

Global correlation of Upper Campanian– Maastrichtian successions using carbon-isotope stratigraphy: development of a new Maastrichtian timescale

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With 9 figures and 3 tables

Abstract. Carbon-isotope stratigraphy has proven to be a powerful tool in the global correlation of Cretaceous successions. Here we present new, high-resolution carbon-isotope records for the Global Boundary Stratotype Section and Point (GSSP) of the Maastrichtian stage at Tercis les Bains (France), the Bottaccione and Contessa sections at Gubbio (Italy), and the coastal sections at Norfolk (UK) to provide a global δ^{13} C correlation between shelf-sea and oceanic sites. The new δ^{13} C records are correlated with δ^{13} C-stratigraphies of the boreal chalk sea (Trunch borehole, Norfolk, UK, Lägerdorf-Kronsmoor-Hemmoor section, northern Germany, Stevns-1 core, Denmark), the tropical Pacific (ODP Hole 1210B, Shatsky Rise) and the South Atlantic and Southern Ocean (DSDP Hole 525A, ODP Hole 690C) by using an assembled Gubbio δ^{13} C record as a reference curve. The global correlation allows the identification of significant high-frequency δ^{13} C variations that occur superimposed on prominent Campanian-Maastrichtian events, namely the Late Campanian Event (LCE), the Campanian-Maastrichtian Boundary Event (CMBE), the mid-Maastrichtian Event (MME), and the Cretaceous-Paleogene transition (KPgE). The carbon-isotope events are correlated with the geomagnetic polarity scale recalculated using the astronomical ⁴⁰Ar/³⁹Ar calibration of the Fish Canyon sanidine. This technique allows the evaluation of the relative timing of base occurrences of stratigraphic index fossils such as ammonites, planktonic foraminifera and calcareous nannofossils. Furthermore, the Campanian-Maastrichtian boundary, as defined in the stratotype at Tercis, can be precisely positioned relative to carbon-isotope stratigraphy and the geomagnetic polarity timescale. The average value for the age of the Campanian-Maastrichtian boundary is 72.1 ± 0.1 Ma, estimated by three independent approaches that utilize the Fish Canyon sanidine calibration and Option 2 of the Maastrichtian astronomical timescale. The CMBE covers a time span of 2.5 Myr and reflects changes in the global carbon cycle probably related to tectonic process rather than to glacio-eustasy. The duration of the high-frequency δ^{13} C variations instead coincides with the frequency band of long eccentricity, indicative of orbital forcing of changes in climate and the global carbon cycle.

Key words. Carbon-isotope stratigraphy, Campanian, Maastrichtian

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1. Introduction

The latest Cretaceous witnessed the transition of the Earth's system from the mid-Cretaceous greenhouse to the cooler Cenozoic climate, accompanied by several perturbations in the global carbon cycle. Repeated intervals of climatic warming and cooling occurred superimposed on this Campanian-Maastrichtian longterm cooling trend (e.g. Sheldon et al. 2010). Distinct changes in foraminiferal δ^{18} O and δ^{13} C data, recognized on a global scale, indicate substantial deep-water cooling and reduced rates of organic-carbon burial across the Campanian-Maastrichtian transition (e.g. Barrera et al. 1997, Li and Keller 1998, Barrera and Savin 1999). Periods of intermittent warming occurred during the Maastrichtian, and are seen as initiators for the first wave of late Cretaceous extinctions (Frank and Arthur 1999, Keller 2001, Bralower et al. 2002). Furthermore, prominent, high-amplitude sea-level falls, recorded in both hemispheres, argue for a highly variable environmental system (Kominz et al. 2004, Crampton et al. 2006, Simmons et al. 2007). Our understanding of causal mechanisms is still limited during this time. At present, the spatial and temporal interactions of tectonic, climatic and oceanographic processes over large areas of the Cretaceous world are not well constrained, mainly because of the limitations of the available temporal framework. A critical element in developing a rigorous time scale is the formal identification of stage boundaries by the establishment of Global Stratotype Sections and Points (GSSPs) as an essential prerequisite for the global correlation of boundaries (Gradstein et al. 2004).

In 1984, a meeting in Copenhagen recognized that the base of the Maastrichtian Stage needed formal definition, and identified a series of six biostratigraphical datum levels (two based on ammonites, one based on a belemnite, two based on planktonic foraminifera, and one based on a nannofossil) that could possibly serve to mark the Campanian-Maastrichtian boundary (Birkelund et al. 1984), with three less suitable possible markers also proposed. In 2001, the abandoned quarry section at Tercis, near Dax in Landes, SW France was proposed as the GSSP for the Campanian-Maastrichtian boundary, defined by the arithmetic mean of the thickness of sediment encompassing eight of the nine previously proposed biostratigraphic datum levels (Odin and Lamaurelle 2001, Odin 2001). This level fell closest to the base occurrence of the ammonite Pachydiscus neubergicus, which therefore effectively marks the boundary, and

serves as a guide event. The GSSP was ratified in 2002.

Although selection of the base of the ammonite Pachydiscus neubergicus at Tercis provided a working definition of the Campanian-Maastrichtian boundary, correlation with other successions worldwide has not proved easy, partly due to insufficient preservation and inaccurate identification of microfossils, and partly due to the endemism and diachroneity of Campanian-Maastrichtian marine faunas of this age. The indifferent preservation and rarity of some microfossil taxa at Tercis means that the widely-used Tethyan biostratigraphical scheme, based on nannofossils (CC, NC, and UC divisions: Sissingh 1977, Perch-Nielsen 1985, Bralower et al. 1995, Burnett et al. 1998) and planktonic foraminifera (e.g. Premoli Silva and Sliter 1995, Huber et al. 2008), are difficult to apply in detail to the section containing the GSSP. Additionally, P. neubergicus itself shows striking diachroneity, in the order of at least 1 Myr, in its base occurrence relative to belemnite biostratigraphy across northern Europe, from the Ukraine to Denmark (Christensen et al. 2000). It is therefore presently impossible to identify accurately the base of the Maastrichtian in marine successions in many parts of the globe. For example, in ODP/IODP oceanic sites worldwide, and in the boreal chalk succession, extending from Central Asia to the UK, rarity, preservation, endemism and poor taxonomy of biostratigraphical marker renders precise correlation to the GSSP impossible. Furthermore, diagenetic alteration of calcite in the Tercis succession precludes the use of high-resolution strontium isotope stratigraphy (McArthur et al. 1992, 1993, 1994), and does not permit certain identification of magnetic chrons in the proximity of the boundary and above. The present situation not only hampers accurate correlation of successions, but limits the possibilities of developing an orbital timescale for this interval.

The problems of Campanian–Maastrichtian correlation, and specifically of the boundary itself, can be resolved by combining high-resolution carbon-isotope stratigraphy using bulk carbonate with the most recent and macro-, micro- and nannofossil zonations. This technique has proved successful in recent studies that show the Campanian–Maastrichtian boundary to be accompanied by a distinct negative carbon isotope excursion, known from the deep sea and the boreal chalk (Clauser 1994, Jarvis et al. 2002, Voigt et al. 2010, Thibault et al. 2012). Here, we present new, high-resolution carbon-isotope records for the GSSP at Tercis les Bains, France, the Bottaccione and Contes-

sa sections at Gubbio, Italy, and the coastal section at Norfolk, UK, and provide a correlation to the tropical Pacific (ODP Hole 1210B; Jung et al. 2012), the South Atlantic and Southern Ocean (DSDP Hole 525A and ODP Hole 690C: Li and Keller 1998, Friedrich et al. 2009) and boreal chalk sections (Trunch borehole, Norfolk, UK: Jenkyns et al 1994, Lägerdorf-Kronsmoor-Hemmoor, N-Germany: Stenvall 1997, Voigt et al 2010, Stevns-1 core, Denmark: Thibault et al. 2012). The new carbon-isotope records allow for the direct calibration of carbon-isotope events with palaeomagnetic reversals and the microfossil zonation of planktonic foraminifera and calcareous nannofossils at Gubbio (Gardin et al. 2012) and with the position of the Campanian-Maastrichtian boundary at Tercis.

We additionally provide an improved timescale for the Maastrichtian, by using three independent approaches. The first approach rests on the recalculation of palaeomagnetic reversals at Gubbio relative to the Fish Canyon sanidine calibration (Kuiper et al. 2008) and carbon-isotopic correlation of these ages to the GSSP section. The second approach uses the recalibrated late Campanian-Maastrichtian ⁴⁰Ar/³⁹Ar dates from Western Interior bentonites (Obradovich 1993) relative to the Fish Canyon sanidine, and their correlation to the GSSP section by use of inoceramid bivalves that occur in both successions (Walascyzcyk et al. 2002a). The third approach uses the astronomically tuned Maastrichtian time scale (Husson et al. 2011). Each approach gives a very similar age of the Campanian-Maastrichtian boundary and provides a temporal framework within which to consider carboncycle changes and related tectonic and climatic processes.

2. Methods

Stable-isotope measurements were performed in the stable-isotope laboratory at Oxford University for the sections from Tercis, Bottaccione and Norfolk and in the stable-isotope laboratory at Frankfurt University for samples from the Contessa section. In Oxford, samples were analysed isotopically for δ^{13} C and δ^{18} O using a VG Isogas Prism II mass spectrometer with an on-line VG Isocarb common acid bath preparation system. Samples were cleaned with hydrogen peroxide (H₂O₂) and acetone [(CH₃)₂CO] and dried at 60 °C for at least 30 minutes. In the instrument, they were reacted with purified phosphoric acid (H₃PO₄) at 90 °C. Calibration to V-PDB standard via NBS-19 was made daily using the Oxford in-house (NOCZ) Carrara marble standard.

In Frankfurt, stable carbon-isotope analyses of bulk carbonates were performed at a reaction temperature of 72 °C using a Finnigan MAT 253 with Gasbench. All isotope values are reported in ‰ V-PDB. The reproducibility of repeated standard measurements was better than 0.1‰ for both oxygen and carbon isotopes, in both laboratories. Data are available under http://doi.pangaea.de/10.1594/PANGAEA.778972.

The δ^{13} C records of the studied sections were correlated to each other using the software package Analy-Series 2.0 (Paillard et al. 1996). The assembled Gubbio δ^{13} C record was used as the reference for relative depth tuning, since it represents a deeper water depositional environment and is considered to reflect relatively small changes in sedimentation rate (see chapter 'Gubbio' for details). The correlation of the individual δ^{13} C records with Gubbio rests on the definition of δ^{13} C features as tie-points (Table 1). Most of the δ^{13} C events



Fig. 1. Palaeogeographic map of the early Maastrichtian (70 Ma) showing position of localities used in this study (after Hay et al. 1999).

Tie-point	Description
1	δ^{13} C rise towards the K/Pg boundary
2	δ^{13} C minimum a short distance below the C29r/C30n boundary
3a	δ^{13} C maximum within C30n, transition towards long-term δ^{13} C decline
3b	C30r/C31n boundary
4	δ^{13} C maximum within C31n, upper δ^{13} C max of the mid-Maastrichtian δ^{13} C plateau
5	inflection towards the lower δ^{13} C maximum of the mid-Maastrichtian δ^{13} C plateau, a short distance above the C31n/C31r boundary
6	inflection towards the long-term Maastrichtian δ^{13} C rise in C31r, top of the CMBE, top occurrence of <i>Tranolithus orionatus</i>
7	base of a δ^{13} C maximum at the C31r/C32n1n boundary, the top occurrence of <i>Broinsonia parca</i> subsp. is just below this point
8	δ^{13} C minimum at the C32n1n/C32n1r boundary, top occurrence of <i>Uniplanarius trifidus</i> is between tie- points 7 and 8
9a	base of a prominent δ^{13} C maximum in C32n2n, base of the CMBE
9b	δ^{13} C minimum at the C32n2n/C32r1r boundary, base occurrence of <i>Gansserina gansseri</i> is just below this point
10	δ^{13} C minimum in C32r1r
11	top of the LCE δ^{13} C minimum in C33n
12	base of the LCE δ^{13} C minimum, base occurrences of <i>Radotruncana calcarata</i> and <i>Uniplanarius trifidus</i> are below this point
13	inflection towards a δ^{13} C maximum, base occurrence of <i>Praediscosphaera stoveri</i> is above, and the base occurrence of <i>Belemnitella mucronata</i> is below this point

Table 1 Description of tie-points for carbon-isotope correlation.

are small scale and not unique in shape or magnitude. Therefore, the tie-points are further justified by palaeomagnetic reversals and their relative position to biostratigraphic events. Additional improvements in the carbon-isotope correlation are achieved by smaller adjustments between tie-points. Carbon-isotope correlation of the upper Campanian successions below the Late Campanian Event (LCE) rests on the δ^{13} C record of northern Germany (Lägerdorf-Kronsmoor) as reference for relative depth tuning. Finally, the relative depth scale was converted into time by using the recalculated ages of geomagnetic reversals based on the astronomical Fish Canyon sanidine calibration (Kuiper et al. 2008) above the Chron C32/C33 boundary and the cyclostratigraphic age model of Lägerdorf-Kronsmoor (Voigt and Schönfeld 2010) below the Chron C32/C33 boundary (Table 2).

3. Sections

3.1 Tercis les Bains, Landes, SW France

The stratotype section for the base of the Maastrichtian Stage at Tercis is situated in the Aquitaine Basin. During Late Cretaceous times, sedimentation occurred in a shallow to deep-shelfal depositional setting. The Tercis

Ages of geomagnetic reversals between the K/Pg
boundary and top Chron C33n, and of long eccen-
tricity cycles below top Chron C33 versus relative
depths (Voigt and Schönfeld 2010).

	Age (Ma) GTS 2004	Age (Ma) Kuipers et al. 2008	Age (Ma) Husson et al. 2011 (Option 2)
Base Chron			
K/Pg bdy	65.50	65.95	66.00
C29r	65.86	66.28	66.30
C30n	0/./U	08.13	68.20 68.20
C_{21n}	07.81	08.24 60.17	08.32 60.22
C31n	00.75	09.17	09.22
C_{22n1n}	70.90	71.41	71.40
C_{32n1r}	71.25	71.03	71.04
C_{32n2n}	72.03	73 30	73.60
C32n2n	73.23	73.70	73.90
C32r1n	73.32	73 79	74.00
C32r2r	73.58	74.05	74.10
405 kyr cycle			
Base UCa13		74.05	74.10
Base UCa8		76.07	76.13
Mid UCa2		78.30	78.35

section consists of an upper Campanian-lower Maastrichtian succession of marly and glauconitic limestones and flint-bearing limestones (Odin 2001). The biostratigraphy of Tercis was extensively discussed in a dedicated volume (Odin 2001). Accordingly, the Tercis section is zoned by ammonites (Küchler and Odin 2001, Odin et al. 2001b), inoceramids (Walaszczyk et al. 2002b), planktonic foraminifera (Odin et al. 2001a, Ion and Odin 2001, Arz and Molina 2001), calcareous nannofossils (Gardin et al. 2001, Melinte 2001, von Salis 2001, Gardin and Monechi 2001) and dinoflagellates (Antonescu et al. 2001). The Campanian-Maastrichtian boundary is defined as the arithmetic mean of twelve biotic datum levels and occurs at level 115.2 m (Odin and Lamaurelle 2001). The base occurrence of the ammonite *Pachydiscus neubergicus*, the preferred guide event for the base of the Maastrichtian Stage lies above, and the top occurrence of the ammonite Notoceras hyatti lies below this level. The preservation of micro- and nannofossils is strongly affected by diagenetic alteration, which complicates their unequivocal identification and results in the placement of different base and top occurrences at different levels in the section by different authors (Fig. 2). Therefore, it is difficult to apply the Tethyan planktonic and calcareous nannofossil zonation schemes to the Campanian-Maastrichtian boundary data at Tercis.

A first schematic stratigraphic log was published by Hancock (1993). More detailed lithological information was provided by Odin (2001). These logs are rather sketchy and do not include sedimentary details, such as burrows, graded bedding, fossil accumulation horizons or the size evolution of flint. Therefore, the Tercis section was re-measured and re-logged for this study (Fig. 2). Samples were taken at 0.5 m spacing in section PII, which is the reference section for levels 0 to 101 (Odin 2001), and from section PIV, the reference section for levels 100-162. The original marks, applied to indicate levels 0 to 162, are strongly weathered today and could only be detected at some horizons. All preserved marks are indicated on the new lithological log (Figs. 2, 3). No level numbers were detectable above Odin's level 123. Therefore, the levels are estimated from their relative position to marker beds in the upper part of the section. Of special importance are distinct marl-flint associations at levels 120 (107.8 m), 129 (117.4 m) and 136 (125.7 m), as well as strongly weathered beds (levels 140 to 142, below level 146, and between levels 149 and 152).

The sedimentary succession at Tercis is marl-dominated in the lower part (up to level 98, equals our 90.5 m), where it comprises alternating marls and marly limestones, as well as glauconitic fossiliferous limestones, and is limestone-dominated in the upper part, where white limestones with abundant flint bands and single marl seams occur (Figs. 2, 3). This facies change is interpreted as indicating the transition from an inner to outer shelf environment. The newly logged section reveals a series of sedimentary features that allow an interpretation of the succession in terms of filled and unfilled accommodation space, and thus, for relative sea-level change that can be set against carbon-isotope stratigraphy. Shallowing and progradation in the lower, marly, part of the succession is indicated by the increase in thickness of limestone beds, which also become progressively more fossiliferous, more enriched in glauconite and more intensely burrowed. Fine-grained detrital sediments were bypassed and deposited in more basinward settings. Retrogradational sedimentary stacking and increased accommodation is indicated by the development of distinct marl-limestone rhythms, which are poor in glauconite and macrofossils. The increase of accommodation space and landward displacement of facies belts favoured the deposition of detrital marls.

The shift towards the flint-yielding limestones above level 98 likely records a distinct landward displacement of the coastline and/or deepening of the basin, with deposition of detrital clays closer to the shoreline. The absence of clays generated a suitable facies for the diagenetic formation of chalcedonic quartz and ultimately of flint bands (Ehrmann 1986). Sediment progradation is suggested by the occurrence of distinct marl seams with increasing bed thickness, while retrogradation resulted in the deposition of pure limestones with flint formation. Increased current strength and winnowing can result in fossil accumulation in lags during maximum flooding when the site of highest sediment accumulation is displaced to the most landward position. According to this sedimentary model, seven retrogradation-progradation cycles can be identified in the sedimentary succession at Tercis (Fig. 3). The overall facies evolution indicates a progressive deepening of the basin, either by increased basinal subsidence, by rising sea level or by a combination of both.

3.2 Gubbio, Umbria-Marche, Italy

A Campanian–Maastrichtian pelagic succession of pink micritic limestone is exposed in the Bottaccione Gorge and along the Contessa Highway at Gubbio, Umbria–Marche Basin, Italy. The two sections are





mark the sample positions for this study.

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Fig. 3. Campanian–Maastrichtian carbon-isotope stratigraphy, biostratigraphic datum levels, biozones and sedimentary cycles of the Global Boundary Stratotype and Section Point at Tercis les Bains, SE France. (1) Lithostratigraphic units are from Odin (2001). (2) Ammonite zones and data are from Küchler and Odin (2001) and Odin et al. (2001b). (3) Inoceramid zones and datum levels are from Walaszczyk et al. (2002b). (4) Nannofossil zones and datum levels are from Melinte and Odin (2001)^a, von Salis (2001)^b, Gardin and Monechi (2001)^c and Gardin et al. (2001)^d. (5) Planktonic foraminiferal zones and datum levels are from Odin et al. (2001a), Ion and Odin (2001)^e, and Arz and Molina (2001)^f, and, (6) Palaeomagnetic data are from Lewy and Odin (2001). Numbers and dots beside the lithological log mark levels of Odin (2001). Numbers in italics are estimated according to their relative positions on Odin's log (2001; see text). Sedimentary cycles indicate progradational (P) and retrogradational (R) stacking patterns (this study). Grey shaded areas mark the Late Campanian (LCE) and Campanian–Maastrichtian (CMBE-1 to CMBE-5) carbon-isotopic events (see text).

considered to represent an environment of continuous pelagic deposition, although intrastratal folds and normal faults occur locally (Arthur and Fischer 1977). The Bottaccione section has been studied since the eighteen-hundreds and became famous for its record of the organic-rich black shale (Livello Bonarelli) at the Cenomanian–Turonian boundary and for the iridium anomaly at the Cretaceous–Palaeogene (K/Pg) boundary (Bonarelli 1891, Alvarez et al. 1977, Alvarez 2009, Bernoulli and Jenkyns 2009). Detailed logging and magnetostratigraphic work was performed by Arthur and Fischer (1977), Lowrie and Alvarez (1977, 1981), Channell et al. (1978) and Lowrie et al. (1982). Calcareous nannofossils and planktonic foraminifera provide a biostratigraphic context (Monechi and Thierstein 1985, Premoli Silva and Sliter 1995). More recently, the upper Campanian–Maastrichtian Contessa Highway section has been studied as a complementary section (Gardin et al. 2012). Logging, biostratigraphic and geomagnetic data were provided by Chauris et al. (1998) and Gardin et al. (2012). The two sections at Gubbio serve as a subtropical reference.

Upper Campanian to Maastrichtian strata of the Bottaccione and Contessa sections were sampled every 25 cm for carbon-isotope stratigraphy. The upper Maastrichtian part of Contessa is not accessible, so sampling of this section was performed from the base occurrence of the planktonic foraminifera *Radotruncana calcarata* (285 m, Chron 33n) up to 366 m (within Chron C31n; Fig.4). The Bottaccione section was sampled from the K/Pg boundary down to the base of the *R. calcarata* Zone (373–285 m).



Fig. 4. Upper Campanian–Maastrichtian carbon-isotope stratigraphy, geomagnetic polarity reversals and biostratigraphic datum levels (base and top occurrences of calcareous nannofossils and planktonic foraminifera) of the Bottaccione Gorge and Contessa Highway sections at Gubbio, Umbria–Marche Basin, Italy. Palaeomagnetic data are from Arthur and Fischer (1977), Lowrie and Alvarez (1977, 1981) and Gardin et al. (2012). Planktonic foraminifera and calcareous nannofossil datum levels are from Monechi and Thierstein (1985), Premoli Silva and Sliter (1995), and Gardin et al. (2012). Arrows mark the transition between Contessa and Bottaccione sections for the composite Gubbio δ^{13} C curve. Grey shaded areas mark the Late Campanian (LCE), Campanian–Maastrichtian (CMBE-1 to CMBE-5), Mid-Maastrichtian (MME-1 to MME-3) and Cretaceous-Palaeogene (KPgE-1 to KPgE-3) carbon-isotopic events (see text).

Between chrons C31n and C32r, the δ^{13} C curves of both sections show similar features and can be precisely correlated relative to the established geomagnetic zonation, as well as to nannofossil and planktonic foraminiferal events (Fig. 4). Results of geomagnetic correlation indicate a 10 m hiatus in Chron C31n in the Bottaccione section (353 m, Fig. 4) that is represented by the 353 m to 363 m interval at Contessa (Gardin et al. 2012). The loss of section at Bottaccione is related to faulting and the δ^{13} C correlation confirms the stratigraphic position of this gap.

Below Chron C32r the correlation of the two sections is more difficult. A prominent negative δ^{13} C excursion that occurs between the base and top of *R.calcarata* at Contessa is missing at Bottaccione. This stratigraphic mismatch indicates another gap in the Bottaccione section, and it is likely that parts of the succession were lost during Neogene faulting and overthrusting. To obtain the most complete δ^{13} C record, the δ^{13} C curves of Bottaccione and Contessa were assembled using the Contessa record for the upper Campanian–lower Maastrichtian part (below 365.75 m) and the Bottaccione record for the upper Maastrichtian part (above 357.5 m; Fig.4).

3.3 Trunch borehole and coastal outcrops, Norfolk, eastern England

A cored borehole through the entire Cretaceous Chalk at Trunch in north-eastern Norfolk (Wood et al. 1994, Jarvis et al. 2002, 2006) provides a high-resolution carbon and oxygen isotope record for the Cenomanian-lowermost Maastrichtian interval, and the $\delta^{13}C$ data for the upper Campanian to Maastrichtian interval are replotted here (Fig. 5). The strontium-isotope stratigraphy of the succession has been published by McArthur et al. (1993). The nannofossil biostratigraphy, originally worked out by Burnett (1990), was updated in this study to reflect the zones of Burnett et al. (1998). The Maastrichtian part of the core overlaps the succession exposed on the Norfolk coast, between 3 and 8 km north and north-west of Trunch. Here, Maastrichtian chalk masses, emplaced in glacial till, are exposed in cliffs and foreshore adjacent to the villages of Sidestrand, Overstrand and Trimingham, east of Cromer. General descriptions have been published by Brydone (1906) and Peake and Hancock (1961, 1970), who provided a detailed lithological log and biostratigraphical data. Biostratigraphies have been provided by belemnites (Schulz 1979), brachiopods (Johansen and Surlyk 1990) and benthic foraminifera

(Hart and Swiecicki 2003). Jenkyns et al. (1994) published an outline carbon- and oxygen-isotope curve, based on only a few samples. The lithostratigraphy was revised by Mortimore et al. (2001), who identified the approximate position of the Campanian–Maastrichtian boundary in the large chalk mass at Overstrand, based on belemnite biostratigraphy. A new correlation of the successions in the various erratic blocks at Overstrand, Sidestrand and Trimingham is provided as a new composite succession of just over 35 m of chalk in Fig. 5. Samples were collected every 0.25 m for isotopic analysis. The highest 10 m of succession (Beacon Hill Grey Chalk Member, *sumensis* Belemnite Zone) are not exposed and could not be sampled.

3.4 Lägerdorf-Kronsmoor and Hemmoor quarries, northern Germany

The Coniacian-Maastrichtian chalk successions of Lägerdorf-Kronsmoor and Hemmoor are the standard outcrop sections for the boreal white chalk of northern Germany and have been the subject of stratigraphic research for more than five decades. The 570 m-thick Coniacian-Maastrichtian succession is composed of foraminifera-bearing nannofossil chalk with carbonate content > 85–95% (Ehrmann 1986, Schulz et al. 1984, Voigt and Schönfeld 2010). Biostratigraphies using belemnites and echinoids, brachiopods, benthic foraminifera, and calcareous nannofossils have been established (Ernst and Schulz 1974, Schulz 1978, Schulz et al. 1984, Surlyk 1984, Burnett 1990, McLaughlin et al. 1995, Schönfeld et al. 1996, Niebuhr et al. 2011), as has a strontium-isotope stratigraphy (McArthur et al. 1992, McLaughlin et al. 1995). The application of the macrofossil, and in part the nannofossil zonation (Campanian and upper Maastrichtian), are confined to the higher latitudes of northern Europe a (e.g. Schönfeld and Burnett 1991, Christensen 1999). The Coniacian-lower Maastrichtian carbon-isotope stratigraphy of the Lägerdorf-Kronsmoor section can be correlated with accuracy to the δ^{13} C curve of the Trunch borehole (Voigt et al. 2010) and represents, together with the Maastrichtian δ^{13} C curve of Hemmoor (Stenvall 1997), the first nearly continuous carbon-isotope record of the boreal chalk for the last 20 million years of the Cretaceous. Originally, the Kronsmoor and Hemmoor sections were considered to overlap, because of a correlation of a migration event of belemnellids from eastern Europe (Schulz et al 1979). However, the δ^{13} C record indicates an exposure gap within the lower Maastrichtian sumensis Zone (Fig. 6; Voigt et al. 2010).



3.5 Stevns-1 core, eastern Denmark

A 450 m-thick succession of chalk was drilled in 2005 adjacent to the coastal cliff of Stevns Klint (Stevns-1 core) from the Lower Danian down into the upper Campanian of the Upper Cretaceous Chalk Group (Stemmerik et al. 2006). The core represents one of the most expanded uppermost Campanian–Maastrichtian successions worldwide and in particular through the chalk of NW Europe (compare with Surlyk and Birkelund 1977, Bailey et al. 1983, Schulz et al. 1984, Villain 1988, Schönfeld et al. 1996, Surlyk and Håkansson 1999). The nannofossil biostratigraphy of the core has been published by Sheldon (2008). High-resolution carbon-isotope stratigraphy was performed by Schovsbo et al. (2008) and Thibault et al. (2012).

3.6 ODP Hole 1210B, tropical Pacific Ocean

In the tropical Pacific, 130 m of upper Campanian-Maastrichtian sediments were recovered from Shatsky Rise at ODP Hole 1210B (Bralower et al. 2002). Based on palaeotectonic reconstructions, Shatsky Rise was situated in tropical latitudes (~10° N) during the Campanian-Maastrichtian interval (Fig.1; Larson et al. 1992). The sediments from Hole 1210B are interpreted as being deposited at upper abyssal (2500 m) water depths during the Maastrichtian (Kaiho 1999). The sediments consist of foraminifera-bearing nannofossil ooze and chalk, with repeated levels of chert layers. The biostratigraphy of Hole 1210B is based on planktonic foraminifera (Bralower et al. 2002, Petrizzo et al. 2011) and calcareous nannofossils (Lees and Bown 2005). High-resolution carbon-isotope stratigraphy was performed by Jung et al. (2012).

3.7 DSDP Hole 525A, South Atlantic and ODP Hole 690C, Southern Ocean

In the South Atlantic and Southern oceans, Maastrichtian calcareous chalks and oozes were recovered from the Walvis Ridge at DSDP Hole 525A (Moore and Rabinowitz et al., 1984). The Walvis Ridge is an aseismic volcanic ridge between Africa and the Mid-Atlantic Ridge that was situated, during the Maastrichtian at 36° S palaeolatitude at a water depth of approximately 1,000–1,550 m (Moore et al. 1984, Li and Keller 1999). A biostratigraphy of Hole 525A has been published, using planktonic foraminfera (Li and Keller 1998) and calcareous nannofossils (Manivit 1984, Henriksson 1993, Thibault and Gardin 2007). Palaeomagnetic data are provided by Chave (1984). Carbon-isotope records of benthic foraminifera have been published by Li and Keller (1998) and Friedrich et al. (2009).

ODP Hole 690C recovered uppermost Campanian– Maastrichtian sediments on the southwestern flank of Maud Rise, an aseismic volcanic ridge located in the eastern Weddell Sea (Fig. 1; Barker and Kennett et al., 1990, Barrera and Savin, 1999). The succession was deposited at a palaeolatitude of about 65 °S in an estimated water depth of 1,800 m (Huber 1990, Thomas 1990). Biostratigraphies have been published, based on planktonic foraminifera and calcareous nannofossils (Barrera and Huber 1990, Huber 1990, Pospichal and Wise 1990); and magnetostratigraphic investigations indicate an interval between chrons C32n and C31r (Hamilton 1990). A high-resolution carbon-isotope record of the benthic foraminiferan *Gavelinella* was published by Friedrich et al. (2009).

4. Carbon-isotope events

4.1 Late Campanian Event (LCE)

The Late Campanian Event (LCE) is characterized by a distinct negative carbon-isotope excursion, of variable magnitude, that lasted ~450 kyr (Figs. 3–7). The excursion is well represented in the shelf-sea sections at Trunch and Lägerdorf (both with a magnitude of 0.4%), Gubbio (magnitude of 0.3%) and Tercis (magnitude of 0.8%). The LCE carbon-isotope excursion is not distinctly developed in the tropical Pacific (Hole 1210B, Fig. 7). At Tercis, the LCE is accompanied by glauconitic fossiliferous limestones (levels 63 to 68)

Fig. 5. Carbon-isotope correlation of the composite section of glacial masses along the Norfolk coast (this study) and the Trunch borehole (Jenkyns et al. 1994, Jarvis et al. 2002). Biostratigraphic data at Norfolk are based on belemnites (Schulz 1979), brachiopods (Johansen and Surlyk 1990), and benthic foraminifera (Swiecicki 1980, Hart and Swiecicki 2003). The lithostratigraphy is by Mortimore et al. (2001). The labeling of flint bands ('B' to 'W') was performed by Peake and Hancock (1961, 1970). Trunch stratigraphy is based on the compilation by Jarvis et al. (2002), with data of Burnett (1990), Wood et al. (1994), Peake and Hancock (1970). Abbreviations: *la-pobt – lanceolata-pseudobtusa* Zone, *obt – obtusa* Zone, *sum – sumensis* Zone, T – Trimingham Member, LMP – Little Marl Point Member, LOM – Lower Overstrand Marl, UOM – Upper Overstrand Marl, *Bl – Belemnella lanceolata*.



and represents the uppermost part of a shallowing-upward sequence (Fig. 3). At Gubbio, Trunch and Lägerdorf, no distinct sedimentary features can be observed.

Biostratigraphically, the LCE excursion lies in the Radotruncana calcarata Zone at the Contessa Highway section in Gubbio, but above the R. calcarata Zone at Tercis (see discussion chapter 5.3). At Gubbio, the base occurrences of Uniplanarius trifidus and *R. calcarata* lie below the LCE, and the top occurrence of *R*. calcarata above it. The base occurrence of Globotruncanella havanensis occurs within the LCE (Fig. 4). At Tercis, the base of U. trifidus is, similarly to Gubbio, a short distance below the LCE, and the base occurrence of G. havanensis lies within the LCE (Fig. 3). Furthermore, the minimum of the LCE carbon-isotope excursion corresponds to the base occurrences of the inoceramid "Inoceramus" altus and the ammonite N. hyatti at Tercis (Walaszczyk et al. 2002b, Küchler and Odin 2001). In northern Germany, the LCE lies around the boundary of the Galerites vulgaris and Nostoceras (Bostrychoceras) polyplocum zones (Schulz 1978, Schulz et al. 1984). The base occurrence of *N*. *polyplocum* is associated with the δ^{13} C minimum of the LCE at Lägerdorf.

4.2 Upper Campanian interval between the LCE and CMBE

Between the carbon-isotope events LCE and CMBE lies a plateau-like interval of relatively stable δ^{13} C values with a variability of less than 0.2% (Figs. 3–7). This interval encompasses upper Chron C33n to lower Chron C32n2n (Fig. 4). At Gubbio and at Hole

1210B, the top occurrence of *R. calcarata* falls in the lower part of this interval and thus appears to be isochronous (Figs. 4, 7). A small negative inflection is visible in the middle part of this interval at Trunch, Lägerdorf, Tercis, Gubbio and Site 1210B and corresponds to the lower part of Chron C32r2r (correlation point 10 in Fig. 6). At Gubbio, the base occurrence of *Globotruncana aegyptiaca* lies a little way below the inflection at Contessa and corresponds to it at Bottaccione. The top occurrence of the nannoconid group of nannofossils lies a little way above this inflection point at Tercis (Fig. 3), where the base occurrence of *"I." oblongus* is found at the same level.

4.3 Campanian-Maastrichtian Boundary Event (CMBE)

The CMBE is a prominent negative carbon-isotope excursion that lasted around 2.5 Myr and is recorded from shelf-sea settings (e.g. Lägerdorf-Kronsmoor: Voigt et al. 2010, Stevns-1 core: Thibault et al. 2012) and oceanic sites in the Pacific, Southern and South Atlantic oceans (e.g. Site 305, Hole 525A, Hole 690C, Site 869: Larson and Moberly et al. 1975, Jenkyns et al. 1995, Frank and Arthur 1999, Barrera and Savin 1999, Friedrich et al. 2009, Voigt et al. 2010). The magnitude of the CMBE carbon-isotope excursion (CMBE-CIE) is highly variable and ranges from 0.3 to 0.4‰ at Gubbio, Tercis and 1210B to values up to 1.0% at Kronsmoor, Norfolk coast, Stevns-1 and in the bottom-waters of Holes 690C (Maud Rise) and 525A (Walvis Ridge) (Figs. 6, 7). Furthermore, the CMBE-CIE displays superimposed detailed structure

Fig. 6. Correlation of the upper Campanian–Maastrichtian δ^{13} C records of Norfolk (coastal outcrops and Trunch borehole), Lägerdorf-Kronsmoor-Hemmoor (northern Germany; Voigt et al. 2010), Stevns-1 core (Thibault et al. 2012), Tercis and Gubbio, showing the Late Campanian (LCE), Campanian-Maastrichtian (CMBE-1 to CMBE-5), Mid-Maastrichtian (MME-1 to MME-3) and Cretaceous-Palaeogene (KPgE-1 to KPgE-3) carbon-isotopic events (shaded grey). Subsidiary carbon-isotope events are indicated by grey stippled lines. All δ^{13} C records are plotted against age after they have been tuned to the relative depth scale of Gubbio (between the K/Pg boundary and the LCE) and Lägerdorf (below the LCE). Encircled numbers show the position of tie-points used for correlation (see Table 1 for explanation). The age model is based on the ages of geomagnetic reversals, which have been recalibrated to the astronomically calibrated age of the Fish Canyon sanidine (Kuiper et al. 2008). Below Chron C32, age control is provided by long eccentricity cycles detected in the Lägerdorf section (Voigt et al. 2010). Anchor points are the base of Chron C32, the base of a small δ^{13} C maximum below the LCE in the vulgaris Zone, and the base of the δ^{13} C peak lying just above the base of the basiplana/spiniger Zone at Lägerdorf. References: 1) Arthur and Fischer (1977); 2) Gardin et al. (2012); 3) Premoli-Silva and Sliter (1995); 4) Monechi and Thierstein (1985); 5) Küchler and Odin (2001); 6) Walaszczyk et al. (2002b); 7) Gardin et al. (2001); 8) Schulz (1979); 9) Burnett (1990); 10) Sheldon et al. (2010); 11) Schulz et al. (1984). Abbreviations: Rc - Radotruncana calcarata, Gh - Globotruncanella havanensis, Ga - Globotruncana aegyptiaca, Rf - Racemiguembelina fructicosa, Ph - Plummerita hantkeninoides,la – lanceolata, obt – obtusa, c/s – conica/senior, bas/spin – basiplana/spiniger, polyp – polyplocum, gr/gr – grimmensis/ granulosus, c - cimbrica, f - fastigata, tj - tegulatus/junior, ar/jun - argentea/junior, bd - baltica/danica, E.e. - Eiffellithus eximius, U.t. – Uniplanarius trifidus, B.p.c. – Broinsonia parca constricta, T.o. – Tranolithus orionatus, L.q. – Lithraphidites quadratus, M.m. – Micula murus, M.p. – Micula prinsii.



in the form of five small-scale positive excursions. Two of them are prominent and numbered here as peaks CMBE-1 and CMBE-4 (Figs. 6, 7). The maximum of CMBE-1 marks the onset of the CMBE-CIE, while CMBE-4 lies in the minimum trough of the CMBE. The peaks of CMBE-2 and CMBE-3 are not distinctly developed in any section. Peak CMBE-2 lies at the interval of major $\delta^{13}C$ decrease of the CMBE and is suppressed in some cases (e.g. Gubbio, Norfolk, Tercis). Peak CMBE-3 lies on the rise towards peak CMBE-4 and is well developed in Tercis, Gubbio, northern Germany, and Site 305 and Hole 1210B. The small-scale positive excursions appear as cyclic 400-500-kyr-long variations during the long-term δ^{13} C decrease and may record orbitally forced changes in the global carbon cycle.

Within the CMBE lies the Campanian–Maastrichtian boundary. Its position, in the GSSP at Tercis, lies above the CMBE δ^{13} C decrease (between peaks CMBE-2 and 3; Figs. 3, 6, 7). The base occurrence of the belemnite *Belemnella lanceolata*, the boreal index fossil for the base of the Maastrichtian (Schulz 1979), corresponds to peak CMBE-1 in Kronsmoor and Norfolk and is of latest Campanian age, with reference to the GSSP. Other important biostratigraphic markers associated with the CMBE are as follows (Figs. 3–7):

- The base occurrence of *Gansserina gansseri* (planktonic foraminifera) lies below the onset of the CMBE at Gubbio and Hole 1210B; however, the base of *G.gansseri* at Hole 525A is in CMBE-5 (Fig. 7).
- The top occurrence of *Nostoceras hyatti* (ammonite) lies on the δ¹³C decrease associated with peak CMBE-2 at Tercis, just below the Campanian–Masstrichtian boundary (Fig. 3).

- The base occurrence of *Pachydiscus neubergicus* (ammonite) is between peaks CMBE-2 and 3 at Tercis (Fig. 3).
- The base occurrences of the inoceramids "Inoceramus" redbirdensis and Endocostea typica are associated with the base of CMBE-2 and CMBE-3, respectively, at Tercis (Fig. 3).
- The top occurrence of *Uniplanarius trifidus* lies between peaks CMBE-3 and 4 at Tercis (Fig. 3), Gubbio (Fig. 4), and Hole 1210B (Fig. 7).
- The top occurrence of *B.p. constricta* lies between peaks CMBE-4 and 5 at Tercis (Fig. 3), within CMBE-5 at Gubbio (Fig. 4), and Stevns-1 (Fig. 6), but in CMBE-3 at Hole 1210B (Fig. 7).
- The top of *Tranolithus orionatus* lies at the end of the CMBE-CIE at Gubbio (Fig.4) Stevns-1 (Fig.6) and Site 1210B (Fig.7), and in CMBE-5 at Kronsmoor (Fig.6).

Relative to the palaeomagnetic polarity scale of Gubbio, the CMBE encompasses mid-Chron C32n2n to mid-Chron C31r. The Campanian–Maastrichtian boundary lies within the uppermost Chron C32n.2n.

4.4 Mid-Maastrichtian Event (MME)

Above the CMBE, carbon-isotope values rise towards high values around the mid-Maastrichtian. The δ^{13} C increase is more prominent than the δ^{13} C decrease during the CMBE. It has an overall magnitude of 0.6% to 1.5% and appears as a two-step rise. Most of the δ^{13} C increase occurred in the first phase in less than 0.5 Myr as shown in the records of Gubbio, Stevns-1 (Fig.6), Hole 525A and Hole 1210B (Fig.7). In the mid-Maastrichtian, the δ^{13} C values form a plateau-like high (Mid-

Fig. 7. Correlation of upper Campanian–Maastrichtian δ^{13} C records of oceanic sites to Gubbio, showing the Late Campanian (LCE), Campanian-Maastrichtian (CMBE-1 to CMBE-5), Mid-Maastrichtian (MME-1 to MME-3) and Cretaceous-Palaeogene (KPgE-1 to KPgE-3) carbon-isotopic events (shaded grey). The δ^{13} C data from Hole 1210B (Shatsky Rise) are derived from bulk carbonate (Jung et al. 2012) and the δ^{13} C records from Hole 525A and Hole 690C are derived from benthic foraminifera (Li and Keller 1998, Friedrich et al. 2009). Encircled numbers show the position of tie-points used for correlation (Table 1). Black triangles mark the position of documented magnetic reversals. The age model is based on the ages of geomagnetic reversals recalibrated to the astronomically calibrated age of the Fish Canyon sanidine (Kuiper et al. 2008). References: 1) Premoli Silva and Sliter (1995); 2) Gardin et al. (2012); 3) Monechi and Thierstein (1985); 4) Huber (1990); 5) Huber (1992); 6) Pospical and Wise (1990); 7) Manivit (1984); 8) Henriksson (1993); 9) Thibault and Gardin (2007); 10) Li and Keller (1998); 11) Lees and Bown (2005); 12) Bralower et al. (2002). Abbreviations: Rc - Radotruncana calcarata, Gh – Globotruncanella havanensis, Ga – Globotruncana aegyptiaca, Gg – Gansserina gansseri, Rf – Racemiguembelina fructicosa, Ph – Plummerita hantkeninoides, Cc – Contusotruncana contusa, Pi – Pseudotextalria intermedia, Pha – Pseudoguembelina hariaensis, Pp – Pseudoguembelina palpebra, E.e. – Eiffellithus eximius, U.t. – Uniplanarius trifidus, B.p.c. – Broinsonia parca constricta, T.o. – Tranolithus orionatus, L.q. – Lithraphidites quadratus, M.m. – Micula murus, M.p. – Micula prinsii, B.c. – Biscutum coronum, B.m. – Biscutum magnum, A.m. – Abathomphalus mayaroensis, G.g. – Gansserina gansseri, C.c. – Contusotruncana contusa.

Maastrichtian Event, MME), that lasted ~570 kyr and comprises a lower and an upper maximum, with a negative inflection in between at Hemmoor, Stevns-1, Gubbio and Site 1210B (MME Peaks 1-3; Figs. 6, 7). In terms of boreal macrofossil biostratigraphy, the MME covers the cimbrica, fastigata and lower half of the tegulatus/junior belemnite zones. The rise towards the upper maximum starts at the base of the tegulatus/ junior Zone and correlates with a major transgression in the North Sea and adjacent basins (Niebuhr 1995). In terms of nannofossil biostratigraphy, the MME corresponds to the interval between the top occurrence of Reinhardtites levis to, possibly, the base of Lithraphidites quadratus (Sheldon 2008), although the base of L. quadratus is recorded in Hemmoor above the MME (McLaughlin et al. 1995).

At Gubbio, the base occurrence of *L.quadratus* corresponds to the maximum of peak MME-1, which is below its base occurrence in the higher latitudes. The base occurrence of the planktonic foraminifer *Abathomphalus mayaroensis* is recorded below the MME at Gubbio (Fig. 4) but above the MME at Hole 1210B (Fig. 7). In the South Atlantic Ocean, the base of *Abathomphalus mayaroensis* is much older and is related to CMBE-5 at Hole 525A (Huber 1990, 1992). The MME lies just above the base of Chron C31n.

4.5 Interval between the MME and the K/Pg boundary

Above the MME, the δ^{13} C records show a long-term decrease towards the K/Pg boundary with superimposed short-term variations (Figs. 4, 6, 7). One prominent short-term feature is a positive inflection that marks a transition from gently decreasing, towards more steeply decreasing, δ^{13} C values. The inflection is located in the argentea/junior Zone of the boreal macrofossil biostratigraphy and corresponds to an acme occurrence of Nephrolithus frequens at Stevns-1 (Thibault et al. 2012), as well as the base of N.frequens at Hemmoor (McLaughlin et al. 1995). At Hole 1210B, the δ^{13} C inflection is associated with the base of Ceratolithoides kamptneri (Lees and Bown 2005; Fig. 7). At Gubbio, the δ^{13} C inflection is represented by a distinct positive peak that corresponds to the base of Micula murus within Chron C30n. The base occurrence age of *M.murus* is delayed at Gubbio and is associated with MME-Peak 2 (Chron C31n) in the tropical Pacific.

Further $\delta^{13}C$ features are two negative-positive oscillations in the uppermost Maastrichtian (KPg-1 to

KPg-3) that are well developed at Stevns-1 (Figs. 6, 7), Hemmoor (Fig. 6), Gubbio (Figs. 4, 6, 7), Hole 525A (Fig. 7) and in part at Hole 1210B (Fig. 7) as well. The correlation of the Hemmoor section remains uncertain; this section does not reach the K/Pg boundary, so it remains unclear whether the prominent δ^{13} C negative excursion represents the lower or the upper of the two negative oscillations. The lower of the two negative peaks lies in uppermost Chron C30n, with the C29r/ C30n boundary related to the δ^{13} C increase of KPg-2. The K/Pg boundary lies within the second δ^{13} C maximum at Gubbio, Stevns-1 and at Hole 525A above KPg-3 (Figs. 4, 6, 7). The uppermost Maastrichtian of Hole 1210B falls within a hiatus and only the lower δ^{13} C minimum of KPg-1 is preserved. The base of Micula prinsii is associated with the KPg-1 minimum at Hole 1210B (Fig. 7) and at Gubbio (Figs. 4, 6, 7). In the South Atlantic at Hole 525A, the top occurrence of G. gansseri is related to KPg-2.

5. Discussion

5.1 Significance of carbon-isotope correlation

Carbon-isotope stratigraphy has proven to be a powerful correlative tool for Cretaceous marine sediments in those instances where the δ^{13} C curves have characteristic shape and form (e.g. Scholle and Arthur 1980, Jenkyns et al. 1994, Voigt et al. 2000, Jarvis et al. 2006). However, the results of this study also demonstrate the limitations of the method. Changes in Campanian-Maastrichtian global carbon cycling are small, causing only minor changes in the inorganic δ^{13} C record. The overall δ^{13} C variability can be lower than 0.5% (e.g. Gubbio) and does not exceed 1.0% (Norfolk, northern Germany, Holes 525A and 690C). None of the above-described δ^{13} C features is diagnostic in terms of absolute value, magnitude of excursion or position of a short-term event on a long-term trend. The absolute value is, to some degree, influenced by burial diagenesis and thus the degree of lithification. In general, increased burial diagenesis causes a dampening of the amplitudes of δ^{13} C variability (Schrag et al. 1995). Repeated lithological changes support processes related to differential diagenesis, which might dampen or amplify primary carbon-isotopic changes. Prominent processes include the differential alteration of carbonate species, variable degree of organic-matter degradation, or changes in porosity and permeability that modify pathways for diagenetic fluid migration. The correlation between Gubbio and Holes 525A and 690C is, therefore, independently justified by palaeomagnetic data (Fig. 7). The correlation between Gubbio and Shatsky Rise rests mainly on the strong similarity of both carbon-isotope curves, and is supported by biostratigraphic events. The correlation between Gubbio and Tercis is justified by calcareous nannofossil dates, and the correlation of Gubbio with sections of the boreal chalk is based mainly on the diagnostically large magnitude of the carbon-isotope events. In general, pelagic carbonate successions of the outer shelf and open ocean are the archives best suited to preserve primary fluctuations of the global carbon cycle.

5.2 Diachroneity and isochroneity of biostratigraphic datum levels

The identification of distinct events in Campanian-Maastrichtian carbon-isotope stratigraphy, relative to the palaeomagnetic polarity scale and anchored to biostratigraphic datum levels, allows for greater precision in the inter-site calibration of micro- and macrofossil zonations. In addition, the duration of carboncycle changes can be considered relative to the Campanian-Maastrichtian time-scale. In the following, the diachronous and isochronous occurrences of biostratigraphic index taxa will be discussed, relative to carbon-isotope stratigraphy. Diachrononeity in base and top occurrences of biostratigraphic index taxa are a well-acknowledged problem in Campanian-Maastrichtian biostratigraphy (e.g. Huber 1990, Pospichal and Wise 1990, Schönfeld and Burnett 1991, Li and Keller 1998, Burnett et al. 1998, Christensen et al. 2000, Lees 2002, Huber et al. 2008). The reasons are sometimes complex but commonly are related to taxonomic and preservational effects (e.g. rare/non-occurrences of index taxa, poor preservation, different species concepts or lack of experience of various authors) as well as true biogeographic provincialism.

Macrofossils

Nostoceras polyplocum: The stratigraphic range of the ammonite *N.polyplocum* is used as a zonal index taxon in European shelf-sea sections. A series of studies relate a pronounced regression to this biozone and discuss it in terms of eustatic sea-level change (Jarvis et al. 2002, Steuber et al. 2005, Niebuhr et al. 2007, Wendler et al. 2011). However, the occurrence range of *N.polyplocum* appears to be diachronous, relative to

the carbon-isotope stratigraphy presented here: single occurrences of the ammonite are below the LCE at Tercis and within the LCE at Lägerdorf. Accordingly, the age discrepancy of the base occurrence of *N.polyplocum*, between Tercis and Lägerdorf, is about 2.5 Myr.

Pachydiscus neubergicus: The base occurrence of the ammonite *P. neubergicus* is just above the Campanian–Maastrichtian boundary at Tercis (Odin and Laumarelle 2001) between CMBE-2 and CMBE-3 (Fig. 6). Single records of *P. neubergicus* at Kronsmoor place its base occurrence in the *pseudoobtusa* to lower *obtusa* belemnite zones (Niebuhr and Esser 2003, Niebuhr et al. 2011), and thus between CMBE-2 and CMBE-3. Accordingly, its base occurrence at Kronsmoor is also very close to the Campanian–Maastrichtian boundary. However, the calibration of base occurrences of *P. neubergicus* relative to the boreal belemnite zonation suggests a diachronous appearance of this species in Europe (Christensen et al. 2000).

Belemnella sumensis: The base occurrence of the belemnite *B. sumensis* lies consistently at the base of CMBE-4 at Norfolk (Figs. 5, 6) and Kronsmoor (Fig. 6) and thus is a good biostratigraphic marker for boreal chalk zonation.

Planktonic foraminifera

Radotruncana calcarata: The R.calcarata Zone is defined by the total range of the species, which is considered to occur globally synchronously within a brief time span of 0.5 Myr (Premoli Silva and Sliter 1995, Huber et al. 2008). The calibration of the base and top occurrences of R. calcarata to carbon-isotope stratigraphy does not necessarily support this assumption. At Tercis, the calcarata Zone appears to lie below the LCE, although a longer range of R. calcarata at Tercis has been reported by Ion and Odin (2001), based on two poorly preserved specimens, and so the LCE may be within this zone there (Fig. 3). At Gubbio, the LCE is within the biozone. The preservation of microfossils at Tercis is rather poor, such that the total range of this zone is possibly not recorded. At Shatsky Rise, the LCE is not well resolved; however, a small-scale δ^{13} C minimum occurs at the appropriate stratigraphic position, within the range of *R*. calcarata and above the base occurrence of Uniplanarius trifidus. A recently published δ^{13} C record from southern Tibet places the *R*. calcarata Zone, similar to its level at Tercis, entirely below the LCE (Wendler et al. 2011). At present, it is not clear whether the range of *R*.calcarata is diachronous, or whether the total range is not fully recovered due to

deleterious preservational effects, and/or possible problems with identifying the species in poorly preserved sediments. However, the calibration of the *calcarata* zone to carbon-isotope stratigraphy suggests a longer duration of the species range of more than 1 Myr.

Globotruncana aegyptiaca: The base occurrence of G.aegyptiaca lies a short distance below the C32r2r/C33n boundary, below the base of Globotruncana gansseri at Gubbio, and in a δ^{13} C minimum below the CMBE at 1210B, coincident with the base of G.gansseri. The temporal discrepancy between the two sites covers the duration of Chron C32r based on carbon-isotope correlation, and is possibly related to the inconsistent identification of the true base age of G.aegyptiaca.

Globotruncana gansseri: The base occurrence of *G.gansseri* is frequently used as a marker for the base of the Maastrichtian Stage (e.g. Li and Keller 1998, Barrera and Savin 1999). Relative to carbon-isotope stratigraphy, its base lies at a similar levels below the CMBE at Gubbio and Hole 1210B. The record at Gubbio shows this level to be in the upper Chron C32r1r. The good agreement between the two sites argues for a synchronous appearance of *G.gansseri* at low latitudes. At higher latitudes, the base of *G.gansseri* appears to be delayed, lying above CMBE-4 at Hole 525A.

Contusotruncana contusa: The base occurrence of *C.contusa* is reported to lie shortly below the base of *Racemiguembelina fructicosa* (Huber et al. 2008), or to coincide with the base of *R.fructicosa* (e.g. Bralower et al. 2002). The results of this study indicate that the base of *C.contusa* appears to be diachronous, occurring at a level between CMBE-2 and CMBE-3 at Tercis (at level 116), at the level of the δ^{13} C rise above the CMBE at Gubbio and Hole 525A, and just below the base of MME-1 at Hole 1210B. This range covers a time-span of 3 Myr, does not show a latitudinal pattern, and might be related to species misidentification (Gardin et al. 2012). Accordingly, the base of *C.contusa* is currently not a reliable marker for biostratigraphic zonation.

Racemiguembelina fructicosa: The base of *R.fructicosa* correlates with the upper third of Chron C31r in southern Europe (Premoli Silva and Sliter 1995, Robaszynski and Caron 1995) and in the subtropical North Atlantic Ocean (Huber et al. 2008). Relative to carbon-isotope stratigraphy, the base of *R.fructicosa* corresponds to the δ^{13} C rise above the CMBE at Gubbio. However, the base of *R.fructicosa* lies within the middle of Chron C31n in the mid-latitude South Atlantic Ocean (Li and Keller, 1998), which corresponds to the MME (Fig. 7): this pattern has been linked to poleward diachroneity in the distribution of this taxon (Huber et al. 2008). A similar position is indicated by shipboard biostratigraphic data from Hole 1210B, where the base of *R.fructicosa* is associated with the MME (MME-1) and the lowest consistent occurrence of the nannofossil *Cribrocorona gallica* (Lees and Bown 2005). Whether this higher occurrence age in the tropical Pacific is related to true diachroneity or to the lower resolution of shipboard data remains to be resolved.

Abathomphalus mayaroensis: At Gubbio (Bottaccione Gorge and Contessa Highway), the base of A.mavaroensis lies just below the base of Chron C31n (Gardin et al. 2012), and thus a short distance below the base of the MME. At Hole 1210B, it lies somewhat higher, above MME-3, which might reflect the lower resolution of the shipboard data (Bralower et al. 2002). At high latitudes, the base of A. mayaroensis correlates with the middle of Chron C31r, as summarized from several circum-Antarctic palaeomagnetic reference sections (Huber 2002), a horizon equivalent to the level of the post-CMBE δ^{13} C rise at Hole 690C (Fig.7). This diachroneity has been attributed to migration of this species from polar to lower latitudes. In the South Atlantic Ocean, the lowest base occurrence of A. mavaroensis has been recorded. The lowest base occurrence of A. mayaroensis has been recorded In the South Atlantic Ocean: at Hole 525A, the datum level lies in the lower Chron C31r (Li and Keller 1998), a level that corresponds to CMBE-4 (Fig. 7). At present state of knowledge, no arguments can be offered for the diachroneity among different high-latitude sites.

Calcareous nannofossils

Uniplanarius trifidus: The base occurrence of U.trifidus lies consistently below the LCE at Gubbio and Tercis and so appears to be a reliable biostratigraphic event at low latitudes. Its base at Hole 1210B is lower than at Gubbio and Tercis, lying below the base of *R.calcarata* (outside the lower range of Fig.7). However, the δ^{13} C curve at Shatsky Rise is not well constrained because the LCE is not such a distinct carbonisotope event. Therefore, it is not clear whether the lower base of U.trifidus at Shatsky Rise is a true temporal offset or a result of erroneous carbon-isotopic correlation. Eiffellithus eximius: The top of E. eximius appears to be diachronous, straddling an interval of about 3 Myr. It lies below the base occurrences of U.trifidus and R. calcarata at Bottaccione, and is not recorded from the Contessa Highway section (Gardin et al. 2012). At Hole 1210B, it lies between the base and top of R. calcarata (Bralower et al. 2002, Lees and Bown 2005), below the suspected LCE. At Tercis, the top occurrence of *E. eximius* appears to be somewhere between levels 99 and 124.5 (Fig. 3), thus sitting below or in the lower part of the CMBE. A base occurrence below the CMBE is also recorded from Kronsmoor. The different positions of this datum level may relate to a preservational or taxonomic problem. The top age of *E. eximius* appears not to be a reliable biostratigraphic marker at all latitudes.

Uniplanarius trifidus: The top occurrence of U.trifidus is used as a nannofossil marker in the low-latitudes, where it occurs at the top of Subchron C32n.1n in the lowermost Maastrichtian (Husson et al. 2001, Gardin et al. 2012). Relative to carbon-isotope stratigraphy, it lies at the top of CMBE-3 at 1210B and Gubbio, and between CMBE-3 and CMBE-4 at Tercis. This level is relatively synchronous within < 400 kyr, making this low-latitude datum level reasonably useful for the identification of the Lower Maastrichtian.

Broinsonia parca subsp.: The top occurrence of *B.parca* subsp. lies in a relative narrow interval in the upper half of the CMBE. It is recorded at the top of CMBE-3 at 1210B, from CMBE-4 to CMBE-5 at Tercis, in CMBE-5 at Gubbio, between CMBE-4 and CMBE-5 at Kronsmoor, and below the top of CMBE-5 at Stevns-1 (Figs. 6, 7). Thus, this appears to be a relatively good datum level for biostratigraphic correlation.

Tranolithus orionatus: The top of *T. orionatus* lies above CMBE at Stevns-1 and 1210B, and in CMBE-5 at Kronsmoor. At Gubbio, this datum level is on top of CMBE-5 at Contessa and above the CMBE at Bottaccione (Fig. 4; Gardin et al. 2012). This discrepancy is possibly related to preservational effects in limestones, and the upper top occurrence at Bottaccione is considered as the more reliable. The top of *T. orionatus* lies above what is exposed in Tercis. This datum level is a relatively reliable marker for biostratigraphic correlation.

Reinhardtites levis: The top of *R. levis* seems also to be synchronous, relative to carbon- isotope stratigraphy. This datum level lies within the initial δ^{13} C increase above the CMBE at Gubbio and Stevns-1. In northern

Germany, it lies in an interval with no outcrop exposure between the Kronsmoor and Hemmoor sections that is equivalent to the initial post-CMBE δ^{13} C increase.

Lithraphidites quadratus: The base of *L. quadratus* is close to the base of MME-1 in Gubbio, in MME-2 at Stevns-1, at the base of MME-3 at 1210B, and above the MME in Hemmoor (northern Germany). The temporal range of FO ages is in the order of 700 kyr.

Micula murus: The base of M. murus is considered as a synchronous datum level in the lower latitudes (Burnett et al. 1998), and is recorded from Chron 30n at the Bottaccione and Contessa Highway sections in Gubbio (Gardin et al. 2012). There, it is associated with the inflection towards more steeply declining δ^{13} C values above the MME (Fig. 4). At 1210B, however, the base of *M.murus* is significantly older and lies between MME-2 and MME-3 just above the base of L. quadratus. This interval falls into Chron C31n according to the carbon-isotope correlation of this study. The temporal discrepancy is of the order of 1 million years and might reflect a true diachroneity. Although not well constrained by palaeomagnetic data, an older base age of *M.murus* has also been suggested from low-latitude sites in the tropical Atlantic (Thibault et al. 2010).

Micula prinsii: The base of *M. prinsii* lies in a narrow range close to the chron C29/C30 transition. It lies in KPgE-1 at 1210B, and in KPgE-2 at Gubbio and at Hole 525A. The temporal offset is less than 300 kyr according to the age model of this study used for carbon-isotope correlation. This datum level is a relatively reliable marker for stratigraphic correlation.

5.3 Age of the Campanian–Maastrichtian boundary

Based on our global correlation of carbon-isotope events, the Campanian–Maastrichtian boundary as defined in the low-latitude, shelfal, global stratotype section in Tercis, can be independently traced into open-oceanic and high northern-latitude shelf-sea settings. Furthermore, it can be more precisely positioned, relative to the geomagnetic polarity time-scale. All Late Cretaceous radiometric ages are based on the 40 Ar/ 39 Ar ages from bentonites in the Western Interior Basin, USA (Obradovich 1993). Here, these radiometric ages are recalculated to adjust them to the astronomical calibration of the Fish Canyon sanidine (28.201 ± 0.046 Ma; Table 3; Kuiper et al. 2008). More recent ages for the Fish Canyon sanidine differ

Ammonite Zone	Age (Ma) Obradovich (1993), Cobban et al. (2006)	Age (Ma) Kuiper et al. (2008)
Baculites clinolobatus	69.59	70.03
Baculites grandes	70.00	70.45
Baculites eliasi	71.98	72.44
Baculites reesidei	72.94	73.41
Baculites compressus	73.52	73.99
Didymoceras cheyennense	74.67	75.15
Exiteloceras jenneyi	75.08	75.56
Didymoceras nebrascense	75.19	75.67
Baculites scotti	75.08	75.56
Baculites scotti	75.84	76.32

Table 3Radiometric ${}^{40}\text{Ar}/{}^{39}\text{Ar}$ ages of bentonites in the
Western Interior Basin.

from this value (28.305 Ma: Renne et al. 2010, 27.93 Ma: Channell et al. 2010). However, the age of the Fish Canyon sanidine as provided by Kuiper et al. (2008) has been independently confirmed by Rivera et al. (2011) by cross-calibration of 40 Ar/³⁹Ar ages of tephra sanidines with the directly astronomically tuned Faneromeni section in Crete. Consequently, the age of Kuiper et al. (2008) is used in this study.

The age of the Campanian-Maastrichtian boundary can be estimated by three independent methodologies. The first approach is based on the results of carbon-isotope correlation between Tercis (the GSSP) and Gubbio (the standard section for the geomagnetic polarity time-scale; Channell et al. 1978, Lowrie and Alvarez 1981, Gardin et al. 2012). At Tercis, the Campanian-Maastrichtian boundary lies on top of CMBE-2 at the top of the CMBE δ^{13} C decrease. This level can be precisely be correlated to Gubbio, where it falls in the upper part of Chron C32n2n (at around 322 m in the Contessa Highway section). The ages of geomagnetic reversals have been adjusted for the Fish Canyon sanidine calibration, in the same way as the radiometric bentonite ages (Table 2). Accordingly, the Campanian-Maastrichtian boundary has an age of 72.14 Ma.

The second approach is based on the biostratigraphic correlation of inoceramid zones between Tercis and the Western Interior Basin (Fig. 8). The Campanian–Maastrichtian boundary lies in the upper third of the "*Inoceramus*" *redbirdensis* Zone, between the base occurrences of "*I*." *redbirdensis* and *Endocostea typica*, in Tercis (Walaszczyk et al. 2002a). These datum levels are also recorded in the Western Interior Basin, with the base of "*I*." *redbirdensis* occurring at the same level as

the base of the ammonite *Baculites eliasi* (Walaszczyk et al. 2002b). The newly-calibrated radiometric ⁴⁰Ar/³⁹Ar age within the *B.eliasi* Zone is 72.44 ± 0.31 Ma (Cobban et al. 2006; Table 3). Based on the inoceramid correlation, the Campanian–Maastrichtian boundary is supposed to be high (upper third) in the *Baculites eliasi* and "*I.*" *redbirdensis* zones (Walaszczyk et al., 2002b). This placement of the Campanian–Maastrichtian boundary in the Western Interior Basin results in an approximate age of 72.2 ± 0.2 Ma.

The third approach is similar to the first one, but uses Option 2 of the recently published astronomical time-scale for the age of magnetochron boundaries (Table 2; Husson et al. 2011). The astronomical timescale is based on the identification of the long eccentricity signal in high-resolution measurements of magnetic susceptibility and grey-level variations in different deep-sea sections. Option 2 places the K/Pg boundary near 66 Ma, following the recalculated age using the new standard of Kuiper et al. (2008) and the recent cyclostratigraphic studies of the Paleogene (Westerhold et al. 2009, Hilgen et al. 2010). Husson et al. (2011) placed the Campanian-Maastrichtian boundary in a long eccentricity minimum at 72.75 Ma that is 0.75 Ma before the top of U.trifidus at ODP Hole 762C. Using the astronomically calibrated ages of magnetochron boundaries and the carbon-isotopic correlation of the Campanian and Maastrichtian stages in this study, the boundary age is 71.98 Ma, and thus nearly corresponds to the top occurrence age of U. trifidus at Site 762C. This age is nearly identical with the age produced in approach 1, and shows the good agreement between the new ⁴⁰Ar/³⁹Ar calibration and cyclostratigraphic interpretation.

All three ages fall into a narrow range. To date, the largest uncertainty in the definition of the Campanian-Maastrichtian boundary age was related to uncertainties in global biostratigraphic correlation. This problem is still evident in the Geological Timescale 2004 (Gradstein et al. 2004), where a large error is given for the boundary age $(70.6 \pm 0.6 \text{ Ma})$ or from the new estimate utilizing the astronomical time-scale and the position of the LO of U.trifidus in Site 762C (72.75 Ma; Husson et al. 2011). Finally, Husson et al. (2011) placed the Campanian-Maastrichtian boundary in the middle of Chron C32n2n instead of the upper part, as indicated in this this study. The high-resolution, global correlation of carbon-isotope events significantly reduces the uncertainties related to erroneous correlation of some biostratigraphic datum levels. The small discrepancies between the three ages are





either related to the error of radiometric age determination, or to a small uncertainty in the correlation of the base age of "*Inoceramus*" redbirdensis. Averaging all three ages, the age of the Campanian–Maastrichtian boundary can be fixed at 72.1 ± 0.1 Ma.

5.4 Late Campanian–Maastrichtian carbon-isotope variability

The most prominent features of the upper Campanian– Maastrichtian carbon-isotope record are the negative excursions of the LCE and CMBE, the rapid increase after the CMBE and the short-term variations of the CMBE, MME and KPgE. The negative δ^{13} C excursions of the LCE and CMBE are often discussed as being a result of eustatic sea-level fall that would promote lowland erosion and organic-matter oxidation (e.g. Jarvis et al. 2002, 2006). Eustatic sea-level fall has been explained in terms of thermal contraction of the water column due to global cooling and glacio-eustasy, with the negative CMBE δ^{13} C excursion linked with a sequence boundary recorded in shallow-water settings along the New Jersey margin (Barrera et al. 1997, Miller et al. 1999, 2005, Browning et al. 2008).

This study shows that the CMBE covers a time span of ~2.5 Myr with the negative δ^{13} C excursion lasting \sim 1 Myr. This time-span is too long to explain the carbon-cycle change solely by a glacio-eustatic mechanism that usually operates on time-scales of < 100 kyr. Even if shorter periods within the CMBE were accompanied by glacial periods, the presently available database does not provide unequivocal evidence for glaciations at time-scales of < 100 kyrs. Further negative evidence comes from the sequence stratigraphic interpretation of the Tercis section. The sedimentary succession shows a distinct facies shift from shallowmarine to outer-shelf environments across the CMBE, which is indicative of a retrogradational stacking pattern, equivalent to a transgression (Figs. 2, 3). Even if the transgression were controlled by local processes, such as increased rates of basin subsidence, a glacioeustatic sea-level fall with a magnitude of > 10 m, as



Fig. 9. Comparison of the age-scaled δ^{13} C records of Gubbio and Hole 1210B with the astronomical solution of eccentricity, La2010d (green curve; Laskar et al. 2011). The blue and red curves are 405 kyr and ~2.4 Myr orbital eccentricity cycles, respectively extracted by low-pass filtering. The age model is based on the ages of geomagnetic reversals recorded at Gubbio, which have been recalibrated to the astronomically calibrated age of the Fish Canyon sanidine (see Table 2; Kuiper et al. 2008).

is supposed for the CMBE (Miller et al. 1999), should have equally affected the facies architecture at Tercis, and this is not the case. The long duration of the LCE, CMBE and MME, in combination with a rather slow initiation, is more indicative for large-scale earth processes such as changes in plate-tectonic configuration (Müller et al. 2008a), mid-ocean ridge subduction (Müller et al. 2008b) or hot-spot volcanism (Frey et al. 2000, Coffin et al. 2002), as triggering mechanisms. Tectonic processes could have caused changes in oceanic bathymetry and shelf-sea size and/or in the exchange of deep waters between oceanic basins, and thus could have influenced the magnitude of the ¹²C transfer into reduced (i.e. organic-carbon-rich) sedimentary reservoirs.

The short-term δ^{13} C events (CMBE-1 to CMBE-5, MME-1 to MME-3 and KPgE-1 to KPg-3) occurred superimposed on long-term shifts in the carbon cycle and were likely related to orbitally forced climate changes. The high-frequency δ^{13} C events are especially well developed in tropical surface waters (Gubbio, Hole 1210B). Their duration falls in the frequency band of long eccentricity, as is evident from the $\delta^{13}C$ plot relative to the astronomical solution "La2010d" (Fig. 9; Laskar et al. 2011). Although the δ^{13} C records of Hole 1210B and Gubbio are not astronomically tuned, this first approximation shows a high sensitivity of the δ^{13} C signal during periods of high eccentricity amplitudes. A high-frequency δ^{13} C variability, paced by long eccentricity, is well known from Cenozoic successions (e.g. Cramer et al. 2003, Pälike et al. 2006). Proposed mechanisms imply climatically induced oscillations in biosphere productivity with the long residence time of carbon in the ocean amplifying longer forcing periods (Pälike et al. 2006); the size of terrestrial wetlands related to seasonal land-sea heating, and distribution of precipitation (Zachos et al. 2010); and variations in the size of the methane hydrate reservoir, due to changes in ocean temperature and productivity (Dickens 2003, 2011). Similar mechanisms could have exerted control on Campanian-Maastrichtian carbon-cycle changes.

Furthermore, the depositional environment seems to have had an effect on the magnitude of the Campanian–Maastrichtian δ^{13} C excursions, especially to those of the LCE and CMBE. The largest magnitude excursion, with values up to 1‰, is recorded from shelf-sea sections in the North Sea and the deep waters of the Southern Ocean. Small excursions are typical for oceanic surface waters, such as those of the Tethys or the tropical Pacific. This pattern is also evident at Tercis, where the succession shows the overall deepening trend. Accordingly, upper Campanian shallowwater deposits are related to a larger magnitude of the LCE, while the outer shelf deposits of the uppermost Campanian–Maastrichtian are associated with a small CMBE, whose magnitude is similar to those of open oceanic sites.

The reasons for the higher sensitivity of deep-water and shallow-marine carbon reservoirs are complex and might be related in some degree to diagenetic effects, but also to a more effective carbon partitioning of dissolved inorganic carbon in smaller water-bodies. Different patterns of carbon partitioning between shelves and oceanic basins are poorly understood today. However, the global exchange of water-masses may have been limited in Late Cretaceous time with shelf-sea reservoirs becoming smaller and affected by regional organic-carbon sources and sinks. In particular sealevel change would have exerted control on the shelf size, the amount of supplied isotopically light carbon and the degree of water-mass exchange with the open ocean by the development of shelf-frontal systems. Likewise, the deep waters of the Southern Ocean could have formed a restricted reservoir with only limited exchange with other deep-water bodies, due to the plate-tectonic configuration and the small size of the South Atlantic Ocean

6. Summary

New, high-resolution carbon-isotope records for the GSSP of the Campanian-Maastrichtian boundary at Tercis les Bains (France), the Bottaccione and Contessa sections at Gubbio (Italy), and the coastal sections at Norfolk (UK) have been generated to provide a global δ^{13} C correlation between boreal and Tethyan shelf-sea as well as low- and higher latitude oceanic sites. The new $\delta^{13}C$ records are correlated with $\delta^{13}C$ stratigraphy of the boreal chalk sea (Trunch borehole. Norfolk, UK; Lägerdorf-Kronsmoor-Hemmoor section, northern Germany), the tropical Pacific (ODP-Hole 1210B, Shatsky Rise) and the Southern Ocean (DSDP Hole 525A, ODP Hole 690C) by using a composite Gubbio δ^{13} C record as reference. The global correlation allows the identification of significant, high-frequency δ^{13} C variations occurring superimon Campanian-Maastrichtian posed long-term changes such as the Late Campanian Event (LCE), the Campanian-Maastrichtian Boundary Event (CMBE), the mid-Maastrichtian Event (MME) and the Cretaceous-Paleogene Event (KPgE). The carbon-isotope events are related to the geomagnetic polarity scale, recalculated for the astronomical ⁴⁰Ar/³⁹Ar calibration of the Fish Canyon sanidine. This recalculation allows the evaluation of the relative timing of base occurrence ages of stratigraphic fossil datum levels, such as ammonites, planktonic foraminifera and calcareous nannofossils. Furthermore, the Campanian-Maastrichtian boundary, as defined in the stratotype at Tercis, can be precisely positioned relative to carbon-isotope stratigraphy and the geomagnetic polarity timescale. The average value for the age of the Campanian-Maastrichtian boundary is 72.1 ± 0.1 Ma, estimated by three independent approaches that consider the Fish Canvon sanidine calibration and Option 2 of the Maastrichtian astronomical time-scale. The CMBE lasted 2.5 Myr and most probably reflects carbon-cycle changes related to tectonic processes. The duration of high-frequency δ^{13} C variations instead coincides with the frequency band of long eccentricity, and suggests orbital forcing of climate-induced carbon-cycle changes.

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