (Wade et

284 al., 2011).

285 These age assignments, however, cannot be correct, because calcareous nannofossil

286 biohorizons that lie below or within C22n (top of *T. orthostylus*, top of *Toweius*, base of *D.*

287 *sublodoensis*) occur above 388 mbsf (Mita, 2001). Instead, there must be a significant hiatus

288 or condensed interval at the chert horizon, and the above noted intervals of normal polarity

289 are C23n and C24n.1n.

290

291 **4 Methods**

292

293 **4.1 Samples for isotopes and foraminifera**

294

295 The three sites provide a good stratigraphic background and key existing data for

296   understanding the temporal link between the EECO, carbon isotope perturbations and

297   planktic foraminiferal evolution. Our analytical aim was to obtain comparable data sets

298   across the sites. More specifically, a need existed to generate stable isotope and planktic

299   foraminiferal assemblage records at the Possagno section, to generate stable isotope records

300   at DSDP Site 577, and to generate planktic foraminiferal assemblage records at ODP Site

301   1051.

302   In total, 298 samples were collected from the originally exposed Possagno section in

303   2002-2003 for isotope analyses. The sampling interval was 2 to 5 cm for the basal 0.7 m, and

304   at variable spacing from 20 to 50 cm for the interval between 0.7 m and 66 m. Bulk sediment

305   samples previously were examined for their calcareous nannofossil assemblages (Agnini et

306   al., 2006). One hundred and ten of these samples were selected for the foraminiferal study.

307   Aliquots of the 110 samples were weighed, and then washed to obtain foraminifera using

308   two standard procedures, depending on lithology. For the indurated marly limestones and

309   limestones, the cold-acetolyse technique was used (Lirer, 2000; Luciani and Giusberti, 2014).

310   This method disaggregates strongly lithified samples, in which foraminifera otherwise can be

311   analyzed only with thin sections (Fornaciari et al., 2007; Luciani et al., 2007). For the marls,

312   samples were disaggregated using 30 % hydrogen peroxide and subsequently washed and

313   sieved at 63 µm. In most cases, gentle ultrasonic treatment (e.g., low-frequency at 40 kHz for

314   30–60 seconds) improved the cleaning of the tests.

315   Relative abundance data of planktic foraminiferal samples were generated for 65 samples

316   at Site 577 (Lu, 1995; Lu and Keller, 1995). We collected new samples for stable isotope

317   measurements that span their previous effort.

318   Fifty samples of Eocene sediment were obtained from Hole 1051A between 452 to 353

319   mbsf. Sample spacing varied from 2.0 m to 0.5 m. As the samples are ooze and chalk, they

320   were prepared using disaggregation using distilled water and washing over 38 µm and 63 µm

13

321 sieves. Washed residues were dried at <50°C.

322

323 **4.2 Stable Isotopes**

324

325 Carbon and oxygen stable isotope data of bulk sediment samples from the Possagno section

326 and Site 577 were analysed using a Finnigan MAT 252 mass spectrometer equipped with a

327 Kiel device at Stockholm University. Precision is within ±0.06 ‰ for carbon isotopes and

328 within ±0.07 ‰ for oxygen isotopes. Stable isotope values were calibrated to the Vienna Pee

329 Dee Belemnite standard (VPDB) and converted to conventional delta notation ($\delta^{13}C$ and

330 $\delta^{18}O$).

331

332 **4.3 Foraminifera analyses**

333

334 The mass percent of the >63 μm size fraction relative to the mass of the bulk sample,

335 typically 100 g/sample was calculated for the 110 Possagno samples. This is referred to as the

336 weight percent coarse fraction, following many previous works. Due to the consistent

337 occurrence of radiolarians at Site 1051, the coarse fraction cannot give information on

338 foraminiferal productivity.

339   Relative abundances for both Possagno and Site 1051 have been determined from about

340 300 complete specimens extracted from each of the 110 samples investigated in the >63 μm

341 size fraction from random splits.

342   The degree of dissolution, expressed as the fragmentation index (*F* index) was evaluated

343 according to Petrizzo et al. (2008) on ca. 300 elements, by counting planktic foraminiferal

344 fragments or partially dissolved tests versus complete tests. These data are expressed in

345 percentages. Fragmented foraminifera include specimens showing missing chambers and

346 substantial breakage. The taxonomic criteria for identifying planktic foraminifera follows the

347 work by Pearson et al. (2006).

348

349 **5 Results**

350

351 **5.1 Carbon isotopes**

352

353 *Possagno*

354 Carbon isotopes of bulk carbonate at Possagno vary between +1.8 and -0.3 ‰ (**Figure 4,**

355 **Table S1**). Overall, $\delta^{13}C$ decreases from 1.8 ‰ at the base of the section to about 0.6 ‰ at 14

356 m. Generally, values then increase to 1.5 ‰ at 24 m, and remain between 1.5 ‰ and 0.8 ‰

357 for the remainder of the studied interval.

358   Superimposed on these trends are a series of negative CIEs. The most prominent of these

359 (~1.5 ‰) occurs at the 0 m level, and marks the PETM (Agnini et al., 2009). However, other

360 negative CIEs lie above this marker and within the lowermost 21.4 m, albeit some are only

361 defined by one data point (**Figure 4, Table S1**). The lower two at ~8 m and ~12.5 m probably

362 represent the H-1/ETM-2 and J event, respectively, as they lie at the appropriate stratigraphic

363 horizons in relation to polarity chrons. The K/X event may lie at 14.8 m, although this height

364 marks the start of the condensed interval.

365   The complex interval between 15.5 m and 24 m broadly corresponds to all of Chron

366 C23n and the bottom half of Chron C22r. A series of CIEs occur in that interval on the order

367 of 1.4 ‰, superimposed on a background trend of increasing $\delta^{13}C$ values (about 0.7 ‰). We

368 tentatively label these CIEs with even numbers for internal stratigraphic purposes (**Figure 4**),

369 as will become obvious below; their magnitudes range between 0.9 and 0.3 ‰ (**Table S1**).

370 However, the sample spacing through this interval varies from 20 to 50 cm. The precise

371 magnitudes and positions certainly could change with higher sample resolution, given the

372 estimated compacted sedimentation rate of ~0.5 cm/kyr for this part of the section (Agnini et

373 al., 2006).

374   Above Chron C22r, the Possagno δ^{13}C record contains additional minor CIEs (**Figure 4**).

375 The most prominent of these CIEs, at least relative to baseline values (~1.2 ‰), occurs within

376 Chron C21n. More important to understanding the EECO, a ~0.6 ‰ CIE nearly coincides

377 with the base of *D. sublodoensis* within the lower part of Chron C22n.

378

379 *DSDP Site 577*

380 The δ^{13}C record of bulk carbonate at DSDP Site 577 from just below the PETM through

381 Chron C22n ranges between 2.3 and 0.6 ‰ (**Figure 5; Table S2**). Overall, δ^{13}C decreases

382 from 1.4 ‰ at 84.5 mcd to about 0.6 ‰ at ~76 mcd. Values then generally increase to 2.1 ‰

383 at ~68 mcd, and remain between 2.3 ‰ and 1.6 ‰ for the rest of the studied interval. Thus,

384 the ranges and general trends in δ^{13}C for the two sections are similar, but skewed at DSDP

385 Site 577 relative to Possagno by about +0.6 ‰.

386   Like at Possagno, the early Eocene δ^{13}C record at DSDP Site 577 exhibits a series of

387 CIEs (**Figure 5**). The portion of this record from the PETM through the K/X event has been

388 documented and discussed elsewhere (Cramer et al., 2003; Dickens and Backman, 2013). The

389 new portion of this record, from above the K/X event through Chron C22n, spans the

390 remainder of the EECO. Within this interval, where background δ^{13}C values rise by ~1.5 ‰,

391 there again occur a series of minor CIEs with magnitudes between 0.3 and 0.5 ‰ (**Table S2**).

392 Here, however, multiple data points define most of the CIEs. We again give these an internal

393 numerical labelling scheme. A ~0.4 ‰ CIE also nearly coincides with the base of *D.*

394 *sublodoensis* within the lower part of C22n.

395

5.2 Oxygen isotopes

Possagno

Oxygen isotopes of bulk carbonate at Possagno range between -3.3 and 0.8 ‰ with a mean value of -1.7 ‰ (**Figure 4, Table S1**). In general, considerable scatter exists across the data set with respect to depth, as adjacent samples often display a difference in $\delta^{18}O$ that exceeds 0.5 ‰. Nonetheless, some of the more prominent lows in $\delta^{18}O$ show a clear correspondence with negative $\delta^{13}C$ values (CIEs) and vice versa. This correspondence occurs across the PETM and other known hyperthermals, as well as within and after the EECO. Indeed, the main phase of the EECO corresponds with a broad has the lowest $\delta^{18}O$ values.

DSDP Site 577

The $\delta^{18}O$ record at Site 577 noticeably deviates from that at Possagno (**Figure 5, Table S2**). This is because values range between -1.1 ‰ and 0.2 with an average value of -0.4 ‰. Thus, relative to Possagno, the record at Site 577 has less scatter, and an overall shift of about -1.3 ‰. There is ~~exists~~ again a modest correlation between decreases in $\delta^{18}O$ and negative $\delta^{13}C$ values, as well as a general low in $\delta^{18}O$ across the main phase of the EECO.

5.3 Coarse fraction

The coarse fraction of samples from Possagno shows two distinct trends (**Figure 6, Table S3**). Before the EECO, values are 10.4 % ± 2.67 %. However, from the base of the EECO and up through the section, values decrease to 5.3 ± 1.3 %.

5.4 Foraminiferal preservation and fragmentation

421

422   Planktic foraminifera are consistently present and diverse throughout the studied intervals at

423   Possagno and at ODP Site 1051. Preservation of the tests at Possagno varies from moderate

424   to fairly good (Luciani and Giusberti, 2014). However, planktic foraminiferal tests at

425   Possagno are recrystallized and essentially totally filled with calcite. Planktic foraminifera

426   from samples at Site 1051 are readily recognizable throughout the studied interval. Planktic

427   foraminifera from Site 577, at least as illustrated by published plates (Lu and Keller, 1995),

428   show a very good state of preservation (albeit possibly recrystallized).

429   The *F* index record at Possagno (**Figure 6, Table S3**) displays large amplitude variations

430   throughout the investigated interval. The highest values, up to 70 %, were observed between

431   16 and 22 m. In general, highs in *F* index values correspond to lows in the $\delta^{13}C$ record.

432   The *F* index record at Site 1051 (**Figure 8, Table S4**) shows less variability compared to

433   that at Possagno, although some of this may reflect the difference in the number of samples

434   examined at the two locations. A maximum value of 60 % is found in Zone E5, just below an

435   interval of uncertain magnetostratigraphy (Norris et al., 1998), but corresponding to the J

436   event (Cramer et al., 2003). Relatively high *F* index values, around 50 %, also occur in

437   several samples below this horizon. The interval across the EECO generally displays low *F*

438   index values (<20 %).

439

440   **5.5 Planktic foraminiferal quantitative analysis**

441

442   *Possagno*

443   Planktic foraminiferal assemblages at Possagno show significant changes across the early to

444   early middle Eocene (**Figure 6, Table S3**). Throughout the entire section, the mean relative

445   abundance of *Acarinina* is about 46 % of the total assemblage. However, members of this

446 genus ~~show~~ exhibit peak abundances of 60-80 % of the total assemblage ~~occur~~ across several

447 intervals, often corresponding to CIEs. Particularly prominent is the broad abundance peak of

448 *Acarinina* coincident with the main phase of the EECO.

449   The increases in *Acarinina* ~~acarininid~~ relative abundance typically are counterbalanced

450 by transient decreases of subbotinids (that include both *Subbotina* and *Parasubbotina* genera;

451 **Figure 6**). This group also shows a general increase throughout the section. Below the EECO

452 the relative abundances of subbotinids average ~24 %. Above the EECO, this average rises to

453 ~36 %.

454   The trends of *Acarinina* ~~acarininids~~ and subbotinids contrast with that of *Morozovella*

455 ~~morozovellids~~ (**Figure 6**), which exhibit a major and permanent decline within Zone E5. This

456 group collapses from mean abundances ~24 % in the 0-15 m interval to <6 % above 15 m.

457 Qualitative examination of species shows that, in the lower part of Zone E5, where relatively

458 high *Morozovella* ~~morozovellids~~ abundances are recorded, there is no dominance of any

459 species. *M. marginodentata*, *M. subbotinae* and *M. lensiformis* are each relatively common,

460 and *M. aequa, M. aragonensis, M. formosa* and *M. crater* are each less common. By contrast,

461 in the upper part of Zone E5, where low abundances of *Morozovella* ~~morozovellids~~ occur, *M.*

462 *aragonensis, M. formosa, M. crater* and *M. caucasica* are the most common species. The

463 general decrease of *Morozovella* ~~morozovellids~~ abundances appears unrelated to the

464 disappearance of a single, dominant species.

465   At Possagno, *Morozovella* ~~morozovellids~~ never recover to their pre-EECO abundances.

466 This is true even if one includes the morphologically and ecologically comparable genus

467 *Morozovelloides* (Pearson et al., 2006), which first appears in samples above 36 m.

468   Other planktic foraminiferal genera are always less than 15 % of the total assemblages

469 throughout the studied interval at Possagno (**Figure S1, Table S3**).

470

471   *ODP Site 577*

472   Samples from Site 577 were disaggregated in water and washed through a >63 sieve (Lu,

473   1995; Lu and Keller, 1995). They determined relative abundances of planktic foraminifera

474   from random splits of about 300 specimens (Lu, 1995; Lu and Keller, 1995). The resulting

475   data are shown in **Figure 7**, placed onto the composite depth scale by Dickens and Backman

476   (2013). Major changes in planktic foraminiferal assemblages are comparable to those

477   recorded at Possagno. Such changes include indeed a distinct decrease of *Morozovella*

478   ~~morozovellids~~ within Zone E5. The decrease at Site 577 is from mean values of 26.6 % to 6.7

479   % (**Table S4**). This marked drop occurs at ca. 78 mcd close to the J event and at the start of

480   the EECO. Like at Possagno, *Morozovella* ~~morozovellids~~ never recover to their pre-EECO

481   abundances.

482   The *Morozovella* ~~morozovellids~~ decrease is counter balanced by the trend of *Acarinina*

483   ~~acarininid~~ abundances that increase from mean values of 30.4 % to 64.8 % in correspondence

484   to the level of the *Morozovella* ~~morozovellid~~ collapse. Subbotinids fluctuate in abundance

485   throughout the interval investigated from 1 % to 18 %, with a mean value of ca. 8 %.

486

487   *ODP Site 1051*

488   Planktic foraminifera show distinct changes in abundance at Site 1051 (**Figure 8, Table S5**).

489   The changes of the main taxa are similar to the variations observed at Possagno. The genus

490   *Acarinina* displays an increase in mean relative abundance from 35 % (base to ca. 450 mbsf)

491   to around 50 % (ca. 430 mbsf), with maximum values of about 60 %. The relatively low

492   resolution used here does not permit comparison to the early Eocene CIEs at Site 1051

493   (Cramer et al., 2003), or how the relative abundance of planktic foraminiferal genera varies

494   with respect to CIEs.

495   The abundance of subbotinids shows ~~little~~ small variations around mean values of 20 %

496   at Site 1051. Like at Possagno, samples from Site 1051 also record a slight increase in

497   abundance toward the end of the EECO and above.

498   The major change in planktic foraminiferal assemblages at Site 1051 includes a distinct

499   decrease of *Morozovella*, from mean values around 40 % to 10 % in the middle part of Zone

500   E5 (**Figure 7**). Similar to Possagno, the lower part of Zone E5 with the higher percentages of

501   *Morozovella* ~~morozovellids~~ does not record the dominance of selected species, but at Site

502   1051 *M. aragonensis* and *M. formosa* besides *M. subbotinae* are relatively common whereas

503   *M. marginodentata* is less frequent. Within the interval of low ~~morozovellids~~ *Morozovella*

504   abundances, *M. aragonensis* and *M. formosa* are the most common taxa. The general decline

505   of *Morozovella* ~~morozovellids~~ does not appear therefore related, both at Possagno and at Site

506   1051, to the extinction or local disappearance of a dominant species.

507

508   **6 Discussion**

509

510   **6.1 Dissolution, recrystallization, and bulk carbonate stable isotopes**

511

512   The bulk carbonate stable isotope records within the lower Paleogene sections at Possagno

513   and at Site 577 need some reflection ~~thought~~, considering how such records are produced and

514   modified in much younger strata dominated by pelagic carbonate. In open ocean

515   environments, carbonate preserved on the seafloor principally consists of calcareous tests of

516   nannoplankton (coccolithophores) and planktic foraminifera (Bramlette and Riedel, 1954;

517   Berger, 1967; Vincent and Berger, 1981). However, the total amount of carbonate and its

518   microfossil composition can vary considerably across locations because of differences in

519   deep-water chemistry and in test properties (e.g., ratio of surface area to volume;

520   mineralogical composition). For regions at low to mid latitudes, a reasonable representation

521   of carbonate components produced in the surface water accumulates on the seafloor at

522   modest (<2000 m) water depth. By contrast, microfossil assemblages become heavily

523   modified in deeper water, because of increasingly significant carbonate dissolution (Berger,

524   1967). Such dissolution preferentially affects certain tests, such as thin-walled, highly porous

525   planktic foraminifera (Berger, 1970; Bé et al., 1975; Thunell and Honjo, 1981).

526   The stable isotope composition of modern bulk carbonate ooze reflects the mixture of its

527   carbonate components, which mostly record water temperature and the composition of

528   dissolved inorganic carbon (DIC) within the mixed layer (<100 m water depth). The stable

529   isotope records are imperfect, though, because of varying proportions of carbonate

530   constituents, and "vital effects", which impact stable isotope fractionation for each

531   component (Anderson and Cole, 1975; Reghellin et al., 2015). Nonetheless, the stable isotope

532   composition of bulk carbonate ooze on the seafloor can be related to overlying temperature

533   and chemistry of surface water (Anderson and Cole, 1975; Reghellin et al., 2015).

534   Major modification of carbonate ooze occurs during sediment burial. This is because,

535   with compaction and increasing pressure, carbonate tests begin to dissolve and recrystallize

536   (Schlanger and Douglas, 1974; Borre and Fabricus, 1998). Typically within several hundred

537   meters of the seafloor, carbonate ooze becomes chalk and, with further burial, limestone

538   (Schlanger and Douglas, 1974; Kroencke et al., 1991; Borre and Fabricus, 1998). Carbonate

539   recrystallization appears to be a local and nearly closed system process, such that mass

540   transfer occurs over short distances (i.e., less than a few meters) (above references and Matter

541   et al., 1975; Arthur et al., 1984; Frank et al., 1999).

542   In pelagic sequences with appreciable carbonate content and low organic carbon content,

543   bulk carbonate $\delta^{13}C$ records typically give information of paleoceanographic significance

544   (Scholle and Arthur, 1980; Frank et al., 1999). Even when transformed to indurated

545   limestone, the $\delta^{13}C$ value for a given sample should be similar to that originally deposited on

546 the seafloor. This is because, for such sediments, almost all carbon within small ~~sedimentary~~

547 volumes exists as carbonate. Bulk carbonate $\delta^{18}O$ records are a different matter, especially in

548 indurated marly limestones and limestones (Marshall, 1992; Schrag et al., 1995; Frank et al.,

549 1999). This is because pore water dominates the total amount of oxygen within an initial

550 parcel of sediment, and oxygen isotope fractionation depends strongly on temperature. Thus,

551 during dissolution and recrystallization of carbonate, significant exchange of oxygen isotopes

552 occurs. At first, carbonate begins to preferentially acquire ^{18}O, because shallowly buried

553 sediment generally has lower temperatures than surface water. However, with increasing

554 burial depth along a geothermal gradient, carbonate begins to preferentially acquire ^{16}O

555 (Schrag et al., 1995; Frank et al., 1999).

556

557 **6.2 Carbon isotope stratigraphy through the EECO**

558

559 Stratigraphic issues complicate direct comparison of various records from Possagno and Site

560 577. The two sections have somewhat similar multi-million year sedimentation rates across

561 the early Eocene. However, the section at Possagno contains the condensed interval, where

562 much of C23r spans a very short distance (Agnini et al., 2006), and the section at Site 577 has

563 a series of core gaps and core overlaps (Dickens and Backman, 2013).

564　　　An immediate issue to amend is the alignment of Cores 8H and 9H in Hole 577* and

565 Core 8H in Hole 577A (**Figure 5**). On the basis of GRAPE density records for these cores,

566 Dickens and Backman (2013) initially suggested a 2.6 m core gap between Cores 8H* and

567 9H*. However, a 3.5 m core gap also conforms to all available stratigraphic information. The

568 newly generated $\delta^{13}C$ (and $\delta^{18}O$) records across these three cores show the latter to be correct.

569　　　Once sedimentation rate differences at Possagno are recognized and coring problems at

570 Site 577 are rectified, early Eocene $\delta^{13}C$ records at both locations display similar trends and

571   deviations in relation to polarity chrons and key microfossil events (**Figures 4, 5**). Moreover,

572   the δ^{13}C variations seemingly can be correlated in time to those found in bulk carbonate δ^{13}C

573   records at other locations, including Site 1051 (**Figure 8**) and Site 1258 (**Figure 9**). As noted

574   previously, such correlation occurs because the bulk carbonate δ^{13}C signals reflect past global

575   changes in the composition of surface water DIC, even after carbonate recrystallization.

576   For the latest Paleocene and earliest Eocene, nominally the time spanning from the base

577   of C24r through the middle of C24n, detailed stable carbon isotope records have been

578   generated at more than a dozen locations across the globe (Cramer et al., 2003; Agnini et al.,

579   2009; Galeotti et al., 2010; Zachos et al., 2010; Slotnick et al., 2012; Littler et al., 2014;

580   Agnini et al., in review). These records can be described consistently as a long-term drop in

581   δ^{13}C superimposed with a specific sequence of prominent CIEs that include those

582   corresponding to the PETM, H-1, and J events. In continuous sections with good

583   magnetostratigraphy and biostratigraphy, there is no ambiguity in the assignment of CIEs

584   (Zachos et al., 2010; Littler et al., 2014; Slotnick et al., 2012, 2105a; Lauretano et al., 2015).

585   This "δ^{13}C template" can be found at the Possagno section and at Site 577 (**Figure 9**); it is

586   found at Site 1051 for the depth interval where carbon isotopes have been determined

587   (**Figure 8**).

588   After the J event and across the EECO, very few detailed δ^{13}C records have been

589   published (Slotnick et al., 2012, 2015a; Kirtland-Turner et al., 2014). Moreover, the available

590   records are not entirely consistent. For example, the K/X event in Clarence River valley

591   sections manifests as a prominent CIE within a series of smaller δ^{13}C excursions (Slotnick et

592   al., 2012, 2015a), whereas the event has limited expression in the δ^{13}C record at Site 1258

593   (Kirtland-Turner et al., 2014; **Figure 9**).

594   The new records from Possagno and Site 577 emphasize an important finding regarding

595   bulk carbonate δ^{13}C records across the EECO. Between the middle of C24n and the upper

596 part of C23r, there appears to be a sequence of low amplitude, low frequency CIEs. (Note

597 that this portion of the record is missing at Possagno because of the condensed interval;

598 **Figure 9**). However, near the C23r/C23n boundary, a long-term rise in $\delta^{13}C$ begins, but with

599 a series of relatively high amplitude, high frequency CIEs (Kirtland-Turner et al., 2014;

600 Slotnick et al., 2014). The number, relative magnitude and precise timing of CIEs within this

601 interval remain uncertain. For example, the CIE labelled "4" appears to occur near the top of

602 C23r at Site 577 but near the bottom of C23n.2n at Site 1258 and at Possagno. Additional

603 $\delta^{13}C$ records across this interval are needed to resolve the correct sequence of CIEs and to

604 derive an internally consistent labelling scheme for these perturbations. It is also not clear

605 which of these CIEs during the main phase of the EECO specifically relate to significant

606 increases in temperature, as clear for the "hyperthermals" in the earliest Eocene. Nonetheless,

607 numerous CIEs, as well as an apparent change in the mode of these events, characterize the

608 EECO (Kirtland-Turner et al., 2014; Slotnick et al., 2014).

609 The causes of $\delta^{13}C$ changes during the early Paleogene lie at the crux of considerable

610 research and debate (Dickens et al., 1995, 1997; Zeebe et al., 2009; Dickens, 2011; Lunt et

611 al., 2011; Sexton et al., 2011; De Conto et al., 2012; Lee et al., 2013; Kirtland Turner et al.,

612 2014). Much of the discussion has revolved around three questions: (1) what are the sources

613 of ^{13}C-depleted carbon that led to prominent CIEs, especially during the PETM? (2) does the

614 relative importance of different carbon sources vary throughout this time interval? and, (3)

615 are the geologically brief CIEs related to the longer secular changes in $\delta^{13}C$? One might

616 suggest, through several papers, a convergence of thought as to how carbon cycled across

617 Earth's surface during the early Paleogene, at least between the late Paleocene and the K/X

618 event (Cramer et al., 2003; Lourens et al., 2005; Galeotti et al., 2010; Hyland et al., 2013;

619 Zachos et al., 2010; Lunt et al. 2011; Littler et al., 2014; Lauretano et al., 2015; Westerhold et

620 al., 2015). Changes in ~~tectoniesm~~, tectonism, volcanism, and weathering drove long-term

621 changes atmospheric $p\text{CO}_2$ (Vogt, 1979; Raymo and Ruddiman, 1992; Sinton and Duncan,

622 1998; Demicco, 2004; Zachos et al., 2008), which was generally high throughout the early

623 Paleogene, but increased toward the EECO (Pearson and Palmer, 2000; Fletcher et al., 2008;

624 Lowenstein and Demicco, 2006; Smith et al., 2010; Hyland and Sheldon, 2013). However, as

625 evident from the large range in $\delta^{13}\text{C}$ across early Paleogene stable isotope records, major

626 changes in the storage and release of organic carbon must have additionally contributed to

627 variability in atmospheric $p\text{CO}_2$ and ocean DIC concentrations (Shackleton, 1986; Kurtz et

628 al., 2003; Komar et al., 2013). When long-term increases in $p\text{CO}_2$, perhaps in conjunction

629 with orbital forcing, pushed temperatures across some threshold, such as the limit of sea-ice

630 formation (Lunt et al., 2011), rapid inputs of ^{13}C-depleted organic carbon from the shallow

631 geosphere served as a positive feedback to abrupt warming (Dickens et al., 1995; Bowen et

632 al., 2006; DeConto et al., 2012).

633   Our new $\delta^{13}\text{C}$ records do not directly address the above questions and narrative

634 concerning early Paleogene carbon cycling. However, they do highlight two general and

635 related problems when such discussion includes the EECO. First, surface temperatures appear

636 to stay high across an extended time interval when the $\delta^{13}\text{C}$ of benthic foraminifer (**Figure 1**)

637 and bulk carbonate (**Figure 9**) increase. Second, numerous brief CIEs mark this global long-

638 term rise in $\delta^{13}\text{C}$. Whether the aforementioned views need modification or reconsideration

639 (Kirtland Turner et al., 2014) is an outstanding issue, one that depends on how long-term and

640 short-term $\delta^{13}\text{C}$ changes relate across the entire early Paleogene.

641   The overall offset between bulk carbonate $\delta^{13}\text{C}$ values at Possagno and Site 577 may hint

642 at an important constraint to any model of early Paleogene carbon cycling. Throughout the

643 early Eocene, $\delta^{13}\text{C}$ values at Site 577 exceed those at Possagno by nominally 0.8 ‰ (**Figure**

644 **9**). This probably does reflect recrystallization or lithification, because similar offsets appear

645 across numerous records independent of post-depositional history but dependent on location

646   (Schmitz et al., 1996; Cramer et al., 2003; Slotnick et al., 2012, 2015a; Agnini et al.,

647   submitted). In general, absolute values of bulk carbonate $\delta^{13}C$ records increase from the

648   North Atlantic and western Tethys (low), through the South Atlantic and eastern

649   Tethys/Indian, to the Pacific (high), although suggestively with a latitudinal component to

650   this signature.

651

652   **6.3 Stable oxygen isotope stratigraphy across the EECO**

653

654   Bulk carbonate $\delta^{18}O$ values for Holocene sediment across the Eastern Equatorial Pacific

655   relate to average temperatures in the mixed layer (Shackleton and Hall, 1995; Reghellin et al.,

656   2015). Indeed, values are close to those predicted from water chemistry ($\delta^{18}O_w$) and

657   equilibrium calculations for calcite precipitation (e.g., Bemis et al., 1998) if vital effects in

658   the dominant nannoplankton increase $\delta^{18}O$ by nominally 1‰ (Reghellin et al., 2015).

659   Site 577 was located at about 15°N latitude in the eastern Pacific during the early

660   Paleogene. Given that sediment of this age remains "nannofossil ooze" (Shipboard Scientific

661   Party, 1985), one might predict past mixed layer temperatures from the $\delta^{18}O$ values with

662   three assumptions: early Paleogene $\delta^{18}O_w$ was 1.2 ‰ less than that at present-day to account

663   for an ice-free world; local $\delta^{18}O_w$ was equal to average seawater, similar to modern chemistry

664   at this off-Equator location (LeGrande and Schmidt, 2006); and, Paleogene nannoplankton

665   also fractionated $\delta^{18}O$ by 1.0 ‰. With commonly used equations that relate the $\delta^{18}O$ of

666   calcite to temperature (Bemis et al., 1998), these numbers render temperatures of between

667   16°C and 21°C for the data at Site 577. Such temperatures seem too cold by at least 10°C,

668   given other proxy data and modelling studies (e.g., Pearson et al., 2007; Huber and Caballero,

669   2011; Hollis et al., 2012; Pross et al., 2012; Inglis et al., 2015). At low latitudes, bottom

670   waters are always much colder than surface waters. Even during the EECO, deep waters

671 probably did not exceed 12°C (Zachos et al., 2008). The calculated ~~tepid~~ temperatures likely

672 indicate partial recrystallization of bulk carbonate near the seafloor. Examinations of

673 calcareous nannofossils in Paleogene sediment at Site 577 show extensive calcite

674 overgrowths (Shipboard Scientific Party, 1985; Backman, 1986). Relatively low $\delta^{18}O$ values

675 mark the H-1 and K/X events, as well as the main phase of the EECO (**Figure 5**). Both

676 observations support the idea that the bulk carbonate $\delta^{18}O$ at Site 577 represents the

677 combination of a primary surface water $\delta^{18}O$ signal and a secondary shallow pore water $\delta^{18}O$

678 signal.

679   Lithification should further impact bulk carbonate $\delta^{18}O$ records (Marshall, 1992; Schrag

680 et al., 1995; Frank et al., 1999). Because this process occurs well below the seafloor, where

681 temperatures approach or exceed those of surface water, the $\delta^{18}O$ values of pelagic marls and

682 limestones should be significantly depleted in ^{18}O relative to partially recrystallized

683 nannofossil ooze. This explains the nominal 2‰ offset in average $\delta^{18}O$ between correlative

684 strata at Possagno and at Site 577. While temperature calculations using the $\delta^{18}O$ record at

685 Possagno render reasonable surface water values for a mid-latitude location in the early

686 Paleogene (26-31°C, using the aforementioned approach), any interpretation in these terms

687 more than likely reflects happenstance. The fact that planktic foraminifera are completely

688 recrystallized and totally filled with calcite at Possagno supports this inference.

689   One might suggest, at least for the Possagno section, that meteoric water might have also

690 impacted the $\delta^{18}O$ record. This is because rainwater generally has a $\delta^{18}O$ composition less

691 than that of seawater. However, samples were collected at Possagno in 2002-2003 from fresh

692 quarry cuts.

693   As observed at Site 577, however, horizons of lower $\delta^{18}O$ at Possagno may represent

694 times of relative warmth in surface water. This includes the broad interval between 16 and

695 22.5 m, which marks the main phase of the EECO, as well as many of the brief CIEs, at least

696   one that clearly represents the PETM (**Figure 4**). That is, despite obvious overprinting of the

697   original $\delta^{18}O$ signal, early to early middle Eocene climate variations appear manifest in the

698   data.

699

700   **6.4 The EECO and planktic foraminiferal abundances**

701

702   Bulk carbonate $\delta^{13}C$ records, especially in conjunction with other stratigraphic markers,

703   provide a powerful means to correlate early Paleogene sequences from widely separated

704   locations (**Figure 9**). They also allow for placement of planktic foraminiferal assemblage

705   changes into broader context.

706   The most striking change in planktic foraminiferal assemblages occurred near the start of

707   the EECO. Over a fairly short time interval and at multiple widespread locations, the relative

708   abundance of *Acarinina* ~~acarininids~~ increased significantly whereas the relative abundance of

709   *Morozovella* ~~morozovellids~~ decreased significantly. This switch, best defined by the decline

710   in *Morozovella* ~~morozovellids~~, happened just before the condensed interval at Possagno

711   (**Figure 6**), just above the J event at Site 577 (**Figure 7, Table S4**), and during the J event at

712   Site 1051 (**Figure 8**). At the Farra section, cropping out in the same geological setting of

713   Possagno at 50 km NE of the Carcoselle quarry, it also appears to have occurred close to the J

714   event (**Figure 10**). Indeed, the maximum turnover in relative abundances may have been

715   coincident with the J event at all locations. Importantly, the relative abundance of subbotinids

716   only changed marginally during this time.

717   The *Morozovella* ~~morozovellid~~ decline across the start of the EECO did not rebound

718   afterward. At Possagno, at Site 1051, and at Site 577, it was coupled with the gradual

719   disappearances of several species, including *M. aequa, M. gracilis, M. lensiformis, M.*

720   *marginodentata,* and *M. subbotinae*. Furthermore, the loss of *Morozovella* ~~morozovellids~~ was

721 not counterbalanced by the appearance of the *Morozovelloides* genus, which shared with

722 *Morozovella* the same ecological preferences. This latter genus appeared in C21r, near the

723 Ypresian/Lutetian boundary, and well after the EECO (Pearson et al., 2006; Aze et al., 2011),

724 including at Possagno (Luciani and Giusberti, 2014; **Figure 6**). Though *Morozovelloides*

725 were morphologically similar to *Morozovella*, they probably evolved from *Acarinina*

726 (Pearson et al., 2006; Aze et al., 2011; **Figure 1**).

727   At Possagno, higher abundances of *Acarinina* acarininids also correlate with pronounced

728 negative $\delta^{13}C$ perturbations before and after the EECO (**Figure 6**). This includes the H-1

729 event, as well as several unlabelled CIEs during C22n, C21r and C21n. Such increases in the

730 relative abundances of *Acarinina* acarininids have been described for the PETM interval at

731 the nearby Forada section (Luciani et al., 2007), and for the K/X event at the proximal Farra

732 section (Agnini et al., 2009). Unlike for the main switch near the J event, however, these

733 changes are transient, so that relative abundances in planktic foraminiferal genera are similar

734 before and after the short-term CIEs.

735

736 **6.5 The impact of dissolution**

737

738 Carbonate dissolution at or near the seafloor presents a potential explanation for observed

739 changes in foraminifera assemblages. Some studies of latest Paleocene to initial Eocene age

740 sediments, including laboratory experiments, suggest a general ordering of dissolution

741 according to genus, with *Acarinina* acarininds more resistant than *Morozovella*

742 morozovellids, and the latter more resistant than subbotinids (Petrizzo et al., 2008; Nguyen et

743 al., 2009, 2011).

744   Carbonate solubility horizons that impact calcite preservation and dissolution on the

745 seafloor (i.e., the CCD and lysocline) also shoaled considerably during various intervals of

746 the early Eocene. The three most prominent hyperthermals that occurred before the main

747 phase of the EECO (PETM, H-1, K/X) were clearly marked by pronounced carbonate

748 dissolution at multiple locations (Zachos et al., 2005; Agnini et al., 2009; Stap et al., 2009;

749 Leon-Rodriguez and Dickens, 2010). A multi-million year interval characterized by a

750 relatively shallow CCD also follows the K/X event (Leon-Rodriguez and Dickens, 2010;

751 Pälike et al., 2012; Slotnick et al., 2015b).

752   Should changes in carbonate preservation primarily drive the observed planktic

753 foraminiferal assemblages, it follows that the dominance of *Acarinina* ~~acarininids~~ during the

754 EECO and multiple CIEs could represent a taphonomic artefact. Limited support for this idea

755 comes from our records of fragmentation (*F* index). In general, intervals with relatively high

756 abundances of *Acarinina* ~~acarininids~~ (and low $\delta^{13}C$) correspond to intervals of fairly high

757 fragmentation at Possagno and at Site 1051 (**Figures 6, 8**). This can suggest carbonate

758 dissolution, because this process breaks planktic foraminifera into fragments (Berger, 1967;

759 Hancock and Dickens, 2005).

760   Carbonate dissolution can cause the coarse fraction of bulk sediment to decrease (Berger

761 et al., 1982; Broecker et al., 1999; Hancock and Dickens, 2005). This happens because whole

762 planktic foraminiferal tests typically exceed 63 μm, whereas the resulting fragments often do

763 not exceed 63 μm. The decrease in CF values at the start of the EECO at Possagno (**Figure 6**)

764 may therefore further indicate loss of foraminiferal tests. However, relatively low CF values

765 continue to the top of the section, independent of changes in the *F* index. The CF record

766 parallels the trend of *Morozovella* ~~morozovellids~~ abundance, and thus might also suggest a

767 loss of larger *Morozovella* ~~morozovellids~~ rather than carbonate dissolution.

768   The cause of the long-term rise in carbonate dissolution horizons remains perplexing, but

769 may relate to reduced inputs of ^{13}C-depleted carbon into the ocean and atmosphere (Leon-

770 Rodriguez and Dickens, 2010; Komar et al., 2013). Should the *Morozovella* ~~morozovellids~~

771 decline and amplified *F* index at the Possagno section mostly represent dissolution, it would

772 imply considerable shoaling of these horizons in the western Tethys, given the inferred

773 deposition in middle to lower bathyal setting. As with open ocean sites (Slotnick et al.,

774 2015b), further studies on the Eocene lysocline and CCD are needed from Tethyan locations.

775 One idea is that remineralization of organic matter intensified within the water column,

776 driven by augmented microbial metabolic rates at elevated temperatures during the EECO;

777 this may have decreased pH at intermediate water column depths (Brown et al., 2004;

778 Olivarez Lyle and Lyle, 2006; O'Connor et al., 2009; John et al., 2013, 2014).

779   Despite evidence for carbonate dissolution, this process probably only amplified primary

780 changes in planktic foraminiferal assemblages. The most critical observation is the similarity

781 of the abundance records for major planktic foraminiferal genera throughout the early Eocene

782 at multiple locations (**Figures 6-8**). This includes the section at Site 1051, where carbonate

783 appears only marginally modified by dissolution according to the *F* index values (**Figure 7**).

784 Subbotinid abundance also remains fairly high throughout the early Eocene. One explanation

785 is that, in contrast to laboratory experiments (Nguyen et al., 2009, 2011), subbotinids are

786 more resistant to dissolution than *Morozovella* ~~morozovellids~~ (Boersma and Premoli Silva,

787 1983; Berggren and Norris, 1997), at least once the EECO has transpired. In the proximal

788 middle-upper Eocene section at Alano, Luciani et al. (2010) documented a dominance of

789 subbotinids within intervals of high fragmentation (*F* index) and enhanced carbonate

790 dissolution. The degree of dissolution across planktic foraminiferal assemblages may have

791 varied through the early Paleogene, as distinct species within each genus may respond

792 differently (Nguyen et al., 2011). So far, data on dissolution susceptibility for different

793 species and genera are limited for early and early middle Eocene times (Petrizzo et al., 2008).

794   There is also recent work from the Terche section (ca. 28 km NE of Possagno) to

795 consider. This section is located in the same geological setting as Possagno, but across the H-

796   1, H-2 and I1 events, there are very low *F* index values and marked increases of *Acarinina*

797   ~~acarininids~~ coupled with significant decreases of subbotinids (D'Onofrio et al., 2014).

798   Therefore, although the Possagno record may be partially altered by dissolution, an increase

799   of warm water *Acarinina* ~~acarininids~~ concomitant with decrease of subbotinids seems to be a

800   robust finding during early Paleogene warming events in Tethyan settings.

801

802   **6.6 A record of mixed water change**

803

804   The switch in abundance between *Morozovella* ~~morozovellids~~ and *Acarinina* ~~acaraninids~~ at

805   the start of the EECO supports a hypothesis whereby environmental change resulted in a

806   geographically widespread overturn of planktic foraminiferal genera. During the PETM and

807   K/X events, *Acarinina* ~~acarininids~~ became dominant over *Morozovella* ~~morozovellids~~ in a

808   number of Tethyan successions. This has been interpreted as signifying enhanced

809   eutrophication of surface waters near continental margins (Arenillas et al., 1999; Molina et

810   al., 1999; Ernst et al., 2006; Guasti and Speijer, 2007; Luciani et al., 2007; Agnini et al.,

811   2009), an idea consistent with evidence for elevated (albeit more seasonal) riverine discharge

812   during these hyperthermals (Schmitz and Pujalte, 2007; Giusberti et al., 2007; Schulte et al.,

813   2011; Slotnick et al., 2012; Pułjalte et al., 2015). Increased nutrient availability may also have

814   occurred at Possagno during the early part of the EECO, given the relatively high

815   concentration of radiolarians, which may reflect eutrophication (Hallock, 1987).

816   However, the fact that the major switch at the start of the EECO can be found at Sites

817   1051 (western Atlantic) and Site 577 (central Pacific) suggests that local variations in

818   oceanographic conditions, such as riverine discharge, was not the primary causal mechanism.

819   Rather, the switch must be a consequence of globally significant modifications related to the

820   EECO, most likely sustained high temperatures, elevated pCO$_2$, or both. Given model

821   predictions for our Earth in the coming millennia (IPCC, 2014), indirect effects also could

822   have contributed, especially including increased ocean stratification and decreased pH.

823   An explanation for the shift may lie in habitat differences across planktic foraminiferal

824   genera. Although both *Morozovella* ~~morozovellids~~ and *Acarinina* ~~acaraninids~~ likely had

825   photsymbionts, *Morozovella* ~~morozovellids~~ may have occupied a shallower surface habitat

826   than the latter genus as indicated by minor variations in their stable isotope compositions

827   (Boersma et al., 1987; Pearson et al., 1993; 2001).

828   One important consideration to any interpretation is the evolution of new species that

829   progressively appear during the post-EECO interval. In good agreement with studies of lower

830   Paleogene sediment from other low latitude locations (Pearson et al., 2006), thermocline

831   dwellers such as subbotinids and parasubbotinids seem to proliferate at Possagno (Luciani

832   and Giusberti, 2014). These include *Subbotina corpulenta, S. eocena, S. hagni, S. senni, S.*

833   *yeguanesis, Parasubbotina griffinae,* and *P. pseudowilsoni*. The appearance of the radially-

834   chambered *Parasubbotina eoclava*, considered to be the precursor of the truly clavate

835   chambered *Clavigerinella* (Coxall et al., 2003; Pearson and Coxall, 2014), also occurs at 19.8

836   m, and in the core of the EECO (Luciani and Giusberti, 2014). *Clavigerinella* is the ancestor

837   of the genus *Hantkenina* that successfully inhabited the sub-surface and surface waters during

838   the middle through late Eocene (Coxall et al., 2000).

839   A second consideration is the change in planktic foraminiferal assemblages during the

840   Middle Eocene Climate Optimum (MECO), another interval of anomalous and prolonged

841   warmth ca. 40 Ma (Bohaty et al. ~~and Zachos~~, 2009 ~~2003~~). At Alano (**Figure 11**) and other

842   locations (Luciani et al., 2010; Edgar et al., 2012), the MECO involved the reduction in the

843   abundance and test size of large *Acarinina* ~~acarininids~~ and *Morozovelloides*. This has been

844   attributed to "bleaching" and the loss of photosymbionts resulting from global warming

845   (Edgar et al., 2012), although related factors, such as a decrease in pH, a decrease in nutrient

846   availability, or changes in salinity, may have been involved (Douglas, 2003; Wade et al.,

847   2008). The symbiotic relationship with algae is considered an important strategy adopted by

848   muricate planktic foraminifera during the early Paleogene (Norris, 1996; Quillévéré et al.,

849   2001). Considering the importance of this relationship in extant species (Bé, 1982; Bé et al.,

850   1982; Hemleben et al., 1989), the loss of photosymbionts may represent a crucial mechanism

851   to explain the relatively rapid decline foraminifera utilizing this strategy, including

852   *Morozovella* ~~morozovellids~~ at the start of the EECO.

853   Available data suggest that the protracted conditions of extreme warmth and high pCO$_2$

854   during the EECO were the key elements inducing a permanent impact on planktic

855   foraminiferal evolution, and the decline of *Morozovella* ~~the morozovellids~~. Even ~~during~~ the

856   PETM, the most pronounced hyperthermal, did not adversely affect the genus *Morozovella*

857   ~~morozovellids~~ permanently. While "excursion taxa" appeared, *Morozovella* ~~morozovellids~~

858   seem to have increased in abundance in open ocean settings (Kelly et al., 1996; 1998, 2002;

859   Lu and Keller, 1995; Petrizzo, 2007); only in some continental margin settings did a transient

860   decrease in abundance occur (Luciani et al., 2007).

861

862   **6.7 Post-EECO changes at Possagno**

863

864   Several small CIEs appear in the δ^{13}C record at Possagno during polarity chrons C22n, C21r,

865   and C21n. Some of these post-EECO excursions coincide with planktic foraminiferal

866   assemblage changes similar to those recorded in lower strata. Specifically, there are marked

867   increases of *Acarinina* ~~acarininids~~ (**Figure 6**). These "post-EECO" CIEs are concomitant

868   with δ^{18}O excursions and coupled to distinct modifications in the planktic foraminiferal

869   assemblages comparable to those recorded across known hyperthermals in Tethyan settings

870   (Luciani et al., 2007; Agnini et al., 2009; D'Onofrio et al., 2014). Additional hyperthermals,

871 although of less intensity and magnitude, may extend through the entirety of the early and

872 middle Eocene, as suggested previously (Sexton et al., 2006; 2011; Kirtland-Turner et al.,

873 2014). Whether these imply different forcing and feedback mechanisms compared to the

874 PETM remains an open discussion.

875

876 **7 Summary and conclusions**

877   The symbiont-bearing planktic foraminiferal genera *Morozovella* and *Acarinina* were

878 among the most important calcifiers of the early Paleogene tropical and subtropical oceans.

879 However, a remarkable and permanent switch in the relative abundance of these genera

880 happened in the early Eocene, an evolutionary change accompanied by species reduction of

881 *Morozovella* and species diversification of *Acarinina*. We show here that this switch probably

882 coincided with a carbon isotope excursion (CIE) presently coined J. Although the Early

883 Eocene Climatic Optimum (EECO), a multi-million year interval of extreme Earth surface

884 warmth, lacks an accepted definition, we propose that the EECO is best defined as the

885 duration of time between the J event and the base of *D. sublodoensis* (about 53 Ma to 49 Ma

886 on the 2012 GTS).

887   Our conclusion that the planktic foraminferal switch coincides with the start of the

888 EECO derives from the generation of new records and collation of old records concerning

889 bulk sediment stable isotopes and planktic foraminiferal abundances at three sections. These

890 sections span a wide longitude range of the low latitude Paleogene world: the Possagno

891 section from the western Tethys, DSDP Site 577 from the central Pacific Ocean, and ODP

892 Site 1051 from the western Atlantic Ocean. Importantly, these locations have robust

893 calcareous nannofossils and polarity chron age markers, although the stratigraphy required

894 amendment at Sites 577 and 1051.

895   An overarching problem is that global carbon cycling was probably very dynamic during

896   the EECO. The interval appears to have been characterized not only by numerous CIEs, but

897   also a major switch in the timing and magnitude of these perturbations. Furthermore, there

898   was a rapid shoaling of carbonate dissolution horizons in the middle of the EECO. A key

899   finding of our study is that the major switch in planktic foraminiferal assemblages happened

900   at the start of the EECO. Significant, though ephemeral, modifications in planktic

901   foraminiferal assemblages coincide with numerous short-term CIEs, before, during and after

902   the EECO. Often, there are marked increases in the relative abundance of *Acarinina*

903   ~~acarininids~~, similar to what happened permanently across the start of the EECO.

904   Although we show for the first time that the critical turnover in planktic foraminifera

905   clearly coincided with the start of the EECO, the exact cause for the switch (aka the decline

906   of *Morozovella* ~~morozovellids~~) remains elusive. Possible causes are multiple, and may

907   include temperature effects on photosymbiont-bearing planktic foraminifera, changes in

908   ocean chemistry, or even interaction with other microplankton groups such as radiolarians,

909   diatoms or dinoflagellates that represented possible competitors in the use of symbionts or as

910   symbiont providers. For some reason, a critical threshold was surpassed at the start the

911   EECO, and this induced an unfavourable habitat for continued *Morozovella* ~~morozovellid~~

912   diversification and proliferation but a favourable habitat for the genus *Acarinina* ~~acarinids~~.

913

931

932 **References**

933

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1511

1512   **Figure Captions**

1513

1514   **Figure 1.** Evolution of climate, carbon cycling, and planktic foraminifera across the middle

1515   Paleogene on the GPTS 2012 time scale. Left side shows polarity chrons, and smoothed

1516   oxygen and carbon isotope records of benthic foraminifera, slightly modified from

1517   Vandenberghe et al. (2012). Original oxygen and carbon isotope values come from

1518   compilations by Zachos et al. (2008) and Cramer et al. (2009). Middle of the figure indicates

1519   planktic foraminiferal biozones by Wade et al. (2011) with three modifications. The lower

1520   boundary for Zone E7a is now based on the first occurrence of *Astrorotalia palmerae* due to

1521   diachroneity in the first appearance of the previously selected marker *Acarinina*

1522 *cuneicamerata* (Luciani and Giusberti, 2014). The base of Zone E5, identified by the first

1523 appearance of *Morozovella aragonensis,* occurs within the middle of C24n instead of lower

1524 C23r (see text). A question marks the top of *Morozovella subbotinae* because there is

1525 diachroneity for this occurence (see text). Right side shows a partial view of *Morozovella*

1526 ~~morozovellid~~ and *Acarinina* ~~acarininid~~ evolution as envisioned by Pearson et al. (2006) and

1527 Aze et al. (2011). It does not include several "root taxa" that disappear in the earliest Eocene

1528 (e.g., *M. velascoensis*) or "excursion taxa" that appear during the Paleocene-Eocene Thermal

1529 Maximum (PETM) (e.g., *M. allisonensis*). Superimposed on these records are key intervals of

1530 climate change, including the Early Eocene Climatic Optimum (EECO), the Middle Eocene

1531 Climatic Optimum (MECO) and the three well documented early Eocene hyperthermal

1532 events. The extent of the EECO is not precise, because of stratigraphic issues (see text). Red

1533 and blue triangles= top and base of the *Morozovella* ~~morozovellid~~ and *Acarinina* ~~acarininid~~

1534 zonal markers.

1535

1536 **Figure 2.** Approximate locations of the three sites discussed in this work during the early

1537 Eocene. Also shown is Site 1258, which has a bulk carbonate $\delta^{13}C$ record spanning the

1538 EECO. Base map is from http://www.odsn/de/services/paleomap.html with paleolatitudes

1539 modified for Sites 577, 1051 and 1258 according to www.paleolatitude.org model version 1.2

1540 (Van Hinsbergen et al., 2015). Possagno paleolatitude is referred to the

1541 http://www.odsn.de/odsn/services/paleomap/adv_map.html model since it is not yet

1542 available at http://www.odsn.de/services/paleomap.html.

1543

1544 **Figure 3**. The Possagno section. Upper panel: geological map (modified from Braga, 1970).

1545 1 = Quaternary deposits; 2, 3 = Calcarenite di Castelcucco (Miocene); 4 = glauconitic

1546 arenites (Miocene); 5 = siltstones and conglomerates (upper Oligocene-lower Miocene); 6 =

1547   Upper Marna di Possagno (upper Eocene); 7 = Formazione di Pradelgiglio (upper Eocene); 8

1548   = Marna di Possagno (upper Eocene); 9 = Scaglia Cinerea (middle-upper Eocene); 10 =

1549   Scaglia Rossa (upper Cretaceous-lower Eocene); 11 = faults; 12 = traces of stratigraphic

1550   sections originally studied by Bolli (1975); red circle = the Carcoselle quarry. Lower panel:

1551   the exposed quarry face during Summer 2002 (Photo by Luca Giusberti).

1552

1553   **Figure 4.** Lithology, stratigraphy, and bulk sediment stable-isotope composition of the

1554   Possagno section aligned according to depth. Litholologic key: 1 = limestone; 2 = marly

1555   limestone and calcareous marl; 3 = cyclical marl-limestone alternations, 4 = marl; 5= Clay

1556   Marl unit (CMU). Planktic foraminiferal biozones follow those of Wade et al. (2011), as

1557   modified by Luciani and Giusberti (2014). Magnetostratigraphy and key calcareous

1558   nannofossil events come from Agnini et al. (2006); NP-zonation is from Martini (1971).

1559   Nannofossil events are shown as red triangles (tops), blue triangles (bases), and purple

1560   diamonds (evolutionary crossovers); *S. rad.* = *Sphenolithus radians*; T.c./T.o. = *Tribrachiatus*

1561   *contortus/ Tribrachiatus orthostylus; D. lod.* = *Discoaster lodoensis*; Tow. =*Toweius*; *T. orth.*

1562   = *Tribrachiatus orthostylus*; *D. sublod.* = *Discoaster sublodoensis*. Stable isotope records

1563   determined in this study. Established early Eocene "events" are superimposed in light red;

1564   suggested carbon isotope excursions (CIEs) within the EECO are shown with numbers.

1565

1566   **Figure 5.** Cores, stratigraphy, and bulk sediment stable isotope composition for the early

1567   Eocene interval at Deep-Sea Drilling Project (DSDP) Site 577 aligned according to

1568   composite depth (Dickens and Backman, 2013). Note the increased length for the gap

1569   between Core 577*-8H and Core 577*-9H (see text). The Wade et al. (2011) E-zonation,

1570   partly modified by Luciani and Giusberti (2014), has been applied to Site 577 given

1571   assemblages presented by Lu (1995) and Lu and Keller (1995). Note that: (a) the base of

1572 Zone E3 (top of *Morozovella velascoensis*) lies within a core gap; (b) the E4/E5 zonal

1573 boundary (base of M. *aragonensis*) occurs within C24n, in agreement with Luciani and

1574 Giusberti (2014); (c) the E5/E6 zonal boundary is problematic because the top of *M.*

1575 *subbotinae* occurs in middle C24n, much earlier than the presumed disappearance in the

1576 upper part of C23n (Wade et al., 2011). We have therefore positioned the E5/E6 boundary at

1577 the lowest occurrence of *Acarinina aspensis,* according to the original definition of Zone E5

1578 (Berggren and Pearson, 2005); (d) we cannot differentiate between Zone E6 and Zone E7a

1579 due to the absence of *Astrorotalia palmerae* and to the diachronous appearance of *A.*

1580 *cuneicametrata* (Luciani and Giusberti, 2014)*.* Magnetostratigraphy and key calcareous

1581 nannofossil events are those summarized by Dickens and Backman (2013). For the latter and

1582 beyond that noted for **Figure 4**: *F.* spp. = *Fasciculithus spp.*; *D. dia. = Discoaster diastypus.*

1583 Stable isotope records: black - Cramer et al. (2003), red and blue - this study. Early Eocene

1584 "events" are the same as those in **Figure 4**.

1585

1586

1587 **Figure 6.** The Possagno section and its δ^{13}C record (**Figure 4**) with measured relative

1588 abundances of primary planktic foraminiferal genera, fragmentation index (*F* index) and

1589 coarse fraction. The subbotinid abundance includes both *Subbotina* and *Parasubbotina*

1590 genera. Note that a significant increase in *Acarinina* abundance marks the EECO and several

1591 carbon isotope excursions (CIEs). Note also the major decline in abundance of *Morozovella*

1592 at the start of the EECO. Filled yellow hexagons show occurrences of abundant radiolarians.

1593 Lithological symbols and early Eocene "events" are the same as those in **Figure 4**.

1594

1595 **Figure 7.** The early Eocene succession at DSDP Site 577 and its δ^{13}C record (**Figure 5**) with

1596 relative abundances of primary planktic foraminiferal genera (Lu, 1995; Lu and Keller,

1597   1995). Note the major switch in *Morozovella* and *Acarinina* abundances approximately

1598   coincides with the J-event, the top of polarity chron C24n, and the start of the EECO. Early

1599   Eocene "events" are the same as those in **Figure 4**.

1600

1601   **Figure 8.** Stratigraphy, bulk sediment $\delta^{13}C$ composition, relative abundances of primary

1602   planktic foraminiferal genera, and fragmentation index (*F* index) for the early Eocene interval

1603   at ODP Site 1051. Planktic foraminiferal biozones follow those of Wade et al. (2011), as

1604   modified by Luciani and Giusberti (2014; see **Figure 1** caption). Magnetostratigraphy and

1605   positions of key calcareous nannofossil events come from Ogg and Bardot (2001) and Mita

1606   (2001), but with an important modification to polarity chron labelling (see text and Cramer et

1607   al., 2003). Calcareous nannofossil horizons are the same as in previous figures. Foraminferal

1608   information comes from this study; subbotinids include both *Subbotina* and *Parasubbotina*.

1609   Early Eocene "events" are the same as those in **Figure 4**.

1610

1611   **Figure 9.** Carbon isotope and paleomagnetic records across the early Eocene for the

1612   Possagno section, DSDP Site 577, and ODP Site 1258 (Kirtland-Turner et al., 2014). This

1613   highlights the overall framework of carbon cycling in the early Eocene, but also stratigraphic

1614   problems across the EECO at each of the three sites. At Possagno, the coarse resolution of

1615 $\delta^{13}C$ records and the condensed interval makes correlations difficult. At ODP Site 1258 the

1616   prominent K/X event seems missing. At DSDP Site 577, the entire record is compressed in

1617   the depth domain. Nonetheless, a major shift in frequency and amplitude of carbon isotope

1618   excursions (CIEs) appears to have happened during the EECO. CIEs that suggestively

1619   correlate within the EECO are shown with numbers.

1620

1621   **Figure 10.** Records of magnetostratigraphy, bulk sediment $\delta^{13}C$, $CaCO_3$ content, *F* index and

1622   abundance patterns for primary planktic foraminiferal taxa at the Farra section, which crops

1623   out 50 km NE of Possagno. All data are from Agnini et al. (2009). Note that the switch in

1624   abundance between *Morozovella* and *Acarinina* occurs close the J event.

1625

1626   **Figure 11**. Records of *Morozovella* ~~morozovellids~~ and large *Acarinina* ~~acarininids~~ (>200

1627   micron) in the western Tethyan setting from the Possagno section (this paper) and the Alano

1628   section (Luciani et al., 2010), plotted with generalized δ^{13}C and δ^{18}O curves for benthic

1629   foraminiferal on the GTS2012 time scale (as summarized by Vandenberghe et al., 2012;

1630   slightly modified). These records suggest that the long-lasting EECO and MECO intervals of

1631   anomalous warmth mark two main steps in the decline of *Morozovella*, *Morozovelloides*

1632   ~~morozovellids~~ and *Acarinina* ~~acarininids~~. The plankic foraminferal biozones follow those

1633   presented by Wade et al. (2011), as partly modified by Luciani and Giusberti (2014).

1634

1635   **Supplementary material**

1636

1637   **Table S1**. Carbon and oxygen isotopes from the Possagno section.

1638

1639   **Table S2**. Carbon and oxygen isotopes from DSDP Site 577.

1640

1641   **Table S3**. Foraminiferal abundances, fragmentation index (%) and coarse fraction (%) from

1642   the Possagno section.

1643

1644   **Table S4**. Foraminiferal abundances from DSDP Site 577.

1645

1646   **Table S5**. Foraminiferal abundances from ODP Site 1051.

1647

1648   **Figure S1.** The Possagno $\delta^{13}C$ data and relative abundance of minor planktic foraminiferal

1649   genera and selected species plotted against lithology and fragmentation index (*F* index) data.

1650   Magnetostratigraphy is from Agnini et al. (2006). The planktic foraminferal biozonal scheme

1651   is from Wade et al. (2011), as modified by Luciani and Giusberti (2014). Various symbols are

1652   the same as in **Figure 4**.

1653

1654   **Appendix A: Taxonomic list of planktic foraminiferal species cited in text and figures**

1655

1656   *Globanomalina australiformis* (Jenkins, 1965)

1657   *Morozovella aequa* (Cushman and Renz, 1942)

1658   *Morozovella gracilis* (Bolli, 1957)

1659   *Morozovella lensiformis* (Subbotina, 1953)*,*

1660   *Morozovella marginodentata* (Subbotina, 1953)

1661   *Morozovella subbotinae* (Morozova, 1939)

1662   *Parasubbotina eoclava* Coxall*,* Huber and Pearson, 2003

1663   *Parasubbotina griffinae* (Blow, 1979)

1664   *Parasubbotina pseudowilsoni* Olsson and Pearson, 2006

1665   *Subbotina corpulenta* (Subbotina, 1953)

1666   *Subbotina eocena* (Gümbel, 1868)

1667   *Subbotina hagni* (Gohrbandt, 1967)

1668   *Subbotina senni* (Beckmann, 1953)

1669   *Subbotina yeguanesis* (Weinzierl and Applin, 1929*)*

1670   *Planoglobanomalina pseudoalgeriana* Olsson & Hemleben, 2006

1671

Appendix B: Taxonomic list of calcareous nannofossil taxa cited in text and figures

Globanomalina australiformis (Jenkins, 1965)

Morozovella aequa (Cushman and Renz, 1942)

Morozovella gracilis (Bolli, 1957)

Morozovella lensiformis (Subbotina, 1953)*,*

Morozovella marginodentata (Subbotina, 1953)

Morozovella subbotinae (Morozova, 1939)

Parasubbotina eoclava Coxall*,* Huber and Pearson, 2003

Parasubbotina griffinae (Blow, 1979)

Parasubbotina pseudowilsoni Olsson and Pearson, 2006

Subbotina corpulenta (Subbotina, 1953)

Subbotina eocena (Gümbel, 1868)

Subbotina hagni (Gohrbandt, 1967)

Subbotina senni (Beckmann, 1953)

Subbotina yeguanesis (Weinzierl and Applin, 1929*)*

Planoglobanomalina pseudoalgeriana Olsson ~~&~~ and Hemleben, 2006

Appendix B: Taxonomic list of calcareous nannofossil taxa cited in text and figures

Discoaster diastypus Bramlette and Sullivan, 1961

Discoaster lodoensis Bramlette and Sullivan, 1961

Discoaster sublodoensis Bramlette and Sullivan, 1961

Fasciculithus Bramlette and Sullivan, 1961

Fasciculithus tympaniformis Hay and Mohler in Hay et al., 1967

1697   *Sphenolithus radians* Deflandre in Grassé, 1952

1698   *Toweius* Hay and Mohler, 1967

1699   *Tribrachiatus contortus* (Stradner, 1958) Bukry, 1972

1700   *Tribrachiatus orthostylus* (Bramlette and Riedel, 1954) Shamrai, 1963