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Variability in climate and productivity during the Paleocene–Eocene Thermal Maximum in the western Tethys (Forada section)

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Abstract. The Forada section (northeastern Italy) provides a continuous, expanded deep-sea record of the Paleocene–Eocene Thermal Maximum (PETM) in the central-western Tethys. We combine a new, high-resolution, benthic foraminiferal assemblage record with published calcareous plankton, mineralogical and biomarker data to document climatic and environmental changes across the PETM, highlighting the benthic foraminiferal extinction event (BEE). The onset of the PETM, occurring  esposkky after a precursor event, is marked by a thin, black, barren clay layer, possibly representing a brief pulse of anoxia and carbonate dissolution. The BEE occurred within the 10 cm interval including this layer. During the first 3.5 kyr of the PETM, several agglutinated recolonizing taxa show rapid species turnover, indicating a highly unstable, CaCO₃-corrosive environment. Calcareous taxa reappeared after this interval, and the next 9 kyr were characterized by rapid alternation of peaks in abundance of various calcareous and agglutinated recolonizers. These observations suggest that synergistic stressors, including deepwater CaCO₃ corrosiveness, low oxygenation, and high environmental instability caused the extinction. Combined faunal and biomarker data (BIT index, higher plant n-alkane average chain length) and the high abundance of the mineral chlorite suggest that erosion and weathering increased strongly at the onset of the PETM, due to an overall wet climate with invigorated hydrological cycle, which led to storm flood events carrying massive sediment discharge into the Belluno Basin. This interval was followed by the core of the PETM, characterized by four precessionally paced cycles in CaCO₃ %, hematite %, δ¹³C, abundant occurrence of opportunistic benthic foraminiferal taxa, and calcareous nanofossil and planktonic foraminiferal taxa typical of high-productivity environments, radiolarians, and lower δD₇₇-alkanes. We interpret these cycles as reflecting alternation between an overall arid climate, characterized by strong winds and intense upwelling, and an overall humid climate, with abundant rains and high sediment delivery (including refractory organic carbon) from land. Precessionally paced marl–limestone couplets occur throughout the recovery interval of the carbon isotope excursion (CIE) and up to 10 m above it, suggesting that these wet–dry cycles persisted, though at declining intensity, after the peak PETM. Enhanced climate extremes at mid-latitudes might have been a direct response to the massive CO₂ input in the ocean atmosphere system at the Paleocene–Eocene transition, and may have had a primary role in restoring the Earth system to steady state.

1 Introduction

The Paleocene–Eocene Thermal Maximum (PETM) has over the last 24 years attracted intensive study by the scientific community as one of the most dramatic and rapid climatic disruptions of the Cenozoic (e.g., Kennett and Stott, 1991; Zachos et al., 2001; Sluijs et al., 2007a; McInerney and Wing, 2011; Littler et al., 2014). During the PETM (~55.6 Ma), the Earth’s surface temperature increased by ~5 °C in a few thousand years (McInerney and Wing, 2011;
Dunkley-Jones et al., 2013; Zeebe et al., 2014; Bowen et al., 2015), and remained high for 100 to 170–200 kyr (e.g., Röhl et al., 2007; Giusberti et al., 2007; Murphy et al., 2010). The PETM is recognized in terrestrial and marine settings by a negative carbon isotope excursion (CIE; e.g., Kennett and Stott, 1991; Bowen et al., 2004), with variable magnitude ranging from ~2–4.5‰ in marine carbonates (e.g., Thomas and Shackleton, 1996; Bains et al., 1999; Thomas et al., 2002; Zachos et al., 2005; Handley et al., 2008; McCarren et al., 2008) to 4–7‰ in marine and terrestrial organic carbon and leaf waxes (e.g., Kaiko et al., 1996; Bowen et al., 2004, 2015; Pagani et al., 2006a; Smith et al., 2007; Handley et al., 2008; McCarren et al., 2008). This CIE is attributed to a massive, rapid input of isotopically light carbon into the ocean–atmosphere system which destabilized the global carbon cycle and led to rapid and extreme global warming (e.g., Dickens et al., 1997; Thomas and Shackleton, 1996; Pagani et al., 2006b; Panchuk et al., 2008; Dickens, 2011; DeConto et al., 2012). Both the source(s) of the carbon and the triggering mechanism(s) of the emissions are still strongly debated (e.g., Meissner et al., 2014), in part because the pattern and size of the CIE does not necessarily simply reflect the size and isotopic signature of the carbon input, but is affected by biotic and sedimentary processes (e.g., Kirtland Turner and Ridgwell, 2013). Despite these debates, the onset of the CIE is an outstanding global correlation tool (McInerney and Wing, 2011; Stassen et al., 2012b), formally used to define the base of the Eocene (Aubry et al., 2007).

The carbon cycle perturbation of the PETM led to acidification of surface ocean waters (Penman et al., 2014) and severe shallowing of the calcite compensation depth (CCD; Zachos et al., 2005; Kelly et al., 2010; Hönsch et al., 2012). Widespread carbonate dissolution coincided with the base of the CIE (e.g., Thomas and Shackleton, 1996; Thomas, 1998; Hancock and Dickens, 2005; McCarren et al., 2008). The paleoceanographic changes affected primary and export productivity (e.g., Thomas, 2007; Winguth et al., 2012; Ma et al., 2014), which in general increased in marginal basins and along continental margins but decreased in open oceans (e.g., Gibbs et al., 2006; Stoll et al., 2007; Speijer et al., 2012). The higher ocean temperatures may have led to increased remineralization of organic matter in the oceans due to increased metabolic rates (John et al., 2013, 2014; Boscolo Galazzo et al., 2014; Ma et al., 2014). The combination of increased remineralization, higher temperatures and increased ocean stratification led to a decrease in oxygen levels in bottom waters regionally, especially along continental margins (including the Arctic Ocean) and in the Atlantic Ocean (e.g., Benjamin, 1992; Speijer et al., 1997, 2012; Gavrilov et al., 1997; Thomas, 2007; Chun et al., 2010; Winguth et al., 2012; Nagy et al., 2013; Wieczorek et al., 2013; Dickson et al., 2014; Pälike et al., 2014; Post et al., 2016), while oxygen minimum zones in open oceans expanded globally (Zhou et al., 2014), including at Forada (Luciani et al., 2007). The increased primary productivity in marginal basins has been linked to increased influx of nutrients from the continents, caused by increased erosion and weathering due to intensification of the hydrological cycle, because precipitation is correlated to globally averaged surface temperatures (e.g., Pierrehumbert, 2002). A widespread increase in kaolinite in PETM sediments has been related to the global increase in precipitation and intensity of chemical weathering (e.g., Robert and Chamley, 1991; Robert and Kennett, 1994; Kaiko et al., 1996; Gibson et al., 2000), as also suggested by Os-isotope evidence (Ravizza et al., 2001; Wieczorek et al., 2013). However, reconstruction of hydrological changes from clay mineral assemblages is complex, and additional evidence is needed (Thiry, 2000; Schmitz and Pujalte, 2003, 2007; Egger et al., 2003, 2005; Handley et al., 2012).

The severe climatic perturbations of the PETM profoundly affected terrestrial and marine ecosystems, triggering faunal and floral radiations and migrations (e.g., Kelly et al., 1996; Bralower, 2002; Gingerich, 2003; Wing et al., 2005; Slijthere et al., 2007a; Jaramillo et al., 2010; McInerney and Wing, 2011). Deep-sea benthic foraminifera experienced the most severe extinction of the Cenozoic, the benthic foraminiferal extinction event (BEE) (Thomas, 1989, 1990, 1998; Kennett and Stott, 1991; Thomas and Shackleton, 1996; Alegret et al., 2009a, b; 2010). The BEE was rapid (<10 kyr; Thomas, 1989, 2003, 2007) and wiped out the Cretaceous bathyal and abyssal “Velasco-type fauna” (Berggren and Aubert, 1975; Tjalsma and Lohmann, 1983; Thomas, 1998, 2007), marking a significant step towards the establishment of modern benthic foraminiferal fauna (Thomas, 2007). The extinction was far less severe in shelf environments (Gibson et al., 1993; Speijer et al., 2012; Stassen et al., 2015).

The cause of this global extinction remains under debate, because neither anoxia nor higher or lower productivity nor carbonate dissolution occurred globally at bathyal to abyssal depths in the deep sea, the largest habitat on Earth (e.g., Thomas, 2003, 2007; Alegret et al., 2010), and benthic foraminifera are highly efficient dispersers (Alve and Goldstein, 2003). The link between the environmental changes during the PETM and the benthic foraminiferal extinction event thus remains poorly understood. A common obstacle to perform detailed high-resolution studies of the PETM in deep-sea sediments is the fact that many records are condensed or discontinuous, especially across the few thousand years (Zeebe et al., 2014) of the onset of the carbon isotope excursion. The Forada section (northeastern Italy) represents an outstanding exception in that it contains an expanded deep-sea record of the PETM, which has been extensively studied because of its continuity and cyclostratigraphy (Agnini et al., 2007; Giusberti et al., 2007; Luciani et al., 2007; Tipple et al., 2011; Dallanave et al., 2012). Carbonate dissolution is less severe at Forada than in many other sections, with calcareous benthic foraminifera present for most of the interval characterized by the CIE (>4 m; Giusberti et al., 2007). Given the limited number globally of complete
and expanded deep-sea PETM sections, the Forada section represents an invaluable opportunity to investigate the environmental impacts of the PETM and repercussions on deep-sea fauna.

We provide a high-resolution benthic foraminiferal record for the Forada section in order to reconstruct the progression (tempo and mode) of environmental and biotic changes during the PETM. These data allow us to reconstruct the environmental disruption and the benthic foraminiferal response to PETM warming in detail and document the community recovery. Benthic foraminiferal data are integrated with sedimentological and geochemical data (Giusberti et al., 2007; Tipple et al., 2011), and data on calcareous plankton communities (Agnini et al., 2007; Luciani et al., 2007), providing perhaps the most complete reconstruction across the PETM in Europe to date.

We pay homage to research by the Italian researchers (Di Napoli Alliata et al., 1970; Braga et al., 1975) who first described the benthic foraminiferal turnover across the Paleocene–Eocene transition in Italy.

2 Materials and methods

2.1 The Forada section

The Forada section (46.036083° N, 12.063975° E) is exposed along the Forada creek, ~2 km east of the village of Lentiai (Fig. 1) in the Venetian Prealps (NE Italy). It consists of ca. 62 m of Scaglia Rossa, pink-reddish limestones and marly limestones, locally rhythmically bedded, and encompassing the Upper Cretaceous through the lower Eocene (Fornaciari et al., 2007; Giusberti et al., 2007). The upper Paleocene–lower Eocene succession is interrupted by the clay marl unit (CMU; Giusberti et al., 2007), which marks the PETM and correlates with clay-rich units on other continental margins (e.g., Schmitz et al., 2001; Crouch et al., 2003; John et al., 2008; Nicolo et al., 2010). The investigated interval has been subdivided into four sub-intervals based on the δ13C record in bulk rock (Giusberti et al., 2007). From bottom to top, these are the pre-CIE, the main CIE, the CIE recovery and the post-CIE (Fig. 2). The main CIE (Giusberti et al., 2007; Figs. 2, 3) occurs in the >3 m thick CMU, within which are recorded the short-lived occurrences of the calcareous plankton “excursion taxa” (Kelly et al., 1996, 1998) and the BEE (Agnini et al., 2007; Giusberti et al., 2007; Luciani et al., 2007). Sedimentation rates in the CMU were 5 times higher than in the upper Paleocene, indicating increased continental weathering and run off, which led to increased sediment influx in the Belluno Basin (Giusberti et al., 2007).

2.2 Benthic foraminifera

Benthic foraminiferal assemblages were studied in 54 samples from the same set studied by Luciani et al. (2007) across an ~11 m thick interval straddling the PETM (~467 to +591.5 cm; Fig. 2), which reflects ~800 kyr (Giusberti et al., 2007). In this study the planktic foraminifera fragmentation index (F index) of Luciani et al. (2007) is used as a proxy for dissolution (Figs. 2, 3) (Hancock and Dickens, 2005). The sample spacing for benthic foraminiferal assemblage analysis was determined based on biostratigraphic and cyclostratigraphic data (Agnini et al., 2007; Giusberti et al., 2007; Luciani et al., 2007). A sampling interval of 3–5 cm was used across the onset of the CIE (~42.5 to +50 cm interval), a 25 cm sample interval over the main CIE (from +75 to 335 cm). Below ~42.5 cm and above 335 cm we adopted a spacing between 20 and 50 cm. Samples were collected excluding, to the extent possible, bioturbated material. Further selection and removal of bioturbated material was carried out in the laboratory before sample processing. Data previously collected from the Forada section indicate that significant bioturbation effects are not present (e.g., Agnini et al., 2007; Giusberti et al., 2007; Luciani et al., 2007).

Foraminifera were extracted from the indurated marls and limestones using the “cold acetolysis” technique of Lirer (2000), following Luciani et al. (2007). Soft marly and clayey samples (mostly from the CMU interval) were disaggregated using a 10–30 % solution of hydrogen peroxide. The samples with the lowest content of CaCO3 (e.g., clays of basal CMU) were treated with diluted hydrogen peroxide (10 %) in order to prevent possible additional breakage of tests (especially of planktic foraminifera). For more details on the comparison between the two methods of preparation (cold acetolysis versus hydrogen peroxide), we refer the reader to Luciani et al. (2007).
Figure 2. Faunal and geochemical variations across the PETM at the Forada section plotted against chronostratigraphy, precessional cycles, lithology, recognized benthic foraminiferal assemblages (A to F) and isotopic intervals. % agglutinated: agglutinated to agglutinated and calcareous hyaline ratio; % infaunal taxa: infaunal to infaunal and epifaunal ratio; simple diversity and Fisher-α diversity index; N g⁻¹ = number of benthic foraminifera per gram (faunal density) in the > 63 mm size fraction; coarse fraction (CF) calculated according to Hancock and Dickens (2005) as the weight percent of the > 63 µm size fraction relative to the weight of the bulk sample; fragmentation index (F index) is from Luciani et al. (2007). The gray bands indicate intervals of carbonate dissolution. α: pre-CIE dissolution, β: burndown layer; BFDI: benthic foraminiferal dissolution interval.

Figure 3. Summary of the main mineralogical, geochemical and cyclostratigraphic features recognized across the Paleocene–Eocene boundary and in the clay marl unit (CMU) of the Forada section and radiolarian abundance plotted against isotopic intervals and recognized benthic foraminiferal assemblages (A to F). N g⁻¹ for the radiolarians refers to the number of radiolarians (> 125 µm fraction) per gram of dry sediment. F index from Luciani et al. (2007). VPDB – Vienna Pee Dee belemnite standard. Modified from Giusberti et al. (2007).

The quantitative study of benthic foraminifera was based on representative splits (using a Jones-type micro-splitter, Geneq Inc.) of approximately 200–400 individuals > 63 and < 500 µm (Table S1 in Supplement). The use of the small size fraction is time-consuming and presents difficulties in taxonomic determination, but we preferred to avoid the loss of small taxa, which are important for paleoecological investigations (e.g., Thomas 1985; Boscolo Galazzo et al., 2013,
2015), especially directly after the BEE, when small species are dominant (Thomas, 1998; Foster et al., 2013). Between 0 and −222 cm (uppermost Paleocene), the fraction ≥125 µm of at least one-fourth of the residue was carefully scanned for large specimens of the extinction taxa, here labeled “cosmopolitan extinction taxa” (CET) (see Thomas, 1998, 2003). These CET records have been treated qualitatively (Fig. 4). The extinction taxa include Anomalinoïdès rubiginosus, Angulogavelinella avnimelechi, Aragonia velascoensis, Bolivinoides delicatus, Cibicidoides dayi, C. hyphalus, C. velascoensis, Clavulina amorpha, Clavulinoides trilaterra, Clavulinoides globulifera, Coryphostoma midwayensis, Dorothea beloides, D. bulbetta, D. pupa, D. retusa, Neoeponides megastoma, Gavelinella beccariiformis, Gyroidinoides globosus, G. quadratus, Marsonella indentata, Neoflabellina jarvisi, N. semireticulata, Nuttallinella florealis, Osangularia velascoensis, Paralabamina hillebrandti, Pullenia coryelli, and Remeëella varians (e.g., Beckmann, 1960; Von Hillebrandt, 1962; Tjalsma and Lohmann, 1983; Speijer et al., 1996; Thomas, 1998), each of which is present at Forada.

We identified most common taxa at the species level (Table S2). Taxa with high morphological variability and/or variable preservation were identified at generic or higher taxonomic level. Specimens of the most representative taxa were imaged using the SEM at the C.U.G.A.S. (Centro Universitario Grandi Apparecchiature Scientifiche) of Padova University (Plates 1–4). Relative abundances of the taxa and taxon groups, along with faunal indices such as the calcareous/agglutinated ratio, the infaunal/epifaunal ratio, and bi-triserial percentage, were calculated (Figs. 2, 5–7 and S1). The absolute abundance (N g⁻¹: number of benthic foraminifera per gram of bulk dried sediment) was calculated for both the ≥63 and ≥500 µm fractions. Faunal diversity indices (species diversity and Fisher-α; Fig. 2) were calculated using the PAST package (Hammer et al., 2001). Segments belonging to tubular/branched agglutinated forms (e.g., Rhizaminina, Rhabdammina, Bathysiphon) were counted, but they were excluded from calculations because there is no reliable method to convert the abundance of multiple fragments into that of single individuals (Ernst et al., 2006).

We assigned species to epifaunal and infaunal morphotypes by comparing their test morphology to the morphotypes in Corliss (1985), Jones and Charnock (1985), Corliss and Chen (1988), Kaminski and Gradstein (2005), Hayward et al. (2012), and Mancin et al. (2013). However, caution is needed in applying taxonomic uniformitarianism due to our limited knowledge of the biology and ecology of the highly diverse living species. Even for many living species, the relation between test morphology and microhabitat has not been directly observed but is extrapolated from data on other taxa (e.g., Jorissen, 1999). The assignment of modern foraminifera to microhabitats based on their morphology may be accurate in only 75% of cases (Ernst et al., 1993): comparisons between past and recent environments thus need careful evaluation as well as cross correlation between benthic foraminiferal and other proxy data. The ecology as evaluated from the literature (Table 1) is shown for selected benthic foraminiferal taxa from the PETM interval at Forada.
The ecological preferences of selected benthic foraminifera, as evaluated from the literature, are summarized in Table 1.

Table 1: Summary of the known ecological preferences of selected benthic foraminifera, as derived from the literature. Common in Pleistocene.

<table>
<thead>
<tr>
<th>Foraminifera Species</th>
<th>Ecological Preferences</th>
</tr>
</thead>
<tbody>
<tr>
<td>Spiroplectammina navarroana</td>
<td>Typically found in oligotrophic environments, blooming during hyperthermal events.</td>
</tr>
<tr>
<td>Oridorsalis umbonatus</td>
<td>Common in high-productivity continental margins.</td>
</tr>
<tr>
<td>Haplophragmoides spp.</td>
<td>Abundant in anoxic OAE2 black shales in abyssal sections.</td>
</tr>
<tr>
<td>Osangularia spp.</td>
<td>Opportunistic species, repopulating the seafloor during short-term re-oxygenation phases.</td>
</tr>
<tr>
<td>Eobigenerina variabilis</td>
<td>Found in high-productivity areas worldwide.</td>
</tr>
<tr>
<td>Karrerulina conversa</td>
<td>Resistant to carbonate dissolution, common in OAE2 black shales.</td>
</tr>
</tbody>
</table>

Further references:
- Gooday et al. (2008), Kuhnt and Kaminski (1993)
2.3 Age model

The age model used for calculating the longevity of benthic foraminiferal assemblages (see below) follows Luciani et al. (2007), with the lower Eocene chronology based on the cyclostratigraphic age model of Giusberti et al. (2007; Fig. 3). The duration of each precessional cycle has been assumed to be 21 kyr. Sedimentological and geochemical parameters oscillate cyclically within the main CIE, in at least five complete precessional cycles (Figs. 2, 3). The CIE recovery interval is composed of six distinct, precessional marl–limestone couplet cycles (Fig. 3). The recognition of 11 cycles in the combined CIE and recovery interval implies an estimate of the total duration of the CIE of ca. 230 kyr (Fig. 3). Giusberti et al. (2007) and Röhl et al. (2007) disagree on the duration of the main CIE and recovery interval (179 ± 17 and 231 ± 22 kyr, respectively). The main difference between these two chronologies is the assignment of different numbers of precessional cycles within the main body and recovery interval (Tippel et al., 2011). A 3He-based chronology for Site 1266 (Walvis Ridge) suggests a total PETM duration of 234 ± 48/−34 kyr (Murphy et al., 2010), in line with the age model of Giusberti et al. (2007).

Lithological cycles have not been firmly identified in the Paleocene part of the section, and sedimentation rates are interpolated between the base of the PETM at ±0 cm and the lowest occurrence of the calcareous nannofossil Discoaster multiradiatus at ca. −12.5 m (Giusberti et al., 2007), using a duration of the time between these events of 1.238 Myr (Westerhold et al., 2007). In this age model, the investigated portion of the Forada section spans ca. 800 kyr.

3 Results

Benthic foraminiferal assemblages are generally dominated by calcareous hyaline taxa (85–90 %; Fig. 2), but agglutinated taxa significantly increase in abundance within the CMU (25–90 %; Fig. 2). Infaunal taxa strongly dominate the assemblage throughout the studied interval (~ 80 %). Faunal diversity is fairly high, particularly in the upper Paleocene (Fig. 2), and preservation is generally moderate, though poor within the lowermost centimeters of the Eocene. Most foraminiferal tests at Forada are recrystallized and either totally or partially filled with calcite.

Composition and abundance of the assemblages change prominently across the ca. 11 m thick interval investigated (Figs. 2, 5–7) coeval with the geochemical signature of the PETM, and broadly coincident with the main lithological changes. We recognized six successive benthic foraminiferal assemblages (labeled A to F; Figs. 2, 5–8), mainly based on changes in abundance of the taxa listed in Table 1. Assemblages A and B are characteristic of the dominantly reddish calcareous marls mottled by greenish “flames” of the uppermost Paleocene, separated by the thin, barren clay layer from assemblages C, D and E, which occur in the first half of the main excursion of the CIE (lowermost Eocene), within the CMU (basal green laminated clays overlain by mottled reddish clays, marly clays and marls). Assemblage F characterizes the marls of the upper half of the CMU, as well as the CIE recovery interval and the overlying post-exursion interval of reddish limestone–marl couplets (Giusberti et al., 2007).

3.1 Assemblage A: the upper Paleocene fauna

Assemblage A (~467.5 to −37.5 cm, with an estimated duration > 430 kyr) has a high diversity, with abundant infaunal taxa (ca. 70–80 %; Fig. 2). Small bolivinids (< 125 µm) of the Bolivinoides crenulata group (Plate 3, Figs. 7–9) and smooth-walled Bolivina spp. together comprise 50–60 % of the > 63 µm fauna (Fig. 5), with Siphogenerinoides brevispinosa (~ 10 %) and other bulimulids less common (Figs. 5, 6). Epifaunal morphotypes are mainly represented by small cibicides (10 %), Anomalolinoides spp. (5 %) and Cibicidoides spp. (usually < 5 %; Fig. 5). Rare taxa include reussellids, angulogerinids, nodosariids, dentalinids, gyroinids, valvabaminids and unilocular hyaline taxa (Fig. S1 in Supplement). Agglutinated taxa are mainly represented by Spiroplectammina spectabilis, Trochamminoides spp., Paratrochamminoides spp., Reophax spp. and Subreophax spp. The Paleocene CET (Plate 1) are not a major component of the assemblage > 63 µm (< 10 %; Fig. 6), but are common to abundant in the size fraction > 125 µm (> 20 %). Many of these have large, heavily calcified tests. The most common taxa include Gavelinella beccariiformis, Pullenia coryelli and Coryphostoma midwayensis (Table S1). CET such as Clavulinoides globulifera, Cibicidoides dayi and Cibicidoides velascoensis are common in the > 500 µm size fraction, together with trochamminids and large lituolids (Plate 1, Fig. 19, 6–8; Plate 4, Figs. 7, 8, 14, 20). The latter occur up to the top of the Paleocene, but are absent in the Eocene. At ~261.5 cm, the CET peak at 15 %, their maximum abundance in the studied section (Fig. 6). At the same level, peaks of large, stout, heavily calcified taxa (e.g., Cibicidoides and anomalolinids) co-occur with agglutinated taxa (Glomospira, Spiroplectammina and Haplophragmoides; Figs. 6, 7), whereas small, thin-walled forms such as bolivinids, Siphogenerinoides brevispinosa and cibicides decline markedly in relative abundance (Figs. 5–7). Faunal density (N g⁻¹), diversity and the percentage abundance of infaunal morphotypes decrease (Fig. 2), as do δ¹³C and CaCO₃ %, whereas the planktonic foraminiferal fragmentation index (F index) increases significantly (Fig. 2). The upper boundary of this assemblage is defined by the increase in abundance of the opportunistic taxa Tappanina selmen-sis and Siphogenerinoides brevispinosa, marking the onset of Assemblage B.
3.2 Assemblage B: the pre-CIE Paleocene fauna

Assemblage B occurs at -31 to 0 cm, estimated duration ~34 kyr. At about -20 cm the lithology shifts from reddish to greenish marls with Zoophycos and Chondrites (intervals Pa I and II of Giusberti et al., 2007). In this assemblage, Siphogenerinoides brevispinosa and Tappanina sel.

mensis increase in relative abundance compared to Assemblage A (> 10% at ~ -27 and -12 cm; Figs. 6, 7). Between the two peaks of S. brevispinosa (at about ~ -20 cm; Figs. 6, 7), there is a transient negative carbon isotope excursion of about 1‰, a drop in CaCO$_3$ from 60 to 40%, a decline in the coarse fraction to 2%, and a peak in the F index (85-90%; Figs. 2, 3). Small and thin-walled taxa such as bolinids, cibicids and S. brevispinosa decreases markedly in relative abundance, whereas big, heavily calcified taxa (e.g., Cosmopolitan Extinction Taxa, Cibicidoides spp., Natallides truempyi) and agglutinated forms increase (Figs. 5–7). In addition, faunal density drops, as does the percentage of infaunal taxa (from 90 to 50%), and diversity increases (Fig. 2). From -4.5 cm upwards, the preservation of benthic foraminifera deteriorates, while the F index reaches 100% (Figs. 2, 3). At -1.5 cm preservation worsens and most bistriral taxa decline in abundance drastically, whereas benthic foraminiferal absolute abundance and CaCO$_3$ % both decrease (Fig. 2). Faunal diversity peaks, while anomalinids, Cibicidoides spp., N. truempyi, O. umbonatus, and agglutinated forms increase markedly in relative abundance (Figs. 2, 5, 6). In the uppermost Paleocene sample, we see the highest occurrence of most CET (Figs. 4, 6). Few CET (e.g., Aragonia velascoensis) disappear below this sample (Fig. 4). These are generally rare, occurring discontinuously throughout the Paleocene, even in large samples of residue > 125 µm (Fig. 4). The uppermost occurrence of the CET defines the upper boundary of this assemblage, at the base of the black clay layer (Figs. 4, 6).

3.3 The black clay

The lowermost Eocene is a thin, black clay layer (0 to +0.3 cm), slightly enriched in organic carbon, and carbonate-free (Giusberti et al., 2007; Figs. 3, 8). This clay marks the base of the CMU, and contains a few specimens only, agglutinated benthic foraminifera of the genera Haplophragmoides and Recurvoides (10 specimens in 22 g washed sediment). It probably was deposited over less than a millennium, in view of its small thickness and place within the precessationally paced cycles in the PETM.

3.4 Assemblage C: basal CIE agglutinated fauna

We label BFDI (benthic foraminiferal dissolution interval) the lowermost Eocene interval (lowermost 10 cm of laminated green clays of CMU; estimated duration 3.5 kyr) with low CaCO$_3$ wt% (~15%), and the most negative δ$^{13}$C values in bulk carbonate (~2‰). Assemblage C is dominated by agglutinated taxa (about 90%; Fig. 2) with badly preserved and deformed tests. Tests of calcareous-hyaline forms are rare, partially dissolved and fragmented. Assemblage C has minimum values of faunal density (<5), diversity, and wt% coarse fraction (Fig. 2). Infaunal morphotypes have their lowest abundance (ca. 36% ; Figs. 2, 6). Agglutinated
3.5 Assemblage D: lowermost CIE fauna

In Assemblage D (+10 to +35 cm, lithologically characterized by laminated green clays; estimated duration ~9 kyr), calcareous-hyaline forms are consistently present and badly preserved, with dominant taxa having dwarfed and thin-walled tests, e.g., Globocassidulina subglobosa (25 %), Tappendina selmensis (20 %), and Osangularia spp. (~11 %; Figs. 6, 7; Plate 2, Figs. 13–16). A specific assignment of basal PETM osangulariids at Forada is not possible because of their very small size and poor state of preservation. From +30 cm upwards, relative abundances of G. subglobosa and Osangularia spp. drastically decline, whereas T. selmensis reaches its maximum abundance (ca. 33 %; Figs. 6, 7). Minor components are the “other buliminids” group (up to 10 % at the top of the assemblage; see Figs. 5 and 5 – related caption), Pleurostomella spp., Oridorsalis umbonatus, anomalinids and stilostomellids (Figs. 5, 6 and S1). Agglutinated forms remain abundant, up to 50 %. At +20 cm, calcified radiolarians become abundant, dominating the microfossil association up to +2 m above the base of CMU (Luciani et al., 2007; Figs. 3, 8). Within the interval of Assemblage D, δ¹³C shifts from ~2 to ~1 ‰, and the CaCO₃ wt % recovers to ~40 %, despite strong dilution with terrigenous sediments (Fig. 3). The upper boundary of this assemblage is defined by the consistent decrease in T. selmensis (to <5 %).

3.6 Assemblage E: main CIE fauna I

In this interval (+35 to +185 cm; lithologically characterized by green and reddish clays and marls; estimated duration ca. 42 kyr) benthic foraminiferal preservation improves, and calcareous-hyaline forms dominate the assemblages again (Fig. 2). Siphogenerinoides brevispinosa is consistently present again, with two peaks up to 20 % (Figs. 6, 7). Pleurostomella spp. increase to up to >10 %, and Bolivinoides crenulata and smooth-walled Bolivina spp. to up to 30–40 % (Figs. 5, 6). Calcareous-hyaline epifaunals such as cibicids and anomalinids reappear at <5 % (Fig. 5). Faunal density and diversity gradually increase upwards, whereas agglutinated taxa markedly decrease in abundance (<20 %) at ~30–70 cm (Fig. 2). The upper boundary of this assemblage is defined by the marked drop in relative abundance of S. brevispinosa (to <5 %).
3.7 Assemblage F: main CIE fauna II, CIE recovery and post CIE fauna

Assemblage F characterizes the upper half of the CMU (reddish marls), from about +185 cm up to its top (+337.5 cm), and the overlying interval (red marly limestone couplets) up to +649 cm; estimated total duration > 281 kyr). The relative abundance of *Siphogenerinoides brevispinosa* is low (<5 %), whereas *Bulimina taxpamensis* and *Nuttallides truempyi* increase in abundance, respectively to 5 and 10 %, and show cyclical variations in relative abundance (Figs. 6, 7). Pleurostomellids (~10 %), the “other buliminids” group (~10 %; Fig. 5), cibicids (~10 %), *Oridorsalis umbonatus* (~5 %), stilostomellids (~5 %) and *Abyssammina* spp. (~5 %) are common (Figs. 5, 6). Relative abundance of infaunal taxa (mostly bolivinids) and faunal density (N g⁻¹) returns to Paleocene values (75–80 %; Fig. 2). Diversity increases (simple diversity up to 60, Fisher-α diversity up to 20; Fig. 2) but remains lower than in the Paleocene. All faunal indices show cyclical variation (Fig. 2), as do the relative abundance of benthic foraminifera as well as planktic foraminiferal and calcareous nannofossil assemblages (Agnini et al., 2007; Luciani et al., 2007). In the lower third of the interval in which this assemblage occurs, just above the CMU (ca. +337.5 cm), the relative and absolute abundance of radiolarians decrease markedly and agglutinated taxa such as *Glomospira* spp., *Eobigenerina variabilis* and *Karrerulina* spp. slightly increase in relative abundance (~+2–3 %) (Figs. 2, 3, 6, 7).

4 Discussion

4.1 Paleodepth of the Forada section

Based on benthic foraminifera in the >125 µm size fraction, Giusberti et al. (2007) suggested a paleodepth between 600 and 1000 m for the Forada section. Our data on the >63 µm size fraction suggest a somewhat greater paleodepth, i.e., upper lower bathyal, between 1000 and 1500 m (van Morkhoven et al., 1986). Representatives of the bathyal and
Figure 7. Enlargement of the interval from −1 to +2 m across the Paleocene–Eocene boundary at Forada showing the relative abundance of selected benthic foraminifera plotted against biostratigraphy, precessional cycles, lithology, $\delta^{13}$C bulk record, recognized benthic foraminiferal assemblages (A to F) and isotopic intervals. Benthic foraminiferal biozonation after Berggren and Miller (1989). The gray bands indicate intervals of carbonate dissolution. $\alpha$: pre-CIE dissolution interval; $\beta$: burnout layer; BFDI: benthic foraminiferal dissolution interval.

Figure 8. Summary of main calcareous plankton (calcareous nannofossils and planktonic foraminifera) and benthic foraminiferal events and inferred environmental conditions (from Agnini et al., 2007; Luciani et al., 2007 and present work), isotopic intervals, thickness, precessional cycles and benthic foraminiferal assemblages (A to F) recognized in this work. The stratigraphic intervals containing assemblages A and B, C, and D to F are considered as pre-extinction, extinction and repopulation intervals, respectively. Benthic foraminiferal zonation after Berggren and Miller (1989). CNET: calcareous nannofossil excursion taxa (see Agnini et al., 2007).
Plate 2. SEM micrographs of the most representative species of the Eocene postextinction faunas occurring at Forada. 1. Ammobaculites agglutinans (BRI+10); 2. Eobigenerina variabilis (BRI+50); 3. Eobigenerina variabilis (BRI+50); 4. Karrerulina conversa (BRI+50); 5. Karrerulina hordida (BRI–25.5); 6. Spiroplectammina navarroana (BRI–33/7); 7. Spiroplectammina spectabilis (BRI+50); 8. Rashnovammina munda (BRI–50.5); 9. Haplophragmoides cf. kirki (BRI+5); 10. Saccammina placenta (BRI25.5); 11. Glomospira irregularis (BRI+50); 12. Glomospira charoides (BRI+75.5); 13. Osangularia sp. (BRI+15); 14. Globocassidulina subglobosa (BRI+15); 15. Tappanina selmensis (BRI+15); 16. Tappanina selmensis (BRI–9); 17. Siphogenerinoides brevispinosa (BRI–11); 18. Siphogenerinoides brevispinosa (BRI–365); 19. Bulimina taxpamensis (BRI+150); 20. Bulimina taxpamensis (BRI+150); 21. Pleurostomella sp. (BRI+150); 22. Bolivina sp. costate (BRI+385); 23. Nutallides truempyi (BRI+150); 24. Oridorsalis umbonatus (BRI–135.5); 25. Aragonia aragonensis (BRI–105); 26. Abyssammina pougi (TAL7B).


abyssal Velasco-type fauna (Berggren and Aubert, 1975), such as Aragonia velascoensis, Cibicoides velascoensis, Gyroidinoides globosus, Nutallides truempyi, Nutallinellina florealis, Osangularia velascoensis and Gavelinella beccariiformis are common at Forada. The faunas across the uppermost PETM interval and higher are similar to the PETM fauna in the upper abyssal Alamedilla section (southern Spain; Alegret et al., 2009a) and at Walvis Ridge at 1500 m paleodepth (Thomas and Shackleton, 1996; Thomas, 1998). Abyssammina spp. and Nutallides truempyi (upper depth limit at 1000 and 300 m, respectively; Van Morkhoven et al., 1986; Speijer and Schmitz, 1998) increase in abundance by more than a factor of 2 during the PETM at Forada, as typical for PETM deep-sea benthic foraminiferal records (e.g., Thomas, 1998, 2007; Thomas and Shackleton, 1996; Alegret...

et al., 2009a, 2010; Giusberti et al., 2009). In these deliberations we excluded the bolivinids, because we consider that their high abundance is due to the “delta-depression effect” (see below).

4.2 Environmental reconstruction during the late Paleocene

4.2.1 The Belluno Basin Paleocene deep-sea environment (Assemblage A)

Throughout most of the investigated section, infaunals strongly dominate over epifaunals, mainly due to the high abundances of bolivinids (Figs. 2, 5). Such dominance of bolivinids is common in lower and middle Eocene hemipelagic Scaglia sediments in the Belluno Basin (Agnini et al., 2009; Boscolo Galazzo et al., 2013). Presently, bolivinids are common along continental margins, and at bathyal depths, at the interception of the oxygen minimum zone (OMZ) with the seafloor, typically between 200 and 1000 m in modern oceans (Levin, 2003). High abundances of bolivinids commonly correlate with high organic matter flux and/or oxygen depletion (e.g., Murray, 1991; Gooday, 1994; Bernhard and Sen Gupta, 1999; Schmiedl et al., 2000; Thomas et al., 2000; Jorissen et al., 1995, 2007; Thomas, 2007). We see high abundances of such taxa typically at greater depths than usual in regions with significant organic matter input from rivers, the so-called “delta-depression effect” first described in the Gulf of Mexico (Pflum and Frerichs, 1976; Jorissen et al., 2007). Such lateral inputs of organic matter thus result in (partial) decoupling between the food supply to the benthos and local primary productivity (e.g., Fontanier et al., 2005; Arndt et al., 2013).

At Forada, there is neither geochemical nor sedimentological evidence for persistent suboxic conditions at the seafloor (Giusberti et al., 2007), and the high benthic foraminiferal faunal diversity likewise does not indicate low-oxygen conditions. The upper Paleocene calcareous plankton is dominated by morozovellids, indicating oligotrophic surface water conditions (Luciani et al., 2007; Fig. 8). The calcareous nannofossil assemblage is dominated by the generalist taxa *Toweius* and *Coccolithus*, with high percentages of *Sphenolithus* and *Fasciculithus* (Agnini et al., 2007; Fig. 8), supporting that surface waters were oligotrophic. We thus think that environments in the Belluno Basin, close to a continental margin (Agnini et al., 2007), were characterized by the delta-depression effect, in which hemipelagic sedimentation incorporated significant laterally transported terrigenous organic matter to serve as food for the benthos (e.g., Fontanier et al., 2005; Arndt et al., 2013).

The occurrence of large, epifaunal (> 500 µm) species (Assemblage A and B) has been related to an optimum food supply, but also to very low food supply, since a lack of food keeps individuals from reproducing successfully and leads to continued test growth (Boltovskoy et al., 1991; Thomas and Gooday, 1996).

Overall, Assemblage A indicates oligo-mesotrophic surface waters, with bolivinids probably exploiting refractory, laterally advected organic matter. The high faunal diversity suggests that seasonal to periodical increases in primary pro-
ductivity may have occurred (e.g., Gooday, 2003; Fontanier et al., 2006a, b, 2014), allowing a species-rich, highly diverse infauna and epifauna to inhabit the seafloor, and co-occur with the benthivivins in the sedimentary record.

At Forada, the relative abundance of Paleocene CET is low (average < 10%; Fig. 6), due to the large number of Bolivinacea dominating the fine size fraction used for this study (>63 µm). Many CET are epifaunal morphotypes, commonly larger than 125 µm, as also noted elsewhere (e.g., Giusberti et al., 2009). Similarly low percentages (12–15%) of CET have been recorded in Scaglia sediments of the Contessa section (Giusberti et al., 2009) and at ODP Site 690 by Thomas (2003), where infaunal morphotypes (buliminids and uniserial calcareous taxa) are abundant in the >63 µm fraction.

4.2.2 The precursor warming event (Assemblage B)

The onset of Assemblage B, about 34 kyr before the onset of the CIE (~−30 cm), is marked by increase in relative abundance of opportunistic taxa such as Tappanina selmensis and Siphogenerinoides brevispinosa (Figs. 6, 7; Table 1). The arrival of Tappanina selmensis, an upper bathyal to outer shelf species in the Maastrichtian (Frenzel, 2000), at greater depths might indicate warming of deep waters before the beginning of the PETM, as also reflected in the migration of warm-water planktonic species to high southern latitudes (Thomas and Shackleton, 1996; Table 1). The benthic foraminiferal changes roughly coincided with a significant increase in acarininids % (planktonic foraminifera, >50%), likely indicating warming of surface waters (Luciani et al., 2007; Fig. 8). The foraminiferal assemblages hence suggest warming throughout the water column and increased surface nutrient availability and deepwater food availability, whereas no changes in productivity in calcareous nannofossils are recorded (Agnini et al., 2007; Luciani et al., 2007; Fig. 8). The foraminiferal evidence for warming is associated with an increase in δD24alkanes and TEX86 values (Fig. 9), suggesting increased aridity and sea surface temperature prior to the onset of the CIE (Tipple et al., 2011).

Multiple proxies thus indicate that climatic and oceanographic conditions started to change ~30 kyr before the onset of the CIE, pointing to a PETM precursor event, reflected by a <5 cm thick dissolution interval at ~−22 cm, coinciding with a negative shift in bulk δ13C (~−1‰; Figs. 2, 3). Within this interval, dissolution-sensitive benthic foraminifera (e.g., S. brevispinosa and small bolivinids) markedly decrease in abundance, while more robust and agglutinated taxa increase (Figs. 2, 5–8), as does the F index of planktic foraminifera (to ~85–90%; Luciani et al., 2007; Fig. 3). This dissolution level may thus reflect a brief episode of rising lysocline/CCD (<5 kyr) in response to a precursor-ery of isotopically light carbon (Bowen et al., 2015). Similar precursor events have been observed worldwide (e.g., Sluijs et al., 2007b, 2011; Secord et al., 2010; Kraus et al., 2013; Garel et al., 2013; Bornemann et al., 2014; Bowen et al., 2015), indicating that disturbance of the global carbon cycle started before the PETM, as potentially also reflected in the occurrence of hyperthermals in the Paleocene (Thomas et al., 2000; Cramer et al., 2003; Coccioni et al., 2012).

At the top of Assemblage B (uppermost 4.5 cm), just prior to the onset of the CIE, carbonate preservation declined markedly, as reflected in F index, CaCO3 %, and foraminiferal preservation. In this interval, representing the “burndown” layer (BL; e.g., Thomas and Shackleton, 1996; Thomas et al., 1999; Giusberti et al., 2007; Figs. 4, 7, 8), CET remained present. Dissolution in the upper BL removed most thin, dissolution-prone calcareous tests (e.g., Siphogenerinoides brevispinosa and small bolivinids), concentrating the more heavily calcified and the agglutinated taxa (including CET; Figs. 5–7). Benthic foraminiferal assemblages in the topmost Paleocene at Forada thus cannot be interpreted with confidence due to the severe dissolution.

4.3 Climate and marine life during the PETM

4.3.1 The black clay: a desert below the CCD

This very thin, carbonate-free interval is somewhat enigmatic. The virtually barren sediment may have been deposited during the maximum rise of the CCD, under environmental conditions so unfavorable that benthic life was excluded, a “dead zone” (sensu Harries and Kauffman, 1990) during the earliest phase of the PETM. Geochemical redox indices in the black clay and the underlying and overlying samples suggest persistently oxygenated bottom waters (Giusberti et al., 2007), but may reflect diagenesis during re-oxygenation of bottom waters after a short period of anoxia, as commonly observed for Mediterranean sapropels (Higgs et al., 1994; van Santvoort et al., 1996). The presence of the thin black clay without microfossils is thus highly suggestive of a brief pulse of anoxia, as supported by a single peak value of organic carbon (0.6 wt%; Giusberti et al., 2007). The high value of biogenic barium (3151 ppm) in the black clay (Fig. 3), despite the fact that barite is generally not preserved under anoxic conditions (Paytan and Griffith, 2007; Paytan et al., 2007), may represent reprecipitation at theoxic–anoxic sediment interface after dissolution under anoxic conditions (Giusberti et al., 2007), and/or high rates of organic remineralization in the water column, during which the barite forms (Ma et al., 2014).

4.3.2 The early peak PETM (assemblages C and D)

The 10 cm of sediment directly overlying the Paleocene–Eocene boundary (i.e., the base of the CIE; Figs. 7, 8) was deposited in strongly CaCO3-corrosive waters, below the lysoclone and close to or below the CCD. The rapid rise of the CCD/lysoclone during the PETM is a predicted consequence.
of massive input of carbon (CO$_2$ or CH$_4$) in the ocean–atmosphere system on a millennial timescale (e.g., Dickens et al., 1997; Thomas, 1998; Zachos et al., 2005; Zeebe et al., 2009, 2014; Hönisch et al., 2012). The carbonate dissolution at Forada is consistent with observations at many other deep-sea sites (e.g., Schmitz et al., 1997; Thomas, 1998; Zachos et al., 2005; Kelly et al., 2010). The benthic foraminiferal extinction event (BEE) at Forada (i.e., corresponding to the BB1–BB2 zonal boundary of Berggren and Miller, 1989) occurs within this 10 cm thick interval, between the top of the CET-bearing burndown layer and the base of Assemblage D, where benthic calcareous taxa reappear (Figs. 4, 7, 8). The concentration of CET in the burndown layer, and the reappearance of calcareous hyaline taxa only 10 cm above the onset of the PETM at Forada, confirms that the CET extinction occurred over 3.5 kyr or less in the central-western Tethys, similar to evaluations of this timing from carbon cycle modeling (Zeebe et al., 2014).

Sediment just above the black clay, reflecting a first slight deepening of the CCD, contains a low-diversity fauna of mostly agglutinated, dwarfed (close to 63 µm in diameter) benthic foraminifera as well as calcareous nanofossils with signs of dissolution, with planktic foraminifera virtually absent (Agnini et al., 2007; Luciani et al., 2007; Fig. 8). This first wave of benthic pioneers recolonized the seafloor during the peak CIE, in CaCO$_3$-undersaturated waters, and reflects a highly stressed environment (Assemblage C; Figs. 6–8). Among the pioneers, Eobigenerina variabilis is peculiar of the PETM of the Forada section (Figs. 6, 7). Eobigenerina is a recently erected genus in the Textulariopsidae, including non-calcareous species previously assigned to Bigenerina (Cetean et al., 2011), and it is known to behave opportunistically during Cretaceous Oceanic Anoxic Event 2 (OAfrican; Table 1). A major component of the upper part of Assemblage C is Karrerulina conversa (Fig. 7). The species dominates the lowermost Eocene deposits in the Polish Carpathians (Bałk, 2004), commonly occurring in the Paleocene–Eocene of the central North Sea and Labrador margin, and in Morocco (Kaminski and Gradstein, 2005).

Modern Karrerulina (e.g., K. apicularis ≡ K. conversa) lives in oligotrophic abyssal plains, with well-oxygenated bottom and interstitial waters (Table 1). However, the test morphology of Karrerulina, combined with its abundant occurrence in the doubtless stressed environment of the basal PETM at Forada and Zumaia (Table 1), suggests that this genus may also act opportunistically.

After ca. 4 kyr, a further deepening of CCD allowed a consistent increase in abundance of benthic calcareous taxa (ca. 50%; Assemblage D; Fig. 2), coinciding with the lowest recovery of bulk carbonate $\delta^{13}$C values, from $-2$ to $-1$‰ (Giusberti et al., 2007; Tipple et al., 2011; Fig. 7). These calcareous recolonizers included dwarfed and thin-walled forms of G. subglobosa, Tappanina selmensis, Osangularia spp. and Oridorsalis umberatus (Figs. 6, 7). A similar peak in small Osangularia also occurs in the basal PETM at the Contessa section, as documented for the first time in the present paper (Fig. S2). Representatives of the genus Osangularia (Osangularia spp.) behaved opportunistically in the PETM.
of the Tethyan Alameda section (Alegret et al., 2009a). Moreover, Boscolo Galazzo et al. (2013) found small-size Osangularia within organic-rich levels immediately following the Middle Eocene Climatic Optimum in the Alano section (in northeastern Italy). During the Cretaceous OAEs, Osangularia spp. opportunistically repopulated the seafloor during short-term re-oxygenation phases (see references in Table 1). Although Osangularia is generally referred to as preferring stable, well-oxygenated environments (e.g., Murray, 2006; Alegret et al., 2003), we suggest that some extinct species of this genus could actually behave as opportunists and recolonizers.

Assemblage D contains almost equal abundances of calcareous and agglutinated taxa, indicating that factors other than bottom water CaCO₃ concentration were controlling faunal variability within this assemblage (Figs. 6, 7). Possibly, strongly enhanced runoff and sediment delivery can explain the abundance of agglutinated taxa (40–60 %), such as Glomospira spp. (e.g., Arreguín-Rodríguez et al., 2013, 2014), above the first 10 cm of the CMU. We thus recognize a rapid succession of recolonizer taxa during the first 12 kyr of the CIE (assemblages C–D). The small size of both the agglutinated and hyaline recolonizers is indicative of r-strategist species which reproduce quickly and can thus quickly repopulate stressed environments, as soon as conditions improve slightly (e.g., Koutsoukos et al., 1990; Thomas, 2003). The rapid pace at which different populations of recolonizers succeeded each other indicates a highly unstable environment, with marked fluctuations in the amount, timing and quality of the food reaching the seafloor. Sediment deposition during this interval may have occurred in rapid pulses, e.g., following intense rainstorms, carrying refractory organic matter to the deep-sea environment. Pauses between events may have allowed the benthic foraminifera to recolonize the sediment, profiting from the abundance of food. This is consistent with calcareous nanofossil assemblages showing an increase in Ericsonia and declines in abundance of Sphenolithus, Octolitithus, Zygrhablithus and Fasciculithus, indicating an unstable and nutrient-rich upper water column (Agnini et al., 2007; Fig. 8). Archaeal biomarkers show a large influx of terrestrial, soil-derived organic matter (branched and isoprenoid tetraethers, or BIT index) from the onset of the PETM up to ~ +10 cm (Tipple et al., 2011). Higher plant n-alkane average chain length (ACL) decreased immediately after the onset of the CIE, consistent with increased humidity (Fig. 9; Tipple et al., 2011). The abundance of the clay mineral chlorite indicates enhanced physical erosion (Robert and Kennett, 1994) during deposition of the lower 50 cm of the CMU, rapidly decreasing upward (Fig. S3).

The greenish marly clays containing assemblages C and D show primary lamination, indicating that macrobenthic invertebrates were absent, as at the Dee and Mead Stream sections (New Zealand; Nicolo et al., 2010) and Zumaya (Spain; Rodríguez-Tovar et al., 2011). The presence of benthic foraminifera, however, indicates that bottom and pore waters were not permanently anoxic. Pore waters may have become dysoxic periodically due to high temperatures, decomposing organic matter and possibly enhanced water column stratification, leading to the absence of metazoans and stressed benthic foraminiferal assemblages. Low-pH seafloor conditions may have also played a significant role in excluding macrobenthic fauna in this early phase of PETM at Forada. Deep-sea animals are highly sensitive to even modest but rapid pH changes (Seibel and Walsh, 2001), which are harmful even for infaunal deep-sea communities (Barry et al., 2004).

### 4.3.3 The core of the CIE and recovery (assemblages E, F)

The benthic foraminiferal assemblage changes significantly from Assemblage D to Assemblage E, coinciding with the gradual reappearance of motting (as thin reddish “flames” in the green sediment). Bolivinids return as a major faunal component (50 %), and agglutinated taxa decrease in abundance. Peaks of tapered elongate calcareous forms, including Siphogenerinoides brevispinosa, the “other buliminids” group, pleurostomellids and stilostomellids, replace the recolonizers (Figs. 5, 6). These groups could have been functioned as opportunistic taxa, able to flourish when food supply was periodically high (e.g., Table 1). Coinciding with Assemblage E, planktic foraminifera return to be a significant component of the microfossil assemblage (e.g., Luciani et al., 2007; Fig. 8), while radiolarians remain abundant throughout the CMU (Giusberti et al., 2007; Luciani et al., 2007). The planktic foraminiferal assemblage is dominated by acarininids, with a double peak of the excursion species Acarinina sibaiyaensis and A. africana, which, combined with the high percentages of the nanofossil Ericsonia, indicates warm and eutrophic surface waters (e.g., Ernst et al., 2006; Guasti and Speijer, 2007; Agnini et al., 2007; Luciani et al., 2007; Fig. 8).

Detrital hematite sharply increased in concentration at the onset of Assemblage E (Giusberti et al., 2007; Dallanave et al., 2010, 2012; Fig. 3). Hematite forms in soils under warm and dry conditions, and an increase in hematite in marine sediments is considered indicative of an arid climate over the adjoining land, with increased wind strength (Larasoaña et al., 2003; Zhang et al., 2007; Itambi et al., 2009), or humid to subhumid climates with seasonal drying (Torrent et al., 2006). It is delivered to the deep-sea environment through river runoff or as aeolian dust (e.g., Zhang et al., 2007; Itambi et al., 2009). Within the CMU, hematite shows cyclical fluctuations with a ~ 21 kyr periodicity, but other terrigenous components (quartz and phyllosilicates) do not co-vary in abundance after a ~ 15 % increase at the onset of the CMU (Fig. 3). To explain the different abundance patterns, we interpret hematite as wind-delivered and silicate minerals as runoff-delivered.
The hematite % peaks may be indicative of cyclical variability in wind-delivered material, rather than the earlier prevailing consistently humid climate. The lithological anomaly of the CMU, the 5-fold increase in sedimentation rates and increase in reworked Cretaceous nannofossils (Agnini et al., 2007; Fig. 8), and the silicate mineral and hematite % records all indicate marked fluctuations in the hydrological regime throughout this interval. High hematite % may reflect the presence of high-pressure cells over land, during an overall dry climate phase, with increased wind strength and dust delivery to the sea (Larrasoña et al., 2003; Zhang et al., 2007; Itambi et al., 2009). In contrast, low values of hematite % may indicate periods of greater humidity and enhanced precipitation. Such alternation of wet and arid phases favored deeper soil erosion on the continental areas surrounding the Belluno Basin (Thiry, 2000; Schmitz and Pujalte, 2003), causing major washouts during the wet phases, which may explain the 5-fold increase in sedimentation rates and 15 % increase in phyllosilicate abundance in the CMU (Fig. 3).

The hematite % cycles are in phase with cycles in CaCO\textsubscript{3} %, radiolarian abundance, and bulk carbonate δ\textsubscript{13}C, slightly preceding the others stratigraphically (Fig. 3). During the arid climate phase, enhanced wind strength may have generated intense surface water mixing and offshore nutrient upwelling, inducing increases in primary productivity and phytoplankton blooms. The blooms in primary productivity resulted in deposition of abundant algal biomass, leading to the occurrence of peaks of pleurostomellids, stilostomellids and Siphogenerinoides brevispinosa in Assemblage E. Productivity may have remained fairly high during the wet periods, as indicated by consistently high biogenic barium throughout the CMU (Giusberti et al., 2007; Paytan et al., 2007). During the rainy periods, upwelling rates may have been lower, with nutrients mostly supplied in river runoff. The delivery of food to the seafloor may have been more continuous, but with more important input of refractory organic matter from land.

In contrast to these proxies, which show cyclicity at precessional periods throughout the CMU, higher plant n-alkanes average chain length (ACL) and δD vary only in its lowermost 50 cm (Tippel et al., 2011; Fig. 9). Possibly, the sedimentary n-alkanes were derived from a pool of plant material produced during subsequent wet and dry phases, so that ACL and δD may represent averaged records of leaf wax n-alkanes produced during different mean climate states in the upper CMU. Even so, the δD values within the CMU are on average ~ 15 ‰ lower than above and below (Fig. 9), as reported for the Cigogna section (10 km away; Krishnan et al., 2015), possibly reflecting more humid conditions/higher precipitation during the PETM wet times (e.g., Sachse et al., 2006; Smith and Freeman, 2006), or greater productivity of plant material during the wet phases. Alternatively, it may reflect a primary change in the isotopic composition of meteoric waters (Krishnan et al., 2015).

In the following benthic foraminiferal Assemblage F (upper CMU, recovery phase), Siphogenerinoides brevispinosa and Tappanina selmensis are less abundant, whereas Bulimina taxapamensis, Abyssammina spp., and Nuttalides truempyi increase in relative abundance (Figs. 6, 7). These are typical deep-sea, open-ocean taxa which thrive under more oligotrophic conditions (e.g., Thomas, 1998), and might indicate progressively less intense or shorter primary productivity blooms during the arid phases, and/or mark the return to fully oxygenated seafloor and pore water conditions. Less intense eutrophy at the transition from Assemblage E to F is further supported by calcareous plankton data, showing a decrease in the planktic foraminiferal excursion species, and among nannofossils, a decrease in Ericsonia (Agnini et al., 2007; Luciani et al., 2007; Fig. 8). Coinciding with the top of the CMU, there were marked changes in calcareous plankton assemblages, although benthic foraminiferal Assemblage F persisted. Among calcareous nannofossils the abundance of Zygrhablithus, Sphenolithus and Octolithus increased, whereas that of reworked taxa decreased (Fig. 8). In the planktic foraminiferal assemblage, Acarinina species declined in abundance, and the fauna became more diverse, with fluctuations modulated by lithology in the marl–limestone couplets overlying the CMU (Fig. 8).

The lithological unit above the CMU consists of an alternation of limestones and marls at precessional frequencies (~ 21 kyr; Fig. 2). These limestone–marl couplets persist for up to 8 m above the CMU (well beyond the top of the studied interval; Giusberti et al., 2007; Luciani et al., 2007), then gradually become less clearly expressed, fading upwards. The marl–limestone couplets may reflect the persistence of wet (marl)-arid (limestone) cycles for ~ 800 kyr after the end of the CMU deposition, though at an amplitude declining over time. This persistence resembles the extended (650 kyr) humid period, starting at the onset of PETM, recognized in the sediment record at Site 401 of eastern North Atlantic (Bornemann et al., 2014). Our benthic foraminiferal data agree with this interpretation, showing substantially unchanged seafloor conditions up to +650 cm (uppermost sample analyzed).

4.4 Clues from Forada on PETM climate change

The integrated data set collected at Forada supports the occurrence of enhanced climatic contrasts and productivity changes in the western Tethys during the PETM, and agrees with previous studies suggesting intense weather extremes at mid- to subtropical latitudes (Fig. 10; Table S3). At the onset of the PETM, middle to subtropical latitudes may have been characterized by intense, monsoonal-type rainfall, followed by a succession of wet and arid phases, possibly precessionally paced, during the core of the PETM (e.g., Collinson et al., 2007; Kraus and Rigging, 2007; Egger et al., 2009; Foreman et al., 2014; Stassen et al., 2012a, b; 2015; Fig. 10 and
The Forada record allows for distinct recognizing the temporal successions among these distinct climatic phases up to 800 kyr after the onset of the PETM, as well as directly relating them to the progression of the CIE, its recovery and termination. The climatic conditions inferred from the Forada section and other records at similar latitudes differ from those derived from the subtropical net evaporation zone (15–35° N), (e.g., from the Tremp–Graus Basin – Pyrenees), which document a generally much drier climate with a brief interval of increased storminess and intense flash flood events at the onset of the PETM (Schmitz and Pujalte, 2007). Records from subtropical to mid-latitudes also differ from those derived from the northern rice belt and into the Arctic Basin (> 50° N), which suggests that humid conditions may have been more persistent there, with increased rates of precipitation and on average moister conditions during the PETM (Pagani et al., 2006b; Sluijs et al., 2006; Harding et al., 2011; Dypvik et al., 2011; Kender et al., 2012; Wieczorek et al., 2013; Fig. 10; Table S3).

The combination of all these climatic records (Fig. 10; Table S3) suggests that the net result of increased weather extremes during the core of the PETM might have been to decrease rainfall at subtropical to mid-latitudes, and increase moisture transport toward the high latitudes, as originally suggested by Pagani et al. (2006b). Few tropical records exist, so that precipitation changes here are less clear. Rainfall in coastal Tanzania may have decreased during the early PETM but it was combined with violent precipitation events and floodings (Handley et al., 2008; 2012; Aze et al., 2014; Table S3). In Central America, conditions during the PETM may have shifted to more continuously humid (Jaramillo et al., 2010). The long-lasting cyclicity and precise chronology at Forada suggest that this enhanced climate variability at subtropical to mid-latitudes may have lasted for several hundred thousand years after the onset of the CIE. Despite the possible decrease in net rainfall, these weather extremes persisting over several 10^5 kyr may have significantly enhanced the rate of erosion and weathering, through the occurrence of alternating wet–dry periods. The weathering may have led to a decrease in atmospheric CO\textsubscript{2} levels by consumption of CO\textsubscript{2} during weathering reactions. The increased supply of cations through enhanced weathering–erosion would have
driven ocean pH up and atmospheric CO₂ down (Broecker and Peng, 1982; Raymo et al., 1988; Zachos et al., 2005). Enhanced seasonal extremes across large geographical areas (the subtropical to mid-latitudinal belt) thus might have been a response to the large CO₂ input at the Paleocene–Eocene transition, and may have had a primary role in restoring the carbon cycle to steady state.

5 Conclusions

The continuous and expanded record of benthic foraminifera across the PETM at Forada, integrated with the extensive data sets previously generated across this interval, may provide the most complete reconstruction of ecological and climatic changes during the Paleocene–Eocene Thermal Maximum in Europe. Coupled sedimentological, molecular and micropaleontological records highlight a complex sequence of environmental and climatic changes during the time period across the CIE:

- Climatic and oceanographic conditions started to change ~30 kyr before the onset of the PETM, with a possible precursor event.

- Our high-resolution benthic foraminiferal record combined with the established chronology lets us infer that the BEE in the central-western Tethys occurred over a time interval of not more than 4 kyr. At the onset of the PETM, combined de-oxygenation, acidification and environmental instability may have synergistically impacted deep-sea life.

- Four benthic foraminiferal assemblages occur (C–E and lower F) within the CMU (coinciding with the main phase of CIE). Assemblage C is characterized by successive peaks of different agglutinated recolonizers. Calcareous recolonizers return in the following assemblage, after calcium carbonate saturation increased. The complex succession of peaks of agglutinated and hyaline recolonizers in these two assemblages (C, D; 12.5 kyr) suggests multiple repopulation episodes. The benthic foraminiferal data integrated with molecular and mineralogical data point to increased precipitation and strong continental erosion during this short initial stage of the PETM.

- Within the core of the CIE, δ¹³C and mineralogical properties such as hematite and calcium carbonate wt % vary at precessional periodicity. Combined with data on radiolarian abundance and benthic foraminiferal assemblage composition, this variability suggests an alternation of overall wetter and drier periods. Enhanced weather extremes during most of the PETM may have led to a decrease in total precipitation over the central-western Tethys.

- The benthic foraminiferal assemblage at Forada did not significantly change with the onset of the deposition of marl–limestone couplets unit above the CMU (middle and upper third of Assemblage F). This suggests that the enhanced climatic variability at precessional timescales persisted well after the end of the CIE recovery. We argue that enhanced seasonal extremes at mid-latitudes might have been a direct climate response to the huge CO₂ input at the Paleocene–Eocene transition, and may have had a primary role in restoring carbon cycle steady state through links with the water cycle and weathering rates.

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