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Palaeoenvironmental controls of the distribution of organic matter within a C_{org}-rich marker bed (Faraoni level, uppermost Hauterivian, central Italy)

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Key-words: Black shales, oxygen-deficient environments, organic geochemistry, palynofacies, Lower Cretaceous, Umbria-Marche Basin

ABSTRACT

The Faraoni Level is an uppermost Hauterivian marker-bed within the Umbria-Marche Cretaceous sequence, characterised by an ammonite-rich limestone bed sandwiched between black shales. This marker level displays a remarkable similarity of facies and thickness at a regional scale. Mineralogical and geochemical data establish that the black shales contain up to 25 wt.% organic carbon (average 8%). Geochemical and palynological data suggest that the organic matter is of Type II, the amorphous material being the dominant component.

The regular and widespread bedding pattern of the Faraoni Level is interpreted as due to high frequency changes, possibly related to climatic oscillations, that affected the entire Umbria-Marche Basin. These changes probably induced fluctuations in the intensity of oxygen depletion on the basin floor. Medium levels of fluorescence of the amorphous organic matter, as well as the medium S/C ratio, argue for a dysoxic depositional environment rather than an entirely anoxic one.

In contrast to the facies and thickness consistency of the Faraoni Level, this marker horizon displays significant lateral variations in the relative abundance of the various components in the individual black shale layers, particularly its organic carbon content. This lateral variability is interpreted as the result of local palaeogeographic –mainly topographic– factors which controlled the intensity or duration of oxygen-deficient conditions on the sea floor of the Umbria-Marche Basin.

RESUME

Le Niveau Faraoni est un niveau repère de l'Hauterivien supérieur au sein de la succession pélagique du bassin des Marches-Ombrie (Italie centrale). Il est caractérisé par un banc calcaire très riche en ammonites, encadré par plusieurs black shales. Ce niveau repère présente une remarquable homogénéité de faciès et d'épaisseur à l'échelle régionale. Les données minéralogiques et géochimiques révèlent que les black shales contiennent jusque 25% de carbone organique, avec une moyenne de 8%. Les données géochimiques et palynologiques suggèrent que la matière organique est de Type II (phytoplanctonique); le matériel amorphe étant le constituant dominant des palynofaciès.

La stratonomie régulière et la grande étendue régionale du Niveau Faraoni est interprétée comme le résultat de changements environnementaux à haute fréquence, très probablement d'origine climatique, qui ont affectés le bassin des Marches-Ombrie et peut-être l'ensemble de la Téthys méditerranéenne à la fin de l'Hauterivien. Ces changements ont probablement induit des fluctuations dans l'intensité de la déficience en oxygène des eaux de fond, favorisant ainsi la préservation de la matière organique. La fluorescence moyenne de la matière organique amorphe, ainsi que le rapport S/C modéré, plaident en faveur d'un environnement dysoxique plutôt qu'anoxique dans le bassin des Marches-Ombrie.

Malgré la constance des faciès et des épaisseurs du Niveau Faraoni, quelques variations latérales sont notées au sein des coupes étudiées, notamment dans la proportion de matière organique de tel ou tel black shale. Ces variations latérales sont interprétées comme le résultat de facteurs paléogéographiques locaux, notamment topographiques, qui ont contrôlé l'intensité ou la durée des conditions dysoxiques sur le fond du bassin des Marches-Ombrie.

1. Introduction

During the Cretaceous a combination of factors, such as high primary productivity and/or good preservation, led to the widespread deposition of C_{org}-rich sediments in a large variety of depositional settings. In the pelagic Umbria-Marche Basin (central Italy), characteristic C_{org}-rich levels occur within the Cretaceous succession that are recognisable at regional to global scale (Arthur & Premoli Silva 1982; Coccioni et al. 1987, 1989; Cecca et al. 1994). These C_{org}-rich marker beds are as follows, from oldest to youngest (Fig. 1): Faraoni Level in the uppermost Hauterivian of the Maiolica Formation, the Selli, 113, and Urbino Levels in the Aptian-Albian of the Marne a Fucoidi Formation, and the Cenomanian-Turonian boundary Bonarelli Level of the Scaglia Bianca Formation.

These levels usually display broad ranges in organic con-

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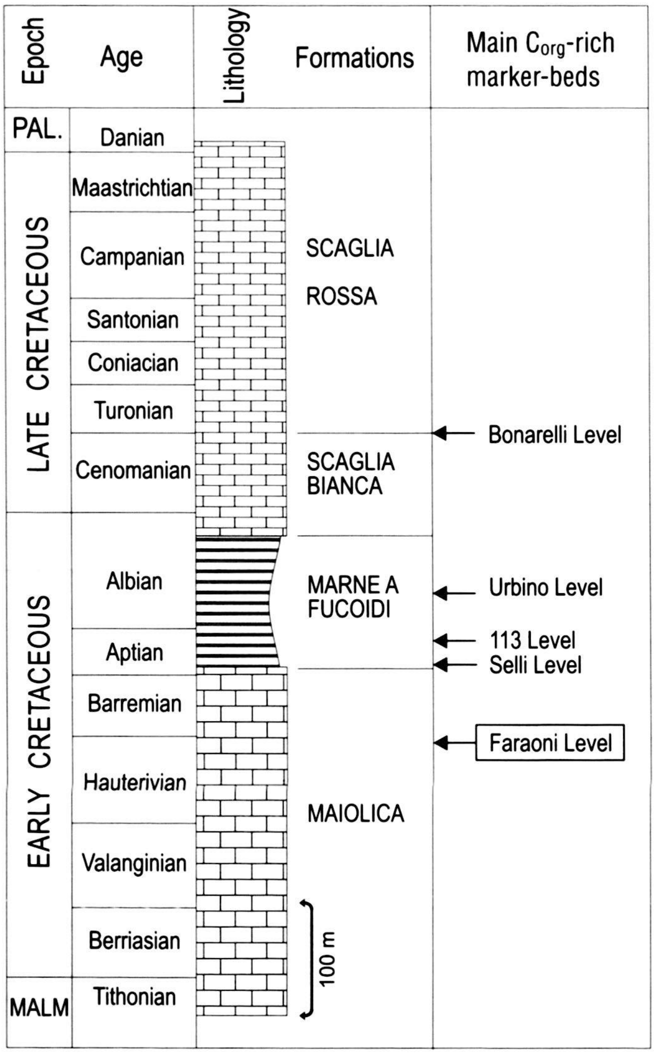

Fig. 1. Stratigraphic distribution of the major organic carbon-rich marker-beds within the Cretaceous succession of the Umbria-Marche Basin, including the Faraoni Level.

tent on a regional scale, as wide as 0.5 to 25% for one given level (see Baudin et al. 1998a for a short review). Vertical variations of organic carbon content are obvious and usually explained by changes in productivity and/or preservation conditions through time (Weissert 1981; Arthur & Premoli Silva 1982; Coccioni et al. 1987; 1989; Cecca et al. 1994; Baudin et al. 1998a). Lateral variations are also obvious, but rarely illustrated and poorly explained. Determining the lateral variation in organic matter in a single bed is also important as it may serve to indicate how representative an individual section is of a wider setting and may assist in the reconstruction of the palaeoenvironment and palaeogeography.

This paper documents the vertical and lateral variations of the Faraoni Level in terms of organic matter content, and discusses the palaeogeographic control of these variations.

2. Palaeogeographical setting

The Umbria-Marche Basin was part of the Mediterranean Tethys (Dercourt et al. 1993), particularly of the so-called Adria promontory (Channell et al. 1979) or Apulian block (Dercourt et al. 1985), which is described as a northern promontory of the African continent. Early Jurassic rifting produced numerous isolated and shallow Bahamian-type carbonate platforms surrounded by pelagic basins (Bernoulli 1972; Bernoulli & Jenkyns 1974). The Umbria-Marche region corresponds to a pelagic basin, whose palaeotopography was irregular due to the occurrence of numerous pelagic seamounts and surrounding furrows (Bernoulli 1972; Cecca et al. 1990; Santantonio 1993).

The palaeotopographic gradients, inherited from the Jurassic, persisted in the Early Cretaceous (Lowrie & Alvarez 1984), although they were less pronounced and slowly attenuated by the draping and smoothing pelagic sedimentation (Centamore et al. 1971). Successions 30–80 metres thick are recorded above Jurassic swells whereas 450 metres in thickness are commonly measured in areas of Jurassic furrows. The Lower Cretaceous succession is mainly represented by the Maiolica Formation, which ranges from the uppermost Tithonian to the lowermost Aptian (Fig. 1). It consists mainly of rhythmically bedded whitish micritic limestones and chert nodules or layers. Thin black shales appear in the upper Hauterivian, and their frequency and thickness increase markedly towards the contact with the overlying Marne a Fucoidi Formation (Arthur & Premoli Silva 1982). One of these prominent black shale intervals is the Faraoni Level, deposited in the latest Hauterivian.

3. Lithostratigraphical description of the Faraoni level

The Faraoni Level is characterised by an ammonite-rich bed sandwiched between cm-thick black shales (Cecca et al. 1994). It has been identified in 26 sections in the Umbria-Marche Basin (Fig. 2), where it shows the same lithological and stratigraphic organisation. Its lower boundary is generally characterised by the occurrence of a more or less continuous black chert layer. Its thickness ranges from 25 to 42 cm and it generally exhibits seven layers, lettered A to G (Fig. 3). They are from base to top:

A 2–6 cm of laminated black shale with marcasite nodules directly situated above the black chert,
B 1–6 cm of micritic limestone, locally containing ammonites,
C 0.5–1 cm of laminated black shale,
D 18–20 cm of micritic limestone with abundant ammonites. This bed is a marker-bed in all the outcrops. It commonly contains marcasite nodules which produce a characteristic red-coloured weathering. The ammonite fauna indicate the *Pseudothurmannia catulloi* Subzone from the upper part of the *Pseudothurmannia angulicostata* Zone which is the last Hauterivian ammonite zone. The microfacies is charac-

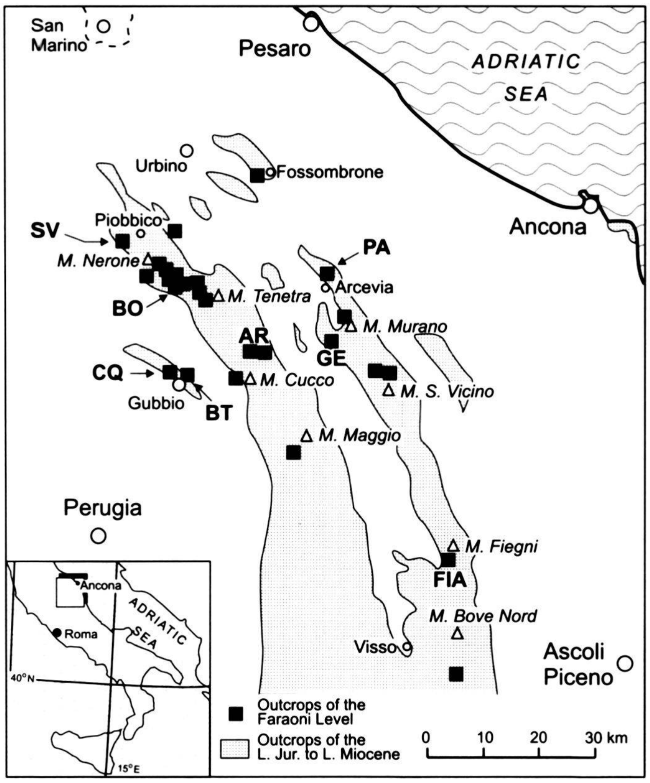

Fig. 2. Location of Faraoni Level outcrops and studied sections. AR: Arceviese road (km 59.4); BO: Bosso; BT: Bottacione; CQ: Contessa quarry; FIA: Lago del Fiastrone; GE: Genga; PA: Palazzo d'Arcevia; SV: Sette Vene (Apecchiese road km 33.7).

terised by abundant radiolarians and small-sized globular planktonic foraminifera (*Gorbachikella* spp., Coccioni et al. 1998),

E 1–2.5 cm of laminated black shale,

F 1–5 cm of micritic limestone which may occur as thin lenses in some sections,

G 1.5–3.5 cm of laminated black shale, with marcasite nodules.

Macroscopic observations establish a remarkable stratigraphic and lithologic similarity for sections located 85 km apart (Fig. 3). Nevertheless, minor lateral differences exist within the succession of the Faraoni Level (Cecca et al. 1994). Those with reference to the section studied here are as follows:

In the Arceviese road section, bed A reaches its maximum thickness of about 10 cm and shows a limestone layer intercalated between two laminated black shale layers. This limestone layer presents an undulated base and contorted structures, indicating that it is redeposited.

The middle part of bed F contains a thin film of black shale in the Palazzo d'Arcevia section.

The bed F is absent in the Genga section.

The upper boundary of the Faraoni Level usually coincides with the uppermost black shale layer. However, this bed (G) is reduced by compaction and/or stylolitisation to a thin film in the Sette Vene section.

4. Material and methods

Fifty-seven samples were obtained from 8 of the 26 sections where the Faraoni Level has been identified (Fig. 2). Samples were open-air dried and crushed in an agate mortar. Their carbonate content was measured using a calcimetric bomb. Total

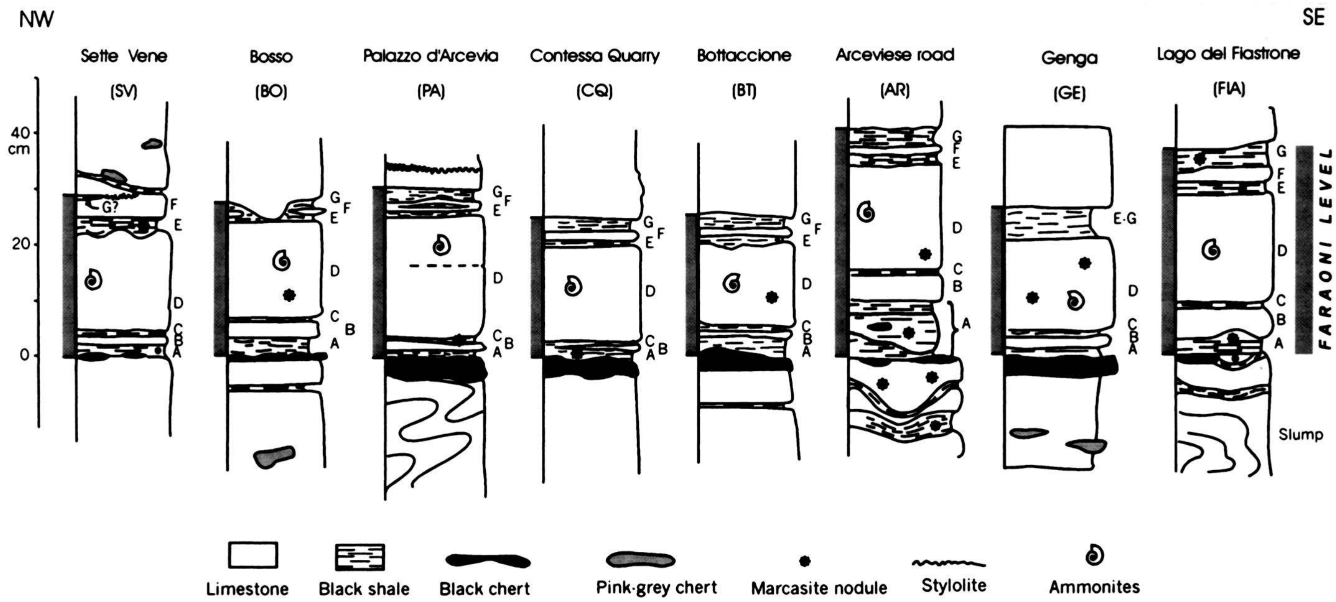

Fig. 3. Lithological profiles of the studied sections. See the text for a description of beds A-G.

		ORIGIN	GROUP	CONSTITUENT
CONTINENTAL	Allochthonous	Higher-plant debris	Phytoclasts	Palynomacerals
		Pollen & spores	Sporomorphs	Bisaccate pollen
				Non saccate pollen & spores
		Degraded higher plant	Amorphous OM	Non fluorescente AOM
MARINE	Autochthonous	Degraded phytoplankton		Fluorescent AOM
		Marine phytoplankton		Dinocysts & acritarchs
				Other marine algae
		Foraminifera		Chitinous foram. linings

Fig. 4. Palynofacies classification used in this study (modified from Steffen & Gorin, 1993)

carbon and sulphur contents were measured using a LECO IR-112 analyser. Total organic carbon contents were calculated by difference, assuming that all the carbonate is pure calcite. Rock-Eval pyrolyses were performed on a RE II instrument according to the method described by Espitalié et al. (1985–86). Standard notations are used: $CaCO_3$, total sulphur (S) and total organic carbon (TOC) contents are expressed in weight %, whereas Tmax is expressed in °C. The hydrogen index (HI) and oxygen index (OI) values are expressed in mg HC per g of TOC and mg CO_2 per g of TOC, respectively. Results are given in Table 1.

Forty-three kerogens were concentrated by HF-HCl maceration. Some of them were selected and their elemental composition (C, H, O, N, S and Fe) measured using the method summarised by Durand & Monin (1980). Kerogen types were recognised using the classical van Krevelen diagram, which plots H/C *vs.* O/C (atomic ratios).

The overall kerogen composition was determined by palynological analysis using counts of 300 clasts per slide (*sensu* Tyson 1995). The scheme used to classify sedimentary organic particles in transmitted light microscopy is derived and simplified from that suggested by Whitaker (1984), which is quite efficient for palaeoenvironmental studies (e.g. Gorin & Steffen 1991; Steffen & Gorin 1993; Pittet & Gorin 1997; Bombardiere & Gorin 1998; Wood & Gorin 1998). Five groups of constituents are distinguished (Fig. 4):

Phytoclasts: comprise all opaque and translucent land-plant debris. Different subgroups can be observed within this fraction and they were distinguished during the counts. Nevertheless, no significant information can be extracted from such a subdivision, and only phytoclasts as a whole will be used in the diagrams illustrating the results.

Sporomorphs: comprise the land-derived pollen grains and spores.

Amorphous organic matter (AOM): comprise all particulate organic components that appear structureless at the scale of light microscopy. On the basis of optical observation, two types of AOM are distinguished in our samples: granular, orange to dark-brown translucent AOM, here described as clotted (*sensu* Combaz 1980); and fluffy AOM, yellowish to dark-brown, here named spongeous.

Marine phytoplankton: comprise all dinoflagellate cysts (or dinocysts), acritarchs, and other marine algae. The latter two groups being very rare, they have been always grouped with the dinocysts.

Foraminifera: comprise the chitinous linings of some foraminifera, mostly from benthic forms (Tyson 1995).

5. Results and discussion

5.1 Sediment composition and its vertical and lateral variations

The studied samples can be regarded as a mixture of four components: carbonate, silicate (including clay minerals and biogenic silica), iron-sulphide (mainly pyrite and marcasite that have not been differentiated), and organic matter. The concentration of organic carbon describes the quantity of organic matter, although it should be kept in mind that organic carbon may represent only 60–75 wt.% of the total organic matter.

The limestone layers consist mainly of calcium carbonate (96–100%), whereas the organic carbon content is low (< 0.05 %). The remainder (< 4%) is represented by clay minerals and biogenic silica, derived from radiolarian tests. The black shale layers also have a high but variable carbonate content. The $CaCO_3$ content ranges between 0 and 89%, with a mean value around 50%. Their total organic carbon content is also highly variable, ranging from 1 to 25 % with an average around 8% TOC. The remainder consists of clay minerals, biogenic silica and iron-sulphide.

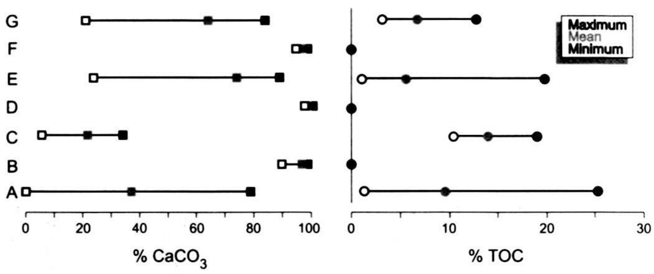

Fig. 5. Lateral and vertical variation in the carbonate and organic carbon content within single beds of the Faraoni Level.

Section Bed	Lithology	$CaCO_3$ %	S %	TOC %	Tmax °C	HI	OI
Arceviese road (km 59.4)							
G	Brownish laminated shales	54	0.96	10.47	417	267	91
F	White micritic limestone	99	nd	0.05	-	-	-
E	Laminated black shale	88	0.48	3.26	415	308	79
D	White micritic limestone	99	nd	0.02	-	-	-
C	Laminated black shale	34	1.60	11.65	419	283	95
B	White micritic limestone	98	nd	0.03	-	-	-
A top	Brownish laminated shales	35	3.84	10.36	416	270	86
A middle	Gray laminated limestone	97	0.15	0.42	420	444	36
A silex	Black chert	5	nd	0.27	420	288	18
A base	Laminated black shale	80	0.21	3.84	417	440	46
Bosso							
G	Brown laminated shale	73	0.38	3.45	432	281	85
F	White micritic limestone	97	nd	0.04	-	-	-
E	Brown laminated shale	89	0.05	1.04	432	255	106
D	White micritic limestone	99	nd	0.02	-	-	-
C	Dark brown laminated shale	6	0.70	11.21	433	283	85
B	White micritic limestone	99	nd	0.02	-	-	-
A	Brown laminated shale	75	0.08	2.31	433	318	87
Bottacione							
G	Brown laminated shale	81	0.32	5.27	419	540	22
F	White micritic limestone	99	nd	0.03	-	-	-
E	Brown laminated shale	86	0.17	3.93	421	510	18
D	White micritic limestone	100	nd	0.01	-	-	-
B	White micritic limestone	98	nd	0.02	-	-	-
A	Brownish laminated shale	36	0.05	1.36	421	463	65
Contessa Quarry							
G	Laminated black shale+limestone	80	0.40	5.76	420	544	19
F	White micritic limestone	99	nd	0.02	-	-	-
E	Laminated black shale+limestone	84	0.25	4.34	420	548	21
D	White micritic limestone	99	nd	0.03	-	-	-
C	*Brown laminated shale*	32	0.85	10.53	423	552	24
B	White micritic limestone	99	nd	0.02	-	-	-
A	Brown laminated shale	32	0.12	1.63	422	582	45
Genga							
E-G	Brown laminated shale	73	0.65	6.92	411	555	38
D	White micritic limestone	99	nd	0.02	-	-	-
C	Laminated black shale	14	1.45	17.24	412	506	30
B	White micritic limestone	98	nd	0.03	-	-	-
A	Laminated black shale	0	1.22	21.09	411	501	17
Lago del Fiastrone							
G	Brown laminated marlstone	84	0.14	3.22	423	345	73
G base	Black coat	nd	nd	5.72	433	281	59
F	White micritic limestone	99	nd	0.02	-	-	-
E	Brown laminated shale	82	0.13	2.92	427	315	79
D	White micritic limestone	100	nd	0.00	-	-	-
C	Laminated black shale	20	1.10	14.06	423	422	84
B	White micritic limestone	99	nd	0.02	-	-	-
A	Brown laminated shale	52	0.75	10.27	417	392	65
Palazzo d'Arcevia							
G	Brown laminated shale	56	0.49	6.37	429	297	88
F middle	Laminated black shale	35	0.35	4.65	431	244	94
F base	White micritic limestone	99	nd	0.02	-	-	-
E	Brown laminated shale	53	0.46	6.62	427	279	92
D	White micritic limestone	98	nd	0.05	-	-	-
B	White micritic limestone	96	nd	0.02	-	-	-
A	Brown laminated shale	0	0.55	9.83	425	275	102
Sette Vene (Apecchiese road km 33.7)							
G ?	Laminated black shale	21	5.56	12.81	418	469	37
F	White micritic limestone	96	nd	0.05	-	-	-
E	Laminated black shale	24	3.20	19.97	413	516	31
D	White micritic limestone	99	nd	0.03	-	-	-
C	Laminated black shale	23	5.80	19.10	413	503	33
B	White micritic limestone	98	nd	0.03	-	-	-
A	Laminated black shale	23	3.37	25.33	415	582	26

Note : nd: not determined and - no meaning

Tab. 1. Data on carbonate and sulphur content, and Rock-Eval pyrolysis of the Faraoni Level.

As expected, the vertical variations in bulk mineralogy of the Faraoni Level follow the primary lithological contrast of the rhythmic black shale-limestone interbedding (Fig. 5 and Tab. 1). It appears that bed C is depleted in carbonate (21% in average) compared to the other black shales layers (averaging 37, 74 and 64% for beds A, E and G respectively). Conversely, bed C is enriched in organic carbon (average 14%), whereas bed E is organic-lean (average 5.7%). However, this general trend is not observed in each section. For example, beds C and E show an equal organic carbon content (up to 19%) in the Sette Vene section (Tab. 1). At a basin scale, this section is singular by its richness in organic matter compared to the other sections. All the black shale layers are carbonate-poor (< 25% $CaCO_3$) and display the highest TOC values measured in the different sections (Tab. 1).

The lateral variations of the relative components of black shale layers are important as well (Fig. 5). The range in TOC and carbonate is wide for bed A and narrow for bed C. Bed E shows also a wide range in TOC compared to bed G.

5.2 Relationship between organic carbon and other components

The carbonate content in the Faraoni Level reflects mainly the relative abundance of carbonate micro- and nannofossils, more particularly of planktonic foraminifera and coccoliths (Coccioni et al. 1998). The TOC-$CaCO_3$ crossplot shows a negative correlation for carbonate content above 40% (Fig. 6). Such a negative relationship can be interpreted as reflecting the dilution of both organic matter and terrigeneous clays by carbonate input (Ricken 1993). The variations in carbonate input may be due to variations in primary productivity or dissolution, both of which repeatedly diluted or concentrated the organic

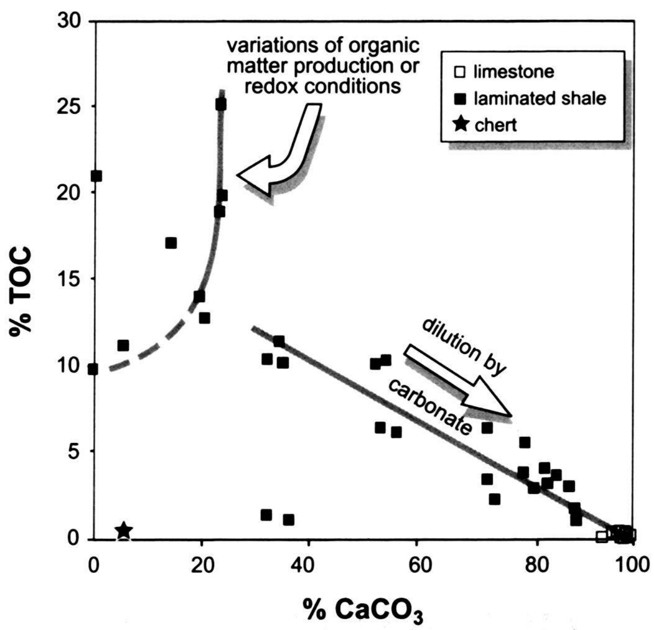

Fig. 6. Carbonate *vs.* organic-carbon content within the Faraoni Level.

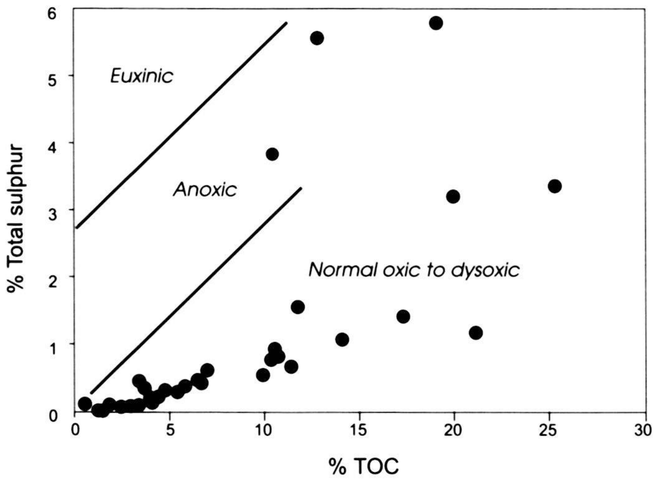

Fig. 7. Sulphur *vs.* TOC content of the black shales from the Faraoni Level in a Berner & Raiswell (1984) diagram.

carbon of the sediment. The relationship is less clear for carbonate content below 40%, as wide variations are noted in the organic carbon content (Fig. 6). This may reflect either occasional prolific algal blooms or fluctuations in the redox conditions, during periods of overall low carbonate productivity.

Three samples have an unusual position in this diagram. One corresponds to a black chert from the middle part of bed A in the Arceviese road section. As expected, this sample has low TOC and $CaCO_3$ contents. The two others correspond to bed A in the Bottaccione and Contessa Quarry sections. They contain only 1.3 to 1.6% TOC, for a carbonate content not higher than 36%. This depletion in organic matter compared to other samples with the same carbonate content is not fully understood, but may result from recent weathering. Nevertheless, it should be noted that bed A lies directly above the chert layer, and shows the most extreme concentrations for individual components: 0 to 80% for $CaCO_3$ and 1.3 to 25% for TOC.

The total sulphur content of the black shale layers varies between 0.1 and 5.8% (Tab. 1). According to elemental analyses of kerogen concentrates, this sulphur is essentially associated with iron sulphides, which are the products of syndepositional bacterial sulphate reduction in interstitial anoxic waters. Sulphate reduction is only sustained as long as sufficient metabolizable organic matter and sulphate are available. Furthermore, formation of pyrite depends on the availability of reactive iron. These processes are well established by Leventhal (1983) and Berner & Raiswell (1984), and sulphur-organic carbon relationships have been widely used for the interpretation of the oxygen content of bottom water in ancient depositional environments. The sulphur-TOC relationship for the Faraoni Level exhibits a positive correlation with a low gradient, suggesting dysoxic rather than anoxic conditions for most sections (Fig. 7). Only few samples from the Sette Vene section are located in the anoxic part of the diagram.

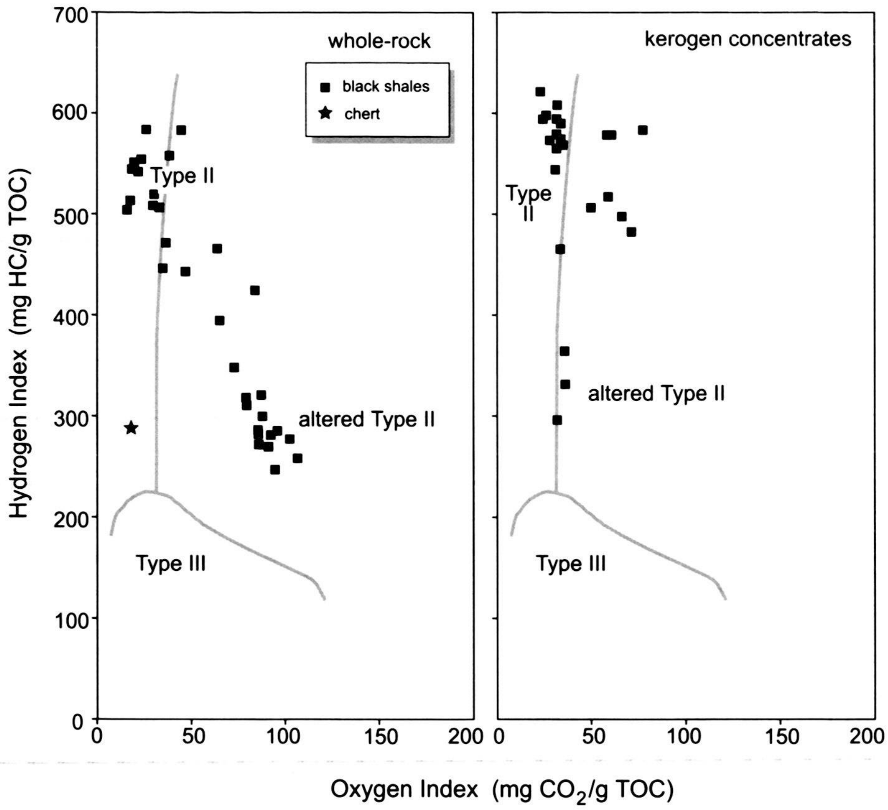

Fig. 8. Hydrogen *vs.* oxygen index diagram of the organic-rich samples and kerogen concentrates from the Faraoni Level.

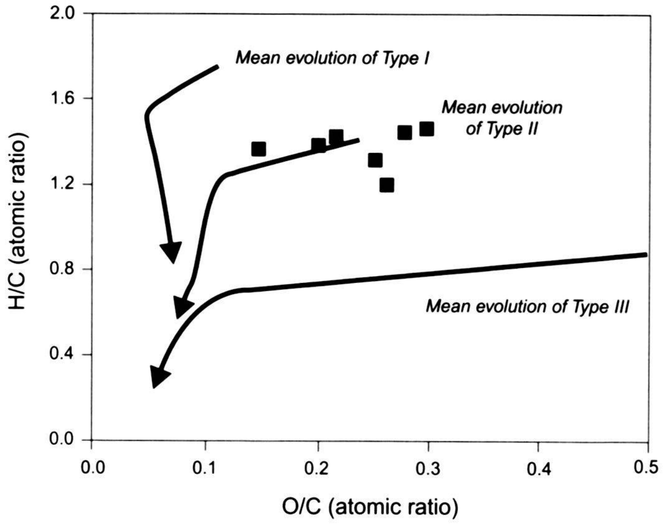

Fig. 9. Van Krevelen diagram with selected samples from the Faraoni Level.

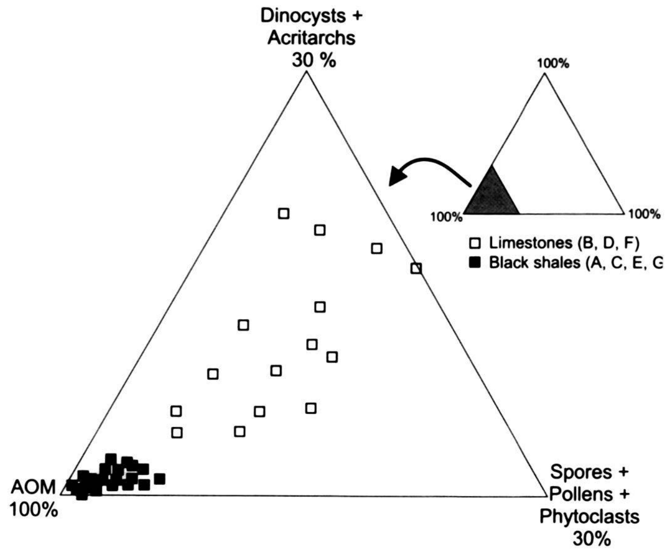

Fig. 10. Ternary diagram of palynofacies components from black shales and limestone samples from the Faraoni Level. Note that black shales are enriched in sporomorphs, whereas limestones are enriched in dinocysts.

5.3 Organic matter source

Temperatures of maximum pyrolytic yield (Tmax) for samples having an organic richness higher than 0.25% are in the range of 411–433 °C with an average around 423 °C. This indicates that the organic matter did not experience high temperatures during burial. All studied sections are located below or at the beginning of the oil-window regarding petroleum generation. Consequently, the source of organic matter from the Faraoni Level can be estimated from pyrolysis data on the basis of hydrogen and oxygen index-values (Espitalié et al. 1985–86).

Hydrogen index-values range from 240 to 580 mg HC/g TOC, whereas oxygen index-values range from 20 to 100 mg CO_2/g TOC (Tab. 1). In the HI-OI diagram (Fig. 8A), the organic-rich samples are clearly located in the area of Type II organic matter. However, many grey to brown laminated samples have HI-values ranging between 400 and 250 (mean value of 370), suggesting a mixture of Type II and Type III organic matter or selective degradation of marine material. The relatively high OI-values of these samples is consistent with the oxidation of the organic matter.

Nevertheless, low HI-values and high OI-values can be due to mineral matrix effect (e.g., adsorption on clay mineral surfaces of hydrocarbons generated during pyrolysis and production of CO_2 due to carbonate cracking, Espitalié et al. 1985–86). Most of the kerogen concentrates prepared for palynological investigation were analysed using Rock-Eval. HI-values for kerogen concentrates consistently reveal a slight increase compared to HI-values for whole rock, but a good correlation exists between the two (Fig. 8B). The OI-values of kerogen concentrates are lower than those for whole rock sample, but they confirm that some samples contain an altered Type II organic matter (Fig. 8).

Elemental analysis of 7 kerogen concentrates reveals that the organic carbon represents 60 to 70% of the total organic matter. H/C ratios range between 1.15 and 1.41, whereas O/C ratios range between 0.15 and 0.3. Most of the kerogen concentrates cluster close to path II in the van Krevelen diagram (Fig. 9). This confirms the marine origin of the organic matter for the Faraoni Level.

5.4 Palynofacies characterisation

Palynofacies of the black shales are dominated by amorphous organic matter (AOM) which constitutes more than 95% of the organic particles (Fig. 10). The AOM is also dominant in the limestone layers, making from 70 to 91% of the organic particles. Clotted AOM is always prevalent in both black shale and limestone layers. Nevertheless, small quantities of spongeous AOM (up to 5%) were observed in some samples, especially at the base and middle part of the redeposited bed A of the Arceviese road section and in many limestones beds. A large part of clotted AOM shows weak fluorescence and spongeous AOM is non-fluorescent. Qualitative variations in AOM fluorescence are provided for the Arceviese road section, for which a detailed sampling was carried out (Fig. 11). As the thermal maturation of the black shales is low, the weak fluorescence of AOM is probably a result of oxidation of the organic matter during settling. The absence of fluorescence of spongy AOM which predominates in organic-poor samples, equally supports this hypothesis.

The very high abundance of AOM in the studied material makes it difficult to correctly evaluate the origin of the organic matter. Any trend in variation in the relative abundance of

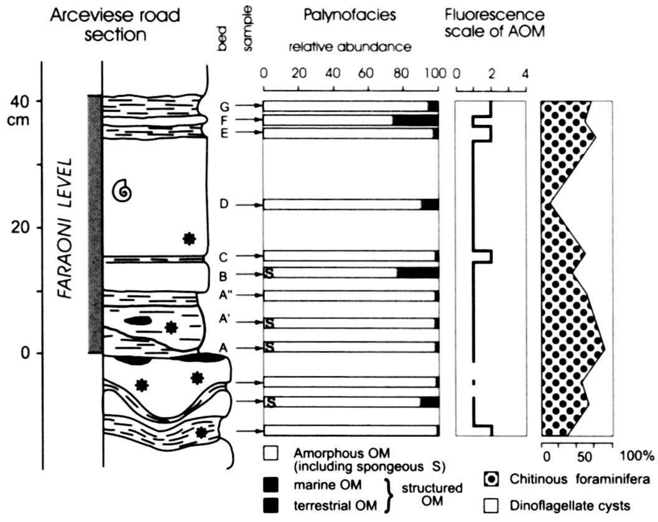

Fig. 11. Vertical variations of palynofacies, fluorescence of amorphous organic matter, and relative abundance of selected palynomorphs around the Faraoni Level of the Arceviese road section.

structured organic components (dinocysts, sporomorphs and phytoclasts, for which the marine or terrestrial origin can be easily inferred), would therefore be obscured by the amount of AOM in the samples. However, some considerations can still be made regarding the origin of AOM from the Faraoni Level. Among structured organic components, phytoclasts are never an important constituent of the palynofacies (<5%). Those found in the studied samples are very small (< 50 μm), indicating a size-sorting typical for an environment situated far from the source of terrigeneous input. This would indicate a marine origin for the AOM of the black shales from the Faraoni Level, which is in agreement with the geochemical data.

By contrast, higher percentages of marine palynomorphs are observed in the limestone layers where lower concentrations of AOM have been recorded (Fig. 11). Higher relative abundance of dinocysts characterises the carbonate-rich layers, whereas black shale layers display higher abundance of sporomorphs (spores, non saccate and bisaccate pollen grains). The different palynomorph content of the carbonate-poor versus carbonate-rich layers reflects probably different hydrodynamic regimes. High abundance of foraminiferal linings were recorded in the black shale layers; this group sometimes dominates the palynomorph assemblage. Melia (1984) reported higher abundance of foraminiferal linings (800 specimens per gram of dry sediment) offshore West Africa, concentrated along the upwelling area. This might be an indication of higher primary productivity during the deposition of the black-shale layers. However, in ancient pelagic sediments, significant abundance of foraminiferal linings appear to be strongly correlated with redeposition (Tyson 1984). The significance of such abundant foraminiferal linings in the studied material is still not entirely clear and it has been postulated that they might represent an adaptation of certain foraminifera, not forming mineralised parts of their test into an oxygen-depleted environment (Galeotti 1995).

6. Depositional controls of the Faraoni level

Interpretation of what chiefly contributes to the pelagic black shales/limestone rhythms is a complex problem (de Boer 1991; Arthur & Dean 1991; van Buchem et al. 1995 and references therein). In the case of Faraoni Level, there is evidence of high-frequency environmental changes that led to the interbedding of limestone and black shale layers. A simple explanation is provided by the theory of astronomically induced climatic changes. Such forcing mechanisms have been recognized throughout the Cretaceous succession of Umbria-Marche Basin, more particularly in the Maiolica Formation (Fischer & Arthur 1977; de Boer 1991).

The abundance and diversity of fossils found in the limestone layers of the Faraoni Level suggest nutrient-rich surface waters (Cecca et al. 1994; Coccioni et al. 1998). No clear evidence exists, so far, indicating whether productivity was indeed higher during the deposition of black shale, which stored abundant organic matter. Nevertheless, it is very likely that the difference in the palynological composition between black shale and limestone layers reflect changes of hydrodynamic and nutrient regimes. Different mechanisms may be responsible for the observed differences in palynomorphs between carbonate-poor and carbonate-rich layers. On one hand, higher quantities of dinocysts in the carbonate-rich layers might simply reflect dinoflagellates blooms during intervals of limestone deposition. On the other hand, relative abundance of sporomorphs in the black shales may also coincide with periods of higher continental runoff or enhanced winds, supplying the Umbria-Marche Basin with clay minerals and terrestrial organic residues. This explanation has already been assumed for the deposition of Aptian-Albian black shales within the Marne a Fucoidi Formation (Pratt & King 1986; Hochuli et al. 1999).

Whatever the forcing mechanism, the organic-matter flux determines oxygen consumption in the water column. In a situation close to an equilibrium between the supply of oxygen and of organic matter to deep waters, small shifts of external variables may cause a cyclic pattern of black shales and oxygenated limestones. It is clear from the lithological, geochemical and palynological data that oxygen-deficient conditions were developed on the bottom of the basin during the deposition of the black-shale layers. However, the medium fluorescence of amorphous organic matter, as well as the S/C ratio, argue for a dysoxic depositional environment rather than an entirely anoxic one.

To summarise, the vertical trend of the Faraoni Level is probably the result of high-frequency climatic variations, which modulate the organic-matter production and preservation by pushing the sensitive sedimentary system over certain threshold conditions (Fig. 12). A stratified water column with stagnant dysoxic waters was probably developed during intervals of changing circulation patterns or fresh-water inputs. This

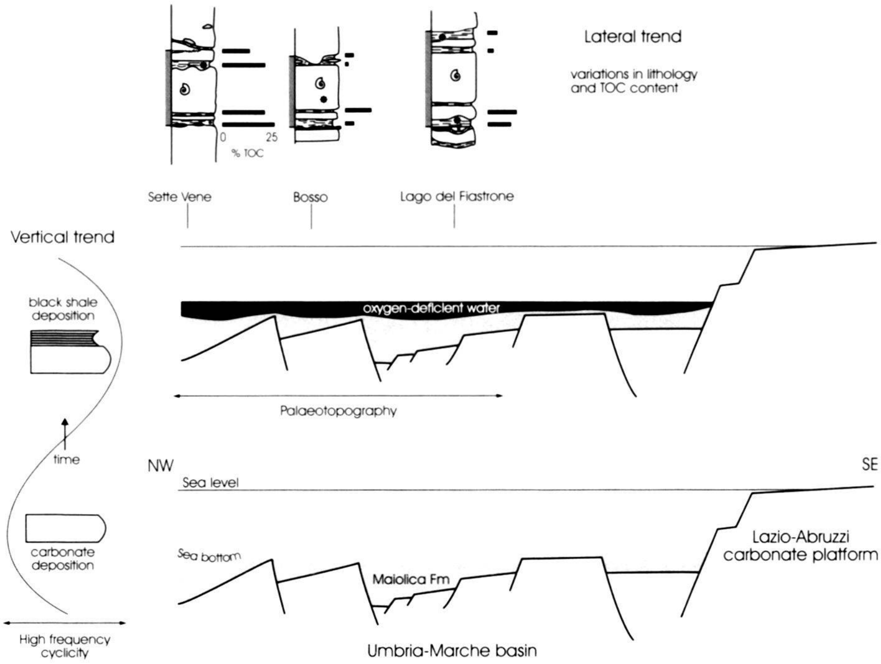

Fig. 12. Idealised restoration profile of the Umbria-Marche Basin during the Late Hauterivian and main factors controlling the deposition of limestone and black shale layers of the Faraoni Level.

phase of reduced oxygen supply to the sea floor have led to high TOC levels and good preservation of Type II organic matter, as observed in the black shale layers A, C, E and G of the Faraoni Level.

In addition, a lateral variation is superposed to the vertical one. Variations in sediment composition of individual black layers are obvious from section to section (Fig. 5). Such variations may be the result of i) differential weathering of outcrops, ii) variations in thermal diagenesis of organic matter or iii) lateral changes of depositional conditions.

Although the samples originate from natural outcrops, lateral variations do not seem to be the result of weathering. The similar composition of every black shale and limestone layers (Tab. 1) of the closely-spaced Contessa Quarry and Bottacione sections (Fig. 2) can simply not be the result of a random weathering process. A difference due to variation of thermal diagenesis is also unlikely, because the maturation stage is more or less the same throughout the Umbria-Marche Basin, all samples being immature with respect to the oil window.

These lateral variations rather correspond to primary difference during the time of deposition. Lateral disparities in sediment composition are common in turbiditic environments. As some pelagic black shales are regarded as fine-grained turbidites originating from the oxygen-minimum zone (Stow 1987), this could be also the case of the Faraoni Level. Although slightly expressed at the end of the Hauterivian, the irregular topography of the Umbria-Marche Basin probably determined a more or less undulated sea-floor. Such a configuration could be prone to erosion and resedimentation. Nevertheless, apart from the redeposited bed A in the Arceviese road section, nothing else indicates a turbiditic sedimentation for the black shales. Moreover, if some slumps exist below the Faraoni Level, the time of deposition of this marker-bed appears as tectonically quiescent within the Umbria-Marche Basin. It is more likely that the small undulations of the sea floor were suitable to low water recycling and hence to oxygen depletion. Preservation of organic matter varies accordingly; and its best quality and highest quantity occurs in the deeper parts of the various sub-basins, created as troughs since the Jurassic (Fig. 12).

It is likely that this depositional model is equally valid for other organic carbon-rich marker beds of the Umbria-Marche Basin. Similar high frequency pulses of oxygen-deficient conditions, partly controlled by paleotopography, were demonstrated for the upper Cretaceous black cherts of the Scaglia Bianca Formation (Beaudoin et al. 1996). They still need to be demonstrated for the so-called Selli, 113, Urbino and Bonarelli Levels (Fig. 1).

7. Possible causes of the "Faraoni Event"

In spite of its different lithological signature, the Faraoni Level has similar paleontological and geochemical signatures as the Selli and Bonarelli Levels, which are well established regional records of major palaeoceanographic events, the so-called Oceanic Anoxic Events (Schlanger & Jenkyns 1976, Jenkyns 1980, 1999). No OAE has been described from the uppermost Hauterivian, till now, but various types of evidence suggest the existence of, at least, a western Mediterranean palaeoceanographic event at that time. Precise equivalents of the Faraoni Level have been recognised in the Trento Plateau of southern Alps (Cecca et al. 1996, Faraoni et al. 1997, Baudin et al

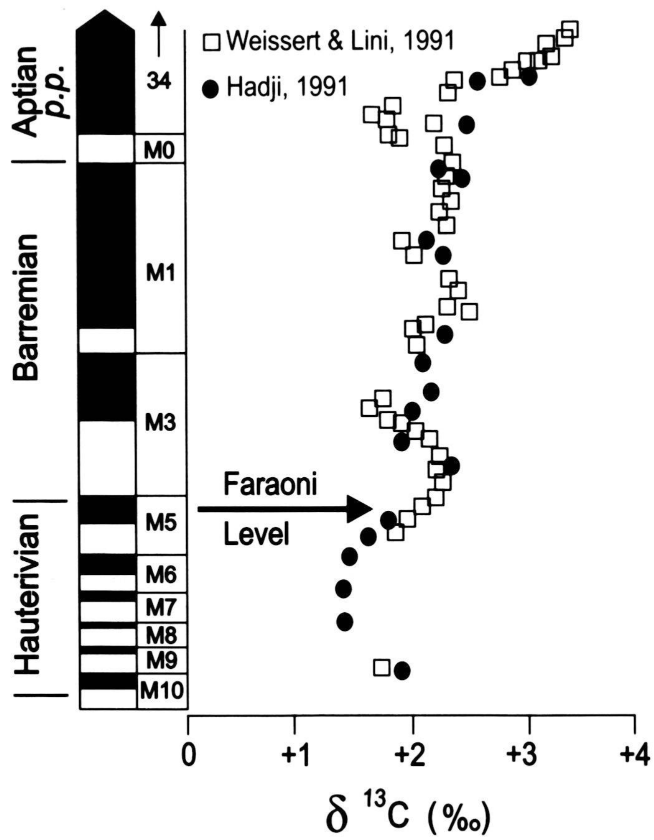

Fig. 13. Carbon-isotope curves for the Hauterivian-early Aptian interval of the Maiolica succession from northern and central Italy, showing the small positive excursion accompanying the Faraoni Level.

1998b), in the Vocontian Basin of Southeastern France (Baudin et al. 1999) and possibly in Subbetic Basin (work in progress). All of them display most of the palaeontological and geochemical features of the Faraoni Level of Umbria-Marche Basin, suggesting that the event causing this marker-bed is not limited only to the single Umbria-Marche Basin.

The faunal and nannofloral signatures of the Faraoni Level are quite similar, although less pronounced, compared to those of other OAE's. The Faraoni Level coincides with an increase in abundance of radiolaria and planktonic foraminifera (the small-sized globular *Gorbachikella* spp.), whereas benthic foraminferal abundance decreases drastically (Coccioni et al. 1998, Baudin et al. 1999). On the contrary, nannoconid abundance falls during the Faraoni Level, which is a common feature of other OAE's, and more particularly of the Early Aptian OAE 1a (Erba 1994, Erba et al. 1999).

The organic-matter enrichment of the Faraoni Level is of the same order as that of the Selli, Urbino and Bonarelli Levels and displays the same palynological and organic geochemical signatures (see Baudin et al. 1998a for a short review).

A small positive excursion of $\delta^{13}C$ (0.5 ‰) is recorded in the Umbria-Marche Basin (Hadji, 1991) and in the southern Alps (Wiessert et al. 1985, Weissert & Lini, 1991) close to the stratigraphic position of the Faraoni Level (Fig. 13). Interpretations of carbon-isotope anomalies in the Cretaceous have sought to correlate positive $\delta^{13}C$ excursion in carbonates to unusual storage of organic carbon in marine environment (Scholle & Arthur, 1980, Weissert et al. 1985, Arthur et al. 1990, Weissert et al 1998), especially during time of OAE's (Schlanger et al 1987, Arthur et al 1990, Arthur & Sageman 1994) Following the same interpretation, the Late Hauterivian excursion recorded in Umbria-Marche and Southern Alps basins suggest an enhanced organic-carbon preservation at least at a regional scale.

Nevertheless, there is no exact stratigraphic correspondence between black-shale deposition and the carbon-isotope response. It seems that the small positive excursion mentioned above is registered immediately after the Faraoni Level (Fig. 13). A comparable time-lag is known between the Selli Level, which is C_{org}-rich and shows constant $\delta^{13}C$ values, and the subsequent Early Aptian carbon-isotope positive excursion (Bralower et al 1993, Marconi et al 1994, Menegatti et al 1998, Larson & Erba 1999). The isotopic uniformity during black-shale deposition may reflect an equilibrium between C_{org} burial and increased rate of ^{12}C recycling and nutrient-rich intermediate water resulting from intensification of oceanic thermohaline circulation. Moreover, a pronounced negative carbon-isotope excursion is locally recorded just before the Selli Level (Menegatti et al 1998) and interpreted as the release of methane (which is isotopically very negative) due to the dissociation of gas hydrates (Jahren et al. 2001). Such a negative excursion is not yet recorded just before or in coincidence with the Faraoni Level, but the Hauterivian isotopic record is less documented than that of the Barremian-Albian interval.

All these data suggest that the Faraoni Level may correspond to a short-lived oxygen-deficient event in the Mediterranean Tethys. However, in order to better constrain the palaeoceanographic significance of this event, high-resolution isotopic studies are necessary along sections where the Faraoni Level has been clearly identified.

An intense debate exist on the causal global mechanisms of OAE's. Many studies have recently focused on the mid-Cretaceous Period, because of the co-occurrence of several palaeobiological, palaeoclimatic and palaeocanographic events (Erbacher & Thurow 1997, Weissert et al 1998, Larson & Erba 1999, Jenkyns 1999, Jenkyns & Wilson, 1999, Sanfourche & Baudin 2001, Jahren et al 2001, Jones & Jenkyns, 2001). It is likely that mid-Cretaceous OAE's were associated with an increased flux of CO_2 into the atmosphere, induced by rapid seafloor spreading and emplacement of mid-oceanic plateaus, which may have produced warmer and humid conditions. The resulting greenhouse effect may have led to an intensified nutrient supply from continents to the ocean, and simultaneous thermohaline stratification of the water-column.

Although strong mantle dynamics apparently do not characterise Late Hauterivian times (Larson 1991), a similar concatenation of forcing mechanisms can be invoked to explain the Faraoni Level record. According to available data, this level originated as a consequence of increased primary productivity, although highly eutrophic conditions were not reached (Coccioni et al 1998). The high-productivity model requires an efficient circulation and nutrient recycling. This may be achieved by local upwelling installation, related to peculiar conditions such as circulation patterns and basin geometry. The patchy distribution of the Faraoni Level in Mediterranean Tethyan pelagic basins seems in agreement with local upwelling systems. The resulting higher organic carbon fluxes from surface waters would have been responsible for the oxygen-depletion of deep waters. However, another direct mechanism may also be inferred.

The black shales deposited during OAE'S are commonly related to transgressive events, especially with first- or second-order flooding surfaces, as recognised by Haq et al (1987) and Hardenbol et al (1998). Flooding of landmasses and creation of shelf seas may enhance productivity, with the expansion of oxygen-minimum zone as a result (Jenkyns 1980) and the drowning of carbonate platforms surrounding the pelagic basins where dysoxia/anoxia was being developed (Jenkyns 1991, Weissert et al 1998).

Indeed, according to the Hardenbol et al. (1998) eustatic chart, the *Pseudothurmania catulloi* ammonite subzone precisely corresponds to a major flooding surface (mfs Ha 6). It is quite likely that this sea-level rise provoked and sustained enhanced productivity on shelf seas, as well as the development of a condensed interval in offshore environments, with a resulting accumulation of great concentrations of organic matter (Loutit et al. 1988). Because this sea level rise is rapidly followed by a regressive trend and a sequence boundary in the final Hauterivian (Ha7 of Hardenbol et al. 1998), it could explain the lower thickness of the Faraoni Level compared to other Umbria-Marchean C_{org}-rich marker beds.

8. Conclusions

The uppermost Hauterivian Faraoni Level corresponds to an important episode of organic-matter preservation in the upper part of the Maiolica Formation of the Umbria-Marche Basin. This level is characterised by an alternation of organic-rich black shales with organic-poor micritic limestones, which can be traced all over the basin. This regular bedding pattern is interpreted as due to high-frequency climatic cycles that affected organic-matter storage on the sea-floor, by the development of high primary productivity and oxygen-deficient conditions. Superimposed on this general pattern, local morphological factors controlled the intensity of oxygen-deficient conditions on the bottom of the Umbria-Marche Basin, and then the local organic richness of the black shales.

It appears that several pelagic basins within the Mediterranean Tethys were prone to dysoxia/anoxia during the late Hauterivian, and recorded a possible short-lived anoxic event. It is likely that similar forcing mechanisms responsible for global OAE's, such as high sea level, increasing productivity and an overall warm climate, operated during this time interval. It is most likely that the local/regional triggering factors – mainly the palaeogeographic configuration of basins – in coincidence with a high sea-level interval, had a strong influence on the appearance and characteristics of the Faraoni Level.

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