Clay mineralogical and geochemical expressions of the “Late Campanian Event” in the Aquitaine and Paris basins (France): Palaeoenvironmental implications

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A B S T R A C T

Campanian sediments from two French sedimentary basins were studied, using clay mineralogy and stable isotope (δ13C and δ18O) geochemistry, in order to investigate the Late Campanian Event. The clay fraction of the Campanian sediments from the Tercis-les-Bains section (Aquitaine Basin) and from the Poigny borehole (Paris Basin) is mainly composed of smectite. This background sedimentation was, however, interrupted during the Upper Campanian in the two basins by a substantial increase in detrital inputs, including illite, kaolinite, and chlorite at Tercis-les-Bains, and illite at Poigny. This detrital event, resulting from the enhanced erosion of nearby continental areas triggered by increasing runoff, has also been recognized in the Tethys and South Atlantic oceans. It coincided with a global negative carbon isotope excursion, the Late Campanian Event (LCE). Carbon isotope stratigraphy was used to correlate the two basins with previously studied sections from distant areas. Spectral analysis of the bulk δ13C from Tercis-les-Bains suggests a duration of ca.

1. Introduction

The Cretaceous is a “greenhouse” period with maximum sea-surface temperatures recorded around the Cenomanian to Turonian interval (Jenkyns et al., 1994; Clarke and Jenkyns, 1999; Pucéat et al., 2005; Friedrich et al., 2012). Following this climatic optimum, isotopic data highlight a long-term cooling during the remainder of the Late Cretaceous (Huber et al., 1995; Clarke and Jenkyns, 1999; Friedrich et al., 2012; Linnert et al., 2014). This cooling trend accelerated during the beginning of the Campanian (Friedrich et al., 2012, Linnert et al., 2014), but its mechanisms and dynamics are not yet well understood. The Campanian is also characterized by significant fluctuations of the sea level (Haq et al., 1987; Barrera et al., 1997; Jarvis et al., 2002), a major shift in the δ15N of marine organic matter (Algeo et al., 2014), clay mineralogical changes, and the occurrence of positive and negative carbon isotope events: the Santonian–Campanian Boundary Event (SCBE) (Jarvis et al., 2002, 2006), the Mid Campanian Event (MCE) (Jarvis et al., 2002, 2006; Thibault et al., 2012a), the conico Event (Perdiou et al., 2015), the Late Campanian Event (LCE) (Jarvis et al., 2002, 2006; Voigt et al., 2012; Thibault et al., 2012a, b), the Epsilon Event (EE) (also called C1- Event) (Thibault et al., 2012a, 2015), and the Campanian–Maastrichtian Boundary Event (CMBE) (Barrera, 1994; Barrera and Savin, 1999; Friedrich et al., 2009; Jung et al., 2012; Voigt et al., 2012; Thibault et al., 2012a, 2015). Mineralogical changes expressed by detrital inputs of kaolinite and illite have been observed in many sedimentary basins, including the South Atlantic Ocean (Chamley et al., 1984), the Umbria-Marche Basin (Deconinck, 1992), the Saharan Platform (Li et al., 2000), and in the Paris Basin (Deconinck et al., 2005). As these mineralogical changes are stratigraphically poorly constrained, they cannot be associated with isotopic events. The objective here is to better understand the Campanian palaeoclimatic cooling by an integrated study of clay mineralogy and
isotope geochemistry ($\delta^{13}C$ and $\delta^{18}O$). In addition, a cyclostratigraphic study was conducted in order to estimate the duration of the clay mineral change during the LCE.

We focus on two French sedimentary basins, the Aquitaine and the Paris basins. Isotopic data from the Tercis-les-Bains section (Aquitaine Basin) published by Voigt et al. (2012) are compared with new clay mineralogical data, while clay mineralogical data from the Poigny borehole (Paris Basin) published by Deconinck et al. (2005) are compared with new isotopic data. The whole data set is used to better constrain the timing of clay mineralogical changes and isotopic events that occurred in both basins.

2. Palaeogeography and geological settings

During the Campanian, the Atlantic Ocean was widening, while the Tethys Ocean was in the process of closing due to the counterclockwise motion of Africa (Smith, 1971; Dewey et al., 1973; Blakey, 2008). This period corresponded to the development of epicontinental seas in the Tethyan realm. Western Europe was an archipelago, whose islands corresponded to emergent Hercynian massifs (e.g., Armorican, Central, and Rhenian Massifs) separated by epicontinental seas (Fig. 1). These emergent lands locally contributed to terrigenous sedimentation, although most Campanian sediments in the studied basins are composed of chalk and bioclastic limestone beds.

2.1. The Tercis-les-Bains section

The studied section is located in an abandoned quarry near Tercis-les-Bains (north-west of Dax) and belongs to the Aquitaine Basin (south-west France, Fig. 1). This basin was in an intermediate position between the North Atlantic and the Tethyan oceans (Fig. 1). The Tercis-les-Bains quarry, opened on the side of a diapir, shows vertically oriented Late Campanian to Maastrichtian beds (Bilotte et al., 2001; Odin, 2001). The 116-m-thick Campanian succession is composed of bioclastic limestone beds with common glauconitic horizons, flint nodules, and occasional marly levels (Fig. 2). The relatively homogeneous facies, microfacies, and faunal associations reflect deposition on the outer shelf in lower offshore environments (Berthou et al., 2001). The section is defined as the Global boundary Stratotype Section Point (GSSP) of the base Maastrichtian Stage (Odin, 2001), ensuring a well-defined magnetostratigraphic and biostratigraphic framework for the middle and upper part of the Campanian (Fig. 3).

2.2. The Poigny borehole

A thick succession of chalk (about 700 m), deposited from the Cenomanian to the Campanian, was drilled at Poigny, south-east of Paris (Craie 700 project, Mégien and Hanot, 2000; Fig. 1). The Paris Basin was surrounded by the London-Brabant Massif to the north, by the Massif Central to the south and by the Armorican Massif to the west (Fig. 1). During the Late Cretaceous, the Paris Basin was an epicontinental sea where chalk accumulated. It was connected with the Tethys to the south-east, with the Boreal Ocean to the north and with the North Atlantic to the west. The lithological description of the borehole includes marker beds and biostratigraphic data based on benthic foraminifera, dinoflagellates, ostracods, nannofossils, and bivalves (Fig. 4), which allowed a detailed stratigraphic framework of the ~250-m-thick Campanian succession to be established (Robaszynski et al., 2005). Unfortunately, planktonic foraminifera cannot be studied in the Campanian succession of the Poigny borehole due to their poor preservation.

Fig. 1. Studied sites located on (A) a geographic map and (B) on a palaeogeographic map of the Western Peri–Tethyan Realm during the Early Campanian (modified after Philip and Floquet, 2000).
but some markers are present in the Cenomanian–Turonian interval. Through bio- and lithostratigraphic arguments, the Santonian–Campanian boundary has been identified between 285 and 290 m (Robaszynski et al., 2005). A major erosion event during the early Tertiary was responsible for the absence of uppermost Campanian and Maastrichtian chalk deposits (Mettraux et al., 1999; Guillocheau et al., 2000; Lasseur, 2007).

3. Materials and methods

3.1. Clay mineralogy

In the Tercis-les-Bains quarry, bulk-rock samples were collected at a sample interval of 50 cm (Fig. 2). Mineralogical analyses were performed at the Biogéosciences Laboratory, University of Bourgogne Franche-Comté, Dijon, France. Clay mineral assemblages of 212 samples devoid of matrix were identified by X-ray diffraction (XRD) on oriented mounts of non-calcareous clay-sized particles (<2 μm). The procedure described by Moore and Reynolds (1997) was used to prepare the samples. Diffractograms were obtained using a Bruker D4 Endeavor diffractometer with CuKα radiations with LynxEye detector and Ni filter, under 40-kV voltage and 25-mA intensity. Three preparations were analyzed, the first after air-drying, the second after ethylene-glycol solvation, and the third after heating at 490 °C for 2 h. The goniometer was scanned from 2.5° to 28.5° for each run. Clay minerals were identified by the position of their main diffraction peaks on the three XRD runs, while semi-quantitative estimates were produced in relation to their area (Moore and Reynolds, 1997). Areas were determined on diffractograms of glycolated runs with MacDiff 4.2.5. Software (Petschick, 2000). Beyond the evaluation of the absolute proportions of the clay minerals, the aim was to identify their relative fluctuations along the section. Peaks close to 14 Å in air-dried conditions and 17 Å after ethylene-glycol solvation are random R0 type illite/smectite mixed-layers (60–80% of smectite sheets on average according to Inoue et al., 1989 and Moore and Reynolds, 1997). In the result and discussion sections, the term smectite, as classically employed by sedimentologists, is used to refer to these minerals (Chamley et al., 1990; Deconinck et al., 2005, Pellenard and Deconinck, 2006). The smectite/illite ratio (S/I) corresponds to the ratio between the 17-Å peak area and the 10-Å peak area (defined as illite), after ethylene-glycol solvation.

3.2. Stable isotope geochemistry

Wherever possible, samples were recovered every meter from the Poigny borehole for geochemical analyses. Isotopic analyses (δ13C and
δ¹⁸O) were performed on 243 samples of bulk sediment along the whole section from the Santonian–Campanian boundary to the Late Campanian. Unfortunately, due to the Tertiary erosion, the uppermost part of the Campanian is missing and the yellowish chalk succession observed in the topmost part of the core probably indicates the circulation of meteoric fluids (Fig. 4). Isotope analyses were performed at the

![Image](https://example.com/image.png)

**Fig. 3.** Clay mineralogy of the Campanian at Tercis-les-Bains compared with the carbon isotope stratigraphy from Voigt et al. (2012), with the two different isotopic interpretations of Voigt et al. (2012) and of Thibault et al. (2012a, 2012b). References: (1) Odin and Lamaurelle (2001), Lewy and Odin (2001); (2) Odin et al. (2001b); (3) Walaszczyn and Odin (2002); (4) Odin et al. (2001a); (5) Melinte and Odin (2001); (6) Gardin et al. (2001); (7) Voigt et al. (2012); Odin (2001). Abbreviations: E.e. = Eiffellithus eximius; G.ventricosa/G. rugosa = Globotruncana ventricosa/Globotruncana rugosa; P. = Pseudoxybeloceras sp.; R.m. = Racinolithus magnus; U.g. = Uniplanarius gothicus; U.t. = Uniplanarius trifidus; Gl = glauconite.

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**Fig. 4.** Isotope data of the Poigny borehole compared to the mineralogical data from Deconinck et al. (2005). The dashed line corresponds to the mineralogical change identified by Deconinck et al. (2005). The figure caption used to describe the lithology is the same as Fig. 2. Abbreviations: A. = Areoligera; B.a. = Bolivinoides australis; B.d. = Bolivinoides decoratus; B.in. = Bolivina incrassata; B.p.c. = Broinsonia parca constricta; B.st. = Bolivinoides strigillatus; E.e. = Eiffellithus eximius; G.h. = Globotruncanella havanensis; G.m. = Gavelinella monterelensis; G.s. = Gavelinella stelligera; S.p. = Semonia sphera protrusa; LCE = late Campanian event; MCE = mid Campanian event; SCBE = Santonian–Campanian Boundary Event.
Spectral analyses were conducted on isotopic data (δ13C) to detect orbital cycles and to estimate the duration of the LCE in the Tercis-les-Bains section. The δ13C series shows a marked negative shift in δ13C values (1%) previously identified as the LCE (Voigt et al., 2012). This shift and the long-term trends were removed by applying a best-fit piecewise linear regression; the series was then standardized (mean = 0; standard deviation = 1). The spectrum of the AR(1) pre-whitened series was calculated using the multi-taper method, applying three 20 tapers (Thomson, 1982, 1990). The low-spec method was then used to calculate the spectrum background of the pre-whitened series and the confidence levels (Meyers, 2012, 2014). A time-frequency weighted fast Fourier transform (T-F WFFT) applying 30-m-width windows was performed to follow the evolution of the main periods throughout the δ13C series (Martínez et al., 2013, 2015). The method consists in dividing the series into a series of intervals of 30-m width separated from each other by 0.5 m. In each interval, the local trend in the average is removed by subtracting a linear regression from each interval of the series. Each of the intervals is then weighted using one Slepian sequence and a Fast Fourier Transform is calculated on each of the weighted signals. The result is a 3-dimensional spectrum, called spectrogram, showing in blue the spectrum background and in red the highest powers. A Taner band-pass filter was then applied to isolate the cycles of interest (Taner, 2003).

4. Results

4.1. Tercis-les-Bains

At Tercis-les-Bains, the clay mineral assemblages are predominantly (more than 80%) composed of random illite/smectite mixed-layers (IS R0), hereafter referred to as smectite (Fig. 3). Other clay minerals, including illite (generally less than 20%) and traces of chlorite (less than 5%), occur in most samples (Fig. 3). Kaolinite is absent except within the interval from 33 to 62.5 m. In this interval, the kaolinite content significantly increases up to 8% together with more abundant illite and chlorite. This major change in the clay mineral assemblages matches preliminary data (Odin, 2001) and is the most striking feature of the section. The kaolinite-bearing interval starts concomitantly to the occurrence (FO) of the calcareous nannofossil

δ18O values that coincides with the lowest recorded δ13C values.

5. Discussion

5.1. Influence of diagenesis

Before interpreting the clay mineral successions in terms of palaeoenvironments, it is necessary to evaluate the influence of diagenesis. In both studied sedimentary successions, the occurrence of smectite indicates negligible influence of burial diagenesis, as these minerals are very sensitive to temperature increase, with illitization starting at about 60 °C (Mrdo, 2009). According to the geological history of the Paris Basin (Brunet and Le Pichon, 1982), a 500-m-burial depth was estimated at Poigny, not deep enough to trigger incipient illitization. In this borehole, clay minerals are thus mainly considered as detrital (Deconinck et al., 2005; Fig. 4). Similarly, the occurrence of smectite at Tercis-les-Bains indicates a minor influence of burial diagenesis (Fig. 3). The presence of glauconitic granules, generally less than 1% (Odin and Amorosi, 2001; Fig. 3), implies that illite identified by XRD (10-A peak) consists of a mixture of a major detrital component with minor authigenic glauconitic minerals formed during early diagenesis. The amounts of glauconite reaching 1–3% only occur in three thin well-identified horizons indicative of lower sedimentation rates (Fig. 3).

In the Poigny borehole, the upper part of the chalk is yellowish, in marked contrast to the remainder of the core. This yellowish chalk has a low Mg content and has therefore been interpreted as alteration caused by meteoric fluids during the post-Cretaceous emersion (Le Calloncne et al., 2000). The isotopic results for the upper part of the Poigny core (59–50 m; Fig. 5) are characterized by a negative excursion of about 1% in δ18O values that coincides with the lowest δ13C values. A crossplot of carbon- and oxygen-isotope data from the Poigny borehole (Fig. 6) highlights the difference in isotopic composition between the samples from the yellowish chalk (59 to 50 m; red dots) and those from the remainder of the section (309 to 59 m; blue dots). The data were therefore treated as two separate subsets, and a Spearman’s coefficient was computed for each of these subsets to test the existence of any correlation within the data. This coefficient was 0.3.
chosen because of the non-linear nature of the relationship between the two variables, $\delta^{13}C$ and $\delta^{18}O$. Values of 1 indicate perfect correlation, values of $-1$ indicate perfect anti-correlation, while 0 indicates absence of correlation. From 309 m to 59 m, $\delta^{13}C$ and $\delta^{18}O$ values are negatively correlated, with a Spearman coefficient of $r_s = -0.85$, whereas $\delta^{13}C$ and $\delta^{18}O$ values are positively correlated from 59 to 50 m, with a Spearman coefficient of $r_s = +0.42$. In marine carbonates, the positive correlation of $\delta^{13}C$ and $\delta^{18}O$ values can be explained by several mechanisms, such as kinetic effects recorded within the shells of some organisms (McConnaughey, 1989a, 1989b; McConnaughey et al., 1997; Wenzel et al., 2000; Auclair et al., 2003; Gillikin et al., 2006), or co-variations of seawater temperature and local primary productivity, associated with remineralization of organic matter at depth (e.g., Kirby et al., 1998). However, positive correlation can also be observed as the result of diagenesis, when it is extensive enough to affect carbon isotope composition, which is usually less prone to diagenetic alteration. The influence of meteoric fluids during telogenesis is one form of diagenesis known to produce yellowish alteration of chalk (Le Callonnec et al., 2000). The telogenesis hypothesis can be justified for the first data subset (59–50 m) by two factors: (1) the specific aspect of the chalk and (2) the positive correlation between $\delta^{13}C$ and $\delta^{18}O$ values.
contrast, no positive correlation was observed for the second data subset (309–59 m), which suggests a negligible diagenetic effect on δ13C values for most of the core.

5.2. Identifying C-isotopic events and correlation with other sections in Europe

At Poigny, the Santonian–Campanian boundary is located between 290 m and 285 m, based on benthic foraminiferal bioevents (Robaszynski et al., 2005). This is consistent with the 0.3‰ δ13C positive excursion recorded in this interval, which is therefore attributed to the SCBE (Fig. 4).

A slight increase of δ13C occurring between 164.25 and 138.25 m is tentatively attributed to the MCE (Fig. 4), and the clear trend toward lower values of δ13C, from 112 to 59 m, corresponds to the lower part of the LCE. Several biostratigraphic criteria preclude the attribution of this negative excursion to the CMBE because (1) among benthic foraminifera considered to be good markers of the Campanian–Maastrichtian boundary, Neofibulina reticulata is not present, and (2) the LO of Effelliella eximius occurs at the base of the Late Campanian R. calcarata zone in the Kalaat Senan section (Tunisia; Robaszynski et al., 2000), within the uppermost part of Chron C33n and just below the base of the R. calcarata zone in the Bottaccione section (Italy, Gardin et al., 2012), and in the Late Campanian polyplocum zone in the Lägerdorf–Kronsmoor section (Voigt and Schönfeld, 2010). This nannofossil biostratigraphic marker is also recorded in the early Late Campanian in Norfolk (England, Jarvis et al., 2002), Tercis-les-Bains (Gardin et al., 2001) and in the ODP Hole 762C (Thibault et al., 2012b).

Carbon isotopes are widely used to correlate sections around the globe (e.g., Scholle and Arthur, 1980) and may be a useful tool if diagenetic influences on δ13C are carefully considered (Wendler, 2013). A large number of carbon isotope events have been widely recognized from the Coniacian to the Maastrichtian and defined in the δ13C reference curve of the English chalk (Jarvis et al., 2002). In the GSSP of Tercis-les-Bains, Thibault et al. (2012a) have notably recognized the three-step negative shifts of the δ13C of the CMBE (CMBa, CMBb, CMBc). In the Late Campanian, the EE (or C1 Event; Thibault et al., 2012a, 2015) defined as a slight negative excursion of the carbon-isotope curve is recorded in the Stevns-1 borehole (Danish Basin; Thibault et al., 2012a), in the ODP 762C Hole (Indian Ocean; Thibault et al., 2012b), and in the chalk of Lägerdorf–Kronsmoor section (Boreal Ocean, Thibault et al., 2012b; Figs. 3 and 7).

In the high-resolution carbon–isotope data of Tercis-les-Bains, the LCE appears as a major excursion of —1‰ between 53 and 70 m but is immediately preceded by a sharp —0.4‰ excursion between 45 and 53 m, that we name here pre-LCE following Perdiou et al. (2015; Fig. 3). These authors identified two significant stepwise negative shifts prior to the LCE in the North Sea that they defined as pre-LCE, correlated to Lägerdorf–Kronsmoor and interpreted as an amplification of the pacing of the carbon cycle by the 405-kyr eccentricity preceding the main LCE (Perdiou et al., 2015). We propose here to position the pre-LCE and the LCE excursions based on the most positive value recorded

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**Fig. 7.** Proposed correlation of carbon isotope curves and 405-kyr cycles between the Tercis-les-Bains section and Lägerdorf–Kronsmoor section. This correlation is supported by the tie-points of isotopic curves highlighted by the colored circles. Comparison of the 405-kyr cycles suggests a number of hiatuses at Tercis-les-Bains, which is line with the sequence stratigraphic interpretation; 72.1 Ma indicates the absolute age of the Campanian–Maastrichtian boundary. (1) Voigt et al. (2012), (2) Voigt and Schönfeld (2010), (3) Conica Event as defined by Perdiou et al., 2015, (4) Voigt et al., 2010. Abbreviations: bas./spin. = basiplana/spiniger; con./sen. = conica/senonencis; gr. = grimmensis/granulosis; G.v./G.r. = Globotruncana ventricosa/Globotruncana rugosa; P = Pseudothybeloceras sp.; R.c. = Radotruncana calcarata; G. havanensis = Globotruncana havanensis.
at the base and at the top of each negative excursion. Correlation with Lägerdorf–Kronsmoor suggests that a number of pre-LCE excursions may be also identified there between 145 and 170 m (Fig. 7). Correlation of 405-kyr cycles and carbon isotope variations between the Lägerdorf–Kronsmoor section and the North Sea core Adda-3, where Perdieu et al. (2015) define their pre-LCE excursions, suggests that the pre-LCE interval at Lägerdorf–Kronsmoor includes the two small stepwise 0.2 to 0.3% negative shifts at ca. 155 and 165 m (Fig. 7). Therefore, the sole −0.4% excursion recorded in Tercis-les-Bains that precedes the LCE is correlated here to the whole interval that includes these two negative shifts at Lägerdorf–Kronsmoor (Fig. 7). At Poigny, a rather similar record is observed with the occurrence of pre-LCE stepwise negative excursions between 112 and 59 m, immediately preceding the lower part of the LCE marked by a transient progressive 0.8% negative excursion from 100 to 85 m (Fig. 4). The positive shift that constitutes the upper half of the LCE is hindered at Poigny by the level of intense alteration that shifts the δ13C and the δ18O toward very negative values (highlighted in gray in Fig. 4).

In the lowermost part of the Upper Campanian, a long-lasting excursion of ca. −0.3% immediately follows the MCE at Tercis-les-Bains between 35 and 20 m. This excursion is characterized by steady δ13C values around 2.25‰, a sharp negative shift at the base following the MCE, and a sharp positive shift at the top, coinciding with the FO of R. calcarea. This excursion correlates with a similar trend at Lägerdorf–Kronsmoor observed from 93 to 120 m within UC15c88 (Figs. 3 and 7). This excursion has also been identified in the Adda-3 borehole (North Sea) and recently defined as the conico Event by Perdieu et al. (2015).

5.3. Duration of the LCE

In the Geological Time Scale 2012 (Gradstein et al., 2012), the duration from the FO of Uninplanarius sissinghii to the LO of B. parca constricta is proposed as 5.59 Myr. The uncertainty of the age of the two bioevents is calculated using a Compound Poisson Gamma law, applied to the problem of time-scale uncertainty (Haslett and Parnell, 2008; De Vleeschouwer and Parnell, 2014; Martinez and Dera, 2015, see supplementory information). The age uncertainty (2σ) of the FO of U. sissinghii is assessed as 0.54 Myr, and the age uncertainty (2σ) of the LO of B. p. constricta as 0.78 Myr. The duration from the FO of U. sissinghii to the LO of B. p. constricta with error margins is thus 5.59 ± 1.24 Myr (Table S3). In the Tercis-les-Bains section, the thickness between these two bioevents is 115.08 m, equivalent to an average sedimentation rate of 20.59 m/Myr, ranging from 16.65 to 26.95 m/Myr within the error margins. The 8.6-m wavelength, the highest-amplitude cycle in the δ13C data (Fig. 5C), would thus correspond to an average period of 0.42 ± 0.1 Myr (Table S3), which is close to the period of the 405-kyr eccentricity (Laskar et al., 2004, 2011). The filtered signal on the band of 5–8 m allows the Tercis-les-Bains section to be divided into sequences of 405 kyr (Fig. 5B). The long-eccentricity cycle (405 kyr) identified at Tercis-les-Bains was also recognized using CaCO3 data in the Lägerdorf–Kronsmoor section (northern Germany; Voigt and Schönfeld, 2010), from sediment gray level variations in the ODP Hole 762C (Exmouth Plateau; Thibault et al., 2012b) and in bulk carbonate δ13C in the Bottaccione section (central Italy; Sprovieri et al., 2013).

Based on the 405-kyr sequences identified in the interval that spans the carbon–isotope negative shift (Figs. 5, 7), a total duration from the base of the pre-LCE to the top of the LCE is estimated as 1.3 Myr at Tercis-les-Bains. In the Lägerdorf–Kronsmoor record, the LCE as defined by Voigt et al. (2010) spans approximately two and a half 405-kyr cycles (UCa10, UCa9, and half of UCa8) and thus corresponds to a duration of ca. 1 Myr, while pre-LCE excursions span UCa7 and the upper half of UCa6 (Fig. 7). Here, we have attempted a correlation of 405-kyr eccentricity cycles identified from the δ13C of Tercis-les-Bains to the 405-kyr eccentricity identified from the CaCO3 of Lägerdorf–Kronsmoor (Fig. 7). This attempt is constrained by the correlation of carbon isotope events between the two sections, and in particular the conico Event, the pre-LCE excursions, the LCE and the CMBE (Fig. 7). From this correlation, it appears that several 405-kyr cycles are lacking at Tercis-les-Bains, supporting the inference that this section was affected by some short-term hiatuses, notably at the bottom and at the top of the LCE interval. However, the duration of the LCE appears rather similar at Tercis-les-Bains, spanning 800 to 900 kyr, while the pre-LCE spans another 400 kyr. Thus, the duration of the large perturbation of the carbon cycle affecting the Campanian, including both the pre-LCE and LCE, is estimated as ca. 1.3 Myr.

5.4. Relationship between clay mineralogy and δ13C

The clay mineral assemblages measured at Tercis-les-Bains and at Poigny are dominantly composed of smectite. This feature has been commonly observed in the Late Cretaceous clay sedimentation and attributed to hot semi-arid climatic conditions, high sea level, and volcanic activity (Chamley et al., 1990; Deconinck and Chamley, 1995; Jeans, 2006). Occasional detrital inputs are however prominent during the Campanian. A rise in detrital mineral content, including chlorite, illite, and kaolinite, is recorded in coincidence with the pre-LCE–LCE excursions (Fig. 3). At Poigny, the clay fraction of chalk shows an increase in detrital illite content at the expense of smectite from 11.5 to 61.5 m in depth (Deconinck et al., 2003; Fig. 4). Illite and chlorite are considered primary minerals originated from ancient rocks while kaolinite may be either reworked from the same detrital sources or from pedogenic blankets. As the three minerals fluctuate similarly, they probably have a common origin and their synchronous rise thus reflects increasing runoff. In that case, kaolinite cannot be taken as a good proxy of hydrolyzing conditions. This change in clay mineralogy starts within the pre-LCE interval and is fully expressed in the LCE interval. Perdieu et al. (2015) suggested that the pre-LCE interval corresponds to an amplification of the response of the carbon cycle to Milankovitch forcing prior to the LCE but did not discuss the main forcing environmental factors of these excursions. Here we show that, in both basins, the pre-LCE and LCE occurred during a period of increasing detrital inputs, which reflect enhanced erosion on continental massifs. Two hypotheses have been proposed to explain the illite input during the Late Campanian at Poigny: a climatic origin (cooling) or a tectonic episode (Riedel's Peine in the Paris Basin, Mortimore and Pomerol, 1997). More intensive erosion may also result from a sea-level fall. Eustatic variations indirectly affect the carbon cycle through changes in rates and sources of erosion. During the Campanian, sea-level changes have been correlated to δ13C excursions (Jarvis et al., 2002, 2006). High sea level recorded during the Cretaceous led to the formation of many shelf seas and shallow environments. Such environments are associated with enhanced primary productivity (phytoplankton) and/or enhanced preservation of organic matter (OM) in anoxic environments. These conditions could have promoted OM burial, which may well explain the relationship between positive δ13C excursions (SCBE and MCE) and transgressions (Jarvis et al., 2002, 2006). In contrast, negative δ13C excursions during the Late Cretaceous have been associated to sea-level falls and, more specifically, the negative shift of the LCE has been associated with the polyplacocum regression (Jarvis et al., 2002). Regression would have promoted erosion of the continents and the oxidation of OM by reworking continental and marine OM-rich levels (Jarvis et al., 2002, 2006; Voigt et al., 2012; Martinez and Dera, 2015), bringing isotopically light carbon to the oceans. It is therefore possible to propose a scenario for the LCE consistent with the observed isotopic and mineralogical data changes, considering that a drop in sea level was responsible for both a negative isotope excursion and a coeval detrital input. Further mineralogical and geochemical studies should be conducted on a wider scale to estimate the spatial extent of these changes.
5.5. Palaeotemperature trends

Bulk $\delta^{18}$O values should be considered with caution, as the original signal can be easily altered by diageneisis. Calculations of temperature based on the equation of Anderson and Arthur (1983) using a $\delta^{18}$O of seawater of $-1\%e$ to account for the absence of a well-developed ice-sheet (Shackleton and Kennett, 1975) would yield values ranging from 27 °C at the base of the Campanian to 18 °C in the upper part of the Campanian at Poigny. These temperatures are lower than temperatures calculated by Linnert et al. (2014) from TeX$_{sp}$ data at the same latitude in subtropical marine environments, which probably indicates an offset of the bulk-rock $\delta^{18}$O from original values during diageneisis. Nonetheless, temperature fluctuations can still be preserved in the trend of bulk-rock $\delta^{18}$O if the studied series displays a homogenous lithology (Jenkyns et al., 1994; Pellenard et al., 2014). As the Poigny borehole shows no major lithological change, the cooling trend potentially recorded here throughout the Campanian would be in agreement with the cooling of surface waters (Linnert et al., 2014) and bottom-waters (Friedrich et al., 2012) during this period.

5.6. Palaeoenvironmental scenario for the LCE

The additional Late Campanian isotopic and mineralogical data shown here offer new insights regarding the nature of the LCE. The increase in detrital flux in several sedimentary basins (Aquitaine Basin, Paris Basin, Umbria–Marches Basin, Saharan Platform) probably resulted from the intensification of continental erosion, and we estimate here a total duration of about 1.3 Myr for the large perturbation that comprises the LCE and pre-LCE. Geochemical data, notably the increase in the $^{87}$Sr/$^{86}$Sr ratio recorded during the R. calcarata zone in the Postalam section (Wagner et al., 2012), and during the LCE in the Lägerdorf–Kronsnoo section (McArthur et al., 1993), are also consistent with an intensification of continental weathering.

Higher levels of continental weathering can result from tectonic activity and/or sea-level changes. The Paris Basin experienced inversion processes during the Late Turonian (NE–SW compression), after an extensional period of subsidence (Albian to Turonian). The resulting deformation occurring during the Late Cretaceous (Guillocheau et al., 2000; Mansy et al., 2003) may have led to the observed stronger continental erosion. Otherwise, a global sea-level fall may have lowered base levels and enhanced erosion of continental areas as proposed by Jarvis et al. (2002, 2006) during the polyplucm event.

Continental erosion favored by newly exposed continental areas during the time interval spanned by the LCE would have ultimately led to consumption of atmospheric CO$_2$ through silicate weathering. The possible resulting decrease in the atmospheric pCO$_2$ could explain both the expression of the LCE and have contributed to the global cooling identified during the Campanian (Friedrich et al., 2012; Linnert et al., 2014).

This scenario can actually be tested through a simple isotope mass balance calculation. The postulated intensification of continental weathering lasted about 1.3 Myr, as estimated here from the duration of the interval from pre-LCE to LCE (i.e., the interval of significant change in clay mineralogy). It is assumed here that the ca. 1.0% negative carbon isotope excursion recorded in the carbonate rocks was caused by a significant increase in continental weathering and by the oxidation of OM (Kump, 1991; Kump and Arthur, 1999; Jarvis et al., 2002, 2006). We have adapted here the simple model of Kump and Arthur (1999), with initial (pre-LCE) steady-state atmospheric pCO$_2$ estimated at 1200 ppmv (Hong and Lee, 2012), and isotopic composition of oceanic carbonates ($\delta^{13}$C$_{Carb}$) estimated at +2.2‰. The more intense mid-ocean-ridge spreading of the Late Cretaceous leads us to increase the volcanic and metamorphic input of carbon by 30% (with regard to Phanerozoic average values proposed by Kump and Arthur, 1999). To reproduce the average 1.0‰ negative excursion of the LCE observed at Tercis–les-Bains and Poigny, the continental erosion needs to be multiplied by 1.5. As a consequence, total carbon burial would have increased by 50%.

In such a scenario, the increase in continental weathering significantly affects atmospheric pCO$_2$, which shows a rapid decrease from 1200 ppm to ~660 ppm. This drop in the atmospheric pCO$_2$ may have induced a cooling after the LCE, which is consistent with the isotopic data from the El Kef section (Tunisia; Jarvis et al., supplementary material, 2002) and from the Shuqualak–Evans borehole (Mississippi, USA, Linnert et al., 2014). This mechanism may explain the cooling phase observed in the Late Campanian–Maastrichtian (Friedrich et al., 2012; Linnert et al., 2014).

6. Conclusions

The integrated use of data from clay mineralogy and stable isotope geochemistry ($^{18}$O, $^{13}$C) reveals that a significant increase in detrital inputs of ilite and/or chlorite and kaolinite occurred in the Paris and Aquitaine basins during the time interval spanning the $\delta^{13}$C pre-LCE and LCE. This finding argues for the intensification of the hydrological cycle and/or of continental erosion at that time.

Based on cyclostratigraphic analyses performed on the $\delta^{13}$C of the Tercis–les-Bains section, the duration of the interval from the start of pre-LCE to LCE is estimated as ca. 1.3 Myr at least. The duration of LCE sensu stricto is estimated here as 0.8–0.9 Myr.

The more intense weathering of continental areas during the LCE was favored by a vast exposure of continents via the post-Turonian tectonic activity and enhanced by the polyplucm regression. Intense weathering is probably responsible for a pCO$_2$ decrease, which would have contributed to a global cooling in the Late Campanian.

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Appendix A. Supplementary data

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References


