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1 **Continental weathering as a driver of Late Cretaceous cooling: new insights from clay**
2 **mineralogy of Campanian sediments from the southern Tethyan margin to the Boreal**
3 **realm**

4
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20

21 Abstract

22

23 New clay mineralogical analyses have been performed on Campanian sediments from
24 the Tethyan and Boreal realms along a palaeolatitudinal transect from 45° to 20°N (Danish
25 Basin, North Sea, Paris Basin, Mons Basin, Aquitaine Basin, Umbria-Marche Basin and

26 Tunisian Atlas). Significant terrigenous inputs are evidenced by increasing proportions of
27 detrital clay minerals such as illite, kaolinite and chlorite at various levels in the mid- to upper
28 Campanian, while smectitic minerals predominate and represented the background of the Late
29 Cretaceous clay sedimentation. Our new results highlight a distinct latitudinal distribution of
30 clay minerals, with the occurrence of kaolinite in southern sections and an almost total
31 absence of this mineral in northern areas. This latitudinal trend points to an at least partial
32 climatic control on clay mineral sedimentation, with a humid zone developed between 20°
33 and 35°N. The association and co-evolution of illite, chlorite and kaolinite in most sections
34 suggest a reworking of these minerals from basement rocks weathered by hydrolysis, which
35 we link to the formation of relief around the Tethys due to compression associated with
36 incipient Tethyan closure. Diachronism in the occurrence of detrital minerals between
37 sections, with detrital input starting earlier during the Santonian in the south than in the north,
38 highlights the northward progression of the deformation related to the anticlockwise rotation
39 of Africa. Increasing continental weathering and erosion, evidenced by our clay mineralogical
40 data through the Campanian, may have resulted in enhanced CO₂ consumption by silicate
41 weathering, thereby contributing to Late Cretaceous climatic cooling.

42

43 Keywords: Campanian, Late Cretaceous cooling, clay minerals, carbon isotope stratigraphy,
44 climatic belt, continental weathering

45

46 1. Introduction

47

48 The Late Cretaceous is characterised by a long-term global climatic cooling from the
49 Turonian onward, with a marked acceleration during the Campanian ([Huber et al., 1995](#);
50 [Puc at et al., 2003](#); [Friedrich et al., 2012](#); [Linnert et al., 2014](#)). This period shows evidence of

51 an overall decrease in atmospheric CO₂ levels that likely contributed to this global cooling,
52 although the data remain scarce and do not allow to identify the timing and phases of CO₂
53 decline within the Late Cretaceous (Royer et al., 2012; Wang et al., 2014; Franks et al., 2014).
54 Decreasing mantle degassing linked to variations in seafloor production rates and continental
55 arc magmatism has been invoked to explain the observed decline in CO₂ levels during the
56 Late Cretaceous (Berner, 2004; Cogné and Hummler, 2006; Van Der Meer et al., 2014;
57 McKenzie et al., 2016).

58 Continental silicate weathering also governs atmospheric CO₂ on a multi-million year
59 time scale (Berner, 1990, 2004; Raymo and Ruddiman 1992; Dessert et al., 2003), but this
60 process remains poorly explored for the Late Cretaceous. Yet this period was marked by
61 major geodynamic changes, which included the initiation of Tethys Ocean closure (Dercourt
62 et al., 1986). This event, which is linked to the convergence of Africa toward Eurasia, was
63 associated with the development of topographic relief around the Tethys, including
64 lithospheric folds in Morocco (Frizon de Lamotte et al., 2011), and the 5000 km long chain of
65 relief developed across North Africa and the Middle East by the Ayyubid orogeny (Şengör
66 and Stock, 2014). This increased relief might be expected to have induced an increase in
67 continental silicate weathering, enhancing CO₂ consumption.

68 Clay minerals assemblages may be used to assess fluctuations in continental
69 weathering intensity (Chamley, 1989; Hermoso and Pellenard, 2014). Numerous clay
70 minerals data exist for the Cenomanian, Turonian and Maastrichtian stages but they remain
71 scarce for the Santonian–Campanian interval although many significant environmental
72 changes occurred at that time. A major reorganization of oceanic circulation took place during
73 the Santonian–Campanian, evidenced by neodymium isotope data (Robinson et al., 2010;
74 Martin et al., 2012; Moiroud et al., 2016). The Campanian stage was also characterised by a

75 significant cooling, as indicated by $\delta^{18}\text{O}$ values of benthic foraminifera and TEX_{86} data
76 (Friedrich et al., 2012; Linnert et al., 2014; O'Brien et al., 2017).

77 At a regional scale, preliminary clay minerals data from the Tercis-les-Bains section
78 (Aquitaine basin, North Atlantic influenced) and from the Poigny borehole (Paris basin,
79 Boreal influenced) have shown increasing inputs of detrital illite and kaolinite during the
80 Campanian (Chenot et al., 2016), coinciding with a global carbon-isotope negative excursion,
81 the so-called “Late Campanian Event” of Jarvis et al. (2002). This suggests that changes in the
82 carbon cycle at this time were accompanied by an increase of continental weathering. In order
83 to explore the spatial and temporal extent of these modifications in continental weathering, we
84 have acquired new data for clay minerals assemblages from Campanian sediments through 6
85 different sections and boreholes from the Tethyan and Boreal realms, along a transect from
86 $\sim 18^\circ$ to $\sim 42^\circ\text{N}$ palaeolatitude. Combined with previously published data sets, our work
87 provides the first constrains at the Tethyan scale on variations in continental weathering
88 induced by tectonic uplift during the Campanian.

89

90 2. Geodynamic framework and global palaeogeography of the studied sites

91

92 From the mid-Cretaceous onward, global plate tectonic changes induced modification of
93 the tectonic stress field in Europe. The Tethys Ocean began to close due to the anticlockwise
94 movement of Africa, with: (1) at the northern margin, the opening of the Bay of Biscay and
95 active subduction zones in Apulia, the Dinarides and Hellenids; and (2) at the southern
96 margin, the development of an intra-oceanic orogenic belt (Smith, 1971; Dewey et al., 1973;
97 Charvet, 1978; Bárdossy and Dercourt., 1990; Faccenna et al., 2001; Blakey, 2008; Kley and
98 Voigt, 2008; Voigt et al., 2008; Fig. 1).

99 To the north, the interplay of extensional and compressional tectonics, resulting from the
100 west-central Europe's thin lithosphere pinch between Baltica's and Africa's cratonic
101 lithospheres, induced NW–SE striking thrust faulting on the European plate (Kley and Voigt,
102 2008). These processes caused the development of subsiding basins (e.g. Sorgenfrei-Tornquist
103 Zone), and inversions of former depocentres (e.g. Mid-Polish Trough), with different rates of
104 subsidence (Kley and Voigt, 2008; Voigt et al., 2008). These newly created reliefs provided
105 detrital particles into the adjacent seas (Fig. 1). The geology of the southern margin of Central
106 Europe resulted from the convergence linked to the subduction zone between the European
107 and African plates.

108 Campanian palaeogeographic reconstructions show that the northern Tethyan margin was
109 partly covered by epicontinental seas (Fig. 1), with emerged lands representing remnants of
110 Variscan relief (e.g. Armorican, Central, Iberian, Ebro, Welsh, Rhenish, Bohemian massifs),
111 the inverted Mid-Polish Anticline (Voigt et al., 2008), and regional shoals (Dalmatian shoal,
112 High Karst; Charvet, 1978). The southern passive continental margin of the Tethys Ocean was
113 a wide platform (e.g. Saharan platform, Syrte basin), influenced by several detrital sources,
114 including locally emerged land masses such as Kasserine Island in Tunisia (Kadri et al., 2015)
115 and the north-western African craton (Fig. 1). However, the palaeogeography of Central
116 Europe is more difficult to reconstruct because of the large area affected by erosion during the
117 Late Cretaceous inversion (Voigt et al., 2008; Wolfgring et al., 2016; Neuhuber et al., 2016).

118

119 2.1. Boreal Realm

120

121 2.1.1. Danish North Sea and eastern Danish basin: Stevns-2 and Adda-3 boreholes

122

123 The Chalk Group of the Danish North Sea is well studied for its hydrocarbons reservoir
124 properties (Hardman, 1982; Megson and Tygesen, 2005). During the Late Cretaceous, the
125 Danish basin was bordered by the Baltic Shield to the northeast, the Grampian High to the
126 northwest, and the Rhenish–Bohemian Massif to the south (Fig. 1).

127 The 350 m long Stevns-2 core was drilled in Boesdal Quarry (55°15'31''N 12°24'04''E)
128 located in the eastern part of the Danish basin (palaeolatitude ~42°N; Philip and Floquet,
129 2000; Fig. 1). The core recovers a complete succession of upper Campanian to Maastrichtian
130 chalks, with a distinctive interval of alternating chalk-marl in the upper Campanian, a feature
131 that is also observed in the nearby Stevns-1 core and appears to characterise the whole Stevns
132 peninsula (Thibault et al., 2016a). The calcareous nannofossils biostratigraphy, high-
133 resolution carbon- and oxygen-isotope chemostratigraphy, and sedimentology of the Stevns-2
134 core have been described by Boussaha et al. (2016, 2017; Fig. 2).

135 The Adda-3 well in the Danish Central Graben (55°47'50''N 04°53'26''E) is located in the
136 southern part of the North Sea rift system (palaeolatitude ~45°N; Philip and Floquet, 2000;
137 Fig. 1). The Campanian interval, composed of bioturbated white chalks with occasional thick
138 flint bands and marly layers, occurs between 2200.8–2260.8 m depth. The calcareous
139 nannofossils biostratigraphy and stable-isotope geochemistry ($\delta^{13}\text{C}$, $\delta^{18}\text{O}$) have been
140 presented by Perdiou et al. (2016; Fig. 2).

141

142 2.1.2. Mons Basin: Cbr-7 borehole

143

144 The Mons basin (southern Belgium) was a transitional area between the North Sea and the
145 Paris basin to the WSW, bordered by the Rhenish Massif to the ENE (palaeolatitude ~37°N;
146 Philip and Floquet, 2000; Fig. 1). Based on lithology, the Campanian chalk of the Mons basin
147 has been subdivided into three formations (Cornet and Briart, 1870; Briart and Cornet, 1880):

148 the Trivières Chalk (white to grey marly chalk without flint); the Obourg Chalk (fine white
149 chalk with flint in the north of the Mons basin); and the Nouvelles Chalk (fine white chalk
150 without flint).

151 The 75 m-deep Cbr-7 borehole was drilled on the northern margin of the Hainaut-Sambre
152 quarry (50°25'10"N 04°1'33"E) located in the southeast of the Mons basin (Fig. 1).
153 Robaszynski and Anciaux (1976) subdivided the succession into the: Trivières Chalk (75.0–
154 46.5 m depth); Obourg Chalk (46.5–30.5 m); and Nouvelles Chalk (30.5–2.4 m depth).
155 Biostratigraphic data are scarce, but the uppermost part of the Trivières Chalk, the Obourg
156 Chalk and Nouvelles Chalk have been attributed to the lower part of the upper Campanian
157 according to the vertical distribution of foraminifera, belemnites and echinoids (Robaszynski
158 and Christensen, 1989; Fig. 3). A hardground exhibiting high Mn concentrations, acquired by
159 inductively coupled plasma-atomic emission spectrometry (ICP-AES) in the original study of
160 Richard et al. (2005) and an important $\delta^{13}\text{C}$ negative excursion of 0.5‰ amplitude occurs in
161 the uppermost part of the Trivières Chalk.

162

163 2.2. Tethyan Realm

164

165 2.2.1. Umbria-Marche basin: Gubbio – la Bottaccione and Furlo – Upper Road sections

166

167 A thick succession of Upper Cretaceous pelagic carbonates was deposited in the Umbria-
168 Marche basin in central Italy. During the Late Cretaceous, this deep basin was surrounded by
169 the High-Karst to the northeast (Charvet, 1978; palaeolatitude ~25°N; Philip and Floquet,
170 2000; Fig. 1). The symmetric NE–SW anticline of the Gubbio – la Bottaccione section
171 (43°21'45"N 12°34'57"E; Fig. 1) exposes a succession of pelagic carbonates from the Upper
172 Jurassic to the Palaeocene (~400 m-thick), followed by the first terrigenous turbidites within

173 the Miocene (Arthur and Fisher, 1977). The Campanian–Maastrichtian Scaglia Rossa
174 Formation is composed of pelagic carbonates with small quantities of iron oxides, including
175 magnetite and hematite, responsible for the pink colour of the limestones (Lowrie and
176 Alvarez, 1977; Channell et al., 1982; Lowrie and Heller, 1982).

177 The Campanian Scaglia Rossa Formation of Gubbio – la Bottaccione shows many
178 prominent 5 to 10 cm-thick cherty beds in the lower Campanian, overlain by a 5 m-thick
179 marly interval (Fig. 4). Many stratigraphic studies, including pioneering magnetostratigraphy
180 (Lowrie and Alvarez, 1977) and biostratigraphy based on foraminifera (Premoli Silva, 1977),
181 have been published for the section. In this paper we use the stratigraphic data from Coccioni
182 and Premoli Silva (2015) who recently revised the Upper Albian–Maastrichtian bio- and
183 magnetostratigraphy of this Tethyan reference section.

184 The Furlo – Upper Road section (43°38'29"N 12°42'36"E) located north of the Umbria-
185 Marche basin (Fig. 1) exposes pelagic carbonate deposits from the Jurassic to the Palaeocene
186 (~300 m-thick). A well-defined magnetostratigraphy (Alvarez and Lowrie, 1984) and a U/Pb
187 age from a bentonite layer identified within chron C33r (Mattias et al., 1988; Bernoulli et al.,
188 2004), provide a stratigraphic framework for the Campanian–Maastrichtian interval. Slope
189 deposits are expressed in the upper part of the section by the occurrence of more than 70 (10
190 to 100 cm-thick) white-coloured turbidites, and by a 12 m-thick slump at the base of chron
191 C33n (Fig. 4).

192

193 2.2.2. Tunisian Atlas: El Kef – El Djebil section

194

195 During the Campanian–Maastrichtian, the Saharan platform was located at ~18°N on the
196 southern margin of the Tethys Ocean (Philip and Floquet, 2000; Fig. 1). The closure of the
197 Tethys generated tectonic compressive and extensive domains; the Saharan platform belonged

198 to an external domain of intracontinental deformation, far from the intra-oceanic deformation
199 zone to the north (Aris et al., 1998; Boutib et al., 2000; Frizon de Lamotte et al., 2009; Bey et
200 al., 2012).

201 The 500-m thick El Kef – El Djebil section, located in the Tunisian Atlas immediately to
202 the north of the Saharan platform (36°10'37"N 08°44'05"E) corresponds to the Abiod Chalk
203 Formation. The Campanian–Maastrichtian Abiod Formation comprises three members
204 (Burollet, 1956; Jarvis et al., 2002; Fig. 3), from base to summit: white chalks with occasional
205 calciturbidites (lower chalk 'bar' or Haraoua Member); bioturbated marls with common
206 limestone beds (middle marl or Akhdar Member); and a yellow-greyish bioturbated chalky
207 unit (upper chalk 'bar' or Ncham Member). The biostratigraphic framework is mainly based
208 on well-preserved planktonic foraminifers, complemented by carbon isotope
209 chemostratigraphy (Robaszynski et al., 2000; Jarvis et al., 2002; Mabrouk El Asmi, 2014).

210

211 3. Materials and Methods

212

213 3.1. Oxygen and carbon isotopes

214

215 New stable-isotope data were generated for the two Italian studied sections
216 (Supplementary Data A, B). Wherever possible, samples were recovered for geochemical
217 analyses every metre from the Gubbio – la Bottaccione section and every half metre from the
218 Furlo – Upper Road section. Stable-isotope analyses of carbonate ($\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$) were
219 performed on bulk rocks collected along the whole of each section, from the Santonian–
220 Campanian boundary to the Campanian–Maastrichtian boundary (Fig 4). Isotopic analyses
221 were carried out at the Leibniz–Laboratory für Altersbestimmung und Isotopenforschung,
222 Christian–Albrechts University, Kiel, Germany. Samples devoid of macrofossils were crushed

223 in an agate mortar and pestle into fine and homogeneous calcite powders, which were reacted
224 with 100% phosphoric acid at 70°C and analysed using a ThermoScientific MAT253 mass
225 spectrometer, connected to a Kiel IV preparation device. Eleven samples from the Gubbio – la
226 Bottaccione section were additionally analysed at the Biogéosciences Laboratory, University
227 of Bourgogne Franche-Comté, Dijon, France. Here, calcite was reacted with 100% phosphoric
228 acid at 90°C using a Multiprep online carbonate preparation line connected to an Isoprime
229 mass spectrometer.

230 All isotopic values are reported in the standard δ -notation in per mil relative to V-PDB
231 (Vienna Pee Dee Belemnite) by assigning a $\delta^{13}\text{C}$ value of +1.95‰ and a $\delta^{18}\text{O}$ value of
232 –2.20‰ to NBS19. External reproducibility as determined by replicate analyses of laboratory
233 standards was $\pm 0.08\text{‰}$ (2σ) for oxygen isotopes in both laboratories and $\pm 0.05\text{‰}$ (2σ) for
234 carbon isotopes at Leibniz – Laboratory and $\pm 0.04\text{‰}$ at the Biogéosciences Laboratory.

235

236 3.2. Clay mineralogy

237

238 All bulk-rock samples were collected in the field and from the core storage facility with
239 a regular sample spacing (Figs 2–4). Mineralogical analyses were performed at the
240 Biogéosciences Laboratory, University of Bourgogne Franche-Comté. Clay minerals
241 assemblages were identified by X-ray diffraction (XRD) on oriented mounts of non-
242 calcareous clay-sized particles ($< 2\ \mu\text{m}$). The procedure described by Moore and Reynolds
243 (2009) was used to prepare all samples to better compare the integrity of the dataset and to
244 avoid discrepancies due to the process of quantification. Diffractograms were obtained using a
245 Bruker D4 Endeavor diffractometer employing CuK_α radiation with a LynxEye detector and
246 Ni filter, under 40 kV voltage and 25 mA intensity. For each sample, three preparations were

247 analysed: after air-drying; after ethylene-glycol solvation; and after heating at 490°C for 2
248 hours. The goniometer was scanned from 2.5° to 28.5° 2 θ for each run.

249 Clay minerals were identified by the positions of their main diffraction peaks on the
250 three XRD runs (Table 1), while semi-quantitative estimates were produced in relation to their
251 peak areas (Moore and Reynolds, 2009). Peak areas were determined on diffractograms of
252 glycolated runs with MacDiff 4.2.5 software (Petschick, 2010). The percentages of kaolinite
253 and chlorite were determined by deconvolution of the d(002)_{kaolinite} and d(004)_{chlorite} peak
254 areas that appear respectively at 3.57 and 3.52Å and using the 7.1Å peak area common to
255 both minerals. Beyond the evaluation of the absolute proportions of the clay minerals, the aim
256 was to identify their relative fluctuations through the sections.

257 Measurement of the relative proportions of smectite and illite layers in the R0 mixed
258 layers were performed on the diffractograms, following two methods: the procedure of Moore
259 and Reynolds (2009) and the determination of the saddle Index after Inoue et al. (1989). In
260 most cases, the procedure of Moore and Reynolds was performed on reflections 001/002 and
261 002/003. However, the presence of illite at 10Å stretches the 001/002 reflection and modifies
262 the true position of 001/002. In this case, estimation of the smectite layers was determined
263 only using the reflection 002/003.

264

265 3.3. Correlation of the studied sites: $\delta^{13}\text{C}_{\text{carb}}$ isotopic events

266

267 We have used global and local carbon-isotope events previously recognised in the
268 Campanian to correlate sections and boreholes. During the Campanian, seven isotopic events
269 have been identified.

270 - The Santonian–Campanian Boundary Event (SCBE), consisting of a global positive
271 shift of $\delta^{13}\text{C}_{\text{carb}}$, with a varying amplitude of 0.3‰ to 2.9‰ (Jarvis et al., 2002; Gale et

272 al., 2008; Wendler, 2013; Thibault et al., 2016b; Dubicka et al., 2017), has been
273 widely recognised in successions throughout the Boreal, Tethyan, North Pacific and
274 Central Atlantic realms (Table 2, Event 1). This event coincides with the C34/C33r
275 chron boundary and the Highest Occurrence (HO) of the crinoid *Marsupites*
276 *testudinarius*, both defining the Santonian–Campanian boundary (Italy, Gubbio – la
277 Bottaccione, Premoli Silva and Sliter, 1994; Texas, Waxahachie Dam Spillway, Gale
278 et al., 2008; Poland Boceniec, Dubicka et al., 2017).

279 - The *papillosa* Zone Event (PZE) is a positive excursion of ~0.2‰ coincident with a
280 medium-term $\delta^{13}\text{C}$ maximum, occurring in the mid-Lower Campanian *papillosa* zone
281 at Lägerdorf and in the uppermost *Globotruncana elevata* zone (Chron 33r/33n
282 boundary) on the Bottaccione section. Its stratigraphic significance still needs to be
283 tested by additional high-resolution data sets (Thibault et al., 2016b; Sabatino et al.,
284 2018; Table 2, Event 2).

285 - The Mid-Campanian Event (MCE), first described by Jarvis et al. (2002) on the
286 $\delta^{13}\text{C}_{\text{carb}}$ curves of El Kef (Tethyan realm, Tunisia), Bidart (North Atlantic realm,
287 France) and Trunch (Boreal realm, England, Jenkyns et al., 1994), is defined by a
288 positive excursion of 0.3‰ occurring near the base of *Globotruncana ventricosa*
289 planktonic foraminifera zone and the base of the upper Campanian (Table 2, Event 3).
290 At Tercis-les-Bains, this event occurs at the base of chron C33n, comprising the
291 Lowest Occurrence (LO) of *Rucinolithus magnus* and *Uniplanarius gothicus*
292 nannofossils. This event has been recognised at a larger scale by Perdiou et al. (2016)
293 in the North Sea.

294 - The Conica Event (CE; Perdiou et al., 2016) is a small negative excursion of $\delta^{13}\text{C}_{\text{carb}}$
295 of ~0.4‰, occurring at the base of the *conica-senior* macrofossil zone, in the Boreal
296 and North Atlantic realms (Table 2, Event 4).

- 297 - The Late Campanian Event (LCE) is a global event, identified in the Tethyan, Boreal,
298 Central Pacific, Indian Ocean and North Atlantic realms (Jarvis et al., 2002, 2006;
299 Voigt et al., 2010, 2012; Thibault et al., 2012b; Sabatino et al., 2018). It consists of a
300 marked negative excursion of $\delta^{13}\text{C}_{\text{carb}}$ with an amplitude varying between 0.3‰ and
301 1.3‰. It is located within the middle of chron C33n (Table 2, Event 5). This event
302 occurs within (Gubbio, Voigt et al., 2012; Sabatino et al., 2018) and/or immediately
303 above (El Kef, Jarvis et al., 2002; Tercis-les-Bains, Voigt et al., 2012, Chenot et al.,
304 2016) the *Radotruncana calcarata* planktonic foraminifera zone (UC15 d-e
305 nannofossil zone) in the Tethyan realm and in the mid-*Belemnitella mucronata*
306 macrofossil zone in the Boreal realm. Perdiou et al. (2016) identified two steps in this
307 isotopic event, called the pre-LCE and the main-LCE, which coincide with an increase
308 of illite, kaolinite and chlorite in the Aquitaine basin (Chenot et al., 2016) and illite in
309 the Paris basin (Deconinck et al., 2005).
- 310 - The Epsilon event (EE or C1-) was defined by Thibault et al. (2012a) as a negative
311 excursion of $\sim 0.25\text{‰}$, followed by a positive shift of 0.2‰ and a second negative
312 excursion of 0.1‰ , which imparts to this event a sharp resemblance to the Greek letter
313 ϵ . The event occurs slightly above the HO of *Eiffellithus eximius* in the Boreal,
314 Tethyan and North Atlantic realms (Table 2, Event 6).
- 315 - The Campanian–Maastrichtian Boundary Event (CMBE, Voigt et al., 2010, 2012) is a
316 global negative event with an amplitude ranging from 0.3‰ to 1‰ , occurring in the
317 reverse chron C32n2n, and identified in many sedimentary basins from the Tethyan to
318 Pacific realms (Table 2, Event 7). This isotopic event was divided into three steps by
319 Thibault et al. (2012a): CMBa (negative excursion of 0.6‰); CMBb (positive
320 excursion of 0.2‰); CMBc (negative excursion of 0.4‰). The definition of the
321 CMBE by Voigt et al. (2012) is different as it comprises the whole long-term decrease

322 in carbon isotopes ranging from the middle of chron C32n2n up to the lower half of
323 chron C31r and is further described by 5 distinct small positive peaks that are
324 superimposed on the long-term trend (CMBE1 to CMBE5).

325

326 3.4. Calcareous nannofossils biostratigraphy

327

328 The El Kef – El Djebil section constitutes one of the reference isotopic curves for the
329 Campanian of the Tethys (Jarvis et al., 2002) but so far, no calcareous nannofossils
330 biostratigraphy was available for that section. In this study, we analysed 17 samples from the
331 archive of Jarvis et al. (2002) in order to establish a coarse calcareous nannofossils
332 biostratigraphy that can be directly correlated to the already available isotope curve. Standard
333 smear-slides were prepared following the methodology described in Bown (1998) and the
334 biostratigraphy is based solely on presence/absence of individual taxa observed in crossed
335 nicols at a magnification of x1000 on a Leica DM750P optical microscope. The CC
336 nannofossil zonation of Sissingh (1977) modified by Perch-Nielsen (1985) and the UC^{TP}
337 (Tethyan Province) zonation of Burnett (1998) have been applied (Fig. 5).

338

339 3.5. Calcium carbonate content

340

341 Calcimetry was performed on the Gubbio – la Bottaccione and Cbr-7 samples at the
342 Biogéosciences Laboratory, University of Bourgogne Franche-Comté, using a Bernard
343 calcimeter. The samples were treated with hydrochloric acid and the CO₂ released was used to
344 quantifying the percentage of CaCO₃. The data are reported on Fig. 3 and Supplementary
345 Data C for the Cbr-7 borehole and in Supplementary Data B for the Gubbio – la Bottaccione
346 section.

347

348 4. Results

349

350 4.1. Clay mineralogy

351

352 The clay fraction of the six studied sites is composed predominantly (often more than 80%)
353 of R0 random illite/smectite mixed-layers; hereafter referred to as smectitic minerals (not
354 represented on Figs. 2–4 to emphasize other clay minerals variations; Supplementary Data A,
355 B; C, D, E, F). This result was expected, since Upper Cretaceous sediments are characterised
356 by an abundance of these minerals, considered to be the background of the clay sedimentation
357 (Deconinck and Chamley, 1995; Deconinck et al., 2005; Jeans, 2006; Chenot et al., 2016).
358 Other clay minerals, occurring in significant proportions, include illite, chlorite, kaolinite and
359 palygorskite.

360

361 4.1.1. Stevns-2 and Adda-3 boreholes

362

363 Illite (less than 10% of the clay fraction) occurs in most samples from the Stevns-2
364 borehole. Traces of chlorite (less than 2%) are recorded between 349.9 and 251.7 m depth
365 corresponding to beds of alternating chalk-marl and higher gamma-ray values of the upper
366 Campanian; this interval is associated with a warm optimum preceding the early
367 Maastrichtian cooling (Thibault et al., 2016a; Boussaha et al., 2017). Chlorite is absent above
368 the Campanian–Maastrichtian boundary (Fig. 2; Supplementary Data D). The clay minerals
369 assemblage of the Adda-3 borehole includes traces of illite and traces of kaolinite in all
370 samples (Fig. 2; Supplementary Data E). Based on the methods of Moore and Reynolds
371 (2009) and Inoue et al. (1989), estimation of the smectite layers in the IS R0 comprises

372 between 50 and 80% in the Stevns-2 borehole (Supplementary Data D) and between 85 and
373 95% in the Adda-3 borehole (Supplementary Data E). The evolution of the percentage does
374 not show any trend.

375

376 4.1.2. Cbr-7 borehole

377

378 According calcium carbonate content data, the percentage of the clay fraction is estimated
379 to range between 5% from 75 to ~40 m depth to < 3% from ~40 m to the top of the core (Fig.
380 3; Supplementary Data C).

381 In the Trivières Chalk, at the base of the core, beside smectitic minerals, the clay
382 assemblages consist of 20% illite, with small quantities of kaolinite that decrease from the
383 base to the top of the formation (from less than 10% to traces). Kaolinite disappears in the
384 overlying Obourg Chalk, while illite occurs in small proportions in this formation, together
385 with traces of fibrous clays (palygorskite), clinoptilolite and talc. The Nouvelles Chalk is
386 characterised by increasing proportions of illite with varying quantities of fibrous clays,
387 clinoptilolite and talc (Fig. 3).

388 To estimate the percentage of smectite layers in the IS R0, the method of Moore and
389 Reynolds (2009) could not be performed on the Cbr-7 borehole diffractograms because
390 reflections 001/002 and 002/003 are poorly expressed. However, the method of Inoue et al.
391 (1989) shows an evolution of the Saddle Index similar to the trend of illite. First, from 80 to
392 45 m depth, the saddle index displays a decreasing trend from 0.7 to 0.3, which means an
393 increase of smectite layers in the IS R0 (from ~60 to ~80%). In a second part, the saddle index
394 records an increasing trend between 45 – 10 m depth from 0.3 to 1, interrupted by highest
395 values around 35 m, corresponding to smectite layers ranging from ~80 to ~50%. At the top

396 of the borehole, from 7 to 3 m depth, the saddle index displays the lowest values around 0.5,
397 equivalent to about 70% of smectite layers (Supplementary Data C).

398

399 4.1.3. Gubbio – la Bottaccione and Furlo – Upper Road sections

400

401 In the Gubbio – la Bottaccione section, from the base to 45 m, beside abundant smectitic
402 minerals, the clay fraction consists of illite (10%) with occasional traces of kaolinite and
403 chlorite (Fig. 4; Supplementary Data B), while from 50 to 70 m, kaolinite and chlorite occur
404 systematically and increase up to maxima of >5% along with abundant illite (50%). In the
405 uppermost part of the section, the proportions of kaolinite and chlorite decrease and kaolinite
406 essentially disappears from 80 m upwards. From the base to 50 m, the percentage of smectite
407 layers in the IS R0 is estimated to ~70%, whereas from 50 to 80 m, it decreases until ~50%.
408 From 80 m to the top of the section, smectite layers increase again to ~65 % (Supplementary
409 Data B).

410 From the base of the Furlo section to 44 m, the percentage of illite is around 10% with
411 traces of kaolinite (Fig. 4; Supplementary Data A). From 57 m, kaolinite increases
412 significantly, rising upwards to >10% at 75 m, together with more abundant illite and traces
413 of chlorite. Interestingly, the onset of this major mineralogical change coincides with the
414 appearance of turbidites above a 12 m-thick slump. The percentage of smectite layers in the
415 IS R0 follows an opposite trend compared to illite until 80 m: from the base of the section to
416 44 m the highest percentage of smectite layers is estimated to ~70% and then progressively
417 decreases down to ~50%. However, from 80 m to the top of the section, the percentage of
418 smectite layers in IS R0 seems to increase again up to ~70% (Supplementary Data A).

419

420 4.1.4. El Kef – El Djebil section

421

422 In the sediments from El Kef – El Djebil section, the evolution of the clay fraction, again
423 dominated by smectitic minerals, is divided into several mineralogical zones (Fig. 3;
424 Supplementary Data F). Traces of illite are recorded in most samples. By contrast, kaolinite
425 and chlorite are more abundant in 3 intervals. In the first interval, from 0–60 m, the
426 percentages of kaolinite and chlorite diminish and these minerals disappear upward. In the
427 second interval, from 80–330 m, the proportion of kaolinite increases to a maximum of 10%
428 around 180 m, and then decreases progressively to 330 m. Chlorite shows a similar trend,
429 with maximum values of around 5% at 180 m. Interestingly, the highest percentages of
430 chlorite and kaolinite occur at the transition between the lower white chalks and the
431 bioturbated marls. In the third interval, from 410–430 m, the proportion of kaolinite again
432 rises up to 10%, along with traces of chlorite, and then falls from 430–465 m. However, the
433 values in the third interval must be interpreted with caution, because of the high proportion of
434 Si which distorts the peak area on the diffractogram required for percentage calculation (Si/Al
435 ratio determined by an ICP-AES by Mabrouk El Asmi, 2014; Fig. 3).

436

437 4.2. Isotope analyses

438

439 4.2.1. Gubbio – la Bottaccione section

440

441 Bulk-rock $\delta^{13}\text{C}$ values range from about 2.1 to 2.7‰ through the Gubbio – la Bottaccione
442 section (Fig. 4; Supplementary Data B). Three moderate isotopic excursions may be
443 identified. From 1.2–9.5 m, a 0.3‰ positive shift is observed at the Santonian–Campanian
444 transition and the C34/C33r chron boundary. A second positive shift of 0.4‰ starts at 23.8 m
445 and ends at 40.3 m, before the first exposure gap, at the base of the *Contusotruncana*

446 *plummerae* zone. After increasing from 2.3 to 2.6‰ up to 61 m, $\delta^{13}\text{C}$ values decrease down to
447 about 2.4‰ up to 83 m. This decrease mostly occurs in the *Radotruncana calcarata* zone and
448 the upper part of chron C33n. From 95.8 m to the top of the section, the bulk-rock $\delta^{13}\text{C}$ values
449 display a decreasing trend of 0.2‰, which coincides with the Campanian–Maastrichtian
450 transition, within the *Gansserina gansseri* zone.

451 Bulk-rock $\delta^{18}\text{O}$ values display an increasing trend from values of about -3.0‰ at the base
452 of the Gubbio – la Bottaccione section, to values of about -2.2‰ around 40 m, and do not
453 show any further trend for the remaining of the section (Fig. 4; Supplementary Data B).

454

455 4.2.2. Furlo – Upper Road section

456

457 Bulk-rock $\delta^{13}\text{C}$ values range from about 2.0 to 2.9‰ in the Furlo – Upper Road section
458 (Fig. 4; Supplementary Data A). Between 1.3 and 5.9 m, a first two-step positive excursion of
459 0.4‰ coincides with the C34/C33r chron boundary and the Santonian–Campanian transition.
460 This excursion is followed by a negative excursion of about 0.4‰ with $\delta^{13}\text{C}$ values reaching
461 2.2‰ at 17 m. An increase in $\delta^{13}\text{C}$ values is then recorded, with maximum values of about
462 2.9‰ at about 30 m. From about 30 m to 44 m, the data show an overall decreasing trend
463 from values of about 2.9‰ to values of about 2.6‰, with one sample at the top of the interval
464 displaying a value of 2.4‰. It can be noted that this interval coincides with the first
465 occurrence of turbidites.

466 Above the 12 m-thick slump in the middle part of the studied section, bulk-rock $\delta^{13}\text{C}$ data
467 decrease from values of 2.6‰ at 56.7 m to values on average 2.3‰ at 92 m. This trend is
468 interrupted by a two-step negative excursion of 0.5‰, from 57.7 to 65.5 m, and from 65.5 to
469 72.5 m, occurring in the middle of chron C33n. Above 92 m, following a small positive
470 excursion of about 0.2‰, $\delta^{13}\text{C}$ values decrease again upwards, from $\delta^{13}\text{C}$ values of 2.5‰ at

471 93 m, and down to minimum values of 2‰ at the top of the section. The onset of this decrease
472 coincides with the C33n1r/C32n2n chron boundary.

473 The $\delta^{18}\text{O}$ values remain quite stable in the lower part of the section, from the base to 44 m,
474 with values around -2‰ on average (Fig. 4; Supplementary Data A). In the upper part of the
475 section, they display a slight increase, from values of about -2‰ to values of about -1.5‰
476 around 80 m, before decreasing again to about -2‰ at the top of the section.

477

478 4.3. Nannofossils bioevents and biozonation of El Kef

479 A total of 17 samples of the El Kef section were studied here. Sample 81 (98.55 m) was
480 barren, but the remainder of the samples yielded numerous late Cretaceous specimens. In total
481 61 individual calcareous nannofossil species were recognized in this study, a number which is
482 rather low for the Campanian as this stage is characterised by the highest species richness of
483 the whole Mesozoic (Bown, 2004). The low diversity recorded in El Kef section is likely the
484 result of the preservation of the assemblage that is at best moderate. *Watznaueria barnesiae*,
485 *Cribrosphaerella ehrenbergii*, *Prediscosphaera cretacea* and *Zeugrhabdotus bicrescenticus*
486 are common.

487 The studied samples belong to the interval from zone CC18-CC19 to CC24 of Sissingh
488 (1977) and UC14a^{TP}-UC15b^{TP} to UC18 of Burnett (1998) due to the presence of *Broinsonia*
489 *parca constricta* in the first sample (K1, 2.2 m) and the HO of *Tranolithus orionatus* in
490 sample K447 (474.6 m). According to this applied biozonation, the age of the studied interval
491 should range from the late early Campanian to early Maastrichtian. However, the
492 inconsistency in the order of last occurrences recorded in the upper part of the section with
493 respect to global schemes as well as with respect to other stratigraphic considerations,
494 suggests that the zonations are hardly applicable to the section and that the whole studied

495 succession remains restricted to the Campanian only (see Discussion). A summary of the
496 main results is provided in Fig. 5 (range chart with zonation and bioevents).

497

498 5. Discussion

499

500 5.1. Influence of diagenesis

501

502 A prerequisite for the use of clay minerals for palaeoenvironmental reconstructions is
503 that they should mainly have a detrital origin. Smectitic minerals, which are the background
504 of clay sedimentation in most latest Cretaceous open marine environments, are very sensitive
505 to burial diagenesis. Illitization processes start when the temperature reaches about 60°C,
506 (Kübler and Jaboyedoff, 2000; Kübler and Goy-Eggenberger, 2001) and IS R0 are
507 progressively transformed into I/S R1, then R3, and finally into illite (Środoń, 2009). In each
508 studied section, the illitization process is considered as subsidiary because of the high
509 abundance of IS R0 throughout the successions (Supplementary Data A, B, C, D, E, F;
510 Delissanti et al., 2010). In addition, low T_{\max} values, comprised between 402° and 433°C,
511 have been determined in previous studies for the “Bonarelli level” (OAE 2) of the Furlo
512 section (Scaglia Rossa Formation), that corresponds to the immature zone of the organic
513 matter, thereby suggesting negligible burial diagenesis consistent with persistence of IS R0
514 (Mort et al., 2007; Delissanti et al., 2010).

515 The $\delta^{13}\text{C}$ values of sediments from the Gubbio – la Bottaccione and Furlo – Upper Road
516 sections comprised between 2.0 and 3.0‰, match $\delta^{13}\text{C}$ values typically observed in latest
517 Cretaceous marine sediments of the Boreal and Tethyan realms (Jenkyns et al., 1994; Jarvis et
518 al., 2002; Voigt et al., 2012; Figs. 6, 7). A cross-plot between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values shows
519 close agreement for carbon between Furlo and Gubbio but a difference in oxygen, with

520 samples from Gubbio presenting lower values. A Spearman's coefficient was computed for
521 each dataset to test the existence of a correlation within the data of each section. This method
522 was chosen because of the non-linear nature of the relationship between the two variables,
523 $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ (Chenot et al., 2016). A value of 1 indicates a perfect correlation, a value of -1
524 a perfect anti-correlation, while 0 indicates an absence of correlation. The two datasets
525 generate Spearman's coefficients of $r_s = -0.66$ for Furlo and $r_s = +0.11$ for Gubbio (Fig. 6),
526 pointing to an absence of a significant correlation between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values within each
527 dataset.

528 The isotope data do not exhibit any inverted J curve in $\delta^{13}\text{C}$ – $\delta^{18}\text{O}$ space that could
529 reflect diagenesis involving fluid-rock interactions in addition to physical mixture of different
530 diagenetic mineral phases (e. g. Bishop et al., 2014). This supports an absence of extensive
531 diagenesis affecting both isotopic systems, and argues in favour of a preservation of $\delta^{13}\text{C}$
532 values. By contrast, the markedly lower $\delta^{18}\text{O}$ values recorded at Gubbio, which presents a
533 deeper depositional environment compared to Furlo, likely indicate an impact of diagenesis
534 on oxygen isotopes, although probably limited considering that the values fall within the
535 range of shallow buried pelagic carbonate successions of comparable age elsewhere (e.g.
536 Voigt et al., 2010; Fig. 7).

537

538 5.2. Identification and redefinition of carbon-isotopic events

539

540 In order to correlate the studied sections and boreholes, we used seven carbon-isotopic
541 events, namely the: (1) Santonian–Campanian Boundary Event; (2) *papillosa* Zone Event; (3)
542 Mid-Campanian Event; (4) Conica Event; (5) Late Campanian Event; (6) Epsilon Event and
543 (7) Campanian–Maastrichtian Boundary Event, described in Section 3.3 (Fig. 8; Table 2).

544

545 5.2.1. Identifications of carbon-isotopic events in the Gubbio – la Bottaccione and Furlo –
546 Upper Road sections

547

548 A 0.4‰ positive shift in the basal Gubbio – la Bottaccione section (between 1.2 and 9.5 m)
549 and a two-step positive carbon-isotopic excursion of 0.4‰, recorded on the Furlo – Upper
550 Road section (between 1.3 and 5.9 m), coincide with the Santonian–Campanian transition and
551 the C34/C33r chron boundary respectively and are therefore attributed to the SCBE (Figs. 4;
552 8).

553 A sharp increase of 0.4‰ in the $\delta^{13}\text{C}$ curve identified within the mid-Lower Campanian in
554 the Gubbio – la Bottaccione section, starting at the base of the chron C33n and ending at the
555 base of the *Contusotruncana plummerae* zone, is assigned to the *papillosa* Zone Event (cf.
556 Thibault et al., 2016b; Sabatino et al., 2018). This excursion is not visible in the Furlo –
557 Upper Road section profile where it may be obscured or missing due to the slump interval
558 (Figs. 4, 8).

559 The MCE has been previously well-defined on the Bottaccione section, occurring above
560 the LOs of *Rucinolithus magnus* and *Uniplanarius gothicus*, but is not recorded on the
561 isotopic curve of this study, maybe because of the low amplitude of this isotopic event,
562 around 0.2‰ (Thibault et al., 2016b; Sabatino et al., 2018).

563 The LCE is well defined in the Gubbio – la Contessa section (Voigt et al., 2012) and in the
564 Gubbio – la Bottaccione section (Sabatino et al., 2018; Figs. 4, 8) with an amplitude of around
565 0.4‰. However, in this study, the LCE seems to be poorly expressed on our Gubbio – la
566 Bottaccione profile, likely because of the occurrence of several gaps in the sedimentary record
567 of this section. However, the decreasing trend of 0.25‰ from 61 to 83 m identified in the la
568 Bottaccione section in the upper part of the chron C33n, including the *Radotruncana*
569 *calcarata* zone (Figs. 4, 8), is tentatively associated to the target horizon where the LCE could

570 be expected. In the Furlo – Upper Road section, the two-step negative excursion of 0.5‰,
571 above the 12 m-thick slump and occurring in chron C33n may be ascribed to the pre-LCE and
572 main-LCE (Figs. 4, 8).

573 The onset of the CMBE is well represented in both the Gubbio – la Bottaccione (starting at
574 98.5 m) and Furlo – Upper Road (starting at 94.1 m) sections (Figs. 4, 8).

575

576 5.2.2. Identification and attribution of carbon-isotope events from other sections

577

578 5.2.2.1. El Kef – El Djebil section

579

580 The calcareous nannofossil record of the upper part of the El Kef section (335 to 496
581 m) is difficult to interpret. Voigt et al. (2012, Fig. 6) documented a consistent succession of
582 closely spaced nannofossil HOs that occur in the Lower Maastrichtian at Gubbio and
583 Lägerdorf-Kronsmoor-Hemmoor, all within the upper part (their intervals 3 – 5) of the
584 CMBE: *U. trifidus*, *B. parca constricta*, *T. orionatus*. The consecutive datum levels lie on a
585 marked falling trend in the $\delta^{13}\text{C}$ curve, which precedes a sharp rise in values towards the Mid-
586 Maastrichtian Event (MME), above. The same succession of nannofossil disappearances at El
587 Kef are spread through >100 m of the Ncham Chalk, largely within an interval of relatively
588 high $\delta^{13}\text{C}$ values (Figs. 3, 8) above the negative excursion defining the LCE.

589 The HO of *U. trifidus* has been recently considered as the best nannofossil marker for
590 the Campanian/Maastrichtian boundary and was used to subdivide UC zone 16 into subzones
591 UC16a^{TP} and UC16b^{TP} (Thibault, 2016). The HO of *U. trifidus* at 403 m would thus suggest
592 that this boundary lies close to that stratigraphic height, and that the negative excursion
593 recorded between 339 and 390 m corresponds to the CMBE, not the LCE. However, the HO
594 of *B. parca constricta*, which marks the top of UC18 and should thus be recorded higher up at

595 El Kef was found here below the HO of *U. trifidus*. Moreover, a narrow range of the curved
596 spine nannolith at El Kef, straddling the negative excursion, contradicts other nannofossil
597 results and the identification of this excursion as the CMBE. Indeed, the HO of the curved
598 spine is generally restricted to the late Campanian, and correlates to the top of the LCE
599 (Thibault et al., 2012a; Thibault, 2016; M.R. Razmjooei and N. Thibault, pers. comm.,
600 unpublished results from Iran). It is therefore possible that the HOs of *B. parca constricta* and
601 *U. trifidus*, which are nearly identical between El Kef (this study) and Kalaat Senan
602 (Robaszynski et al., 2000) are recorded in the region of the Kasserine Island much earlier than
603 in the rest of the Tethys due to local environmental conditions.

604 The negative $\delta^{13}\text{C}$ excursion between 339 and 390 m at El Kef lies in the upper part of
605 the Akhdar Marl Member *sensu* Jarvis et al. (2002). Robaszynski et al. (2000) recorded the
606 Upper Campanian ammonite *Nostoceras (Nostoceras) hyatti* above this in the lowest part of
607 the overlying Ncham Chalk Member. The index ammonite taxon *Nostoceras*
608 (*Bostrychoceras polyplacum*) ranges through the lower part of the Akhdar Marl, overlapping
609 with the base of the *G. calcarata* foraminifera Zone, above. The negative $\delta^{13}\text{C}$ excursion
610 occurs above the top of the *G. calcarata* Zone and below the base *N. hyatti*. This stratigraphic
611 relationship between the excursion and the macrofossil datum levels is essentially identical to
612 that recorded for the Late Campanian Event (LCE) in the Tercis-les-Bains GSSP by Voigt et
613 al. (2012), and is well below the base of the CMBE. These data support the original
614 interpretation of Jarvis et al. (2002) that the negative $\delta^{13}\text{C}$ excursion at El Kef is the LCE not
615 the CMBE.

616 The planktonic foraminifera biostratigraphy is not definitive: *G. gansseri* occurs very
617 infrequently at its lowest occurrence at Kalaat Senan (Robaszynski et al. 2000, Figs. 3, 8), and
618 is only recorded consistently in the boundary interval between the Akhdar and Ncham
619 Members ('Gorbeuj Member' of Robaszynski et al. 2000). The first consistent occurrence

620 correlates to above the negative excursion at El Kef (Jarvis et al. 2002, Fig. 2). Unfortunately,
621 *G. gansseri* was not identified in the low-resolution foraminifera study of El Kef (El Djebil)
622 by De Cabrera (in Jarvis et al. 2002).

623 Finally, attributing the main negative excursion at El Kef to the CMBE rather than the
624 LCE would necessitate major changes in sedimentation rate within the section that are not
625 supported by coincident lithological variation, and/or challenges the stratigraphic
626 completeness of the logged succession. In conclusion, we retain here the original carbon
627 isotope event assignments of Jarvis et al. (2002), while acknowledging the challenges of the
628 stratigraphy that warrant further study.

629

630 5.2.2.2. Cbr-7 borehole

631

632 The short 1‰ negative excursion of $\delta^{13}\text{C}$ measured by Richard et al. (2005) in the Cbr-7
633 borehole (Figs. 3, 8) was not ascribed to any recognized isotopic event. The Mn peak
634 observed in this borehole corresponds to a glauconite-bearing hardground, which points to the
635 existence of a condensed interval and/or hiatus, consistent with Mn accumulation. In this case,
636 the 1‰ excursion recorded in $\delta^{13}\text{C}$ values may correspond to the LCE, according to its
637 occurrence within the lower part of the upper Campanian.

638 However, the hardground at the summit of the Trivières Chalk occurs in the basal
639 *Belemnitella mucronata* zone (Robaszynski et al., 2001; Richard et al., 2005), i.e. very close
640 to the base of the Upper Campanian as calibrated in NW Europe. Interestingly, a succession
641 of hardgrounds occurs at a very similar stratigraphic level in the Trunch borehole of Eastern
642 England, called the Trunch Hardgrounds by Jarvis et al. (2002). The MCE is placed
643 immediately above this condensed interval in the Trunch borehole. If these hardgrounds are
644 equivalent to the Trivières hardground, then the negative $\delta^{13}\text{C}$ excursion in the Cbr-7 borehole

645 correlates to the negative excursion immediately below the Trunch Hardgrounds, and the
646 small peak above the negative excursion in Cbr-7 would be the MCE. Decreasing-upward
647 values at the top of Cbr-7 in the Nouvelles Chalk can then be interpreted to represent the
648 lower part of the LCE, which we used in this study for the stratigraphic correlation (Figs. 3,
649 8).

650

651 5.3. Clay minerals origin in the Tethyan realm

652

653 The origin of smectitic minerals in oceanic sediments is controversial and has been
654 debated by Chamley (1989) and Thiry (2000). The abundance of these minerals in Cretaceous
655 sediments, especially from cores drilled in the Atlantic Ocean, is interpreted differently by the
656 authors. Smectitic minerals are considered to be detrital in some studies, based on their
657 chemical composition (Al-Fe beidellite), which is similar to smectitic minerals formed in
658 soils, on their REE element profiles, and on their strontium isotope composition (see
659 Chamley, 1989; Chamley et al., 1990). Alternatively, a volcanogenic or early diagenetic
660 origin of smectitic minerals has been proposed based on the common occurrence of the
661 paragenesis smectite opal-CT-clinoptilolite (Pomerol and Aubry, 1977; Christidis, 1995;
662 Madsen and Stemmerik, 2010). The controversy also arises from the coeval occurrence of
663 smectite-poor continental successions and smectite-rich marine sediments. This paradox has
664 been tentatively explained by the massive neof ormation of smectitic minerals in oceanic
665 basins or by the transformation of detrital clay particles into smectitic minerals (Thiry and
666 Jacquin, 1993). Other authors have suggested that differential settling processes are
667 responsible for the smectitic minerals enrichment in (hemi-) pelagic environments (Chamley,
668 1989).

669 In the clay fraction of the Upper Cretaceous Chalk, it is now clear that the smectitic
670 minerals correspond to a mixture of detrital I/S, authigenic lathed smectitic minerals
671 preferentially formed in slowly deposited sediments, and smectitic minerals deriving from the
672 submarine weathering of volcanic glass shards (Deconinck and Chamley, 1995; Jeans, 2006).
673 Environmental conditions during the Late Cretaceous favoured dominantly smectitic
674 sedimentation: high sea level; low topographic relief on continental areas; relatively
675 seasonally contrasted humid climate; low sedimentation rates; and recurrent volcanism
676 expressed by the common occurrence of bentonite layers (Chamley et al., 1990; Deconinck
677 and Chamley, 1995).

678 Chlorite and illite are generally directly reworked from igneous or metamorphic
679 continental basement rocks and are therefore commonly considered to be primary minerals
680 (Chamley, 1989; Weaver, 1989; Ruffell et al., 2002). Consequently, their proportions increase
681 either during tectonic rejuvenation, sea-level fall, or under dry climate when chemical
682 weathering is reduced. By contrast, the pedogenic formation of kaolinite occurs under warm
683 and humid conditions suitable for high rates of chemical weathering (Ruffell et al., 2002).
684 However, this mineral may be also reworked together with primary minerals (e.g. illite,
685 chlorite) from ancient kaolinite-bearing sedimentary rocks (Deconinck and Vanderaveroot,
686 1995). In that case, kaolinite cannot be used as evidence for the existence of humid climate.
687 Coupling or decoupling in the variation of primary minerals and of kaolinite, however, may
688 help differentiate between pedogenic vs. reworked origins, thereby indicating the reliability of
689 kaolinite as a palaeoclimate indicator.

690 An important feature that is highlighted by clay mineral assemblages of the studied
691 sections is the absence of significant amounts of kaolinite in the northernmost sections located
692 in the Danish basin and the North Sea (Fig. 9). By contrast, kaolinite occurs in the three
693 sections located south of 30°N (El Kef – El Djebil and Umbria-Marche localities). The two

694 sections located latitudinally in-between, in the Paris and Mons basins, present intermediate
695 characteristics with kaolinite present in the Cbr-7 borehole, but absent in the Poigny borehole.
696 This broad latitudinal zonation may result from a climatic control, as humid/arid conditions
697 are known to impact clay mineral assemblages (Ruffell et al., 2002; Dera et al., 2009). The
698 common occurrence of bauxite reported in the general area of Gubbio, Furlo, and Tercis-les-
699 Bains (Bárdossy and Dercourt, 1990), supports at least a partial pedogenic origin of kaolinite
700 observed in these sections (Fig. 10).

701 Based on the distribution of kaolinite and bauxite along the palaeolatitudinal transect,
702 climatic zones are proposed in the Tethyan/Boreal realms during the Campanian (Figs, 10,
703 11). Between 20° and 35°N, the clay fractions are characterised by the occurrence of kaolinite
704 and iron (oxyhydro)oxides (e.g. hematite in the Scaglia-Rossa Formation; Channell et al.,
705 1982), corresponding to intense weathering under warm and humid climate conditions. These
706 climatic conditions are also highlighted by the distribution of Late Cretaceous bauxites which
707 are abundant in southern France and in the Dinarides at palaeolatitudes below 35°N (Charvet,
708 1978; Bárdossy and Dercourt, 1990; D'Argenio and Midszenty, 1995; Fig. 10). North of
709 35°N, clay assemblages are mainly dominated by smectitic minerals and illite (with traces of
710 chlorite and kaolinite), and the absence of bauxite indicates limited chemical weathering,
711 consistent with seasonally contrasted climate.

712 However, variations in primary mineral proportions (chlorite and illite) mirror those of
713 kaolinite, pointing to the additional presence of reworked kaolinite. A similar co-variation of
714 primary minerals and kaolinite is identified at El Kef, again suggesting the predominance of
715 kaolinite reworked from old rocks that cropped out on nearby emergent landmasses. To our
716 knowledge, no major bauxite deposits are reported for the Upper Cretaceous in that area
717 (Tardy et al., 1991; Monsels, 2016), which was located in an arid zone according to the
718 distribution of vegetation (Otto-Bliesner and Upchurch, 1997; Chumakov, 2004). These

719 observations suggest that kaolinite is dominantly reworked together with primary minerals in
720 all sections south of 30°N. Consequently, clay sedimentation was not solely controlled by
721 climate but involved an additional factor, which could be linked to the tectonic instability of
722 the southern margin of the Tethys, or to the long-term trend of sea-level fall that characterises
723 the Late Cretaceous.

724

725 5.4. Diachronous fluctuations of detrital clay minerals with palaeolatitude

726

727 A striking feature arising from the overall clay mineralogical results of all studied
728 sections is the earlier occurrence of kaolinite on southern Tethyan margin sites compared to
729 the northern domain (Fig. 9). At El Kef, three intervals of increased detrital minerals (chlorite
730 and kaolinite) are recorded: from the transition between the lower to middle Campanian
731 (within CC19 and the lower part of CC20 nannofossil biozones); from the base of CC21 to the
732 top of the *Radotruncana calcarata* foraminifera zone; and from the base to the top of CC23
733 (Fig. 3). In the Umbria-Marche basin, the onset of detrital input of clay minerals occurs in
734 youngest sediments dated to the upper part of the middle Campanian (Fig. 4). At Tercis-les-
735 Bains, illite and kaolinite contents increase at the base of the pre-LCE, while a coeval increase
736 of illite is recorded in the Paris Basin (Poigny borehole; Chenot et al., 2016, Fig. 9).
737 Consequently, the studied sections point to a general increase in detrital input during the
738 Campanian that is diachronous, with an earlier onset in the southern sections (El Kef, Furlo,
739 Gubbio) than those immediately to the north (Tercis-les-Bains, Poigny).

740 In the northernmost sections (Poigny, Adda-3 and Stevns-2), no clear trend is apparent
741 in the abundance of detrital clay minerals, except for Cbr-7 (Fig. 9). At that site, an increase in
742 detrital input is recorded by an increase in illite and talc proportions in the upper part of the
743 borehole. This contrasts with the other sites where the increase in detrital inputs is marked by

744 increasing kaolinite proportions. This difference may result from a different climatic context
745 that is evidenced by the transition from kaolinite-bearing clay fraction at the base of the
746 borehole toward sediments containing palygorskite, suggesting an evolution from relatively
747 humid to more semi-arid conditions. It is however worth noting that a hardground separates
748 the two intervals in Cbr-7, pointing to the existence of a significant hiatus, the duration of
749 which cannot be evaluated because of the lack of biostratigraphic markers. The large
750 uncertainties on the stratigraphic framework preclude further temporal comparison of the
751 depicted increase in detrital inputs with that recorded at the other studied sites (Fig. 9).

752

753 5.5. A tectonic versus climatic control of clay sedimentation

754

755 Although the long-term sea-level fall recorded during the Late Cretaceous could have
756 contributed to the general increase in detrital inputs depicted here (Haq et al., 1988;
757 Hardenbold et al., 1998; Haq, 2014), their diachronism between sites strongly suggests the
758 existence of additional processes at play.

759 The Late Cretaceous was characterised by compressive events around the Tethys, linked
760 to the northward motion of Africa toward Eurasia (Kley and Voigt, 2008; Frizon de Lamotte
761 et al., 2011; Jolivet et al., 2016). During the Campanian – Maastrichtian interval, large areas
762 of emerged land and newly created relief in central Europe and in the western Tethyan realm
763 (e.g. southern Carpathians, east-Pyreneans, inverted Mid-Polish Anticline, High-Karst)
764 delivered detrital material to the adjacent sedimentary basins (Charvet, 1978; Willingshofer et
765 al., 2001; Kley and Voigt, 2008; Voigt et al., 2008; Melinte-Dobrinescu and Bojar, 2010;
766 Oms et al., 2016; Figs. 1, 10, 12).

767 On the southern Tethyan margin, evidence of tectonic instability as early as the earliest
768 Campanian, is suggested by the occurrence of synsedimentary faults, slumps and slope

769 instability features in sedimentary successions from northeastern Tunisia (Boutib et al., 2000;
770 Bey et al., 2012). This tectonically driven sedimentation persisted throughout the Campanian
771 and possibly into the Maastrichtian, as illustrated by syndepositional faulting and gravity flow
772 deposits in the Abiod Chalk (Bouaziz et al., 2002; Dlala 2002; Negra, 2016). In the
773 southernmost section of El Kef, the detrital influence occurring in the middle Campanian
774 might be linked to a tectonic rejuvenation of nearby continental areas (e.g. Kasserine Island,
775 Kadri et al., 2015; Figs. 1, 10, 12).

776 At El Kef, several kaolinite- and chlorite-enriched detrital intervals are recognised,
777 separated by intervals devoid of typical detrital minerals. This may reflect distinct tectonic
778 pulses affecting this segment of the southern margin (Figs. 3, 12). The kaolinite-rich interval
779 within the *G. ventricosa* zone at El Kef is coeval with the occurrence of rudist-bearing
780 olistolith beds on the NE margin of Kasserine Island (Negra et al., 2016), further highlighting
781 a phase of platform destabilisation.

782 In central Italy (Umbria-Marche basin), detrital input of kaolinite and chlorite started
783 later, in the earliest part of chron C33n (Figs. 4, 12). Volcanic activity in the area is
784 highlighted by the preservation of a bentonite layer in the *G. elevata* zone (more precisely in
785 the lower Campanian CC18 nannofossil biozone; Mattias et al., 1988; Fig. 4). The ashfall
786 originated from an active volcanic centre related to a subduction zone located to the east of
787 the Umbria-Marche basin, in the Dinarides (Charvet, 1978), and points to active tectonism in
788 this region (Bernoulli et al., 2004; Schmid et al., 2008). In the Furlo – Upper Road section, the
789 occurrence of the bentonite layer is followed by slope destabilisation features including
790 decimetric turbidite beds and a 12 m-thick slump, suggesting long-lasting tectonic instability
791 (Fig. 4).

792 Within the same tectonic and climatic context, additional differences in clay mineralogy
793 between geographically close sections are evidenced in the Umbria-Marche basin. Indeed, a

794 comparison between the Gubbio – la Bottaccione and Furlo – Upper Road sections shows that
795 kaolinite is more abundant at Furlo, which was located on a slope at shallower depth than
796 Gubbio (Fig. 4). This difference is emphasised by the occurrence of numerous slope deposits
797 including slumps and turbidites at Furlo, while basinal deposits characterised the Gubbio
798 section. This clay mineral distribution is attributed to differential settling of kaolinite which
799 likely originated from the east, possibly from the High Karst (Gibbs, 1977; Charvet, 1978;
800 Figs. 1, 12).

801 A local tectonic influence coinciding with the LCE is also recorded in the Tercis-les-
802 Bains section where chlorite, illite and kaolinite increase (Figs. 9–12). This detrital event is
803 likely due to a tectonic pulse linked to the Pyrenean compressional phase between Iberia and
804 southern Europe (Laurent et al., 2001; Vergés et al., 2002; Oms et al., 2016). A coeval event
805 characterised by increasing proportions of illite is recorded in the Paris Basin (Poigny
806 borehole, Deconinck et al., 2005; Figs. 9–12), potentially related to a compressive event
807 (Mortimore and Pomerol, 1997). Slump deposits recorded in the Anglo-Paris Basin in the
808 middle Campanian (Gale et al., 2015) could also relate to local tectonic instability.

809 By contrast, in the northern part of the studied transect, there are no clear detrital events
810 similar to those recorded in the southern sections, despite the presence of active tectonism in
811 this region affected by inversions at the time (Kley and Voigt, 2008; Voigt et al., 2008),
812 probably because this area was too far away from emerged areas (Figs. 9–12). Thus, the
813 climatic control on clay sedimentation is more clearly expressed at these sites. In the Mons
814 basin, during the late Campanian, the progressive decreasing proportion of kaolinite was
815 followed by the occurrence of palygorskite, which reflects the establishment of increasingly
816 semi-arid climatic conditions (Figs. 3, 9–12). However, traces of talc present in the late
817 Campanian could be related either to a change of detrital sources or to the generation of newly
818 exposed areas by the extensive coeval tectonism recorded in the basin (Vandycke and

819 [Bergerat, 1989](#)), which is consistent with the coevally increase of illite in this basin. In the
820 Danish North Sea, similar semi-arid to semi-humid climatic conditions are consistent with the
821 abundance of IS R0, which constitutes the entire clay fraction of uppermost Campanian
822 sediments (Figs. 2, 9–12).

823

824 5.6. Carbon cycle and continental weathering

825

826 Carbon cycle changes and continental weathering had been tentatively linked through
827 the observed correspondence between the carbon isotope excursion defining the LCE and
828 enhanced terrigenous inputs, identified by a coeval increase in kaolinite, chlorite and illite
829 proportions at Tercis-les-Bains and Poigny ([Chenot et al., 2016](#)). The new results presented
830 here, which include data from sections located over a wider range of palaeolatitude, show that
831 this relationship does not hold. Indeed, the diachronous nature of detrital supplies evidenced
832 here, interpreted as reflecting the northward progression of tectonic deformation, results in a
833 decoupling between the carbon-isotope excursions and the evolving clay minerals
834 assemblages.

835 However, our new clay mineralogical data highlight enhanced continental weathering
836 throughout the whole Campanian stage, with a diachronous onset from the south to the north
837 (Figs. 10, 12). As silicate weathering is known to promote atmospheric CO₂ drawdown
838 ([Berner, 1990, 2004](#); [Berner and Kothavala, 2001](#)), our new data hint to a potentially major
839 role of incipient orogenic processes on Late Cretaceous long-term cooling. The data compiled
840 by [Royer et al. \(2012\)](#) and [Franks et al. \(2015\)](#) highlight lower pCO₂ levels in the Campanian
841 and Maastrichtian than during the Albian to Turonian interval, but the temporal resolution is
842 not sufficient to discuss correlations between pCO₂ fluctuations, tectonic phases and the
843 evolution of clay minerals assemblages depicted in our study. Decreasing atmospheric CO₂

844 levels during the Late Cretaceous have been repeatedly associated to reduced outgassing CO₂
845 flux from mid-ocean ridge volcanism and arc magmatism (Berner et al., 1983; Jones et al.,
846 1994; Godderis and François, 1995; Berner, 2004; McKenzie et al., 2016). Without excluding
847 the role of reduced CO₂ volcanic outgassing, our new results highlight an additional important
848 mechanism that may have contributed to the Late Cretaceous cooling. Although initiation of
849 the Tethyan closure began during the mid-Cretaceous, the Santonian–Campanian is marked
850 by a change of direction and faster motion of African toward Eurasia (Bosworth et al., 1999;
851 Guiraud and Bosworth, 1999; Frizon de Lamotte et al., 2011; Jolivet et al., 2016). The
852 Campanian is also characterised by an acceleration of the long-term climatic cooling recorded
853 during the Late Cretaceous (Cramer et al., 2009; Friedrich et al., 2012; Linnert et al., 2014),
854 which coincides with enhanced detrital inputs depicted by clay minerals in our study. This
855 temporal coincidence between an acceleration of cooling, tectonic pulses, and clay detrital
856 input evolution further argues for a significant impact on climate of plate tectonics linked to
857 Africa-Eurasia convergence. Although the importance of this process remains to be tested
858 quantitatively using geochemical models, our work opens new perspectives on the
859 understanding of the Late Cretaceous climate cooling.

860

861 6. Conclusion

862

863 New results on clay minerals assemblages of Campanian sediments from six sections
864 and boreholes ranging from the southern Tethyan margin to the Boreal realm, provide the first
865 insights on the evolution of continental weathering at the Tethyan scale during a long
866 underexplored time interval (11 Myr).

867 We propose a climatic zonation in the west Tethyan to Boreal realms during the late
868 Campanian: (1) a semi-humid climate belt north of 35°N, based on the sporadic occurrence of

869 kaolinite, the occurrence of palygorskite and the high percentage of smectitic minerals; (2) a
870 warm and humid climate belt, based on the occurrence of kaolinite and bauxite between 35°–
871 20°N; and maybe (3) a semi-arid zone south of 20°N, adjacent to the Saharan platform, based
872 on the abundance of smectitic minerals, with kaolinite interpreted here as being reworked
873 from the basement.

874 Superimposed on this latitudinal climate distribution, we have identified detrital events
875 in several basins. These events resulted from weathering of emerged continental areas, that we
876 relate to the main tectonic active zones, and more specifically to the large subduction zone of
877 the central Tethyan realm between African and Eurasian plate (Umbria-Marche basin, Furlo
878 and Gubbio sections), the extensional basins in southern Tethys (Saharan platform margin, El
879 Kef section), and the collision between the Iberian and the Eurasian plates (Aquitaine basin,
880 Tercis-les-Bains section). These tectonic instabilities, associated with a warm and humid
881 climate, likely led to enhanced chemical weathering of the newly created continental relief.

882 The northward migration of an enhanced detrital flux evidenced by our new records,
883 linked to the progression of compressional deformation, reflects the closure of the Tethys due
884 to the northward motion of Africa. As chemical weathering of silicate induces CO₂
885 consumption, we suggest that Late Cretaceous cooling was partly linked to enhance
886 continental weathering.

887

888 7. Acknowledgments

889

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892 calibration, and Anne-Charlotte Guillet for her contribution in the lab.

893

894

FIGURES

895 **Figure 1:** Palaeogeography of the Campanian–Maastrichtian of the west Tethyan and south
896 Boreal realms (modified from Philip and Floquet, 2000). Location of the sites studied: [A]
897 Adda-3 borehole; [C] Cbr-7 borehole; [F] Furlo – Upper Road section; [G] Gubbio – la
898 Bottaccione section; [K] El Kef – El Djebil section; [P] Poigny borehole; [S] Stevns-2
899 borehole; [T] Tercis-les-Bains section. The green squares represent previously published data
900 (Deconinck et al., 2005; Chenot et al., 2016) used in this study.

901 **Figure 2:** Stratigraphy and clay mineralogy of the Campanian–Maastrichtian in northern
902 boreholes. (A) Clay mineralogical data (this study) of the Adda-3 borehole compared to the
903 gamma-ray and carbon- and oxygen-isotopic data (Perdiou et al., 2016). (B) Clay
904 mineralogical data (this study) from Stevns-2 borehole compared to gamma-ray and carbon-
905 and oxygen- isotopic data (Boussaha et al., 2016).

906 **Figure 3:** Stratigraphy and clay mineralogy of the Campanian in the Mons basin, Belgium
907 and Campanian–Maastrichtian at El Kef, Tunisia. (A) Clay mineralogical data from the Cbr-7
908 borehole and carbonate content (this study) compared to carbon- and oxygen-isotope data and
909 manganese contents (ppm) (Richard et al., 2005). (B) Clay mineralogical data from the El Kef
910 – El Djebil section (this study), compared to carbon- and oxygen- isotopic data (Jarvis et al.,
911 2002), carbonate content and Si/Al ratio (Mabrouk El Asmi, 2014). The foraminiferal
912 biostratigraphic data of El Kef – El Djebil section are established from the lithostratigraphic
913 comparison with Kalaat Senan section (see Jarvis et al., 2002), while the calcareous
914 nannofossil biostratigraphic data have been performed on the El Kef – El Djebil samples.

915 *G. aegyptica* = *Globotruncana aegyptica*; *G. elevata* = *Globotruncanita elevata*; *G. f.* =
916 *Globotruncana falsostuarti*; *G. h.* = *Globotruncanella havanensis*; *G. g.* = *Gansserina*
917 *gansseri*; *G. ventricosa* = *Globotruncana ventricosa*; *R. calcarata* = *Radotruncana calcarata*

918 *A. c. var NT* = *Arkhangelskiella cymbiformis*; *A. c. var W* = *Arkhangelskiella cymbiformis var*
919 *W*; *B. p. c.* = *Broinsonia parca constricta*; *C. spine* = curved spine nannolith; *E. e.* =
920 *Eiffellithus eximius*; *M. cf. M. p.* = *Micula cf M. premolisilva*; *R. a.* = *Reinhardites*
921 *anthrophorus*; *T. o.* = *Tranolithus orionatus*; *U. g.* = *Uniplanarius gothicus*; *U. s.* =
922 *Uniplanarius sissinghii*; *U. t.* = *Uniplanarius trifidius*.

923 **Figure 4:** Stratigraphy and clay mineralogy of the Campanian in the Umbria-Marche basin,
924 Italy. Clay mineralogical data (this study) compared to carbon- and oxygen- data (this study)
925 from (A) Gubbio – la Bottaccione and (B) Furlo – Upper Road sections. Magnetostratigraphy
926 of Gubbio – la Bottaccione from [Lowrie and Alvarez \(1977\)](#); Furlo – Upper Road from
927 [Alvarez and Lowrie \(1984\)](#). Gubbio biostratigraphic data from [Coccioni and Premoli Silva](#)
928 [\(2015\)](#).

929 *A. m.* = *Archaeoglobigerina minimus*; *D. a.* = *Dicarinella asymetrica*; *E. e.* = *Eiffellithus*
930 *eximius*; *G. a.* = *Globotruncana aegyptica*; *G. e.* = *Globotruncanita elevata*; *G. h.* =
931 *Globotruncanella havanensis*; *G. gansseri* = *Gansserina gansseri*; *G. st.* = *Globotruncanita*
932 *stuarti*; *G. v.* = *Globotruncana ventricosa*; *R. l.* = *Reinhardites levis*; *R. m.* = *Rucinolithus*
933 *magnus*; *U. g.* = *Uniplanarius gothicus*; *U. t.* = *Uniplanarius trifidus*.

934 **Figure 5:** Occurrence of selected calcareous nannofossils taxa in El Kef – El Djebil section.

935 **Figure 6:** Cross-plot of carbon- and oxygen-isotope bulk-rock data of the Gubbio – la
936 Bottaccione and Furlo – Upper Road sections.

937 **Figure 7:** Cross-plot of carbon- and oxygen-isotope bulk-rock data of the Gubbio – la
938 Bottaccione (▲) and Furlo – Upper Road (▲) sections compared with isotopic data from
939 several sites in the Tethyan Realm (A) to the north of 35 °N and (B) to the south of 35 °N.

940 **Figure 8:** Correlation of the $\delta^{13}\text{C}$ profiles across the Campanian between Adda-3 ([Perdiou et](#)
941 [al., 2015](#)), Stevns-2 ([Boussaha et al., 2016](#)), Cbr-7 ([Richard et al., 2005](#)), Poigny borehole
942 ([Chenot et al., 2016](#)), Tercis-les-Bains section ([Voigt et al., 2012](#)), Gubbio – la Bottaccione

943 section (this study), Furlo – Upper Road section (this study) and El Kef – El Djebil section
944 ([Jarvis et al., 2002](#)).

945 **Figure 9:** Comparison of clay mineralogical assemblages of Campanian sediments of the sites
946 studied (smectitic minerals are not represented). Correlation based on carbon-isotope events,
947 along a palaeolatitudinal transect from ~20°N to ~40°N. Adda-3 borehole, clay mineralogical
948 data (this study) compared to the carbon- isotopic data ([Perdiou et al., 2016](#)); Stevns-2
949 borehole, clay mineralogical data (this study) compared to carbon- isotopic data ([Boussaha et](#)
950 [al., 2016](#)); Cbr-7 borehole, clay mineralogical data (this study) compared to carbon- isotopic
951 data ([Richard et al., 2005](#)); Poigny borehole, clay mineralogical data compared to carbon-
952 isotopic data ([Chenot et al., 2016](#)); Tercis-les-Bains section, clay mineralogical data ([Chenot](#)
953 [et al., 2016](#)) compared to carbon–isotopic data ([Voigt et al., 2012](#)); Gubbio – la Bottaccione,
954 clay mineralogical data compared to carbon- isotopic data (this study); Furlo – upper Road
955 section, clay mineralogical data compared to carbon- isotopic data (this study); El Kef – El
956 Djebil section, clay mineralogical data (this study) compared to carbon – isotopic data ([Jarvis](#)
957 [et al., 2002](#)).

958 **Figure 10:** Geodynamic framework of Campanian Tethyan bauxites and the clay minerals of
959 the studied sections (modified from [Bárdossy and Dercourt, 1990](#); [Jolivet et al. 2016](#)).
960 Location of the sites studied: [A] Adda-3 borehole; [C] Cbr-7 borehole; [F] Furlo – Upper
961 Road section; [G] Gubbio – la Bottaccione section; [K] El Kef – El Djebil section; [P] Poigny
962 borehole; [S] Stevns-2 borehole; [T] Tercis-les-Bains section.

963 Name of the Campanian bauxites localities: (1) Alpilles (France), (2) Haut-Var (France), (3)
964 Markusovce (Slovakia), (4) La Boissière (France), (5) Nurra (Sardinia), (6) Villeveyrac basin
965 (France), (7) Bédarieux (France), (8) Tyrol Brandenburg and Salzburg (Austria), (9)
966 Unterlaussa (Austria), (10) Sümeg (Hungary), (11) Halimba (Hungary), (12) Ihakut-
967 Némethbanya (Hungary), (13) Grméc Hill (Bosnia Herzegovina), (14) Jajce (Bosnia

968 Herzegovina), (15) Grebnik (Kosovo), (16) Küçük Koras, Sebimlkoy (Turkey), (17) Payas,
969 Islaye (Turkey), (18) Sohodol, Cimpeni (Romania), (19) Euboea Island (Greece).

970 **Figure 11:** (A) Palaeolatitudinal topographic profile for the studied sites and their positions
971 relative to the main orogenic belt during the studied time interval (see also Fig. 10) compared
972 with (B) a reconstruction of the latitudinal climatic zonation with the locations of major
973 bauxites. These provide a potential source of hydrolytic minerals into the adjacent
974 sedimentary basin during the Santonian–Maastrichtian interval. (C) Palaeolatitudinal transect,
975 from the base of the Late Campanian Event to the top of the end of late Campanian, of the
976 mean clay minerals assemblages from sites studied in the Tethyan and Boreal realms.
977 Location of the sites studied: [A] Adda-3 borehole; [C] Cbr-7 borehole; [F] Furlo – Upper
978 Road section; [G] Gubbio – la Bottaccione section; [K] El Kef – El Djebil section; [P] Poigny
979 borehole; [S] Stevns-2 borehole; [T] Tercis-les-Bains section. There is no direct
980 correspondence between A and B frames; the formation of bauxite and kaolinite occurred in
981 humid tropical conditions. The synthetic “orogenic belt” (defined in A) represents both the
982 Pyrenees and Alpine units that have their own geodynamic histories.

983 **Figure 12:** Scenario of the African block rotation modification, placed on a palaeogeographic
984 map of the west Tethyan – south Boreal realm, from the end of the Santonian to the late
985 Campanian. Histograms illustrate the proportion of clay mineral species (excluding the
986 smectitic minerals background sedimentation) at the sites studied, during the (A) end
987 Santonian, (B) early Campanian to early mid-Campanian, (C) later mid-Campanian, (D) Late
988 Campanian Event, and (E) late Campanian.

989

990

TABLES

991 **Table 1:** Peak positions used for the recognition of clay minerals ($>2 \mu\text{m}$) of the insoluble
992 residue.

993 **Table 2:** Synthesis of carbon-isotope events recognised in the Campanian.

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SUPPLEMENTARY DATA

Supplementary Data A: Carbon and oxygen isotopes of the bulk rock, lithology and clay mineralogy of the Furlo – Upper Road section.

Supplementary Data B: Carbon and oxygen isotopes of the bulk rock, calcium carbonate content, lithology and clay mineralogy of the Gubbio – la Bottaccione.

Supplementary Data C: Carbon and oxygen isotopes of the bulk rock, calcium carbonate content, lithology and clay mineralogy of the Cbr-7 borehole.

Supplementary Data D: Carbon and oxygen isotopes of the bulk rock, lithology and clay mineralogy of the Stevns-2 borehole.

Supplementary Data E: Carbon and oxygen isotopes of the bulk rock, lithology and clay mineralogy of the Adda-3 borehole.

Supplementary Data F: Carbon and oxygen isotopes of the bulk rock, calcium carbonate content, lithology and clay mineralogy of the El Kef – El Djebil section.

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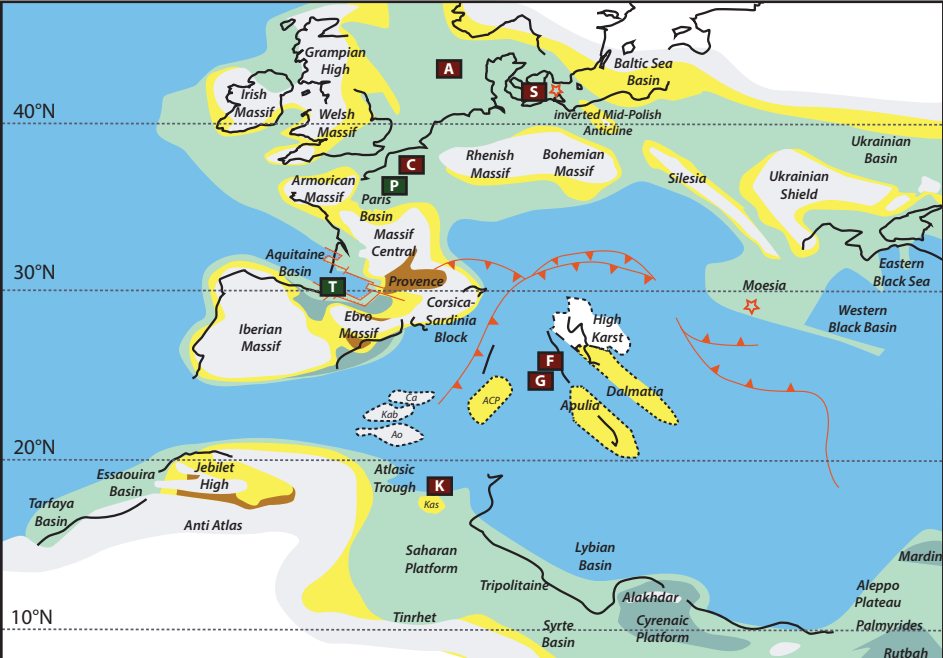
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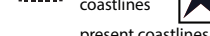
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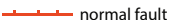
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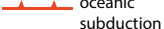
supposed coastlines



present coastlines of continents



normal fault



oceanic subduction



volcano



emerged land

fluviatile/lacustrine sediments

eoian sediments

coastal marine, shallow marine (terrigenous)

shallow marine carbonate deposits (platform, build-up and ramps)

deep(er) carbonate, (hemi)pelagic oozes

deep marine

new data sets

previously published data sets

ACP = Apennine Carbonate Platform

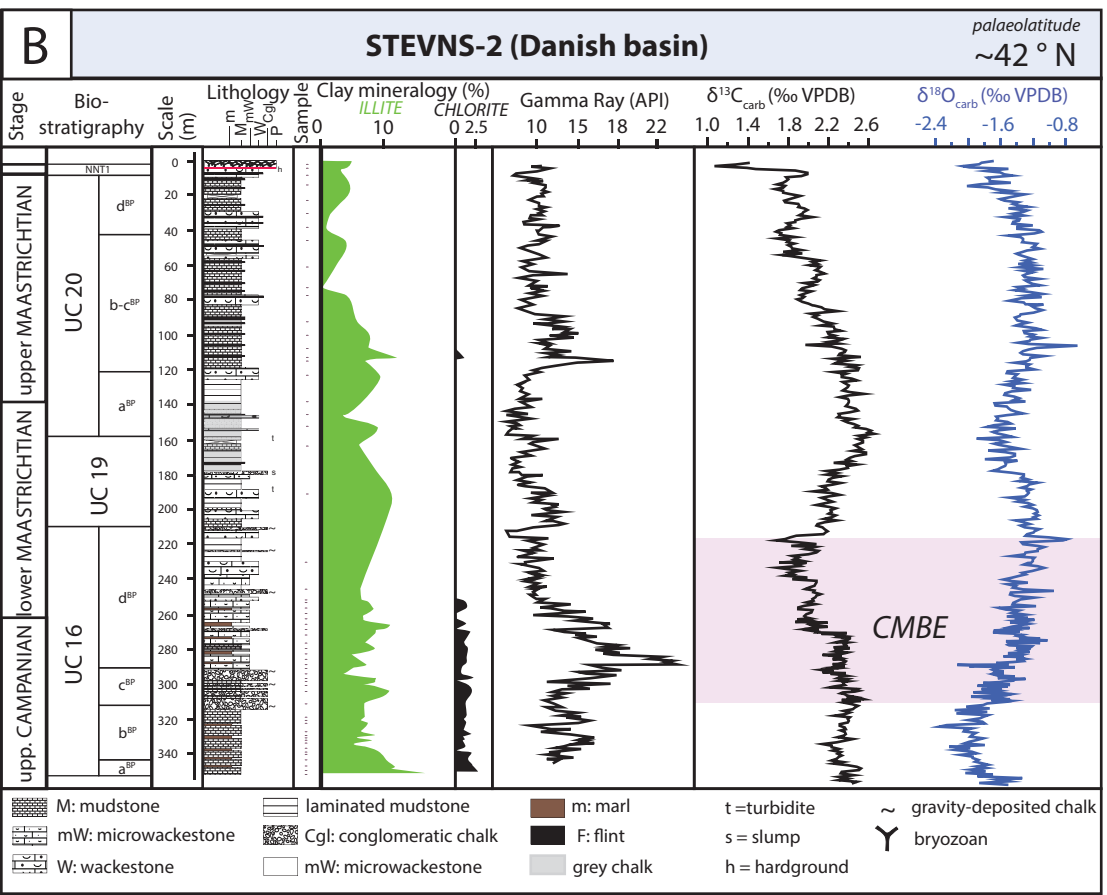
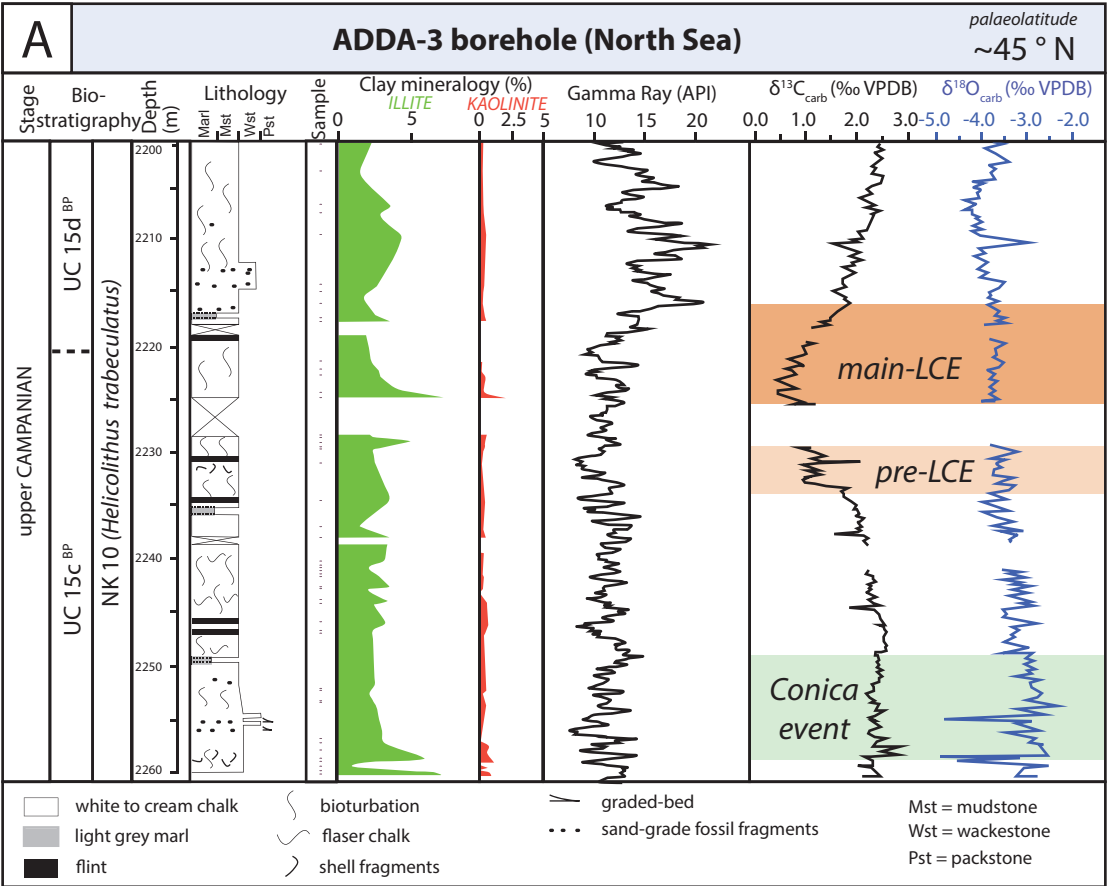
Ao = Alboran

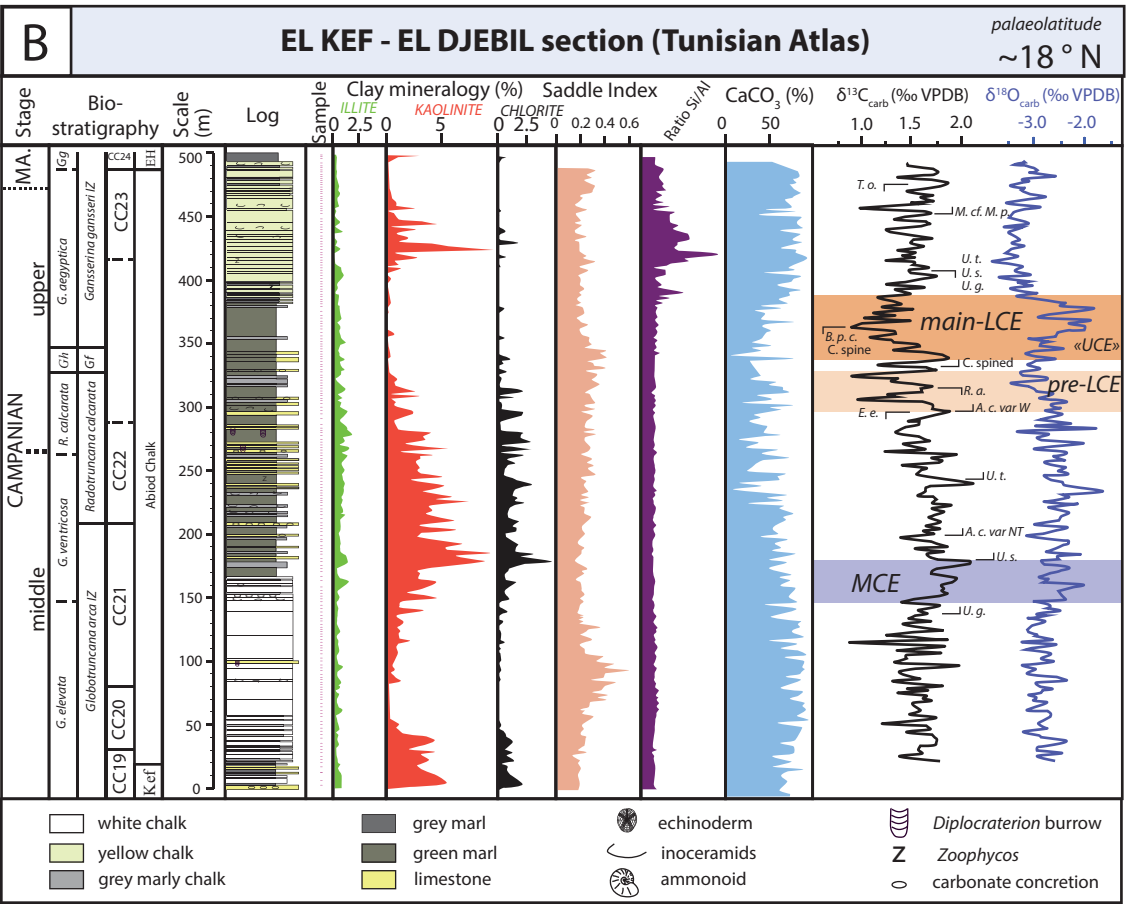
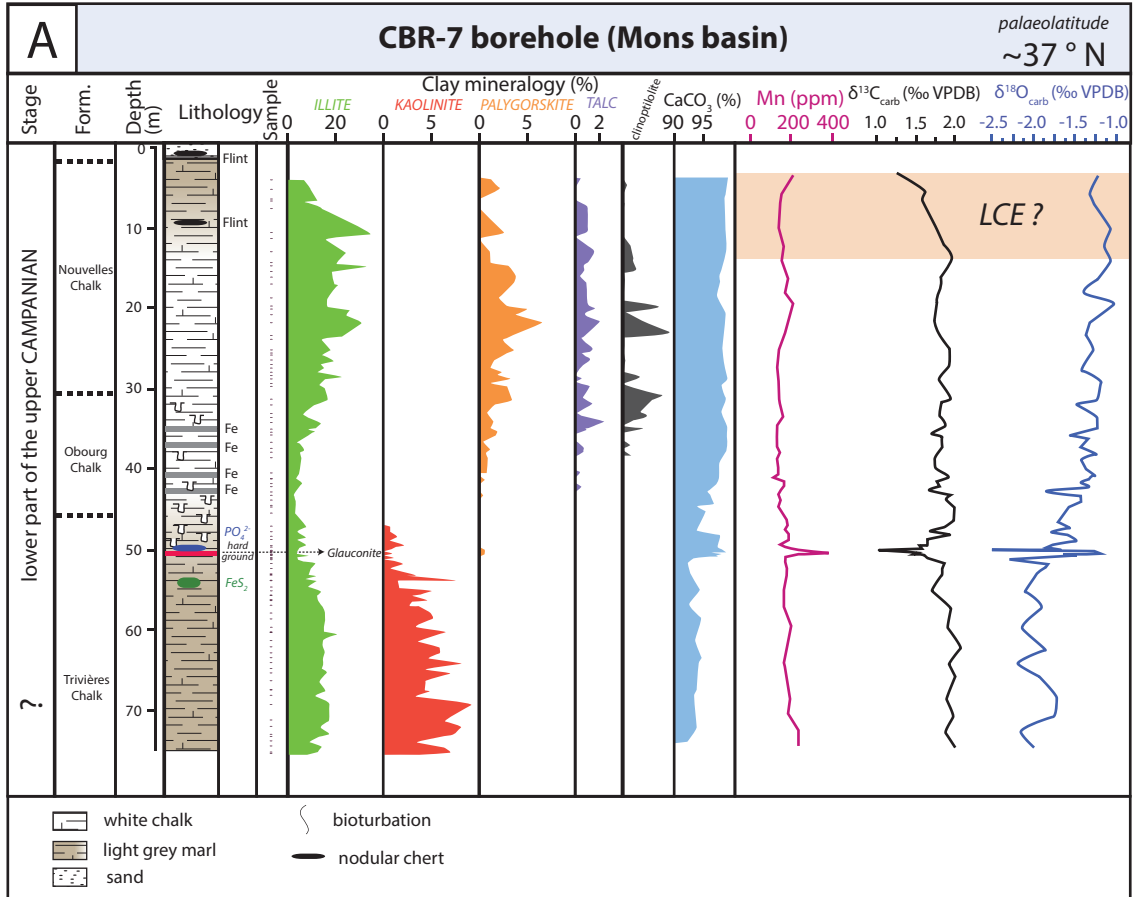
Ca = Calabria

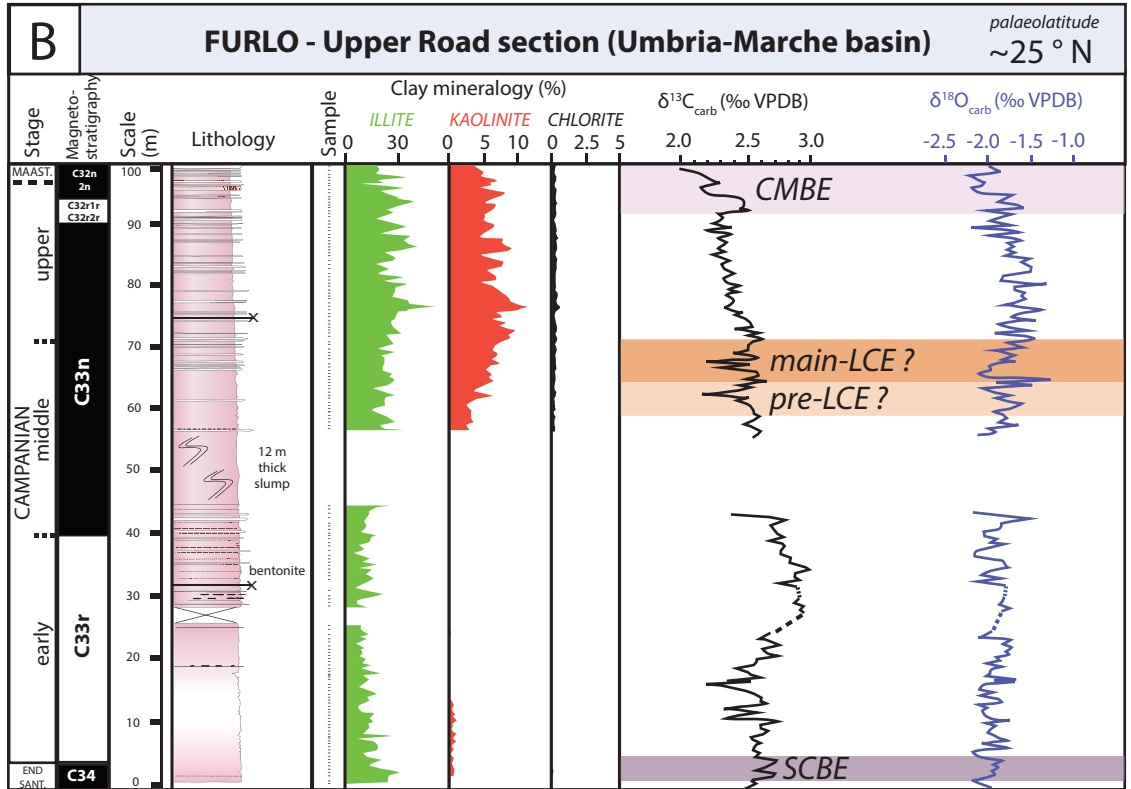
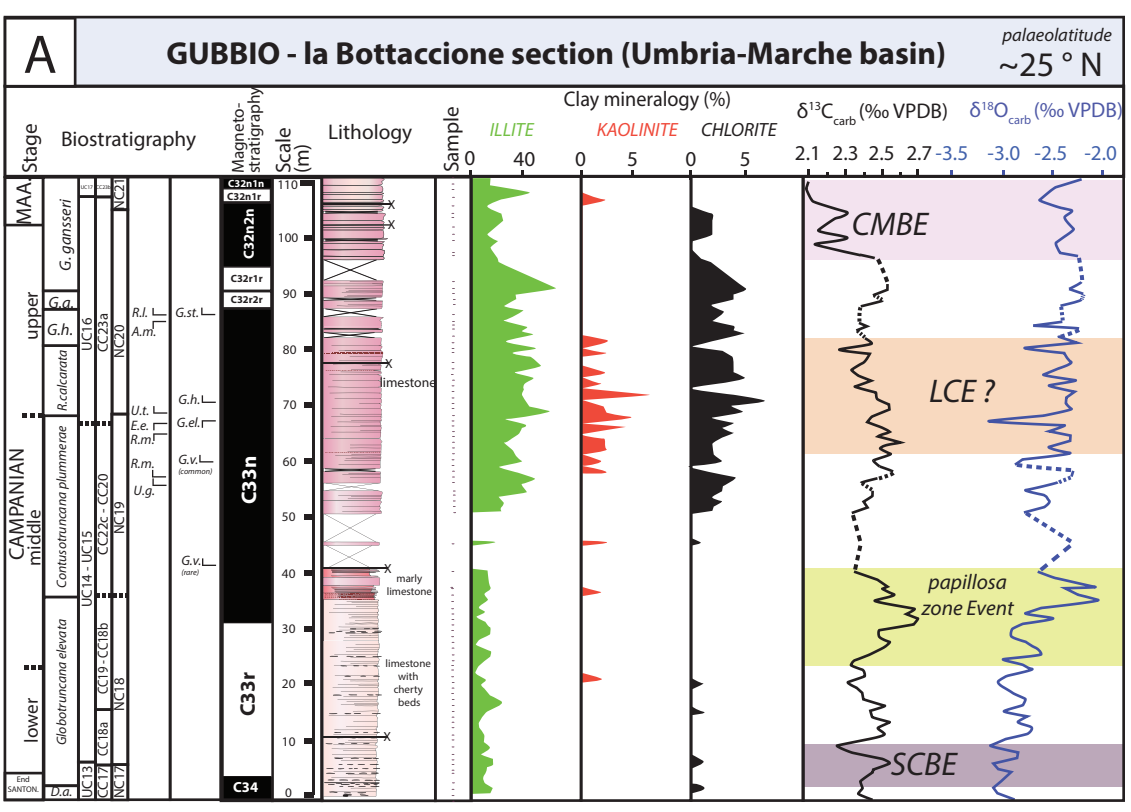
Kab = Kabyliya

Kas = Kasserine Island

0 250 500 km



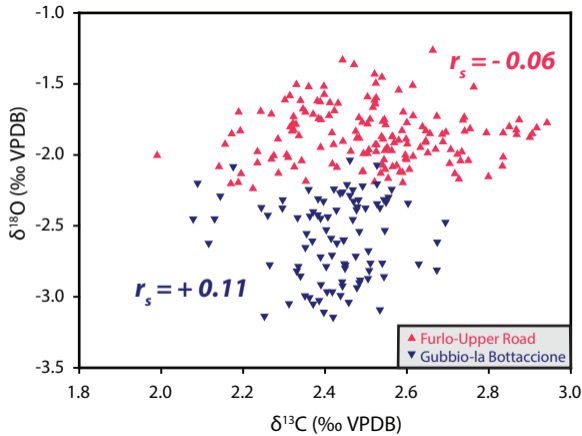




- light pink limestone
- pink limestone chalk
- marlstone

- marly layers
- nodular chert
- fault

- slump



$\delta^{13}\text{C}$ (‰ VPDB)

