Age and synchronicity of planktonic foraminiferal bioevents across the Cenomanian–Turonian boundary interval (middle Cretaceous)

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Abstract

The upper Cenomanian–lower Turonian is a key-stratigraphic interval, as it encompasses the mid-Cretaceous supergreenhouse and a major perturbation of the global carbon cycle (i.e., Oceanic Anoxic Event 2) as evidenced by a global positive carbon isotope excursion and by the nearly world-wide deposition of organic-rich marine facies. A turnover in planktonic foraminiferal
assemblages and in other marine organisms is documented across this stratigraphic interval, but
reconstruction of the timing and identification of the cause and effect relationships between
environmental perturbations and organism response require a highly-resolved stratigraphic
framework. The appearance and extinction levels of planktonic foraminiferal species generally
allow accurate intra- and supra-basinal correlations. However, bioevents cannot be assumed to be
globally synchronous, because the stratigraphic and geographic distribution of species is modulated
by the ecological preferences exhibited by each taxon and controlled by the oceanic circulation,
often resulting in earlier or delayed events in certain geographic areas.

The aim of this study is to test the synchronicity of the planktonic foraminiferal bioevents
recognized across the C/T boundary and to provide the most reliable sequence of events for
correlation of mid- to low latitude localities. For this purpose, we have completed a highly-resolved
biostratigraphic analysis of the European reference section for the C/T boundary at Eastbourne, Gun
Gardens (UK), and of core S57, (Tarfaya, Morocco) and correlated the sequence of bioevents
identified with those recorded in other coeval sections available in the literature, including the
GSSP section for the base of the Turonian Stage at Rock Canyon, Pueblo (Colorado), where we
calculate reliable estimates of planktonic foraminiferal events that are well-constrained by
radioisotopically and astrochronologically dated bentonite layers.

Results indicate that the extinctions of *Thalmanninella deeccki, Thalmanninella greenhornensis, Rotalipora cushmani* and “*Globigerinelloides* bentonensis” in the latest Cenomanian are reliable bioevents for correlation. In addition, our analysis highlights other promising lowest occurrences (LOs) that, however, need to be better constrained by bio- and chemostratigraphy, including the LO *Marginotruncana schneegansi* falling close to the C/T boundary. By contrast, the appearance of *Helvetoglobotruncana helvetica* and of some *Dicarinella*
species, the onset of the “Heterohelix” shift and the extinction of anaticinellids are clearly
diachronous across mid-low latitude localities. Finally, our study suggests that different species
concepts among authors, different sample size and sampling resolution, as well as species
paleoecology are important factors that control the stratigraphic position at which bioevents are
identified.

Keywords: Cenomanian–Turonian, stratigraphy, mid-low latitude correlations, planktonic
foraminifera, Pueblo, Eastbourne.

1. Introduction

The Cenomanian–Turonian boundary interval (middle Cretaceous) represents one of the most
interesting case-studies for investigating the evolution of the marine biota under the intense
environmental perturbations that occurred during Oceanic Anoxic Event 2 (e.g., Schlanger and
Jenkyns, 1976; Scholle and Arthur, 1980; Schlanger et al., 1987). In fact, OAE 2 is globally
recognized as a time of increased sea-surface productivity under greenhouse climate conditions
interrupted by a brief cooling episode (i.e., the “Plenus Cold Event”, see Gale and Christensen,
1996; Forster et al., 2007; Sinninghe Damsté et al., 2010; Jarvis et al., 2011; Jenkyns et al., 2017
among others) that may correspond to a re-oxygenation event of bottom waters in the Western
Interior Seaway (WIS) (i.e., the “benthonic zone”: Eicher and Worstell, 1970; Elderbak and Leckie,
2016). Across OAE 2, planktonic foraminiferal assemblages underwent a substantial re-
organization that led to the extinction of the single-keeled rotaliporids with umbilical supplementary
apertures (genera Rotalipora and Thalmanninella) and to the appearance and progressive
diversification of double-keeled taxa (genera *Dicarinella* and *Marginotruncana*), that dominated the assemblages until the Santonian (e.g., Robaszynski et al., 1990, 1993; Premoli Silva and Sliter, 1999; Leckie et al., 2002; Falzoni et al., 2013, 2016a; Petrizzo et al., 2017). However, correlating stratigraphic sequences, extrapolating global from local signals, and reconstructing the cause and effect relationships between environmental changes and organism response require a reproducible and highly-resolved stratigraphic framework. Unfortunately, the C–T boundary interval lacks magnetostratigraphic control, being within the Cretaceous Normal Superchron (e.g., Gradstein et al., 2012). Nevertheless, this interval is accompanied by a $\sim+2\%$ excursion in both the $\delta^{13}C_{\text{carb}}$ and $\delta^{13}C_{\text{org}}$ resulting from the burial of organic matter during OAE 2 (e.g., Jenkyns, 2010). The shape of the $\delta^{13}C$ profile with its typical peaks and troughs represents one of the most reproducible features of this stratigraphic interval, and being likely synchronously registered in the marine and continental records, it represents a powerful tool for global correlation (e.g., Pratt and Threlkeld, 1984; Tsikos et al., 2004; Jarvis et al., 2006, 2011; Jenkyns, 2010).

Planktonic foraminiferal bioevents are routinely applied to correlate pelagic and hemipelagic successions, and their contribution to implement the accuracy and resolution of the Geologic Time Scale is particularly important since the Early Cretaceous. However, despite this group of pelagic organisms having a wide distribution, each living/fossil species possesses ecologic preferences that may control its geographic and stratigraphic distribution. Consequently, planktonic foraminiferal bioevents cannot be assumed to be globally synchronous and their reliability for correlation requires testing through other relative dating techniques. For instance, the identification of the Cenomanian/Turonian boundary based on planktonic foraminiferal events is problematic. In fact, the base of the Turonian Stage is formally defined by the lowest occurrence (LO) of the ammonite *Watinoceras devonense* at the GSSP section at Rock Canyon, Pueblo, Colorado (Kennedy et al.,
However, ammonites are often rare or absent in hemipelagic and pelagic successions, thus the identification of the C/T boundary in the absence of the primary marker is based on secondary bioevents, including the LO of *Helvetoglobotruncana helvetica* among planktonic foraminifera. However, the appearance of *H. helvetica* is known to be an unreliable event to approximate the base of the Turonian, because of its diachronous occurrence, rarity in the lower part of its stratigraphic distribution, very transitional evolution from its ancestor *Helvetoglobotruncana praehelvetica*, and absence or very rare occurrence in epicontinental margin settings (e.g., Hart and Carter, 1975; Carter and Hart, 1977; Hart and Weaver, 1977; Hart and Bigg, 1981; Leckie, 1985; Hilbrecht et al., 1986; Jarvis et al., 1988; Lipson-Benitah et al., 1988; Robaszynski et al., 1990; Kuhnt et al., 1997; Keller et al., 2001; Luciani and Cobianchi, 1999; Tur et al., 2001; Petrizzo, 2001; Holbourn and Kuhnt, 2002; Caron et al., 2006; Mort et al., 2007; Desmares et al., 2007; Hart, 2008; Gebhardt et al., 2010; Huber and Petrizzo, 2014). Further complication is introduced by inconsistencies in the stratigraphic position of planktonic foraminiferal events, including the identification of LOs and HOs of marker taxa (e.g., *Rotalipora cushmani, H. helvetica*), when the same section is studied by different authors (Pueblo: Eicher and Diner, 1985; Leckie, 1985; Leckie et al., 1998; Keller and Pardo, 2004; Caron et al., 2006; Desmares et al., 2007; Elderbak and Leckie, 2016; Eastbourne: Paul et al., 1999; Keller et al., 2001; Hart et al., 2002; Tsikos et al., 2004).

The aim of this study is to select the most reliable and replicable sequence of planktonic foraminiferal bioevents across the C–T boundary interval, by distinguishing between the most trustworthy isochronous bioevents from those that instead are more regional or diachronous in nature. The section of Eastbourne at Gun Gardens and core S57 (Tarfaya Basin) were here re-studied at high-resolution to complement the planktonic foraminiferal data published in Tsikos et al.
Firstly, we developed a well-constrained age-model of the Pueblo (Rock Canyon) to obtain numerical estimates of planktonic foraminiferal events recognized in the GSSP section. Subsequently, we tested the synchronicity of each bioevent by performing graphic correlations between Pueblo (Rock Canyon) and Eastbourne (Gun Gardens) and between Pueblo (Rock Canyon) and Tarfaya (Core S57), and by comparing the stratigraphic position of each bioevent respect to the peaks and troughs of the $\delta^{13}$C profile in other mid-low latitude localities available in the literature, selected among those yielding the most complete stratigraphic record and a highly resolved $\delta^{13}$C profile. Italian sections are only briefly discussed, because of the absence of planktonic foraminifera in the black shale layers (e.g., Premoli Silva and Sliter, 1995; Premoli Silva et al., 1999; Coccioni and Luciani, 2004, 2005; Coccioni and Premoli Silva, 2015), and of the stratigraphic gap across the Bonarelli Level and time equivalent organic-rich facies (Gambacorta et al., 2015).

2. Materials and Methods

To document the sequence of planktonic foraminiferal bioevents across the C–T boundary interval, we have examined samples from (1) Eastbourne, Gun Gardens, UK, and (2) core S57 drilled in the Tarfaya Basin (Morocco), and paleogeographically located in the Anglo-Paris Basin and central Atlantic Ocean, respectively (Tsikos et al., 2004; Fig. 1). The Eastbourne section yields the most expanded C–T boundary interval of the English Chalk and represents the European reference section for the C/T boundary (Paul et al., 1999). Planktonic foraminifera have been object of a number of studies (Paul et al., 1999; Keller et al., 2001; Hart et al., 2002; Tsikos et al., 2004) and the section at Gun Gardens has been restudied to verify the discrepancies observed in the identification of species and position of the bioevents (including zonal markers). The sampling resolution adopted here for the biostratigraphic analysis is 20 cm throughout the section at
Eastbourne, and between 20 and 50 cm at Tarfaya. Rock samples from core S57 (Tarfaya) and from
the Plenus Marls Member (Eastbourne) have been processed with peroxide water to obtain washed
residues. Novelty of this study compared to Tsikos et al. (2004) is introduced by the disaggregation
of chalk samples from the Grey Chalk, Ballard Cliff and Holywell Members (Eastbourne) with
acetic acid (80%) and water (20%) to obtain washed residues yielding well-preserved isolated
specimens (see Lirer, 2000 and Falzoni et al., 2016b for detailed procedure), a procedure also used
by Elderbak and Leckie (2016) for the hard limestones at the Rock Canyon section.

In order to compare our biostratigraphic results with those from other localities, we have
selected the most complete stratigraphic sequences spanning the C–T boundary interval and having
detailed planktonic foraminiferal biostratigraphic data, as well as a highly-resolved $\delta^{13}$C$_{\text{carb}}$ or
$\delta^{13}$C$_{\text{org}}$ profile: WIS: (1) Rock Canyon, Pueblo, Colorado (Eicher and Diner, 1985; Leckie, 1985;
Leckie et al., 1998; Keller and Pardo, 2004; Caron et al., 2006; Desmares et al., 2007; Elderbak and
Leckie, 2016); Vocontian Basin: (2) Clot Chevalier (Falzoni et al., 2016b) and (3) Pont d’Issole
(Grosheny et al., 2006), SE France; Tethyan Ocean: (4) wadi Bahloul, Tunisia (Caron et al., 2006);
Indian Ocean: (5) Gongzha, Tibet (Bomou et al., 2013) (Fig. 1). In addition, planktonic
foraminiferal bioevents identified in these localities are briefly discussed by comparing their
stratigraphic position with other classic C/T boundary sections, where the $\delta^{13}$C profile is not
available. The published litho-, bio-, and chemostratigraphic data of Clot Chevalier, Pont d’Issole,
wadi Bahloul, and Gongzha are reproduced in the Supplementary Materials (Supplementary Figs.
A-D). Source of data for each section and the methodology applied to study the planktonic
foraminifera (thin sections, washed residues or a combination of both) are listed in Table 1.

Taxonomic concepts for planktonic foraminiferal species identification follow their original
descriptions and illustrations, the online taxonomic database for Mesozoic planktonic foraminifera
Genera attribution is according to the taxonomic revision by Gonzalez Donoso et al. (2007) for rotaliporids and Haynes et al. (2015) for biserial taxa. Species mentioned in the text and/or in the figures are listed in the Taxonomic Appendix. The planktonic foraminiferal biozonation is according to Sliter (1989) and Robaszynski and Caron (1995).

3. Remarks on the planktonic foraminiferal record at Pueblo (Colorado)

The GSSP for the base of the Turonian Stage is located at the Rock Canyon section at Pueblo (Colorado). The primary marker for the identification of the base of the Turonian is the LO of the ammonite *Watinoceras devonense* in bed 86 (Kennedy et al., 2000, 2005) (Fig. 2). According to the GSSP definition, additional secondary bioevents include the LO of the calcareous nannofossil *Quadrum gartneri*, which almost coincides with the C/T boundary as defined by ammonite stratigraphy at Pueblo (Tsikos et al., 2004), and the LO of the planktonic foraminifera *Helvetoglobotruncana helvetica*.

Planktonic foraminifera at Pueblo have been studied numerous times over the last 45 years with different sampling resolution (Eicher and Worstell 1970; Eicher and Diner, 1985, Leckie, 1985, Leckie et al., 1998; Keller and Pardo, 2004; Keller et al., 2004; Caron et al. 2006; Desmares et al., 2007; Elderbak and Leckie, 2016). Almost all the above-mentioned studies [with the exception of Eicher and Worstell (1970), where the planktonic foraminiferal biozonation is not discussed] assigned the sedimentary succession outcropping at Rock Canyon to the three planktonic foraminiferal biozones according to the subtropical biozonation by Sliter (1989) and Robaszynski and Caron (1995): *R. cushmani, Whiteinella archaeocretacea* and *H. helvetica* Zones. However, some discrepancies can be found in the identification of the zonal markers, as follows: the HO of *R.
cushmani is identified in Bed 65 (Kennedy et al., 2005 after Eicher and Diner, 1985), in Bed 66 (Keller and Pardo, 2004), and within Bed 68 (Leckie, 1985; Caron et al., 2006). Desmares et al. (2007) identified atypical morphotypes of R. cushmani (i.e., with a “discrete peripheral keel, which is sometimes not expressed on each chamber or is even totally absent”) up to Bed 85. Leckie (1985) also reported a single occurrence of R. cushmani as high as the upper part of Bed 85, but with a significant stratigraphic gap between this and the presumed HO of R. cushmani in Bed 68 (below Bentonite A). The 3.5-m gap between relatively rare but consistent presence of R. cushmani up to Bed 68, followed by no specimens, and then extremely sparse presence in the upper part of Bed 85 begs a question about reworking.

There are also major inconsistencies with regard to the position of the LO of H. helvetica, which is identified in Bed 86 by Desmares et al. (2007), in Bed 89 by Keller and Pardo (2004) and Kennedy et al. (2005) after Eicher and Diner (1985), in Bed 102 by Caron et al. (2006), and in limestone Bed 103 by Elderbak and Leckie (2016). It should be noted here that three-dimensional specimens of foraminifera were extracted and analyzed from calcareous shales, marlstones, and limestones in the study by Elderbak and Leckie (2016). Based on the above, we placed the top of the R. cushmani Zone according to Leckie (1985), representing the youngest record of the species, with the exception of the possibly reworked specimens within Bed 85, and the base of the H. helvetica Zone according to Elderbak and Leckie (2016) (Fig. 2).

4. Re-interpretation of A, B, and C peaks on the δ¹³C profile

Several δ¹³C_carb and δ¹³C_org records have been generated for the Rock Canyon section and for cores drilled nearby over the last 30 years (Pratt and Threlkeld, 1984; Pratt, 1985; Pratt et al., 1993; Keller et al., 2004; Bowman and Bralower, 2005; Caron et al., 2006; Sageman et al., 2006). In Fig.
we have plotted the $\delta^{13}C_{\text{carb}}$ obtained from outcrop samples at the GSSP section (Caron et al., 2006) and the $\delta^{13}C_{\text{org}}$ profiles obtained from the PU-79 core (Pratt and Threlkeld, 1984; Pratt, 1985). The $\delta^{13}C_{\text{org}}$ curve by Pratt and Threlkeld (1984) and Pratt (1985) was later reproduced by other authors including Kennedy et al. (2005) in the paper where the GSSP for the base of the Turonian Stage was defined, with some discrepancies compared to the original version (see Caron et al., 2006 for discussion).

Pratt and Threlkeld (1984) and Pratt (1985) described peaks A, B, C as follows: “A = initial rapid increase in values and first peak; B = notch caused by brief decrease in values; C = second increase and plateau of values”, meaning that Pratt and Threlkeld (1984) originally interpreted peak A as a maximum, peak B as a trough and peak C as the entire plateau of positive (= less negative) values above B rather than a single point of the $\delta^{13}C$ profile (Fig. 2). However, different criteria have been successively adopted for the identification of the peaks first identified by Pratt and Threlkeld (1984). For instance, the position of the carbon isotope peaks in the Demerara Rise record (Erbacher et al., 2005; Leg 201, Central Atlantic Ocean) has been interpreted as follows: peaks A and B are troughs, C is the positive peak, and a fourth maximum point (named D) is recognized below the decrease of the $\delta^{13}C$ to pre-excursion values. By contrast, Jarvis et al. (2006, 2011) named A, B and C the three $\delta^{13}C_{\text{carb}}$ maxima across the C–T boundary interval in a composite isotope curve of the English chalk, with peak C falling very close to the C/T boundary. Voigt et al. (2007, 2008) adopted the same criteria but also recognized a fourth positive peak above the C/T boundary that they named D. These latter schemes were followed by a number of authors in recent years (e.g., Pearce et al., 2009; Westermann et al., 2010; Bomou et al., 2013; Eldrett et al., 2015; Falzoni et al., 2016b) resulting in the common practice of approximating the C/T boundary to point C (as interpreted by Jarvis et al., 2006) in the absence of *W. devonense*. Other authors preferred to
number the observed maxima of the $\delta^{13}$C profile as I, II, and III (e.g., Caron et al., 2006; Grosheny et al., 2006).

Based on the observations above, the position of the carbon isotope peaks A, B, and C is here summarized in order to univocally compare and correlate the planktonic foraminiferal bioevents across the stratigraphic sections discussed in this study. Therefore, considering previous interpretations and according to Jarvis et al. (2006, 2011), we identify three positive points (A, B, C) and a plateau of high $\delta^{13}$C$_{\text{carb}}$ and $\delta^{13}$C$_{\text{org}}$ values, having a small offset, between B and C, and specifically, A is the initial rapid increase in values and first peak (as originally defined by Pratt and Threlkeld, 1984, and Pratt, 1985), B is the second positive peak of $\delta^{13}$C, following a decrease in values, and beginning of the plateau and is usually represented by multiple $\delta^{13}$C points, and C is the last positive peak of the plateau before the carbon-isotope profile gradually decreases to pre-excursion values. Nevertheless, uncertainties might remain on the identification of A, B, C peaks in some localities, because of the presence of additional peaks and troughs due to local variations of the $\delta^{13}$C content and/or to diagenesis that might complicate the apparently simple structure of the $\delta^{13}$C profile and/or to a different sampling resolution. For instance, point C (i.e., the last positive peak of the plateau before the $\delta^{13}$C decreases to pre-excursion values) at Eastbourne might be placed in two different positions, i.e., (1) at the transition between the Ballard Cliff and the Holywell Member according to Jarvis et al. (2006), or (2) near the top of the Ballard Cliff Member according to Voigt et al. (2008) (Fig. 3). Moreover, slight discrepancies in the stratigraphic position of peaks and troughs on the $\delta^{13}$C$_{\text{carb}}$ and $\delta^{13}$C$_{\text{org}}$ profiles are often observed in case both curves are available for the same section (e.g., Pueblo, Eastbourne).

5. Results
5.1. Eastbourne, Gun Gardens (UK)

Planktonic foraminiferal events identified at Eastbourne (Gun Gardens) in this study and those available in the literature (Paul et al., 1999; Keller et al., 2001; Hart et al., 2002; Tsikos et al., 2004) are combined with the available carbon isotope records (Paul et al., 1999; Tsikos et al., 2004) and plotted against stratigraphy (Fig. 3).

The HO of *R. cushmani* (Fig. 4, 1a–c) is recorded at top of Bed 3 (Keller et al., 2001) or within Bed 4 of the Plenus Marls Member (Paul et al., 1999; Hart et al., 2002; Tsikos et al., 2004), in agreement with this study, while the LO of *H. helvetica* is identified at the top of the Ballard Cliff (Keller et al., 2001) or at the base of the Holywell Member (Hart et al., 2002) (Fig. 3). By contrast, we do not record the occurrence of the latter species throughout the section, in agreement with Paul et al. (1999) and Tsikos et al. (2004). Consequently, the succession studied is assigned to the *R. cushmani* (from 0 to 11.4 m) and to the overlying *W. archaeocretacea* Zone (from 11.4 to 26 m) (Fig. 3), according to Tsikos et al. (2004).

Based on our biostratigraphic analysis, *Praeglobotruncana algeriana* (Fig. 4, 2a–c), *Dicarinella hagni* (Fig. 4, 3a–c), and *Dicarinella imbricata* (Fig. 4, 4a–c) occur from the base of the section, therefore their LOs likely fall in older stratigraphic intervals. Additional planktonic foraminiferal events identified in the Grey Chalk are listed below in stratigraphic order: 1) the LO of *H. praehelvetica* (Fig. 4, 5a–c) at 2.4 m above the base of the section; 2) the LO of *Dicarinella canaliculata* at 3.2 m; and 3) the LO of *Dicarinella elata* (Fig. 4, 6a–c) at 4.0 m. The following events are identified in the Plenus Marls Member: 1) the HO of *Thalmanninella brotzeni* (Fig. 4, 7a–c) at 7.2 m above the base of the section; 2) the HO of *Thalmanninella greenhornensis* (Fig. 4, 8a–c) and of 3) *Thalmanninella deeckei* (Fig. 4, 9a–c) at 8.2 m within Bed 1; 4) the LO of
Praeglobotruncana oraviensis (Fig. 4, 10a–c) at 8.8 m, and 5) the HO of Rotalipora montsalvensis (Fig. 5, 1a–c) at 9.2 m within Bed 2; 6) the HO of Rotalipora praemontsalvensis (Fig. 5, 2a–c) at 10 m within Bed 3; and 7) the HO of “Globigerinelloides” bentonensis at 13 m within Bed 7 (Fig. 3). Specimens that fall in the range of variability of W. archaeocretacea (Fig. 5, 3a–c) are identified from 0.6 m above the base of the section, but occur rarely in the assemblage and show an extremely scattered stratigraphic distribution, therefore their first appearance at 0.6 m may not correspond to its LO in the English Chalk. No noteworthy planktonic foraminiferal bioevents have been identified in the White Chalk Formation. The C/T boundary is here placed at the base of the W. devonense Zone according to Gale et al. (2005), however, it is worth mentioning that the ammonite species W. devonense is not identified at Eastbourne and the W. devonense Zone is recognized based on the occurrence of other coeval ammonite species (Paul et al., 1999; Gale et al., 2005).

5.2 Tarfaya (core S57)

The sedimentary succession studied is assigned to the R. cushmani (from the base of the core to 50.96 m) and to the overlying W. archaeocretacea Zone (from 50.96 m to the top of the cored interval), according to Tsikos et al. (2004). The occurrence of H. helvetica is not recorded throughout the stratigraphic interval examined (Fig. 6). Planktonic foraminiferal bioevents identified in this study are listed in stratigraphic order: (1) LO of H. praehelvetica (54.91 m) (Fig. 5, 4a–c), (2) HO of Th. deeckeii (54.16 m) (Fig. 5, 5a–c), (3) HO of Th. greenhornensis (53.96 m) (Fig. 5, 6a–c), and (4) HO of “G”. bentonensis (50.16 m) (Fig. 5, 7a–c). The “Heterohelix shift” (abundance of biserial taxa > 50% sensu Leckie et al., 1998) is recorded from 50.16 m. Biserial taxa, mainly Planoheterohelix moremani (Fig. 5, 8a–b) and Planoheterohelix paraglobulosa (Fig. 5, 9a–b) dominate the assemblage up to the top of the core. Praeglobotruncana algeriana (Fig. 5,
10a–c), *Dicarinella hagni* (Fig. 5, 11a–c), and *Dicarinella imbricata* occur from the bottom of the core. The C/T boundary is here approximate to fall between peak C on the δ⁰¹³Corg profile and the LO of *Q. gartneri* according to Tsikos et al. (2004).

6. Discussion

6.1 Age-depth model for the Pueblo section

Bentonites occurring in the Portland Core (Pueblo) were accurately and precisely dated by intercalibrating radioisotopic and astrochronologic time scales (Meyers et al., 2012). This study also concluded that bentonites found in the same ammonite biozone in different localities of the WIS across the C–T boundary interval have a common eruptive origin and are isochronous. Therefore, we used the age of bentonites obtained by Meyers et al. (2012) to calculate a reliable estimate of planktonic foraminiferal species first and last appearances data in the Rock Canyon section (Fig. 7). The age of the bentonites used to develop the age-model and the age of the bioevents extrapolated in this study are listed in Table 2. The calculated ages for the LO of *P. algeriana*, *D. hagni*, *D. elata*, *D. canaliculata*, *D. imbricata*, for the HO of *Th. deeckeii*, and for the LO of *H. praehelvetica* (base of the section) and for the LO of *M. marianosi* and *H. helvetica* (top of the section) include a higher margin of error, because these events fall outside the interval constrained by bentonites, although they are aligned to the line of correlation. The age of *M. sigali* was not calculated, because it falls in an interval where the sedimentation rate might have been significantly different (see Fig. 2). The HO of *Th. multiloculata* and of *R. planoconvexa* and of the atypical *R. cushmani* could not
be calculated because of the unavailability of the precise sample depth at which these events are recognized.

The age obtained for the HO of *R. cushmani* (94.29 Ma) is 10 kyr younger than those derived from the work by Robaszynski et al. (1998) in the Anglo-Paris Basin and reported in the GTS 2012 (Gradstein et al., 2012). Because the age estimate in the GTS 2012 was not well calibrated due to the uncertain HO of *R. cushmani* at Gubbio and in the Moroccan record (see Anthonissen and Ogg, 2012), our calculated age for this event represents a more reliable estimate of its extinction, as it falls very close to bentonite A. This age estimate represents a good approximation for the HO of *R. cushmani* across mid-low latitudes, with the exception of the Moroccan record where the interpretation of its apparently delayed extinction requires further studies (see discussion in paragraphs 6.1.2 and 6.3.1). The LO of *H. helvetica* at 93.48 Ma is also slightly younger (40 kyr) than previously estimated (GTS 2012; Gradstein et al., 2012), but because of the clearly diachronous nature of this event, this age cannot be applied to other localities. The LO of *D. imbricata* clearly precedes the extinction of *R. cushmani* at Rock Canyon and in other localities (see discussion below) and the age derived for its appearance in the GSSP section (94.51 Ma) is significantly older (310 kys) than estimated in the GTS 2012 (Gradstein et al., 2012), where this events is reported to occur above the HO of *R. cushmani*. Finally, it is worth noting that the extinction of *Th. greenhornensis* is significantly delayed at Pueblo compared to other mid-low latitude records (see paragraph 6.4.1 and Fig. 9), therefore the age obtained in this study has to be recalibrated in other localities.

6.2. Graphic correlations
To test the synchronism of events and the accuracy of correlations among sections, we performed graphic correlations of Pueblo vs. Eastbourne (Fig. 8a) and Pueblo vs. Tarfaya (Fig. 8b) (e.g., Shaw, 1964; Sadler, 2004; Paul and Lamolda, 2009; Petrizzo et al., 2011). To increase the number of common events, we integrate planktonic foraminiferal datums with the calcareous nannofossil, ammonite, and chemostratigraphic events (peaks A, B and C of the $\delta^{13}$C profile) available in the literature. Moreover, we considered the two interpretations regarding the position of peak C at Eastbourne (i.e., according to Jarvis et al., 2006 and to Voigt et al., 2008), in order to verify which option provides the highest correlation coefficient of the best-fit regression line. Depth of events used to constrain the graphic correlations and their source are listed in Table 3. For the Pueblo section, we considered the youngest record for extinctions and the oldest record for appearances in case the same event was recognized in different positions by different authors, with the exception of the HO of R. cushmani, which is according to Leckie (1985).

6.1.1. Pueblo vs. Eastbourne

The graphic correlation shows that the LO of H. praehelvetica is delayed at Pueblo or fall in an earlier stratigraphic interval at Eastbourne, therefore this event was excluded from the calculation of the regression line (Fig. 8a). We also excluded the HO of Th. greenhornensis, because this event falls in between peaks A and B of the $\delta^{13}$C curve at Pueblo and below peak A at Eastbourne, thus is delayed in the former section.

The values of the correlation coefficient of the regression line ($R^2$) calculated using all the other common events identified at Pueblo and at Eastbourne are similar when considering peak C placed according to Jarvis et al. (2006) ($R^2=0.90849$) and according to Voigt et al. (2008) ($R^2=0.92105$). However, the graphic correlation highlights a possible variation in the sedimentation rate in one or
both sections from around peak B, as testified by a change in the inclination of the line joining the events in the upper right of Fig. 8a. A significant decrease in the sedimentation rate in the upper part of the Eastbourne section is in agreement with the age model developed by Keller et al. (2001) and is likely due to a drop in the terrigenous input starting from the transition between the Plenus Marls Member (deposited during a sea-level low-stand) and the White Chalk Formation (deposited during a high-stand). A slight decrease in the sedimentation rate was also identified at Pueblo approximately at the same stratigraphic level (near the base of the ammonite *Neocardioceras juddii* Zone) (Meyers et al., 2001) in agreement with the age-depth model build in this study. Based on the observations above, we calculated two regression lines as follows: (1) from the base of the sections to the HO of “*G.* bentonensis” and (2) from peak B to the top of the sections, both having a correlation coefficient significantly high ($R^2 = 0.97$ and $R^2 = 0.96$, respectively) (Fig. 8a). In the latter case, we used the position of peak C as identified by Voigt et al. (2008), because it falls much closer to the other events and we maintained this interpretation in the discussion below.

6.1.2. Pueblo vs. Tarfaya

The graphic correlation highlights many differences in the position of the events (Fig. 8b). Firstly, the $\delta^{13}$C peaks are not perfectly aligned, suggesting a decrease in the sedimentation rate from peak B to peak C at Pueblo or an increase at Tarfaya, and/or an erroneous interpretation of their position on the $\delta^{13}$C$_{org}$ profile that may have been affected by diagenetic alteration as observed for the $\delta^{13}$C$_{carb}$ record (Tsikos et al., 2004). The presence of a 3-m thick coring gap and only two common events in the upper part of the sections, including the LO of *Q. gartneri* that likely falls within the non-recovery interval, complicate its interpretation (Fig. 8b).
Discrepancies are also found in the planktonic foraminiferal and calcareous nannofossil data, and the only events that appear trustable for correlation between Pueblo and Tarfaya are the LO of *H. praehelvetica* and the extinctions of *Th. deekei* and *Th. greenhornensis*, as well as the onset of the “Heterohelix” shift. By contrast, the HO of *A. albianus* is delayed at Pueblo or falls in an earlier stratigraphic interval at Tarfaya, while the opposite is true for the HO of *R. cushmani* and of “*G.* bentonensis”. These three events were not used to calculate the regression line, as they might be diachronous between the two sections. Although we do not observe any sedimentological or stratigraphic evidence of interruption of sedimentation, an alternative explanation to the apparently delayed extinctions of *R. cushmani* and of “*G.* bentonensis” at core S57 invokes the presence of a short hiatus between the HO of “*G.* bentonensis” and the onset of the “Heterohelix” shift, that would explain why these events are stratigraphically closer in this section compared to Pueblo and would move down the HO of *R. cushmani* and “*G.* bentonensis” to the correlation line. Moreover, a short hiatus in this position would complicate the interpretation of peaks A and B of the $\delta^{13}C_{org}$ profile adding further uncertainties to the correlation between Pueblo and Tarfaya. Because of the uncertainties regarding the position of the $\delta^{13}C$ peaks and little number of common events at the top of the stratigraphic interval studied, the evaluation of the reliability of planktonic foraminiferal events for correlation between Pueblo and Tarfaya requires further study and comparison with sections elsewhere.

6.3. Testing the accuracy of mid-low latitude correlations using planktonic foraminifera

In order to compare all the sections available and with the attempt to test the reliability of bioevents for correlating mid-low latitudes localities, we used the $\delta^{13}C$ isotope excursion, assuming that it was synchronously registered in the sedimentary successions. We have plotted in Fig. 9 the planktonic
foraminiferal bioevents herein identified at Eastbourne and Tarfaya and those documented from the selected stratigraphic sections (Pueblo, Clot Chevalier, Pont d’Issole, wadi Bahloul and Gongzha) against a schematic $\delta^{13}$C profile. The LO of $H. helvetica$ at Pueblo is plotted in Fig. 9 according to all authors. A summary of the most reliable sequence of planktonic foraminiferal bioevents resulting from our study is reproduced in Fig. 10.

6.3.1 Reliability of zonal markers

The extinction of $R. cushmani$ at Pueblo is recorded in slightly different stratigraphic intervals: from slightly below A to in between A and B of the $\delta^{13}$C curve. However, robust data based on both thin sections and washed residues place the HO of $R. cushmani$ in the trough of $\delta^{13}$C values above peak A in the $\delta^{13}$C curve (Table 1; Leckie, 1985; Leckie et al., 1998; Caron et al., 2006; Elderbak and Leckie, 2016). Remarkable is the identification of atypical $R. cushmani$ up to peak C (Leckie, 1985; Desmares et al., 2007), representing the youngest record of morphotypes falling within the range of variability of $R. cushmani$ documented in the literature (Fig. 9). The HO of $R. cushmani$ is diachronous from south to north within the WIS (Leckie, 1985; Desmares et al., 2007; Lowery et al., 2014), which is not surprising because of the local variations in the salinity, sea-surface temperatures and productivity, and the relatively shallow water depth (e.g., Caldwell and Kauffman, 1993; Arthur et al., 1985; Pratt, 1984, 1985; Leckie, 1985; Leckie et al., 1998; Pagani and Arthur, 1998; West et al., 1998; Keller et al., 2004; Corbett and Watkins, 2013; Lowery et al., 2014; Elderbak et al., 2014; Elderbak and Leckie, 2016, among many others) that might have hindered the migration of pelagic organisms throughout the basin.

In the other sections examined, the HO of $R. cushmani$ is typically recorded in between peak A and B, with the exception of Clot Chevalier and Tarfaya. In the former section, this event falls in
an earlier stratigraphic interval (below A) due to a combination of causes: (1) the presence of hiatus at the base of the Thomel Level and of an overlying condensed stratigraphic interval, and (2) the rarity of *R. cushmani* toward the top of its stratigraphic range, so that a possible hiatus or low sedimentation rate in this interval might considerably bias the position of its HO (Falzoni et al., 2016b). A more reliable HO of *R. cushmani* in the Vocontian Basin is identified at Pont d’Issole in between A and B (Grosheny et al., 2006; Grosheny et al., 2017).

The extinction of *R. cushmani* is recorded at Tarfaya (core S57) a few cm above the supposed peak B, while it falls at peak B in other cores drilled in the Tarfaya Basin (core S75: Kuhnt et al., 2005). Because of the remarkable similarity of these Moroccan core sections, and in the absence of sedimentological and micropaleontological evidences supporting reworking of older sediments at core S57, the extinction of *R. cushmani* appear to be slightly delayed in this area of the central Atlantic Ocean, as testified by the graphic correlation between Tarfaya and Pueblo. However, confirming the delayed HO of *R. cushmani* in this locality requires further studies, because of the potentially diagenetically altered $\delta^{13}$C$_{org}$ record of core S57 and the possible presence of intervals of interruption of sedimentation.

The appearance of *H. helvetica* is recorded across a 3 m-thick stratigraphic interval at Pueblo (Fig. 2) corresponding to the interval from slightly above peak C to the point where the $\delta^{13}$C$_{carb}$ returns to pre-excursion values (Fig. 9). The appearance of *H. helvetica* at Pont d’Issole and Gongzha falls slightly above point C, in agreement with its oldest identifications at Pueblo. According to Tsikos et al. (2004) and in this study, *H. helvetica* does not occur at Eastbourne and at Tarfaya, and it is not identified at Clot Chevalier (Falzoni et al., 2016b). Our study confirms the unreliability of the LO of *H. helvetica* as a marker event for the base of the Turonian, as explained in the Introduction section.
6.4. Secondary planktonic foraminiferal bioevents for mid-to-low latitude correlations

6.4.1. Reliable bioevents

Extinctions across the C–T boundary interval follow a well-defined scheme that is reproducible in all the stratigraphic sections examined, as listed below in stratigraphic order: 1) HO of *Th. deeckeii*, 2) HO of *Th. greenhornensis*, overlaid by the HO of *R. cushmani*, and by 3) the HO of “G”. bentonensis* (Figs. 9-10).

1) The extinction of *Th. deeckeii* always falls below peak A on the $\delta^{13}C$ curve, approximately around the point where the $\delta^{13}C$ profile begins to increase (Clot Chevalier, Pont d’Issole and Gongzha) or in a slightly younger stratigraphic interval where the $\delta^{13}C$ profile changes its grade and increases more distinctly (Pueblo, Eastbourne, Tarfaya).

2) The HO of *Th. greenhornensis* falls slightly below or at peak A, with the exception of Pueblo, where it falls in between A and B and together with the HO of *R. cushmani*, suggesting a significantly delayed extinction in the WIS compared to the other mid-low latitude localities.

Limited discrepancies in the HO of *Th. greenhornensis* and *Th. deeckeii* might be related to differences in sampling resolution and/or rareness of both species towards the top of their stratigraphic distribution. In other localities, the extinction of both species is recorded in the uppermost *R. cushmani* Zone (Blake Nose Plateau: Huber et al., 1999; Austria: Gebhardt et al., 2010; Switzerland: Westermann et al., 2010), with the exception of an apparently earlier HO of *Th. deeckeii* in Tunisia (Robaszynski et al., 1993) and Japan (Hasegawa, 1999) and of *Th. greenhornensis* in Morocco (Keller et al., 2008). It is worth mentioning that different *Th. deeckeii* species concepts might have been applied in the literature. For instance, Pessagno (1967) retained
Th. deeckeï a possible junior synonym of Th. greenhornensis, while it has been identified as distinct
species by subsequent authors (e.g., Robaszynski et al., 1979; Ando and Huber, 2007).

3) The extinction of “G”. bentonensis is recorded either slightly below (Pueblo, Eastbourne
and Clot Chevalier) or immediately above the supposed peak B (Tarfaya). Because this species was
not identified at wadi Bahloul and Gongzha, its extinction level in the eastern Tethyan realm cannot
be assessed. Based on the available data and pending further biostratigraphic studies in sections
belonging to this paleogeographic area, the HO of “G”. bentonensis appears a very reliable marker
for the latest Cenomanian. Its apparently slightly delayed extinction at Tarfaya should be verified
by further studies (see discussion in paragraph 6.3.1 regarding the HO of R. cushmani). Further
support to its validity is provided by its identification always some centimeters to few meters above
the extinction of R. cushmani in Spain (Lamolda et al., 1997) and Morocco (Keller et al., 2008),
while the apparently synchronous extinction of the single-keeled rotaliporids and of “G”.
bentonensis in several Italian sections (Bottaccione-Contessa: Premoli Silva and Sliter, 1995,
Coccioni and Premoli Silva, 2015; Antruiles, Dolomites: Luciani and Cobianchi, 1999; Calabianca-
Guidaloca: Scopelliti et al., 2004; Valdagno: Coccioni and Luciani, 2005) is due to the absence of
planktonic foraminifera in the Corg-enriched layers of the Bonarelli Level (or equivalent) and/or to
the stratigraphic gap across the C/T boundary (Gambacorta et al., 2015).

In addition, the LOs of D. hagni and D. imbricata that are usually recognized below the
beginning of the δ¹³C isotopic excursion, also appear reliable for correlation as discussed below
(Figs. 9-10):

4) The LO of D. hagni is recorded from the base of the sections in most of the localities
examined, with the exception of Pont d’Issole (i.e., across the δ¹³C rise below A) and wadi Bahloul
(i.e., slightly above B). However, its appearance level is well documented in the mid-to-upper R.
*cushmani* Zone of other geographic localities including Tunisia (Robaszynski et al., 1993), Morocco (Keller et al. 2008), Italy (Premoli Silva and Sliter, 1995; Luciani and Cobianchi, 1999; Mort et al., 2007; Coccioni and Premoli Silva, 2015), Spain (Lamolda et al., 1997), Austria (Gebhardt et al., 2010), Switzerland (Westermann et al., 2010), Blake Nose Plateau (Huber et al., 1999), and Japan (Hasegawa, 1999). We believe that its delayed occurrence at Pont d’Issole and wadi Bahloul might rely on its rarity in these localities and/or availability of small-sized samples. This latter hypothesis is supported by the fact that planktonic foraminifera from Pont d’Issole and wadi Bahloul were studied in thin sections in layers characterized by a particularly indurated lithology. Thin sections represent a smaller sized sample compared to washed residues and their study reduces the likelihood of encountering rare species. Discrepancies in the LO of *D. hagni* at low latitudes are found in the WIS: at Pueblo, the LO of *D. hagni* is at the extinction level of rotaliporids (Leckie, 1985), while in south Texas, the LO lies above the extinction level of rotaliporids (Frush and Eicher, 1975; Lowery and Leckie, 2017) suggesting an ecologic control at the southern aperture of the WIS relative to sites to the north in the core of the seaway (recorded as *P. difformis*, Eicher and Worstell, 1970; Eicher and Diner, 1985), indicating that the LO of this species is likely diachronous for sections within the WIS. This diachronous pattern in the WIS is similar to that of the HO of *R. cushmani*, which is also from south to north (Leckie, 1985).

5) The LO of *D. imbricata* is identified from the base of the sections or in the lowermost samples at Pueblo, Eastbourne, Tarfaya, and Clot Chevalier. Its LO appears delayed in the sections that have been partially studied in thin section as follows: at Pont d’Issole (close to excursion A), at Gongzha (slightly above excursion B), and at wadi Bahloul (in between B and C). In sections elsewhere, its LO is documented in the *R. cushmani* Zone (Italy: Premoli Silva and Sliter, 1995; Luciani and Cobianchi, 1999; Coccioni and Luciani, 2004; Mort et al., 2007; Coccioni and Premoli...
Silva, 2015; Spain: Lamolda et al., 1997; Japan: Hasegawa, 1999; Morocco: Keller et al., 2008) and in the *W. archaeocretacea* Zone (Austria: Gebhardt et al., 2010; Switzerland: Westermann et al., 2010). Despite some discrepancies in the LO of *D. imbricata* might be related to subjective species concepts, we believe that its apparent diachronism might rely on the sample size, as this species is often uncommon at the beginning of its stratigraphic range. Overall, in our opinion the appearance of *D. imbricata* can be considered a trustable bioevent for correlation in cases where the size of the samples studied is large enough to encounter rare species.

6.4.2. Bioevents potentially useful but requiring further investigations

1) The LO of *P. algeriana* is an upper Cenomanian event falling in the mid-upper *R. cushmani* Zone below the $\delta^{13}$C isotope excursion A (Pueblo, Pont d’Issole, wadi Bahloul), whereas the occurrence of this species is recorded at Eastbourne, Tarfaya, Clot Chevalier and Gongzha from the base of the section, so that its LO cannot be precisely determined (Fig. 9). However, the appearance of *P. algeriana* is documented in the lower *R. cushmani* Zone (Italy: Premoli Silva and Sliter, 1995; Luciani and Cobianchi, 1999; Spain: Lamolda et al., 1997; Blake Nose Plateau: Huber et al., 1999). Accordingly, several authors identified a *P. algeriana* Subzone defined as the stratigraphic interval between the LO of *P. algeriana* and the HO of *R. cushmani* (Bottaccione-Contessa: Premoli Silva and Sliter, 1995; Coccioni and Premoli Silva, 2015; Eastbourne: Keller et al., 2001). This diachronous appearance likely reflects different species concepts among authors as testified by its accommodation either in the genus *Praeglobotruncana* (Caron, 1966) or *Dicarinella* (Robaszynski et al., 1979). Recently, its distinctive morphological features have been clarified to promote its identification and calibrate its appearance level at a regional to global scale (see Falzoni et al., 2016b).
2) The occurrence of *P. oraviensis* is rarely recorded in the literature with few exceptions (Tunisia: Robaszynski et al., 1990; Spain: Lamolda et al., 1997; Crimea: Kopaevich and Vishnevskaya, 2016; Clot Chevalier: Falzoni et al., 2016b) and its species concept has been differently interpreted by authors including its generic attribution, because of the unavailability of SEM images of the type material (see Falzoni et al., 2016b for taxonomic details). Possibly because of these taxonomic uncertainties, the appearance of *P. oraviensis* is recorded in different levels within the *W. archaeocretacea* Zone (Robaszynski et al., 1990; Lamolda et al., 1997; Kopaevich and Vishnevskaya, 2016; Falzoni et al., 2016b). At Eastbourne, we identify the LO of *P. oraviensis* at the top of *R. cushmani* Zone, representing the oldest record of this species documented in the literature (Figs. 9-10). The delayed occurrence of *P. oraviensis* at Clot Chevalier, in the middle-upper *W. archaeocretacea* Zone likely results from a combination of sedimentologic (hiatus and condensed stratigraphic interval at the top of the *R. cushmani* Zone) and ecologically-related (very rare occurrence of planktonic foraminifera within the lower *W. archaeocretacea* Zone) causes. *Praeglobotruncana oraviensis* does not occur at Tarfaya or at Pueblo, suggesting that some ecologic features (e.g., water depth, trophic regime) might have controlled its geographic distribution at least at the beginning of its stratigraphic range, therefore the reliability of its LO requires further investigation and calibration with other sections.

2) The LO of *M. schneegansi* is recorded slightly above peak C at Pont d’Issole (Grosheny et al., 2006), while ancestor morphotypes named *M. cf. schneegansi* do occur at Clot Chevalier in a slightly younger stratigraphic interval where δ¹³C returns close to pre-excursion values (Falzoni et al., 2016b) (Fig. 9). The LO of *M. schneegansi* is identified in sediments of approximately coeval age either slightly below (Japan: Hasegawa, 1999) or above the LO *H. helvetica* (Tunisia: Robaszynski et al., 1990; Italy: Premoli Silva and Sliter, 1995; Coccioni and Premoli Silva, 2015;
Texas: Lowery and Leckie, 2017). Unfortunately, the synchronicity of the appearance of *M. schneegansi* in these sections cannot be accurately tested, lacking the δ¹³C profile. In addition, *M. schneegansi* is not documented in the lowermost Turonian of the other localities examined and it is absent in the southern mid-high latitudes (Petrizzo, 2000, 2001), suggesting that its geographic distribution might be confined to the tropical-subtropical latitudinal bands. Consequently, the reliability of its LO requires further study, but it might represent a useful bioevent falling close to the C/T boundary at least at low latitudes (Fig. 10).

3) The LO of *M. sigali* is detected well above peak C, but within the lower *H. helvetica* Zone, at Pueblo and at Gongzhaxia, although it seems to be delayed in the former section. Possible ancestor morphotypes of *M. sigali* do occur at Clot Chevalier approximately across the same stratigraphic interval (above C) (Fig. 9). This species is absent at Tarfaya, Eastbourne, Pont d’Issole, and wadi Bahloul, but it is usually documented to first occur slightly below (Furlo: Mort et al., 2007, south Texas: Lowery and Leckie, 2017), or above the LO of *H. helvetica* (Tunisia: Robaszynski et al., 1990; Italy: Premoli Silva and Sliter, 1995; Coccioni and Premoli Silva, 2015; Switzerland: Westermann et al., 2010). Because the appearance level of *M. sigali* is still not documented in several localities, its reliability for mid-to-low latitude correlation requires further studies.

4) LO of other *Marginotruncana* species. The LO of *M. renzi* was identified well above C at the top of the Eastbourne section by Paul et al. (1999) in a slightly younger stratigraphic interval compared to that here re-studied and assigned to the ammonite *Mammites nodosoides* Zone (Figs. 3 and 9). The LO of *M. renzi* is documented slightly above the LO of *H. helvetica* at the Blake Nose Plateau (Huber et al., 1999), in south Texas (Lowery and Leckie, 2017), Italy (Premoli Silva and Sliter, 1995; Coccioni and Premoli Silva, 2015), and Tunisia (Robaszynski et al., 1990). The LO of
M. marianosi is documented at Pueblo and Pont d’Issole falling in the H. helvetica Zone and above C, where the $\delta^{13}$C returns close to pre-excursion values, respectively (Fig. 9). In other localities, the LO of M. marianosi is recorded below (Furlo: Mort et al., 2007) or slightly above (Bottaccione-Contessa: Premoli Silva and Sliter, 1995) the LO of H. helvetica, but this bioevent is significantly delayed in the southern mid-latitudes (Exmouth Plateau: Petrizzo, 2000), as it falls above the extinction of Falsotruncana maslakovae in the late Turonian-early Coniacian. The LO of M. coronata is identified in the lower (Pont d’Issole: Grosheny et al., 2006) or at the top of the H. helvetica Zone in the Tethyan Realm (Tunisia: Robaszynski et al., 1990; Italy: Premoli Silva and Sliter, 1995; Coccioni and Premoli Silva, 2015; Tanzania: Huber and Petrizzo, 2014), and in the southern mid-latitudes (Exmouth Plateau: Petrizzo, 2000). In south Texas, the LO of M. coronata is above the HO of H. helvetica (Frush and Eicher, 1975; Lowery and Leckie, 2017).

Overall, the reliability for correlation of Marginotruncana species needs further investigations and calibration with the carbon isotope record in other localities. Noteworthy, the appearance of marginotruncanids predates the LO of H. helvetica in the southern Indian Ocean (Kerguelen Plateau: Petrizzo, 2001), potentially representing a powerful tool to correlate low-to-high latitude records.

6.4.3. Misleading bioevents

1) The LO of H. praehelvetica has been recorded from below peak A (well below the beginning of the $\delta^{13}$C isotopic excursion) to below peak B (Fig. 9) and in different stratigraphic levels at Pueblo and Eastbourne (Figs. 2-3). This bioevent is identified in the literature, either in the R. cushmani Zone (Bottaccione: Premoli Silva and Sliter, 1995; Furlo: Mort et al., 2007; Antruiles, Dolomites: Luciani and Cobianchi, 1999; Blake Nose Plateau: Huber et al., 1999; Tarfaya: Keller et al., 2008),
at the extinction level of the rotaliporids (Leckie, 1985), or within the lower *W. archaeocretacea* Zone (Tunisia: Robaszynski et al., 1990; Spain: Lamolda et al., 1997). These discrepancies are likely due to the common occurrence of transitional morphotypes between *H. praehelvetica* and its ancestor *Whiteinella aprica* making the identification of the first representative of the species almost subjective (Huber and Petrizzo, 2014). Because of the observations listed above, we regard the LO of *H. praehelvetica* as an unreliable marker for correlation.

2) *Whiteinella archaeocretacea* occurs from the base (or nearly the base) of the section at Pueblo, Eastbourne, Gongzha and wadi Bahloul, while its LO is recorded slightly below A at Pont d’Issole. Specimens strictly resembling the holotype were not identified at Tarfaya and Clot Chevalier (Fig. 9). At Eastbourne, *W. archaeocretacea* is extremely rare and shows a very discontinuous stratigraphic distribution, suggesting that the identification of its lowest appearance level might be strongly biased by a low sampling resolution or by the analyses of small-sized samples. Discrepancies in its LO might also be due to a subjective species concept, because specimens having a rounded (resembling the holotype) as well as a pinched lateral profile (resembling the paratype) were retained to fall in its range of variability. Pending further taxonomic studies and because of its rarity in the assemblages, we regard the LO of *W. archaeocretacea* as an unreliable bioevent.

3) The extinctions of *R. montsalvensis* and *Th. brotzeni* have been identified at Gongzha well below the δ¹³C excursion. At Eastbourne *Th. brotzeni* disappears in the stratigraphic interval where we observe the first δ¹³C rise, while *R. montsalvensis* becomes extinct slightly above the beginning of the second δ¹³C rise, both below A (Fig. 9). Both bioevents have been recorded to fall in the middle *R. cushmani* Zone in the Bottaccione-Contessa composite section (Coccioni and Premoli Silva, 2015), where *R. montsalvensis* and *Th. brotzeni* show a scattered occurrence toward
the top of their stratigraphic range (as well as at Eastbourne), leading to some uncertainties regarding the position of their extinction level. By contrast, other studies indicate the HO of both species falls in an older stratigraphic interval below the appearance of *R. cushmani* (Hasegawa, 1999; Westermann et al., 2010). Pending further studies, we interpret the HO of *R. montsalvensis* and *Th. brotzeni* as being controlled by local environmental conditions and because of their rarity toward the top of their stratigraphic distribution we discourage to use their extinction level for correlation.

4) The LO of *D. elata* is recorded below excursion A (Pueblo, Clot Chevalier, Eastbourne) and above excursion B (wadi Bahloul and Gongzha) (Fig. 9). Remarkably, *D. elata* is identified co-occurring with *Thalmanninella globotruncanoides* in the middle Cenomanian of Tunisia (Kalaat Senan: Robaszynski et al., 1993), representing its oldest documented record in the literature. Most studies identified its LO in the uppermost *R. cushmani* Zone in Spain (Lamolda et al., 1997), whereas its occurrence is not recognized at Tarfaya and Pont d’Issole, in the Italian sections (Bottaccione section: Premoli Silva and Sliter, 1995; Coccioni and Luciani, 2004; Coccioni and Premoli Silva, 2015; Antruiles, Dolomites: Luciani and Cobianchi, 1999; Furlo: Mort et al., 2007), at the Blake Nose Plateau (Huber et al., 1999), in Morocco (Keller et al., 2008), Switzerland (Westermann et al., 2010) and Japan (Hasegawa, 1999). Although discrepancies in its LO might be related to the rarity of *D. elata* in some environmental settings, so that its occurrence might not be detected in small-sized samples and poorly resolved biostratigraphic studies, the observations listed above support its unreliability for correlation, because of its presumably stenotopic ecology and absence in several localities. Discrepancies in the identification of its appearance level may also relay on different species concepts.
5) The LO of *D. canaliculata* has been recorded to fall in different stratigraphic levels as follows: below the initial $\delta^{13}$C positive excursion (Pueblo and Eastbourne), slightly below excursion A (Clot Chevalier), above excursion C (Pont d’Issole and wadi Bahloul), whereas *D. canaliculata* is absent at Tarfaya and Gongzha (Fig. 9). Discrepancies in its appearance level are found in other localities: its LO is identified in the upper *R. cushmani* Zone (Bottaccione-Contessa: Coccioni and Premoli Silva, 2015; Antruiles, Dolomites: Luciani and Cobianchi, 1999; Japan: Hasegawa, 1999), within the *W. archaeocretacea* Zone (Blake Nose Plateau: Huber et al., 1999), and within the *H. helvetica* Zone (Tunisia: Robaszynski et al., 1990). The sections in south Texas may have experienced conditions of environmental exclusion, very low abundances, and/or poor preservation that result in a much delayed LO of *D. canaliculata* within or at the top of the *H. helvetica* Zone (Lowery and Leckie, 2017). Based on the above and on its distinctive morphology, we interpret this bioevent to be considerably diachronous and likely subject to ecologic control.

6) The genus *Anaticinella* was erected to include ecophenotypes that evolved from the typical single-keeled rotaliporids by losing the peripheral keel and inflating the chambers on both the umbilical and spiral sides (Eicher, 1973); this morphologic adaptation was interpreted as forced by the expansion of the oxygen minimum zone at the onset of the OAE 2 that induced the exploitation of sea-surface habitats by taxa that were deep-dwellers (Wonders, 1980; Leckie, 1985; Desmares et al., 2007). Two species were included in the genus *Anaticinella* (=*Pseudoticinella* Longoria, 1973): *multiloculata* and *planoconvexa* (Longoria, 1973). More recently, *planoconvexa* was accommodated in the genus *Rotalipora*, as it was interpreted to directly evolve from *R. cushmani* (Desmares et al., 2008), while the species *multiloculata* belongs to the *Th. greenhornensis* phyletic lineage, thus was accommodated in the genus *Thalmanninella* (González Donoso et al., 2007; Desmares et al., 2008). *Anaticinella* species have been largely documented in the WIS.
Caron et al., 2006; Desmares et al., 2007; 2008), but their occurrence is also recorded in other low latitudes localities (Eastbourne: Keller et al., 2001; France: Grosheny et al., 2006; Tunisia: Caron et al., 2006; Grosheny et al., 2013; Morocco: Keller et al., 2008; Tibet: Bomou et al., 2013).

Desmares et al. (2007) identified the extinction of *Th. multiloculata* and *R. planoconvexa* at Pueblo as follows: a) HO *Th. multiloculata* in between excursions A and B, and b) HO *R. planoconvexa* slightly above excursion C (Fig. 9). The HO of *Th. multiloculata* is recorded below A at Pont d’Issole (Grosheny et al., 2006) and close to A at Gongzha (Bomou et al., 2013), while the HO of *Anaticinella* species is recorded slightly above C at wadi Bahloul. Morphotypes falling in the range of variability of *Th. multiloculata* and *R. planoconvexa* are not identified at Clot Chevalier (Falzoni et al., 2016b) and neither at Eastbourne and Tarfaya, although *Th. multiloculata* is recognized at Eastbourne by Keller et al. (2001). Specimens resembling *R. cushmani* but having 4 to 5 chambers more inflated chambers and a very weakly developed peripheral keel on the first chambers of the last whorl occur rarely at Eastbourne (here figured in Fig. 5, 2a–c). In our opinion and according to Robaszynski et al. (1993), these specimens closely resemble to the original description and to the drawing of the holotype of *Rotalipora praemontsalvensis* (Ion, 1976), rather than to *R. planoconvexa* (Longoria, 1973). However, such specimens might have been included in the genus *Anaticinella* or in the atypical *R. cushmani* morphotypes by previous authors (e.g., Leckie, 1985; Caron et al., 2006; Desmares et al., 2007), especially when observed in thin sections. Further studies are required to better assess the taxonomic status and phyletic relationship among rotaliporids. On the other hand, the geographic distribution of *Th. multiloculata* and of *R. planoconvexa* (sensu stricto) should be further investigated and their occurrence outside the WIS
should be more robustly supported. However, the extinction of Anaticinella species is clearly diachronous (Fig. 9).

7) The “Heterohelix shift” was first described by Leckie (1985), Leckie et al. (1998), and West et al. (1998) in the WIS as an abrupt change planktonic foraminiferal assemblages, which became dominated by biserial taxa (>50% of the population). Accordingly, it has been interpreted as a period of unstable eutrophic surface water conditions that inhibited the proliferation of the keeled K-strategist taxa. The “Heterohelix shift” is identified below excursion A at Gongzha, between excursions A and B at wadi Bahloul, and between excursions B and C at Pueblo and Tarfaya (Fig. 9). The “Heterohelix shift” is not documented in the Vocontian Basin (Clot Chevalier, Pont d’Issole), but recognized around excursion B at Eastbourne (Keller et al., 2001) as an increase in the abundance of biserial taxa from 40% to >60% of the population in the >63 µm size-fraction. Such a dominance of heterohelicids could not be confirmed in the samples examined during this study that instead revealed an increase in the abundance of calcispheres in the same stratigraphic interval (Fig. 11), as reported by Pearce et al. (2009). The “Heterohelix shift” is documented in the lower-middle W. archaeocretacea Zone of other localities (Italy: Coccioni and Luciani, 2004; Tunisia: Nederbragt and Fiorentino, 1999; Morocco: Keller et al. 2008), with the exception of Huber et al. (1999), who identified this bioevent in the lower H. helvetica Zone at the Blake Nose Plateau. Despite the onset of the “Heterohelix shift” is apparently synchronous between Pueblo and Tarfaya (Fig. 8b), its diachronous occurrence across the other mid-low latitude localities (Fig. 9) discourages its application for interbasinal correlations.

7. Conclusion
A highly-resolved biostratigraphic analysis of planktonic foraminiferal assemblages at Eastbourne and Tarfaya, compared with the record at the Turonian GSSP in Colorado, allowed recognition of a sequence of bioevents that are compared to those recorded in other sections available in the literature and correlated to the $\delta^{13}C$ profile. We calculated reliable estimates of the age of most planktonic foraminiferal events identified in the Pueblo GSSP section, including the extinction of the zonal marker *R. cushmani*. Results of graphic correlations and comparison between the sections analyzed indicate that the extinctions of Cenomanian taxa represent the most reproducible sequence of bioevents at mid-low latitudes and should be considered reliable for supra-basinal correlations. This sequence includes, in stratigraphic order, the HOs of: (1) *Th. deeckeii*, (2) *Th. greenhornensis*, (3) *R. cushmani*, and (4) “G”. *bentonensis*. Few exceptions to this scheme are detected at Pueblo (delayed HO of *Th. greenhornensis*) and at Tarfaya (delayed HO of *R. cushmani* and of “G.” *bentonensis*) (Fig. 10). Also, the extinction of *R. cushmani* is diachronous within areas of the WIS. The LOs of *D. hagni* and *D. imbricata* in the pre-excursion interval may be considered additional trustable bioevents for correlation in case the size of the samples used for planktonic foraminiferal biostratigraphy is large enough to ensure the identification of rare species.

Additional useful bioevents that, however, require further investigation, because of their rare identification in several localities or poor calibration with other bio- and chemostratigraphic data are the LOs of *Praeglobotruncana oraviensis* at the top of the *R. cushmani* Zone, and of *Marginotruncana schneegansi*, the latter being particularly promising to approximate the C/T boundary in low latitudes localities. Little information is presently available to test the synchronicity of the appearance of *P. algeriana* in the mid-upper Cenomanian and of other *Marginotruncana* species (i.e, *M. sigali*, *M. coronata* and *M. marianosi*), in the Turonian, but these
events appear worth of being further investigated. By contrast, the geographic and stratigraphic
distribution of *D. elata*, and *D. canaliculata* were likely ecologically driven; however,
inconsistencies in the application of different species concepts by authors are difficult to assess, but
might have introduced additional discrepancies in the identification of their LOs. A very transitional
evolution from the ancestor species and different species concepts among authors can be invoked as
a cause for the diachronous LO of *H. praehelvetica* and *H. helvetica*, while *W. archaeocretacea, Th.
brotzeni* and *R. montsalvensis* occur too rarely in the stratigraphic interval examined, so that their
appearance/extinction can be misleading to trace correlations. Finally, our study confirms the
unreliability of the LO of *H. helvetica* as a marker for the base of the Turonian and suggests that the
“*Heterohelix* shift” represents a response of the planktonic foraminiferal assemblages to a
local/regional increase in sea-surface productivity. In addition, we highlight that the occurrence of
anaticinellids (sensu strictu) is still poorly documented outside the WIS and, regardless, their
extinctions are clearly diachronous.

To conclude, we remark that further efforts have still to be directed toward the stabilization
of the taxonomic concepts of several planktonic foraminiferal species, in order to assure an univocal
approach during biostratigraphic analyses. Moreover, a small sample size and/or a low sampling
resolution might significantly influence the level at which LO and HO are identified even in case
the bioevent is geologically isochronous and these factors should be taken in consideration when
tracing correlations. On the other hand, we underline that the identification of the $\delta^{13}$C peaks and
troughs is not straightforward and should always be supported by a highly-resolved sequence of
bioevents.

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Agostino Rizzi (CNR, Italy) is thanked for assistance at the SEM. FF carried out this research during a post-doc fellowship of the University of Milan. This study was funded by MIUR-PRIN 2010-2011 (2010X3PP8J_001) to Elisabetta Erba (scientific coordinator).

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Planktonic foraminiferal evolutionary, geochemical and palaeoceanographic patterns.

Sedimentology 64, 252–285.


Taxonomic Appendix

List of planktonic foraminiferal species with authors and years mentioned in the text and/or in the figures.
Dicarinella canaliculata (Reuss, 1854)

Dicarinella elata Lamolda, 1977

Dicarinella hagni (Scheibnerova, 1962)

Dicarinella imbricata (Mornod, 1950)

Falsotruncana maslakovae Caron, 1981

“Globigerinelloides” bentonensis (Morrow, 1934). The genus Globigerinelloides is herein indicated in brackets, as it is currently under revision by the Mesozoic Planktonic Foraminiferal Working Group see Taxonomic notes in Petrizzo et al. (2017).

Helvetoglobotruncana helvetica (Bolli, 1945)

Helvetoglobotruncana praehelvetica (Trujillo, 1960)

Marginotruncana coronata (Bolli, 1945)

Marginotruncana marianosi (Douglas, 1969)

Marginotruncana renzi (Gandolfi, 1942)

Marginotruncana schneegansi (Sigal, 1952)

Marginotruncana sigali (Reichel, 1950)

Planoheterohelix moremani (Cushman, 1938)

Planoheterohelix paraglobulosa (Georgescu and Huber, 2009)

Praeglobotruncana algeriana Caron, 1966

Praeglobotruncana oraviensis Scheibnerova, 1960

Rotalipora cushmani (Morrow, 1934)

Rotalipora montsalvensis (Mornod, 1950)

Rotalipora praemontsalvensis Ion, 1976

Rotalipora planoconvexa (Longoria, 1973)
Thalmanninella brotzeni Sigal, 1948

Thalmanninella deeckeii (Franke, 1925)

Thalmanninella globotruncanoides (Sigal, 1948)

Thalmanninella greenhornensis (Morrow, 1934)

Thalmanninella multiloculata (Morrow, 1934)

Whiteinella aprica (Loeblich and Tappan, 1961)

Whiteinella archaeocretacea Pessagno, 1967

Figure captions

TABLE 1. Source of planktonic foraminiferal bioevents and biostratigraphy, methodology used to process samples, $\delta^{13}$C$_{\text{carb}}$ and $\delta^{13}$C$_{\text{org}}$ profiles available in the literature for each section treated in this study.

TABLE 2. Mean depth and age of the events constrained by the age-depth model for the Pueblo section. Age of bentonites and of the C/T boundary (LO of $W$. devonense) are from Meyers et al. (2012). The ages of the other bioevents are calculated in this study.

TABLE 3. Depth and source of bio- and chemosratigraphic events identified at Pueblo, Eastbourne and Tarfaya that were used to perform the graphic correlations illustrated in Fig. 7.

FIGURE 1. Paleogeographic reconstruction for the late Cenomanian (94 Ma), with location of sections examined during this study (after Hay et al., 1999).
FIGURE 2. Pueblo, Colorado (Western Interior Seaway). The C/T boundary is placed between the top of the *N. juddii* and the base of the *W. devonense* Zone according to Caron et al. (2006). Lithostratigraphy is from Kennedy et al. (2005) and stratigraphic logs with position of bentonites are from Kennedy et al. (2005) and Caron et al. (2006). Bed numbers are according to Cobban and Scott (1972). Ammonite biostratigraphy is after Kennedy et al. (1999, 2000). $\delta^{13}$C$_{carb}$ profile and position of peak I, II and III are after Caron et al. (2006). The $\delta^{13}$C$_{org}$ profile of the nearby PU-79 core (Pratt and Threlkeld, 1984; Pratt, 1985) is correlated with the Rock Canyon outcrop using marker beds. Planktonic foraminiferal bioevents are after Eicher and Diner (1985), Leckie (1985), Leckie et al. (1998), Keller and Pardo (2004), Caron et al. (2006), Desmares et al. (2007) and Elderbak and Leckie (2016). The top of the *R. cushmani* Zone is placed according to Leckie (1985), while the base of the *H. helvetica* Zone is according to Elderbak and Leckie (2016), see text for further explanations. Calcareous nannofossil events are after Tsikos et al. (2004).

FIGURE 3. Eastbourne (UK). On the left: lithostratigraphy, planktonic foraminiferal biostratigraphy and $\delta^{13}$C$_{carb}$ profile (black) after Tsikos et al. (2004), age/stage and ammonite biostratigraphy after Gale et al. (2005). On the right: age/stage, lithostratigraphy, planktonic foraminiferal and ammonite biostratigraphy after Paul et al. (1999). The $\delta^{13}$C$_{carb}$ profile (grey) is according to Paul et al. (1999) and refers to the stratigraphic log on the right. Chemostratigraphic peaks (A, B, C) are after Jarvis et al. (2006) and Voigt et al. (2008) (see text for further details). The samples examined in this study refer to the stratigraphic log by Tsikos et al. (2004) in the left. Erosional basal surfaces are according to Keller et al. (2001). Planktonic foraminiferal events after
Paul et al. (1999), Keller et al. (2001), Hart et al. (2002), Tsikos et al. (2004) and this study.

Calcareous nannofossil events are according to Tsikos et al. (2004).

FIGURE 4. Planktonic foraminiferal specimens from the Eastbourne section. (1a–c) *Rotalipora cushmani*, sample GC-600 (0 m, base of the section). (2a–c) *Praeglobotruncana algeriana*, sample GC-260 (3.4 m). (3a–c) *Dicarinella hagni*, sample WC1240 (26.3 m). (4a–c) *Dicarinella imbricata*, sample GC-480 (1.2 m). (5a–c) *Helvetoglobotruncana praehelvetica*, sample GC-360 (2.4 m). (6a–c) *Dicarinella elata*, sample WC360 (17.5 m). (7a–c) *Thalmanninella brotzeni*, sample GC-340 (2.6 m). (8a–c) *Thalmanninella greenhornensis*, sample GC-260 (3.4 m). (9a–c) *Thalmanninella deeckeii*, sample GC-260 (3.4 m). (10a–c) *Praeglobotruncana oraviensis*, sample PM+280 (8.8 m).

Scale bar = 100 µm.

FIGURE 5. Planktonic foraminiferal specimens from Eastbourne and Tarfaya. Eastbourne: (1a–c) *Rotalipora montsalvensis*, sample GC-500 (1 m). (2a–c) *Rotalipora praemontsalvensis*, sample PM+240 (8.4 m). (3a–c) *Whiteinella archaeocretacea*, sample GC-540 (0.6 m). Tarfaya: (4a–c) *Helvetoglobotruncana praehelvetica*, sample S57/T58, 45–51 cm (depth 57.25 m). (5a–c) *Thalmanninella deeckeii*, sample S57/T57, 61–66 cm (depth 56.50 m). (6a–c) *Thalmanninella greenhornensis*, sample S57/T67, 9–14 cm (depth 59.55 m). (7a–c) “*Globigerinelloides bentonensis*”, sample S57/T59, 38–43 cm (depth 58.16 m). (8a–b) *Planoheterohelix moremani*, sample S57/T58, 45–51 cm (depth 57.25 m). (9a–b) *Planoheterohelix paraglobulosa*, sample S57/T58, 45–51 cm (depth 57.25 m). (10a–c) *Praeglobotruncana algeriana*, sample S57/T57, 61–66 cm (depth 56.50 m). (11a–c) *Dicarinella hagni*, sample S57/T57, 61–66 cm (depth 56.50 m).

Scale bar = 100 µm.
FIGURE 6. Tarfaya, core S57 (Morocco): planktonic foraminiferal biozonation, calcareous nannofossil events, position of the C/T boundary and δ¹³Corg profile after Tsikos et al. (2004) and Jenkyns et al. (2017). Planktonic foraminiferal and chemostratigraphic events according to this study.

FIGURE 7. Age-depth model for the Pueblo section. The age model is constrained by bentonite ages as calculated by Meyers et al. (2012). The linear functions obtained are as follows: 1) Bentonite A to Bentonite B (y=-9.5x+897.16); 2) Bentonite B to LO of W. devonense (y=-8.2353x+778.19); 3) LO of W. devonense to Bentonite C (y=-6.3636x+602.45); 4) Bentonite C to Bentonite D (y=-6.8462x+647.7).

FIGURE 8. Graphic correlations: 8a) Depth-depth plot of Pueblo vs. Eastbourne and 8b) depth-depth plot of Pueblo vs. Tarfaya. Please note that the depth of the LO of Q. gartneri at Tarfaya is represented with the error bar because its precise position is uncertain and likely falls within the coring gap (Tsikos et al., 2004).

FIGURE 9. Planktonic foraminiferal bioevents identified in each section plotted against a simplified δ¹³C profile. The methodology applied to study planktonic foraminifera (washed residues and/or thin sections) is indicated for each locality. Reliable bioevents are in green, potentially useful bioevents are in blue. Misleading bioevents include (1) ecologically controlled bioevents (purple), (2) unreliable bioevents because of taxonomic uncertainties, subjective species concepts and transitional evolution from ancestor species (red), and (3) possibly delayed appearances because of
species rare occurrence, low sampling resolution and/or small sample size (orange). Misleading bioevents are categorized according to the most important factor that in our opinion controlled species diachronism, in case multiple options are possible. The HO of *R. cushmani* at Clot Chevalier is in black, because its position is controlled by the sedimentologic features of the section. See text for references and discussion.

FIGURE 10. Summary of the most reliable sequence of planktonic foraminiferal bioevents for mid-low latitudes correlation across the C–T boundary interval and list of the bioevents that appear potentially useful but require further calibration in other localities. Reliable bioevents are numbered in stratigraphic order from the bottom to the top.

FIGURE 11. SEM images of the washed residues obtained from the rock samples collected at Eastbourne showing the composition of the assemblage in the <125 µm size-fraction with dominant calcispheres and rare biserial taxa. 1) Sample WC300 (16.9 m); 2) sample WC500 (18.9 m); 3) sample WC800 (21.9 m); and 4) sample WC1300 (26.9 m). Scale bar = 200 µm.

Supplementary materials

FIGURE A. Clot Chevalier section (Vocontian Basin, SE France): age/stage, lithostratigraphy, planktonic foraminiferal biostratigraphy and bioevents, δ¹³C_{carb} profile and chemostratigraphic events after Falzoni et al. (2016b).
FIGURE B. Pont d’Issole section (Vocontian Basin, SE France): the chemostratigraphic events are after Jarvis et al. (2006), with the exception of peak C (grey in the figure) that is herein placed based on the definition given in the text. The position of the C/T boundary is here estimated to fall within the interval from the estimated LO of *W. devonense* based on the bio- and chemostratigraphic correlation with Eastbourne (see Jarvis et al., 2011) and the LO of *H. helvetica* and includes peak C (as positioned in this study). Lithostratigraphy is according to Jarvis et al. (2006). Planktonic foraminiferal biostratigraphy and bioevents are from Grosheny et al. (2006). The $\delta^{13}C_{\text{carb}}$ profiles are from Grosheny et al. (2006) and Jarvis et al. (2006).

FIGURE C. Wadi Bahloul section (Tunisia): age/stage, lithostratigraphy, ammonite zonation, planktonic foraminiferal biostratigraphy and bioevents, $\delta^{13}C_{\text{carb}}$ profile and chemostratigraphic events (I, II, and III) from Caron et al. (2006). A, B, and C peaks are here placed according to the definition provided in this study.

FIGURE D. Gongzha section (Tibet): formations, planktonic foraminiferal biostratigraphy and bioevents, $\delta^{13}C_{\text{carb}}$ profile and chemostratigraphic events from Bomou et al. (2013). The position of the C/T boundary is here estimated to fall within the interval from peak C and the LO of *H. helvetica*. 
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<td>6.49</td>
<td><strong>93.66</strong></td>
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<td>Bentonite C</td>
<td>5.60</td>
<td>93.79</td>
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<tr>
<td>LO <em>W. devonense</em> - C/T boundary</td>
<td>4.90</td>
<td><strong>93.90</strong></td>
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<td>Bentonite B</td>
<td>3.50</td>
<td><strong>94.07</strong></td>
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<td>&quot;Heterohelix shift&quot;</td>
<td>2.80</td>
<td>94.14</td>
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<td>Bentonite A</td>
<td>1.60</td>
<td><strong>94.27</strong></td>
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<td>HO &quot;<em>G</em>. bentonensis&quot;</td>
<td>1.55</td>
<td>94.28</td>
</tr>
<tr>
<td>HO <em>Th. greenhomensis</em></td>
<td>1.45</td>
<td>94.29</td>
</tr>
<tr>
<td>HO <em>R. cushmani</em></td>
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<td>94.29</td>
</tr>
<tr>
<td>LO <em>H. prahelvetica</em></td>
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<tr>
<td>LO <em>D. canaliculata</em></td>
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<tr>
<td>LO <em>D. elata</em></td>
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<td>LO <em>D. imbricata</em></td>
<td>-0.75</td>
<td>94.51</td>
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<tr>
<td>LO <em>D. hagni</em></td>
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<td>94.55</td>
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<tr>
<td>LO <em>P. algeriana</em></td>
<td>-1.20</td>
<td>94.55</td>
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Table 2
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<th>Pueblo</th>
<th>Events</th>
<th>mean depth (m)</th>
<th>Source</th>
<th>Eastbourne</th>
<th>Events</th>
<th>mean depth (m)</th>
<th>Source</th>
<th>Tarfaya</th>
<th>Events</th>
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<td>HO Th. deekei</td>
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<td>Eicher and Diner (1985)</td>
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<td>8.30</td>
<td>this study</td>
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<td>HO 7h. greenhornensis</td>
<td>53.845</td>
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<td>HO R. cushmani</td>
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<td>Tsikos et al. (2004)</td>
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<td>δ¹³C peak A</td>
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<td>Pratt and Theelkold (1984)</td>
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<td>HO 7h. greenhornensis</td>
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<td>Leckie (1985)</td>
<td>HO 7h. greenhornensis</td>
<td>8.30</td>
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<td>HO A. albianus</td>
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<td>Tsikos et al. (2004)</td>
<td></td>
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<td>HO R. cushmani</td>
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<td>Leckie (1985)</td>
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<td>Tsikos et al. (2004)</td>
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<td>HO &quot;G.&quot; bentonensis</td>
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<td>Leckie (1985)</td>
<td>HO &quot;G.&quot; bentonensis</td>
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<td>this study</td>
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<td>onset &quot;Heterohelix shift&quot;</td>
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<td>δ¹³C peak C</td>
<td>44.61</td>
<td>this study</td>
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</tbody>
</table>

Table 3
Fig. 6

CENOMANIAN TURONIAN

P. algeriana
D. hagni
D. imbricata
A. albianus

R. cushmani

“Heterohelix shift” (50.16)
“G.” bentonensis (50.16)
R. cushmani (50.96)
Th. greenhornensis (53.96)
A. albianus
Th. deeckeii (54.16)
H. praehevetica (54.91)

Q. gartneri C/T boundary

C? B? A

R. cushmani W. archaeocretacea

C/T boundary

δ^{13}C_{org}
Fig. 7

Rock Canyon, Pueblo (Colorado)

Age (Ma)

depth (m)

- "Heterohelix shift"
- C/T boundary

- planktonic foraminifera
- ammonites
- bentonites
- bentonite A
- bentonite B
- bentonite C
- bentonite D
- LO H. helvetica
- LO M. marianosi
- HO "G." bentonensis
- HO Th. greenhornensis
- LO H. praehelvetica
- HO Th. deeckei
- LO W. devonense
- LO D. canaliculata
- LO D. elata
- LO D. imbricata
- LO P. algeriana
- LO D. hagni

Bentonite A
Bentonite B
Bentonite C
Bentonite D
LO H. helvetica
LO M. marianosi
HO "G." bentonensis
HO Th. greenhornensis
LO H. praehelvetica
HO Th. deeckei
LO W. devonense
LO D. canaliculata
LO D. elata
LO D. imbricata
LO P. algeriana
LO D. hagni

Fig. 7

Rock Canyon, Pueblo (Colorado)
Symbols

- Chemostratigraphic events
- Planktonic foraminiferal events used to calculate the regression line
- Planktonic foraminiferal events not used to calculate the regression line
- Calcareous nannofossil events used to calculate the regression line
- Calcareous nannofossil events not used to calculate the regression line
- Ammonite events

Lines of best-fit

- peak C at Eastbourne according to Voigt et al. (2008)
  base to top section
  \[ y = 2.3219x + 7.148; \quad R^2 = 0.92105 \]

- peak C at Eastbourne according to Jarvis et al. (2006)
  base to top section
  \[ y = 3.6921x + 6.2984; \quad R^2 = 0.97513 \]

- base section to HO "G." bentonensis and HO A. albianus
  \[ y = 1.4576x + 10.563; \quad R^2 = 0.96072 \]

- peak B to top section (C according to Voigt et al. 2008)
  \[ y = 2.4419x + 6.8633; \quad R^2 = 0.90849 \]

Fig. 8
<table>
<thead>
<tr>
<th>Region</th>
<th>Species and Events</th>
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<tr>
<td>Pueblo</td>
<td>M. renzi</td>
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<tr>
<td>Western Interior S.</td>
<td>M. marianosi</td>
</tr>
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<td>Anglo-Paris Basin</td>
<td>H. helvetica</td>
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<tr>
<td>Tarfaya Basin</td>
<td>R. planoconv.</td>
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<tr>
<td>Vocontian Basin</td>
<td>P. algeriana</td>
</tr>
<tr>
<td>Pont d'Issole</td>
<td>H. praehelv.</td>
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<tr>
<td>Vocontian Basin</td>
<td>D. canaliculata</td>
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<tr>
<td>Tethys</td>
<td>M. sigali</td>
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<tr>
<td>Gongzha</td>
<td>D. imbricata</td>
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<tr>
<td>Indian Ocean</td>
<td>D. elata</td>
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</table>

- **Heterohelix shift**
- **G. bentonensis**
- **G. bentonensis**
- **P. oraviensis**
- **R. cushmani**
- **Th. greenh.**
- **Th. deecke**
- **Th. brotzeni**
- **R. montsalv.**
- **R. praemontsalv.**
- **D. canaliculata**
- **D. elata**
- **D. hagni**
- **D. imbricata**
- **W. archaeocret.**
- **P. algeriana**
- **M. sigali**
- **M. renzi**

**Fig. 9**
<table>
<thead>
<tr>
<th>Most reliable sequence of bioevents</th>
<th>Potentially useful bioevents</th>
<th>simplified $\delta^{13}$C profile</th>
</tr>
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<tbody>
<tr>
<td>D. hagni (1)</td>
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<td>Th. deeckei (3)</td>
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<tr>
<td>Th. greenhornensis (4)</td>
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<tr>
<td>“G”. bentonensis (6)</td>
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</tr>
<tr>
<td>R. cushmani (5)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>M. schneegansi</td>
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<td>M. sigali</td>
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<tr>
<td>other marginotruncanids</td>
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</table>

Fig. 10: Diagram showing bioevents and age intervals.