Dear authors,
sorry for the delay, but it took me some times to carefully check your statistical analysis. I was particularly worried by your fig 10. You mention that the PCA analysis allows to discriminate between three different faunal assemblages. You justify this result by the possibility of grouping the data points of fig 10 into three shaded areas. The problem I see is that these three shaded area have been drawn by hand, and do not seem to be delimited by statistical criteria. Using your dataset, I calculated the ellipse containing 95% nd . If you proceed like this, group 3 (PETM) is not clearly distinguishable from group 1. This similarity is sustained by further analyses (e.g., cluster) even if some of the traditional statistical tests attest their significant difference (variance test). The conclusion should then be that the faunal assemblages of group 1 and 3 are similar, although group 3 displays a greater variability. This point must be clarified. I suggest you to use discriminant analysis methods, like the LDA, or variance tests (such as MANOVA). This would really strengthen your conclusions.

In Figure 10, I described my samples as subdividable in three different floral assemblages (calcareous nannoplankton are algae) and I have never claimed these groups represent ellipses containing 95% of the data points. Anyway, this is a good point and I have added 95% ellipses for Paleocene and Eocene samples in Figure 10. As you wrote, the PETM samples are obviously widely dispersed and are not surrounded by any ellipse anymore.

What is still pretty clear from PCA is that Paleocene and post-PETM Eocene are statistical different. This is not obvious from what you can get from the literature available. In fact, statistical analyses performed (see for an instance Gibbs et al., 2006) are conceived to point out extinction and speciation rates during the event rather than changes in the relative abundance of the assemblages. Our approach is based on the fact that the changes in abundance of the major components of the assemblages are as much important as the extinction and speciation rates and give a different perspective to the variations observed in the PETM calcareous nannoplanton datasets.

Regarding the difference between group 1 and group 3, I do not totally agree when you wrote that *"the floral assemblages of group 1 and 3 are similar, although group 3 displays a greater variability"* because most of the samples representing the CORE CIE are in fact distinguible using Component 2 in the PCA analysis. To support my view, I have followed your advice and I have applied a variance analysis (MANOVA) to my dataset. The result is that the three groups are easily distinguishable one from each other.

To address your comment, I have added an additional figure, Figure 10B, which is a scatter graph + biplot plot obtained applying the MANOVA technique. The text has been changed accordingly.

Also, it is not clear whether the "reworking" data should be included in the analysis, since you lost the stratigraphical signal.

As well explained in the available literature, during the late Paleocene and early Eocene hyperthermals, the weathering is enhanced (e.g., Zachos et al. 2006 and reference herein) and, as a consequence, reworked calcareous nannofossils are transported to the ocean and then display a remarkable increase in their relative abundance especially in marginal setting (e.g., Agnini et al. 2007; 2009). This group is thus considered a good proxy for increase weathering and used as a paleoceanographic indicator.

In addition, I fully understand the reasons of log-transforming the raw dataset. However I question the validity of the replacing values at -0.698970004 as 0 is an informative value in your study. This must be clarified. I would finally suggest you to use abundance data, despite their numerous biases, in the calculation of diversity indices of similarity between the different groups. This should help to discuss about the faunal turnover through the PETM.

With regard to your comment, I think that we have fully answered to reviewer 1 and followed her/his suggestions about how to implement the raw data in order to 1) avoid the closed-sum effect and 2) decrease the sensitivity of PCA to zero values (see reviewer 1's comment on this issue). This is essentially the reasons why we have modified our original statistical analysis and this has been done to follow your previous suggestion. Since you understand the reason why we accepted reviewer 1's suggestion, you may

also understand the importance of possibly reducing the zero value effect. These changes do not substantially modify the results obtained with raw data but I would maintain these two mat transformations because even if they decrease the values of PC1 and PC2, they guarantee for a better fit of the dataset to the statistical method applied in this study.

Regarding fig 6, what are the dashed lines representing ?
Good point. Grey and black dash lines are linear trendlines for Paleocene and Eocene samples, respectively. This has been added to the caption of Figure 6.

Finally, the abbreviations used in the data tables are not described in the text and all cells must be filled.
I'm not sure to have understood to which figures you refer to. You may refer to supplementary figures since I do not see this problem in Table 1. In this table, all the abbreviations are explained in the text or as footnotes directly on the table caption.
Table S1. There are no abbreviations.
Table S2. There are no abbreviations. Explanation of the content of each excel sheet is provided in the read me file (READ ME_Supplementary tables.docx).
Table S3. There are no abbreviations. I have added the description of the first column (Sample CIC/07) that was previously missed.
Table S4. There are no abbreviations. I have added the description of the first column (Sample CIC/07) that was previously missed.

I hope to have answered to your extra comments satisfactorily. Looking to hearing from you soon for your final decision.

Best Regards,
Claudia Agnini

1

Stable isotope and calcareous nannofossil assemblage record of the late Paleocene

and early Eocene (Cicogna section)

4

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20

21 **Abstract.** We present records of stable carbon and oxygen isotopes, $CaCO_3$ content, and changes in

22 calcareous nannofossil assemblages across an 81 m thick section of upper Paleocene-lower Eocene

23 marine sedimentary rocks now exposed along the Cicogna Stream in northeast Italy. The studied

24 stratigraphic section represents sediment accumulation in a bathyal hemipelagic setting from

25 approximately 57.5 to 52.2 Ma, a multi-million-year time interval characterized by perturbations in

26 the global carbon cycle and changes in calcareous nannofossil assemblages. The bulk carbonate $\delta^{13}C$

27 profile for the Cicogna section, once placed on a common time scale, resembles that at several other

28 locations across the world, and includes both a long-term drop in $\delta^{13}C$, and multiple short-term

29 carbon isotope excursions (CIEs). This precise correlation of widely separated $\delta^{13}C$ records in marine

30 sequences results from temporal changes in the carbon composition of the exogenic carbon cycle.

31 However, diagenesis has likely modified the $\delta^{13}C$ record at Cicogna, an interpretation supported by

32 variations in bulk carbonate $\delta^{18}O$, which do not conform to expectations for a primary signal. The

33 record of $CaCO_3$ content reflects a combination of carbonate dilution and dissolution, as also

34 inferred at other sites. Our detailed documentation and statistical analysis of calcareous nannofossil

35 assemblages show major differences before, during and after the Paleocene Eocene Thermal

36 Maximum. Other CIEs in our lower Paleogene section do not exhibit such a distinctive change;

37 instead, these events are sometimes characterized by variations restricted to a limited number of

38 taxa and transient shifts in the relative abundance of primary assemblage components. Both long-

39 lasting and short-lived modifications to calcareous nannofossil assemblages preferentially affected

40 nannoliths or holococcoliths such as *Discoaster, Fasciculithus, Rhomboaster/Tribrachiatus,*

41 *Spenolithus* and *Zygrhablithus*, which underwent distinct variations in abundance as well as

42 permanent evolutionary changes in terms of appearances and disappearances. By contrast,

43 placoliths such as *Coccolithus* and *Toweius*, which represent the main component of the

44 assemblages, were characterized by a gradual decline in abundance over time. Comparisons of

2

45   detailed nannofossil assemblage records at the Cicogna section and at ODP Site 1262 support the

46   idea that variations in the relative and absolute abundances, even some minor changes, were

47   globally synchronous. An obvious link is through climate forcing and carbon cycling, although the

48   linkages between variations in calcareous nannoplankton, changes in $\delta^{13}C$ records and

49   oceanography will need additional work.

50

51   **INDEX TERMS**

52   Paleocene, Eocene, calcareous nannofossils, stable isotopes, paleoclimate, Tethys

53

54   **1   INTRODUCTION**

55

56   A remarkable interval of global warming occurred from the middle Paleocene to the early Eocene,

57   between approximately 59 and 51 million years ago (Ma). This inference comes from a variety of

58   proxies (Huber and Caballero, 2011; Hollis et al., 2012), including the stable oxygen isotope ($\delta^{18}O$)

59   composition of benthic foraminifera (**Figure 1**). The precise timing of the long-term temperature

60   rise remains somewhat unconstrained, because absolute ages across the early Eocene remain

61   unsolidified. Throughout this work, we assume that the Option-1 (WO-1) time scale presented by

62   Westerhold et al. (2008) is correct (**Table 1**), but acknowledge that an offset of ca 400 kyr may occur

63   within the time interval of interest (Vandenberghe et al., 2012). Debate also surrounds the

64   magnitude and distribution of the temperature warming. Earth's surface temperatures, at least at

65   high latitudes and in the deep sea, seem to have risen by at least 6°C from ca. 59 to 51 Ma (Zachos

66   et al., 2008; Bijl et al., 2009; Huber and Caballero, 2011; Hollis et al., 2012). Indeed, the latter date

67   marks the acme of the Early Eocene Climatic Optimum (EECO), the warmest sustained time interval

68   of the Cenozoic (Zachos et al., 2008; Cramer et al., 2009; Hollis et al., 2012). Such a rise in

69 temperature is not obvious at low latitudes with current data (Pearson et al., 2007; Huber et al.,

70 2011).

71   Somehow related to long-term global warming were a series of major perturbations in the

72 global carbon cycle, as clearly indicated by stable carbon isotope (δ^{13}C) records in benthic

73 foraminifera (**Figure 1**) and bulk carbonate in numerous marine sediment sequences (Shackleton,

74 1986; Corfield, 1994; Cramer et al., 2003; Zachos et al., 2008; 2010; Westerhold et al., 2011; Slotnick

75 et al., 2012). An overall increase in δ^{13}C occurred through most of the Paleocene, which climaxed in

76 a Cenozoic high at ca. 57.5 Ma (Westerhold et al., 2011), commonly referred to as the Paleocene

77 carbon isotope maximum (PCIM). From this time, δ^{13}C generally decreased to ca. 52.5 Ma. However,

78 when examined at higher temporal resolution, multiple δ^{13}C records show several short-term (<200

79 kyr) negative carbon isotope excursions (CIEs) (Cramer et al., 2003; Lourens et al., 2005; Nicolo et

80 al., 2007; Agnini et al., 2009; Zachos et al., 2010; Slotnick et al., 2012). Some of these CIEs clearly

81 coincided with rapid warming (above references). The most prominent and most widely

82 documented example of these "hyperthermals" was the Paleocene-Eocene Thermal Maximum

83 (PETM) at ca. 55.5 Ma, but other apparently similar events occurred at ca. 53.7 Ma (H1 or Eocene

84 Thermal Maximum 2, ETM-2), and at ca. 52.5 Ma (K/X, sometimes called ETM-3).

85   The early Paleogene in general, and the hyperthermals in particular, have attracted

86 considerable geoscience research. On one level, this is because these time intervals represent a

87 range of possible past analogues for understanding the effects of global warming and massive

88 carbon emissions (cf. Keeling and Whorf, 2004; Zachos et al., 2008). On another level, this is because

89 the long-term and short-term temperature and carbon cycle perturbations provide new

90 perspectives for how systems on Earth's surface operate. The PCIM probably represents a

91 tremendous storage of ^{13}C-depleted carbon somewhere on Earth's shallow surface (Shackleton,

92 1986; Kurtz et al., 2003; Komar et al., 2013). In turn, the CIEs probably signify rapid and large inputs

93   of ^{13}C-depleted carbon into the ocean and atmosphere (Dickens et al., 1997; Lourens et al., 2005;

94   Zeebe et al., 2009). The middle Paleocene through early Eocene shows us that Earth's climate and

95   carbon reservoirs were extremely dynamic during past times of global warmth. However, the

96   composition and whereabouts of large quantities of transferable ^{13}C-depleted carbon (e.g., seafloor

97   methane, peat, permafrost) remain uncertain (above references). Indeed, it is not clear if and how

98   the long-term and short-term carbon cycle perturbations were related to one another, or to Earth

99   surface temperatures.

100   The above context presents a series of basic questions to the geoscience community. Two of

101   these are the focus of our study: (1) What is the correct template for understanding carbon cycling

102   during the early Paleogene? Major changes in fluxes of ^{13}C-depleted carbon to the ocean or

103   atmosphere should give predictable and coherent signals in the δ^{13}C of carbon-bearing phases

104   across Earth, as well as the distribution of carbonate dissolution on the seafloor. This is not yet

105   established. For example, several recently published δ^{13}C records (Kirtland-Turner et al., 2014;

106   Slotnick et al., 2015a; Payros et al., 2015) do not precisely correlate with those at other locations

107   (Cramer et al., 2003; Zachos et al., 2010; Slotnick et al., 2012, 2015b), at least with available

108   stratigraphy. (2) How did marine calcifying organisms respond to major early Paleogene

109   perturbations in temperature and carbon cycling, both in terms of evolution and preservation? The

110   prominent changes in temperature and carbon fluxes almost assuredly caused large variations in

111   seawater pH and carbonate ion concentration (CO_3^{2-}) (Dickens et al., 1997; Zachos et al., 2005; Kump

112   et al., 2009; Zeebe et al., 2009; Leon-Rodriguez and Dickens, 2010; ; Hönisch et al., 2012; Pälike et

113   al., 2012), although the response should depend on location and carbon fluxes involved (Dickens,

114   2000; Zeebe and Westbroek, 2003; Komar et al., 2013). Such changes might also affect the ability of

115   living organisms to calcify (Riebesell et al., 2000, 2008; Kleypas et al., 2006; Iglesias-Rodriguez et al.,

116   2008; Stillman and Paganini, 2015), which might impact the fossil record (Agnini et al., 2006; Raffi

117   and De Bernardi, 2008; Erba et al., 2010; Hönisch et al., 2012).

118   In regards to both questions, calcareous nannoplankton are an obvious group of organisms to

119   focus on. This is because they are a main component of open ocean primary production (Milliman,

120   1993; Winter et al., 1994; Rost and Riebesell, 2004), because they are the dominate the output of

121   carbonate in the ocean (Ziveri et al., 1999; Hay, 2004), and because they exhibit marked changes in

122   species composition from the middle Paleocene through the early Eocene (Romein, 1979; Aubry,

123   1998, Bown et al., 2004; Gibbs et al., 2012). While numerous studies have examined calcareous

124   nannofossils across the PETM from different perspectives (e.g., Bralower, 2002; Stoll and Bains,

125   2003; Gibbs et al., 2006a; 2006b; Agnini et al., 2007a; Mutterlose et al. 2007; Bown and Pearson,

126   2009; Jiang and Wise, 2009, Self-Trail et al., 2012), the relationship between these organisms and

127   carbon cycle perturbations before and after this short-lived warming episode remain poorly

128   documented (Gibbs et al., 2012). It seems possible that the high rate of calcareous nannofossil

129   taxonomic evolution (appearances and extinctions), as well as distinct changes in calcareous

130   nannofossil abundance patterns may provide excellent stratigraphic control across the early

131   Paleogene (Bukry, 1973; Perch-Nielsen, 1985; Backman, 1986, Agnini et al., 2014). Moreover, if the

132   exact relationship between changes in nannofossil assemblages and global carbon cycle

133   perturbations were known, key time intervals could be rapidly identified for more detailed work.

134   Finally, changes in calcareous nannofossils across the early Paleogene provide insights about the

135   response of an important part of the sediment forming marine biota to changes in climate and

136   carbon cycling.

137   Very few stratigraphic sections presently have detailed and coupled records of stable isotopes,

138   carbonate content, and calcareous nannofossil abundances across the broad late Paleocene-early

139   Eocene interval. The two notable exeptions are Ocean Drilling Program (ODP) Site 1262 (southeast

6

140 Atlantic) (Agnini et al., 2007b; Zachos et al., 2010) and Deep Sea Drilling Project (DSDP) Site 577

141 (northwest Pacific) (Shackleton, 1986; Dickens and Backman, 2013) (**Figure 2**). Here we present

142 geochemical records ($\delta^{13}C$, $\delta^{18}O$ and $CaCO_3$ content) and calcareous nannofossil census data from

143 the Cicogna section in northeast Italy (**Figures 2, 3**). These data are compared with similar

144 information from Sites 1262 and 577. We show that the Cicogna section provides an important

145 template for understanding potential relationships between climate, carbon cycling and the biotic

146 evolution of calcareous nannoplankton.

147

148 **2   THE CICOGNA SECTION**

149

150 The Cicogna section crops out along the Cicogna Stream near the village of Tassei in the Belluno

151 Province, northeast Italy (**Figure 3**). From a regional geological perspective, the sedimentary rocks

152 of this section belong to the Belluno Basin. This basin represents part of a paleogeographic domain

153 that formed when Jurassic rifting created a series of N–S oriented structural highs (platforms) and

154 lows (basins), which persisted through much of the Paleogene (Bernoulli and Jenkyns, 1974;

155 Bernoulli et al., 1979; Winterer and Bosellini, 1981). Importantly, from the Cretaceous to the middle-

156 late Eocene, expanded deep sea sediment successions accumulated within the basins at nominally

157 30°N latitude (Stefani and Grandesso, 1991; Agnini et al., 2006; 2011; Zattin et al., 2006).

158   The Cicogna section consists of two lithostratigraphic units **(Figure 3)**. The lower portion is a

159 well-exposed upper Paleocene and lower Eocene unit referred to as Scaglia Rossa *sensu lato* (**Figures**

160 **3, 4**) (Giusberti et al., 2007; Dallanave et al., 2009). Based on benthic foraminiferal assemblages, the

161 early Paleogene marls of this unit represent lithified pelagic and hemipelagic sediment that

162 accumulated at middle to lower bathyal water depth, likely between 600 m and 1000 m and not

163 deeper than 1500 m (Giusberti et al., 2007; 2015). The upper portion is a thick early to middle

164 Eocene unit called the Belluno Flysch (**Figures 3, 4**). This unit represents a synorogenic deposit on

165 the flanks of the former Trento and Friuli platforms (Grandesso, 1976; Doglioni and Bosellini, 1987).

166   Once corrected for bed strike and dip (ca. 315°N; ca. 45°) and bends in the stream, the Scaglia

167 Rossa at Cicogna measures 80 m in terms of stratigraphic height (Dallanave et al., 2009). The Belluno

168 Flysch measures 1 m in the Cicogna section. Furthermore, the section of interest can be subdivided

169 into several subunits (**Figure 4**). The lower 20 m is comprised of distinctive alternating beds of gray-

170 greenish to purple marls and calcareous marls, the latter defined by carbonate contents higher than

171 60% (**Figure 3c**). This is overlain by approximately 9 m of pink-red marls with much less lithologic

172 alternation. At 28.7 m, the sedimentary package is broken sharply by an approximately 3 m thick

173 red to brownish-red interval of clayey marls with sporadic grey-green cm-scale spots and lenses

174 (**Figure 3f, g**). This has been called the Clay Marl Unit (CMU), and records the core of the prominent

175 negative δ^{13}C excursion associated with the PETM at multiple outcrop sites within the Belluno Basin

176 (Agnini et al., 2006; 2007a; Giusberti et al., 2007). Above the CMU, from 31.7 to 39.2 m, the section

177 continues with deposition of rhythmic alternations of marls and calcareous marls (**Figure 3g**). Above

178 this 8.5 m thick interval, at ca. 40.5 m, spatic calcite crystals occur. Generally, couplets of marl and

179 calcareous marl couplets become less evident from 40.5 m until 54 m, where such couplets reoccur

180 (**Figure 3d**). At 75.5 m, a thin calcarenitic bed is encountered, presaging the onset of the Belluno

181 Flysch. This turbidite is followed by a temporary return to hemipelagic sedimentation that ends at

182 80.6 m. Above, sedimentation of the Belluno Flysch begins in earnest (**Figures 3b, 4**).

183   The basic stratigraphy of the Scaglia Rossa in the Cicogna section, including both polarity chron

184 boundaries and key calcareous nannofossil biohorizons has been published (Giusberti et al., 2007;

185 Dallanave et al., 2009). The combined biomagnetostratigraphy indicates that the 81 m of interest

186 spans polarity Chron C25r to Chron 23r, and calcareous nannofossil biozones CP6 to CP10 (Okada

187 and Bukry, 1980) or CNP10 to CNE4 (Agnini et al., 2014). Thus, the section represents a 5.3 million

188   year (Myr) long time interval, from about 57.5 to 52.2 Ma on the W0-1 time scale. This also implies

189   an average sedimentation rate (SR) of ca. 15 m/myr. Although the deposition of hemipelagic

190   sediment might suggest relatively constant SRs over time, the PETM and possibly other

191   hyperthermal events in the Belluno Basin were characterized by higher SRs (Giusberti et al. 2007;

192   Agnini et al., 2009; Tipple et al., 2011; Krishnan et al., 2015).

193   The Scaglia Rossa at Cicogna appears to record fairly continuous sediment accumulation at

194   moderately high deposition rates. This is important because it affords longer time duration than

195   most shallow ocean sites, greater time resolution than most deep ocean sites (**Figure 2**), and an

196   overall different environmental setting. Many early Paleogene records, especially those from paleo-

197   shelf environments, such as in Egypt (e.g, Aubry and Salem, 2012) and New Jersey (Mixon and

198   Powars, 1994; Harris et al., 2010), or from early deep sea drilling expeditions, such as in the Indian

199   Ocean (Slotnick et al., 2015b), are discontinuous, either because of hiatuses or core gaps. Much of

200   the detailed work and current understanding of stable carbon isotope stratigraphy and calcareous

201   nannofossil variations across the broad early Paleogene, therefore, has come from deep-sea drilling

202   sites with multiple holes but slow sedimentation rates, although we note the work in Clarence

203   Valley, New Zealand (**Figure 2**), another area that contains several paleo-slope sections with

204   moderately high sedimentation rates (Nicolo et al., 2007; Slotnick et al., 2012, 2015b; Dallanave et

205   al., 2015). For the Cicogna section, detailed stable isotope and $CaCO_3$ records are currently lacking,

206   as well as detailed calcareous nannofossil assemblage information, which we present here.

207

208   **3   MATERIAL AND METHODS**

209

210   **3.1   Samples**

211

212   A total of 492 samples were chiseled from outcrops along the section. Samples were selected so as

213   to be as fresh and unaltered as possible. This included chipping off weathered surfaces while in the

214   field. Each sample was calibrated to height (**Figure 4**). Samples then were split, with one portion

215   powdered in an agate ball mill, and subsequently freeze-dried.

216

217   **3.2   Geochemistry**

218

219   Each powdered samples was analyzed for bulk sediment stable isotope composition at the Stable

220   Isotope Laboratory, University of Southampton, UK. A known mass (~80 µg) was placed into a

221   headspace vial, dried overnight, and flushed with helium. 10 mL of 100 % phosphoric acid was added

222   to each sample and allowed to react. The liberated CO_2 gas was measured using an EUROPA

223   Scientific GEO 20-20 mass spectrometer fitted with a microCAPS for carbonate analysis. Results are

224   reported in standard delta notation relative to Vienna Pee Dee Belemnite (VPDB). An in-house

225   standard of Carrara Marble, calibrated to NBS-19 Limestone, was measured multiple times to

226   evaluate accuracy and precision. The external analytical precision (1σ), based on these replicate

227   analyses, was 0.028 ‰ for $\delta^{13}C$ and 0.057 ‰ for $\delta^{18}O$.

228   The amount of $CaCO_3$ in each sample was calculated from the beam height response during

229   isotope mass spectrometer measurements (Spofforth et al., 2010). The liberated CO_2 gas, when

230   squeezed up in the bellows, is measured and generates a current, the beam height. The pressure of

231   CO_2 gas is directly proportional to the beam height and therefore the mass of carbonate in the

232   sample. Over 100 samples of pure $CaCO_3$, with masses between 200 and 480 µg, were analyzed to

233   establish a linear relationship between beam height and carbonate content ($CaCO_3$ = mx + b; R^2 =

234   0.94 - 0.99). Results were validated by analyzing 30 samples on a C-H-N-O elemental analyzer.

235

3.3 Calcareous nannofossils

The un-powdered sample split was examined for calcareous nannofossils. Raw sediments were processed to prepare standard smear slides (Bown and Young, 1998). To assess the reproducibility of our counting methods in every single sample, a pivotal sample was prepared 10 times by two different operators. Repeated counts of the identical sample performed by different analysts gave similar results (sd <2-5 %). Particle density estimates (Baccelle and Bosellini, 1965) were not carried out because samples have a high range in the terrigenous content (22 to 90 %). An increase or decrease of the silicoclastic component is mainly related to the major or minor efficiency of the chemical and mechanical weathering on land (Agnini et al., 2009). In the studied sediments, the variation in the amount of the terrigenous content through time has modified the density of the allochemic particle component. Consequently, calcareous nannofossil absolute abundances could not be estimated correctly using a homogeneous/constant particle density or by weighing the same amount of sediment for each smear slide. However, the scope of semi-quantitative counts performed in this study is to recognize the precise position of biostratigraphic biohorizons rather than use these data as a proxy of the paleoproductivity of taxa. Essentially, the identification of the appearance or disappearance of any given taxon is not affected by its stratigraphic abundance pattern, which obviously reduces the negative effect of the variable abundance of the silicoclastic component throughout the section. Samples were examined under a Zeiss light microscope at 1250× magnification. Calcareous nannofossils were determined using taxonomy proposed by Aubry (1984, 1988, 1989, 1990, 1999), Perch-Nielsen (1985) and Bown (2005).

A total of 200 samples were examined, providing an average time resolution of ca. 25 kyr. A preliminary qualitative estimate of the abundance and preservation state of calcareous nannofossil assemblages was performed for all samples. An initial large batch (185) was analyzed primarily to

11

260   provide biostratigraphic control for the Cicogna section, and the basic resultshave been presented

261   by Dallanave et al. (2009). We re-checked and refined the positions of some biohorizons by

262   examining 15 additional samples, primarily across some of the CIEs, such as B1/B2, PETM, H1 and

263   H2, and K/X (Cramer et al., 2003). The calcareous nannofossil biostratigraphic schemes used by

264   Dallanave et al. (2009) were those of Martini (1971) and Okada and Bukry (1980). The new zonal

265   scheme of Agnini et al. (2014) is also used here. Biohorizon nomenclature follows that given by

266   Agnini et al. (2014): Base (B), Base common (Bc), Top (T) and Top common (Tc).

267   Calcareous nannofossil biostratigraphic results are based on semi-quantitative analyses, which

268   is based on counts of the number of specimens of selected taxa present in a prefixed area, 1 mm^2

269   or 3 long traverses (modified after Backman and Shackleton, 1983). Calcareous nannofossil

270   paleoecological results are instead based on relative abundances of calcareous nannofossil taxa

271   (percent of the total assemblage), calculated from counts of at least 300 specimens.

272   To capture changes in calcareous nannofossil assemblages we also use a statistical approach.

273   Principal Component Analysis (PCA) was preferred to other methods, as for instance non-metric

274   dimensional scaling (MDS) procedure for which a small number of axes are chosen prior to the

275   analysis and the data are fitted to these dimensions (Hammer et al., 2001). However, non-metric

276   MDS results were performed and are available as supplementary data (Figure S2). Multivariate

277   ANalysis Of Variance (MANOVA) was carried out on our dataset to determine if significative

278   differences are present among the three groups of samples recognized with PCA analysis.

279   Principal component analysis (PCA) and MANOVA was were performed on the percentages of 15

280   subgroups using the statistical software package, PAST ver. 2.17c (Hammer et al., 2001). Such The

281   former analysis is often used for examining paleontological data (e.g., Buccianti et al., 2006; Kucera

282   and Malmgren, 1998; Watkins and Self-Trail, 1992; Thibault and Gardin, 2010; Marino et al., 2012;

283   Bordiga et al., 2015), as it can point out hypothetical variables (components) that explain much of

12

284 the variance in a multidimensional data set. The first principal component accounts for the most

285 variability in any dataset examined. Each succeeding component has the highest variance possible

286 relative to the preceding components (Hammer et al., 2001). This method also increases the

287 symmetry, homoscedasticity and linearity of the data set (Aitchison, 1986). The chosen subgroups

288 were: *Chiasmolithus, Coccolithus, Ellipsolithus, Discoaster, Ericsonia, Fasciculithus, Girgisia,*

289 *Octolithus, Prinsius, Sphenolithus, Toweius, Rhomboaster/Tribrachiatus, Zyghrablithus,* reworked

290 forms, and "others".

291 **4   RESULTS**

292

293 **4.1   Carbon isotopes**

294

295 The bulk rock $\delta^{13}C$ record for the Cicogna section can be described, in a general sense, as a long-

296 term decrease of approximately 3 ‰, punctuated by a series of negative CIEs (**Figure 4**). The most

297 prominent low in $\delta^{13}C$ coincides with the CMU.

298   Previously established polarity chron boundaries and key calcareous nannofossil biohorizons at

299 the Cicogna section (Dallanave et al., 2009) provide very good stratigraphic framework. Once placed

300 onto a common time scale, in this case WO-1 (Westerhold et al., 2008), the $\delta^{13}C$ record at Cicogna

301 is fairly similar to those generated using upper Paleocene and lower Eocene marine carbonate at

302 other locations (Cramer et al., 2003; Zachos et al., 2010; Slotnick et al., 2012). This includes, for

303 example, bulk carbonate $\delta^{13}C$ records at ODP Site 1262, and DSDP Site 577 (**Figure 5**) The relatively

304 high $\delta^{13}C$ values near the base of the Cicogna section document the late stages of the PCIM, which

305 was centered within C25r (**Figure 1**). The overall drop in $\delta^{13}C$ across the section marks the long-term

306 global decrease in $\delta^{13}C$ that lasted through Chron C24n (**Figure 1**). The record contains multiple

307 negative shifts in $\delta^{13}C$. There is, however, an intriguing difference: across the Cicogna section, the

308 long-term 3 ‰ shift in bulk carbonate $\delta^{13}C$ values is generally offset from that in bulk carbonate

309 $\delta^{13}C$ records at Sites 1262 and 577 by approximately -1 ‰.

310 The superimposed CIEs are considered to correspond to CIEs found in $\delta^{13}C$ records from

311 elsewhere, some of which represent known or inferred hyperthermal events (Cramer et al., 2003;

312 Lourens et al., 2005; Nicolo et al., 2007; Zachos et al., 2010; Slotnick et al., 2012). There are three

313 pairs of CIEs below the CMU (**Figure 4**), and during the initial upper Paleocene long-term decline in

314 $\delta^{13}C$. These correspond with the B1/B2, C1/C2 and D1/D2 CIEs documented by others (Cramer et

315 al., 2003; Zachos et al., 2010). Notably, at Site 1262, the B1/B2 CIEs occur during middle C25n, and

316 the C1/C2 CIEs occur at the start of C24r (**Figure 5**). The same is true at Cicogna. Interestingly, at

317 Cicogna, the B2 and C2 CIEs show greater magnitudes than the B1 and C1 CIEs, and these paired

318 excursions are more pronounced than at all other locations examined to date. An additional paired

319 CIE occurs in the uppermost Paleocene (**Figure 4**). This may correlate to a fourth set of late

320 Paleocene CIEs documented at Site 1262 (Zachos et al., 2010).

321 The lower Eocene portion of the $\delta^{13}C$ record at Cicogna (**Figure 4**) begins at the CMU, which

322 marks the PETM (Giusberti et al., 2007; Dallanave et al., 2009). As at many locations, the PETM is

323 characterized by a prominent negative CIE. The shift in $\delta^{13}C$ at Cicogna is approximately -2.5 ‰, a

324 decrease that begins abruptly at 28.7 m and returns more gradually to near pre-excursion values by

325 about 33 m. From approximately 33 to 54 m, the $\delta^{13}C$ curve shows a relatively smooth trend. At 54

326 m, a pair of CIEs begin, with the first pair having a magnitude of about 1.0 ‰. These are the H1/H2

327 events (Cramer et al., 2003), which occurred in the upper part of Chron C24r (Lourens et al, 2005;

328 Zachos et al., 2010; Dickens and Backman, 2013; Dallanave et al., 2015). Above the H1/H2 CIEs, and

329 within Chron C24n, are a series of smaller (0.4 to 0.6 ‰) CIEs. Those at approximately 60, 65 and 72

330 m, are correlated with the I1/I2, J and K/X events, respectively. In summary, the $\delta^{13}C$ record at

331 Cicogna correlates with that at ODP Site 1262 (Zachos et al., 2010) and DSDP Site 577 (Dickens and

14

332 Backman, 2013) (**Figure 5**), as well as at several other locations (Cramer et al., 2003; Slotnick et al.,

333 2012; 2015b). This is important because it enables comparison and discussion between widely

334 separated sedimentary records within a firm temporal framework.

335

336 **4.2   Oxygen isotopes**

337

338 The $\delta^{18}O$ values range from -1.08 to -3.64 ‰ with a mean value of -1.96 ‰ and a standard deviation

339 (1σ) of 0.50 ‰ (**Figure 4**). However, at the broad scale, $\delta^{18}O$ increases upsection, with Paleocene

340 samples averaging -2.10 ‰ and Eocene samples averaging -1.89 ‰. This trend is noteworthy

341 because $\delta^{18}O$ values should decrease upsection, if the composition of the $CaCO_3$ was principally

342 reflecting rising global temperatures through the early Eocene. The 1σ of $\delta^{18}O$ values also increases

343 upsection, being 0.33 ‰ across Paleocene samples and 0.56 ‰ across Eocene samples.

344   There is virtually no correlation (r^2 = 0.014; r=0.12) between $\delta^{18}O$ and $\delta^{13}C$ values across all

345 samples (**Figure 6**). However, most "short-term" CIEs do display deceases in $\delta^{18}O$ (**Figure 4**). An

346 interval of anomalously low $\delta^{18}O$ values occurs from 39.9 m to 40.9 m, where the spatic calcite was

347 observed.

348

349 **4.3   Carbonate content**

350

351 The $CaCO_3$ content varies between 9.4 and 77.7 % across the sample suite, with a mean value of

352 54.3 % and a 1σ of 8.2 % (**Figure 4**). Two important findings emerge from the $CaCO_3$ content record.

353 First, from 39 m to 54 m, where we find limited variance in the $\delta^{13}C$ curve, $CaCO_3$ content averages

354 52.1 % with a 1σ of 4.9 %. Thus, while the average is similar to that calculated for the entire section,

355 the   standard   deviation   is   much   less.   At   Site   1262,   the   corresponding   time   interval   is   also

15

356 characterized by limited variance in $\delta^{13}C$ values and carbonate contents, the latter inferred from the

357 abundance of Fe counts in XRF scans (Zachos et al., 2010). Second, across all samples, the $CaCO_3$

358 content co-varies somewhat (r=0.29) with $\delta^{13}C$ (**Figure 6**). This is because several lows in $CaCO_3$

359 content coincide with minima in $\delta^{13}C$, as is obvious for the B1/B2, PETM and H1/H2 events (**Figure**

360 **4**).

361

362 **4.4   Calcareous nannofossils**

363

364 Calcareous nannofossils are generally abundant, diverse, and moderately well preserved. The sole

365 exception is across a 10 cm interval from 28.75 to 28.85 m, which corresponds to the onset of the

366 CIE that marks the PETM. The three samples from this interval are virtually barren of calcareous

367 nannofossils.

368   Secondary overgrowth of calcite can partially or wholly blur species-specific morphological

369 features. Such diagenetic alteration, however, only marginally influences the relative, semi-

370 quantitative andabsolute abundance of calcareous nannofossil taxa (Toffanin et al., 2013). Calcite

371 dissolution, on the other hand, can significantly affect the relative abundances of various calcareous

372 nannofossils within a given volume of sediment. This is because the removal of more dissolution

373 susceptible taxa, such as *Toweius* and holococcoliths, necessarily increases the abundance of less

374 dissolution susceptible taxa, such as discoasters (Roth and Thierstein, 1972; Adelseck et al., 1973;

375 Roth, 1983; Bornemann and Mutterlose, 2008; Toffanin et al., 2013). In general, moderate to strong

376 calcite dissolution also decreases the total abundance of calcareous nannofossils within a given

377 volume of sediment (Adelseck et al., 1973; Toffanin et al., 2011). In the Cicogna section, calcite

378 overgrowth on discoasters is the prevalent process affecting calcareous nannofossil assemblages

379 (**Plate I**). Most assemblages display high abundances (>1000 specimens/mm^2) and a high diversity,

380   which include more fragile taxa. It follows that dissolution has not severely altered most

381   assemblages in samples from the Cicogna section. Rather, the calcareous nannofossil record is

382   considered to represent a genuine paleoecological signal.

383   Nannofossil assemblages from the Cicogna section display several general trends (**Figures 7-9**).

384   At the most basic level, there is a decrease in the total number of nannofossils (N/mm^2) with

385   decreasing age. Paleocene samples average approximately 2600 specimens/mm^2, whereas Eocene

386   samples above the H1/H2 events average approximately 1200 specimens/mm^2. This decrease in

387   abundance broadly corresponds to a change in calcareous nannofossil composition, as supported

388   through a series of additional observations at the Cicogna section (**Figures 7-9**):

389   - *Coccolithus* and *Toweius* constitute nearly half of the assemblages considering the entire

390   section. However, these genera show a clear decrease in abundance upsection, with a mean

391   value of 60 % in Paleocene samples and 35 % in Eocene samples;

392   - *Zyghrablithus bijugatus* shows a low mean value of approximately 4 % in the Paleocene,

393   followed by a sharp increase in the basal part of the Eocene, and a mean value of

394   approximately 25 % upsection in the Eocene. Hence, the abundance of this taxon expands

395   on behalf of *Coccolithus* and *Toweius*;

396   - *Sphenolithus* decreases progressively during the Paleocene, suddenly disappears at the

397   onset of the PETM, before returning to and exceeding pre-PETM values in the lower Eocene.

398   Thus, the abundance of sphenoliths also expands on behalf of *Coccolithus* and *Toweius*;

399   - *Fasciculithus* shows a severe decline in abundance and species diversity at the onset of the

400   PETM (28.70 m), leading up to their extinction at 34.73 m;

401   - *Octolithus* is rare throughout most of the studied section, but displays high abundances from

402   approximately 14.7 m to 27.5 m in the upper Paleocene;

17

403   • *Discoaster* does not show any distinct change in abundance except for a single peak at the

404   onset of the PETM;

405   • Several Cretaceous and early Paleocene species constitute minor reworked components

406   throughout the section. Notably, the intervals marked by the PETM, H1/H2 and, to a lesser

407   extent, B1/B2 CIEs are characterized by higher abundances of these reworked components;

408   • Representatives of placolith genera, such as *Prinsius, Ericsonia, Chiasmolithus* and *Girgisia*,

409   are minor components of most samples. *Prinsius* displays a marked permanent decrease in

410   abundance from a mean value of approximately 6 % to 2.5 % across the Paleocene/Eocene

411   boundary. By contrast, *Ericsonia* does not show a prominent difference in abundance

412   between Paleocene and Eocene assemblages, but increases in abundance during known and

413   suspected hyperthermal events;

414   • The Calcareous Nannofossil Excursion Taxa (CNET), which include *Discoaster araneus* and

415   the genus *Rhomboaster* are present during the CIE of the PETM. The evolution of the

416   *Rhomboaster/Tribrachiatus* plexus started at the onset of the PETM, when *Rhomboaster* and

417   *T. bramlettei* first appeared, and continued into the lower Eocene with the successive

418   appearances of *T. contortus* and *T. orthostylus* (Raffi et al., 2005; Agnini et al., 2006; 2007b).

419   Beyond the above variations, evolutionary appearances and extinctions occur during the

420   studied time interval (**Figures 7-9**). Most of these species belong to *Discoaster, Sphenolithus* and the

421   *Rhomboaster/Tribrachiatus* lineage, and include *D. multiradiatus, D. diastypus, D. lodoensis, S.*

422   *radians, S. anarrhopus, T. bramlettei, T. contortus* and *T. orthostylus*. The biohorizons defined using

423   these species are exceptionally useful for biostratigraphy and, interestingly, often occur close to

424   changes in $\delta^{13}C$.

425   All assemblage data were used for PCA analysis. This indicates that PC1 (41.3 %) and PC2 (14.7

426   %) together account for 56 % of the variance in the dataset. The PCA graph (**Figure 10A8; Figure S1**

18

427 **supporting material**) shows that samples can be ~~easily~~ subdivided into three subgroups. The first

428 two populations of samples are distinguished because of their different positions along the x-axis

429 (PC1). The third population <u>is much more dispersed but a possible discrimination</u> ~~can be separated~~

430 from the other two <u>seems to be</u> ~~because~~<u>hypothesized because</u> of its different position along the y-

431 axis (PC2). The use of a different statistical procedure, as for instance MDS, does not substantially

432 change these results (**Figure S2**). <u>To further support the subdivision of the study samples in three</u>

433 <u>subgroup, we applied the MANOVA analysis to our data set (Figure 10B). The result clearly</u>

434 <u>confirmed that that Paleocene, PETM and Eocene samples are in fact isolated one from each other.</u>

435

436 **5   DISCUSSION**

437

438 **5.1   Integrated stratigraphy and a carbon isotope template**

439

440 Polarity chron boundaries and calcareous nannofossil biohorizons (**Table 1; Figure 4**) provide a solid

441 stratigraphic framework for the Cicogna section. Calcareous nannofossil biohorizons, including

442 additional ones defined here, align in same stratigraphic order when compared to other locations,

443 such as ODP Site 1262 and DSDP Site 577 (**Table 1; Figure 11**). The Cicogna section represents

444 sediment accumulation between 57.5 and 52.2 Ma on the WO-1 time scale (Dallanave et al., 2009).

445 The average SR was ca.15.2 m/Myr, although this must have varied (**Figures 3, 11**). The CMU, which

446 marks the "core" of the PETM and ca. 80-100 kyr, showing a higher sedimentation rate than much

447 of the remaining record (Dallanave et al., 2009; Krishnan et al., 2015).

448   Once placed into the above stratigraphic framework, the bulk carbonate δ^{13}C profile

449 documented at Cicogna correlates well to that generated at ODP Site 1262 (**Figure 5**). In fact, it is

450 similar to δ^{13}C profiles generated at multiple locations (**Figure 2, Figure S4**), as long as records have

19

451 been properly calibrated in both the depth and time domains. This includes accounting for core

452 stretching and core gaps at scientific drilling sites, such as at DSDP Site 577 (Dickens and Backman,

453 2013), and accounting for changing strike and dip along land sections, such as done at Cicogna

454 (**Figure 3**). During late Paleocene and early Eocene times, the Cicogna section records the long-term

455 decrease in $\delta^{13}C$. Superimposed on this drop were multiple, often paired, negative CIEs. The PETM

456 definitively represents the most prominent CIE, but several other CIEs occurred before and after.

457 Importantly, the relative positions of polarity chron boundaries, key calcareous nannofossil

458 biohorizons and CIEs at Cicogna align well with those found at other locations (**Table 1**; **Figures 5,**

459 **11**).

460   A clearly recognizable $\delta^{13}C$ pattern spans the late Paleocene through the early Eocene at several

461 locations (Cramer et al., 2003; Nicolo et al., 2007; Galeotti et al., 2010; Zachos et al., 2010; Slotnick

462 et al., 2012, 2015b), although the total number of CIEs remains uncertain. At Cicogna, the problem

463 lies in the interval surrounding the K/X event, which broadly corresponds to the start of the EECO

464 (see discussion in Slotnick et al., 2012). We cannot confirm with our sample resolution whether a

465 series of short-term, small amplitude CIEs mark this time, an idea suggested from $\delta^{13}C$ records of

466 the Clarence Valley sections (Slotnick et al., 2012; 2015b). However, as at other locations, such as

467 Site 1262, no significant CIEs occurred within the 1.6 Myrs between the PETM and the H-1/ETM-2

468 event (**Figure 5**).

469   The time-correlative $\delta^{13}C$ template implies changes in the mean ocean $\delta^{13}C$ of dissolved

470 inorganic carbon (DIC). In turn, these compositional changes very likely represent variations in fluxes

471 of highly ^{13}C-depleted carbon to and from the ocean or atmosphere, such as changes in the release

472 and storage of organic carbon (Shackleton, 1986; Dickens et al., 1997; Kurtz et al., 2003; Deconto et

473 al., 2010; Komar et al., 2013). The $\delta^{13}C$ record at Cicogna offers no direct insight on the location of

474 this carbon (e.g., seafloor methane, permafrost, peat). However, it does support an important

475 concept: the magnitudes of given CIEs appear somewhat related to one another and to the long-

476 term $\delta^{13}C$ record. In particular, the PETM occurred about halfway between the long-term high and

477 low in $\delta^{13}C$, and heralded a relatively long time interval lacking CIEs. A generic explanation is that a

478 very large mass of ^{13}C-depleted carbon was injected from some organic reservoir into the ocean or

479 atmosphere during the PETM, and that the reservoir needed to recharge for considerable time

480 before the next injection (H-1/ETM-2) could occur (Dickens et al., 2003; Kurtz et al., 2003; Lunt et

481 al., 2011; Komar et al., 2013).

482   The overall -1 ‰ offset of the $\delta^{13}C$ curve between the records at Cicogna and at Sites 577 and

483 1262 (**Figure 5**) warrants brief discussion. It probably does not reflect wholesale diagenesis and

484 resetting of the primary signal at any of these sections. Otherwise, a recognizable correlative $\delta^{13}C$

485 record and well-preserved nannofossils (**Plate I**) would not be found at all three locations. In fact, it

486 is difficult to modify the original $\delta^{13}C$ composition of carbonate over appreciable distance (> than

487 several meters) in marine sedimentary sequences dominated by fine grained calcite, even those

488 now exposed on land as lithified rock, such as at Cicogna or in the Clarence Valley. This is because

489 the carbon water/rock ratio remains low, because almost all carbon exists in carbonate, and because

490 temperature minimally influences carbon isotope fractionation (Matter et al., 1977; Scholle and

491 Arthur, 1980; Frank et al., 1999). Instead, the offset in the $\delta^{13}C$ curves probably relates to differences

492 in the composition of the original carbonate, a concept that we return to later.

493   However, local dissolution and re-precipitation of carbonate definitely has occurred in the

494 Cicogna section. This can be observed in the overgrowths of secondary calcite on discoasters and

495 *Rhomboaster/Tribrachiatus* (**Plate I**). This process should dampen the original CIEs, because on the

496 meter-scale, dissolution and re-precipitation of carbonate would involve $\delta^{13}C$ gradients in the DIC

497 of surrounding pore water (Matter et al., 1977; Scholle and Arthur, 1980). This may explain, in part,

21

498 why the magnitude of early PaleogeneCIEs in bulk carbonate records are often muted relative to

499 those found in other carbon-bearing phases (Slotnick et al., 2015b).

500

501 **5.2   Oxygen isotopes and a problem recording past temperatures**

502

503 The $\delta^{18}O$ record at Cicogna is intriguing because many of the CIEs are characterized by negative

504 excursions but absolute values of $\delta^{18}O$ generally and unexpectedly increase upsection (**Figure 4**).

505 Similar results have been documented in bulk carbonate stable isotope records at other locations,

506 such as ODP Site 1215 (Leon-Rodriguez and Dickens, 2010) and Mead Stream (Slotnick et al., 2012).

507 Even the $\delta^{18}O$ record of bulk carbonate at Site 1262 shows minimal long-term change from the late

508 Paleocene to the early Eocene (Zachos et al., 2010), the time when high-latitude surface

509 temperatures and deep ocean temperatures presumably increased by 5-6 °C, and one might expect

510 a >1 ‰ decrease in the $\delta^{18}O$ of marine carbonate.

511   Like previous workers, we cannot discount the notion that temperatures at low and high

512 latitudes responded differently across the early Paleogene (Pearson et al., 2007; Huber and

513 Caballero, 2011). Unlike for carbon isotopes, however, local dissolution and re-precipitation of

514 carbonate should significantly impact the $\delta^{18}O$ of marine carbonate. This is because the oxygen

515 water/rock ratio would be high before lithification, and because temperature strongly influences

516 oxygen isotope fractionation (Matter et al., 1975; Scholle and Arthur, 1980; Frank et al., 1999). In

517 general, as calcite-rich sediments and surrounding pore water are buried to higher temperatures

518 along a geothermal gradient, local dissolution and re-precipitation of carbonate shifts carbonate

519 $\delta^{18}O$ to lower values (above references; Schrag et al., 1995). It is likely that, during sediment burial,

520 the bulk carbonate $\delta^{18}O$ records in many lower Paleogene sections, including at Cicogna, have been

521 modified. We suggest that a signal of surface ocean temperature changes remains in the Cicogna

522 section, which gives rise to short-term $\delta^{18}O$ excursions that coincide with CIEs and several known or

523 suspected hyperthermal events. However, the entire $\delta^{18}O$ record at this location likely has shifted

524 to more negative values preferentially with increasing burial depth and age. This partly explains the

525 observed relationship between bulk carbonate $\delta^{13}C$ and $\delta^{18}O$, which lies along a trajectory expected

526 for diagenesis (**Figure 6**). A potential test of this idea would be to show that the overgrowths on

527 nannofossils (**Plate I**) have a significantly lower $\delta^{18}O$ than the primary core carbonate of nannofossil

528 tests.

529

530 **5.3 Calcareous nannofossil assemblages within the context of correlative stable isotope records**

531

532 A detailed stable carbon isotope curve provides a powerful tool to place past changes in calcareous

533 nannofossil assemblages into a highly resolved framework. This is because, as implied above, truly

534 global changes in the $\delta^{13}C$ composition of the ocean should occur within the cycling time of carbon

535 through ocean, which is <2000 years at present-day and presumably for the entire Cenozoic

536 (Broecker and Peng, 1982; Shackleton, 1986; Dickens et al., 1997).

537   Across the study interval at Cicogna, several calcareous nannofossil taxa appear or disappear

538 (**Table 1**). Moreover, their abundances also change between these horizons (**Figures 7-9**). One might

539 hypothesize that these changes in nannofossil assemblages were related to the established (e.g.,

540 the PETM, H1/ETM-2 and K/X) and potential (e.g., the B1/B2, I1/I2) hyperthermal events that

541 occurred during the late Paleocene and early Eocene (**Figures 1, 5**). However, the timing between

542 recorded evolutionary appearances and extinctions of calcareous nannofossils and perturbations in

543 $\delta^{13}C$ are variable (**Figures 7-9**). For instance, several significant calcareous nannofossil changes

544 observed close to H1/H2 hyperthermals (e.g., B *T. othostylus*, B *S. radians*, B *S. villae*, Tc *D.*

545 *multiradiatus,* T *T. contortus*) predate these events. By contrast, several biotic changes observed

23

546 close to the B1/B2 CIEs (e.g., B *D. delicatus,* Tc *S. anarrhopus,* B *D. multiradiatus*, T *Ericsonia robusta*)

547 happened at the end of these events. The PETM seems to provide the only case when a negative

548 CIE precisely corresponds to major changes in calcareous nannofossil assemblages.

549 Profound changes in calcareous nannofossil assemblages occurred across the PETM in several

550 locations (**Figure 2**), both in terms of relative abundances and increases in origination and extinction

551 rates (Aubry, 1998; Bown et al., 2004; Raffi et al., 2005; Gibbs et al., 2006a; Agnini et al., 2007a; Self-

552 Trail et al., 2012). At Cicogna, the assemblages show remarkable, though mostly transient, relative

553 abundance variations across the PETM, including an increase in *Coccolithus*, a decrease in

554 *Zygrhablithus, Sphenolithus, Toweius* and *Prinsius*, and an extinction of most fasciculith species

555 (**Figure 8**). Not surprisingly, these changes are very similar to those in the Forada section, which is

556 also located in the Belluno Basin (Agnini et al., 2007a).

557 Although these changes in relative abundance of taxa alone represent a notable difference with

558 respect to background conditions, most of the changes are transient and/or local when compared

559 with other datasets (Bralower, 2002; Gibbs et al., 2006b; Agnini et al., 2007b; Angori et al., 2007;

560 Mutterlose et al., 2007). For instance, an increase in abundance of *Discoaster* and *Fasciculithus* was

561 reported for some of the PETM section studied (e.g., Bralower, 2002; Tremolada and Bralower,

562 2004; Raffi et al. 2009), but these assemblage variations were not observed in other sections (e.g.,

563 Gibbs et al. 2006; Agnini et al., 2007a; Self-Trail et al., 2012). The only global calcareous nannofossil

564 assemblage features of the PETM are represented by the evolutionary appearance of

565 *Rhomboaster/Tribrachiatus* lineage, the presence during the CIE of short-lived species such as

566 *Discoaster areneus*, and the disappearance of several species of fasciculiths (Raffi et al., 2005; Agnini

567 et al., 2007a).

568 While changes in calcareous nannoplankton assemblages during the PETM have been

569 investigated at high resolution at different locations (e.g., Bralower, 2002, Gibbs et al., 2006b; Agnini

570 et al. 2007a), the longer-term perspective in which such changes occurred during the early

571 Paleogene has remained uncertain (Gibbs et al., 2012). The record at Cicogna provides this

572 opportunity.

573 The PCA analysis of calcareous nannofossil census data (%) indicates that two principal

574 components (PC1 and PC2) account for most (56.0 %) of the variability in our 15 selected subgroups.

575 Such analysis also permits the studied samples to be subdivided into ~~three~~ two populations and a

576 possible widely dispersed group (**Figure 10A**). The first two populations are distinguished because

577 of a major difference along the -x-axis representing PC1, whereas the third population seems to

578 stand~~s~~ out because of a ~~significant~~ difference along the y-axis representing PC2. Importantly, each

579 of these three populations constitutes a homogeneous group in the time domain: Group 1 includes

580 all upper Paleocene samples (Paleocene samples and B1/B2 events); Group 2 consists of almost all

581 lower Eocene samples (Eocene samples, H1/H2 events and K event); Group 3 comprises samples

582 that span the PETM (both core and recovery), and two samples that come from sediment deposited

583 during the core of the H1 and B2 events (**Figure 10**). These results indicate that late Paleocene

584 calcareous nannofossil assemblages are statistically different in their composition from those of

585 early Eocene samples. ~~Moreover, the~~To check if calcareous nannofossil assemblages across the

586 PETM~~, and the climax of the B2 and H1 events,~~ are -statistically different from those of either the

587 late Paleocene or the early Eocene, we performed a MANOVA analysis, which pointed out that

588 ellipses containing 95% of the data points for each group (late Paleocene, early Eocene and PETM)

589 are virtually not overlapping one to each other suggesting that three statistically different

590 populations are recognized across the PETM, the late Paleocene and the early Eocene background

591 assemblages, and the PETM fossil associations.

592

593   The general shift in the relative abundance of placoliths (i.e., *Coccolithus, Toweius* and *Prinsius*),

594   the major component of the late Paleocene assemblages, to nannoliths/holococcoliths (i.e.,

595   *Sphenolithus* and *Zygrhablithus*), the major component of the early Eocene assemblages, largely

596   explains the PC1 component or Axis 1 (Figure 10). By contrast, the dramatic shift toward negative

597   values in the PC2 component or Axis 2 during the PETM happens because of the increase of *Ericsonia*

598   and reworking and the presence of *Rhomboaster-Tribrachiatus* plexus. Presumably, this relates to

599   peculiar paleoenviromental conditions that developed during the event. One can hypothesize that

600   this may have been a major difference in the physicochemical parameters of sea surface waters,

601   such as higher temperatures, higher nutrient concentration or reduced carbonate saturation state.

602   Statistical analysis of our data from Cicogna does not highlight any prominent short-term

603   changes in calcareous nannofossil assemblages, other than across the PETM and perhaps the B2 and

604   H1 events. However, several biohorizons occur around the B1/B2 events. Specifically, these are the

605   Bc *Z. bijugatus*, the brief high abundance of *Octolithus* spp., the evolutionary onset of the *D.*

606   *delicatus/D. multiradiatus* lineage, the presence of the short-ranged *E. robusta*, the final radiation

607   of late Paleocene fasciculiths (i.e., *F. richardii* group, *F. hayi, F. lilianae, F. alanii*), and the Tc of *S.*

608   *anarrhopus*. All these happened at Cicogna and at Site 1262 within Chron C25n (Agnini et al., 2007b;

609   Dallanave et al., 2009; **Figure 11)**, which spanned only 0.54 Myr (Westerhold et al., 2008). These

610   near-synchronous events are intriguing because while the various nannofossils represent only minor

611   components of late Paleogene assemblages, they were destined to become either an abundant

612   constituent of Eocene populations (e.g., *Z. bijugatus* and the *D. delicatus/D. multiradiatus* lineage),

613   or extinct after having been a distinctive element of Paleocene assemblages (e.g. *Fasciculithus* spp.

614   and *S. anarrhopus).* Following the PCIM, the long-term increase in temperature and decrease in δ^{13}C

615   (**Figure 1**) coincided with a series of minor changes in nannofossil assemblages, which subsequently

616   became important, presumably for evolutionary reasons.

26

617   Similar to the late Paleocene, calcareous nannofossil assemblages after the PETM do not show

618   major rearrangements of common taxa during the early Eocene. Instead, minor components of

619   these assemblages exhibit a sequence of closely spaced biohorizons. The sequence of these

620   biohorizons is: T *Fasciculithus,* B *D. diastypus*, B *T. contortus,* T *T. bramlettei*, Tc *D. multiradiatus,* T

621   *T. contortus*, B *T. orthostylus,* B *S. radians,* T *D. multiradiatus,* B *D. lodoensis,* B *G. gammation* and

622   Bc *D. lodoensis* (**Table 1**). Within the resolution of available paleomagnetic and δ^{13}C data, all these

623   biohorizons are virtually synchronous between the Cicogna section and ODP Site 1262 (**Figure 11**).

624   They also almost all occurred in near synchrony at Site 577 (Dickens and Backman, 2013), although

625   the precise correlation remains uncertain, given problems with coring disturbance and subtleties in

626   age models at this location.

627   Importantly, for stratigraphic purposes, the B and Bc of *D. lodoensis* are approximately coeval

628   at all three locations and spaced apart by about 750 kyr. Unless one examines samples in detail,

629   these two biohorizons can be confused and result in an erroneous assignment of early Eocene ages.

630   The evolutionary appearances and extinctions amongst early Eocene nannofossil assemblages

631   may suggest the presence of uneven communities living in an extreme climate in which alterations

632   of environmental conditions, even minor, might trigger evolutionary changes or prominent

633   variations in abundances of a limited number of taxa that typically do not represent the dominant

634   component of assemblages. explained possible explanation is a generally higher tolerance of

635   cosmopolitan taxa to variations in environmental conditions (Boucot, 1975; Winter et al., 1994). In

636   contrast, highly specialized taxa that are adapted to a particular ecological niche, may display

637   greater sensitivity to modifications in the photic zone environment (MacArthur and Wilson, 1967;

638   Pianka, 1970; Baumann et al., 2005).

639   In summary, several genera of calcareous nannofossils, such as *Rhomboaster, Tribrachiatus,*

640   *Sphenolithus, Discoaster* and *Zygrhablithus* were, at least to some extent, affected during the late

641 Paleocene-early Eocene transition, because they show an increased rate of taxonomic evolution

642 (**Figure 11**). However these genera are all minor groups in terms of overall abundance, at least in

643 most lower Paleogene sediment sequences, and they all belong to nannoliths and holococcoliths. It

644 appears that these organisms were more sensitive to environmental changes than heterococcoliths,

645 for example the cosmopolitan genera *Coccolithus* and *Toweius*.

646

647 **5.4   Early Paleogene calcareous nannofossil evolution**

648

649 Any comprehensive paleoenvironmental interpretation involving early Paleogene calcareous

650 nannofossils remains tentative because many taxa, such as *Rhomboaster/Tribrachiatus*, *Discoaster*,

651 *Sphenolithus* and *Zygrhablithus*, are extinct. Still, some single species or species groups are

652 considered to be useful for reconstructions of paleoenvironmental conditions (Geisen et al., 2004).

653 With that viewpoint, and with an understanding of modern holococcolith/nannolith ecology and

654 classical biogeographical model, we provide a scenario regarding late Paleocene-early Eocene

655 calcareous nannofossil evolution.

656   Modern holococcolithophores have numerous tiny rhombohedral calcite crystallites, and are

657 considered as haploid stages of certain heterococcolithophores, which can live in just about any

658 marine photic zone environment, although higher abundances and diversity are typical in

659 oligotrophic settings (Billard and Inouye, 2004). The most common Paleogene holococcolith was

660 *Zygrhablithus bijugatus.* This taxon has been interpreted as a *K*-specialist more adapted to stable

661 environments and oligotrophic conditions (Aubry, 1998; Bralower, 2002; Tremolada and Bralower,

662 2004; Agnini et al., 2007a; Self-Trail et al., 2012). Nannolith is a term used to describe peculiar

663 morphotypes usually observed in association with coccoliths, but lacking the typical features of

664 heterococcoliths or holococcoliths. *Ceratolithus cristatus,* a modern nannolith, has been observed

28

665 on the same cell together with *Neosphaera coccolithomorpha* (Alcolber and Jordan, 1997),

666 suggesting that the nannolith stage (*C. cristatus*) corresponds to the holococcolith stage in other

667 taxa (Young et al., 2005). Paleogene nannoliths include taxa with peculiar morphologies such as

668 *Discoaster, Fasciculithus* and *Sphenolithus*. These genera often have been associated with warm

669 waters and oligotrophic environments and are almost unanimously interpreted as *K*-specialists (Haq

670 and Lohmann, 1976; Backman, 1986; Wei and Wise, 1990; Bralower, 2002; Gibbs et al., 2004; 2006a;

671 2006b; Agnini et al., 2007a). *K* specialists fluctuate at or near the carrying capacity (K) of the

672 environment in which they thrive (MacArthur and Wilson, 1967), and are usually characterized by

673 long individual life-cycles and low reproductive potential. The *K*-specialist strategy is advantageous

674 in highly stable, typically oligotrophic environments, which allows the evolution of stenotopy and

675 where organisms compete by specialization and habitat partitioning (Hallock, 1987; Premoli Silva

676 and Sliter, 1999). The narrow range of adaptability to changes in habitat or ecological conditions

677 stimulates a rapid speciation.

678   At present, it is commonly accepted that modern holococcoliths and nannoliths are not

679 produced by autonomous organisms; rather, they are stages in the life cycle of coccolithophores.,

680 Moreover, the passage between the two stages may be triggered by environmental factors (Billard

681 and Inouye, 2004).

682   Hence, though Paleogene holococcoliths/nannoliths have no direct descendants in present-day

683 oceans, they may very well have shared similar physiological features and life cycles with modern

684 taxa. Assuming this is the case, the increase in the relative abundance of holococcoliths and

685 nannoliths at the expense of heterococcoliths as well as the higher rates of evolution shown by

686 holococcoliths and nannoliths may suggest conditions in which highly specialized taxa could flourish

687 and rapidly evolve. This scenario is consistent with the idea, based on laboratory and modern ocean

688 data, that the calcareous nannoplankton response to environmental change is species or group

689 specific rather than homogeneous across the entire assemblage (Riebesell et al., 2000; Langer et al.,

690 2006; Iglesias-Rodriguez et al., 2008; Lohbeck et al., 2012). Variations in the thermal and chemical

691 structure of photic zone waters may thus account for the observed changes in the early Paleogene

692 calcareous nannofossil assemblages.

693

694 **5.5   Carbon isotope of surface waters during the early Paleogene**

695

696 Like at Cicogna, well-preserved calcareous nannofossils dominate bulk sediment carbonate contents

697 of early Paleogene strata at Sites 577 and 1262 (Backman, 1986; Zachos et al., 2004; Dickens and

698 Backman, 2013). Given that the nannofossil assemblages are fairly similar (**Figure 11**), a really basic

699 question returns: why is the overall early Paleogene bulk carbonate δ^{13}C record at Cicogna less by

700 approximately 1 ‰?

701 A variety of explanations for the δ^{13}C offset can be offered. For example, sediments at Cicogna had

702 greater amounts of organic matter, and during burial diagenesis, a fraction of this carbon was

703 consistently added so as to decrease the δ^{13}C of pore water DIC. We note, though, that C_{org} contents

704 (wt %) at the proximal Forada section generally have values less than 0.1 wt % (Giusberti et al.,

705 2007). Similar C_{org} contents are found at ODP Site 1262, where values range from 0.0 to 0.3 wt %

706 (Zachos et al., 2004).

707   A cursory examination of early Paleogene bulk carbonate δ^{13}C records from other sites of the

708 North Atlantic/western Tethys region (e.g., Sites 550 and 1051; **Figure 2**) shows a commonality:

709 these locations also display negative 0.5 to 1 ‰ offsets relative to correlative records at Sites 577

710 and 1262 (Cramer et al., 2003). The δ^{13}C of DIC in modern surface waters (<100 m) ranges by about

711 2 ‰, because of the differences in temperature, primary productivity and water mass mixing

712 (Kroopnick, 1985; Tagliabue and Bopp, 2008). Notably, however, gradients in δ^{13}C of surface water

30

713   DIC are gradual, such that large regions have fairly similar values. It is possible that bulk carbonate

714   δ^{13}C values in early Paleogene North Atlantic sections record lower values than locations near the

715   Equator or in southern latitudes because of past ocean circulation.

716

717   **6   SUMMARY AND CONCLUSIONS**

718

719   We generate records of bulk carbonate δ^{13}C and δ^{18}O, CaCO$_3$ content and calcareous nannofossil

720   assemblages from the Cicogna section, a marine sedimentary succession that now crops out along

721   a stream in the Southern Alps of northeast Italy. The combined geochemical and calcareous

722   nannofossil results allow us to generate a detailed stratigraphy for the section, as well as to explore

723   relationships between stable isotope variations and nannofossil assemblages. Most lower

724   Paleogene sections examined to date lack such coupled data sets.

725   The δ^{13}C record and calcareous nannofossil assemblages show that the section spans ~5.3 Myr

726   of the late Paleocene and early Eocene interval, from 57.5 to 52.2 Ma on the WO-1 timescale. This

727   is consistent with previous paleomagnetic information and preliminary calcareous nannofossil

728   biostratigraphy (Dallanave et al., 2009), but provides a more detailed stratigraphic framework, one

729   appropriate for correlations to other locations around the world. In particular, the fairly well

730   resolved δ^{13}C record shows long-term and short variations that correspond to those found in several

731   other sections, including an established series of negative CIEs. The most prominent CIE marks the

732   PETM, while other less pronounced CIEs represent the H-1, K/X and other "events" documented

733   elsewhere. The δ^{13}C variations observed at Cicogna clearly reflect global changes in the fluxes of

734   carbon to and from the ocean and atmosphere.

735   PCA analysis of calcareous nannofossil assemblages shows three distinct sample clusters. Late

736   Paleocene and early Eocene assemblages were distinctly different from each other and from that of

31

737   the PETM. Indeed, the PETM, the most intense hyperthermal during the late Paleocene - early

738   Eocene, was characterized by a unique calcareous nannofossil assemblage composition. This

739   suggests that the brief episode of extreme warming permanently modified the composition of

740   calcareous nannoplankton through an increase in the rate of taxonomic evolution (Gibbs et al.,

741   2006a). Less prominent hyperthermal events do not show significant variations in the main

742   components of assemblages, but rather were characterized by a series of changes affecting a limited

743   number of rare taxa. These taxa may have been less tolerant to environmental changes in their

744   habitats.

745   More common taxa, essentially consisting of placoliths, such as the cosmopolitan *Coccolithus*

746   and *Toweius*, display a progressive long-term decrease interrupted by transient changes in their

747   relative abundance but virtually no extinction or origination events occur in these groups during the

748   pertinent time interval. Species belonging to nannoliths and holococcoliths (*Discoaster,*

749   *Fasciculithus, Rhomboaster/Tribrachiatus, Sphenolithus* and *Zygrhablithus*), generally show a higher

750   rate of evolution and a higher concentration of biohorizons close to $\delta^{13}C$ perturbations. In

751   conclusion, calcareous nannoplankton show a different response of the various components of the

752   assemblages, this is consistent with a species or taxonomic unit sensitivity of calcareous

753   phytoplankton to paleoenviromental perturbations. This evolutionary climate-forced model is

754   supported by data from ODP Site 1262, which demonstrate that these changes are global and

755   synchronous between middle latitudes in the western Tethys region and the South Atlantic.

756

33

8   FIGURE AND TABLE CAPTIONS

765 **Figure 1**. Middle Paleocene to middle Eocene (64 to 48 Ma) stable isotope (δ^{13}C and δ^{18}O) records

766 of benthic foraminifera from multiple locations (Zachos et al., 2008) placed on the Option 1 (W01)

767 time scale of Westerhold et al. (2008). Also shown are positions of polarity chrons and calcareous

768 nannofossil biozones for this time interval, both from the CP Biozone scheme (Okada and Bukry,

769 1980) and the CN Biozone scheme (Agnini et al., 2014). Various "events" are noted within this

770 chronostratigraphic framework, including the Paleocene carbon isotope maximum (PCIM), the

771 Paleocene-Eocene thermal maximum (PETM), the H-1/ETM-2 event, the K/X event, and the Early

772 Eocene Climatic Optimum (EECO). To the right is the general lithologic column and

773 magnetostratigraphy of the Cicogna section (Dallanave et al., 2009).

774

775 **Figure 2.** Paleogeographic map indicating approximate locations at 55 Ma for several key sites with

776 detailed stable isotope records across the late Paleocene and early Eocene. These include (marked

777 with black dots and star) the Cicogna section (NE Italy, this study), DSDP Site 577 (Shatsky Rise,

778 Dickens and Backman, 2013), ODP Sites 1051 (Blake Nose, Ogg and Bardot, 2001), 1215 (central

779 Pacific, Raffi et al., 2005), and 1262 (Walvis Ridge, Westerhold et al., 2008), and the Clarence Valley

780 (CV) sections (New Zealand, Slotnick et al., 2015b). The grey areas represent plate fragments, while

781 the black lines show present-day shorelines. Boxes next to site locations show average

782 sedimentation rates from the base of Chron C25n to the base of Chron C23r (57.20 - 52.36 Ma). The

783 base map is from http://www.odsn.de/odsn/services/paleomap/paleomap.html. Red triangles are

784 locations where a decrease in diversity of *Fasciculithus* spp. has been documented near the PETM.

785 Locations include the Clarence Valley sections, central Pacific (ODP Sites 1215, 1220, 1221), western

786 Pacific (DSDP Site 577 and ODP Site 865), South Atlantic (Walvis Ridge, DSDP Site 527, ODP Sites

787 1262,1263-1267; Maud Rise, ODP Site 690), equatorial Atlantic (Ceara Rise, ODP Site 929, Demerara

788   Rise, ODP Sites 1259, 1260), northwestern Atlantic (New Jersey Margin land sections, ODP Site 1051,

789   IODP Site U1403, U1409), northeastern Atlantic (Bay of Biscay DSDP Sites 401, 549 and 550, Zumaya

790   land section), Indian Ocean (DSDP Site 213; ODP 672; Kerguelen Plateau, ODP Site 1135)(Backman,

791   1986; Aubry, 1999; Bralower, 2002; Dupuis et al., 2003; Tremolada and Bralower, 2004; Bralower

792   and Mutterlose, 1995; Monechi et al., 2000; Gibbs et al., 2004; Raffi et al., 2005; Agnini et al., 2007a;

793   Angori et al., 2007, Mutterlose et al., 2007; Jiang and Wise, 2009; Shamrock, 2010; Norris et al, 2014;

794   Dallanave et al., 2015).

795

796   **Figure 3**. The location and representative photographs of the Cicogna section in northeast Italy. (a)

797   Geographic map showing the main Late Cretaceous–early Paleogene paleogeographic domains of

798   the Italian Southern Alps (modified after Cati et al., 1989); (b) Geological map of the local area

799   (modified after Costa et al, 1996 indicating also the location of the Cicogna section (red asterisk); (c)

800   Alternating beds of Paleocene gray-green marls and calcareous marls (0-20 m); (d) The Scaglia Rossa

801   *sensu latu* overlain by the Belluno Flysch; (e) Marl/calcareous marl couplets in the lower Eocene

802   portion of the section (approximately 40.0-70.0 m); (f) The base of the Clay Marl Unit, which denotes

803   the onset of the PETM (approximately 28.7-29.3 m); (g) The brownish-red interval of clayey marls

804   with sporadic grey-green cm-scale spots and lenses, the CMU, overlain by prominent rhythmic

805   alternations of marls and calcareous marls (approximately 28.7-33.0 m).

806

807   **Figure 4.** The Cicogna section with records of bulk carbonate $\delta^{13}C$ and $\delta^{18}O$ data, and $CaCO_3$ content.

808   Calcareous nannofossil biostratigraphy (CP and NP biozones) and magnetostratigraphy are after

809   Dallanave et al. (2009), CN biozones are also reported. Orange and yellow bands mark major $\delta^{13}C$

810   excursions. Dashed lines indicate minor CIEs that have been labeled elsewhere (e.g., E1/E2, F and

811 G; Cramer et al., 2003), whereas dotted lines indicate minor changes in δ^{13}C that appear to occur

812 also at ODP Site 1262 (see also **Figure 5**).

813

814 **Figure 5.** Stratigraphic correlation between upper Paleocene and lower Eocene sections at Cicogna,

815 ODP Site 1262 (Zachos et al., 2010), and DSDP Site 577 (Cramer et al., 2003; Dickens and Backman,

816 2013). All three sites have independently derived nannofossil biohorizons, polarity chrons and δ^{13}C

817 records, which account for subtle temporal offsets. . Color bands and symbols are the same as in

818 Figure 4. Note the missing record at Site 577 that corresponds with known core gaps.

819

820 **Figure 6.** Plots of (a) bulk carbonate δ^{13}C versus bulk oxygenδ^{18}O, and (b) bulk carbonate δ^{13}C versus

821 $CaCO_3$ contentfor samples from the Cicogna section. The black arrow shows the expected effect of

822 burial diagenesis. Grey and black dash lines are linear trendlines for Paleocene and Eocene samples,

823 respectively. - Note the clear distinction in δ^{13}C for Paleocene and Eocene samples, which relates

824 to a long-term decrease in δ^{13}C (**Figure 5**).

825

826 **Figure 7.** Relative (%) and semi-quantitative (N/mm^2) abundances of selected calcareous

827 nannofossil genera across the Cicogna section. Also shown are the lithostratigraphy,

828 magnetostratigraphy, biostratigraphy and carbon isotope (δ^{13}C) stratigraphy at the Cicogna section

829 (**Figure 4**). Color bands and symbols are the same as in previous figures.

830

831 **Figure 8.** Relative (%) and semi-quantitative (N/mm^2) abundances of selected, mainly late

832 Paleocene, calcareous nannofossil taxa across the Cicogna section. Also shown are the

833 lithostratigraphy, magnetostratigraphy, biostratigraphy and carbon isotope (δ^{13}C) stratigraphy at

834 the Cicogna section (**Figure 4**). Color bands and symbols are the same as in previous figures.

36

835 **Figure 9.** Relative (%) and semi-quantitative (N/mm^2) abundances of selected, mainly early Eocene,

836 calcareous nannofossil taxa across the Cicogna section. Also shown are the lithostratigraphy,

837 magnetostratigraphy, biostratigraphy and carbon isotope (δ^{13}C) stratigraphy at the Cicogna section

838 (**Figure 4**). Color bands and symbols are the same as in previous figures.

839

840 **Figure 10.** Statistical Analyses of calcareous nannofossil percentage data of the Cicogna section.

841 Calcareous nannofossils are subdivided into 15 subgroups (*Chiasmolithus, Coccolithus, Ellipsolithus,*

842 *Discoaster, Ericsonia, Fasciculithus, Girgisia, Octolithus, Prinsius, Sphenolithus, Toweius,*

843 *Rhomboaster/Tribrachiatus, Zyghrablithus*, reworking, others). **A)** Principal Component Analysis

844 (PCA) ~~of calcareous nannofossil percentage data of the Cicogna section. Calcareous nannofossils are~~

845 ~~subdivided into 15 subgroups (*Chiasmolithus, Coccolithus, Ellipsolithus, Discoaster, Ericsonia,*~~

846 ~~*Fasciculithus, Girgisia, Octolithus, Prinsius, Sphenolithus, Toweius, Rhomboaster/Tribrachiatus,*~~

847 ~~*Zyghrablithus*, reworking, others).~~ Scatter plot of percentage data of calcareous nannofossil taxa of

848 samples from the Cicogna section in terms of the first and second component. Each sample is

849 represented by a circle and labelled. ~~Different colors serve to separate sub-sets of samples having~~

850 ~~the same age.~~Green and blue shaded areas are the ellipses containing 95% of the data points of

851 Paleocene Group and Eocene group, respectively; B) Multivariate ANalysis Of VAriance (MANOVA).

852 Scatter graph and biplot. Each sample is represented by a black symbol (quadrangle and diamond)

853 and labelled. Green, red and blue shaded areas are the ellipses containing 95% of the data points of

854 Paleocene Group, PETM group and Eocene group, respectively.

855

856 **Figure 11.** Comparison of δ^{13}C profiles and semi-quantitative abundance patterns of selected

857 calcareous nannofossil taxa from the Cicogna section and ODP Site 1262. Calcareous nannofossil

858 biohorizons from DSDP Site 577 are reported in the right part of the figure. Orange and yellow bands

859 mark CIEs shown in previous figures.). Color bands and symbols are the same as in previous figures.

860

861 **Plate I**. Images of selected calcareous nannofossil taxa from samples of the Cicogna section. Scale

862 bar 10µm.

863 **1.** *Discoaster lodoensis* Bramlette and Riedel, 1954. Parallel light. Parallel nicols. Sample CIC/07-492.

864 **2-3.** *Girgisia gammation* (Bramlette Sullivan, 1961) Varol 1989. Crossed nicols. Sample CIC/07-437.

865 **4-5.** *Chiphragmalithus calathus* Bramlette and Sullivan, 1961; 4.Parallel light; 5. Crossed nicols.

866 Sample CIC/07-447. **6-7.** *Sphenolithus radians* Deflandre in Grassé, 1952. 6. Crossed nicols 0°; 7.

867 Crossed nicols 45°. Sample CIC/07-437. **8.** *Tribrachiatus orthostylus* Shamrai, 1963. Parallel light.

868 Sample 208-1262A-11H- 1, 149. Sample CIC/07-447. **9-10.** *Tribrachiatus contortus* (Stradner 1958)

869 Bukry 1972. Parallel light. Sample CIC/07-335. **11-13.** *Zyghrablithus bijugatus* (Deflandre in

870 Deflandre and Fert, 1954) Deflandre, 1959. Crossed nicols. Sample CIC/07-437. **14.** *Discoaster*

871 *salisburgensis* Stradner, 1961. Parallel light. Sample CIC/07-335. **15.** *Discoaster diastypus* Bramlette

872 and Sullivan, 1961. Parallel light. Sample CIC/07-335. **16.** *Fasciculithus tympaniformis* Hay and

873 Mohler in Hay et al. 1967. Crossed nicols. Sample CIC/07-335. **17.** *Octolithus multiplus* (Perch-

874 Nielsen, 1973) Romein, 1979. Crossed nicols. Sample CIC/07-122. **18.** *Discoaster multiradiatus*

875 Bramlette and Riedel 1954. Parallel light. Sample CIC/07-122. **19.** *Toweius pertusus* (Sullivan, 1965)

876 Romein, 1979. Crossed nicols. Sample CIC/07-122. **20.** *Toweius occultatus* (Locker, 1967) Perch-

877 Nielsen, 1971. Crossed nicols. Sample CIC/07-122. **21.** *Toweius eminens* (Bramlette and Sullivan,

878 1961) Perch-Nielsen, 1971. Crossed nicols. Sample CIC/07-029. **22.** *Toweius eminens* (Bramlette and

879 Sullivan, 1961) Perch-Nielsen, 1971. Crossed nicols. Sample CIC/07-029. **23.** *Toweius eminens*

880 (Bramlette and Sullivan, 1961) Perch-Nielsen, 1971. Crossed nicols. Sample CIC/07-122. **24.** *Prinsius*

881 *bisulcus* (Stradner, 1963) Hay and Mohler, 1967. Crossed nicols. Sample CIC/07-029. **25.** *Ericsonia*

882 *robusta* Bramlette and Sullivan 1961. Crossed nicols. Sample CIC/07-029. **26-27.** *Sphenolithus*

883 *anarrhopus* Bukry and Bramlette, 1969. 24. Crossed nicols 0°; 25. Crossed nicols 45°. Sample CIC/07-

884 029. **28-29.** *Zyghrablithus bijugatus* (Deflandre in Deflandre and Fert, 1954) Deflandre, 1959.

885 Crossed nicols. Sample CIC/07-122. **30.** *Thoracosphaera saxea* (Stradner, 1961). Crossed nicols.

886 Sample CIC/07-122.

887

888 **Table 1**. Stratigraphic heights and ages of polarity chron boundaries, key calcareous nannofossil

889 datums, and CIEs at the Cicogna Section and ODP Site 1262.

890

891 **9. SUPPLEMENTARY INFORMATION**

892 **S12. Additional information on statistical analysis**

893 Nannofossil data from the Cicogna section (NE Italy) were subjected to statistical analysis using the

894 program PAST.

895 For PCA analysis, we additionally provide the biplot and the loading graphs of Component 1 and

896 Component 2 (**Figure S1**).

897 For non-metric multidimensional scaling (MDS) analysis, the species counts were combined to

898 produce a matrix of 15 genera. A square root transformation, used to standardize the matrix, was

899 chosen to minimize the influence of dominant taxa on the ordination (Schneider et al., 2011). Non-

900 metric multidimensional scaling (MDS), using the Bray–Curtis distance metric (**Figure S2**) was

901 applied in order to avoid assumptions as much as possible and guarantee the preservation of the

902 relative differences between the samples (McCune and Grace, 2002).

903 **References**
904 Schneider L.J., Bralower, T.J., Kump, L.R.: Response of nannoplankton to early Eocene ocean
905   destratification, Palaeogeogr., Palaeoclimatol., Palaeoecol., 310, 152-162, 2011.
906 McCune, B., Grace, J.B.: Analysis of Ecology Communities. MjM Software Design, Gleneden Beach,
907   Oregon, 2002.
908

909 **S2. Further explanation regarding biostratigraphic calcareous nannofossil counts**

910 The high abundance, widespread distribution and rapid evolution of calcareous nannofossils make

911 them one of the most powerful tool to date Cenozoic marine sediments. The use of semi-

912 quantitative counting and the gathering of high resolved datasets greatly enhance their correlation

913 potential (Backman et al., 2012; Agnini et al., 2014).

914 The methodology used in this study for samples of ODP Site 1262 is that proposed by Backman

915 and Shackleton (1983), which consists in counting the number of calcareous nannofossils belonging

916 to a specific taxon present in a prefixed area (1 mm^2). Because of significant dilution by terrigenous

917 material in samples from the Cicogna section, we extended the study area to 9 mm^2. To further

918 appreciate the importance of semi-quantitative estimates and high-resolution sampling, we

919 compare the Top *D. multiradiatus* and Base *D. lodoensis* as recorded from the Cicogna section, ODP

920 Site 1262 and DSDP Site 550 (**FigureS3**). At Cicogna and ODP Site 1262, we provide detailed

921 abundance patterns of these two taxa. *Discoaster multiradiatus* shows a first decrease in abundance

922 preceding the H1 event and a definitive disappearance just before the onset of the I1 event.

923 *Discoaster lodoensis* displays a first presence in the I1 event, which is followed by an interval of

924 absence that eventually leads to its continuous and common presence close to the onset of the X

925 event (**Figure S3**). Datasets from the Cicogna section and ODP Site 1262 allow a very detailed

926 characterization of these two biohorizons and the recognition of peculiar features that are not

927 present in the low-resolution qualitative biostratigraphic data available for DSDP Site 550. As a

928 consequence, the stratigraphic position of Top *D. multiradiatus* and Base *D. lodoensis* at DSDP Site

929 550 are inaccurate. We hope that this simple exercise could serve to emphasize the crucial

930 importance of producing high-resolution semi-quantitative data to obtain the most reliable

931 biostratigraphic results.

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References

Ali, J. and Hailwood, E.: Magnetostratigraphic (re)calibration of the Paleocene/Eocene boundary interval in Holes 550 and 549, Goban Spur, eastern North Atlantic, Earth Planet. Sci. Lett., 161, 201-213, 1998.

Agnini, C., Fornaciari, E., Raffi, I., Rio, D., Röhl, U., and Westerhold, T.: High-resolution nannofossil biochronology of middle Paleocene to early Eocene at ODP Site 1262: implications for calcareous nannoplankton evolution. Mar. Micropaleontol., 64, 215-248, doi:10.1016/j.marmicro.2007.05.003, 2007b.

Agnini, C., Fornaciari, E., Raffi, I., Catanzariti, R., Pälike, H., Backman, J., and Rio, D.: Biozonation and biochronology of Paleogene calcareous nannofossils from low and middle latitudes, Newslett. Stratigr., 47, 131-181, doi:10.1127/0078-0421/2014/0042, 2014.

Backman, J., Raffi, I., Rio, D., Fornaciari, E. and Pälike, H., 2012: Biozonation and biochronology of Miocene through Pleistocene calcareous nannofossils from low and middle latitudes, Newsl. Stratigr., 45, 221-244.

Cramer, B. S., Wright, J. D., Kent, D. V., and Aubry, M.-P.: Orbital climate forcing of $\delta^{13}C$ excursions in the late Paleocene–early Eocene (Chrons C24n–C25n), Paleoceanography, 18 (4), 1097, doi:10.1029/2003PA000909, 2003.

Dallanave, E., Agnini, C., Muttoni, G., and Rio, D.: Magneto-biostratigraphy of the Cicogna section (Italy): implications for the late Paleocene-early Eocene time scale, Earth Planet. Sci. Lett., 285, 39-51, doi:10.1016/j.epsl.2009.05.033, 2009.

Müller, C.: Biostratigraphic and paleoenvironmental interpretation of the Goban Spur region based on a study of calcareous nannoplankton, Deep Sea Drill. Project, Initial Rep., 80, 389-414, 1985.

Westerhold, T., Röhl, U., Raffi, I., Fornaciari, E., Monechi, S., Reale, V., Bowles, J., and Evans, H. F.: Astronomical calibration of the Paleocene time, Palaeogeogr. Palaeoclimatol. Palaeoecol., 257, 377-403, doi:10.1016/j.palaeo.2007.09.016, 2008.

Zachos, J. C., Kroon, D., Blum, P., et al.: Early Cenozoic Extreme Climates: The Walvis Ridge Transect, Proc. Ocean Drill. Program, Initial Rep., 208 doi:10.2973/odp.proc.ir.208.2004, 2004.

S3. Looking through "frosty glass": Comparison to records at ODP Site 690

We have presented fairly detailed records of bulk carbonate $\delta^{13}C$ and quantified calcareous nannofossil assemblages for the lower Paleogene section at Cicogna, and compared these records with those at the only two locations with similar information. From this comparison, we suggest that a very detailed template exists for the alignment of $\delta^{13}C$ records and calcareous nannofossil assemblage counts across the early Paleogene **(Figure 11)**, one with much higher resolution than given in most previous work, and one most likely related to changes in past global carbon cycling, oceanography, and calcareous nannoplankton evolution.

Significant variations in calcareous nannofossil abundances definitely happened at multiple locations during the PETM (Bralower, 2002, and references noted in main text). However, it is by no

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972 means clear whether such changes extended across the broader early Paleogene, nor how such

973 changes might compare to those across the PETM. One can certainly speculate that variations in

974 calcareous nannofossil abundance records and bulk carbonate $\delta^{13}C$ records might correlate in fine

975 temporal detail across widely distributed sites throughout the early Paleogene, given well-

976 established calcareous nannofossil biozone schemes (Martini, 1971; Okada and Bukry, 1980; Agnini

977 et al., 2014), and a growing appreciation of a very dynamic carbon cycle over this time interval.

978 Nonetheless, the generation of detailed and coupled multi-million year records for quantified

979 calcareous nannofossil abundances and bulk carbonate $\delta^{13}C$ perplexed one of the referees for this

980 paper, who insisted that we needed to make comparisons with existing work at ODP Site 690 and

981 to explain discrepancies.

982   The lower Paleogene record at Site 690 provides a very good example in which to highlight the

983 basic background and importance of our work. Three holes were drilled and cored at ODP Site 690

984 on Maud Rise (South Atlantic; Figure 1) in 1987 using the advanced piston corer (APC) (Barker et al.,

985 1988). Sediment recovery within each core was nearly 100 %, although some cores were shorter

986 than the full 9.7 m. However, most of the lower Paleogene sequence was retrieved in only one of

987 the holes, 690B (Barker et al., 1988). This is important, because m-scale gaps generally occur

988 between successive cores during APC operations (Ruddiman et al., 1987; Lisiecki and Herbert, 2007).

989 The early Paleogene section at Site 690 is, almost assuredly, incomplete, with "missing" portions at

990 each core break.

991   Sediment from Core 690B-19H has been the focus of numerous papers, as it contains the PETM

992 (Kennett and Stott, 1991; Bains et al., 1999; Bralower, 2002). However, correlating this core to the

993 surrounding sedimentary record at Site 690, and the latter to early Paleogene records at other

994 locations is problematic, at least with any detail. For example, using Hole 690B records, Cramer et

995 al. (2003) estimated that 1.4 Myr occurred between the PETM and the H-1 event. This is incorrect,

996 as the duration is close to 1.8 Myr (Westerhold et al., 2008). Beyond the aforementioned core gaps,

997 there are major issues with the paleomagnetic record of early Paleogene sediments in Hole 690B

998 (Ali et al., 2000). Indeed, Ali et al. (2000) recommend using calcareous nannofossil records for

999 correlation purposes of this interval.

1000 Records of bulk carbonate $\delta^{13}C$ (Cramer et al., 2003) and calcareous nannofossil relative

1001 abundances (Pospichal and Wise, 1990) have been generated using sediment at Hole 690B. When

1002 coupled together (**Figure S4**), these records show similarities to those at Cicogna (**Figure 11**). There

1003 is the long-term late Paleocene-early Eocene drop in $\delta^{13}C$ and several superimposed short-term

1004 negative CIEs. There are also closely coeval changes in calcareous nannofossil abundances, such as

1005 the peak in *D. multiradiatus* across the C event, the subsequent peak in *Fasciculithus* spp., and the

1006 cross-over of *T. contortus* and *T. orthostylus* just before the H-1 event. One can also see the problem

1007 with examining nannofossils at low depth/time resolution and qualitatively. We suggest here a

1008 "frosty glass" hypothesis, where details of Earth system change in the distant past are blurred

1009 presently by poorly resolved stratigraphy. This includes basic problems with aligning sections in

1010 depth and time, as well as interpretable quantification of data at high spatial resolution. Despite the

1011 need for additional work at Site 690, we suggest that available records at this location support the

1012 template offered in the main text.

1013 **References**
1014 Agnini, C., Fornaciari, E., Raffi, I., Catanzariti, R., Pälike, H., Backman, J., and Rio, D.: Biozonation and
1015 biochronology of Paleogene calcareous nannofossils from low and middle latitudes, Newslett.
1016 Stratigr., 47, 131-181, doi:10.1127/0078-0421/2014/0042, 2014.
1017 Ali, J. R., Kent, D. V., and Hailwood, E. A.: Magnetostratigraphic reinvestigation of the
1018 Palaeocene/Eocene boundary interval in Hole 690B, Maud Rise, Antarctica, Geophys. J. Int.,
1019 141 (3), 639-646, doi: 10.1046/j.1365-246X.2000.00109.x, 2000.
1020 Bains, S., Corfield, R. M. and Norris R. D.: Mechanisms of climate warming at the end of the
1021 Paleocene, Science, 285, 724-727, 1999.
1022 Barker, P.E, Kennett, J.P., et al.: Proc. ODP, Init. Repts., 113, 1-774,
1023 doi:10.2973/odp.proc.ir.113.1988, 1988.
1024 Bralower, T. .J.: Evidence of surface water oligotrophy during the Paleocene–Eocene thermal
1025 maximum: nannofossil assemblage data from Ocean Drilling Program Site 690, Maud Rise,
1026 Weddell Sea, Paleoceanography, 17, 1029-1042, doi:10.1029/2001PA000662, 2002.

43

1027 Cramer, B. S., Wright, J. D., Kent, D. V., and Aubry, M.-P.: Orbital climate forcing of δ^{13}C excursions
1028   in the late Paleocene-early Eocene (Chrons C24n–C25n), Paleoceanography, 18 (4), 1097,
1029   doi:10.1029/2003PA000909, 2003.
1030 Kennett, J. P. and Stott, L. D.: Abrupt deep-sea warming, palaeoceanographic changes and benthic
1031   extinctions at the end of the Palaeocene, Nature 353, 225-229 doi:10.1038/353225a0, 1991.
1032 Lisiecki, L. E., and Herbert, T. D.: Automated composite depth scale construction and estimates of
1033   sediment core extension, Paleoceanography, 22(4), PA4213, doi:10.1029/2006PA001401,
1034   2007.
1035 Martini, E.: Standard Tertiary and Quaternary calcareous nannoplankton zonation, in: Proceedings
1036   of the 2nd Planktonic Conference, 2, Tecnoscienza, Roma, 739-785, 1971.
1037 Okada, H., and Bukry, D.: Supplementary modification and introduction of code numbers to the low-
1038   latitude coccolith biostratigraphic zonation (Bukry, 1973; 1975), Mar. Micropaleontol., 5, 321-
1039   325, doi:10.1016/0377-8398(80)90016-X, 1980.
1040 Pospichal, J. J., and Wise Jr., S. W.: Paleocene to middle Eocene calcareous nannofossils of Maud
1041   Rise, Weddell Sea, Proc. Ocean Drill. Program, Sci. Results, 113, 613-638,
1042   doi:10.2973/odp.proc.sr.113.205.1990, 1990.
1043 Ruddiman, W. F., Cameron, D., and Clement, B. M.: Sediment disturbance and correlation of offset
1044   holes drilled with the hydraulic piston corer, Proc. Deep Sea Drill. Project, Init. Repts., 94, 615–
1045   634. doi:10.2973/dsdp.proc.94.111.1987, 1987.
1046 Westerhold, T., Röhl, U., Raffi, I., Fornaciari, E., Monechi, S., Reale, V., Bowles, J., and Evans, H. F.:
1047   Astronomical calibration of the Paleocene time, Palaeogeogr. Palaeoclimatol. Palaeoecol.,
1048   257, 377-403, doi:10.1016/j.palaeo.2007.09.016, 2008.
1049

1050 **Supplementary figure captions Figure S1**. PCA plots of calcareous nannofossil data from the Cicogna

1051 section (Italy). A) Loading plot of Component 1; B) Loading plot of Component 2; C) Biplot.

1052 **Figure S2**. Non-metric multidimensional scaling (NMS) plot of calcareous nannofossil data from the

1053 Cicogna section (Italy). Grey dots = barren to virtually barren samples.

1054 **Figure S3**. Abundance patterns of *D. multiradiatus* and *D. lodoensis* from the Cicogna section, ODP

1055 Site 1262 and DSDP Site 550. For these three successions paleomagnetic (Ali and Hailwood, 1998;

1056 Dallanave et al., 2009; Westerhold et al., 2008), carbon isotope (Cramer et al., 2003; Zachos et al.,

1057 2004; this study) and calcareous nannofossil data (Müller, 1985; Agnini et al., 2007, this study) are

1058 available. Top *D. multiradiatus* and Base *D. lodoensis* are clearly recognizable at Cicogna and ODP

1059 Site 1262, where quantitative counts have been performed. By contrast, qualitative data from DSDP

1060 Site 550 do not provide reliable biostratigraphic data P= present; R=rare; F=few; C=common;

1061 A=abundant; V=very abundant.

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1062 **Figure S4**. Carbon isotope data from ODP Site 690 (Cramer et al., 2003) plotted against qualitative

1063 abundance estimates of selected calcareous nannofossil taxa (Pospichal and Wei, 1990). Top

1064 *Fasciculithus* spp. (Aubry et al., 1996)

1065 **Table S1**. Bulk carbonate stable isotopes and carbonate content of samples from the Cicogna

1066 section.

1067 **Table S2**. Calcareous nannofossil assemblage counts for samples from the Cicogna section.

1068 **Table S3.** Dataset used to perform the principal component analysis for calcareous nannofossil

1069 assemblages from the Cicogna section. Calcareous nannofossils are subdivided in 15 subgroups

1070 (*Chiasmolithus, Coccolithus, Ellipsolithus, Discoaster, Ericsonia, Fascicuithus, Girgisia, Octolithus,*

1071 *Prinsius, Sphenolithus, Toweius, Rhomboaster/Tribrachiatus, Zyghrablithus*, reworking, others). In

1072 order to avoid the closed-sum effect that derives from the use of percentage data, we apply a log

1073 transformation of raw data.

1074 **Table S4.** Dataset used to perform the non-metric multidimensional scaling (MDS) for calcareous

1075 nannofossil assemblages from the Cicogna section. Calcareous nannofossils are subdivided in 15

1076 subgroups (*Chiasmolithus, Coccolithus, Ellipsolithus, Discoaster, Ericsonia, Fascicuithus, Girgisia,*

1077 *Octolithus, Prinsius, Sphenolithus, Toweius, Rhomboaster/Tribrachiatus, Zyghrablithus*, reworking,

1078 others). A square root transformation was used to minimize the influence of dominant taxa on the

1079 ordination.

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10   REFERENCES

1082

1083

Adelseck Jr., C. G., Geehan, G. W., and Roth, P. H.: Experimental evidence for the selective dissolution and overgrowth of calcareous nannofossils during diagenesis, Geol. Soc. Am. Bull., 84, 2755-2762, 1973.

Agnini, C., Muttoni, G., Kent, D. V., and Rio, D.: Eocene biostratigraphy and magnetic stratigraphy from Possagno, Italy: The calcareous nannofossil response to climate variability, Earth Planet. Sci. Lett., 241, 815-830, doi:10.1016/j.epsl.2005.11.005, 2006.

Agnini, C., Fornaciari, E., Rio, D., Tateo, F., Backman, J., and Giusberti, L.: Responses of calcareous nannofossil assemblages, mineralogy and geochemistry to the environmental perturbations across the Paleocene/Eocene boundary in the Venetian Pre–Alps, Mar. Micropaleontol., 63, 19-38, doi:10.1016/j.marmicro.2006.10.002, 2007a.

Agnini, C., Fornaciari, E., Raffi, I., Rio, D., Röhl, U., and Westerhold, T.: High-resolution nannofossil biochronology of middle Paleocene to early Eocene at ODP Site 1262: implications for calcareous nannoplankton evolution. Mar. Micropaleontol., 64, 215-248, doi:10.1016/j.marmicro.2007.05.003, 2007b.

Agnini, C., Macrì, P., Backman, J., Brinkhuis, H., Fornaciari, E., Giusberti, L., Luciani, V., Rio, D., Sluijs, A., and Speranza, F.: An early Eocene carbon cycle perturbation at 52.5 Ma in the Southern Alps: Chronology and biotic response, Paleoceanography, 24, PA2209, doi:10.1029/2008PA001649, 2009.

Agnini, C., Fornaciari, E., Giusberti, L., Grandesso, P., Lanci, L., Luciani, V., Muttoni, G., Pälike, H., Rio, D., Spofforth, D. J. A., and Stefani, C.: Integrated bio-magnetostratigraphy of the Alano section (NE Italy): a proposal for defining the middle-late Eocene boundary, Geol. Soc. Am. Bull., 123, 841-872, doi:10.1130/B30158.1, 2011.

Agnini, C., Fornaciari, E., Raffi, I., Catanzariti, R., Pälike, H., Backman, J., and Rio, D.: Biozonation and biochronology of Paleogene calcareous nannofossils from low and middle latitudes, Newslett. Stratigr., 47, 131-181, doi:10.1127/0078-0421/2014/0042, 2014.

Aitchison, J.: The Statistical Analysis of Compositional Data. Chapman and Hall, London - New York, 12, 1-416, 1986.

Alcober, J. A. and Jordan, R. W.: An interesting association between *Neosphaera coccolithomorpha* and *Ceratolithus cristatus,* Eur. J . Phycol, 32, 91-93, 1997.

Angori, E., Bernaola, G., and Monechi, S.: Calcareous nannofossil assemblages and their response to the Paleocene-Eocene Thermal Maximum event at different latitudes: ODP Site 690 and Tethyan sections, Geol. Soc. Am. Spec. Pap., 424, 69-85, doi:10.1130/2007.2424(04), 2007.

Aubry, M.-P.: Handbook of Cenozoic Calcareous Nannoplankton, book 1, Ortholithae (Discoaster), Am. Mus. Nat. Hist. Micropaleontol. Press, New York, 1-263, 1984.

Aubry, M.-P.: Handbook of Cenozoic Calcareous Nannoplankton, book 2, Ortholithae (Holococcoliths, Ceratoliths, Ortholiths and Other), Am. Mus. Nat. Hist. Micropaleontol. Press, New York, 1-279, 1988.

Aubry, M.-P.: Handbook of Cenozoic Calcareous Nannoplankton, book 3, Ortholithae (Pentaliths and Other), Heliolithae (Fasciculiths, Sphenoliths and Other), Am. Mus. Nat. Hist. Micropaleontol. Press, New York, 1-279, 1989.

Aubry, M.-P.: Handbook of Cenozoic Calcareous Nannoplankton, book 4, Heliolithae (Helicoliths, Cribriliths, Lopadoliths and Other), Am. Mus. Nat. Hist. Micropaleontol. Press, New York, 1-381, 1990.

1127 Aubry, M.-P.: Early Paleogene calcareous nannoplankton evolution: a tale of climatic amelioration,
1128   in: Late Paleocene–early Eocene Biotic and Climatic Events in the Marine and Terrestrial
1129   Records, Columbia University Press, New York, 158-201, 1998.
1130 Aubry, M.-P.: Handbook of Cenozoic Calcareous Nannoplankton, book 5, Heliolithae (Zygoliths and
1131   Rhabdoliths), Am. Mus. Nat. Hist. Micropaleontol. Press, New York, 1-368, 1999.
1132 Aubry, M.-P.: Late Paleocene–early Eocene sedimentary history in western Cuba: implications for
1133   the LPTM and for regional tectonic history, Micropaleontol., 45, 5-18, 1999.
1134 Aubry, M.-P. and Salem, R.: The Dababiya Core: A window into Paleocene to early Eocene
1135   depositional history in Egypt based on coccolith stratigraphy, Stratigraphy, 9, 287-346, 2012.
1136 Baccelle, L. and Bosellini, A.: Diagrammi per la stima visiva della composizione percentuale nelle
1137   rocce sedimentary, Ann. Univ. Ferrara, 9, 4, 59-62, 1965.
1138 Backman, J.: Late Paleocene to middle Eocene calcareous nannofossil biochronology from the
1139   Shatsky Rise, Walvis Ridge and Italy, Palaeogeogr. Palaeoclimatol. Palaeoecol., 57, 43-59,
1140   1986.
1141 Backman, J. and Shackleton N. J.: Quantitative biochronology of Pliocene and early Pleistocene
1142   calcareous nannoplankton from the Atlantic, Indian and Pacific Oceans, Mar. Micropaleontol.,
1143   8, 141-170, doi:10.1016/0377-8398(83)90009-9, 1983.
1144 Baumann, K.-H., Andruleit, H., Böckel, B., Geisen, M., and Kinkel, H.: The significance of extant
1145   coccolithophores as indicators of ocean water masses, surface water temperature, and
1146   palaeoproductivity: a review, Paläontol. Zeitsch., 79, 93-112, 2005.
1147 Bernoulli, D., and Jenkyns, H.C.: Alpine, Mediterranean, and Central Atlantic Mesozoic facies in
1148   relation to the early evolution of the Tethys, in: Modern and Ancient Geosynclinal
1149   Sedimentation, Society for Sedimentary Geology (SEPM) Special Publication, 19, 19–160,
1150   1974.
1151 Bernoulli, D., Caron, C., Homewood, P., Kalin, O., and Van Stuijvenberg, J.: Evolution of continental
1152   margins in the Alps, Schweiz. Miner. Petrog., 59, 165-170, 1979.
1153 Bijl, P. K., Schouten, S., Sluijs, A., Reichart, G. J., Zachos, J. C., Brinkhuis, H.: Early Palaeogene
1154   temperature evolution of the southwest Pacific Ocean, Nature, 461, 776–779,
1155   doi:10.1038/nature08399, 2009.
1156 Billard, C. and Innouye, I.: What is new in coccolithophore biology?, in: Coccolithophores - From
1157   Molecular Processes to Global Impact, Springer, Berlin, 1-29, 2004.
1158 Bordiga, M., Henderiks, J., Tori, F., Monechi, S., Fenero, R., and Thomas, E.: The Eocene–Oligocene
1159   transition at ODP Site 1263, Atlantic Ocean: decreases in nannoplankton size and abundance
1160   and correlation with benthic foraminiferal assemblages, Clim. Past Discuss., 11, 1615-1664,
1161   doi:10.5194/cpd-11-1615-2015, 2015.
1162 Bornemann, A. and Mutterlose, J.: Calcareous nannofossil and δ^{13}C records from the Early
1163   Cretaceous of the Western Atlantic Ocean: evidence for enhanced fertilization across the
1164   Berriasian–Valanginian transition, Palaios, 23, 821-832, 2008.
1165 Boucot, A. J.: Evolution and Extinction Rate Controls. Elsevier, Amsterdam, The Netherlands, 250p.
1166   1975.
1167 Bown, P. R.: Paleogene calcareous nannofossils from the Kilwa and Lindi areas of coastal Tanzania:
1168   Tanzania Drilling Project Sites 1 to 10, J. Nannoplankt. Res., 27, 21-95, 2005.
1169 Bown, P., and Pearson, P.: Calcareous plankton evolution and the Paleocene/Eocene thermal
1170   maximum event: New evidence from Tanzania, Mar. Micropaleontol., 71, 60-70, doi:
1171   10.1016/j.marmicro.2009.01.005, 2009.
1172 Bown, P. R. and Young, J. R.: Techniques, in: Calcareous Nannofossil Biostratigraphy, Chapman &
1173   Hall, London, 16-28, 1998.

1174 Bown, P. R., Lees, J. A., and Young, J. R.: Calcareous nannoplankton evolution and diversity through
1175   time, in: Coccolithophores - From Molecular Processes to Global Impact, Springer, Berlin, 481-
1176   508, 2004.
1177 Bralower, T..J.: Evidence of surface water oligotrophy during the Paleocene–Eocene thermal
1178   maximum: nannofossil assemblage data from Ocean Drilling Program Site 690, Maud Rise,
1179   Weddell Sea, Paleoceanography, 17, 1029-1042, doi:10.1029/2001PA000662, 2002.
1180 Bralower,T. J., Mutterlose, J.: Calcareous nannofossil biostratigraphy of ODP Site 865, Allison Guyot,
1181   Central Pacific Ocean: a tropical Paleogene reference section. Proc Ocean Drill Prog Sci Results,
1182   143, 31-72, doi:10.2973/odp.proc.sr.143.204.1995, 1995.
1183 Broecker, W. S. and Peng, T.-H.: Tracers in the Sea, Eldigio Press, LamontDoherty Geological
1184   Observatory, Palisades, New York, 1-690, 1982.
1185 Buccianti, A., Mateu-Figueras, G., Pawlowsky-Glahn, V.: Compositional data analysis in the
1186   geosciences: From theory to practice, Geol. Soc. London, London, Spec. Publ., 264, 1- 12, doi:
1187   10.1144/GSL.SP.2006.264, 2006.Bukry, D.: Low-latitude coccolith biostratigraphic zonation.
1188   Initial Rep. Deep Sea Drill. Proj., 15, 685–703, 1973.
1189 Cati, A., Sartorio, D., and Venturini, S.: Carbonate platforms in the subsurface of the northern
1190   Adriatic area, Mem. Soc. Geol. It., 40, 295-308, 1989.
1191 Corfield, R. M.: Palaeocene oceans and climate: An isotopic perspective, Earth Sci. Rev., 37, 225-252,
1192   doi.org/10.1016/0012-8252(94)90030-2, 1994.
1193 Costa, V., Doglioni, C., Grandesso, P., Masetti, D., Pellegrini, G.B., and Tracanella, E.: Carta Geologica
1194   d'Italia, Foglio 063, Belluno: Roma, Servizio Geologico d'Italia, scale 1:50,000, 1 sheet + 74 p.,
1195   1996.
1196 Cramer, B. S., Wright, J. D., Kent, D. V., and Aubry, M.-P.: Orbital climate forcing of $\delta^{13}C$ excursions
1197   in the late Paleocene– early Eocene (Chrons C24n –C25n), Paleoceanography, 18 (4), 1097,
1198   doi:10.1029/2003PA000909, 2003.
1199 Cramer, B. S., Toggweiler, J. R., Wright, J. D., Katz, M. E., and Miller, K. G.: Ocean overturning since
1200   the Late Cretaceous: Inferences from a new benthic foraminiferal isotope compilation,
1201   Paleoceanogaphy, 24, PA4216, doi:10.1029/2008PA001683, 2009.
1202 Dallanave, E., Agnini, C., Muttoni, G., and Rio, D.: Magneto-biostratigraphy of the Cicogna section
1203   (Italy): implications for the late Paleocene-early Eocene time scale, Earth Planet. Sci. Lett., 285,
1204   39-51, doi:10.1016/j.epsl.2009.05.033, 2009.
1205 Dallanave, E., Agnini, C., Bachtadse, V., Muttoni, G., Crampton, J. S., Strong, C. P., Hines, B. R., Hollis,
1206   C. J., and Slotnick., B. S.: Early to middle Eocene magnetostratigraphy of the southwest Pacific
1207   Ocean and climate influence on sedimentation: insights from the Mead Stream section, New
1208   Zealand, Geol. Soc. Am. Bull., 127, 643-660, doi: 10.1130/B31147.1, 2015.
1209 DeConto, R. M., Galeotti, S., Pagani, M., Tracy, D., Schaefer, K. Zhang, T., Pollard, D., and Beerling D.
1210   J.: Past extreme warming events linked to massive carbon release from thawing permafrost,
1211   Nature, 484, 87-91, doi:10.1038/nature10929, 2012.
1212 Dickens, G. R.: Methane oxidation during the late Palaeocene thermal maximum, Bull. Soc. Geol.
1213   France, 171 (1), 37-49, 2000.
1214 Dickens, G. R.: Rethinking the global carbon cycle with a large, dynamic and microbially mediated
1215   gas hydrate capacitor, Earth Planet. Sci. Lett., 213, 169-183, 2003.
1216 Dickens, G. R. and Backman, J.: Core alignment and composite depth scale for the lower Paleogene
1217   through uppermost Cretaceous interval at Deep Sea Drilling Project Site 577, Newslett.
1218   Stratigr., 46, 47-68, doi:10.1127/0078-0421/2013/0027, 2013.
1219 Dickens, G. R., Castillo, M. M., and Walker, J. C. G.: A blast of gas in the latest Paleocene: Simulating
1220   first-order effects of massive dissociation of oceanic methane hydrate, Geology, 25, 259–262,
1221   1997.

1222

1223 Doglioni, C. and Bosellini, A.: Eoalpine and mesoalpine tectonics in the Southern Alps, Geol.
1224   Rundsch., 77, 734-754, 1987.

1225 Dupuis, C., Aubry, M.-P., Steurbaut, E., Berggren, W.A., Ouda, K., Magioncalda, R., Cramer, B.S., Kent,
1226   D.V., Speijer, R.P., and Heilmann-Clausen, C.: The Dababiya Quarry section: lithostratigraphy,
1227   geochemistry and paleontology. Micropaleontology, 49 (suppl. 1), 41–59. 2003.

1228 Erba, E., Bottini, C., Weissert, H. J., Keller, C. E.: Calcareous nannoplankton response to surface-
1229   water acidification around Oceanic Anoxic Event 1a. Science 329, 428, doi:
1230   10.1126/science.1188886, 2010.

1231 Frank, T. D., Arthur, M. A., and Dean W. E.: Diagenesis of Lower Cretaceous pelagic carbonates,
1232   North Atlantic: paleoceanographic signals obscured, J. Foraminiferal Res., 29, 340–351, 1999.

1233 Galeotti, S., Krishnan, S., Pagani, M., Lanci, L., Gaudio, A., Zachos, J. C., Monechi, S., Morelli, G., and
1234   Lourens, L.: Orbital chronology of early Eocene hyperthermals from the Contessa Road
1235   section, central Italy, Earth Planet. Sci. Lett., 290, 192-200, doi:10.1016/j.epsl.2009.12.021,
1236   2010.

1237 Geisen, M., Young, J. R., Probert, I., Sáez, A. G., Baumann, K.-H., Bollmann, J., Cros, L., De Vargas, C.,
1238   Medlin, L. K., and Sprengel, C.: Species level variation in coccolithophores, in:
1239   Coccolithophores –From Molecular Processes to Global Impact, Springer, Berlin, 327-366,
1240   2004.

1241 Gibbs, S. J., Shackleton, N. J., and Young, J. R.: Orbitally forced climate signals in mid-Pliocene
1242   nannofossil assemblages, Mar. Micropaleontol., 51, 39-56, 2004.

1243 Gibbs, S. J., Bown, P. R., Sessa, J. A., Bralower, T. J., and Wilson, P. A.: Nannoplankton extinction and
1244   origination across the Paleocene-Eocene thermal maximum, Science, 314, 1770-1773,
1245   doi:10.1126/science.1133902, 2006a.

1246 Gibbs, S. J., Bralower, T J., Bown, P. R., Zachos, J. C., and Bybell, L.M.: Shelf and open-ocean
1247   calcareous phytoplankton assemblages across the Paleocene–Eocene thermal maximum:
1248   implications for global productivity gradients, Geology, 34, 233-236, doi:10.1130/G22381.1,
1249   2006b.

1250 Gibbs, S. J., Bown, P. R., Murphy, B. H., Sluijs, A., Edgar, K. M., Pälike, H., Bolton, C. T., and Zachos,
1251   J.C.: Scaled biotic disruption during early Eocene global warming events, Biogeosciences, 9,
1252   4679-4688, doi:10.5194/bg-9-4679-2012, 2012.

1253 Giusberti, L., Rio, D., Agnini, C., Backman, J., Fornaciari, E., Tateo, F., and Oddone, M.: Mode and
1254   tempo of the Paleocene-Eocene Thermal Maximum in an expanded section in the Venetian
1255   Pre-Alps, Geol. Soc. Am. Bull., 119, 391-412, doi:10.1130/B25994.1, 2007.

1256 Giusberti, L., Boscolo Galazzo, F., and Thomas, E.: Benthic foraminifera at the Paleocene/Eocene
1257   thermal maximum in the western Tethys (Forada section): variability in climate and
1258   productivity, Clim. Past Discuss., 11, 4205-4272, doi:10.5194/cpd-11-4205-2015,
1259   2015.Grandesso, P.: Biostratigrafia delle formazioni terziarie del Vallone Bellunese, Boll. Soc.
1260   Geol. Ital., 94, 1323-1348, 1976.

1261 Hallock, P.: Fluctuations in the trophic resource continuum: A factor in global diversity cycles?,
1262   Paleoceanography, 2, 457–471, 1987.

1263 Hammer, Ø, Harper, D. A. T., and Ryan, P. D.: PAST: Paleontological statistics software package for
1264   education and data analysis, Palaeontol. Electron., 4, 9 pp., 2001.

1265 Haq, B. U. and Lohmann, G. P.: Early Cenozoic calcareous nannoplankton biogeography of the
1266   Atlantic Ocean, Mar. Micropaleontol., 1, 119-194, 1976.

1267

1268 Harris, A.D., Miller, K.G., Browning, J. V., Sugarman, P.J., Olsson, R.K., Cramer, B.S. and Wright, J.D.;.
1269   Integrated stratigraphic studies of Paleocene- lowermost Eocene sequences, New Jersey

1270   Coastal Plain: Evidence for glacioeustatic control. Paleoceanography, 25, PA3211,
1271   doi:10.1029/2009PA001800, 2010.Hay, W. W.: Carbonate fluxes and calcareous
1272   nannoplankton, in: Coccolithophores - From Molecular Processes to Global Impact, Springer,
1273   Berlin, 508-528, 2004.

1274 Hönisch, B., Ridgwell, A., Schmidt, D. N., Thomas, E., Gibbs, S.J., Sluijs, A., Zeebe, R., Kump, L.,
1275   Martindale, R. C., Greene, S. E., Kiessling, W., Ries, J., Zachos, J. C., Royer, D. L., Barker, S.,
1276   Marchitto Jr., T. M., Moyer, R., Pelejero, C., Ziveri, P., Foster, G. L., and Williams, B.: The
1277   geological record of ocean acidification, Science, 335, 1058-1063,
1278   doi:10.1126/science.1208277, 2012.

1279 Hollis, C.J., Taylor, K. W. R., Handley, L., Pancost, R. D., Huber, M., Creech, J. B., Hines, B.R., Crouch,
1280   E. M., Morgans, H. E. G., Crampton, J. S., Gibbs, S., Pearson, P. N., and Zachos, J. C.: Early
1281   Paleogene temperature history of the southwest Pacific Ocean: Reconciling proxies and
1282   models, Earth Planet. Sci. Lett., 349-350, 53-66, 2012.

1283 Huber, M., and Caballero, R.: The Early Eocene equable climate problem revisited, Clim. Past, 7, 603-
1284   633, 2011.

1285 Iglesias-Rodriguez, M. D., Halloran, P. R., Rickaby, R. E. M., Hall, I. R., Colmenero-Hidalgo, E., Gittins,
1286   J. R., Green, D. R. H., Tyrrell, T., Gibbs, S. J., Von Dassow, P., Rehm, E., Armbrust, E. V., and
1287   Boessenkool, K. P.: Phytoplankton calcification in a high-CO_2 world, Science, 320, 336-340,
1288   doi:10.1126/science.1154122, 2008.

1289 Jiang, S., and Wise Jr., S.W: Distinguishing the influence of diagenesis on the paleoecological
1290   reconstruction of nannoplankton across the Paleocene/Eocene Thermal Maximum: An
1291   example from the Kerguelen Plateau, southern Indian Ocean, Mar. Micropaleontol., 72, 49-
1292   59, doi: 10.1016/j.marmicro.2009.03.003, 2009.

1293 Keeling, C. D. and Whorf, T. P.: Atmospheric carbon dioxide record from Mauna Loa, in: Oak Ridge
1294   Laboratory Trends: A Compendium of Data on Global Change. Carbon Dioxide Information
1295   Analysis Center, Oak Ridge National Laboratory, U.S. Department of Energy, Oak Ridge,
1296   Tennessee, U.S.A., http://cdiac.esd.ornl.gov/trends/co2/sio-keel-flask/sio-keel-flask.html,
1297   2004.

1298 Kirtland-Turner, S., Sexton, P. F., Charles, C. D., and Norris, R. D.: Persistence of carbon release
1299   events through the peak of early Eocene global warmth, Nature Geosci., 7, 748-751,
1300   doi:10.1038/ngeo2240, 2014.

1301 Kleypas, J. A., Feely, R. A., Fabry, V. J., Langdon, C., Sabine, C. L., and Robbins, L. L.: Impacts of ocean
1302   acidification on coral reefs and other marine calcifiers: A guide for future research, Contrib.
1303   No. 2857, NOAA/Pacific Marine Environm. Lab., 88 pp., 2006.

1304 Komar, N., Zeebe, R. E., and Dickens, G.R.: Understanding long-term carbon cycle trends: The late
1305   Paleocene through the early Eocene. Paleoceanography, 28, 650-662,
1306   doi:10.1002/palo.20060, 2013.

1307 Krishnan, S., Pagani, M., Agnini, C.: Leaf waxes as recorders of paleoclimatic changes during the
1308   Paleocene-Eocene Thermal Maximum: Regional expressions from the Belluno Basin. Organic
1309   Geochemistry, 80, 8-17, doi: 10.1016/j.orggeochem.2014.12.005, 2015.

1310 Kroopnick, P. M.: The distribution of ^{13}C of ΣCO_2 in the world oceans, Deep Sea Res. Part A, 32, 57-
1311   84, 1985.

1312 Kucera, M., Malmgren, B.A.: Logratio transformation of compositional data — a resolution of the
1313   constant sum constraint, Mar. Micropaleontol., 34, 117-120, 1998.Kump, L. R., Bralower, T. J.,
1314   and Ridgwell, A.: Ocean acidification in deep time, Oceanography, 22, 94-107, 2009.

1315 Kurtz, A. C., Kump, L. R., Arthur, M. A., Zachos, J. C., and Paytan, A.: Early Cenozoic decoupling of the
1316   global carbon and sulfur cycles, Paleoceanography, 18, 1090, doi:10.1029/2003PA000908,
1317   2003.

1318 Langer, G., M., Geisen, Baumann, K.-H., Kläs, J., Riebesell, U., Thoms, S., and Young, J. R.: Species-
1319   specific response of calcifying algae to changing seawater carbonate chemistry, Geochem.
1320   Geophys. Geosyst., 7, Q09006, doi:10.1029/2005GC001227, 2006.
1321 Leon-Rodriguez, L. and Dickens, G. R.: Constraints on ocean acidification associated with rapid and
1322   massive carbon injections: The early Paleogene record at Ocean Drilling Program Site 1215,
1323   Equatorial Pacific Ocean, Palaeogeogr. Palaeoclimatol. Palaeoecol., 298, 409-420,
1324   doi:org/10.1016/j.palaeo.2010.10.029, 2010.
1325 Lohbeck, K. T., Riebesell, U., and Reusch, T. B. H.: Adaptive evolution of a key phytoplankton species
1326   to ocean acidification, Nature Geosci., 5, 346-351, doi:10.1038/ngeo1441, 2012.
1327 Lourens, L. J., Sluijs, A., Kroon, D., Zachos, J. C., Thomas, E., Röhl, U., Bowles, J., and Raffi, I.:
1328   Astronomical pacing of late Palaeocene to early Eocene global warming events, Nature, 435,
1329   1083-1087, doi:10.1038/nature03814, 2005.
1330 Lunt, D. J., Ridgwell, A., Sluijs, A., Zachos, J. C., Hunter, S., and Haywood, A.: A model for orbital
1331   pacing of methane hydrate destabilization during the Palaeogene, Nature Geosci., 4, 775-778,
1332   doi:10.1038/ngeo1266, 2011.
1333 MacArthur, R. and Wilson, E .O.: The Theory of Island Biogeography, Princeton University Press, ISBN
1334   0-691-08836-5M, 1967.
1335 Marino, M., Maiorano, P., and Lirer, F.: Changes in calcareous nannofossil assemblages during the
1336   Mid-Pleistocene   Revolution,   Mar.   Micropaleontol.,   69,   70-90,
1337   doi:10.1016/j.marmicro.2007.11.010, 2008.
1338 Martini, E.: Standard Tertiary and Quaternary calcareous nannoplankton zonation, in: Proceedings
1339   of the 2nd Planktonic Conference, 2, Tecnoscienza, Roma, 739-785, 1971.
1340 Matter, A., Douglas, R. G., and Perch-Nielsen, K: Fossil preservation, geochemistry and diagenesis of
1341   pelagic carbonates from Shatsky Rise, northwest Pacific, Init. Rep. DSDP, 32, 891–922, 1975.
1342 Mixon, R.B. and Powars D.S.: Folds and faults in the inner Coastal Plain of Virginia and Maryland:
1343   their effect on the distribution and thickness of Tertiary rock units and local geomorphic
1344   history. In Frederiksen N.O., and Kraft K. (Eds.), Cretaceous and Tertiary Stratigraphy,
1345   paleontology, and structure, southwestern Maryland and northeastern Virginia. American
1346   Association of Stratigraphic Palynologists Field Trip Volume and Guidebook (1984), 112–122,
1347   1994
1348 Milliman, J. D.: Production and accumulation of calcium carbonate in the ocean: Budget of a
1349   nonsteady state. Global Biogeochemical Cycles, 7, 927-957, doi: 10.1029/93GB02524, 1993.
1350 Monechi, S., Angori, E., von Salis, K.: Calcareous nannofossil turnover around the Paleocene/Eocene
1351   transition at Alamedilla (southern Spain). Bull. Soc. Geol. Fr., 171, 477–489, 2000.
1352 Mutterlose, J., Linnert, C., and Norris, R.: Calcareous nannofossils from the Paleocene–Eocene
1353   Thermal Maximum of the equatorial Atlantic (ODP Site 1260B): Evidence for tropical warming,
1354   Mar. Micropaleontol., 65, 13-31, doi:10.1016/j.marmicro.2007.05.004, 2007.
1355 Nicolo, M. J., Dickens, G. R., Hollis, C. J., and Zachos, J. C.: Multiple early Eocene hyperthermals: Their
1356   sedimentary expression on the New Zealand continental margin and in the deep sea, Geology,
1357   35, 699-702, doi:10.1130/G23648A.1, 2007.
1358 Norris, R.D., Wilson, P.A, Blum, P., and the Expedition 342 Scientists: Expedition 342 summary. Proc.
1359   IODP, 342, 1-149, doi:10.2204/iodp.proc.342.2014, 2014.
1360 Ogg, J.G., and Bardot, L.: Aptian through Eocene magnetostratigraphic correlation of the Blake Nose
1361   Transect (Leg 171B), Florida continental margin, Proc. ODP, Sci. Results, 171B, 1-58,
1362   doi:10.2973/odp.proc.sr.171B.104.2001 2001.
1363 Okada, H., and Bukry, D.: Supplementary modification and introduction of code numbers to the low-
1364   latitude coccolith biostratigraphic zonation (Bukry, 1973; 1975), Mar. Micropaleontol., 5, 321-
1365   325, doi:10.1016/0377-8398(80)90016-X, 1980.

1366 Pälike, H., Lyle, M. W., Nishi, H., Raffi, I., Ridgwell, A., Gamage, K., Klaus, A., Acton, G. D., Anderson,
1367   L., Backman, J., Baldauf, J. G., Beltran, C., Bohaty, S. M., Bown, P. R., Busch, W. H., Channell, J.
1368   E. T., Chun, C. O. J., Delaney, M. L., Dewang, P., Dunkley Jones, T., Edgar, K. M., Evans, H. F.,
1369   Fitch, P., Foster, G. L., Gussone, N., Hasegawa, H., Hathorne, E. C., Hayashi, H., Herrle, J. O.,
1370   Holbourn, A. E. L., Hovan, S. A., Hyeong, K., Iijima, K., Ito, T., Kamikuri, S.-I., Kimoto, K., Kuroda,
1371   J., Leon-Rodriguez, L., Malinverno, A., Moore, T. C., Murphy, B., Murphy, D. P., Nakamura, H.,
1372   Ogane, K., Ohneiser, C., Richter, C., Robinson, R. S., Rohling, E. J., Romero, O. E., Sawada, K.,
1373   Scher, H. D., Schneider, L., Sluijs, A., Takata, H., Tian, J., Tsujimoto, A., Wade, B. S., Westerhold,
1374   T., Wilkens, R. H., Williams, T., Wilson, P. A., Yamamoto, Y., Yamamoto, S., Yamazaki, T., and
1375   Zeebe, R. E.: A Cenozoic record of the equatorial Pacific carbonate compensation depth,
1376   Nature, 488, 609-614, doi:10.1038/nature11360, 2012
1377 Payros, A., Ortiz, S., Millán, I., Arostegi, J., Orue-Etxebarria, X., and Apellaniz, E.: Early Eocene climatic
1378   optimum: Environmental impact on the north Iberian continental margin, Geol. Soc. Am.
1379   Bull.,127, 1632-1644doi:10.1130/B31278.1, 2015.
1380 Pearson, P. N., van Dongen, B. E., Nicholas, C. J., Pancost, R. D., Schouten, S., Singano, J. M., and
1381   Wade, B. S.: Stable warm tropical climate through the Eocene Epoch, Geology, 35, 211-214,
1382   2007.
1383 Perch-Nielsen, K.: Cenozoic calcareous nannofossils, in: Plankton stratigraphy, Cambridge Univ.
1384   Press, New York, 427-554, 1985.
1385 Pianka, E. R.: On r and K selection. Amer. Natural., 104, 592-597, doi:10.1086/282697, 1970.
1386 Premoli Silva, I. and Sliter, W.V.: Cretaceous paleoceanography: Evidence from planktonic
1387   foraminiferal evolution, Geol. Soc. Am. Spec. Pap., 332, 301-328, doi:10.1130/0-8137-2332-
1388   9.301, 1999.
1389 Raffi, I., Backman, J., and Pälike, H.: Changes in calcareous nannofossil assemblage across the
1390   Paleocene/Eocene transition from the paleo-equatorial Pacific Ocean, Palaeogeogr.
1391   Palaeoclimatol. Palaeoecol., 226, 93-126, doi:10.1016/j.palaeo.2005.05.006, 2005.
1392 Raffi, I., and De Bernardi, B.: Response of calcareous nannofossils to the Paleocene-Eocene Thermal
1393   Maximum: Observations on composition, preservation and calcification in sediments from
1394   ODP Site 1263 (Walvis Ridge - SW Atlantic), Mar. Micropaleontol., 69, 119-138, doi:
1395   10.1016/j.marmicro.2008.07.002, 2008.
1396 Raffi, I., Backman, J., Zachos, J.C., Sluijs, A.: The response of calcareous nannofossil assemblages to
1397   the Paleocene Eocene Thermal Maximum at the Walvis Ridge in the South Atlantic. Mar.
1398   Micropaleontol., 70, 201-212, doi:10.1016/j.marmicro.2008.12.005, 2009.
1399 Riebesell, U., Zondervan, I., Rost, B., Tortell, P. D., Zeebe, R. E., and Morel, F. M. M.: Reduced
1400   calcification of marine plankton in response to increased atmospheric CO_2, Nature, 407, 364-
1401   367, doi:10.1038/35030078, 2000.
1402 Riebesell, U., Bellerby, R. G. J., Engel, A., Fabry, V. J., Hutchins, D. A., Reusch, K.G., Schulz, T. B. H.,
1403   and Morel, F. M. M.: Comment on "Phytoplankton calcification in a high-CO2 world", Science,
1404   322, 1466b, doi:10.1126/science.1161096, 2008.
1405 Romein, A.J.T.: Lineages in early Paleogene calcareous nannoplankton. Utrecht Micropaleont. Bull.,
1406   22, 1-231, 1979.
1407 Rost, B. and Riebesell, U.: Coccolithophores and the biological pump: responses to environmental
1408   changes, in: Coccolithophores - From Molecular Processes to Global Impact, Springer, Berlin,
1409   99-125, 2004.
1410 Roth, P.H.: Jurassic and Lower Cretaceous calcareous nannofossils in the western North Atlantic (Site
1411   534): biostratigraphy, preservation, and some observations on biogeography and
1412   paleoceanography, DSDP Init. Repts., 76, 587-621, 1983.

1413 Roth, P.H., and Thierstein, H.R.: Calcareous nannoplankton: Leg XIV of the Deep Sea Drilling Project,
1414   DSDP Init. Repts., 14, 421-486, 1972.
1415 Schrag, D.P., DePaolo, D.J., and Richter, F.M.: Reconstructing past sea surface temperatures:
1416   correcting for diagenesis of bulk marine carbonate, Geochim. Cosmochim. Acta, 59, 2265-
1417   2278, 1995.
1418 Scholle P. A., and Arthur, M. A.: Carbon isotope fluctuations in Cretaceous pelagic limestones:
1419   potential stratigraphic and petroleum exploration tool, Amer. Assoc. Pet. Geol. Bulletin, 64,
1420   67-87, 1980.
1421 Self-Trail, J. M., Powars, D. S., Watkins, D. K., and Wandless, G.: Calcareous nannofossil assemblage
1422   changes across the Paleocene-Eocene thermal maximum: Evidence from a shelf setting, Mar.
1423   Micropaleontol., 92-93, doi:10.1016/j.marmicro.2012.05.003, 2012.
1424 Shackleton, N. J.: Paleogene stable isotope events, Palaeogeogr. Palaeoclimatol. Palaeoecol., 57, 91-
1425   102, 1986.
1426 Shamrock, J.L.: Eocene calcareous nannofossil biostratigraphy, paleoecology and biochronology of
1427   ODP Leg 122 Hole 762c, Eastern Indian Ocean (Exmouth Plateau). PhD thesis, University of
1428   Nebraska, 2010.
1429 Slotnick, B. S., Dickens, G. R., Nicolo, M. J., Hollis, C. J., Crampton, J. S., Zachos, J. C., and Sluijs, A.:
1430   Large-amplitude variations in carbon cycling and terrestrial weathering during the latest
1431   Paleocene and earliest Eocene: The record at Mead Stream, New Zealand, J. Geol., 120, 1-19,
1432   doi:10.1086/666743, 2012.
1433 Slotnick, B. S., Lauretano, V., Backman, J., Dickens, G. R., Sluijs, A., and Lourens, L.: Early Paleogene
1434   variations in the calcite compensation depth: new constraints using old borehole sediments
1435   from across Ninetyeast Ridge, central Indian Ocean, Clim. Past, 11, 473-493, , doi:10.5194/cp-
1436   11-473-2015, 2015a.
1437 Slotnick, B. S., Dickens, G. R., Hollis, C. J., Crampton, J. S., Strong, C. P., and Phillips, A.: The onset of
1438   the Early Eocene Climatic Optimum at Branch Stream, Clarence River valley, New Zealand, New
1439   Zeal. J. Geol. Geophys., 58, 262-280 doi: 10.1080/00288306.2015.1063514, 2015b.
1440 Spofforth, D. J. A., Agnini, C., Pälike, H., Rio, D., Fornaciari, E., Giusberti, L., Luciani, V., Lanci, L.,
1441   Muttoni, G., and Bohaty, S. M.: Organic carbon burial following the Middle Eocene Climatic
1442   Optimum (MECO) in the central - western Tethys, Paleoceanography, 25, PA3210,
1443   doi:10.1029/2009PA001738, 2010.
1444 Stefani, C. and Grandesso, P.: Studio preliminare di due sezioni del Flysch bellunese, Rend. Soc. Geol.
1445   It., 14, 157-162, 1991.
1446 Stillman, J. H., and Paganini, A. W., Biogeochemical adaptation to ocean acidifcation, J. Exper.
1447   Biology, 218, 1946-1955, doi:10.1242/jeb.115584, 2015.
1448 Stoll, H.M., and Bains, S.: Coccolith Sr/Ca records of productivity during the Paleocene–Eocene
1449   thermal   maximum   from   theWeddell   Sea.   Paleoceanography,   18(2),   1049,
1450   doi:10.1029/2002PA000875, 2003.
1451 Tagliabue, A., and L. Bopp: Towards understanding global variability in ocean carbon-13, Global
1452   Biogeochem. Cycles, 22, GB1035, doi:10.1029/2007GB003037, 2008.
1453 Thierstein, H. R., Geitzenauer, K. R., Molfino, B., and Shackleton, N. J.: Global synchroneity of late
1454   Quaternary coccolith datum levels: validation by oxygen isotopes, Geology, 5, 400-404, 1977.
1455 Thibault, N. and Gardin, S.: The calcareous nannofossil response to the end-Cretaceous warm event
1456   in the   tropical   Pacific,   Palaeogeogr.   Palaeoclimatol.   Palaeoecol.,   291,   239-252,
1457   doi:10.1016/j.palaeo.2010.02.036, 2010.
1458 Tipple, B. J., Pagani, M., Krishnan, S., Dirghangi, S. S., Galeotti, S., Agnini, C., Giusberti, L., and Rio,
1459   D.: Coupled high-resolution marine and terrestrial records of carbon and hydrologic cycles

1460 variations during the Paleocene-Eocene Thermal Maximum (PETM), Earth Planet. Sci. Lett.,
1461 311, 82-92, doi:10.1016/j.epsl.2011.08.045, 2011.
1462 Toffanin, F., Agnini, C., Fornaciari, E., Rio, D., Giusberti, L., Luciani, V., Spofforth, D. J. A., and Pälike,
1463 H.: Changes in calcareous nannofossil assemblages during the Middle Eocene Climatic
1464 Optimum: clues from the central-western Tethys (Alano section, NE Italy), Mar.
1465 Micropaleontol., 81, 22-31, doi:10.1016/j.marmicro.2011.07.002, 2011.
1466 Toffanin, F., Agnini, C., Rio, D., Acton, G., and Westerhold, T.: Middle Eocene to early Oligocene
1467 calcareous nannofossil biostratigraphy at IODP Site U1333 (equatorial Pacific),
1468 Micropaleontol., 59, 69-82, 2013.
1469 Tremolada, F. and Bralower, T. J.: Nannofossil assemblage fluctuations during the Paleocene–
1470 Eocene Thermal Maximum at Sites 213 (Indian Ocean) and 401 (North Atlantic Ocean):
1471 palaeoceanographic implications, Mar. Micropaleontol., 52, 107-116,
1472 doi:10.1016/j.marmicro.2004.04.002, 2004.
1473 Vandenberghe, N., Hilgen, F. J., and Speijer, R. P.: The Paleogene Period, in: The Geologic Time Scale
1474 2012, Amsterdam, the Netherlands (Elsevier BV), 855–922, 2012.
1475 Watkins, D. K. and Self-Trail, J. M.: Calcareous nannofossil evidence for the existence of the Gulf
1476 Stream during the late Maastrichtian, Paleoceanography, 20, Pa3006,
1477 doi:10.1029/2004pa001121, 1992.
1478 Wei, W., and Wise Jr. S. W.: Biogeographic gradients of middle Eocene–Oligocene calcareous
1479 nannoplankton in the South Atlantic Ocean, Palaeogeogr. Palaeoclimatol. Palaeoecol., 79, 29-
1480 61, 1990.
1481 Westerhold, T., Röhl, U., Raffi, I., Fornaciari, E., Monechi, S., Reale, V., Bowles, J., and Evans, H. F.:
1482 Astronomical calibration of the Paleocene time, Palaeogeogr. Palaeoclimatol. Palaeoecol.,
1483 257, 377-403, doi:10.1016/j.palaeo.2007.09.016, 2008.
1484 Westerhold, T., Röhl, U., Donner, B., McCarren, H.K. and Zachos, J.C.: A complete high - resolution
1485 Paleocene benthic stable isotope record for the central Pacific (ODP Site 1209).
1486 Paleoceanography, 26, PA2216,doi:10.1029/2010PA002092, 2011
1487 Winter, A., Jordan, R.W., and Roth, P.H.: Biogeography of living coccolithophores. in:
1488 Coccolithophores edited by Winter, A., and Siesser, W.G., Cambridge Univ. Press, Cambridge
1489 , 161-177, 1994.
1490 Winterer, E. L. and Bosellini, A.: Subsidence and sedimentation on Jurassic passive continental
1491 margin, Southern Alps, Italy, Am. Assoc. Petr. Geol. Bull., 65, 394-421, 1981.
1492 Young, J. R., Geisen, M., and Probert, I.: A review of selected aspects of coccolithophore biology with
1493 implications for paleobiodiversity estimation, Micropaleontology, 51, 267-288, 2005.,
1494 Zachos, J.C., Kroon, D., Blum, P., et al.: Early Cenozoic Extreme Climates: The Walvis Ridge Transect.
1495 Proc. Ocean Drill. Program, Initial Rep. 208 doi:10.2973/odp.proc.ir.208.2004, 2004.
1496 Zachos, J. C., Röhl, U., Schellenberg, S. A., Sluijs, A., Hodell, D. A., Kelly, D. C., Thomas, E., Nicolo, M.,
1497 Raffi, I., Lourens, L. J., McCarren, H., and Kroon, D.: Rapid acidification of the ocean during the
1498 Paleocene-Eocene Thermal Maximum, Science 308, 1611-1615,
1499 doi:10.1126/science.1109004, 2005.
1500 Zachos, J. C., Dickens, G. R., and Zeebe, R. E.: An early Cenozoic perspective on greenhouse warming
1501 and carbon-cycle dynamics, Nature, 451, 279-283, doi:10.1038/nature06588, 2008.
1502 Zachos, J. C., McCarren, H., Murphy, B., Röhl, U., and Westerhold, T.: Tempo and scale of late
1503 Paleocene and early Eocene carbon isotope cycles: Implications for the origin of
1504 hyperthermals. Earth Planet. Sci. Lett., 299, 242-249, doi:10.1016/j.epsl.2010.09.004, 2010.
1505 Zattin, M., Cuman, A., Fantoni, R., Martin, S., Scotti, P., and Stefani, C.: From middle Jurassic heating
1506 to Neogene cooling: The thermochronological evolution of the southern Alps, Tectonophysics,
1507 414, 191-202, doi:10.1016/j.tecto.2005.10.020, 2006.

1508   Zeebe, R. E. and Westbroek, P.: A simple model for the $CaCO_3$ saturation state of the ocean: The
1509   "Strangelove", the "Neritan", and the "Cretan" ocean, Geochem. Geophys. Geosyst., 4, 1104,
1510   doi:10.1029/2003GC000538, 2003, 2003.
1511   Zeebe, R. E., Zachos, J. C., and Dickens, G. R.: Carbon dioxide forcing alone insufficient to explain
1512   Paleocene-Eocene Thermal Maximum warming, Nature Geosci., 2, 576-580,
1513   doi:10.1038/NGEO578, 2009.
1514   Ziveri, P., Young, J., and van Hinte J. E.: Coccolithophore export production and accumulation rates.
1515   in: On determination of sediment accumulation rates, GeoResearch Forum, Trans Tech
1516   Publications LTD, Switzerland, 5, 41-56, 1999.