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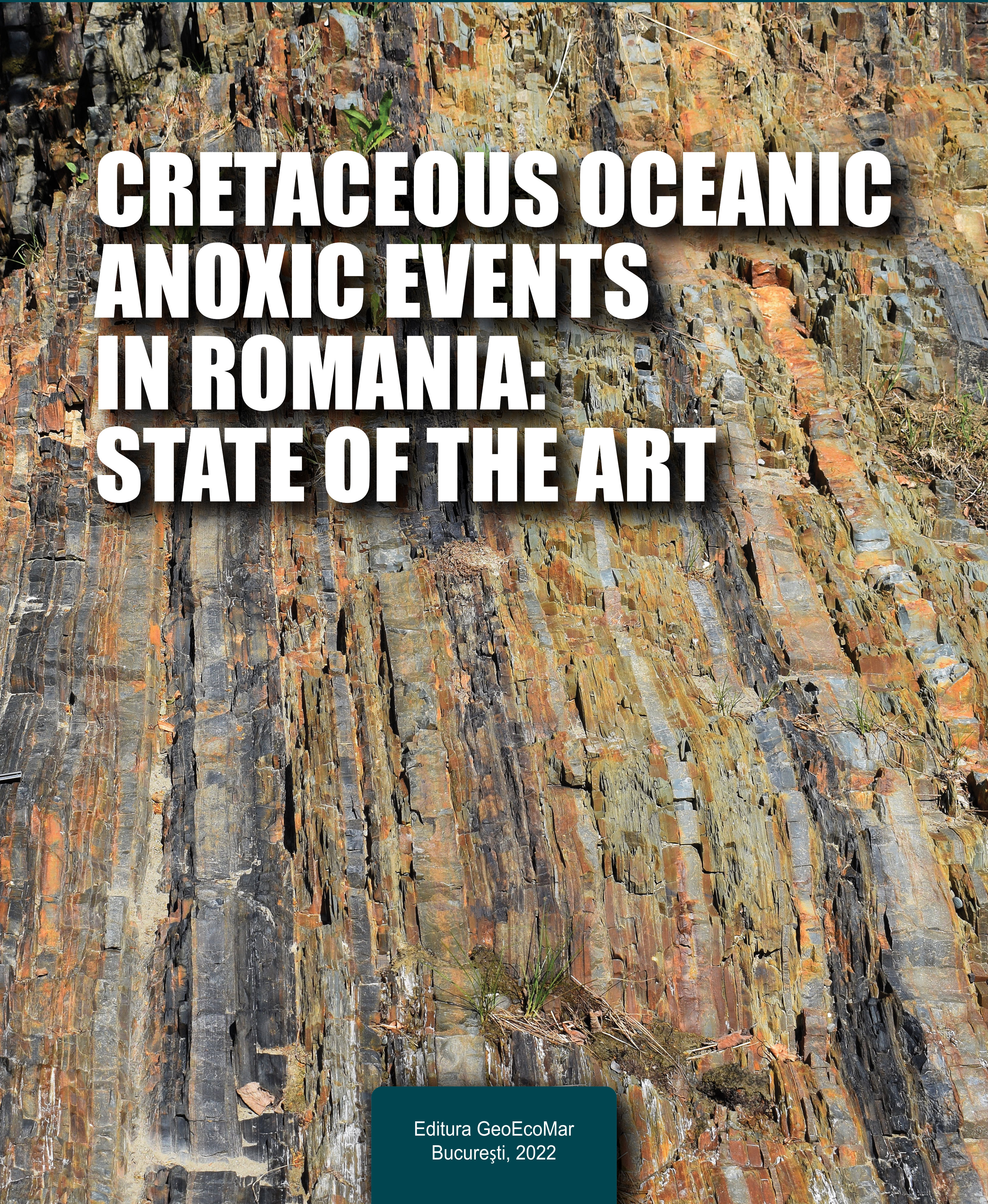

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CRETACEOUS OCEANIC ANOXIC EVENTS IN ROMANIA: STATE OF THE ART

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56

CRETACEOUS OCEANIC EVENTS: CAUSES AND CONSEQUENCES

The recent rise in sea-level (Dangendorf et al., 2017) in response to the increasing levels of atmospheric greenhouse gases and the associated global warming is a significant issue for the contemporaneity. During the Earth history, at some specific intervals, related to the glacial-interglacial events and older, such as the Cretaceous ones, significant sea-level fluctuations, with a higher amplitude that the ones recorded in recent history, are assumed to have taken place (Haq et al., 1987; Haq, 2014; Hu et al., 2017). These modifications, associated with a very warm and humid climate during the entire Cretaceous period and in special within the mid Cretaceous times, are described as 'the Cretaceous Greenhouse' (Sames et al., 2016; Wendler & Wendler, 2016, among many others). These phenomena could predict future climate modifications of our planet (Hay, 2011), if we have a better understanding of the past sea-level changes. Therefore, the Cretaceous climate research has entered in the last decades in an exciting, multidisciplinary phase in which geological, sedimentological, geochemical, mineralogical and palaeontological approaches are integrated, to better constrain the controls on climate change during warmth intervals.

Oceanic Anoxic Events are known to occur in many intervals of our planet history (Jenkyns et al., 1994; Courtillot, 1999; Jarvis et al., 2002, 2006, among many others) such as Cambrian (two OAEs), Ordovician (one OAE), Devonian (one OAE), Permian (one OAE), Jurassic (one OAE in Toarcian), Cretaceous (eight OAEs, including major and minor ones) and the youngest PETM, across the Palaeocene-Eocene (Fig. 1). Most of the OAEs are associated with major emplacement of igneous material due to the LIP (Large Igneous Province) appearance (Percival et al., 2015), while some are also associated with biotic extinctions. However, there are consistent extinction produced in absence of any OAE or LIP, and some known OAEs occurrence seem to be not related to any LIP.

An Oceanic Anoxic Event is primally recognised based on the significant shift of $\delta^{13}C$ isotope, on biotical fluctuation (especially of planktonic organisms) and on its lithological overprint (black shale deposition). The last marker may be present or not, while the first aforementioned two are mandatory in pointing out an OAE. Besides, relating to the geological time scale, an OAE is a short event, developing in a maximum 2 Ma (Arthur &

Premoli Silva, 1982; Arthur et al., 1985; Jarvis et al., 2006; Zorina et al., 2017; Yao et al., 2018).

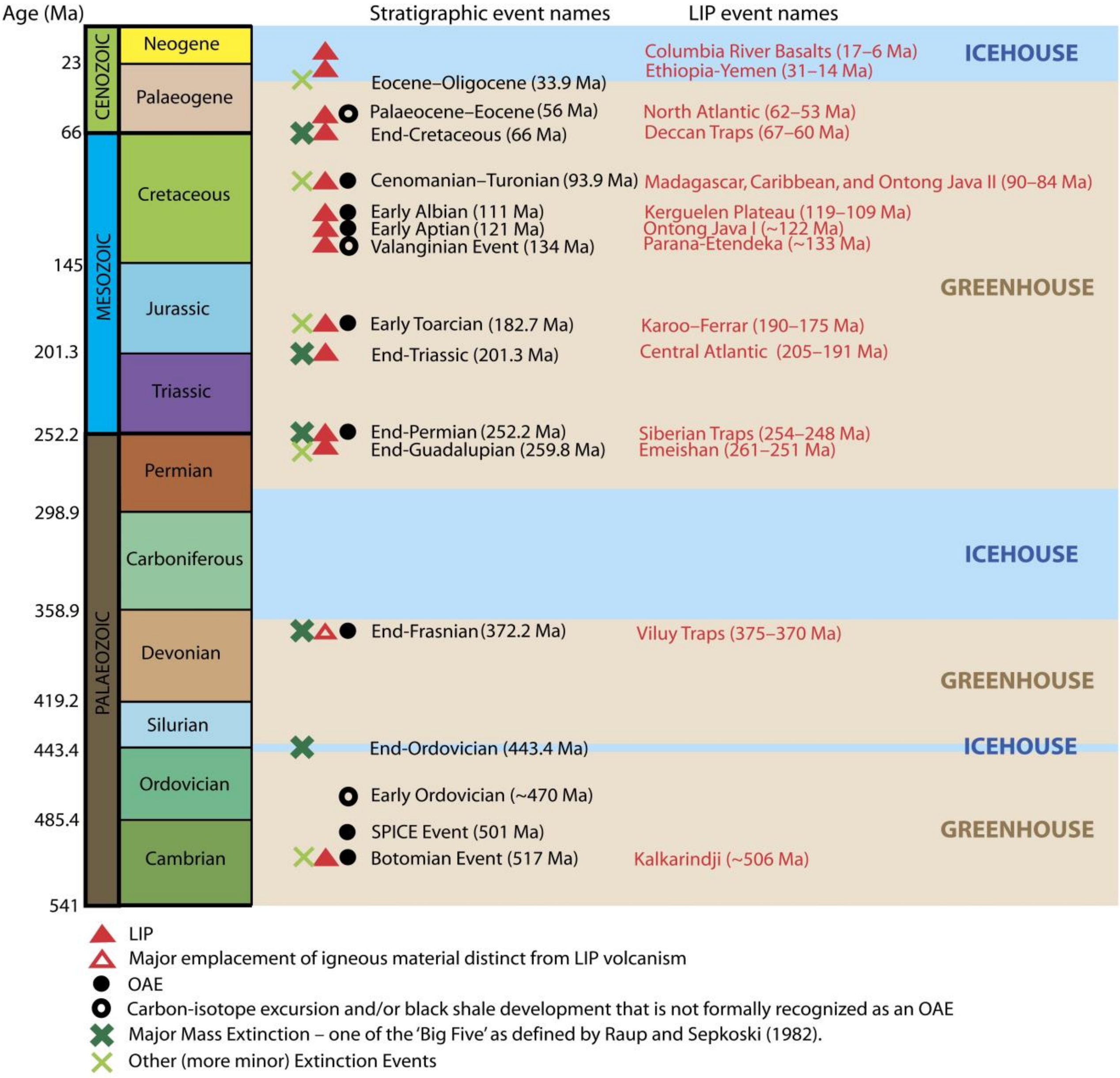

Fig. 1 - Intervals of OAEs (Oceanic Anoxic Events), distribution of LIP (Large Igneous Provinces) and climate modes (after Percival et al., 2015).

The Cretaceous C-isotope records show intermittent negative/positive spikes (Jenkyns, 2004) and consistent patterns of coeval chemostratigraphic curves. Therefore, they show a coeval signal responding to temporal variation of the Earth carbon reservoir.

Most of the OAEs were produced during the mid-Cretaceous, i.e., between the Aptian and Turonian stages (Leckie et al., 2002), an interval known as a super green-house climate mode (Hu et al., 2017), which is characterised by one of the highest sea-level record (Haq, 2014) in our planet history (Fig. 2).

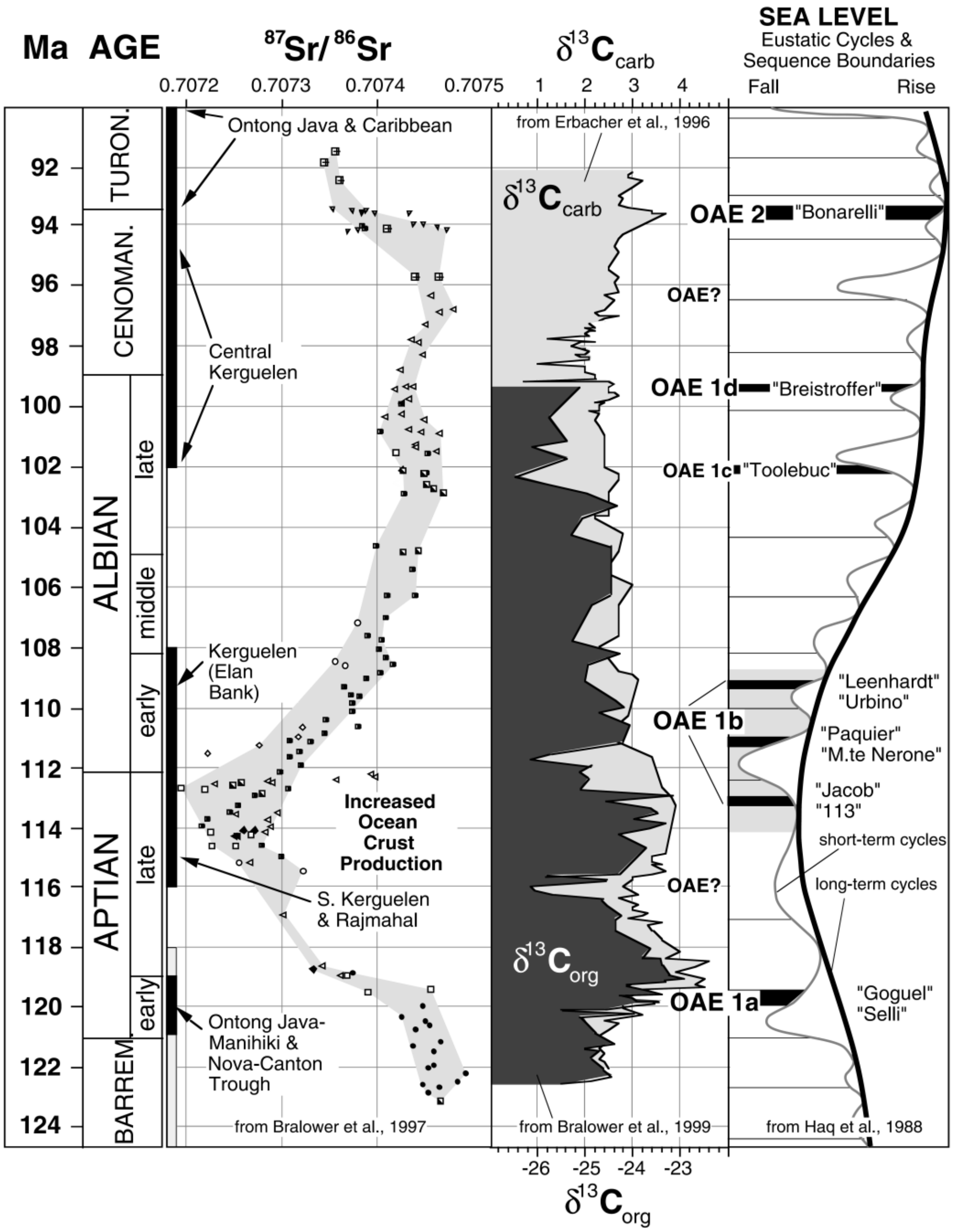

Fig. 2 – Mid-Cretaceous Oceanic Anoxic Events, fluctuation of isotopes ^{13}C and LIP intervals (from Leckie et al., 2002).

LOWER CRETACEOUS OCEANIC EVENTS

The Valanginian Weissert Event

Within the Valanginian, towards its top, the oldest Cretaceous oceanic anoxic event discovered so far, namely the Valanginian Weissert Event, covering around 2,000 Ky, is present (Weissert, 1989; Erba et al., 2004). This event was defined based on the identification of a global positive carbon isotope excursion in the upper Valanginian–lower Hauterivian sediments (Lini et al., 1992; Weissert & Erba, 2004) and coincides with the eutrophication event, proved by the marine planktonic organism distribution pattern (Fig. 3), especially foraminifers and calcareous nannoplankton (Premoli Silva et al., 1989; Erba, 2004, Martinez et al., 2013).

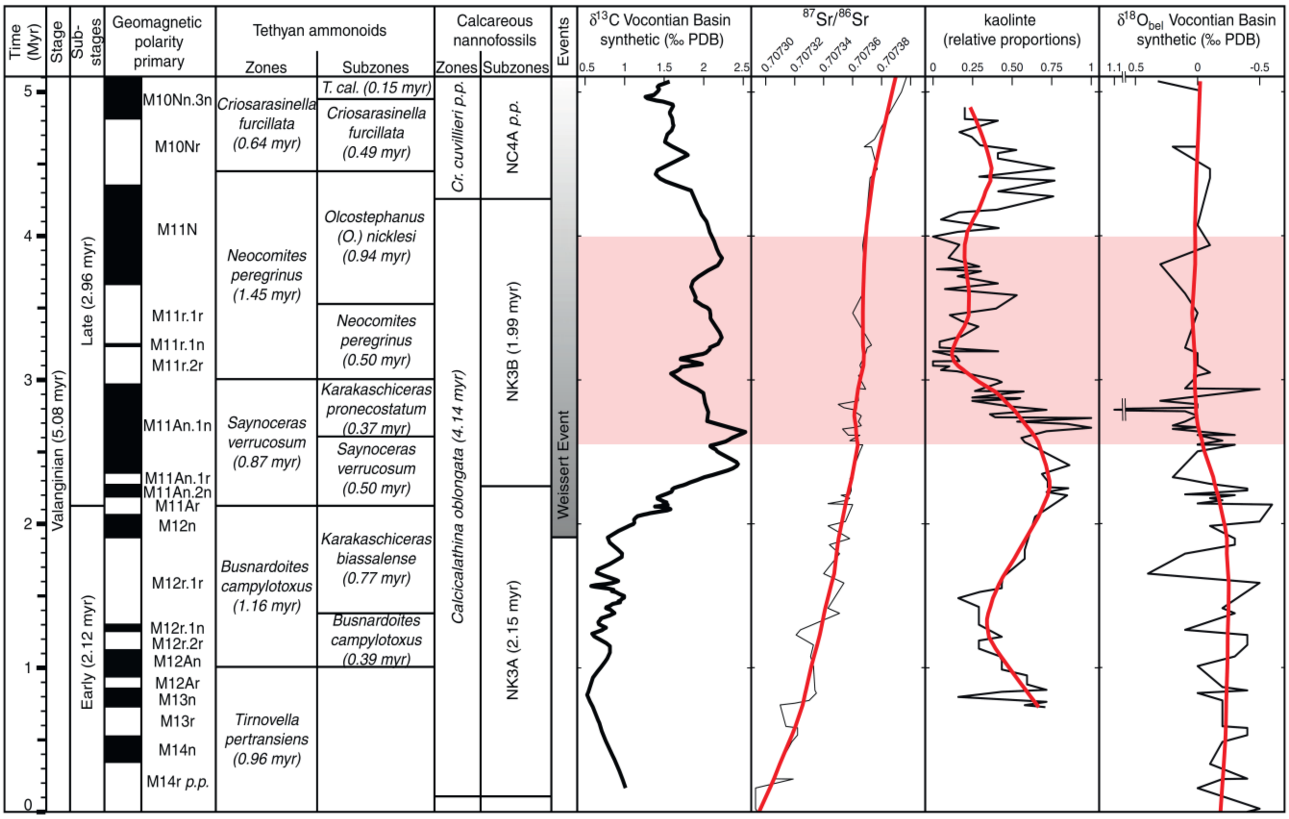

Fig. 3 – Fluctuation of planktonic organisms and geochemical modifications during the Weissert Event in the Vocontian Trough, France (after Martinez et al., 2013).

The late Valanginian drastic shift of Tethyan nannofloras (mainly reflected in the decrease abundance of the nannoconids) was possibly related to a biocalcification crisis, due to CO_2 excess, linked globally to the Parana-Etendeka volcanism and increased rates of oceanic crust production during the Gondwana breakup (Lini et al., 1992; Weissert, 1989; Weissert & Erba, 2004). This hypothesis is also supported by recent evidence that euxinic conditions were developed in the photic-zone, probably where the nannoconids lived during OAEs (Pancost et al., 2004).

The Valanginian Weissert Event in Romania

The oldest Cretaceous OAE globally discovered until now is also the first reported in Romania, identified in the Bucegi Mts. (Fig. 4), within the Mount Lespezi area (Barbu & Melinte-Dobrinescu, 2008), part of the Getic Carbonate Platform as described by Patrulius (1969) and Patrulius & Avram (1976). The studied succession encompasses the late Tithonian–late Valanginian interval, based on identified calpionellid and calcareous nannofossil assemblages.

No significant lithological changes were recorded in the above-mentioned interval, being characterised mainly by the deposition of a pelagic succession, i.e., bioclastic limestones that show strong similitude with the Maiolica Formation of the Lombard basin (Muttoni et al., 2005), possibly linked to the palaeolatitude plate motion.

The whole late Tithonian-early late Valanginian interval contains calcareous nannofossil assemblages clearly dominated by the Tethyan taxa of the *Nannoconus* genus, with over 30% of total assemblages. The nannofossil *Watznaueria barnesiae* abundance (around 30%) is comparable with the nannoconids abundance. Such assemblages characterise low latitudes areas, warm surface waters and relatively poor nutrient contents (Mutterlose & Kessels, 2000; Mutterlose et al., 2005), corresponding to the recent *Emiliania huxleyi* - *Gephyrocapsa oceanica* assemblage from the subtropical/tropical nannofloral bioprovinces (Melinte & Mutterlose, 2001).

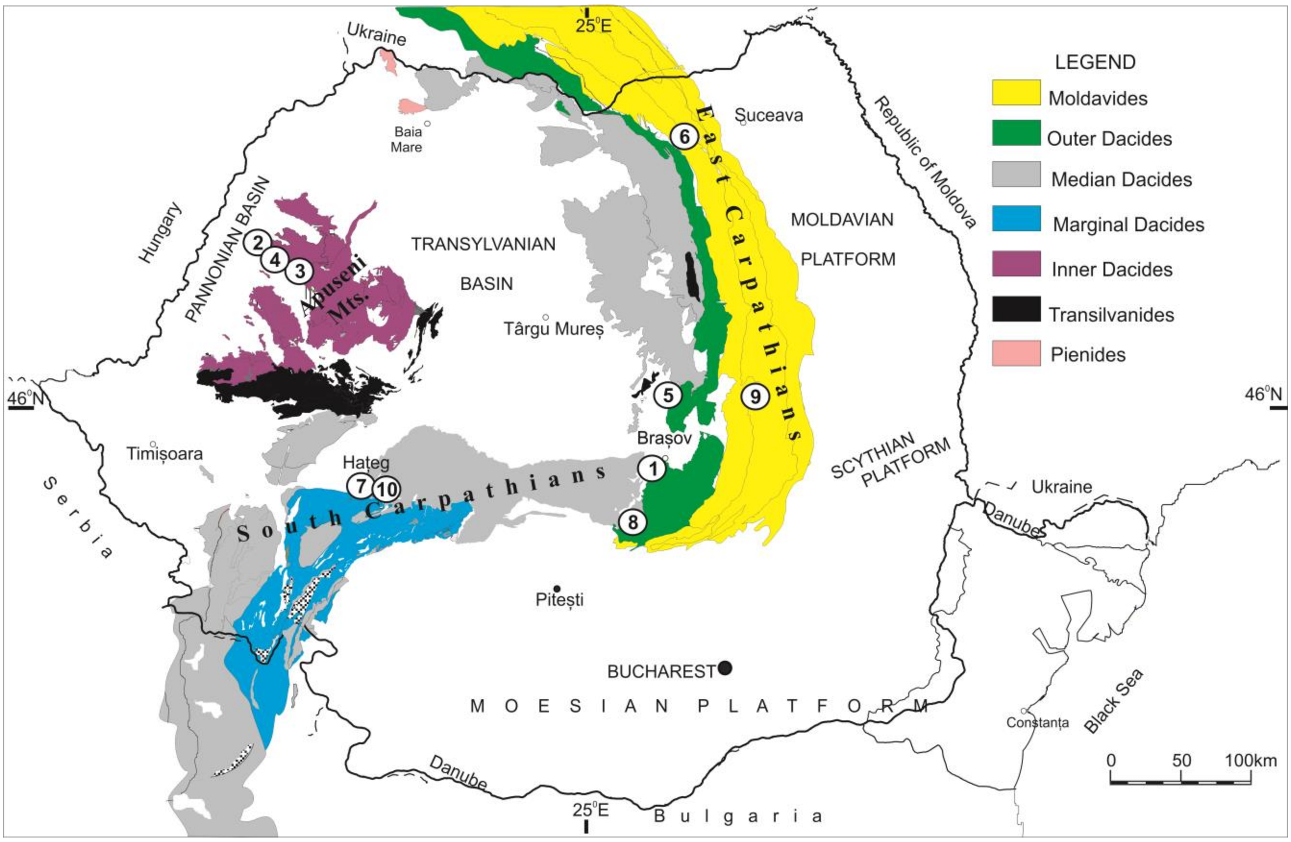

Fig. 4 – Location of the OAEs (Oceanic Anoxic Events) identified so far in Romania. Tectonic map simplified after Săndulescu, 1984. 1 - Valanginian Weissert Oceanic Events (Southern Carpathians, Bucegi Mts., Lespezi); 2 - OAE1a Aptian (Pădurea Craiului, Apuseni Mts.); 3 - OAE1b and OAE1c (Pădurea Craiului, Apuseni Mts.); 4 - OAE1d (Pădurea Craiului, Apuseni Mts.); 5 - OAE1d Albian-Cenomanian Boundary Event (Cernatu, central Eastern Carpathians); 6 - OAE1d (Ostra Valley, N Eastern Carpathians); 7 - OAE2 Cenomanian-Turonian Boundary Event (Ohaba-Ponor, Haţeg Basin, Southern Carpathians); 8 - OAE2 (Stoeneşti – Cetăţeni, Dâmboviţa Valley, S Eastern Carpathians); 9 - OAE2 (Lepşa, Putna Valley, central Eastern Carpathians); 10 - OAE3, Santonian-Campanian Boundary Event (Fizeşti section, Haţeg Basin, Southern Carpathians).

Within the Valanginian-Hauterivian boundary (in the NK3B subzone of Bralower et al., 1989 and NC4 zone of Roth, 1983), the nannoconids recorded a significant decrease, concomitantly with the increase of high-fertility index (Melinte & Mutterlose, 2001; Mutterlose et al., 2005), such as *Diazomatolithus lehmanii*, *Zeugrhabdotus erectus*, *Discorhabdus ignotus* and *Cyclagelosphaera margerelii* (Fig. 5).

Along with the significant variation of the Nutrient Index (NI) of calcareous nannofossil assemblages, in this depositional interval, *Zoophycos* trace-fossils and pyrite framboids, indicating dysaerobic conditions, were observed (Barbu & Melinte-Dobrinescu, 2008). These changes are indicative for a dysaerobic latest Valanginian – earliest Hauterivian episode, coincident with a nutrification episode globally expressing the Valanginian Weissert Event.

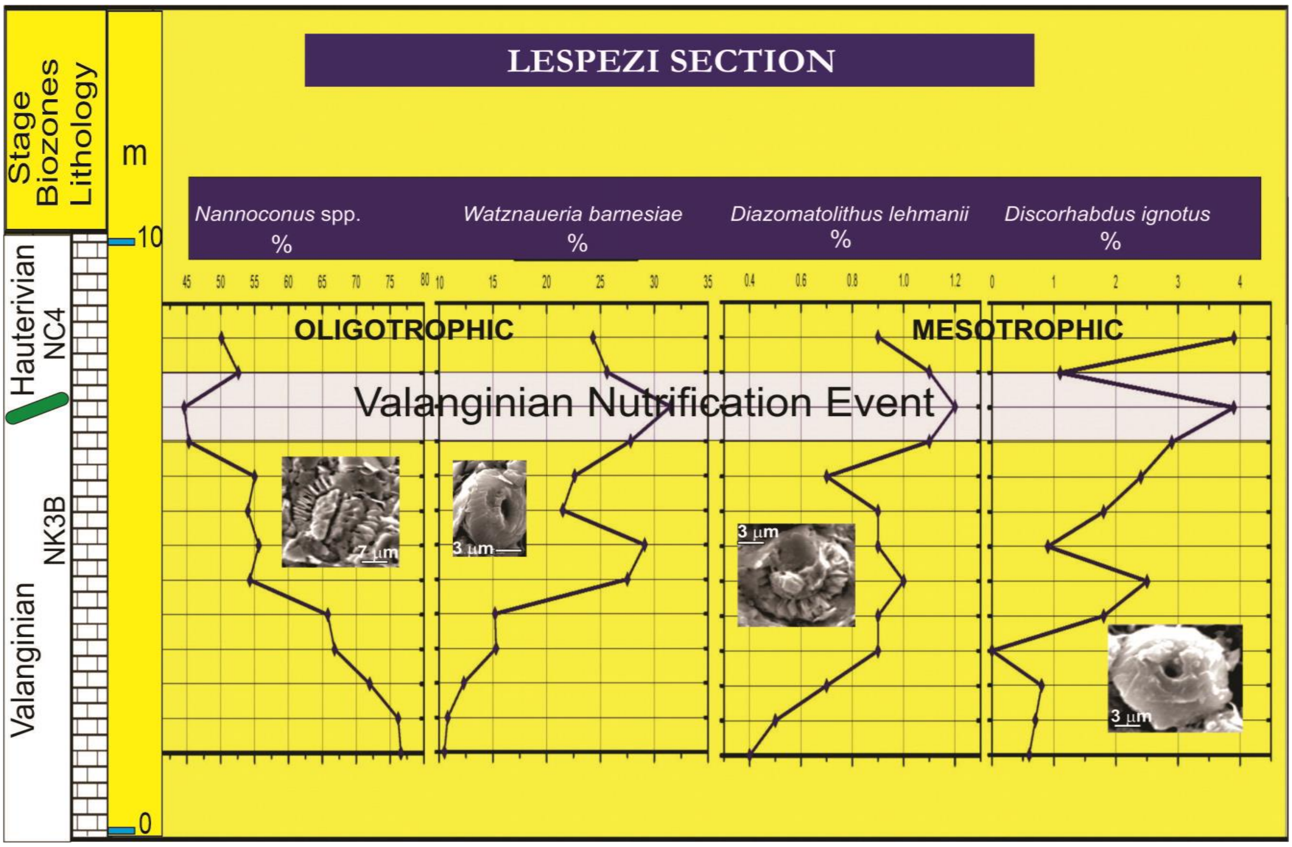

Fig. 5 – Fluctuation of oligotrophic calcareous nannofossil (*Nannoconus* spp. and *Watznaueria barnesiae*) and mesotrophic calcareous nannofossils (*Diazomatolithus lehmanii* and *Discorhabdus ignotus*) at the Valanginian-Hauterivian boundary interval in the Lespezi section.

In the same section, located in the eastern end of the Southern Carpathians, Barbu (2013) identified in the level containing a high abundance of eutrophic calcareous nannofossil taxa and dysaerobic conditions, as observed by Barbu & Melinte-Dobrinescu (2008), a positive $d^{13}C$ excursion. This modification is coeval with low oxygenation conditions leading to organic matter preservation, suggested by the presence of conical and trochospiral inflated benthic foraminiferal specimens and the dominance of small *Spirillina*.

MID CRETACEOUS OCEANIC EVENTS

The Cretaceous is a geological period that lasted from about 145 to 66 My. It is the third and final period of the Mesozoic Era. Actually, the Cretaceous is divided into Early and Late Cretaceous epochs, or Lower and Upper Cretaceous series (Gradstein et al., 2012). The boundary between the Lower Cretaceous and the upper Cretaceous is represented by the boundary between the Albian and Cenomanian stages (https://stratigraphy.org/gssps/).

However, in the international scientific community working on various aspects of the Cretaceous Period, the term mid Cretaceous is unconventionally used. The Aptian-Turonian interval is marked by major palaeoenvironmental changes, including a significant rise of palaeotemperatures of the ocean, this climate mode being described as a hothouse, i.e., “Super greenhouse” (Hallam, 1985; Hay & Floegel, 2012). The mid Cretaceous times are also characterised by a significant sea-level rise and by the presence of most frequent OAEs (Oceanic Anoxic Events) in all the Cretaceous Period (Arthur & Premoli Silva, 1982; Jenkyns et al., 1994; Sames et al., 2016, Wendler & Wendler, 2016).

OAE 1a (Barremian-Lower Aptian)

One of the most studied Cretaceous OAEs was identified in the Aptian, namely OAE1a, lasting around 1,250 Ky and globally described as a high-productivity event (Erbacher et al., 1996; Aguado et al., 2014). The OAE1a setting started within the Barremian-Aptian boundary interval, approximately 125 Ma ago, within a warming interval followed by a cooling one. Globally, the OAE1a is characterised by a pronounced, abrupt and stepped negative carbon isotope excursion, being recorded in sections from Europe, Northern and Central America, Asia and in the deep-sea drillings from Pacific and Atlantic (Bralower et al., 1994; Jenkyns, 2010).

The appearance of significant accumulation rates during the OAE1a setting as observed in sections from southern France (Wissler et al., 2002), northern Germany

(Mutterlose et al., 2009) and Mexico (Núñez-Useche & Barragán, 2012), among many other areas, are probably linked to pronounced weathering and transport of continental materials to the marine realm. Therefore, the nutrient input increased, possibly due to the greenhouse climate mode that was a warm and humid one (Weissert, 1989; Dumitrescu & Brassell, 2006; Melinte-Dobrinescu & Roban, 2011; Giorgioni et al., 2015).

It was suggested that the OAE 1a mirrors the release of ^{13}C-depleted carbon in the ocean-atmosphere Earth System, from either marine volcanism ($\delta^{13}C$ approx. 5‰) or dissociation of methane gas hydrates ($\delta^{13}C$ around 60‰) (Arthur et al., 1985). Several investigations of the isotopes $\delta^{13}C$ and $\delta^{18}O$ shift indicate that, during the OAE1a, there is not only one C cycle-perturbation, but multiple ones (Meyers, 1994; Dumitrescu & Brassell, 2005) most probably linked to the intensive volcanism and the CH_4 gas hydrate dissociation.

A good example of the OAE1a development is the Organyá Basin, NE Spain (Sanchez-Hernandez et al., 2014). Organic-rich levels concurrent with positive excursions in $\delta^{13}C_{org}$ imply enhanced preservation of OM (organic matter) from export production (Fig. 6). Several phases of $\delta^{13}C_{org}$, based on its fluctuation pattern, were observed in the investigated Spanish section, including a pronounced peak, higher than globally reported for OAE1a. This finding could reflect the coexistence of local factors, such as the geomorphology of the basin, added to the global climatic deterioration. The highest $\delta^{13}C_{org}$ is located below the FO (first occurrence) of the nannofossil *Rhagodiscus angustus* (Fig. 6). The abovementioned calcareous nannofossil event is recorded elsewhere above the magnetic Chron M0, within the lowermost Aptian (Erba, 1994). The whole nannofloral assemblages of the investigated succession belong to the NC6 Biozone, coincident with the Chiastozygus litterarius Zone (Roth, 1983). Hence, the maximum of the OAE1a is situated in the latest Barremian, while the final pulses, with smaller amplitudes of this global anoxic event are situated in the Aptian. Additionally, synchronous peaks of TOC, TIC and $\delta^{13}C_{org}$ suggest that enhanced local phytoplankton productivity in the N Spanish Basin simultaneously intensified the oxygen demand in the water column and led to higher burial of organic carbon under oxygen-deficient sub-surface waters.

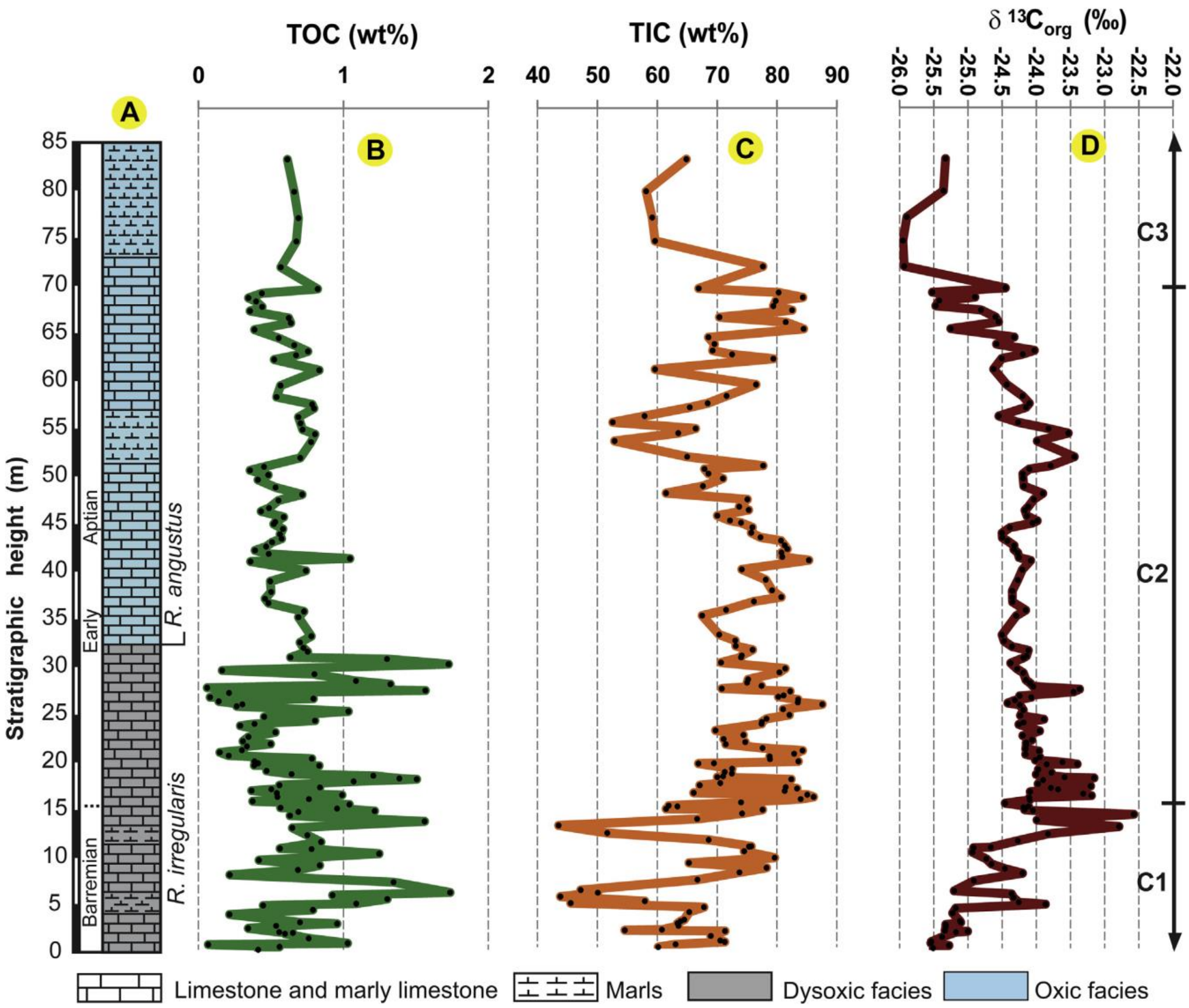

Fig. 6 – The lithological log and geochemical variations recorded in El Pui section (NW Spain).

A: Lithological column including lithological variations, oxic, dysoxic levels, and relevant nannofossil events. **B:** TOC. **C:** TIC. **D:** $\delta^{13}C_{org}$; C1, C2 and C3 subdivisions correspond to the carbon isotopic segments identified in the section (from Sanchez-Hernadez, Y., Maurrasse, F., Melinte-Dobrinescu, M., He, D., Butler, S.K., 2014).

Another significant exposure of OAE1a is placed in the Southern Alps of Italy (Cismon section). There, in the Majolica Formation, several excursions, positive and negative ones were identified as C1, C2, C3, C4 and C5 (Menegatti et al., 1998). The negative excursion of $\delta^{13}C$ was most probably linked to a huge release of light volcanic C during the Ontong Java Plateau (Larson 1991a, b; Erba, 2004; Pancost et al., 2004).

Palaeoclimatic reconstruction based on the clay mineral distribution of OAE1a (Sanchez-Hernandez et al., 2014) is consistent with temperate-warm/humid conditions as proposed by Aguado et al. (2014). The absence of the kaolinite, indicating high temperature in several Tethyan basins, where OAE1a was recorded, is most probably linked to the palaeomorphology of the basin and regional factors.

Interestingly, the overall climate deterioration during one of the most significant mid Cretaceous OAE that is OAE1a did not produce extinctions in the marine phytoplankton world. In the calcareous nannoplankton group, a speciation took place and no extinction was recorded; only some species temporarily disappeared from the record.

Oceanic Anoxic Events OAE1b, OAE1c and OAE1d

Above the OAE1a, from the Lower Aptian up to the Lower Turonian, there are several OAEs, such as OAE1b to OAE1d, which extend from the Aptian up to the Albian-Cenomanian boundary interval. These were regarded as sub-events, as the distribution seem firstly to be spatial limited, but detailed recent investigations confirm their global character.

The OAE1b took place towards the base of the Albian and lasted only 46 Ky, being the expression of the northern Kerguelen Plateau eruption, the OAE1c extends on around 1,000 Ky, but it is not apparently connected to any LIP, while the OAE1d covers around 300 Ky, within the Albian-Cenomanian boundary interval, when the eruption of the central Kerguelen Plateau occurred (Arthur et al., 1990; Larson 1991a; Coffin et al., 2002; Gale et al., 2011).

The OAE1d was originally identified at Gubbio, Umbria-Marche, central Italy (Jenkyns et al., 1994), and subsequently recognised in Mont Risou (Gale et al., 1996) and western North Atlantic (ODP Site1052) (Wilson and Norris, 2001; Petrizzo et al., 2008). More detailed $\delta^{13}C$ records for the Umbria-Marche Basin have been provided by Gambacorta et al. (2015) and Giorgioni et al. (2015). Subsequently, Bornemann et al. (2005 and 2017) made a comprehensive correlation between the OAE1d record from the Boreal Realm and the western Tethys. In the western Tethys, i.e., at Mont Risou, France, where the GSSP for the base of the Cenomanian is located (Kennedy et al., 2004), four episodes of oceanic anoxia have been recognised in the late Albian–lower Cenomanian depositional interval.

According to Gale et al. (1996), the carbon isotope curve of OAE1d registers a broad overall plateau through much of the Risou section and shows four discrete peaks (labelled a–d) separated by sharp, short-lived negative carbon isotope excursions (CIEs). The Monte Petrano section, located in the Umbria-Marche Basin, provided a high-resolution $\delta^{13}C$ record (Gambacorta et al., 2015). In addition, within this $\delta^{13}C$ curve, the four peaks (labelled a–d) defined by Kennedy et al. (2004) are ascertained. Wilson and Norris (2001) identified the OAE 1d at the Albian-Cenomanian boundary (in the Rotalipora appenninica and Rotalipora globotruncanoides foraminiferal zones) at ODP Site 1052E, on Blake Nose in the North Atlantic Ocean.

An interesting finding of OAE1d is located in the Tibet area, where no prominent black shales, the lithological marker of an OAE, occur. In order to determine the age accurately, a detail biostratigraphy was first realized, for giving a precise age of the exposed sediments in the Tingri area (Yao et al., 2018). The above-mentioned authors indicate, based on calcareous nannofossil investigations, that the expanded Youxia section spans a large time interval, i.e., the upper Albian-lower Turonian, including the Albian-Cenomanian boundary (Fig. 7), where the OAE1d is globally reported.

In term of calcareous nannofossils, the boundary between the Albian and Cenomanian lowers in the upper part of the UC0 biozone (Burnett, 1998; Gale et al., 2011). At the GSSP (Global Boundary Stratotype Section and Point) for the base of the Cenomanian, located at Mont Risou, France, the FO of *Prediscosphaera cretacea* was reported above the boundary and below the FO of the nannofossil *Corollithion kennedyi* (the base of UC0) (Kennedy et al., 2004). The same succession of calcareous nannofossils is present in the studied section from Tibet, as globally reported (e.g., Burnett, 1998; Kennedy et al., 2004; Bornemann et al., 2005; Scott et al., 2013; Melinte-Dobrinescu et al., 2015; Bąk et al., 2016).

Nannofossil stratigraphy indicates that the portion of the Youxia section examined in this study spans the upper Albian-lower Turonian depositional interval. The boundary between the Albian and Cenomanian falls within the upper part of the UC0 biozone (Burnett, 1998; Gale et al., 2011).

Following the nannofossil biostratigraphy, the C-isotope geochemistry was applied at the Tibetan section. Additionally, bio- and cyclostratigraphy have been also realised. The investigation of isotope $\delta^{13}C$ values (Fig. 8) indicated the presence of OAE1d (Yao et al., 2018).

The Youxia section is a good example for the presence of an OAE without any lithological marker. The absence of deep anoxia in the area of eastern Tethyan Domain probably is linked to palaeoceanographic conditions, characterised by better oxygenated deep-water palaeoenvironment. A comparison of the French, Italian and Atlantic Ocean curves with the Youxia $\delta^{13}C$ curve suggests that all four carbon isotope peaks can be recognised in the aforementioned section of the Tethyan Himalaya.

The spectral analysis of $CaCO_3$ fluctuation in the succession containing the OAE1d shows that the sedimentary cyclicity was mainly controlled by a short eccentricity (~100 Ky) and precession (22.2 Ky). Based on these facts, the duration of the OAE1d was identified as ~233 Ky (Fig. 8).

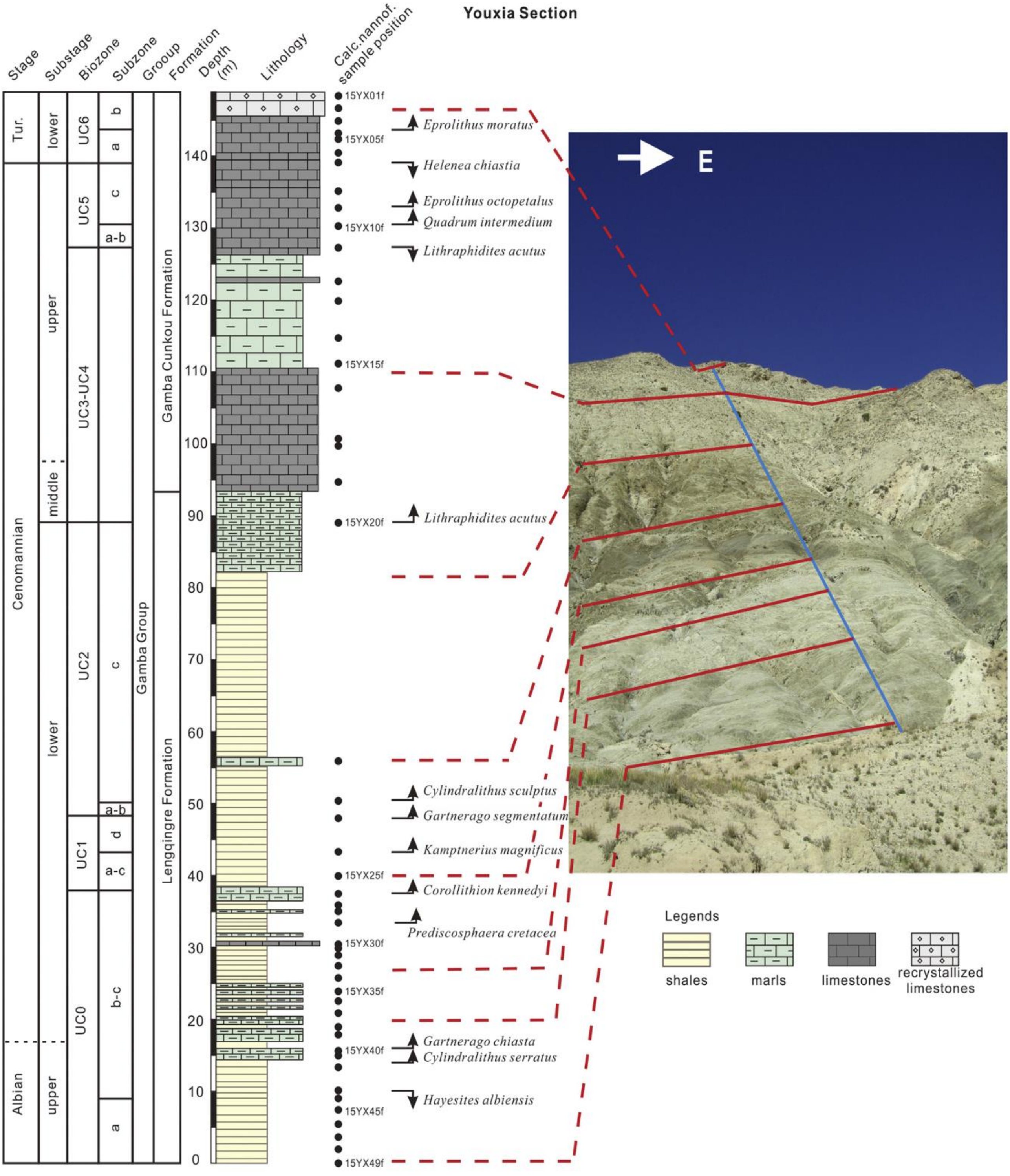

Fig. 7 – Litho- and biostratigraphy of the Youxia section (Tibet) after Yao, H., Chen, X., Melinte-Dobrinescu, M.C., Wu, H., Liang, H. & Weissert, H. (2018).

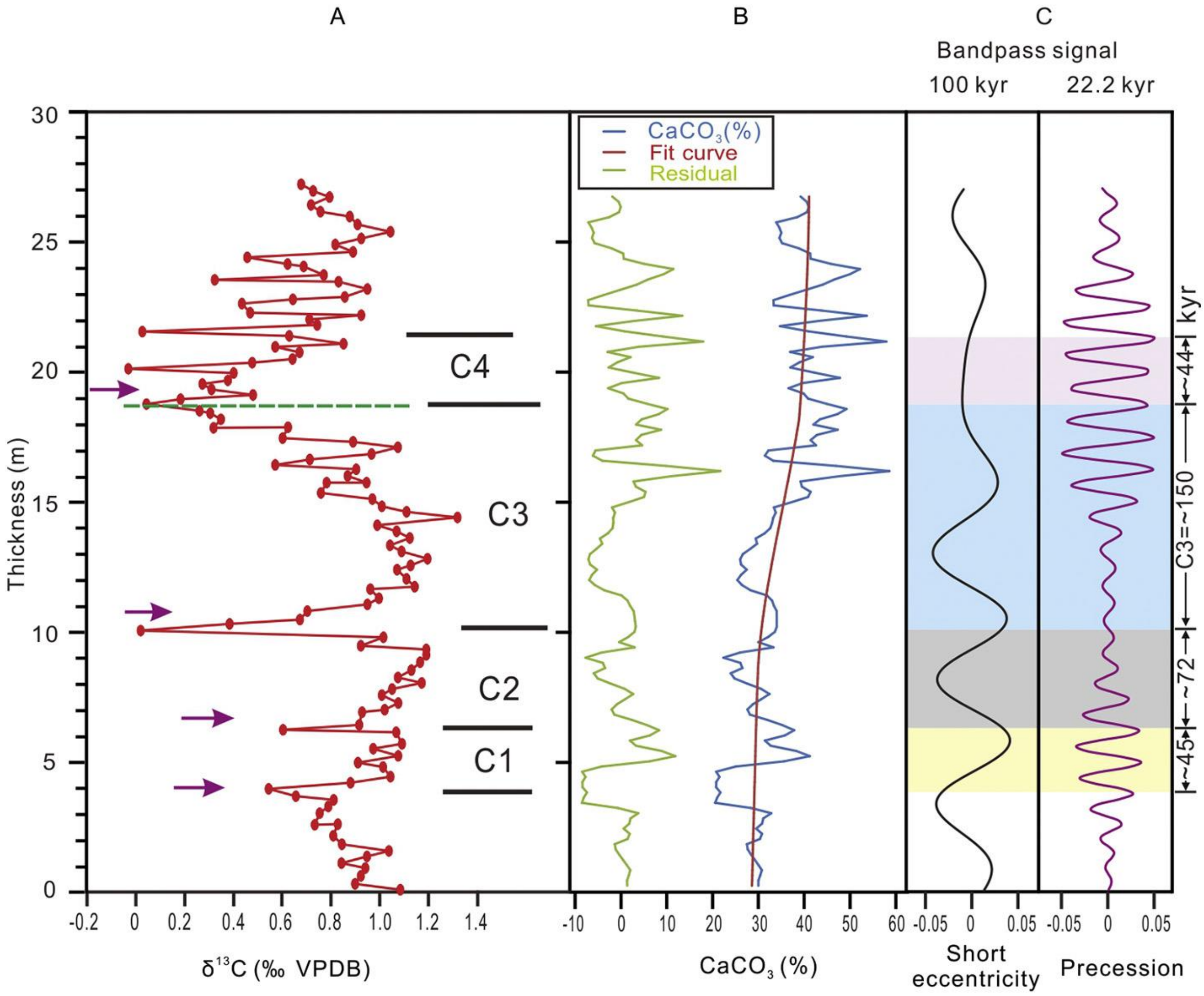

Fig. 8 – Geochemistry and cyclostratigraphy of Youxia section (Tibet) after Yao, H., Chen, X., Melinte-Dobrinescu, M.C., Wu, H., Liang, H. & Weissert, H. (2018).

OAE2 (Cenomanian-Turonian boundary interval)

A prominent oceanic anoxic event of the mid Cretaceous times is OAE2 (Schlanger & Jenkyns, 1976; Arthur et al., 1988 and 1990). It records a rapid global positive excursion of approximately 2‰ ($\delta^{13}C$ enrichment), which occurred from the late Cenomanian to the earliest Turonian interval (Arthur et al., 1988). Extensive data have been published on the Oceanic Anoxic Event 2 (OAE2), known also as CTBE (Cenomanian-Turonian boundary Event) or ‘Livello Bonarelli’, which is recognised as one of the more prominent oceanic anoxic events in the Earth history (Jenkyns, 2010). The OAE2 is also coincident with a significant biotic turnover, as documented in the diversity and composition of the marine planktonic faunas and floras, especially of calcareous nannofossils (Lamolda et al., 1994; Paul et al., 1999; Premoli-Silva et al., 1999; Voigt, 2000; Erba, 2004; Mutterlose et al., 2005; Linnert et al., 2010; Melinte-Dobrinescu et al., 2013).

A very interesting section exposing the OAE2 is situated in N Spain, Asturias region, in Arobes (Lamolda et al., 1997; Melinte-Dobrinescu et al., 2013). The OAE2 is delimited by the positive excursion of $\delta^{13}C$ and changes in the calcareous nannofossil composition (Melinte-Dobrinescu et al., 2013). Based on the planktonic foraminiferal biostratigraphy, with significant events like the LO of *Rotalipora. greenhornensis*, followed by the LO of *R. cushmani* and based on the nannofossils events, such as the FO of *Quadrum gartneri,* we calculated that the OAE2 of Arobes (Phase A and Phase B) lasted around 70 Ky. The shape and magnitude (up to 5.5‰) of the $\delta^{13}C$ isotope curve in Arobes (Fig. 9) is similar to those reported from other successions containing OAE2, such as Wadi El Ghaib of eastern Sinai (5‰: Gertsch et al., 2010), Eastbourne, England (around 5‰: Jarvis et al. 2006), Tarfaya and Agadir, Morocco (3–4‰: Keller et al., 2008), but higher than at Pueblo, Colorado (2.5‰: Keller et al., 2004).

Taking into account the results of the isotope analysis, several phases have been described and correlated for the first time with nannofloral events. Hence, the pre-excursion phase of OAE2 with low $\delta^{13}C$ values is characterised in Arobes by oligotrophic nannofloral assemblages, dominated by *Watznuaeria barnesiae* and *Eprolithus floralis*. The nannofossil *Cyclagelosphaera margerelii* shows a significant abundance distribution pattern, probably related to the palaeogeographical setting of the studied section, placed on middle neritic shelf. During the OAE2 onset (i.e., the first build-up phase), small modifications in calcareous nannofossil composition were remarked, except for the significant increase in the oligotrophic species *E*. *turriseiffelii*, followed by the shift in *Prediscosphaera* spp.

This pattern possibly mirrors the transition from an oligotrophic to a mesotrophic setting towards the end of the first build-up phase. The onset of surface-water mesotrophic conditions is coeval with the Trough Phase of OAE2 in Arobes with the top of *Rotalipora cushmanii* planktonic foraminiferal zone interval, which includes the peaks of the nannofossils *Biscutum constans* and *Prediscosphaera* spp. The diminishing of taxa related to warm water surface and the occurrence of those showing affinities with cooler water surfaces has probably no connection with the setting of an anoxic regime, being rather linked to a widespread rapid cooling of surface waters, in an interval of bottom-water reoxidation.

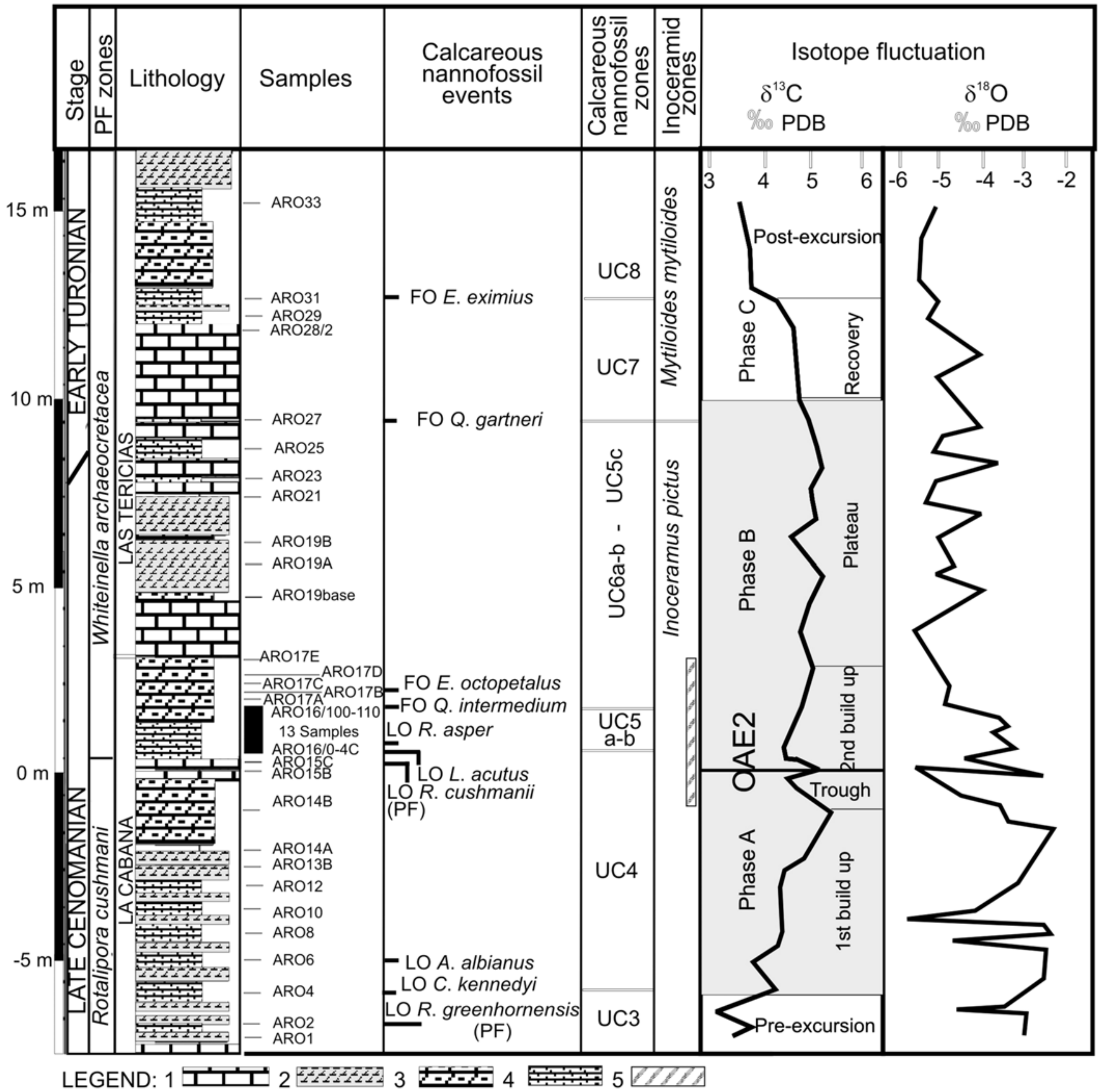

Fig. 9 – Fluctuation of $\delta^{13}C$ and biostratigraphy based on calcareous nannofossils of the Arobes section, N Spain (after Melinte-Dobrinescu et al., 2013).

In Arobes, the LO of *Rotalipora cushmani* took place slightly above the occurrence of cold water serpulid faunas, in an interval of increased $\delta^{18}O$ values and slightly below the peak interval of the nannofossil *Eprolithus floralis*. Therefore, this event is probably related to a cooling phase, which may be correlated with the Plenus Cold Event. The second build-up phase of $\delta^{13}C$ in Arobes contains most of the nannofloral peaks recorded in this section; these are: *Thoracosphaera* spp., followed by *Zeugrhabdotus erectus, Cyclagelosphaera margerelii, Prediscosphaera* spp. and *Eprolithus floralis*. A similar succession of events, except for the bloom in *Thoracosphaera* spp., was previously reported from other northern Spanish sections, spanning the same time interval (Paul et al. 1994), as well as from Italy (Gubbio section: Premoli Silva et al. 1999; Erba, 2004), UK (Dover, Eastbourne: Lamolda et al. 1994; Paul et al. 1999) and in the southern Carpathians (Melinte-Dobrinescu & Bojar 2008). In the later mentioned region, peaks of *E. floralis* are placed in the upper half part of the OAE2. Notably, in the Spanish sections as well as in the English ones, the peaks of *E. floralis* are between 30 and 40%, similar to the values identified in Arobes. The significant enrichment in *E. floralis* presumably indicates that cooler surface waters reached the Tethyan Realm, during the Cenomanian-Turonian boundary interval. This agrees with the global sea level curve, which was supposed to be very high at that time (Haq et al, 1987; Haq, 2014), leading to the appearance of mixed (Tethyan and Boreal) faunas and floras. In fact, *Hepteris septemsulcata*, a component of cold Pennrich faunas, co-occurs with maximum values *E. floralis, below* the LO of foraminifer *R. cushmani*. Furthermore, at Shakespeare Cliff, Dover, maximum percentages of *E. floralis* coincide with the occurrence of the cold macrofaunal events in equivalent levels (Lamolda et al., 1994). Although preservation increases its percentages, as a solution-resistant taxon, this fact does not invalidate the main observed changes. Probably, the second build-up phase is characterised at its beginning by eutrophic conditions that shifted to mesotrophic towards its end.

The Plateau Phase is characterised by minima in both nannofloral abundance and diversity, and by highest percentages of *W. barnesiae*. Therefore, the poor preservation recorded through the whole interval may be a significant cause for the absence of minor components, although it is not a barren zone. The Plateau Phase is characterised by minima in both nannofloral abundance and diversity, and by highest percentages of *Watznaueria barnesiae* (Fig. 10).

As this interval is barren of brachiopods (Melinte-Dobrinescu et al., 2013), probably this is the time when the highest anoxic conditions were established both on the sea floor and in the water-column during OAE2. A significant bioevent is the unusual high relative abundance of the calcareous dinoflagellate *Thoracosphaera* spp., at around 5%, from the base of the studied section. The relatively high percentages of *Thoracosphaera* spp. may indicate restricted or some critical environmental conditions prior to the main development. It implies a diversified calcareous nannofloral ecosystem with noticeable signs of instability. *Thoracosphaera* spp. increased more significantly at the end of the first build-up phase and peaked during the second build-up phase.

The nannofloral turnover, which took place globally across the CTBE, reflects the new palaeoecological conditions (i.e., fluctuations in pH, salinity, and dissolved CO_2 and O_2) of surface waters during OAE2. The phytoplankton record in the Arobes section supports a lower productivity during the OAE2, such as proposed for other localities in northern Spain and southern England. Both palaeoenvironmental and palaeogeographical settings significantly impacted the preservation and distribution pattern of various groups of organisms.

Correlation between the main taxa, such as *Watznaueria barnesiae, Erolithus floralis, Biscutum constans, Zeugrhabdotus erectus*, *Thoracosphaera* spp., *Cyclagelosphaera margerelii, Prediscosphaera* spp. and *Eiffellithus turriseiffelii*, indicates several inconsistencies, according to known proposed trophic behaviour. The extremely unstable ecosystem recorded from the LO of *Axopodorhabdus albianus* until the FO of *Quadrum gartneri* may explain these inconsistencies. Productivity seems to have increased for a short period preceding the critical turnover episode as the ecosystem quickly became starved, with blooms that were less significant, and mesotrophic and eutrophic nannofossils (e.g., *B. constans* and *Z. erectus*) which disappeared from the record. Non-calcareous dinocysts show the same trend but calcareous dinocysts, such as *Thoracosphaera* spp., do not show a record that could complement the missing primary producers.

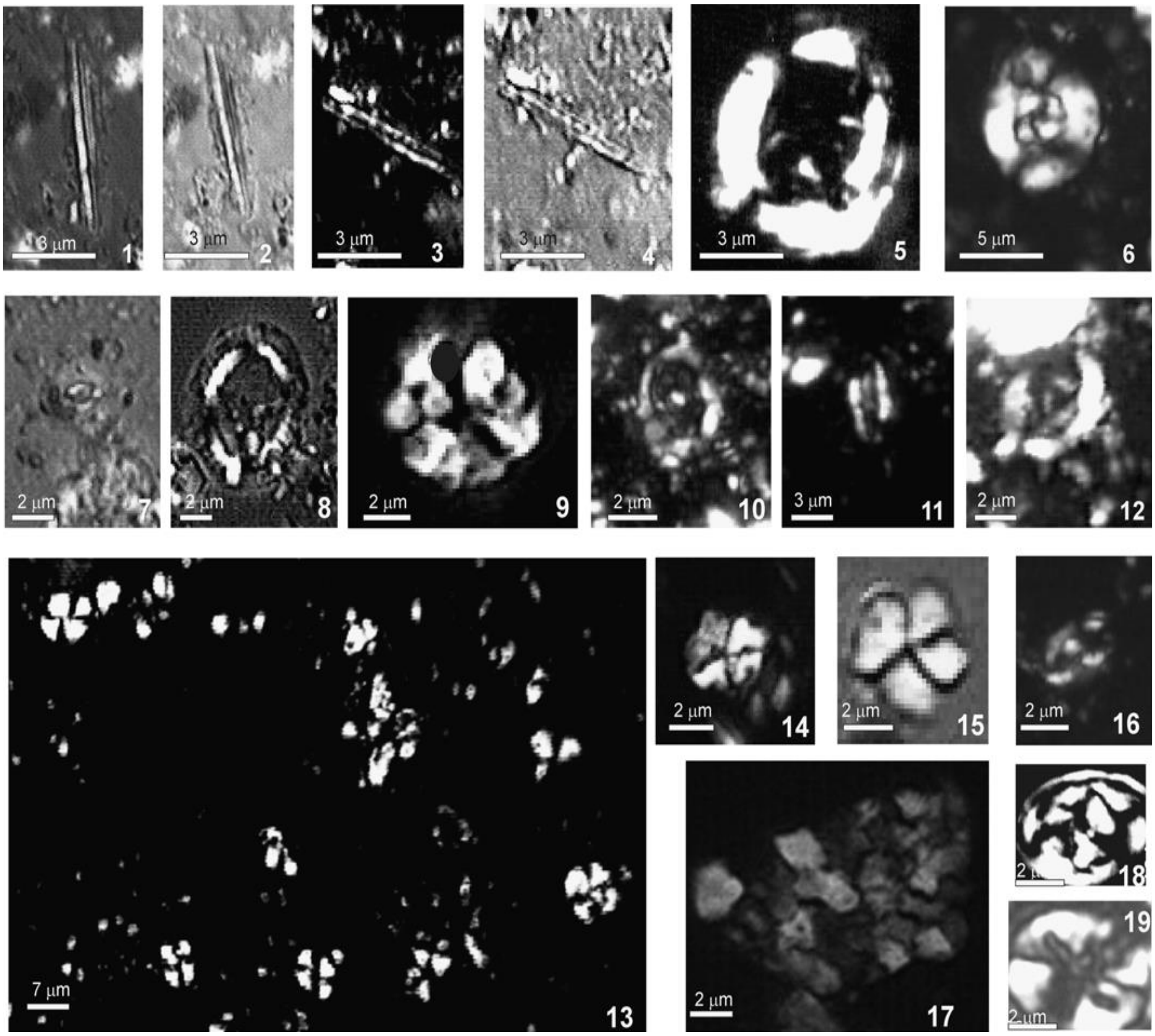

Fig. 10 – Calcareous nannofossils identified in the Spanish Arobes section. All microphotographs taken at LM (light microscope). Figures in crossed-nicols, except 2, 4, 7 and 15 in transmitted light. **1, 2**. *Lithraphidites acutus* Verbeek & Manivit in Manivit et al. 1977**; 3, 4.** *Lithraphidites carniolens* Deflandre 1963; **5.** *Axopodorhabdus albianus* (Black) 1971 Wind & Wise in Wise & Wind, 1977; **6**. *Helenea chiastia* Worsley 1971; **7.** *Biscutum constans* (Gorka 1957) Black in Black & Barnes 1959; **8.** *Manivitella pemmatoidea* (Deflandre in Manivit 1965) Thierstein 1971; **9.** *Eprolithus floralis* (Stradner 1962) Stover 1966; **10.** *Rhagodiscus asper* (Hill 1976) Doeven 1983. **11.** *Rhagodiscus angustus* (Stradner 1963) Reinhardt 1971; **12.** *Rhagodiscus splendens* (Deflandre 1953) Verbeek 1977; **13.** specimens of *Watznaueria barnesiae* (Black in Black & Barnes 1959) Perch-Nielsen 1968; **14.** *Quadrum intermedium* Varol 1992; **15.** *Quadrum gartneri* Prins & Perch-Nielsen in Manivit et al. 1977; **16**. *Zeugrhabdotus erectus* (Deflandre in Deflandre & Fert 1954) Reinhardt 1965; **17.** *Thoracosphaera* sp.; **18.** *Zeugrhabdotus embergeri* (Noël 1958) Perch-Nielsen 1984; **19.** *Eiffellithus eximius* (Stover 1966) Perch-Nielsen1968.

MID CRETACEOUS OCEANIC ANOXIC EVENTS IN ROMANIA

The Oceanic Anoxic Event (OAE)1a in the Apuseni Mts.

The only place in Romania where the OAE1a could be identified was in the NW part of Apuseni Mts., respectively in Pădurea Craiului, an area which belongs to an Upper Jurassic- Lower Cretaceous carbonate platform (Papp et al., 2012). The OAE1a was observed in the Ecleja Formation (Papp & Cociuba, 2012, 2013). Besides, the aforementioned authors have identified, based on a succession of positive and negative shifts of $\delta^{13}C$ (Fig. 11), all the segments C of Menegatti et al. (1998) covering the Barremian-upper Aptian *pro parte*.

The Oceanic Anoxic Event (OAE) 1b, 1c and 1d in the Apuseni Mts.

In the same area situated in NW Romania (Pădurea Craiului of Apuseni Mts.) as above-mentioned, Papp et al. (2012) and Papp & Cociuba (2012) reported the presence of the OAE1b. In the Vârciorog Formation, a significant negative excursion of the $\delta^{13}C$ isotope was recorded. The calcareous algae assemblages of this interval indicate a late Aptian age (Bucur et al., 2010). This event was correlated with OAE1b, as described in the black shale succession of the Vocontian Basin (Bréhéret, 1988).

The upper Albian global anoxic events OAE1c and OAE1d were pointed out by Papp et al. (2012), (2013) and Papp & Cociuba (2012) also in the NW region of the Apuseni Mts. (Pădurea Craiului). In this unit, from the Red Detrital Formation, significant positive and negative excursions of the isotope $\delta^{13}C$ isotope were reported (Fig. 11). In the lower part of the unit, the aforementioned authors observed a positive excursion of $\delta^{13}C$, followed by a decrease and then by another significant increase.

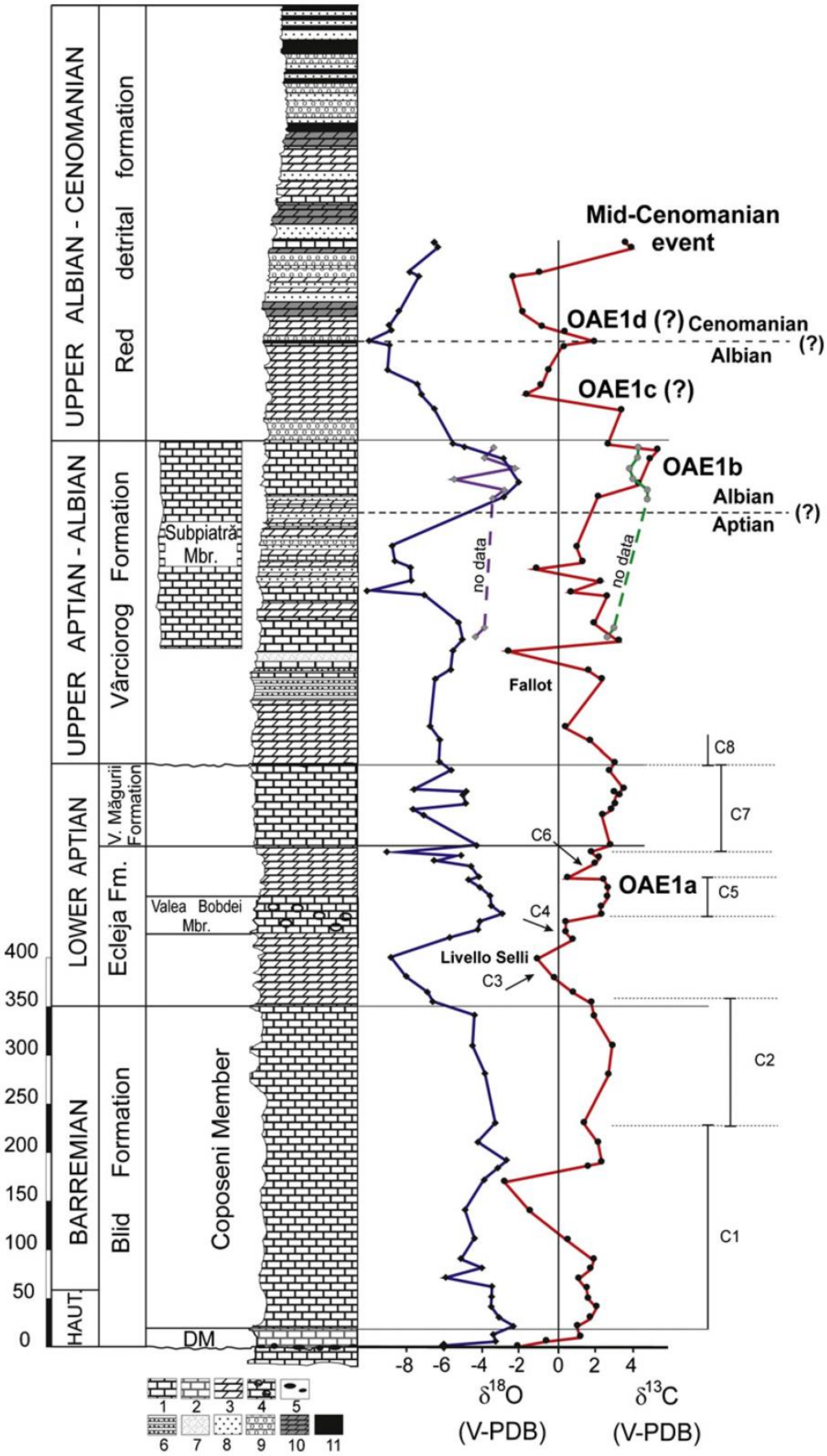

Fig. 11 – Mid Cretaceous Oceanic Anoxic Events observed in NW Romania, Apuseni Mts. in the succession of Pădurea Craiului. Legend: 1 – limestone; 2 – fresh water limestone; 3 – marls; 4 – limestone with cherts; 5 – bauxite; 6 – glauconitic sandstone; 7 – carbonaceous sandstone; 8 – sandstone; 9 – conglomerate; 10 – marly shale; 11 – shale (from Papp et al., 2013).

The Oceanic Anoxic Event (OAE) 1d in the Eastern Carpathians

The OAE1d is the only OAE identified so far in the Eastern Carpathians based on geochemical ($\delta^{13}C$ isotope) fluctuations, accompanied by biostratigraphical ones (Melinte-Dobrinescu et al., 2015). The succession comprising OAE1d is situated in the Cernatu locality (Covasna County), belonging to the Inner Moldavide Nappe System (Săndulescu, 1984), respectively the Teleajen Nappe.

The lower part of the succession containing the OAE1d is composed of grey to blackish and green shales, followed by red shales (Fig. 12). A positive $\delta^{13}C_{org}$ excursion that characterises the OAE1d was observed in the lower part of the Cernatu Valley succession, extending about 2 m in dark-grey and green shales (Fig. 6). The $\delta^{13}C_{org}$ values exhibited an increase of 1.3‰, up to 24.01‰. This positive excursion is situated in the upper Albian, within the occurrence interval of the ammonite *Stoliczkaia notha*. From a nannofossil point of view, the OAE1d in the section falls in the NC10a (=CC9b) calcareous nannofossil subzone (Melinte-Dobrinescu et al., 2015), as globally reported (Bralower et al., 1993; Watkins et al., 2005). Based on planktonic foraminiferal data, the OAE1d is situated within the Parathalmanninella appenninica zone (Coccioni, 2001).

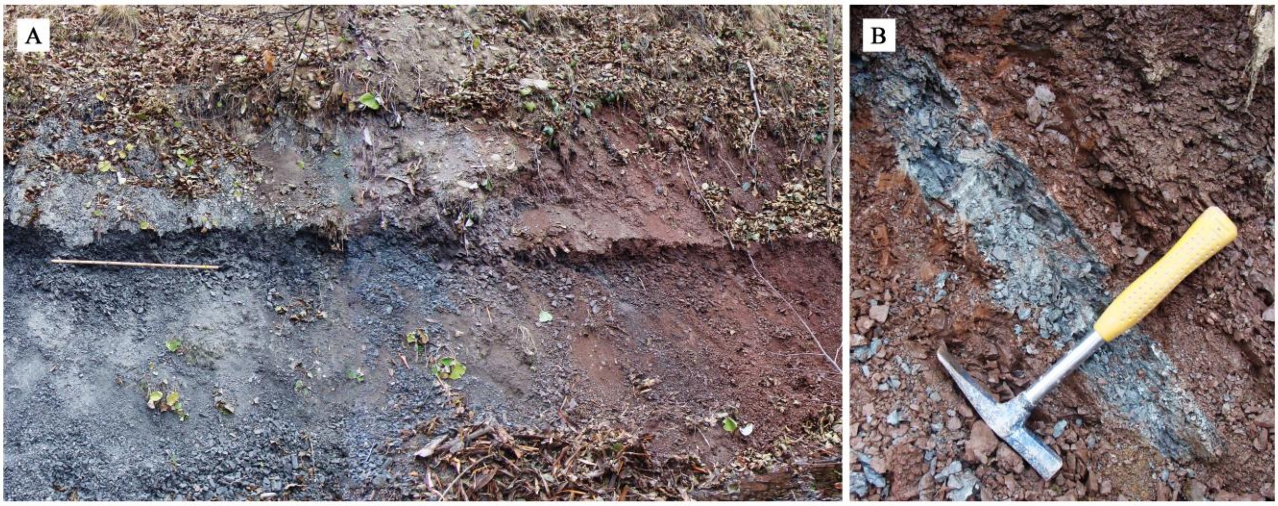

Fig. 12 – Succession of grey-blackish shales (A) followed by red shales (B) in the Cernatu Valley (central western part of the Eastern Carpathians, Teleajen Nappe of the Inner Moldavides).

The positive $\delta^{13}C_{org}$ shift, along with higher values of OM (organic matter) indicates a brief period of increased carbon burial and dysoxic to anoxic bottom water. This hypothesis is supported not only by geochemical fluctuations, but also by biotic features such as a reduction in agglutinated foraminiferal diversity in general, and especially of epifaunal taxa related to the decrease in the oxygen content, while the calcareous nannofossil assemblages are very scarce in diversity and abundance, showing strong dissolution processes (Melinte-Dobrinescu et al., 2015).

The changes in lithology, together with the recovery of macrofaunas towards the upper part of the section, indicate the replacement of the latest Albian anoxic/dysoxic environment with an oxic one within the Albian-Cenomanian boundary interval. This change is possibly related to climate fluctuations.

Therefore, a warm climate of those times could imply the accumulation of red soils on emerged coastal plains, while transgressions could have led to the redeposition of sediments rich in Fe-hydroxides in the marine environment, generating the red beds that are now preserved. The continental origin of the iron is supported by the positive correlation of Fe_2O_3 with Al_2O_3, SiO_2 and TiO_2.

This palaeoenvironmental fluctuation may be also related to the mid Cretaceous tectonic movements that strongly affected the whole Eastern Carpathian belt (Săndulescu, 1984). As a result, the restricted circulation of small silled basins of the Moldavian Trough shifted to an open circulation regime (Ştefănescu & Melinte, 1996; Roban et al., 2017).

Bornemann et al. (2005) studied in detail the OAE1d in sections from the Tethyan and Boreal realms and identified four successive peaks. The oldest one, late Albian in age, situated in the Stoliczkaia dispar ammonite zone, was assigned to the level of “Niveau Breistroffer”. The next successive two, b and c are also late Albian, while the youngest one, is earliest Cenomanian in age, placed in the Mantelliceras mantelli ammonite zone (Fig. 13).

In the section from the Eastern Carpathians, the Cernatu succession displays two distinct peaks of $\delta^{13}C_{org}$. The oldest one, more pronounced and stratigraphically more expanded, is interpreted as the OAE1d Breistroffer Events, being placed in the Stoliczkaia dispar ammonite zone (Ion & Szasz, 1994). The youngest one, probably could be assigned to the final pulse of OAE1d, earliest Cenomanian in age (Fig. 13).

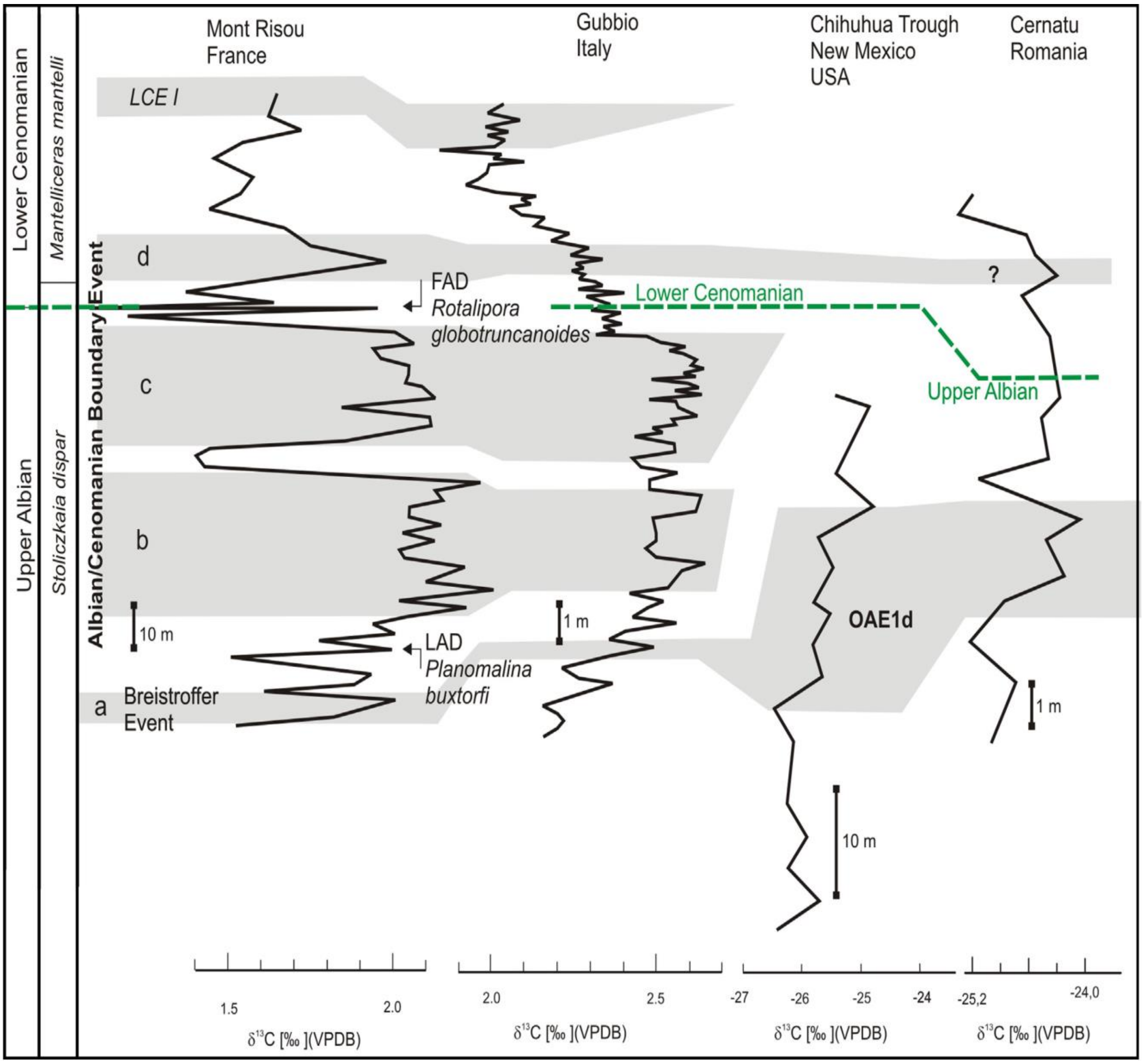

Fig. 13 – Correlation between the OAE1d identified in the Eastern Carpathians, Cernatu section (after Melinte-Dobrinescu et al., 2015) and other Tethyan sections: Mont Risou (France) and Gubbio (Italy) from Kennedy et al. (2004), Bornemann et al. (2005) and Gale et al. (2011); Chihuhua Trough, New Mexico from Scott et al. (2013).

In the northern part of the Eastern Carpathians, there are outcrops of mid Cretaceous successions, especially in the Inner Moldavide structures, such as the Audia Nappe (Săndulescu, 1984; Ştefănescu & Melinte, 1996; Roban et al., 2017). A good exposure is situated in the Ostra Valley (nearby the Stulpicani locality), showing rhythmical alternating calcareous sandstones, grey-greenish claystones and marlstones of the Upper Member of the Audia Formation. Several cm- up to dm-thick black shale levels could be observed (Fig. 14)

Fig. 14 – Albian successions in the Ostra Valley (N Eastern Carpathians), containing cm- up to dm-thick black shale levels.

Similar successions are visible in the Audia Nappe cropping out in the Bistrița Valley (Fig. 15), where the Upper Member of the Audia Formation comprises several cm-thick black and dark-grey shales in turbidite successions.

Fig. 15 – The Upper Member of the Audia Formation exposed in the Bistrița Valley, comprising cm-thick black shale levels.

The identified calcareous nannofossil assemblages of the Audia Formation Upper Member contain, besides long-ranging taxa, significant biostratigraphic species, such as: *Eprolithus floralis*, *Prediscosphaera columnata*, *Axopodorhabdus albianus*, *Eiffellithus turriseifellii and Tranolithus orionatus*. These occurrences indicate an Albian age of the Upper Member included in the Audia unit.

The OAE2 in the Southern Carpathians

The OAE2 was described in the Haţeg Basin (Southern Carpathians). This area is world-wide famous for its rich continental macrofaunas, and particularly for its dinosaur remains (Grigorescu, 1983; Grigorescu et al., 1999). Besides, in the NW and SE parts of the Haţeg Basin, a marine Cretaceous sedimentation occurs, showing fluctuations of the sea-level in this region, from the shallow shelf up to the deep-marine palaeoenvironment (Pop, 1990; Grigorescu & Melinte, 2001; Melinte-Dobrinescu, 2010).

The succession with the OAE2 is exposed in the Ohaba-Ponor village and contains Albian up to Coniacian sediments (Fig. 16). The Albian succession are fresh to brackish water ones (Antonescu et al., 1983), and are followed by shallow-water deposits, with rudists (Lupu, 1966).

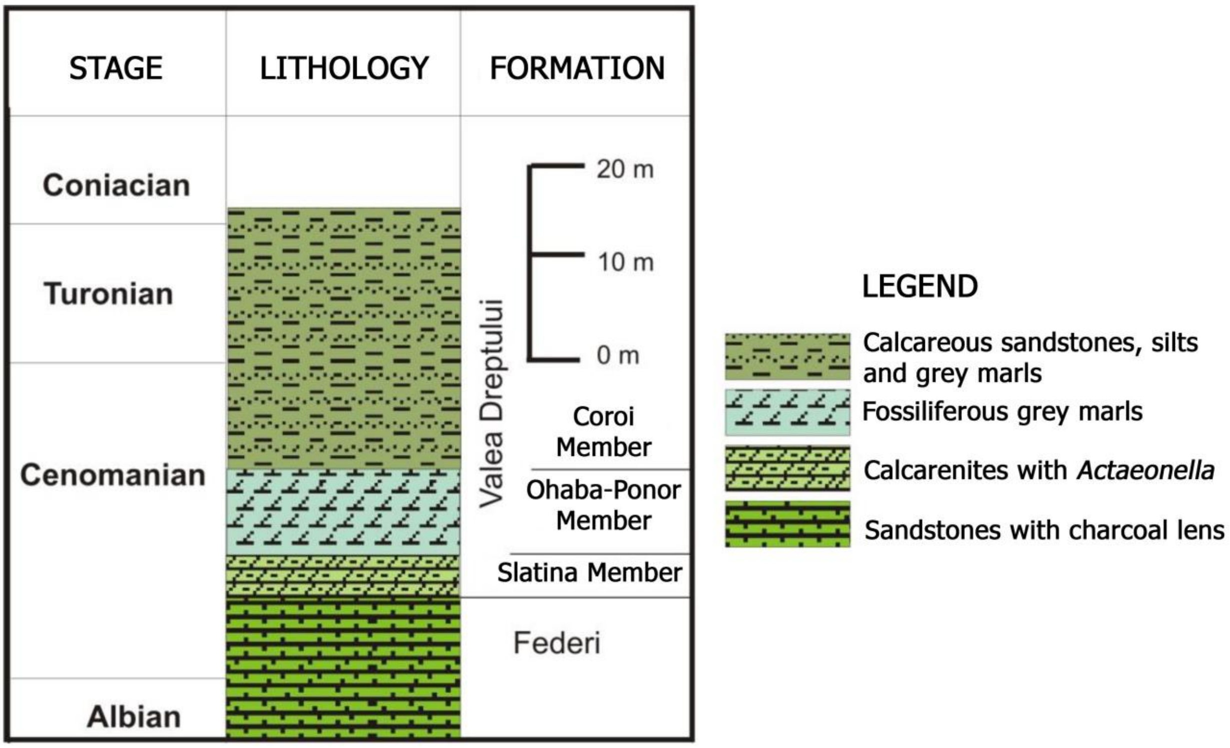

Fig. 16 – Lithostratigraphy of the Albian-Coniacian exposed in the Ohaba-Ponor section.

The section contains rich macrofauna, such as rudists, inoceramids and ammonites (Fig. 17), especially in the Cenomanian-early Turonian interval. In the upper part of the exposed succession, late Turonian-Coniacian in age, the macrofauna is very rare.

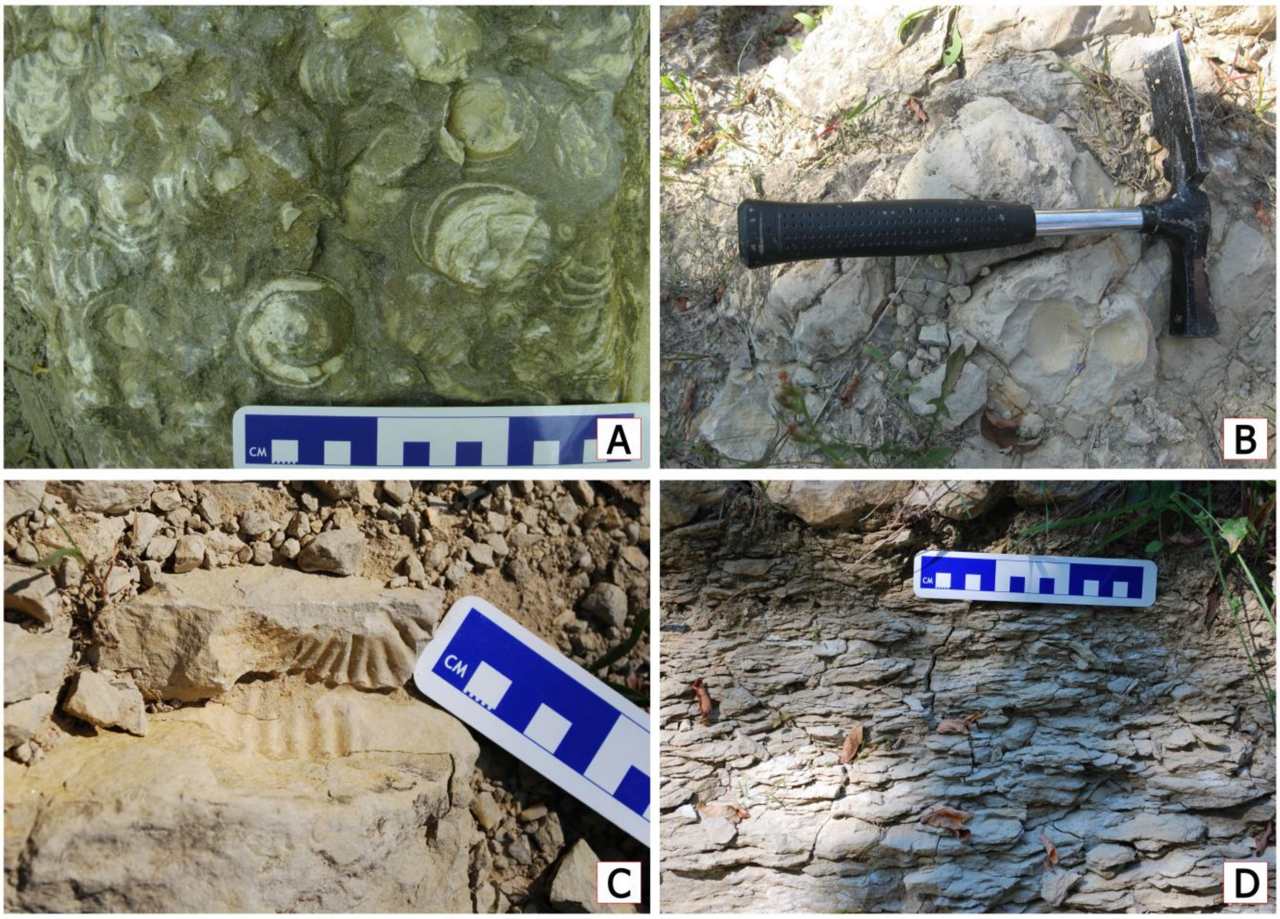

Fig. 17 – Macrofauna encountered in the Cenomanian interval, Valea Dreptului Formation. A - Slatina Member; B and C - Ohaba-Ponor Member; D - Coroi Member.

The OAE2 was identified in the depositional interval covered by the Valea Dreptului Formation (Melinte-Dobrinescu & Bojar, 2008). The lower part of the studied section contains nannofloral assemblages enclosing significant biostratigraphic nannofossils, such as: *Lithraphidites acutus, Eiffellithus turriseiffelii, Corollithion kennedyi* and *Microrhabdulus decoratus*, which refers to the NC11 Zone. Towards the lower part of the upper Cenomanian Pentagonum Ammonite Zone (Szasz, 1976), FO of the nannofossil *Ahmuellerella octoradiata* was observed. This bioevent is followed by the successive last occurrences (LO) of *Corollithion kennedyi*, *Axopodorhabdus albianus* and *Lithraphidites* *acutus*, the last event defining the NC11/NC12 boundary. Within the NC12 nannofossil zone, the successive FOs of *Quadrum intermedium* and *Eprolithus octopetalus* were identified (Fig. 18). All the above-mentioned nannofloral events were observed in the Ohaba-Ponor Member. The FO of the nannofossil *Quadrum gartneri* (the base of the NC13 Calcareous Nannofloral Zone) was observed towards the lower part of the Coroi Member, just below the FO of the inoceramid *Mytiloides labiatus*.

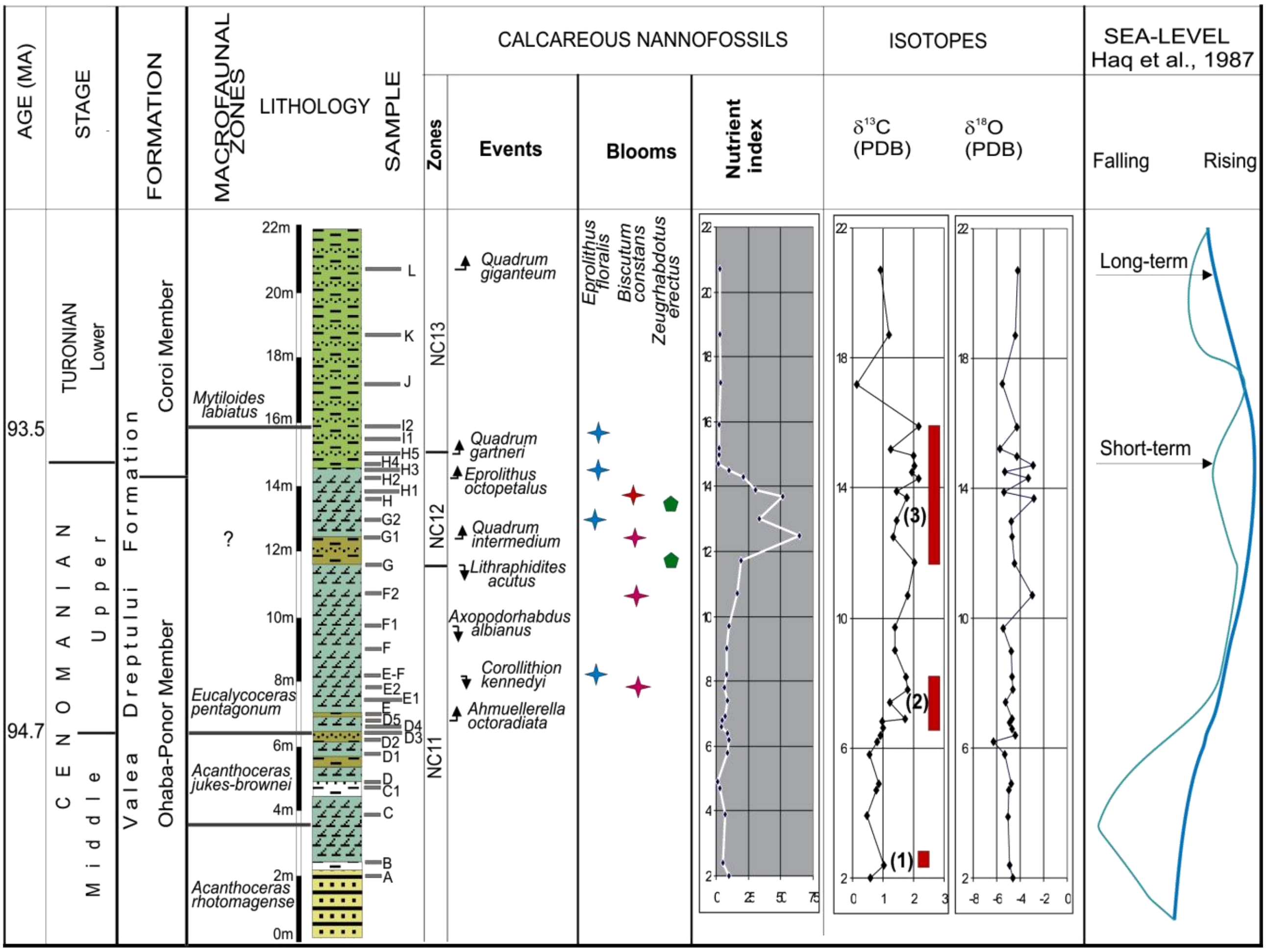

Fig. 18 – The OAE2 geochemical and biostratigraphical records in the Southern Carpathians (Haţeg area) – after Melinte-Dobrinescu & Bojar, 2008.

The $\delta^{13}C$ and $\delta^{18}O$ fluctuations (Fig. 18) identified in the Cenomanian-Turonian boundary interval (Melinte-Dobrinescu & Bojar, 2008) are the following: (1) a positive excursion up to 1.8 ‰ (PDB) within the middle/late Cenomanian boundary, towards the lower part of the Coroi Member, just below the FO of the inoceramid *Mytiloides labiatus* Ammonite Zone boundary interval. This chemostratigraphic event was assigned to the

Middle Cenomanian Event II (MCEII), as described by Jarvis et al. (2006) in sections from Tethys and Boreal realms; (2) An excursion up to 2.2 ‰ (PDB) in the Cenomanian/Turonian boundary interval, within the boundary interval of the ammonite zones Acanthoceras jukes-brownei and Eucalycoceras pentagonum (Pop, 1990; Ion & Szasz, 1994). The shift of the $\delta^{13}C$ is coeval with the OAE2 observed elsewhere, of approximately 2‰ ($\delta^{13}C$ enrichment), which occurred from the late Cenomanian to the earliest Turonian interval (Arthur et al., 1988). The successive peaks of the nannofossils *Biscutum constans*, *Zeugrhabdotus erectus* and *Eprolithus floralis* indicate episodes of cooler surface water and high fertility, which preceded and postponed the Cenomanian/Turonian boundary event. Additionally, some higher values of $\delta^{18}O$ values in between OAE2 and MCEII indicate rapid climate modifications during the early Cenomanian.

The OAE2 in the Eastern Carpathians, southern part

The OAE2 was identified by Cetean et al. (2008), based on the lithological/sedimentological investigations and micropalaeontological analysis, in an Albian-lower Turonian succession of the Stoeneşti - Cetăţeni area (Dâmboviţa Valley), where the Upper Cretaceous sediments are included in the post-tectonic cover of the Outer Dacides (Săndulescu, 1984). The Cenomanian-Turonian boundary interval is made by black shales, followed in the Turonian by variegated (red, green and grey) shales.

Cetean et al. (2008) identified planktonic foraminifers that were assigned to the late Albian-early Turonian, starting with Rotalipora appenninica biozone up to Whiteinella archaeocretacea Zone (Early Turonian). They indicate that, linked to the poor preservation, Rotalipora (= Thalmanninella) reicheli biozone was not been observed.

According to Cetean et al. (2008), the upper Cenomanian sediments including the black shale level lack the nannofossils. Calcareous nannoplankton assemblages were reported from lower Turonian sediments, including, besides long-ranging taxa, the following nannofossils: *Quadrum intermedium, Eprolithus octopetalus, Quadrum gartneri* and *Microrhabdulus decoratus*.

The aforementioned authors identified also rich and diversified agglutinated foraminiferal assemblages, except the supposed level of OAE2 of the Cenomanian-Turonian boundary interval, mainly composed of *Glomospira charoides* and *Haplophragmoides* spp., taxa indicative for an oxygen-depleted palaeoenvironment. An indicator of an anoxic palaeosetting is also the presence of a rich radiolarian-rich level devoid of benthic foraminifera, described by Cetean et al. (2008) from the upper Cenomanian Rotalipora cushmani foraminiferal zone. Above the Cenomanian/Turonian boundary event, the first stage of the recolonization is mirrored by the occurrence of small, thin-walled agglutinated foraminifers.

The OAE2 in the Eastern Carpathians, central part

In the Eastern Carpathians, anoxic depositional intervals are known to occur from the Valanginian up to Cenomanian-Turonian boundary interval. These black shales do not represent a particular Oceanic Anoxic Event, but they reflect a specific palaeosetting of those times. The black shale deposition took place in a very deep environment, on the abyssal plain of the Moldavian Trough (Ştefănescu & Melinte, 1996; Melinte-Dobrinescu & Roban, 2011), near the calcite compensation depth (CCD). Hence, the occurrence of the black shales in the Eastern Carpathians is linked to the existence of a narrow basin, formed on a thinned continental crust of the passive northern margin of western Tethys Realm (Săndulescu, 1984). There, the anoxic to dysoxic bottom waters led to the deposition of organic-rich black shales that accumulated at a low rate, from 2 up to 5 cm/Ky (Ştefănescu & Melinte, 1996).

The Lower Cretaceous setting of the Eastern Carpathians is similar with the Silesian Nappe of the Outer Polish Carpathians. There, a deep, long trough, restricted towards N by the Subsilesian Submerged Ridge and towards S by the Silesian Ridge was settled during Jurassic-Early Cretaceous times. In this trough, the Lower Cretaceous sediments were also deposited around the CCD, being represented by black shales (Ślączka, 1993) and followed by CORBs (Cretaceous Oceanic Red Bes) around the Albian-Cenomanian boundary interval (Bąk., 2006). A similar change in deposition is known from the Alpine area (Northern Calcareous Alps, Wagreich & Krenmayr, 2005), and the Moldavide nappes of the Eastern Carpathians (Melinte-Dobrinescu et al., 2009; Roban & Melinte-Dobrinescu, 2012).

However, despite the large development of the anoxic deposits due to the restricted circulation in the Eastern Carpathian basin, it is possible to recognize some OAEs. Good examples are the Albian-Turonian successions cropping out in the Vrancea Halfwindow (Vrancea Nappe, Outer Moldavides). There, in the Putna and Tişița valleys, the Tisaru Formation is made by Cenomanian-Turonian grey and green shales (Fig. 19), followed by red shales (Melinte-Dobrinescu et al., 2009; Roban et al., 2017)

Fig. 19 – The Tisaru Formation exposed on the left bank of the Putna Valley at Lepșa.

In the Putna Valley, the Tisaru Formation contains a black shale level, cm-thick (Fig. 20). Similarly, black shale levels were encountered in a similar stratigraphic position, being exposed in the nearby Tişița River, upstream the confluence with the Putna River.

There, the calcareous nannofossil assemblages contain, besides long-ranging taxa, significant biostratigraphic species. Several bioevents, such as the successive FO of *Quadrum intermedium*, *Eprolithus octopetalus* and *Quadrum gartneri*, prove the Cenomanian-Turonian boundary interval. The black shale level is barren of calcareous nannoplankton.

The presence of chondritides (e.g., *Chondrites hamatus,* Brustur 2005) indicates a very low oxygen content in the interstitial water of the sediments. Thus, chondritides are considered “index” trace fossils, typical for anoxic conditions.

This finding indicates a possible occurrence of the OAE2 in the central part of the Eastern Carpathians, in the Outer Moldavide structures. This hypothesis needs to be confirmed by the completion of geochemical studies, including fluctuation of $\delta^{13}C$ isotope.

Fig. 20 – The Tisaru Formation (Vrancea Nappe) in the Putna Valley, containing a cm-thick black shale level within the Cenomanian-Turonian boundary interval.

The OAE3 (= The Santonian-Campanian boundary event)

The Santonian-Campanian boundary is globally characterised by a positive $\delta^{13}C$ isotope excursion of 0.3 ‰ up to 2.9 ‰, placed within the Marsupites testudinarius crinoid zone, referred as the Santonian-Campanian Boundary Event (Jarvis et al., 2002). The above-mentioned event was previously reported from many European sections (i.e., N Germany - Schönfeld et al., 1991; East Kent and Sussex - Jenkyns et al., 1994; S France - Jarvis et al., 2002), as well as from Africa (Tunisia - Jarvis et al., 2002), America (US Western Interior - Pratt et al., 1993) and Asia (Li et al., 2006).

Recently, Gale et al. (2008) published an integrated study (geochemistry, stable oxygen and carbon isotopes, as well macro-, micro- and nannofossil events) of the Waxahachie Dam Spillway section, north Texas, a possible boundary stratotype for the base of the Campanian Stage. There, the positive $\delta^{13}C$ isotope excursion starts above the FO of the crinoid *Marsupites testudinarius* and ends above the LO of this species (the bio-event which marks the Santonian - Campanian boundary). Accordingly, the Santonian - Campanian Boundary Event falls within the UC13 calcareous nannofossil zone, situated between the FO of *Arkhangelskiella cymbiformis* and the FO of *Broinsonia parca parca*.

The OAE3 in the Southern Carpathians

For the identification of OAE3, a succession composed of variegated, i.e., red and green hemipelagites (marlstones and claystones) of the Fizeşti Formation, situated in the SE part of the Haţeg basin (Fig. 21) was studied (Melinte-Dobrinescu & Bojar, 2010).

In the studied section, the FO of the ammonite genus *Texanites* was recorded, coincident with the occurrence in the planktonic foraminifer assemblages of the species *Dicarinella asymmetrica*, both bio-events being placed in the Romanian Carpathians within the Coniacian - Santonian boundary interval (Ion & Szasz, 1994; Ion et al., 2000).

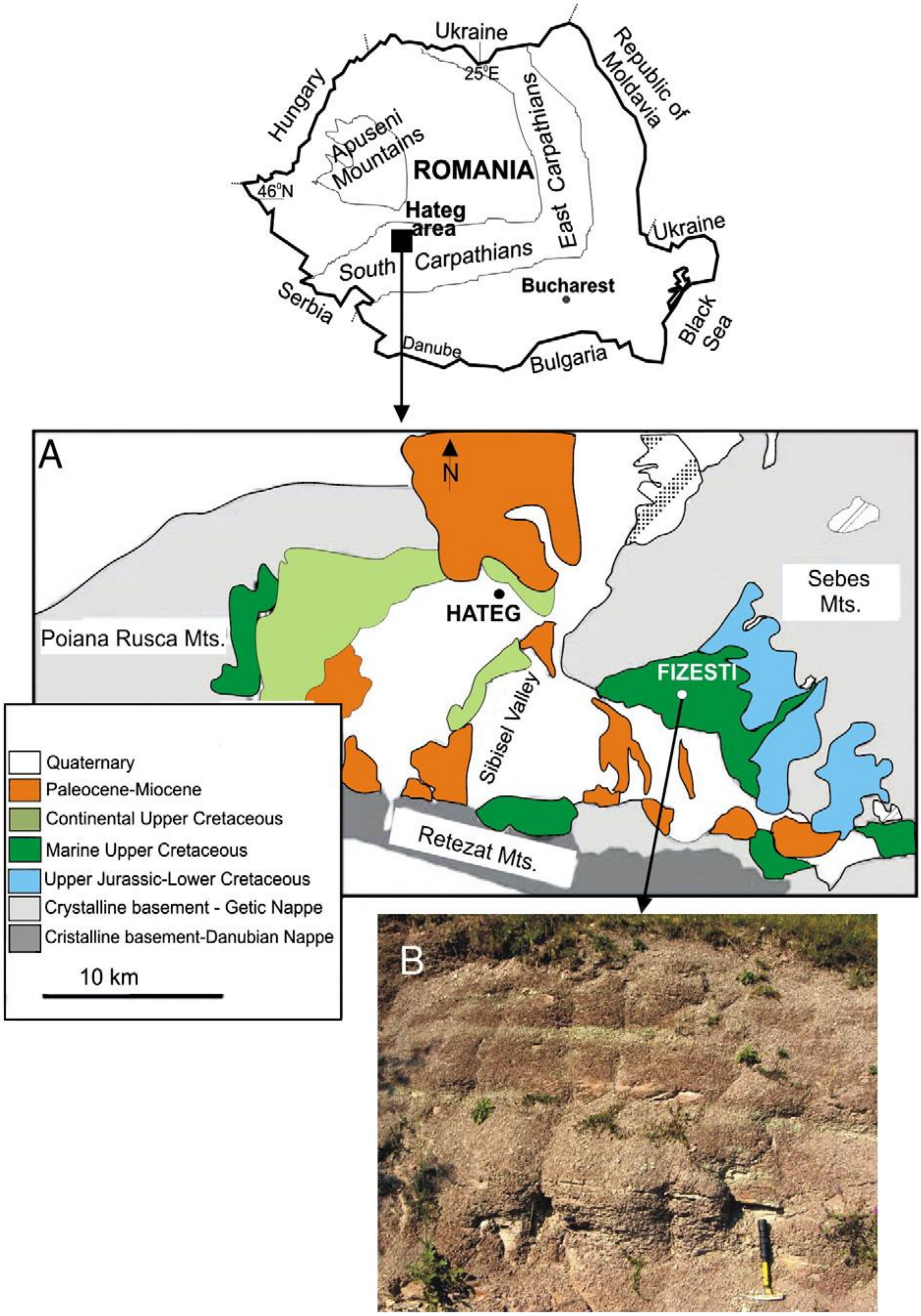

Fig. 21 – Location of the studied section for the OAE3 identification in the SE part of the Hațeg basin (after Melinte-Dobrinescu & Bojar, 2010).

Above the Coniacian - Santonian boundary interval, the succession of the nannofossil events, such as the LO of *Lithastrinus septenarius*, followed by the LO of *Eprolithus floralis*, and by the successive FOs of *Arkhangelskiella cymbiformis, Broinsonia parca parca, Bukryaster hayi, Ceratolithoides verbeekii* and *Ceratolithoides aculeus,* indicates the presence of the calcareous nannoplankton zones (including the subzones) CC16–17, CC18, CC19 and CC20 of Sissingh, 1977, and respectively the zones and subzones UC11c up to UC15b of Burnett, 1998. These biozones cover the Lower Santonian - Lower Campanian interval (Burnett, 1998; Gradstein et al., 2004).

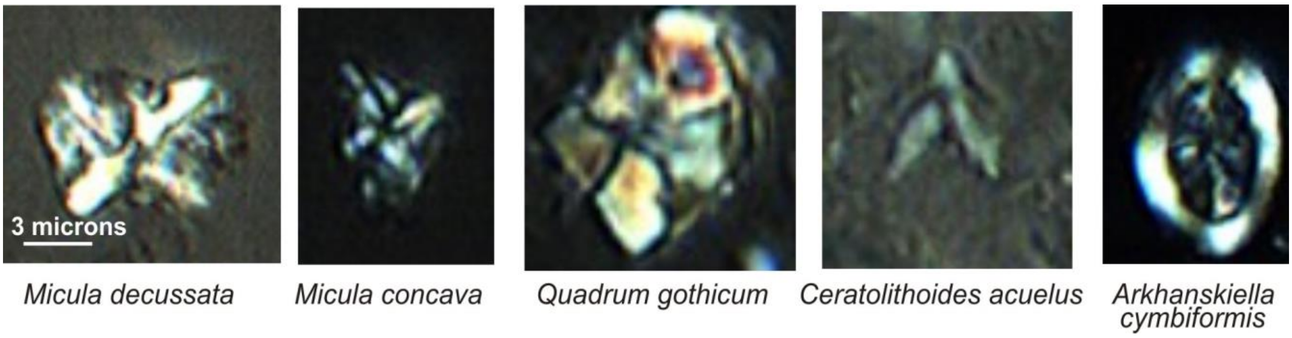

Fig. 22 – Calcareous nannofossil species identified in the Fizeşti Formation, SE Haţeg basin.

The Santonian Campanian boundary is marked in the studied section by the LO of the crinoid *Marsupites testudinarius* (Pop, 1990; Ion & Szasz, 1994; Ion et al., 2000), the primary marker of the above-mentioned boundary (Hancock & Gale, 1996). In the studied Romanian section, an increase of the $\delta^{13}C$ values (from 2.38 ‰ up to 2.58 ‰) was observed above the FO of the nannofossil *Arkhangelskiella cymbiformis*, within the lower part of the UC13 biozone. In the studied sections, two distinct peaks were recorded within the Santonian-Campanian boundary interval (Fig. 23):

(i) The oldest placed just below the FO of the crinoid *Marsupites testudinarius*, in the uppermost Santonian; this event could be correlated with the Hawks Brow Event of Jarvis et al. (2006), placed at the base of the Marsupites testudinarius crinoid zone;

(ii) The youngest placed across the Santonian/Campanian boundary, with the maximum values coincident with the LO of *Marsupites testudinarius*, which represents the globally distributed Santonian/ Campanian Boundary Event of Jarvis et al. (2002, 2006) occurring also in the investigated area.

The level containing the LO of the crinoid *Marsupites testudinarius* yielded the maximum positive excursion of 2.65‰, which is also the maximum recorded in the whole studied succession. Just above the LO of *Marsupites testudinarius*, in the Lower Campanian, the $\delta^{13}C$ values consistently drop down to 2.34‰. The Santonian-Campanian Boundary Event yields, in the Romanian section, a sudden decrease, whereas the same chemostratigraphic event shows a longer-term decrease during the Early Campanian in the English Chalk (Jarvis et al., 2002, 2006).

To explain this difference, Melinte-Dobrinescu & Bojar (2010) addressed a question about the existence of a sedimentary gap. However, both field work observations and the calcareous nannofossils biostratigraphy indicate a continuous sedimentation in the Fizeşti section, which is represented mainly by hemipelagic deposits. Possibly, the Romanian section is condensed, as in the Trunch section (E England) where the Santonian – Campanian Boundary Event extends at around 10 m (Jarvis et al., 2006), while in the Fizeşti section this event covers up to 2 m.

The shape of the $\delta^{13}C$ curve across the Santonian-Campanian Boundary Event in Romania is very similar to the one recorded by Gale et al. (2008) in Texas (USA), differing only by the magnitude of the carbon isotope positive excursion (Fig. 23). This positive excursion is higher in the Fizeşti section than in the Waxahachie Dam Spillway section.

The pattern of $\delta^{18}O$ variation within the Santonian-Campanian boundary interval is also similar for the two above-mentioned sections. In the uppermost Santonian (just below the LO of the crinoid *Marsupites testudinarius*), a shift of $\delta^{18}O$ to fewer negative values, an event followed by a slight and continuous decrease above the Santonian - Campanian boundary, was pointed out. A similar fluctuation pattern of $\delta^{18}O$ was observed (Gale et al., 2008) in Waxahachie Dam Spillway section (Texas, USA), at the same interval.

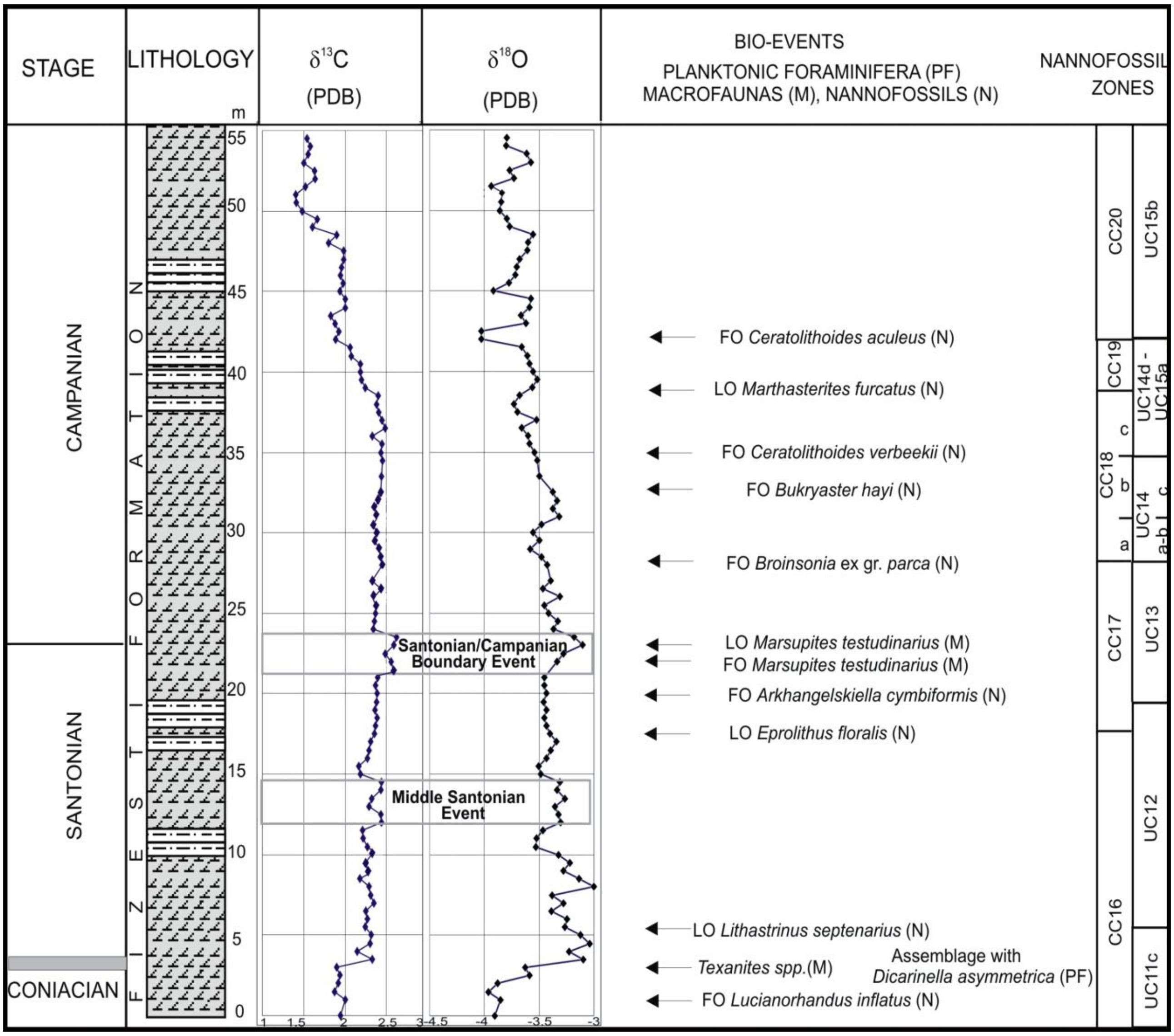

Fig. 23 – Correlation between the $\delta^{13}C$ and $\delta^{18}O$ fluctuations and the main bio-events recorded in the Fizeşti section (after Melinte-Dobrinescu & Bojar, 2010). Planktonic foraminifers and macrofaunas after Pop et al. (1990) and Ion & Szasz (1994); M - Macrofauna; PF - planktonic foraminifera; N - calcareous nannofossil. Sea-level fluctuation after Miller et al. (2005).

Palaeoenvironmental changes related to Oceanic Anoxic Events

Even though a large amount of data was collected and interpreted on OAEs issues since 1976, when they were first discovered, there are still regions which are not yet investigated. It is also possible that some Cretaceous OAEs have not been yet discovered.

The response of the marine phytoplankton in general, and of the calcareous nannoplankton group of organisms in special to the high concentrations of CO_2 amount release in atmosphere and dissolved in the Planetary Ocean during the global anoxic intervals is not yet fully understood. Some fundamental questions still remain:

(i) Why some OAEs induce no modification in the phytoplankton world, while other produce extinction, even though the events are comparable in terms of spatial and temporal amplitude?

(ii) Why is the mass extinction of the Permian-Triassic boundary associated with a LIP and OAE, while the Cretaceous-Palaeogene boundary mass extinction occurs apparently in the absence of any OAE?

(iii) How is the Earth atmosphere and ocean geochemistry restored after long green-house intervals and climate deterioration, lasting over 1,000 Ky?

(iv) Last but not least, will the Earth be able to simply restore back after the contemporaneous green-house climatic period or the Planet will enter in a permanent El Niño?

A detailed and world-wide scientific investigation could give some realistic answers to past, present and future climate deterioration and predict the way our planet will evolve.

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