Palaeoenvironmental changes across the Albian-Cenomanian boundary interval of the Eastern Carpathians

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A B S T R A C T
The studied Cernatu Valley section is situated in the central part of the Eastern Carpathians (Romania) and spans the interval covered by the NC10a (=CC9b) calcareous nannofossil Subzone, as well as the Plectorecurvoides alternans and Haplophragmoides falcatus turulais agglutinated foraminiferal zones. The presence, within the lower part of the section, of the ammonite Stoliczkaia notha indicates a late Albian age, but the section possibly extends to within the Albian-Cenomanian boundary interval, based on the agglutinated foraminiferal assemblages. The deposits are grey to black and green shales that are followed by red shales interbedded with couplets of grey to blackish and green shales. The benthic foraminifers suggest a deep-marine depositional setting, probably lower bathyal or even abyssal, at around 2500 m depth. The deposition was probably near but above the Calcium Compensation Depth (CCD), as very scarce nannofloras and macrofaunas are present. The δ13Corg values vary throughout the section between −25.30‰ and −24.01‰. Within the upper Albian, a positive organic δ13Corg excursion with increases of 1.3‰, up to −24.01‰, is recorded. This positive excursion has been tentatively interpreted as a regional expression of the Oceanic Anoxic Event OAE1d in the Moldavian Trough of the Eastern Carpathian basin. The upper part of the section, belonging to the Haplophragmoides falcatus turulais agglutinated foraminiferal Zone, contains a weak positive δ13Corg excursion marked by an increase in values of about 0.5‰, which is assumed to represent late phases of the Albian-Cenomanian boundary Event. Towards the top of the section, consisting mainly of red shales, calcareous foraminifera also occur, together with more consistent nannofossil assemblages. This biotic change possibly mirrored an alteration of the palaeoenvironment, which shifted from an anoxic/dysoxic setting towards an oxic one. This change is possibly linked to climatic fluctuation, i.e., the onset of a warm and humid climate mode. The intense tectonic activity that took place within the Eastern Carpathians during mid-Cretaceous times could also have been responsible for the environmental changes by modifying the circulation pattern in the Moldavian trough from a restricted circulation to a more open one.

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1. Introduction

The mid-Cretaceous (i.e., the Barremian–Turonian interval, ca. 124–80 Ma) was an interval of significant changes in the Earth History. For this period, a global sea-level highstand, linked to a massive pulse of oceanic crustal production in the Cretaceous Normal Superchron, was assumed (Arthur et al., 1985; Larson, 1991). Most of the known Oceanic Anoxic Events (OAEs) took place during mid-Cretaceous times, being linked to pronounced changes in the Earth system, such as oceanic and atmospheric circulation patterns, modification in carbon, phosphorus and oxygen biogeochemical cycles, as well as fluctuations in marine biotas (Jenkyns, 1980; Erbacher et al., 1996; Larson and Erba, 1999; Hay, 2008). Isotopic investigations of Cretaceous oceanic anoxic events indicate important shifts in the global carbon cycle, most pronounced during the Early Aptian OAE 1a and the Cenomanian-Turonian boundary interval, during OAE2 (Schlanger and Jenkyns, 1976; Arthur et al., 1988), but between the above-mentioned oceanic anoxic events there are several less pronounced OAEs (sub-OAEs), such as OAE1b to OAE1d that extend from the Aptian up to the Albian-Cenomanian boundary interval.
One of the sub-OAEs is OAE1d (Erbacher et al., 1996), described as the Albian-Cenomanian Boundary Event (Jarvis et al., 2006), which correlates with the Niveau Brechstrofer of the Vocontian Basin in south-east France (Bréhéret, 1988) and the Pialli Level of the Umbria-Marche Basin in Italy (Cocciioni et al., 1987; Cocciioni, 2001). The OAE1d has so far been identified in the Western Tethys, North Atlantic and Pacific regions. Therefore a global intensification of thermohaline stratification, playing an important role in preservation of organic carbon at the sea floor in these semirestricted basins, was assumed (Erbacher et al., 2001; Wilson and Norris, 2001; Cocciioni et al., 2001; Bornemann et al., 2005; Robinson et al., 2008; Ben Fadhel et al., 2011; Scott et al., 2013).

Few OAEs have so far been identified in the Romanian Carpathians. OAE2 was described in the Southern Carpathians (Melinte-Dobrinescu and Bojar, 2008) and the southern part of the Eastern Carpathians (Cetean et al., 2008). The Valanginian Weissett Event was observed in the Carpathian bend area, as the regional late Valanginian Nutritition Event (Barbu and Melinte-Dobrinescu, 2008), while several OAEs (i.e., OAE1a, OAE1b and possibly OAE1c and OAE1d) were recognized in the Apuseni Mountains (Papp et al., 2012).

The aim of this paper is to present new lithological/sedimentological, micropalaeontological and geochemical data on the sediments deposited within the Albian-Cenomanian boundary interval in the central part of the Eastern Carpathians. The palaeoenvironmental setting is reconstructed, and possible links to the global oceanic anoxic events of the Albian-Cenomanian boundary interval are discussed.

2. Geological setting

The Eastern Carpathians consist of the western thick- and the eastern thin-skinned tectonic nappes. According to Sândulescu (1975, 1980) and Sândulescu et al. (1981), the following units were recognized (Sândulescu, 1975, 1980; Sândulescu et al., 1981): the Pienides, the Transylvanian Nappes, the Median Dacides (Central East Carpathians Nappes), the Outer Dacides, the Inner Moldavides and the Outer Moldavides, (Fig. 1). Shortening and thrusting of the Eastern Carpathian nappes, as for the Southern Carpathian ones, took place in several periods of deformation from the Late Cretaceous to Quaternary times (Sândulescu et al., 1981; Maţenco and Bertotti, 2000; Răbăgia et al., 2011).

The Inner and Outer Moldavides belonging to the Moldavid Nappe System developed only in the Eastern Carpathians, in the outer (eastern) part of the Flysch Zone (Sandulescu, 1984; Ştefănescu, 1995). Although some effects of the Laramian movements deformed the inner units of these nappes (the inner units), the main movements, Miocene in age, structured the Moldavid nappe pile and thrust it eastwards over the East European Platform (Sândulescu, 1984, 1995; Ştefănescu and Polonic, 1988; Maţenco and Bertotti, 2000). The oldest deposits of the Moldavides are Early Cretaceous in age, developed in hemipelagic and turbidites facies, followed by Upper Cretaceous and Paleogene turbidites and hemipelagic sequences (Sândulescu et al., 1981; Melinte-Dobrinescu and Jipa, 2008; Melinte-Dobrinescu and Roban, 2011), and by huge piles of Neogene shallow marine and continental facies.

One of the major tectonic units of the Moldavides is the Teleajen Nappe, where the section herein presented is situated (Fig. 1). The Teleajen Nappe belongs to the western (innermost) structures of the Inner Moldavides (Dumitrescu et al., 1970). Taking into account its position, in the neighbourhood of the Outer Dacides, two deformation processes were assumed for this nappe, i.e., a first phase with predominant folding and a second phase mainly consisting of overthrusting (Sândulescu et al., 1981). This nappe shows remarkable lithofacial diversity, both along strike (N–S) and transverse to the strike (W–E). The oldest deposits, Valanginian–Barremian in age, are represented by black and grey shales (Bâncilă, 1958; Dumitrescu et al., 1970; Grigorescu, 1971; Grigorescu and Anastasiea, 1976; Antonescu et al., 1978), similar to those recorded in the outer nappes of the Eastern Carpathians (Roban and Melinte-Dobrinescu, 2012). The late Haueterian–late Aptian interval is characterized by the deposition of shaly-sandy turbidites, consisting of grey shales and marls, as well as grey-greenish sandstones (Sândulescu, 1975; Ion, 1975). The Lower Cretaceous sedimentary rocks of the Teleajen Nappe accumulated at a very low rate, in a deep marine environment. Similar depositional conditions lasted until the latest Albian, when, as a result of the mid-Cretaceous deformations, the accumulation rate increased (Ştefănescu and Melinte, 1996).

The most developed lithological unit of the Teleajen Nappe is the upper Aptian–upper Albian Teleajen Formation (Ion, 1978; Melinte-Dobrinescu et al., 2009), composed of massive sandstones accompanied by conglomerates, occurring in several intervals. Towards the end of the Early Cretaceous (i.e., in the latest Albian), an important influx of terrigenous material was delivered by inner (western) sources into the basin where the deposits of the Teleajen Nappe were sedimented (Ştefănescu and Melinte, 1996). In the latest Albian–Coniacian interval, during which the Cernatu Formation was deposited, varying rates of subsidence, combined with a reduced supply of terrigenous material, conditioned the accumulation of condensed hemipelagic deposits (mostly variegated red, grey and green shales), interbedded with grey and black shales (Fig. 2). The red shales are similar to the Cretaceous Oceanic Red Beds (CORBs), described from many Tethyan areas, starting from the lower part of the Upper Cretaceous (Hu et al., 2005a; 2012; Wagrreich and Krenmayr, 2005; Bak, 2007; Yilmaz, 2008; Melinte-Dobrinescu et al., 2009). In general, this type of sedimentation lasted, in the Teleajen Nappe, up to the Coniacian. The post-tectonic cover of the nappe consists of Campanian–Maastrichtian red marlstones, followed by Paleogene flysch and Miocene molasse deposits.

3. Material and methods

The studied section is situated on the right bank of the Cernatu Valley (N 45°59′39.35″E, 25°59′30.72″) (Fig. 1). In total, 14 m of rocks were sampled. Carcaseous nanofossils were analysed using simple smear slides and light microscope techniques (Bown and Young, 1998). In all, 19 samples have been studied. The preservation state was recorded as follows: M – moderate (around 50% of the specimens could be easily identified, while the remaining nanofossils show overgrowth and/or dissolution but the taxonomic identification is not hindered); P – poor (more than 75% of the specimens show dissolution and/or overgrowth and specific identification is often difficult).

The diversity (species richness) was estimated as the number of the total taxa in each sample. The absolute abundance is considered as the average nanofossil number in one field of view, as follows: A – abundant: >1 specimen/field of view (FOV); C – common: 1 specimen/2–10 FOVs; F – few: 1 specimen/11–20 FOVs; R – rare: 1/50 FOVs.

The samples used for carcaseous nanofossil investigations were also used for foraminiferal analysis. The samples (about 0.75 kg each) were first completely dried, for eliminating the intestinal water, then boiled in ‘Glauber’s salt’ (Na2SO4), and exposed to several freeze/thaw cycles. Finally, the samples were washed by using a battery of sieves (63–100–2000 μm). In each sample, the whole washed material resulting from the 750 g of the primary
sample was analyzed, with special focus on the >100 µm fraction that provided the most important microfaunal assemblages. From the aforementioned fraction, all the specimens were collected, while from the 63–100 µm fraction we analyzed and selected specimens from 25% of the washed material. For each sample, a 60-

grid micropalaeontological slide was used, the specimens being glued in position according to their observed abundance and their morphogroup affiliation. The microphotographs of foraminiferal taxa were taken using a ZEISS – Stemi SV11 microscope with an adapted NIKON digital camera. In order to examine the internal
Fig. 2. Litho- and biostratigraphy of the Albian-Coniacian deposits that crop out in the Cernatu Valley (inner part of the Eastern Carpathians, Teleajen Nappe). Legend: 1—sandstones; 2—siderites; 3—grey shales; 4—tuffitic shales; 5—red shales; 6—dark grey and black shales; 7—first occurrence; 8—last occurrence; 9—interval of occurrence; IZ—interval zone.
structures, some of the agglutinated species were studied and illustrated under glycerin immersion. For quantitative studies, the percentage of specimens/taxa from the entire number of specimens recorded in one sample was calculated. The abundance was considered as follows: abundant >20%; frequent = 10%–20%; moderate = 3%–10%; and rare <3%.

To determine the volatile content, the loss on ignition method (LOI) was applied, by means of combustion using a Caloris L1003 type furnace. The samples were first ground and dried in the oven for 5 h at 105 °C. Then, they were weighed and treated at 550 °C for 6 h. The weight loss between 105 °C and 550 °C should be proportional to the amount of organic matter (OM) contained in the sample (Heiri et al., 2001). Besides the organic carbon, the volatiles could contain interstitial water and other gases. Therefore, the calculated organic matter (OM) includes CO2 and other gases, as well as interstitial water, besides the total organic carbon (TOC). Afterwards, the samples were treated for another 3 h at 950 °C. The weight loss between 550 °C and 950 °C is proportional to the amount of the inorganic CO2 content of the sample. Further, the inorganic CO2 and CaCO3 amounts were calculated. X-ray fluorescence (XRF) elemental analyses were performed on a HORIBA XGF 7000 X-ray spectrometer. Analytical precision in the analysis of major elements, (Mg, K, Si, Al, Ca, Fe, P, S and Mn) is 0.01%.

Stable isotopic analyses were performed at Graz University. The stable isotopes were measured on the organic fraction. The results of the stable isotope analyses are reported in per mil to the Pee Dee Belemnite Standard (VPDB).

4. Results

4.1. Lithostratigraphy

Previous studies (Filipescu et al., 1963; Sândulescu and Sândulescu, 1965; Ion, 1975, 1976; 1978 and this study) identified in the Cernatu Valley the following Middle Albain–Coniacian succession:

(i) The Middle Member of the Teleajen Formation (middle Albain–early late Albain in age), comprises massive calcareous sandstones with ripple marks, interbedded with thin grey-green shales and marlstones. The identified agglutinated foraminiferous assemblages contain common Plectocurvoides alternans Noth, Thalmanammina neocomiensis Geroch, Glomospira irregularis (Grzybowski), Gaudryina filiformis Berthelin, Rhizammina spp. and Saccammina spp. which were assigned to the Plectocurvoides alternans-Haplophragmoides gigas minor-Haplophragmoides concavus Interval Zone (Ion, 1976, 1978).

(ii) The Upper Member of the Teleajen Formation (late Albain pro parte), composed of rhythmically alternating thin grey-green calcareous sandstones and grey-green marlstones. At the top of the Teleajen Formation, centimetre-thick siderites are present. The lower and middle parts of this member contain similar agglutinated foraminiferous assemblages to those described above, characteristic of the Plectocurvoides alternans-Haplophragmoides gigas minor-Haplophragmoides concavus Interval Zone. Around the middle part of the unit, the calcareous benthic foraminiferous species Pleurostomella obtusa Berthelin was observed; (Ion, 1975, 1978); this bioevent is followed by the first occurrence of Quadririnophora altomorphophoridae (Reuss) The FO of the planktonic foraminiferous taxon Gonyaulax caesar (Glaessner) Renz, Luterbacher & Schneider was identified within the upper part of the Upper Member. The macrofaunal taxa include the belemnite Neohibolites minimus Miller and the ammonite Pazosia mayoriana d’Orbigny. Towards the top of this member, Ion (1975, 1978) identified the Plectocurvoides alternans and Haplophragmoides falcatusuritalis agglutinated foraminiferal zones. Other foraminifers present in the assemblages are Recurvoides imperfectus (Hanzlikova); Reophax guttifer (Brady), Pseudobolivina variabilis (Vašiček), Gaudryina oblonga (Zaspelova), Caudamina ovula ovula (Grzybowski) Geroch, as well as acme occurrences of Thurammina spp. and Trochammina spp. (Ion, 1978). Rare specimens of the ammonite taxa Stolizkaia dispar d’Orbigny and S. notha Seeley were identified at the top of the Teleajen Formation in the Cernatu Valley (Ion and Szasz, 1994).

(3) The Cernatu Formation (latest Albain–Coniacian in age) comprises grey and blackish shales at the base, followed by red shales interbedded with thin blackish, grey and green shales. The top of this formation is formed of red, dark grey and greenish marlstones and shales. Previously, the age of the Cernatu Formation was mainly assigned on the basis of the macrofaunal and foraminiferous assemblages (Ion, 1976, 1978). Therefore, the deposits of the lower part of the Cernatu Formation belong to the Plectocurvoides alternans-Haplophragmoides falcatusuritalis Interval Zone (as defined by Neagu et al., 1992), which is concomitant with the Triatexa gauldaina Acme Subzone (Fig. 2). Towards the upper part of the Cernatu Formation, the planktonic foraminiferous species Marginotruncana coronata (Bolli) first occurs.

In this paper, we investigated in detail 14 m of sedimentary rocks, composed of grey shales in the lower part and predominantly red shales in its upper part. Towards the top, thin cm grey to blackish and green shales are interbedded with the red ones (Figs. 3A, 3B and 4).

Taking into account the grain size and the CaCO3 content of the investigated deposits, several lithofacies were identified, as follows:

(i) The shaly lithofacies, which prevailed in the studied section, contains grey-green, red and red-green couples, with low CaCO3 content, up to 5%. The lowest part of the section is composed of grey to blackish and green shales, followed by red shales, with interbedded thin centimetre-thick grey to blackish and green shales. Sedimentary structures observed are horizontal parallel lamination and subordinate massive beds.

(ii) The carbonate shaly lithofacies comprises fine-grained deposits with a CaCO3 content between 20 and 25%. Two intercalations of this lithofacies were encountered in the highest part of the section; one of them, up to 40 cm thick, shows a dark grey colour, while the other, close to the top of the section, up to 20 cm in thickness, is formed of red and green millimetre couples.

(iii) The marly lithofacies is present only at the base of the section. The grey marlstones, up to 20 cm in thickness, have a high CaCO3 content of almost 50%.

(iv) The sandstone lithofacies is present only in the lower part of the section as a single bed up to 10 cm in thickness. This sandstone bed shows parallel laminations except at the top, where slightly deformed crossed laminations were observed. The CaCO3 content is 26%, this relatively high value being linked to carbonate cement and clasts.

4.2. Calcareous nannofossils

From 19 samples analyzed, only 10 yielded calcareous nannofossil assemblages. In general, the samples with nannofloras
showed moderate preservation towards the upper part of the section, and poor preservation in the lower part (Table 1). The most abundant species in all assemblages is *Watznaueria barnesiae* (Black in Black & Barnes, 1959) *Perci-Nielsen* (1985), a taxon regarded as one of the nannofossils that are most resistant to dissolution. Assemblages containing more than 40% of *Watznaueria barnesiae* are supposed to be heavily altered (Roth and Krumbach, 1986).

All of the samples are clearly dominated by *Watznaueria barnesiae*, which exceeds 40% of the calcareous nannofossil assemblages, while in samples from the lower part of the section (i.e., samples CE11 and CE12b) it represents over 80% of the total nanofloras. The next common recorded nannofossil is *Eprolithus floralis* (Stradner, 1962) *Stover* (1966), which is also an indicator of nannofloraal preservation, being highly resistant to dissolution (Roth, 1978; Lamolda et al., 1994; Melinte-Dobrinescu et al., 2013). In all the studied samples, *Watznaueria barnesiae* and *Eprolithus floralis* jointly amounted to over 60% of the total nanofloras.

*Eifelithus turriseiffelii* (Deflandre in Deflandre & Fert, 1954) Reinhardt (1965) is present in nanofossil assemblages from the base of the section, hence the entire section is assigned to the NC10 Zone of Roth (1978, 1983), specifically the NC10a Subzone of Bralower et al. (1993), as well as to the CC9 Zone of Sissingh (1977), specifically the CC9b Subzone of *Perci-Nielsen* (1985). The nanofossil *Eifelithus monechiae* Crux (1991) is present in the lower part of the studied section; *Hayesites albiensis* Manivit (1971) which disappears 3 m above the base of Niveau Breistroffer (Late Albian) at Mont Risou, France - the Global Boundary Stratotype Section and Point (GSSP) for the base of the Cenomanian Stage, (Kennedy et al., 2004), is not present in the Cernatu Valley section. Other nannofossils that characterize, according to the aforementioned authors, the GSSP of the base of the Cenomanian, i.e., *Calcultes anfractus* (Jakubowski, 1986) Varol & Jakubowski (1989), with a latest Albian first occurrence (FO) and *Gartnerogona theta* (Black in Black & Barnes, 1959), Jakubowski (1986), with an earliest Cenomanian FO, are not recorded in the studied section.

Recently, Gradstein et al. (2012) indicated that the FO of *Eiffelithus monechiae* is calibrated at 107.5 Ma, while the FO of *Eiffelithus turriseiffelii* is dated at 103.13 Ma, and it is followed, at 100.54 Ma by the LO (last occurrence) of *Hayesites albiensis*. The latter bioevent is slightly below the Albain-Cenomanian boundary, dated as 100.45 Ma. The Cernatu Valley section is between the FOs of *Eiffelithus turriseiffelii* and *Corollithion kennedyi* Crux (1981), and therefore probably spans the time-interval between 107.5 and 100.45 Ma. The absence of *Hayesites albiensis* could be interpreted as a primary signal, or may be linked to the poor preservation state of the nanofloras.

### 4.3. Foraminiferal assemblages

In the investigated section, the foraminiferal assemblages are dominated, at least in the lowest part, by deep water benthic agglutinated taxa (Fig. 5). In the upper part of the section, in sample CE22 and above, some calcareous foraminiferal species also occur, but no planktonic species have been noticed. Preservation of the foraminfera is generally moderate to good, but in the lowest deposits the calcareous benthic foraminfera are very rare and partially dissolved. The abundance (number of specimens/sample) varies between 300 and 500, but the highest samples, especially those collected from the red shales, yielded a higher abundance, more than 1000 specimens/sample.

The foraminiferal associations contain index species, such as *Plecterecurvoidea alternans* Noth, *Haplophragmoides falcatusauralis* Neagu and *Bulbobaculis problematicus* (Neagu), used in the zonation of *Geroch and Novak* (1984), as well as *Neagu* (1990; Neagu et al., 1992) for the Carpathian area to characterize the Late Albian—Early Cenomanian interval. The last two taxa were more numerous towards the top of the section (Fig. 5).

Besides these species, the foraminiferal assemblages frequently contain other agglutinated taxa (Figs. 6 and 7) such as: *Rhabadammina sp.*, *Rhizammina indivisa* Brady, R. sp., *Bathyphysion dubius* (White), B. vitta Nauss, *Kalamopsis grzybowski* (Dylążanka), *Hiplocrepina depressa* Vasić, H. sp., *Psammosphaera fusca* Schultzze, *Saccammina grzybowski* (Schubert), S., *Lagenammina alexanderi* (Loeblich & Tappan), *Hyperammina gaultina* Ten Dam, H. sp., *Ammodiscus cretaceus* (Reuss), *Glomospira gordialis* (Jones and Parker), G. *charoides* (Jones and Parker), G. ex gr. *irregularis* (Grzybowski), *Lituotuba lituiformis* (Brady), L. incerta Franke, *Conglophragmium irregularis* (White), *Caudammina ovula* (Grzybowski), *Rzehalkina sp.*, *Spiropectammina roemeri* Lalicker, *Traxia gaultina canarenta* (Neagu). Most of the samples commonly contain taxa of the genus *Pseudodosinella*, i.e., *P. parvula* (Huss), which is considered by *Geroch and Kaminski* (1995) a deeper water
Fig. 4. Lithology, biostratigraphy and geochemistry of the Albian/Cenomanian boundary interval in the Cernatu Valley section (central part of the Eastern Carpathians). Legend: 1 - shales; 2 - carbonate shales; 3 - marlstones; 4 - sandstones; 5 - Current ripples; m - mud; z - silt; s - sandstone.

Occurrence interval of the ammonite Stoliczkaia dispar NC10a (CC08b)

Plectorecurvoidea alternans
Haplophragmoides falcatusuturalis

Aglutinated foraminiferal zones

Geochemistry

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equivalent of “Reophax” minuta Tappan as well as P. troyeri (Tappan). The former species is more common in the lowest part of the section. Others components of the agglutinated foraminiferal associations are Gerochammina stanislawi Neagu (both shorter and elongated morphotypes), having its first occurrence in the early Albian and its acme and maximum size in the Cenomanian (Neagu, 1968, 1990), as well as Trochammina umiatensis Tappan, which spans the latest Albian-earliest Cenomanian interval (Neagu, 1972). The genus Recurvoides is represented especially by Recurvoides imperfectus (Hanzlikova/C19a), which is more frequent in the upper part of the section, and Recurvoides ex gr. walteri (Grzybowski), which is found particularly in the lowest part of the section.

Calcareous benthic foraminifers appear only towards the upper part of the section (Fig. 5). They are mainly represented by lagenids, such as Nodosaria oligostegia Reuss, Lenticulina pachynota (Reuss), Ramulina novaculeata Bullard, as well as buliminids, such as Pleurostomella obtusa Berthelin and Nodosarella frequens (Storm). The species Osangularia cretacea (Carbonier) also occurs commonly.

4.4. Geochemistry

The organic matter content (OM) fluctuates throughout the section between 2.0 and 3.2 (w %), with the highest values recorded in general towards the lower part of the section, in the blackish, grey and green shales. In the upper part of the section, above the lowest occurrence of the red shales, the OM values are around 2.8%, decreasing to around 2% at the top (Fig. 4).

The CaCO3 content is generally very low, except at the base of the section, where it reaches 50%. High values, around 20–25%, were also identified in the sandstone bed and towards the top of the section. The highest oxide values were identified for SiO₂, between 35 and 60%, followed by Al₂O₃, between 9 and 18%, and Fe₂O₃ which varies from 5 up to 13%, with the maximum value recorded in the red shale beds. Low values were identified for MgO, which fluctuates between 2 and 4%, and K₂O, which shows values between 0.5 and 6%, the highest ones being recorded in the upper part of the section (Fig. 4). Very low amounts of TiO₂ and MnO₂, in general below 1%, characterize the section. Two peaks of MnO₂, synchronous with peaks of CaCO₃, were observed, towards the top of the section (Fig. 4).

4.5. Carbon stable isotopes

The organic δ¹³Corg values vary between −25.30‰ and −24.01‰. Both the lowest and the highest values were recorded in the lowest part of the section. Within the lower part of the section, in the middle part of the grey to blackish and green shales, a ca 2 m interval yielded less negative δ¹³Corg values, between −24.5‰ and −24.01‰ (Fig. 4).
Fig. 5. —Distribution and relative abundance of foraminiferal taxa in the Cernatu Valley section (Eastern Carpathians). For the abundance, the following classes have been used: abundant >20%; frequent = 10%–20%; moderate = 3%–10%, and rare <3%. For the lithology see legend of Fig. 4.
Above, organic δ¹³C values decrease significantly to −25.20‰, than increase again to −24.30‰ in the blackish to grey shales interbedded with the red shales. At the top of the section, the δ¹³C values decrease again to below −25‰ (Fig. 4).

5. Discussion

5.1. Biostratigraphy

A significant biostratigraphic marker for the interval studied is the ammonite Stoliczkaia notha (see Filipescu et al., 1963; Ion and Szasz, 1994), which occurs in the lower part of the section, up to sample CE19 (Fig. 4). Stoliczkaia notha is usually recorded in the uppermost Albian, together with the ammonite Stoliczkaia dispar (Kennedy and Delamette, 1994; Kennedy et al., 2008).

The FAD (first appearance datum) of representatives of the ammonite genus Stoliczkaia, i.e., S. dispar and S. clavigera, was observed in the uppermost Albian deposits of the Col de Palluel section, Hautes-Alpes region, France, in the Mortoniceras rostratum ammonite Zone, while its LAD (last appearance datum) is placed around the Albian-Cenomanian boundary, slightly below the first occurrence of the planktonic foraminiferal species Thalmanninella globotruncanoides (Gale et al., 2011); the latter event marks the base of the Cenomanian (Kennedy et al., 2004).

Based on the foraminiferal assemblages, the lower part of the section was assigned to the upper part of the Plecturecurvoides alternans Zone, late Albian in age (Neagu et al., 1992). The upper part of the section (sample CE19 and above) belongs to the Haplophragmoides falcatosuturalis Zone which covers the Albian-Cenomanian boundary interval (Neagu et al., 1992). According to the aforementioned authors, the Haplophragmoides falcatosuturalis benthic foraminifer Zone corresponds to the upper part of the Parathalmanninella appenninica planktonic foraminifer Zone and the lower part of the Thalmanninella globotruncanoides Zone. In
terms of nannofloral biostratigraphy, the section is situated within the NC10a calcareous nannofossil Subzone (Roth, 1978; Bralower et al., 1993) and the CC9b Subzone (Sissingh, 1977; Perch-Nielsen, 1985). To summarize, the base of the Cernatu Valley section is placed within the latest Albian, while a latest Albian–earliest Cenomanian age could be assumed for its top.

5.2. Palaeobathymetry

Agglutinated and calcareous foraminiferal morphogroups have been used to reconstruct the palaeobathymetry of the Carpathian basin, where the dark grey and green shales, followed by red shales accumulated during the Albian-Cenomanian boundary interval. Formerly, many authors, i.e., Koutsoukos and Hart (1990), Nagy et al. (1995), van den Akker et al. (2000), Holbourn et al. (2001), Kaminski and Gradstein (2005) and Setoyama et al. (2013), used this method to determine the palaeobathymetry of various sedimentary marine basins. This approach was also applied to the Upper Cretaceous pelagic to hemipelagic deposits of the southern Eastern Carpathians (Cetean et al., 2008; Cetean et al., 2011). However, the use of morphogroups in palaeoenvironmental reconstructions has some limitations (Kender et al., 2008), especially when applied to turbidite deposits, where the frequent sediment transport processes imply substantial microfaunal reworking.

The lower part of the Cernatu Valley section, composed mainly of dark grey and green shales, with a single interbedded thin 10 cm-thick sandstone, is characterized by a predominance of tubular morphotypes, i.e.,

- *Rhizammina*, *Rhabdammina* and *Bathysiphon*, frequent globular morphotypes, such as *Psammosphaera*, *Saccammina* and *Caudammina*, as well as flattened planispiral/trochospiral, i.e., *Ammodiscus*, *Glomospira*, *Lituitubus* and *Rzebachina*. The aforementioned morphotypes correspond to the M1, M2a and M3a agglutinated benthic foraminiferal morphotypes described as erect and surficial epifaunas or shallow infaunas with a suspension and/
or passive deposit feeding behaviour (Setoyama et al., 2011, 2013; Cetean et al., 2011). These types of morphotypes indicate a lower bathyal to abyssal marine setting (Fig. 8). Holbourn et al. (2001) showed that assemblages with *Rhizammina* spp and *Plectorecurvoides alternans* characterized the abyssal zones of the North and South Atlantic Ocean during the Late Albian.

These data are in agreement with the calcareous nannofossil distribution. From the base of the section up to sample CE18, the deposits are almost devoid of nannofossils (Table 1), while the few samples where they occur yielded assemblages largely dominated by *Watznaueria barnesiae*, which is highly resistant to dissolution and/or diagenetic processes. We assume that the palaeo-obathymetric setting of the lower part of the section was abyssal up to lower bathyal, and close to the CCD. Because some intervals of the lower part of the section contain a few ammonites and nannofossils, we suppose that sedimentation took place above the CCD, the depth of which was at 2500 m at that time (Tucholke and Vogt, 1979).

In the upper part of the section, especially in the interval from sample CE20 up to sample CE22, an increase was observed in planconvex trochospiral morphotypes (*Trochammina*), rounded streptospiral ones, i.e., *Recuvoides*, rounded planspiral (*Haplophragmoides*), elongate subcylindrical, such as *Gerachammina*; all of them benthic foraminiferal morphotypes belonging to the M2b and M4a-b morphogroups (Fig. 8). Additionally, elongate tapered morphotypes, such as *Bulbobaculites* increased in abundance at the top of the section (between samples CE22 and CE26), while *Pseudonodosinella*, the other genus belonging to the same morphotype, showed a maximum abundance in the middle part of the section, between samples CE14 and CE21.

These morphotypes are characteristic of surficial epifaunas, as well as shallow to deep infaunas, with an active deposit feeding behaviour (Cetean et al., 2011; Setoyama et al., 2013). The aforementioned authors indicate that these morphotypes are present in a variety of marine settings, such as upper bathyal to abyssal, and also shelf to deep marine. In the N and S Atlantic, late Albian assemblages dominated by *Plectorecurvoides alternans*, *Rhizammina* spp. and *Ammodiscus*, such as those observed between samples CE19 and CE22 in the Cernatu Valley section, are linked to lower bathyal down to abyssal palaeoenvironments (Holbourn et al., 2001). The whole interval between samples CE19 and CE22 contains calcareous nannofossil assemblages that still show generally poor preservation (Table 1).

Towards the top of the section, calcareous foraminifers, together with agglutinated ones, are present. The occurrence of such mixed associations probably mirrored changes in bottom conditions, from dysoxic to oxic. It should be noted that the lowest occurrence of calcareous foraminifera in the section, i.e., the genera *Pleurostomella* and *Osangularia*, is located slightly below the oldest red shale occurrence, just above the end of the presumed OAE1d. Possibly, this bioevent is linked to a short-term change, i.e. the shift from a dysoxic setting to an oxic one.

Afterwards, the two above-mentioned taxa show a temporary disappearance up to samples CE22 and CE 24 respectively, as well as an increasing frequency at the section top, probably reflecting stable oxic conditions.

The topmost two samples of the section show increasing frequencies of calcareous foraminifers, as well as abundant and high diversity agglutinated foraminiferal assemblages. Additionally the nannofloras are consistent towards the top of the section, showing a moderate preservation.

The calcareous benthic foraminifera at the top of the section are mainly represented by low trochospiral/bilaterally symmetrical morphotypes (*Osangularia*), pyramidal to conical ones (*Pleurostomella*), as well as cylindrical ones (*Nodosaria, Ramulina* and *Marginulina*) that can be included in the CH-B morphogroup (CH-B 1, CH-B4 and CH-B7 sub-morphogroups), which refers to infaunal taxa with active deposit feeding (Frenzel, 2000). Concerning the agglutinated foraminifers of the samples CE23–CE26, the highest abundance is shown by the genera *Haplophragmoides*, *Bulbobaculites*, *Glosamira* and *Caudammina* which belong to the morphotypes M3 and M4 (Setoyama et al., 2011, 2013), distributed in diverse marine settings, such as inner shelf, upper bathyal, but also in abyssal regions.

Taking into account the above-mentioned data, we can assume that the investigated sedimentary rocks were deposited in a deep marine area, from upper bathyal to abyssal. As the lower part of the section contains a few ammonites and nannofossils, and the upper part of the section yielded moderately preserved calcareous nannofossils, we suppose that all of the sedimentary rocks of the Cernatu section were deposited above the CCD. In general, the identified agglutinated foraminifers, including *Plectorecurvoides alternans*, are typical deep-water assemblages from environments below the CCD, but they were also reported, from both turbidites and hemipelagic deposits in other Carpathian areas, i.e. the Polish Carpathians (Bąk et al., 2005). It was assumed (Bąk, 2000) that the upper to middle bathyal depths created an upper depth limit for the range of *Plectorecurvoides alternans*.

### 5.3. Palaeoenvironment

The lower part of the section is characterized by grey to blackish and green shales, followed by red shales. Based on colour modification, as well as on geochemical features, a shift from the late Albian anoxic/dysoxic conditions to oxic ones within the Albian-Cenomanian boundary interval could be assumed.

$\text{Al}_2\text{O}_3$ is positively correlated with $\text{SiO}_2$ ($R^2 = 0.61$), (Fig. 9A) and $\text{TiO}_2$ ($R^2 = 0.50$). The red shales of the Cernatu section show a higher $\text{Fe}_2\text{O}_3$ content, between 10 and 13%, more than the grey to blackish and green shales in the lower part of the section. The presence of $\text{Fe}_2\text{O}_3$ is possibly linked to haematite clusters. In general, a very weak negative correlation exists between $\text{Fe}_2\text{O}_3$ and MnO$_2$ ($R^2 = 0.33$), the latter oxide displaying generally lower values in the red shales than in the grey to blackish and green shales. A weak negative correlation ($R^2 = 0.31$) was found between MnO$_2$ and SiO$_2$, but a positive one ($R^2 = 0.53$) exists between MnO$_2$ and CaCO$_3$, (Fig. 9B), the latter being probably an indicator for the authigenic origin of the manganese bearing minerals. SiO$_2$ is negatively correlated with CaO ($R^2 = 0.7$) (Fig. 9C).

The weak positive correlation between $\text{Al}_2\text{O}_3$ and $\text{Fe}_2\text{O}_3$ (Fig. 4) suggests a continental origin of the iron, by the alteration of lateritic crusts from emerged areas. These correlations led to the interpretation that the Al, Si, Ti and Fe have mainly a continental origin, while the Mn and Ca are included in authigenic minerals, such as calcite and siderite. On the other hand, the weak correlations among these compounds do not exclude an authigenic origin of the iron.

A high Fe/Al ratio was observed throughout the section, averaging 0.93, but with a higher average, i.e., 0.97, in the red shales (above sample CE17), and a lower one, i.e., 0.81, in the grey to blackish and green shales. It was assumed that the high values of the Fe/Al ratio could be linked to warm and humid conditions resulting from intense weathering of the hinterland rocks (Pattan et al., 2012). Such a setting, together with a high precipitation regime, could have decoupled soluble Fe from the red soils of emerged plains, so that it could be transported into the basin of the Moldavides of the Eastern Carpathians, where the red shales started to accumulate within the Albian-Cenomanian boundary interval.
<table>
<thead>
<tr>
<th>Morphgroup</th>
<th>Morphotype</th>
<th>Test form</th>
<th>Life position</th>
<th>Feeding habitat</th>
<th>Environment</th>
<th>Main species (Recoded in this study)</th>
</tr>
</thead>
<tbody>
<tr>
<td>M1</td>
<td>Tubular</td>
<td>Erect fauna</td>
<td>Suspension feeding</td>
<td>Tranquil bathyal and abyssal with low organic flux</td>
<td>Rhizammina indivisa, Rhabdammina sp., Bathysiphon dubius, Kalamopsis grybowskiinn</td>
<td></td>
</tr>
<tr>
<td>M2</td>
<td>Globular</td>
<td>Shallow infauna</td>
<td>Suspension feeding and/or Passive deposit feeding</td>
<td>Common in bathyal and abyssal</td>
<td>Psammospira fissa, Hyperammina gautiina, Saccoammina grybowskiinn, Caudammina ovulam</td>
<td></td>
</tr>
<tr>
<td></td>
<td>M2a</td>
<td>Rounded trochospiral and streptospiral</td>
<td>Surficial epifauna</td>
<td>Active deposite feeding</td>
<td>Shelf to deep marine</td>
<td>Recurvirovaid ex. gr. walteri, Recurvirovaid imperfecctus, Plecturorevovaid alternans</td>
</tr>
<tr>
<td></td>
<td>M2b</td>
<td>Planeconvex trochospiral</td>
<td>Surficial epifauna</td>
<td>Active deposite feeding</td>
<td>Shelf to deep marine</td>
<td>Trochammina umiatensis, Trochammina sp.</td>
</tr>
<tr>
<td></td>
<td>M2c</td>
<td>Elongate keeled</td>
<td>Surficial epifauna</td>
<td>Active deposite feeding</td>
<td>Shelf to marginal marine</td>
<td>Spiroplectammina roemeri</td>
</tr>
<tr>
<td>M3</td>
<td>Flattened trochospiral</td>
<td>Surficial epifauna</td>
<td>Active and passive deposit feeding</td>
<td>Lagoonal to abyssal</td>
<td>Ammodiscus cretaceous, Glomospira gordialis, Glomospira charoides, Lithothea l. luteiformis, Lithothea incerta, Rehakina sp.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>M3a</td>
<td>Flattened planospiral and streptospiral</td>
<td>Surficial epifauna</td>
<td>Active and passive deposit feeding</td>
<td>Lagoonal to abyssal</td>
<td>not in this study</td>
</tr>
<tr>
<td></td>
<td>M3b</td>
<td>Flattened irregular</td>
<td>Surficial epifauna</td>
<td>Suspension feeding</td>
<td>Upper bathyal to abyssal</td>
<td>not in this study</td>
</tr>
<tr>
<td></td>
<td>M3c</td>
<td>Flattened streptospiral</td>
<td>Surficial epifauna</td>
<td>Active and passive deposit feeding</td>
<td>Upper bathyal to abyssal</td>
<td>Paratrochamminoides irregularis</td>
</tr>
<tr>
<td>M4</td>
<td>Rounded planospiral</td>
<td>Surficial epifauna and/or shallow infauna</td>
<td>Active deposite feeding</td>
<td>Inner shelf to upper bathyal</td>
<td>Haplophragmooides falcatusaurdis</td>
<td></td>
</tr>
<tr>
<td></td>
<td>M4a</td>
<td>Round planospiral</td>
<td>Surficial epifauna and/or shallow infauna</td>
<td>Active deposite feeding</td>
<td>Inner shelf to upper bathyal</td>
<td>Gerochamminoides stanislavii, Tritaxia gautiina carenata</td>
</tr>
<tr>
<td></td>
<td>M4b</td>
<td>Elongate subcylindrical</td>
<td>Active deposite feeding</td>
<td>Inner shelf to upper bathyal with increased organic flux</td>
<td>Pseudonodosinella parvula, Pseudonodosinella troyeni, Reophax cf. deckeri, Bulbofixula problematica, Ammonobaculites sp., Hippocrepina depressa</td>
<td></td>
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</tbody>
</table>

Fig. 8. Relationship between various morphogroups of agglutinated foraminifera and the palaeoenvironment; modified after Cetean et al. (2011) and Setoyama et al. (2013).
The red shales identified in the Cernatu Valley section could be regarded as Cretaceous Oceanic Red Beds (CORBs), which are widely distributed from the Aptian up to the Maastrichtian all over the Tethyan Realm (Hu et al., 2005a). In several regions, such as the Alps (Erba et al., 1999; Wagreich et al., 2009; Hu et al., 2005b; 2012), Carpathians (Skupien et al., 2009; Melinte-Dobrinescu et al., 2009), north Caucasus (Egoyan, 1986), central North Atlantic (Jansa et al., 1979; Weissert, 1991) and Himalayas (Premoli Silva et al., 1991; Hu et al., 2005b), CORBs occurred during mid-Cretaceous times. Since most of the CORBs followed mid-Cretaceous anoxic events such as OAE1a, OAE1b, OAE1c and OAE1d, it was assumed that the deposition of red marine beds is a possible consequence of oceanic anoxic events (Wang et al., 2011).

Based on the sedimentary petrology, CORBs were classified into three main types (Wagreich et al., 2009; Wang et al., 2011): (1) red hemipelagic and pelagic carbonates; (2) deep-water red claystones deposited around the carbonate compensation depth (CCD); and (3) red cherts and radiolarites deposited below the CCD. Additionally, Hu et al. (2012) used geochemistry, i.e., CaO, Al₂O₃ and SiO₂ fluctuations, to classify CORBs, and identified three major categories: (i) Ca-CORBs (red limestones and red marls); (ii) Al-CORBs (red shales); and (iii) Si-CORBs (red cherts and red radiolarites).

According to the above-mentioned classification, the CORBs of the Cernatu Valley section are entirely Al-CORBs (Fig. 9D), deposited slightly above the CCD. Similar to other worldwide distributed Al-CORBs (Hu et al., 2012), the SiO₂ content is high, between 40 and 60%, while the Al₂O₃ content is moderate, averaging 13.5%. The CaCO₃ content globally was found to vary consistently in Al-CORBs, from 1% (Marzak, Czech Republic) to 21% (Albian ODP1049C, North Atlantic), with a maximum of 36% in the Pindos Basin, Greece (Hu et al., 2012), while in the Cernatu Valley section, the average CaCO₃ content is 3%. The lithological and biotic features of the CORBs observed in this section indicate that they could be included in deep-water red shales deposited around (above) the carbonate compensation depth (CCD).

5.4. The Oceanic Anoxic Event 1d

A positive organic δ¹³Corg excursion tentatively attributed to OAE1d was observed in the lower part of the Cernatu Valley section, extending over ca. 2 m in thick dark grey and green shales (Fig. 4). The δ¹³Corg 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, as elsewhere (Bralower et al., 1993; Leckie et al., 2002; Bornemann et al., 2005; Watkins et al., 2005). Globally, OAE1d is situated within the Parathalmanninella appenninica planktonic foraminiferal Zone (Coccioni et al., 2001; Kennedy et al., 2004).

No planktonic foraminifers were observed in the section. However, the agglutinated foraminifers from the lower part of the section (including the interval of the positive δ¹³C excursion) were assigned to the upper part of the Plectorecurvoides alternans agglutinated foraminiferal Zone, corresponding pro parte to the Parathalmanninella appenninica planktonic foraminiferal zone (Neagu et al., 1992). Hence, a Late Albian age may be assigned based on the benthic foraminiferal assemblages.

Interestingly, changes in the agglutinated foraminiferal assemblages have been observed within the OAE1d interval in the section (Fig. 5). The abundance of taxa that belong to the M1 morphogroup (tubular), and in general of the epifaunal species, decreases significantly, while the shallow to deep infaunal species are not so much affected. Within this interval, the temporary disappearance
of some foraminifers, i.e., taxa of the genera *Recurvoides*, *Hyperammina* and *Kalamopsis*, is recorded. This biotic event could be linked to a lowering of bottom oxygen concentration, together with intensification of the organic carbon flux, palaeoenvironmental changes that affected mainly the epifaunal taxa.

It is noteworthy that the genus *Trochammina* first occurs, and in significant numbers, within the OAE1d interval in the section. Former studies (Erbacher et al., 1999) described trochanminids to be the first colonizers after an OAE. In the Cernatu Valley section, *Trochammina* taxa are abundant in the OAE1d interval, above which *Trochammina* decreases in abundance, but then shows another increase towards the upper part of the section (samples CE20–CE24) slightly below the interval (samples CE25–CE26) where the calcareous foraminifers reappear in strength (Fig. 5). The top of the section (between samples CE22 and CE26), where red shales co-occur with grey ones, shows an overall decrease in diversity of the agglutinated foraminifera (Fig. 5), concomitant with a weak (around 0.5‰) positive $\delta^{13}C_{org}$ excursion (Fig. 4). We may hypothesize that the fluctuation pattern of the agglutinated foraminifers is a response to the termination of the Albian-Cenomanian boundary event.

It is noteworthy that, in some of the papers focused on the OAE1d, the authors studied the bulk pelagic carbonate for $\delta^{13}C$ fluctuations (Kennedy et al., 2004; Bornemann et al., 2005; Gale et al., 2011). Other studies (Scott et al., 2013) reported $\delta^{13}C$ measurements of both organic carbon, $\delta^{13}C_{org}$, as we present in this paper, and inorganic carbon $\delta^{13}C_{carb}$, both fluctuations being relatively consistent with the global shift of $\delta^{13}C$ from the end of the Early Cretaceous, i.e. the latest Albian OAE1d.

The positive $\delta^{13}C_{org}$ excursion observed in the lower part of the Cernatu Valley section probably corresponds to the oldest of the above-described peaks, the OAE1d (= Breistroffer level) (Fig. 10). The weak positive $\delta^{13}C_{org}$ excursion identified towards the top of the section could be integrated in the last phases of the Albian-Cenomanian Boundary Event.

6. Conclusions

The sedimentary rocks of the studied Cernatu Valley section, deposited during the late Albian, and probably encompassing the Albian-Cenomanian boundary interval, are mainly composed of hemipelagic deposits, i.e., dark grey to green shales, overlain by red shales interbedded with grey and green shales. These lithologies are typical of the uppermost Lower Cretaceous deposits of the Moldavian Trough that covered large areas that are now included in the Moldavide nappes of the Eastern Carpathians. The deposits of the Moldavian Trough were sedimented in a narrow marine basin, with anoxic to dysoxic deep water, at probably around

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Fig. 10. Correlation between isotope $\delta^{13}C$ fluctuations within the Albian/Cenomanian boundary interval, including the late Albian OAE1d; Mont Risou (France) and Gubbio (Italy) from Kennedy et al. (2004), Bornemann et al. (2005) and Gale et al. (2011); Chihuahua Trough, New Mexico from Scott et al. (2013); Cernatu (Romania) – this paper; FAD – first appearance datum; LAD – last appearance datum.
2000—2500 m depth; this assumption is supported by the finding of agglutinated foraminifers that are mainly related to the lower bathyal to abyssal marine setting. As few calcareous nanofossils are present, together with scarce macrofaunas (ammonites), we conclude that deposition took place near, but above the CCD.

Toward the lowest part of the section, a short positive δ13Corg excursion was observed in the dark grey shales, indicating a brief period of increased carbon burial and dysoxic bottom water. This scenario is supported not only by geochemical fluctuations, i.e., increase in δ13C values and OM, but also by biotic features such as a reduction in agglutinated foraminiferal diversity, and especially of epifaunal taxa related to the decrease in the oxygen content. The weak positive δ13Corg excursion identified in the section is probably the regional expression of the OAE1d. A weaker positive δ13Corg excursion is observed towards the top of the section, where the agglutinated foraminiferal assemblages display the same fluctuation in abundance and diversity as in the OAE1d interval.

The changes in lithology, together with the recovery of macrofaunas towards the upper part of the section indicate the replacement of the late Albian anoxic/dysoxic environment with an oxic one within the Albian–Cenomanian boundary interval. This change is possibly related to climate fluctuations. 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 Fe2O3 with Al2O3, SiO2 and TiO2.

This palaeoenvironmental fluctuation towards the end of the Albian could also have been related to the middle Cretaceous tectonic movements that strongly affected the whole Eastern Carpathian belt. As a result, the restricted circulation of small silled basins of the Moldavian Trough shifted to an open circulation regime in the late Albian—early Cenomanian interval.

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Appendix A. Supplementary data

Supplementary data related to this article can be found at http://dx.doi.org/10.1016/j.cretres.2014.10.010.