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# Carbon cycle history through the Middle Jurassic (AalenianBathonian) of the Mecsek Mountains, Southern Hungary

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1 Carbon cycle history through the Middle Jurassic (Aalenian– Bathonian) of the Mecsek  
2 Mountains, Southern Hungary

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14  
15 **Abstract.** A carbonate carbon isotope curve from the Aalenian–Bathonian interval is presented  
16 from the Óbánya valley, of the Mecsek Mountains, Hungary. This interval is certainly less well  
17 constrained and studied than other Jurassic time slices. The Óbánya valley lies in the eastern part of  
18 the Mecsek Mountains, between Óbánya and Kisújbánya and provides exposures of an Aalenian to  
19 Lower Cretaceous sequence. It is not strongly affected by tectonics, as compared to other sections  
20 of eastern Mecsek of the same age. In parts, a rich fossil assemblage has been collected, with  
21 Bathonian ammonites being especially valuable at this locality. The pelagic Middle Jurassic is  
22 represented by the Komló Calcareous Marl Formation and thin-bedded limestones of the Óbánya  
23 Limestone Formation. These are overlain by Upper Jurassic siliceous limestones and radiolarites of  
24 the Fonyászó Limestone Formation. Our new data indicate a series of carbon isotope anomalies  
25 within the late Aalenian and early-middle Bajocian. In particular, analysis of the Komló Calcareous  
26 Marl Formation reveals a negative carbon isotope excursion followed by positive values that occurs  
27 near the base of the section (across the Aalenian–Bajocian boundary). The origin of this carbon-  
28 isotope anomaly is interpreted to lie in significant changes to carbon fluxes potentially stemming  
29 from reduced run off, lowering the fertility of surface waters which in turn leads to lessened  
30 primary production and a negative  $\delta^{13}\text{C}$  shift. These data are comparable with carbonate carbon  
31 isotope records from other Tethyan margin sediments. Our integrated biostratigraphy and carbon  
32 isotope stratigraphy enable us to improve stratigraphic correlation and age determination of the

33 examined strata. Therefore, this study of the Komló Calcareous Marl Formation confirms that the  
34 existing carbon isotope curves serve as a global standard for Aalenian–Bathonian  $\delta^{13}\text{C}$  variation.

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36 **Key words.** Carbon, isotope stratigraphy, Aalenian, Bajocian, Óbánya, Mecsek, Hungary.

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## Introduction

39 The  $\delta^{13}\text{C}$  curve of the Aalenian–Kimmeridgian interval shows a series of major Jurassic  
40 isotope events within the Aalenian, early-middle Bajocian, Callovian and middle Oxfordian  
41 (Hoffman et al. 1991; Bill et al. 1995; Weissert & Mohr 1996; Jenkyns 1996; Bartolini et al. 1999;  
42 Rey & Delgado 2002; O'Dogherty et al. 2006; Sandoval et al. 2008; Nunn et al. 2009; Price et al.  
43 2016) recorded in both southern and northern Tethyan margin sediments. The potential of these  
44  $\delta^{13}\text{C}$  records for regional and global correlation of ancient marine sediments is evident. Some of  
45 these excursions (e.g. during the Oxfordian) are intrinsically coupled with climatic changes and have  
46 been extensively studied in many parts of the world (e.g. Bill et al. 1995; Jenkyns 1996). With  
47 respect to the Middle Jurassic interval (e.g. the Aalenian–Bathonian) it is the carbon isotope curves  
48 of Bartolini et al. (1999) and Sandoval et al. (2008) that often serve as a global standard (e.g. Ogg &  
49 Hinnov 2012). Major carbon-cycle perturbations in the Middle Jurassic are also recognised in  
50 terrestrial organic matter (fossil wood) (Hesselbo et al. 2003). Although the excursions of the  
51 Middle Jurassic have received only modest attention, they occur on more than one continent and  
52 may thus serve for global correlation of strata (e.g. Wetzel et al. 2013; Hönig & John 2015; Dzyuba  
53 et al. 2017). The goal of this study is to examine Aalenian – Bathonian carbon isotope stratigraphy  
54 from Hungary for comparison. A further aim of this study is to examine linkages between the  $\delta^{13}\text{C}$   
55 record of past global biotic and climatic change.

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## Geological Setting

58 The Lower Jurassic of the Mecsek Mountains of Hungary (Fig. 1) is characterized by coal  
59 bearing continental and shallow marine siliciclastic sediments (Haas et al. 1999). From Late  
60 Sinemurian times onwards, deposition consisted of deeper marine hemipelagic facies with mixed  
61 siliciclastic–carbonate lithologies (the Hosszúhetény and Komló Calcareous Marl formations,  
62 Raucsik & Merényi 2000). The site of this hemipelagic marly and calcareous marly sedimentation  
63 was most probably on or distally beyond the northern outer shelf of the Tethys Ocean (Fig. 2),  
64 whilst shallower conditions occurred towards the western margins (Enay et al. 1993). Although the  
65 precise age of the Komló Calcareous Marl Formation is uncertain, an Aalenian to Bajocian age is  
66 indicated (Forgó et al. 1966, Fig. 3). Overlying the Komló Calcareous Marl Formation, in the Mecsek,  
67 is the pelagic Bathonian–Callovian Óbánya Limestone Formation consisting of thin-bedded  
68 limestones and marls (Galács 1994). The Upper Jurassic is represented by a siliceous limestone and

69 radiolarite (the Fonyászó Limestone) as well as thin-bedded limestone (the Kisújbánya and Márévár  
70 limestones).

71 The Óbánya valley (Fig. 1) lies in the eastern part of the Mecsek Mountains, between  
72 Óbánya and Kisújbánya and provides exposures of the Komló Calcareous Marl Formation. The  
73 succession is not strongly affected by tectonics, as compared to other sections of eastern Mecsek of  
74 the same age (Velledits et al. 1986). The exposed Aalenian and Bajocian sediments (the Komló  
75 Calcareous Marl Formation) can be seen as alternating limestone beds (0.2–0.5m) and laminated  
76 beds consisting of dark grey, spotted, bituminous, micaceous marls (Fig. 4). The laminated beds  
77 become harder upwards with increasing carbonate content (from 36 to 55%, Velledits et al. 1986).  
78 Aside from some bivalve and plant imprints within the lower part of the succession, an ammonite  
79 (*Ludwigia* sp.) has been found indicating an Aalenian age (Velledits et al. 1986). The total thickness  
80 of the Aalenian has been estimated by Velledits et al. (1986) to be ~75 m although only the top  
81 ~25m was exposed. Fossils from the middle part of the succession include the ammonites  
82 *Dorsetensia* (at ~105m) and *Stephanoceras* (indicative of the Humphriesianum Zone), bivalve  
83 moulds together with carbonized plant fragments (Velledits et al. 1986).. Age diagnostic  
84 ammonites (e.g. *Leptosphinctes*, *Adabofoloceras* of the Niortense Zone) are recorded within the  
85 upper part of the Komló Calcareous Marl (Velledits et al. 1986). The total thickness of sediments of  
86 Bajocian age is ~170 m. The Komló Calcareous Marl is overlain by a red calcareous marl and nodular  
87 limestone (Fig 4) rich in age diagnostic ammonites (e.g. *Parkinsonia*, *Morphoceras* and *Procerites*)  
88 and pelagic microfossils (Galácz 1994). This 20m thick formation (the Óbánya Limestone Formation)  
89 is of Bathonian age (Galácz 1994) and was deposited in a pelagic environment. During this time  
90 major flooding events also occur elsewhere in northern Europe, with a peak transgression at the  
91 Bajocian-Bathonian boundary (Hallam 2001).

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### Materials and methods

94 For this study, 273 bulk carbonate samples were derived from the outcrop of the Óbánya  
95 valley. Samples were taken from both marl and limestone lithologies (Fig. 5). The average spacing of  
96 samples was ~0.3m. Subsamples (250 to 400 micrograms) avoiding macrofossils and sparry calcite  
97 veins, were then analysed for stable isotopes using a GV Instruments Isoprime Mass Spectrometer  
98 with a Gilson Multiflow carbonate auto-sampler at Plymouth University. Isotopic results were  
99 calibrated against the NBS-19 international standard. Reproducibility for both  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  was  
100 better than  $\pm 0.1\%$ , based upon duplicate sample analyses.

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## Results

The isotope results are presented in Figures 6 and 7. As isotopic analyses were undertaken from both marl and limestone lithologies a comparison of the isotopic composition of the two lithologies can be made. For the limestone (n = 181) the mean  $\delta^{13}\text{C}$  value is 1.0‰ and -3.5‰ for  $\delta^{18}\text{O}$ . For the marl (n = 92) the mean  $\delta^{13}\text{C}$  value is less positive, 0.4‰ and -4.3‰ for  $\delta^{18}\text{O}$ . The greater number of limestone vs. marl samples analysed reflects the generally better exposure of the limestones and poorer quality of the marl outcrops. It is for this reason that the carbon isotope curve (Fig. 7) is plotted though the limestone data only. Using Student's T-Test the isotopic difference between limestones and marls is also significant (at  $p < .05$ ). These data are also consistent with stable isotope data from Raucsik (1997) who also isotopically analysed both limestone and marlstone carbonate samples from the Komló Calcareous Marl Formation (Fig. 6).

With respect to the carbon isotope stratigraphy (derived from the limestone data) a number of features of the curve are of particular note. Firstly, there is a negative excursion followed by positive values occurring near the base of the section (across the Aalenian–Bajocian boundary). The carbon-isotope values then become more positive, reaching the most positive values seen (at about 100 m height in Fig. 7). Although showing a good deal of scatter, values remain fairly positive, until towards the top of the Komló Calcareous Marl Formation where there is a drop in  $^{13}\text{C}$  values (at 166m). Carbon isotope values then increase again, where they reach a maximum (of 2.4 ‰), within the Bathonian. The carbon isotope data derived from the marls also follow this trend.

The wide range of oxygen isotopes and the low values, possibly points to a diagenetic overprint. Although a temperature control on oxygen isotopes cannot be excluded (see below), deep burial diagenesis and precipitation of calcite cement, commonly results in depleted in  $\delta^{18}\text{O}$  values (Hudson 1977; Weissert, 1989; Hönig & John 2015). The preservation of  $\delta^{13}\text{C}$  values or trends during carbonate diagenesis is, however, quite typical, and is likely due to the buffering effect of carbonate carbon on the diagenetic system, as this is the largest carbon reservoir (e.g., Scholle & Arthur 1980; Weissert, 1989). Hence, with respect to the oxygen isotope data, a diagenetic overprint affecting the samples analysed and results is likely. Although showing some scatter, oxygen isotope values remain fairly negative at the base of the section (the Aalenian) and become increasingly more positive upsection. The most positive oxygen isotope values are identified in the Bathonian (the Óbánya Limestone Formation).

## Discussion

### *Limestone–marl alternations*

The conspicuous limestone–marl alternations of the Komló Calcareous Marl Formation are likely to be caused by temporal variations in environmental parameters. It is generally accepted that the cause of the cyclical alternation of limestone beds and marls represents a direct response to changes in environmental conditions, such as productivity cycles (e.g. Wendler et al. 2002); dilution, i.e. changes in the influx of terrigenous non-carbonate material (e.g. Raucsik 1997; Weedon & Jenkyns 1999) or changes in input of carbonate mud from adjacent shallow-water carbonate factories (e.g. Pittet & Strasser 1998). Based on stable isotope data, Raucsik (1997) suggested that the higher  $\delta^{13}\text{C}$  of the limestones was associated with higher productivity, whilst terrigenous dilution may have formed the limestone-marlstone alternation. Given that the data presented here are consistent with the data of Raucsik (1997), in that the limestones typically record more positive  $\delta^{13}\text{C}$  values (Fig. 6), the same conclusion could be reached. A similar pattern could also be related to relatively short term changes in the export of neritic carbonate mud, as the  $\delta^{13}\text{C}$  of neritic muds, derived from relatively shallow waters, tend to show more positive values than carbonate ooze produced by planktonic organisms (e.g. Swart & Eberli, 2005). Indeed, Bajocian shallow-water carbonate factories on the southern Tethyan shelf (Leinfelder et al. 2002) are likely to show relatively positive carbon isotope values, although are somewhat distal to the study site of hemipelagic sedimentation on northern outer shelf of the Tethys Ocean (Fig. 2). The Bajocian was a time of widespread oolite formation along the Northern (and southern) Tethys margin (Wetzel et al. 2013). Isotope values from these Northern Tethyan oolites (Wetzel et al. 2013) do not show particularly positive values expected for aragonite oolites. Indeed textures indicate that these oolites were calcitic (Wetzel et al. 2013) (i.e. a calcite sea sensu Sandberg, 1983) and therefore this region exporting aragonite during this time appears unlikely. Of note is that the oxygen isotope data for the marls are more negative than the data derived from the limestones (Fig. 6), a pattern consistent with carbonate ooze produced in relatively warm surface waters.

Bodin et al. (2016), have also suggested lithological, rather than oceanographic controls on  $\delta^{13}\text{C}$  trends (e.g. during the earliest Toarcian of Morocco), whereby neritic  $\delta^{13}\text{C}_{\text{micrite}}$  signatures show more positive values than carbonate ooze produced by planktonic organisms. Changes in carbon isotope values in marine carbonate successions have also been attributed to changes in organic matter remineralization and subaerial exposure around hardgrounds and subsequent carbonate precipitation from meteorically influenced fluids (e.g. Immenhauser et al. 2002; Hönig &

165 John 2015). Evidence for subaerial exposure (e.g. signs of palaeokarst) was not observed in the  
166 Óbánya valley.

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### 168 ***Climatic and eustatic influences on carbon cycle changes***

169 The negative excursion followed by positive values that occurs near the base of the section  
170 (across the Aalenian–Bajocian boundary), observed in this study, appears to correlate with a major  
171 carbon cycle perturbation recognized elsewhere (Fig. 8) as a widespread phenomenon on the basis  
172 of its carbon–isotope expression in both oceanic (Bartolini et al. 1996; O'Dogherty et al. 2006;  
173 Sucheras-Marx et al. 2012; Hönig & John 2015) and terrestrial reservoirs (Hesselbo et al. 2003). It  
174 cannot be excluded that an earlier negative excursion followed by positive values occurring in the  
175 Aalenian Concavum Zone at Agua Larga (O'Dogherty et al. 2006; Sandoval et al. 2008), is  
176 correlatable with the lowermost negative excursion. However, this possibility is not favoured as no  
177 sizable negative shift is located above, in the Bajocian part of the succession. Carbon isotopes reach  
178 their most positive values, within the Óbánya succession, during the Early to mid-Bajocian, before  
179 declining across the Bajocian–Bathonian boundary. This same trend, is seen, for example in the  
180 Terminilletto section, Apennines, Italy (Bartolini et al. 1999) and the Betic Cordillera of southern  
181 Spain (O'Dogherty et al. 2006). Although there are evident differences in facies between these  
182 sections, due to deposition under differing conditions across the Tethys Ocean, the  $\delta^{13}\text{C}$  signatures  
183 are similar. The carbon-isotope trends are therefore likely to represent at least supraregional  
184 perturbations in the carbon cycle. Hence, wide scale mechanisms need to be considered to account  
185 for the observed trends.

186 Gradual negative carbon isotope excursions in the geological record have, for example, been  
187 explained by reduced primary production (e.g. Weissert & Channell 1989) whereby, increasingly  
188 oligotrophic conditions, caused by reduced run off and nutrient fluxes to the oceans, lower the  
189 fertility of surface waters which in turn leads to lessened primary production and a negative  $\delta^{13}\text{C}$   
190 shift. Such a mechanism for  $\delta^{13}\text{C}$  decreases has been associated with regressive conditions in the  
191 latest Jurassic Tethyan seaway (e.g. Weissert & Channell 1989; Tremolada et al., 2006). During the  
192 Aalenian–Bajocian boundary interval  $\delta^{13}\text{C}$  decreases have also been correlated with regressive  
193 intervals (Sandoval et al. 2008). O'Dogherty et al. (2006) also point out the coincidence between  
194 carbon cycle perturbations and major changes in marine biota. For example the latest Toarcian–  
195 Early Aalenian is marked by the coexistence of very low radiolarian content, high proportions of the  
196 nanofossil *Schizosphaerella* spp., and moderate proportions of *C. crassus*, indicative of



197 oligotrophic to mesotrophic palaeoceanographic conditions (Aguado et al. 2008). Although, there is  
198 no evidence of a regressive Aalenian-Bajocian boundary interval at Óbánya, a significant regressive  
199 event in Europe took place in Late Aalenian times (e.g. Hardenbol et al. 1998; Haq & Al-Qahtani,  
200 2005) followed by Early Bajocian transgression and deepening (Hallam, 2001). As noted by Hallam  
201 (2001), Underhill & Partington (1993) demonstrated that the Aalenian eustatic sea-level fall in the  
202 Jurassic was in fact a phenomenon of regional tectonics. Within Europe the effects of an Early  
203 Bajocian transgression can be recognised widely, for example, in Morocco (Bodin et al. 2017), north  
204 eastern Spain (e.g. Aurell et al., 2003) and in the Jura Mountains of southern France (e.g. Razin et  
205 al. 1996).

206 Major perturbations in the carbon cycle have also been associated with pulses of  
207 magmatism (e.g. Wignall 2001; Hesselbo et al. 2003; Pálffy et al. 2001). However, the carbon isotope  
208 excursion reported here and any association with a large pulse of magmatism is not clearly  
209 demonstrated. For example, radiometric data from the Karoo basalts indicates that the main  
210 volume of the Karoo Large Igneous Province (LIP) was emplaced between 181 and 184 Ma (i.e.  
211 during the Late Pliensbachian to Early Toarcian) with limited late stage basaltic activity at 176 Ma  
212 (e.g. Jourdan et al. 2008). Younger episodic magmatic activity, associated with the break-up of  
213 Gondwana following the formation of the Karoo LIP, is reported from Patagonia and the Antarctic  
214 Peninsula (Pankhurst et al. 2000). Aalenian–Bathonian volcanism is also reported from the Crimea  
215 (Meijers et al. 2010), the Caucasus region (Odin et al. 1993) and Mexico (Rubio-Cisneros & Lawton  
216 2011). Interestingly, the Aalenian–Early Bajocian interval also overlaps with the birth of the Pacific  
217 Plate and a major pulse of subduction related magmatism (Bartolini & Larson 2001; Koppers et al.  
218 2003). Evidence for the impact of this magmatic activity can be assessed through  $^{87}\text{Sr}/^{86}\text{Sr}$  data. The  
219 Aalenian–Early Bajocian seawater  $^{87}\text{Sr}/^{86}\text{Sr}$  curve shows, however, a flat segment alluding to the  
220 limited impact of this magmatic activity. This contrasts with the relatively rapid fall in the seawater  
221  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio seen through the Late Bathonian and Early Callovian (Wierzbowski et al. 2012). Hence  
222 it appears likely that the volcanogenic  $\text{CO}_2$  associated with these events certainly represents a  
223 potential source for light carbon, although possibly not of sufficient magnitude and sufficiently light  
224 to achieve the observed isotopic change.

225 Alternatively, an injection of isotopically light carbon into the ocean and atmosphere from a  
226 remote source, such as methane from clathrates, wetlands, or thermal metamorphism organic rich  
227 sediments (e.g., McElwain et al. 2005; Bachan et al. 2012) has been considered as means to explain  
228 negative carbon isotope excursions. Similar events have been considered to have been a result of

229 more regional events caused by recycling of isotopically light carbon from the lower water column  
230 (e.g. McArthur et al. 2008). However, that the Aalenian-Bajocian boundary event is observed in  
231 both marine (Fig. 8) and terrestrial settings (e.g. Hesselbo et al. 2003) has been considered to be an  
232 indication that the observed isotopic signals may have recorded a global (rather than regional)  
233 perturbation of the carbon cycle. As noted above, changes in the export of neritic carbonate mud  
234 (e.g. Swart & Eberli 2005) could also conceivably result in a negative isotope excursion in the  
235 geological record (e.g. Bodin et al. 2016; Ait-Ito et al. 2017). Hence a shift in the  $\delta^{13}\text{C}_{\text{micrite}}$  signature  
236 is possible without any relation to variations in the global carbon isotope trend (Bodin et al. 2016;  
237 2017). For this latter mechanism to be considered, sustained changes in the export of neritic mud  
238 are required to reach the study site and affect carbonate factories across Tethys. Furthermore, the  
239 negative excursion occurring in both oceanic and terrestrial reservoirs, provides an additional  
240 challenge for this to be a viable mechanism.

241 In contrast to the Aalenian-Bajocian boundary interval, more positive  $\delta^{13}\text{C}$  values (Fig. 8) in  
242 the Early Bajocian Tethyan seaway could have been linked to warmer climates and rising sea levels,  
243 increased runoff and nutrient fluxes to the oceans, increasing the fertility of surface waters (e.g.  
244 Sandoval et al. 2008; Suchéras-Marx et al. 2012). For example Suchéras-Marx et al. (2012) show  
245 that calcareous nannofossil fluxes increase markedly (mainly related to the rise of *Watznaueria*  
246 genus) from the upper part of the Aalenian to the Early Bajocian, coinciding with a positive shift in  
247 carbon isotope compositions of bulk carbonate. High levels of  $\text{CO}_2$  in the atmosphere could have  
248 also accelerated the transfer of nutrients from the continents to the oceans, through increasing  
249 weathering. Indeed, as noted above, significant injections of  $\text{CO}_2$  have been associated with major  
250 pulses of subduction-related magmatism, linked to the opening of the Pacific Ocean and the  
251 breakup of Pangaea (e.g. Bartolini & Larson 2001). Equally, the evolution of Tethyan seawater  
252 temperatures during the Middle Jurassic period inferred from the oxygen isotopic composition of  
253 belemnite rostra, bivalve shells and from fish teeth (see Brigaud et al. 2009; Price, 2010) reveal  
254 warmth during the Early Bajocian and cooling from late Bajocian times through into the Bathonian.  
255 Also, the oxygen isotope data of this study (Fig. 7) broadly replicate this trend, whereby more  
256 negative values are seen in the lower part of the succession and more positive values are observed  
257 in the upper part of the section and within the Bathonian. Such a pattern of warming and cooling is  
258 consistent with an Early Bajocian transgression noted above.

259 Increasing  $\delta^{13}\text{C}$  values in the Bajocian Tethyan seaway have also been linked to elevated  
260 productivity, as shown by radiolarian assemblages (Bartolini et al. 1999). O'Dogherty et al. (2006)

261 further point out ammonite radiations during the Early Bajocian, concomitant with increasing  $\delta^{13}\text{C}$   
262 values. The Early Bajocian positive excursion has also been correlated in the southern margin of  
263 western Tethys with a “carbonate production crisis” and concomitant with the onset of biosiliceous  
264 sedimentation in several basins (Bartolini et al. 1996). The Komló Calcareous Marl Formation  
265 shows, however, increasing carbonate content upwards (Velledits et al. 1986) rather than any  
266 marked decreases in carbonate. It is the Late Jurassic that sees biosiliceous sedimentation in the  
267 Óbánya valley (Velledits et al. 1986).

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### Conclusions

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Our study of the Komló Calcareous Marl Formation of the Mecsek Mountains of Hungary reveals a negative carbon isotope excursion followed by positive values that occurs near the base of the section (across the Aalenian–Bajocian boundary). The origin of this carbon-isotope anomaly is interpreted to lie in significant changes to carbon fluxes stemming from changes in primary production linked to increasingly oligotrophic conditions, caused for example, by reduced run off and nutrient fluxes to the oceans, lowering the fertility of surface waters which in turn leads to lessened primary production and a negative  $\delta^{13}\text{C}$  shift (e.g. O'Dogherty et al. 2006; Sandoval et al. 2008). That the Aalenian-Bajocian boundary carbon isotope event is observed in both marine and terrestrial settings (e.g. Hesselbo et al. 2003) indicates that the observed isotopic signals record global (rather than regional) perturbation of the carbon cycle. Changes in the export of neritic carbonate mud could also conceivably result in a negative isotope excursion, but this mechanism required sustained changes affecting carbonate factories across Tethys. Furthermore, the negative excursion occurring in both oceanic and terrestrial reservoirs, challenges this as a viable mechanism. In view of the gradual isotopic changes inferred from these Tethyan carbonates, an explanation in terms of the rapid dissociation of gas hydrates also appears unlikely. This study of the Komló Calcareous Marl Formation further confirms that the carbon isotope curves of Bartolini et al. (1999) and Sandoval et al. (2008), do indeed serve as a global standard for Aalenian–Bathonian  $\delta^{13}\text{C}$  variation.

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483 **Figure Captions**

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485 Figure 1 A. Location map showing the location of the Óbánya valley within Hungary. Grey inset box  
486 shows location of the Mecsek Mountains. B. Distribution of Mesozoic volcanic and  
487 sedimentary units within the Mecsek Mountains from Galácz (1994).

488 Figure 2. Middle Jurassic palaeogeographic map of the Western Tethyan realm (modified from Enay  
489 et al. 1993), localities; 1– Óbánya, 2 Wadi Naqab, United Arab Emirates 3 – Southern Iberia,  
490 4 – Umbria-Marche Basin (Central Italy), 5 – Cabo Mondego, Portugal, 6 – Chaudon Norante,  
491 SE France.

492 Figure 3. Lithostratigraphical scheme for the Jurassic deposits of the Mecsek Zone (Hungary)  
493 modified from Némédi Varga (1998) and Fózy (2012).

494 Figure 4 A, B, Sections of the Komló Calcareous Marl Formation in the Óbánya valley, (notebook for  
495 scale). C. The Óbánya Limestone Formation D. Upper Jurassic siliceous limestones and  
496 radiolarites the Fonyászó Limestone Formation.

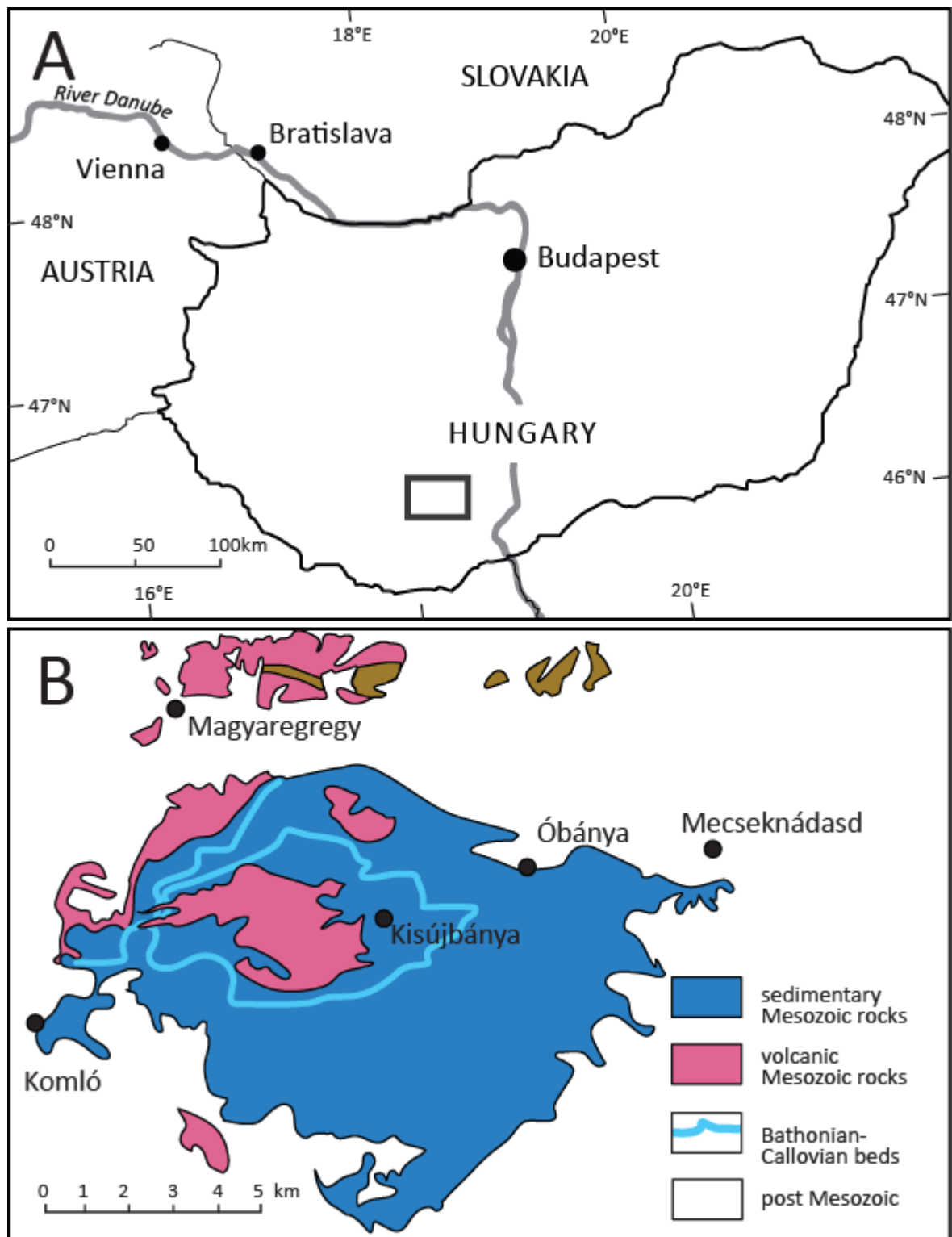
497 Figure 5 A. Photomicrograph of the limestone lithology (from the Komló Calcareous Marl  
498 Formation) dominated by calcite microspar (sample OB115, scale bar 1 mm). Small patches  
499 of coarser sparry calcite may have formed as a cement during diagenesis within primary  
500 porosity or by neomorphism of aragonite. B. Photomicrograph of marl lithology (from the  
501 Komló Calcareous Marl Formation) showing abundant small sparry bioclasts, including  
502 crinoids within a micritic and organic rich matrix (sample OB546, scale bar 0.2 mm). C.  
503 Photomicrograph of the Óbánya Limestone Formation showing abundant sparry small  
504 bioclast fragments within a muddy matrix (sample OB125, scale bar 0.2 mm).

505 Figure 6. Cross plot of  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  data from the Aalenian– Bajocian interval, Óbánya valley, of  
506 the Mecsek Mountains, Hungary. Data from Raucsik (1997) is also shown.

507 Figure 7. Isotopic results ( $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}_{\text{micrite}}$ ) from the Óbánya section. The ammonite data is from  
508 Velledits et al. (1986) and Galácz (1994). Zonal boundaries are not possible to identify  
509 because of very scattered occurrences of diagnostic ammonites. Crosses are the data  
510 derived from marls. The isotope curves (and 5 point running means) are plotted though the  
511 limestone data only.

512 Figure 8. Carbon isotope stratigraphies of the Aalenian– Bathonian interval from Óbánya compared  
513 with Southern Spain (from O'Dogherty et al. 2006), Cabo Mondego, Portugal (from

514 Suchéras-Marx et al. 2012); Chaudon Norante, SE France (from Suchéras-Marx et al., 2013);  
515 Wadi Naqab, United Arab Emirates (Hönig and John 2015) and the Umbria-Marche Basin,  
516 Italy(from Bartolini et al. 1999).  
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Figure 1

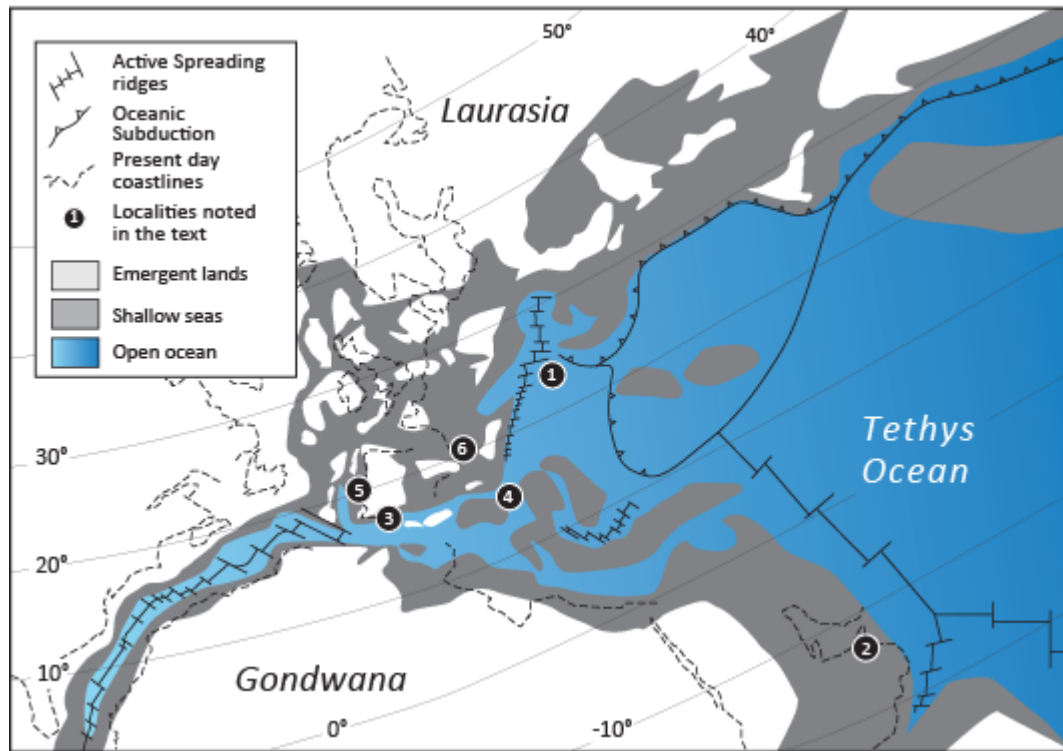


Figure 2

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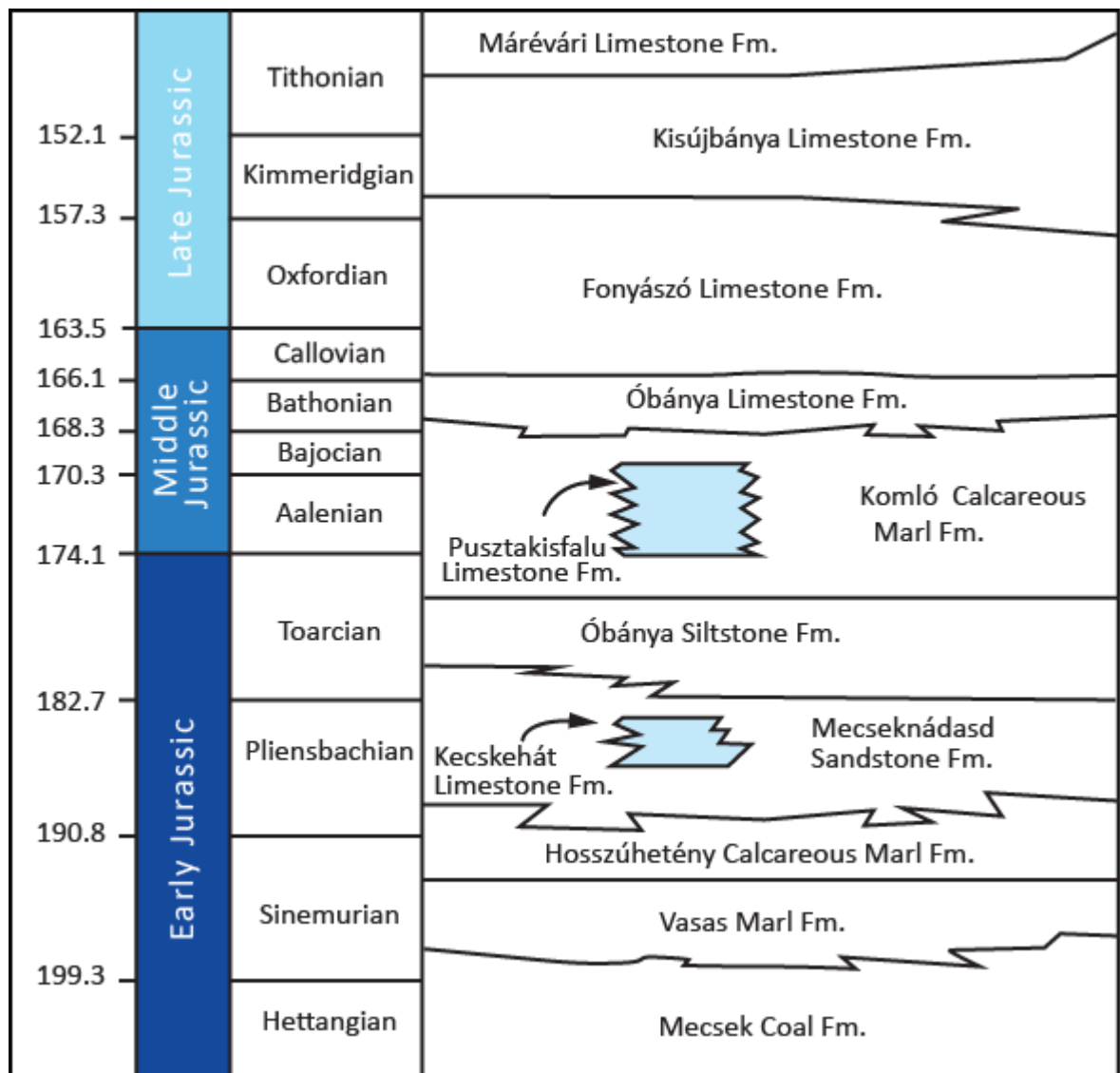
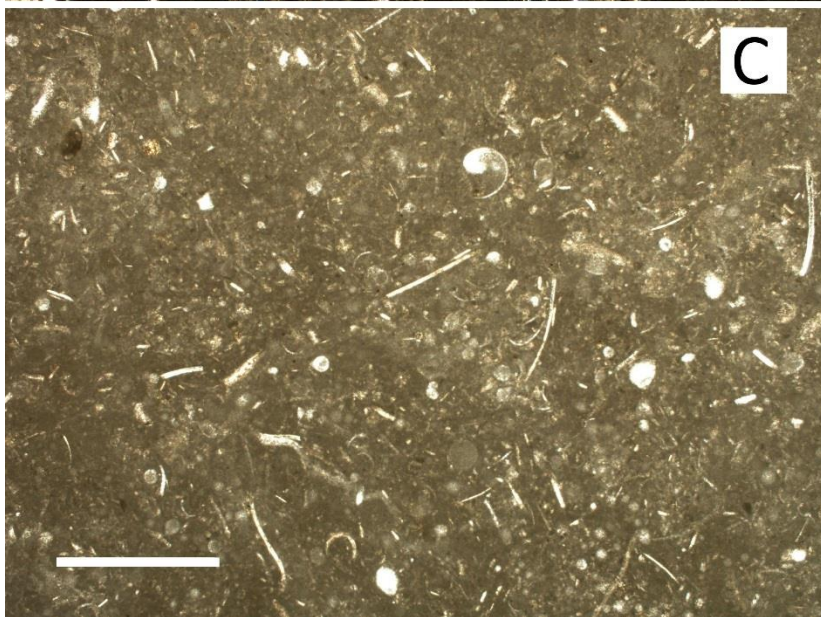
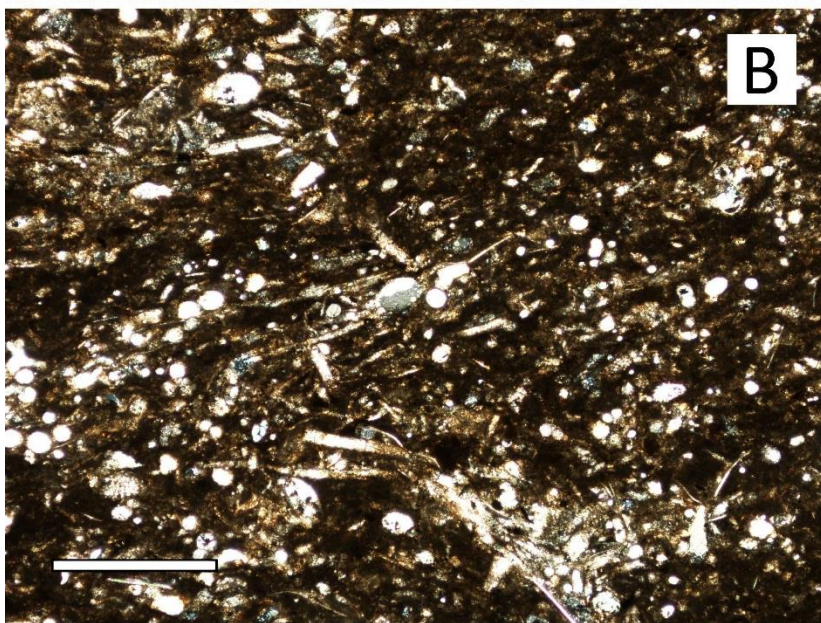
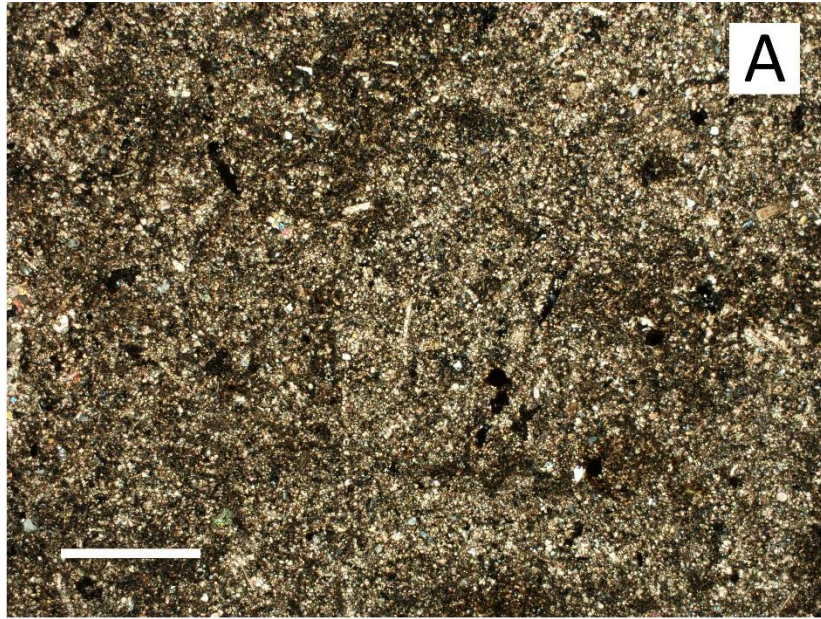


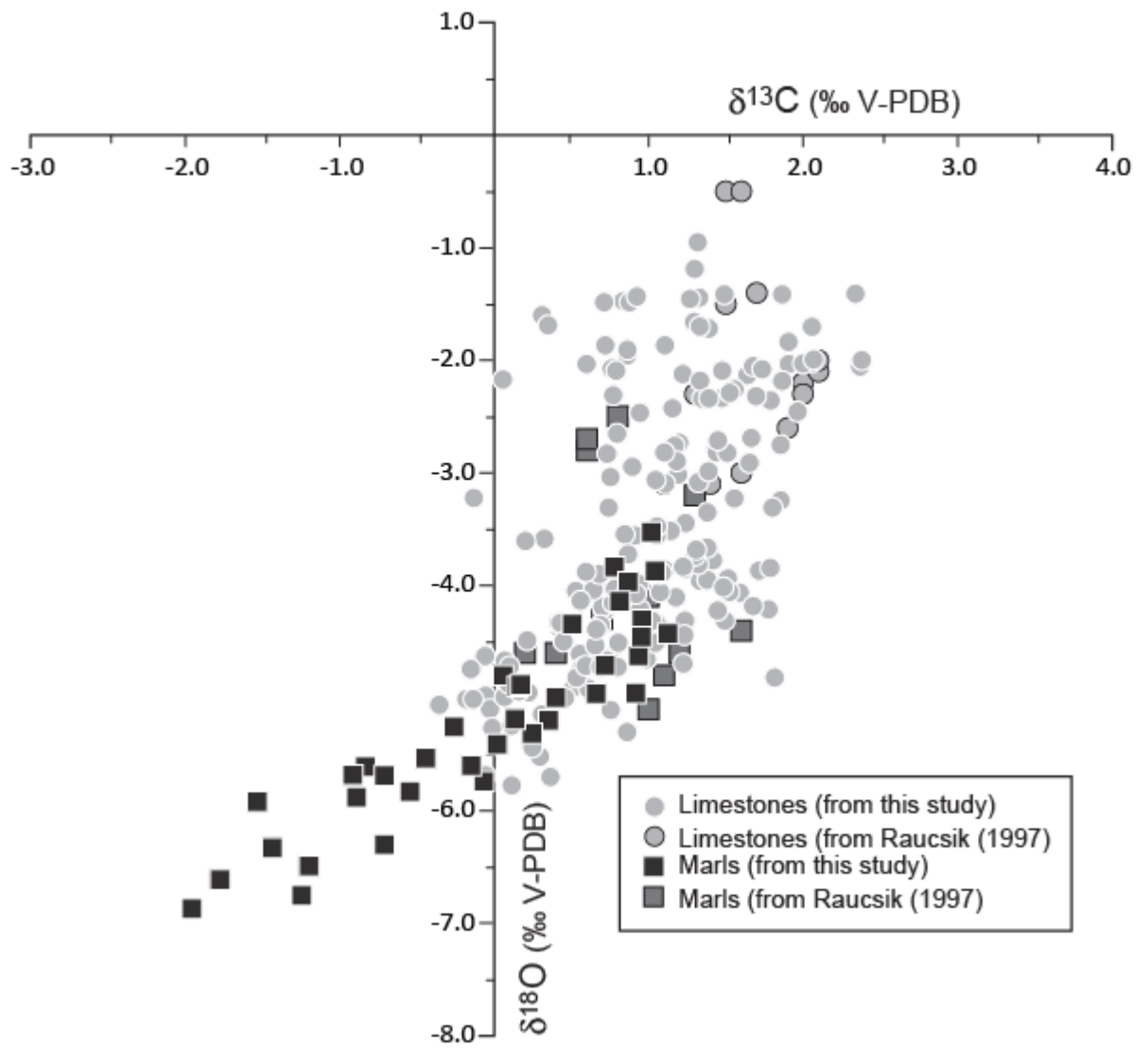
Figure 3

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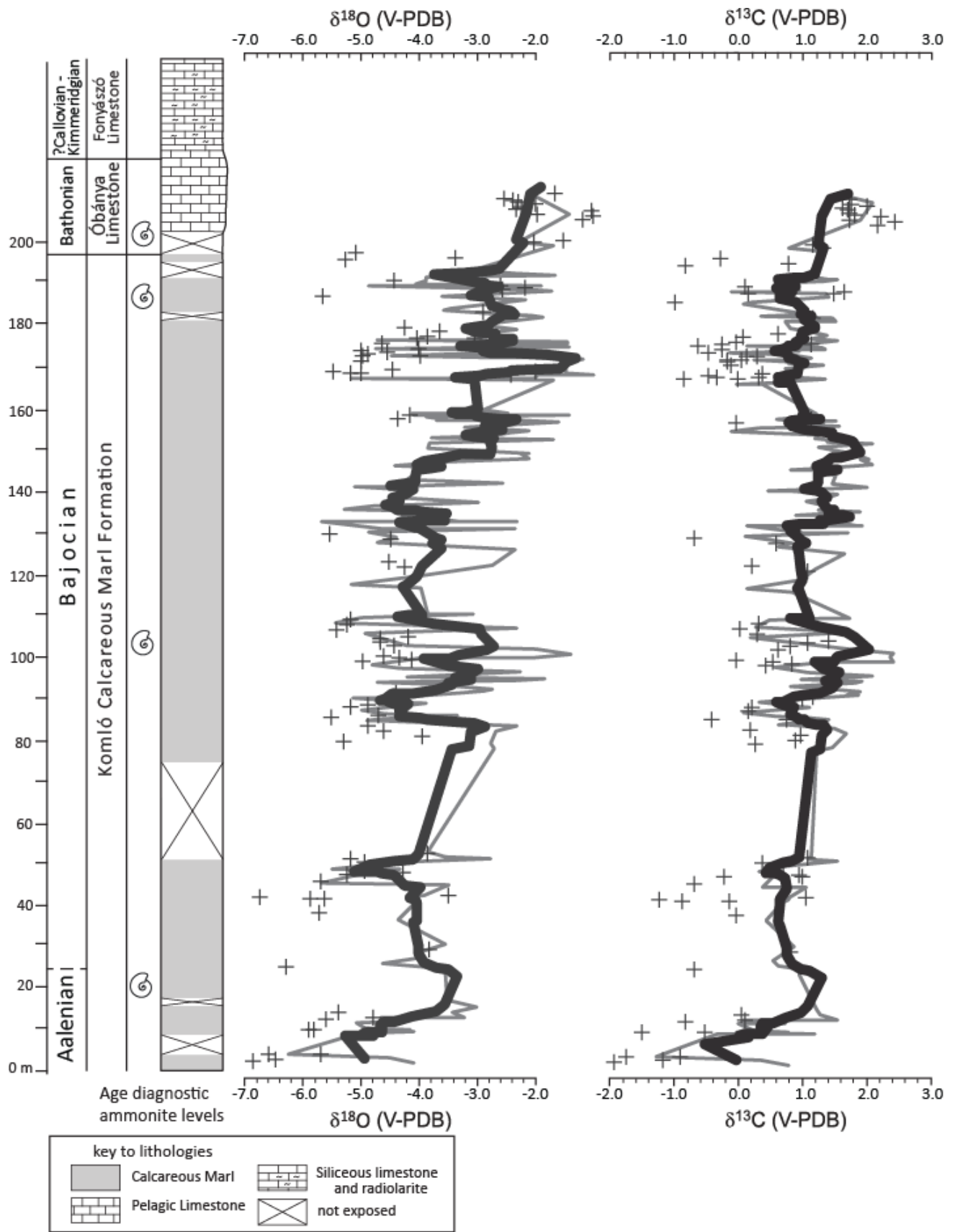






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Figure 6



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Figure 7

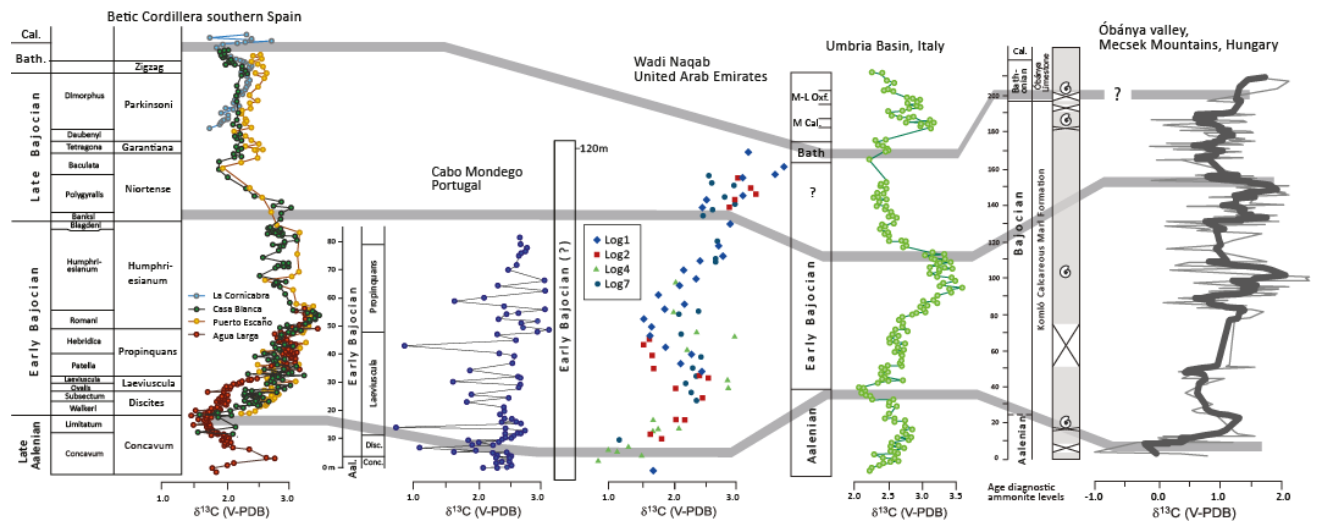


Figure 8

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