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PALAEO
Palaeogeography, Palaeoclimatology, Palaeoecology 233 (2005) 163–186
www.elsevier.com/locate/palaeo

Mid-Cretaceous oceanic red beds in the Umbria–Marche Basin, central Italy: Constraints on paleoceanography and paleoclimate

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Received 13 January 2005; received in revised form 25 September 2005; accepted 3 October 2005

Abstract

Detailed studies of the mid-Cretaceous sedimentary strata in the Umbria–Marche Basin in Italy revealed that aside from several well-known organic-rich "black shale" horizons that record OAEs, several varicolored, mainly reddish horizons indicate oxic conditions at the ocean bottom. Eight such horizons have been identified in Aptian–Cenomanian sequences in the Umbria–Marche Basin. The dysoxic/oxic beds alternate regularly but not cyclically and seem to be the result of non-random processes. Their occurrence is not a completely random process. The duration of deposition of these oceanic red beds (ORBs) varies from ~ 0.13 my (ORB4) recorded in the *Ticinella primula* zone, to ~4.54 my for ORB1, which spans the *Globigerinelloides ferreolensis* zone to the *Ticinella bejaouaensis* zone in the Piobbico core.

Mid-Cretaceous ORBs are not a local phenomenon because they occur in the Tethyan deposits in the Southern and Austrian Alps, the Carpathians, the Northern Caucasus, the central North Atlantic, in northeastern England and in the western Himalayas. They provide evidence for periodic changes in redox conditions at the ocean bottom. Such changes could have been caused by changes in bioproductivity, basin geometry, sedimentation rates, paleocirculation and/or production of bottom waters with higher content of dissolved oxygen in response to changes in paleoclimate. We suggest that the periodic inflow of colder, more oxygenated bottom waters was the probable cause of ORBs development, either as a result of changes in the ocean bottom topography, or as a result of brief cool climate periods. However, reliable proxies for changes in deep ocean circulation are still lacking. If ORBs were the result of the paleoclimate and, therefore, indirectly caused by changes in CO_2 in atmosphere, they document the increased sensitivity and instability of the mid-Cretaceous climate. Thus, changes in ocean dynamics were on a scale of several hundred thousand years to several million years, which has not been considered by most theories of CO_2 cycling, mid-Cretaceous greenhouse paleoclimate and paleocean dynamics.

Keywords: Oceanic red beds; Mid-Cretaceous; Paleoceanography; Paleoclimate; Central Italy

1. Introduction

The mid-Cretaceous was a period of increased tectonic activity, changing paleogeography and increases in atmospheric CO_2 (Larson and Pitman, 1972; Ziegler et al., 1982; Larson, 1991; Larson and Erba, 1999), resulting in global warming that peaked in the early

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doi:10.1016/j.palaeo.2005.10.003

Turonian (Jenkyns et al., 1994; Barron et al., 1995; Huber et al., 1995, 2002; Clarke and Jenkyns, 1999). During this period (121–93.5 Ma, according to Gradstein et al., 1995, time scale) in the western Tethys and central North Atlantic basin, dark grey to black shales were deposited that indicate dysoxic to anoxic ocean-bottom conditions, which were a consequence of a "greenhouse world" (Weissert, 1991; Hochuli et al., 1999; Bice and Norris, 2002). Several events of increased organic matter deposition in the marine environment occurred during this time interval. These events, known as the oceanic anoxic events (OAEs), were apparently restricted to relatively short time intervals of a few million years and less (Jenkyns, 1980; Leckie et al., 2002, for review). The OAEs represent major perturbation of global climate–ocean system that led to diagnostic shifts in the carbon isotopic signature of sedimentary carbon and indicate cooling of the atmosphere via extensive CO_2 drawdown, in response to massive burial of marine organic carbon (Schlanger and Jenkyns, 1976; Arthur et al., 1987, 1990; Bralower et al., 1994; Kuypers et al., 2002). Among the plethora of black shale horizons identified both on land and in the oceans, two organic carbon-enriched horizons–the lower Aptian Selli level (OAE1a) and the uppermost Cenomanian, the Bonarelli level (OAE2)–stand out. Both have not only broad global distribution but are identified by $\delta^{13}C$ isotope positive excursions (e.g., Schlanger et al., 1987; Arthur et al., 1988; Menegatti et al., 1998; Hochuli et al., 1999). Most of the other OAEs were more local, with occurrences confined to the western Tethys–Atlantic domain and were less extreme in terms of global environmental impact (Leckie et al., 2002; Galeotti et al., 2003). Regardless of their origin, deposition of organic carbon ($C_{org.}$)-enriched deep-sea deposits indicates a major change in the mode of carbon burial in western Tethys, from an inorganic carbon buried as pelagic carbonates during Early Cretaceous to organic carbon burial in black shales during mid-Cretaceous. Such data have been frequently used for modelling the greenhouse climate that dominated the mid-Cretaceous (Barron, 1983; Barron et al., 1995; Bice and Norris, 2002).

During recent field work in the Umbria–Marche Basin in Italy, we noted the occurrence of several pinkish, reddish, or maroon bands and zones in the deep-sea mid-Cretaceous Marne a Fucoidi and Scaglia Bianca formations, suggesting variability in paleoceanographic conditions and perhaps indirectly in paleoclimate. Occurrence of some of these red beds was previously noted by Arthur and Fischer (1977), Arthur (1979), Erba (1988), Premoli Silva et al. (1989), Tornaghi et al. (1989), Coccioni et al. (1990), Erbacher (1994), Coccioni (1996), Fiet and Masure (2001), but not much attention was given to them, even though they are prominent in outcrops and cores. Only a brief note on the oxidation state of the Umbria–Marche Cretaceous succession was given by Arthur and Fischer (1977), who simply stated that "the transition from comparatively unoxidized Aptian–Albian sediments to more highly oxidized Cenomanian beds is of more than local significance and reflects worldwide changes in ocean dynamics." Premoli Silva et al. (1989) mentioned that the change in sediment color from grey-green to red and red brown in upper Aptian strata in central Italy reflects a change to a more vigorous circulation resulting from better oxygenated bottom waters.

If the appearance of these red beds is related to changes in the content of dissolved oxygen in bottom waters and, thus, indirectly result to paloclimate change, then the question to ask is—was the mid-Cretaceous "greenhouse" climate relatively stable, as implied by various studies (Douglas and Savin, 1973, 1975; Crowley and North, 1991; Frakes et al., 1992; Jenkyns et al., 1994; Barron et al., 1995; Huber et al., 1995; Bice and Norris, 2002; Otto-Bliesner et al., 2002), or was it much more dynamic with brief cool periods, like the oxygen isotope data suggest for the Early Cretaceous Tethys (Weissert and Lini, 1991; Hochuli et al., 1999; Jenkyns and Wilson, 1999; Puceat et al., 2003; Weissert and Erba, 2004) and sedimentary record may indicate?

This paper concentrates on reddish colored oceanic red beds (ORBs) located between the lower Aptian Selli level (OAE1a) and the uppermost Cenomanian Bonarelli level (OAE2) in the Umbria–Marche Basin in central Italy. We briefly discuss correlative strata in eastern European Tethys and Asia. For ORBs exposed in the Umbria–Marche Basin, we provide a brief lithologic, sedimentological, petrographic, and micropaleontologic composition; we synthesize available data to document their stratigraphic position and compare their isotopic signals and biotic content. We also studied similar mid-Cretaceous ORBs cropping out in other regions of Europe and in the Himalayas, where remnants of the Tethys are exposed, in order to assess the timing and geographical extent of these events and to better understand possible triggering mechanisms of major mid-Cretaceous paleoceanographic changes.

2. Methods

Field studies in the Umbria–Marche Basin document the occurrence of ORBs in the mid-Cretaceous strata.

The studied sections are located near the towns of Gubbio and Piobbico. Detailed outcrop description and sampling was conducted in the Bottaccione Gorge and Vispi Quarry located in the Contessa valley near Gubbio, the Poggio le Guaine and Fiume Bosso sections, the Monte Petrano sections, and the Piobbico Road and Gorgo a Cerbara sections near Piobbico (Fig. 1). We have compiled data from the Piobbico core (Erba, 1988, 1992; Tornaghi et al., 1989) and from the Cismon Apticore located in the Belluno Basin (Erba et al., 1999) to document the occurrence of ORBs in the Marne a Fucoidi, or coeval formations.

Leckie et al. (2002) established a high-resolution geochronology encompassing the late Barremian to late Turonian (123–90 Ma), using the time scale of Gradstein et al. (1995) and the biostratigraphic age model of Bralower et al. (1997). In this paper we use geochronology of Leckie et al. (2002) to establish positions of individual ORBs and their durations in the Umbria–Marche Basin. We also collected samples from the boundary between the Scaglia Bianca and Scaglia Rossa in the Vispi Quarry section for stable carbon and oxygen isotope analyses.

Samples were prepared and stable isotope analyses were performed in the Marine Geology Laboratory, Tongji University. Powdered samples were placed in sample vials in a Finnigan automatic carbonate device (Kiel III), reacted with *ortho*-phosphoric acid at 70 °C to generate CO_2, then transferred to and measured in a Finnigan MAT252 mass spectrometer. Precision was regularly checked with a Chinese national carbonate standard (GBW04405) and international standard NBS19; the standard deviation was 0.07‰ for $\delta^{18}O$ and 0.04‰ for $\delta^{13}C$. Conversion to the international Pee Dee Belemnite (PDB) scale was performed using NBS19 and NBS18 standards.

3. Geological setting and lithology

The Cretaceous pelagic sequence of the complex Umbria–Marche Basin was deposited near the continental margin of the Apulia block. The block moved northward from Africa relative to northern Europe and was strongly influenced by pre-orogenic deformation, such as extensional normal faulting (Marchegiani et al., 1999). The basement of the Umbria–Marche Apennines is continental, with the Upper Jurassic through Lower Miocene pelagic strata overlying a Triassic to Lower Jurassic carbonate platform. During the latest phase of the Alpine–Himalayan orogeny in Miocene time (see Centamore et al., 2002), the basin was involved in tectonic compression and became part of the foreland

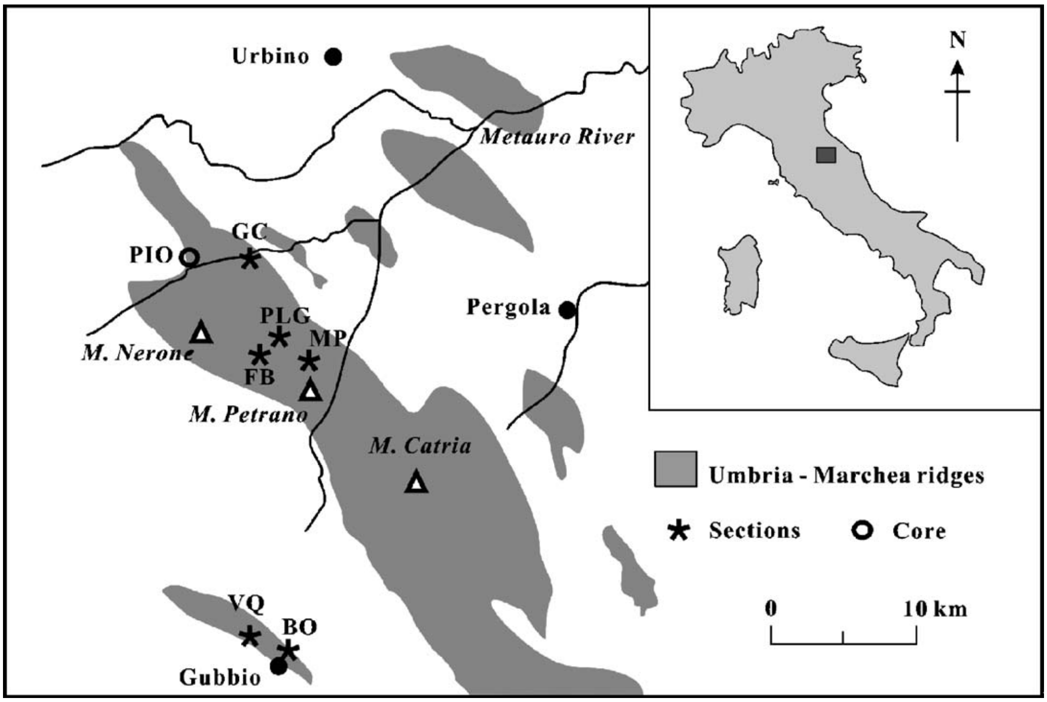


Fig. 1. Location of the studied sections near Gubbio–Piobbico area, the Umbria–Marche Basin, central Italy (modified from Baudin et al., 1998). Sections: BO: Bottaccione Gorge section; CQ: Vispi Quarry section; FB: Fiume Bosso section, GC: Gorgo a Cerbara section; MP: Monte Petrano section; PIO: Piobbico core; PLG: Poggio le Guaine section.

fold-and-thrust belt of the Umbria–Marche Apennines of central Italy. The motion of thrust sheets was from the southwest to the northeast.

The Cretaceous sedimentary succession cropping out in the Umbria–Marche Basin has been subdivided into several discrete formations and members based on color changes, carbonate content and the presence/absence of chert and black shales (see Arthur and Fischer, 1977; Alvarez and Montanari, 1988; Coccioni, 1996). The principal formations in upward stratigraphic order are Maiolica (upper Tithonian–lower Aptian), Marne a Fucoidi (lower Aptian–upper Albian), Scaglia Bianca (upper Albian–lowest Turonian) and Scaglia Rossa (lowest Turonian to middle Eocene). The Maiolica Formation consists primarily of whitish to medium gray pelagic limestones enclosing near the top of the formation, beige to black chert nodules, or layers and dark gray to black, organic-rich horizons with variable carbonate content. The overlying Marne a Fucoidi Formation is a more shaly sequence of dark gray to black calcareous shales, light green-gray marly limestones with intervals of interbedded red and green marlstones and calcareous mudstones. Black shale horizons in the Marne a Fucoidi include the lower Aptian Selli Level (OAE1a), uppermost Aptian “113” Level, lower Albian “Monte Nerone” Level, “Urbino” Level, and the uppermost Albian “Pialli” Level (see Coccioni et al., 1987, 1989, 1990; Coccioni and Galeotti, 1993; Coccioni, 2001; Galeotti et al., 2003). However, only the Selli level is accompanied by a distinct carbon isotope anomaly.

The Marne a Fucoidi Formation grades into the Scaglia Bianca Formation, which is comprised of yellowish-gray to grayish limestones intercalated with a few pink to reddish limestone beds and several greenish-gray to black marlstones and shales. The latter formation was subdivided (Coccioni et al., 1992) into four discrete informal lithologic members. These are (from bottom to top): W1 (lower yellowish-gray member), comprised of yellowish-gray limestone with common nodules and lenses of greenish-grey chert; pink to reddish limestone and greenish-gray to black, marly/shaly layers are intercalated in the lower part of the member; W2 (reddish member), consisting of pink to reddish limestone with subordinate yellowish-gray limestone and rare greenish-gray marly layers; pinkish-gray chert nodules and lenses are distributed throughout; W3 (upper yellowish-gray member), represented by yellowish-gray limestone with common nodules and lenses of greenish-gray chert; and W4 (grayish member), comprised of mostly light-gray to gray limestone with common nodules and lenses of dark-gray to black chert throughout; black marly/shaly layers are interbedded near the top. The Bonarelli horizon (OAE2) is a prominent marker bed near the top of the Scaglia Bianca Formation. It is an ichthyolithic–bituminous–radiolarian unit about 1 m thick, consisting of siliceous, carbonate-free olive-green to black mudstone and black shales with more than 23% organic carbon (Arthur and Premoli Silva, 1982).

The Scaglia Bianca grades upward into the Upper Cretaceous Scaglia Rossa Formation which was described in detail by Arthur and Fischer (1977) and Alvarez and Montanari (1988). It is comprised predominantly of pinkish-reddish marly limestones that generally contain 65–92% $CaCO_3$, except for rare shaly interbeds, which accentuate the bedding. It was subdivided by latter authors into R1 and R2 Members according to the presence/absence of chert. Member R1 (uppermost Turonian–lower Campanian) is characterized by the presence of greenish-gray and less commonly reddish-brown chert nodules and beds, interbedded with red to pink with less common yellow-gray to light pink colored limestone. The Member R2 (lower Campanian–Maastrichtian) generally lacks cherts and is predominantly composed of pink to reddish-brown limestones.

4. Mid-Cretaceous ORBs in the Umbria–Marche Basin

4.1. Stratigraphy and composition

The Cretaceous biostratigraphy in the Gubbio–Piobbico area (Fig. 1) is well known since Renz’s (1936) pioneering study. Detailed studies of the type sections at the Bottaccione Gorge and the Vispi Quarry near Gubbio were published by Luterbacher and Premoli Silva (1962), Premoli Silva (1977), Premoli Silva and Sliter (1994), Coccioni et al. (1987, 1989, 1990, 1995), Tremolada (2002) and Coccioni and Luciani (2004). The Poggio le Guaine and Fiume Bosso sections were described by Coccioni et al. (1990) and by Coccioni and Galeotti (1993). The Monte Petrano section was described by Fiet (1998) and Fiet and Masure (2001) (Fig. 1). The Piobbico core was drilled through the Aptian–Albian Marne a Fucoidi Formation in the early 1980s and was the subject of several detailed multidisciplinary studies including sedimentology, organic and inorganic geochemistry, calcareous nannofossil and foraminiferal biostratigraphy, cyclostratigraphy and paleomagnetism (see Erba, 1988, 1992; Tornaghi et al., 1989; Herbert et al., 1995; and references cited). All those

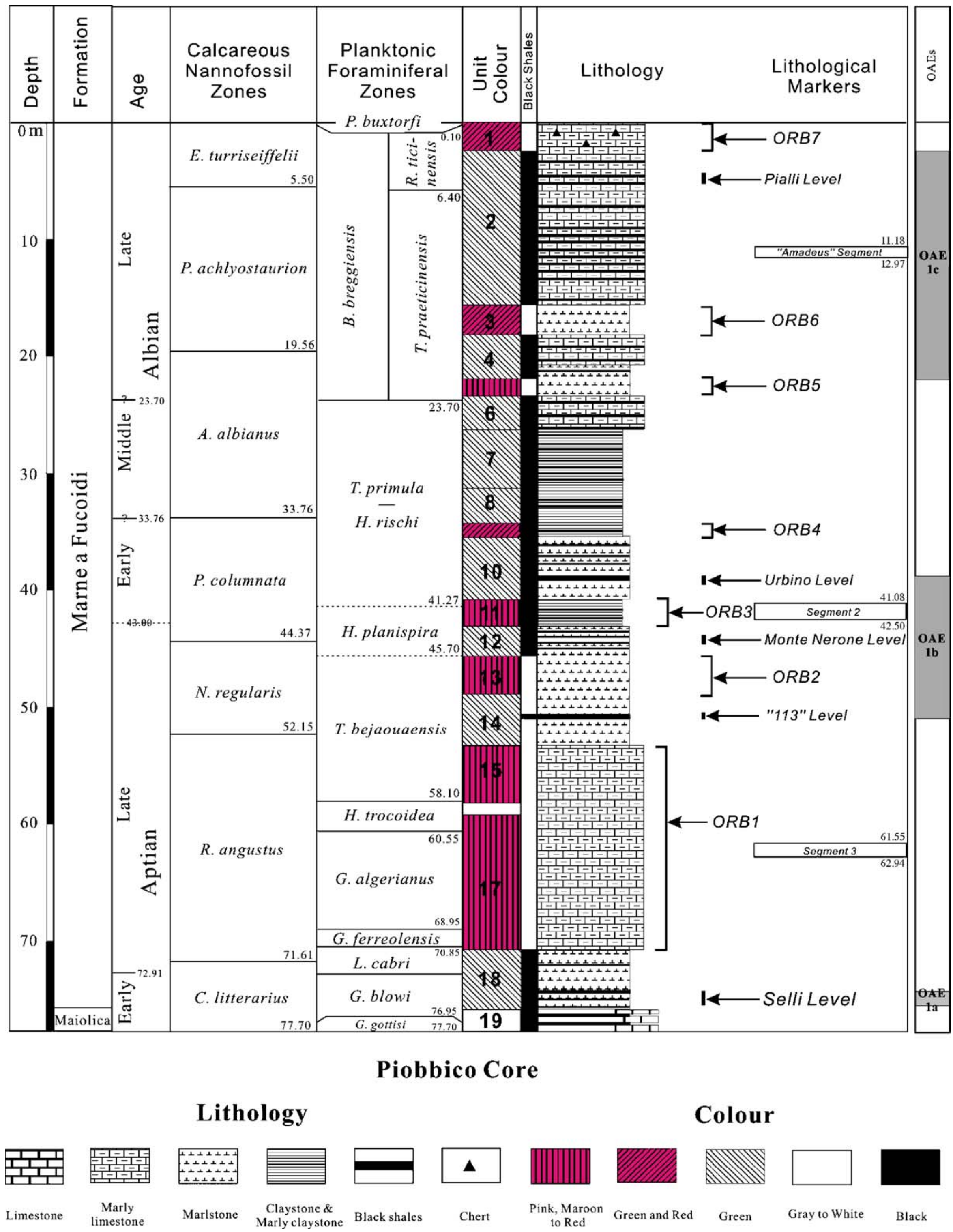
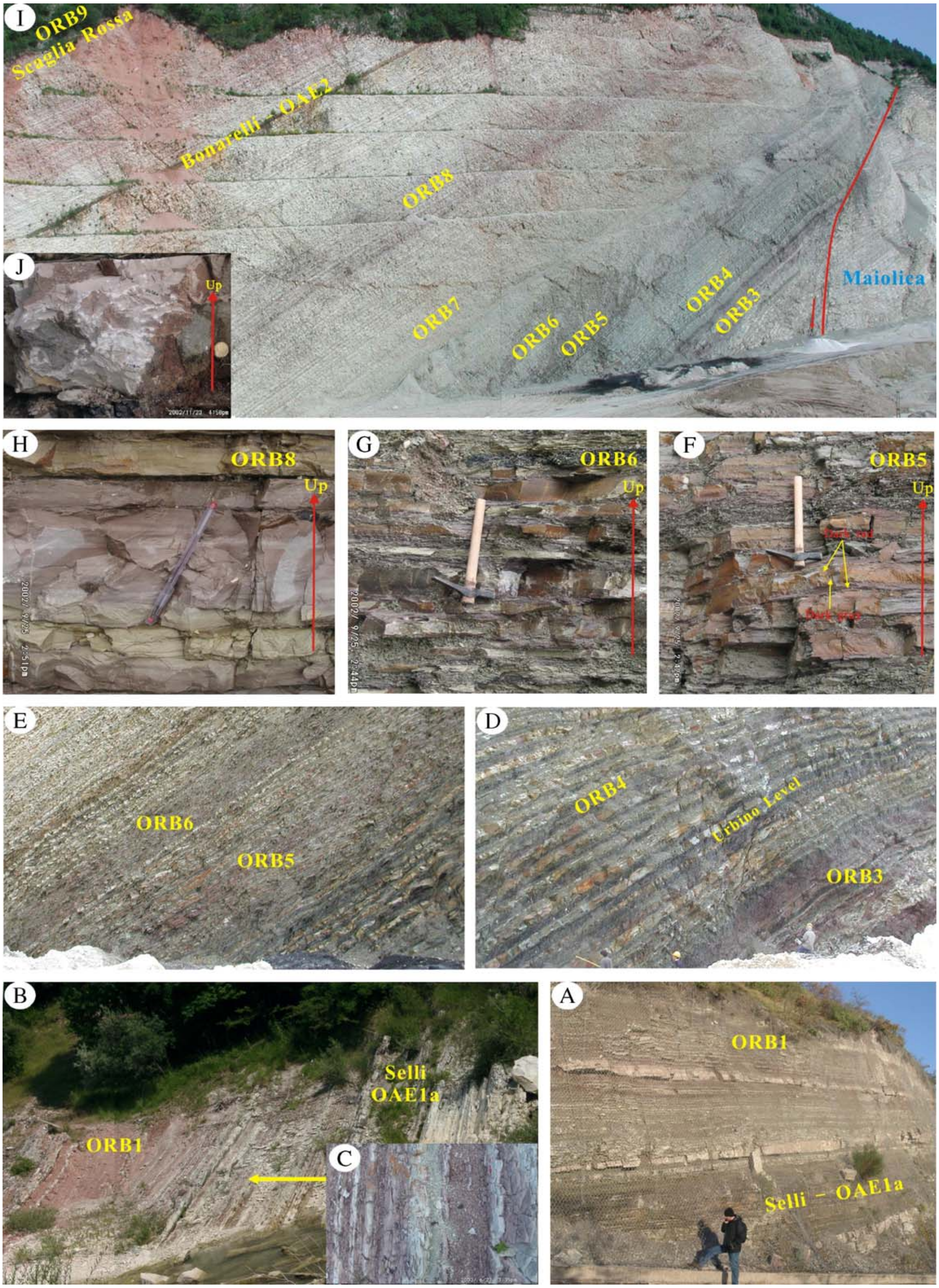


Fig. 2. Synthesis of stratigraphic data for the Piobbico core. Lithostratigraphy and integrated calcareous nannofossil-planktonic foraminiferal biostratigraphy are after Erba (1988), Tornaghi et al. (1989) and Erba (1992). Distinctive regional black shale levels, OAEs, and ORBs are marked in the right column. In the middle column, a more detailed color characteristic of the strata is presented. Also shown are locations of additional, perhaps more local OAEs, e.g., "113" Level, "Monte Nerone" Level, "Urbino" Level and "Pialli" Level (Coccioni et al., 1987, 1989, 1990; Coccioni and Galeotti, 1993; Coccioni, 2001; Galeotti et al., 2003).

studies provide well-established biostratigraphic framework for ORBs in the Aptian–early Turonian of the Umbria–Marche Basin.

In the course of field studies in the Umbria–Marche Basin, we concentrated on the occurrence of reddish colored sedimentary beds in a stratigraphic interval



bounded by the Selli level at the base (OAE1a) and by the Bonarelli level at the top (OAE2). In this sedimentary succession, we identified eight horizons characterized by pinkish and/or reddish colored beds. These "red bed" levels are numbered in an upward stratigraphic order—ORB1 to ORB8 (see Figs. 2, 4, and 5). ORB1 to ORB6 are also present in the Piobbico core (Erba, 1988; Tornaghi et al., 1989). The individual ORB horizons are described in an upward stratigraphic order below.

ORB1 is over 15 m thick in the Gorgo a Cerbara section. It is dominated by dark red marlstone, red marly limestones and red calcareous shales with subordinate gray marlstones and marly limestones. The gray beds locally are interbedded with the red beds in a cyclic manner (Fig. 3B). ORB1 overlies a 2.6-m-thick transitional zone comprised of thinly bedded, gray marly limestones (Fig. 3A, B), enclosing few gray chert lenses that in turn overlie the Selli bed. The red beds are generally less calcareous than the gray beds. The contacts between red beds and gray beds are commonly sharp (Fig. 3C). Within the intercalated gray beds, the color becomes more reddish in a few millimeter thick both near the base and the top. In the Piobbico core, ORB1 corresponds to Units 15 and 17 (53.14–70.65 m) of Erba (1988) (Fig. 2). These two red-colored horizons are separated by Unit 16, which is a 0.91-m-thick green-gray marly limestones, interbedded with dark-brown marly limestones and calcareous marls. In Units 15 and 17, the lithotypes are dark-red and dark-brown marls and marly limestones with few green-gray-colored calcareous marls and marly limestones. Carbonate content varies from 8.9% to 71.5% (Tornaghi et al., 1989). Burrows filled with black marls are widespread throughout, and *Chondrites* is common. Black shales are absent throughout the interval.

ORB2 corresponds to Unit 13 in the Piobbico core (Fig. 2). It is 3.19 m thick and consists of intercalated dark-red to gray-red marls and calcareous clays, locally with some gray-green patches. The carbonate content varies from 8.9% to 62% (Tornaghi et al., 1989). Clayey horizons are very rare, and black shale beds are absent. Radiolarian bands are rare, as are *Chondrites* burrows.

ORB3 corresponds to Unit 11 in the Piobbico core (Fig. 2). It is 2.55 m thick and made up of dark-red, locally varicolored gray-green clays and marly clays, rarely marls. The carbonate content varies from 2.8% to 50% (Tornaghi et al., 1989). ORB3 differs from ORB1 and ORB2 by abundant black-shale layers, deficient in carbonate content. In the Vispi Quarry section, a normal fault between Maiolica and Marne a Focoidi cut out ORB1, ORB2 and most of ORB3 (Fig. 3I, D). The upper part of the ORB3 is composed of red clays and marls with gray, green-white marlstones and clays. Bioturbation is abundant.

ORB4 is 1.56 m thick in the Vispi Quarry section (Fig. 3D), and 0.6 m thick in the Monte Petrano section (Fiet and Masure, 2001) (Fig. 4). In the Vispi Quarry, it consists of gray-green, locally variegated gray-brown, clayey marls and marls with alternating brown, rarely green varicolored clays. The red clays are intercalated with gray-green marls and clays. The contacts between red beds and gray beds are sharp. Bioturbation is extremely abundant in one gray-green marlstone that is overlain by red clays. Three 3–5-cm-thick, carbonate-poor, black shale layers correspond to Unit 9, 1.11 m thick, in the Piobbico core (Fig. 2). Carbonate content varies from 8.9% to 51.2% (Tornaghi et al., 1989).

ORB5 is 1.26 m in the Vispi Quarry section (Fig. 3E) and 1.4 m thick (Fig. 2) in the Monte Petrano section (Fiet and Masure, 2001) (Fig. 4). It consists of slightly bioturbated brown-red marls and clays, intercalated with subordinate white-gray marly limestone beds. Millimeter-thick dark red horizons are near the top and bottom of the white-gray marly limestones (Fig. 3F). In some reddish beds, the middle part (less than 1/3 of the bed's thickness) is dark gray color and more calcareous. It corresponds to Unit 5, which is 1.39 m thick in the Piobbico core. The carbonate content varies from 52% to 71.5% (Tornaghi et al., 1989). Black-shale layers are absent and radiolarian horizons are rare.

ORB6 is 2.05 m in the Vispi Quarry section (Fig. 3E) and 3.1 m thick in the Monte Petrano section (Fiet

Fig. 3. Photographs showing the mid-Cretaceous ORBs in the Umbria–Marche Basin. (A–C) the early Aptian sequence, showing the change from early Aptian OAE1a–Selli Level to the first red bed–ORB1, in the Piobbico Road section (A) and in the Gorgo a Cerbara section (B). In the ORB1, the gray beds are interbedded with the red beds in a cyclic manner (B and C). The contacts between red beds and gray beds are usually sharp (C); (D) early Albian sequence showing the positions of the ORB3, ORB4, and the Urbino level in Vispi Quarry section; (E) late Albian sequence showing the positions of the ORB5 and ORB6 in the Vispi Quarry section; (F) close-up photo of the ORB5 in the Vispi Quarry section, showing millimeter-thick dark red horizons occurring near the top and bottom of the white-gray marly limestones; (G) close-up photo of the ORB6 in the Vispi Quarry section; (H) close-up photo of the ORB8 in the Vispi Quarry section showing that in some beds, a white color occurs at the bottom and gradually becomes into pinkish or dark reddish towards the top; (I) panoramic photo of the Vispi Quarry section showing the positions of the ORBs (from ORB3 to ORB9) and the Bonarelli level; (J) close-up photo of the transitional bed from Scaglia Bianca to Scaglia Rossa, which is whitish in color at the bottom and gradually becomes pinkish towards the top.

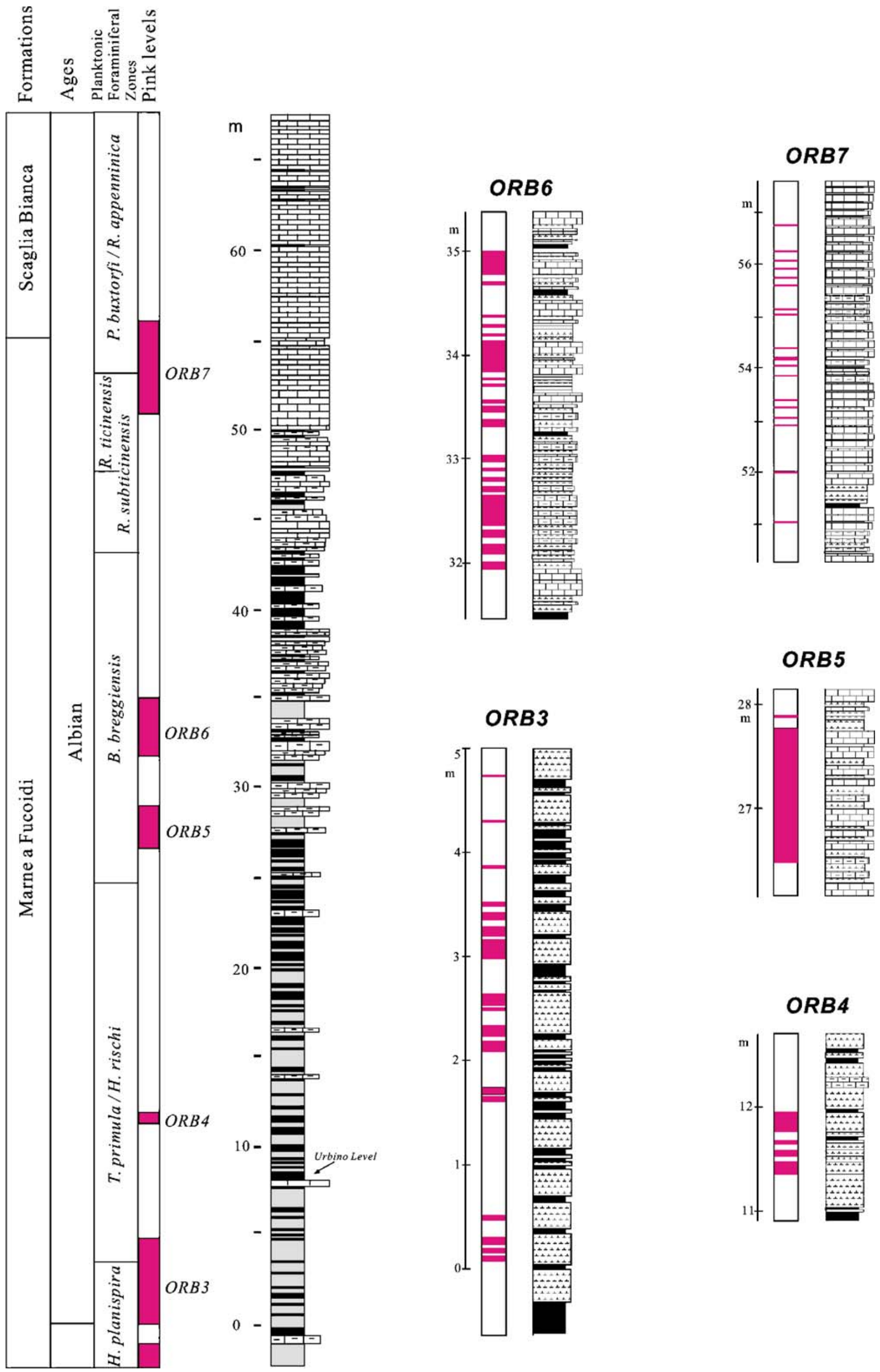


Fig. 4. The detailed lithostratigraphy and biostratigraphy of the Albian part of the Monte Petrano section. On the left side, the detailed lithology of the Albian strata and positions of ORBs are indicated; on the right side, lithology and color variation of five ORBs in the Albian section of the Monte Petrano are shown in more detail (modified after Fiet and Masure, 2001). Legend as in Fig. 2.

and Masure, 2001) (Fig. 4). In The Vispi Quarry section, it is comprised of 2–15-cm-thick, brown-red marls and clays, alternating with 1–10-cm-thick, gray marly limestones. Two, 3–5-cm-thick, black-shale layers occur. The contacts between red and gray beds are sharp (Fig. 3G). Red beds are in general less calcareous than gray beds. In the Piobbico core, ORB6 corresponds to Unit 3, which is 2.35 m thick (Fig. 2). The carbonate content varies from 52% to 71.5% (Tornaghi et al., 1989).

ORB7 is 1.89 m in the Vispi Quarry section (Fig. 3I) and about 5.8 m thick in the Monte Petrano section (Fiet and Masure, 2001) (Fig. 4) and 4.2 m in the Bottaccione Gorge section (Premoli Silva and Sliter, 1994). Its lower part correlates with Unit 1 at the top of the Piobbico core (Erba, 1988). It is thinly bedded, with intercalations of whitish-gray, 3–10-cm-thick, marly limestones and gray and pink marls. Light gray chert lenses and biotrubation are abundant. The pink marls mostly occur as 1–3-cm laminations with sharp contacts to the gray beds.

ORB8 is about 14 m thick in the Bottaccione Gorge section (Arthur and Fischer, 1977; Premoli Silva and Sliter, 1994) and about 19 m in the Vispi Quarry section (Fig. 3I). ORB8 level was named the W2 reddish member by Coccioni et al. (1995). It occurs in the Scaglia Bianca Formation, about 30 m below the Bonarelli Level (Fig. 5). It is comprised of pink to reddish limestones with subordinate gray limestones and rare greenish-gray marly layers. Pinkish-gray chert nodules and lenses are abundant and distributed throughout. In some beds, parts colored white occur at the bottom and gradually become pinkish or dark reddish towards the top (Fig. 3H). In some gray beds, reddish laminations vary from millimeters thick to a few centimeters thick with a transitional color change.

Using the mid-Cretaceous integrated calcareous biostratigraphy and the age of planktonic foraminiferal zones of Leckie et al. (2002), we calculated time durations of individual ORBs in the Piobbico core, in the Monte Petrano section, in the Bottaccione Gorge section, and in the Vispi Quarry section (Table 1; Fig. 9). The duration of these oxic events varied from 0.13 my for ORB4 to 4.54 my for ORB1 (Table 1).

4.2. Biotic signal

Micropaleontological studies document that the mid-Cretaceous was a time of rapid radiation and turnover in marine plankton, benthic foraminifera, molluscs, and terrestrial plants (see Premoli Silva and Sliter, 1999; Leckie et al., 2002). Turnover in the planktonic foraminifera closely tracks that of the radiolaria (Erbacher and Thurow, 1997), and both groups display the greatest rates of turnover at or near the major OAEs. The calcareous nannoplankton seems to be most strongly affected by the early Aptian OAE1a (Selli event) and the Cenomanian/Turonian boundary OAE2 (Bonarelli event) (Leckie et al., 2002; Erba, 2004). Kaiho (1999) concluded that flux of organic carbon to the sediments and dissolved-oxygen levels in the water column were important in the control of benthic foraminiferal test sizes, wall thickness, morphology, and species composition. In the Piobbico core, Tornaghi et al. (1989) noticed that planktonic foraminiferal abundance does not correlate with any special lithotypes in the Piobbico core. Radiolaria in ORB1 are more common in gray and green-gray marly limestones and very rare in the red, more oxidized marl. However, in ORB2, radiolarians appear to occur in larger abundances in some of the red calcareous marls and are less common in some gray lithotypes (Tornaghi et al., 1989). The relationship between the ORBs and marine biotic response to a highly oxygenated deep-sea bottom environment has not been studied in any detail and a more specific study of the microfossils in various lithotypes is needed.

4.3. Stable isotopes

Carbon and oxygen isotopes of Cretaceous rocks in the Umbria–Marche Basin (Jenkyns et al., 1994; Hochuli et al., 1999; Stoll and Schrag, 2000; Galeotti et al., 2003; Weissert and Erba, 2004) and in less detail for the Aptian–early Turonian composite sequence in the Piobbico area (Erbacher and Thurow, 1997) documented close but not complete correspondence between positive shifts on the $\delta^{13}C$ curve and positions of major OAEs, such as those of the OAE1a and OAE2 (e.g., Schlanger et al., 1987; Arthur et al., 1988; Menegatti et al., 1998; Luciani et al., 2001). The first ORB1 in the Piobbico corresponds to the lowest value on the $\delta^{13}C$ curve after a positive excursion identifying the Selli level (Menegatti et al., 1998; Weissert et al., 1998). A negative shift is indicated for the middle Albian at the level of ORB4 in the Cismon core (Erba et al., 1999). Similar trends can be seen in the isotopic data of Jenkyns et al. (1994) and Stoll and Schrag (2000) for ORB8, which correlates with a broad 0.5‰ negative excursion, representing the lowest values in the Cenomanian (Fig. 5). $\delta^{13}C$ progressively increases up to the Turonian with a short positive 0.7‰ excursion at the mid-Cenomanian event (MCE) (Coccioni and Galeotti, 2003). In the Vispi section, no obvious carbon isotopic

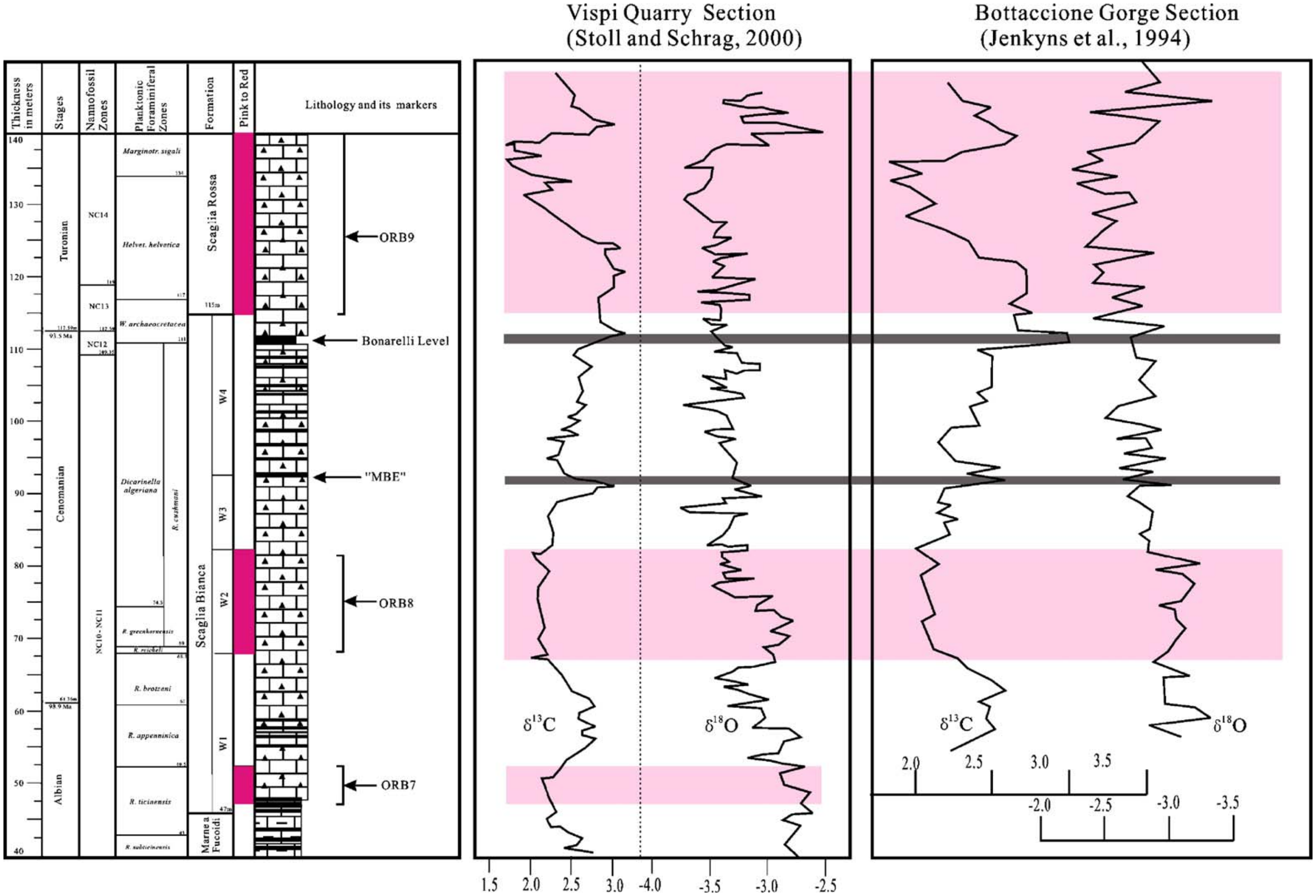


Fig. 5. Lithostratigraphy and biostratigraphy of the Albian–Turonian segment of the Bottaccione Gorge section (revised after Premoli Silva and Sliter, 1994; Tremolada, 2002; Coccioni and Galeotti, 2003) are shown plotted against the late Albian–late Turonian carbon and oxygen isotope record from the Bottaccione Gorge (Jenkyns et al., 1994) and Vispi Quarry (Stoll and Schrag, 2000). Note that the MCE (mid-Cenomanian event) and the Bonarelli event are associated with positive excursions of $\delta^{13}C$, while ORB7 and ORB8 correspond to low values on the carbon isotopes curve. Legend as in Fig. 2.

Table 1
Occurrences of mid-Cretaceous oceanic red beds (ORBs) in the Piobbico core, the Monte Petrano sections, the Bottaccione Gorge and the Vispi Quarry; their stratigraphic position in planktonic foraminiferal zonation, thickness and duration of ORBs

ORBs	Planktonic foraminiferal zones	Piobbico Core		Monte Petrano		Bottaccione Gorge		Vispi Quarry
		Thickness (m)	Estimated duration (Ma)	Thickness (m)	Estimated duration (Ma)	Thickness (m)	Estimated duration (Ma)	Thickness (m)
ORB8	From *R. reicheli* zone to lower part of the *R. cushmani* zone					~14	0.95	~19
ORB7	From upper part of *R. ticinensis* zone to lower part of the *R. appenninica* zone			~5.8	0.87	~4.2	0.63	1.89
ORB6	Within the *R. praeticinensis* zone	2.35	0.35	3.10	0.44			2.05
ORB5	Within the *R. praeticinensis* zone	1.39	0.21	1.40	0.20			1.26
ORB4	Within the *T. primula* zone	1.11	0.28	0.60	0.13			1.56
ORB3	From upper part of the *H. planispira* zone to lower part of the *T. primula* zone	2.36	1.30					
ORB2	From upper part of the *T. bejaouaensis* zone to lower part of the *H. planispira* zone	3.19	0.55					
ORB1	From *G. ferreolensis* zone to *T. bejaouaensis* zone	17.51	4.54					

Planktonic foraminiferal zonation is after Leckie et al. (2002). Data sources: for Piobbico core from Erba (1988) and Tornaghi et al. (1989); for Monte Petrano from Fiet and Masure (2001); and for Bottaccione Gorge from Premoli Silva and Sliter (1994) and Coccioni et al. (1995).

change corresponds to the lithological change from Scaglia Bianca to Scaglia Rossa (Fig. 6).

The oxygen isotopic record from the Vispi section is characterized by a general trend toward more negative $\delta^{18}O$ with up-section from upper Albian to middle Turonian. Several positive excursions of 0.75–1.5‰ are superimposed on pervasive higher frequency, low-amplitude variations (Stoll and Schrag, 2000). A larger 1.5‰ positive excursion is seen in the lower Cenomanian record at the top of the *Rotalipora brotzeni* zone and at the bottom of the *Rotalipora cushmani* zone and extends into the lower part of the ORB8 (Fig. 5), which may indicate a brief cooling period. The $\delta^{18}O$ curve, however, does not show any other clear correlation between ORBs and oxygen isotope shifts (Figs. 5 and 6).

4.4. Transition from Bonarelli level to Scaglia Rossa Formation

About 4.3 m above the Bonarelli Level, the Scaglia Bianca Formation grades into a succession of predominantly pinkish-reddish limestones and marls of the Scaglia Rossa Formation. We studied in detail the transition from light gray limestones of the Scaglia Bianca Formation overlying the Bonarelli level into pink-reddish limestones of the Scaglia Rossa in the Vispi Quarry (Fig. 7).

The first bed of variegated white and pink limestone occurs at 4.34–4.44 m (Fig. 7) above the Bonarelli level; the next millimeter-scale bedded pinkish limestone occurs at 5.72 m. Two transitional beds at 10.35–10.55 m and at 10.63–10.83 m, respectively, are below the completely reddish colored beds (Fig. 6; Fig. 7). Each of these beds is about 20 cm thick, whitish in color at the bottom and gradually becoming pinkish towards the top (Fig. 3J). Above these two transitional beds, the limestone becomes pinkish to reddish at 10.83 m (Fig. 6; Fig. 7). If we apply the 7.4 m/Ma sedimentation rate of the *Helvetoglobotruncana helvetica* zone of Premoli Silva and Sliter (1994), it took about 0.88 Ma for the change from deposition of the white to light gray limestones of the Scaglia Bianca Formation into the predominantly red-colored limestones of the Scaglia Rossa Formation.

It is interesting to note that the first variegated grayish and pinkish colored nodular chert (~10–20 cm thick) is enclosed in whitish limestones, about 3 m above the Bonarelli Level (Fig. 7). The next variegated pinkish and gray, thin laminae (about 2 mm thick) of chert occur at 3.5 m and 3.74 m (Fig. 7) in whitish limestones. The reddish colored chert appears at 3.03–3.74 m above the Bonarelli Level, below the appearance of the first pinkish limestone at 4.34 m. Because the nodular chert formed during late diagenesis, the color variation signifies that redox conditions changed during sediment lithification probably as result of percolating formation water from the overlying "oxic zone." Why iron oxidation did not occur in the overlying pelagic carbonate is yet unanswered. Could be this

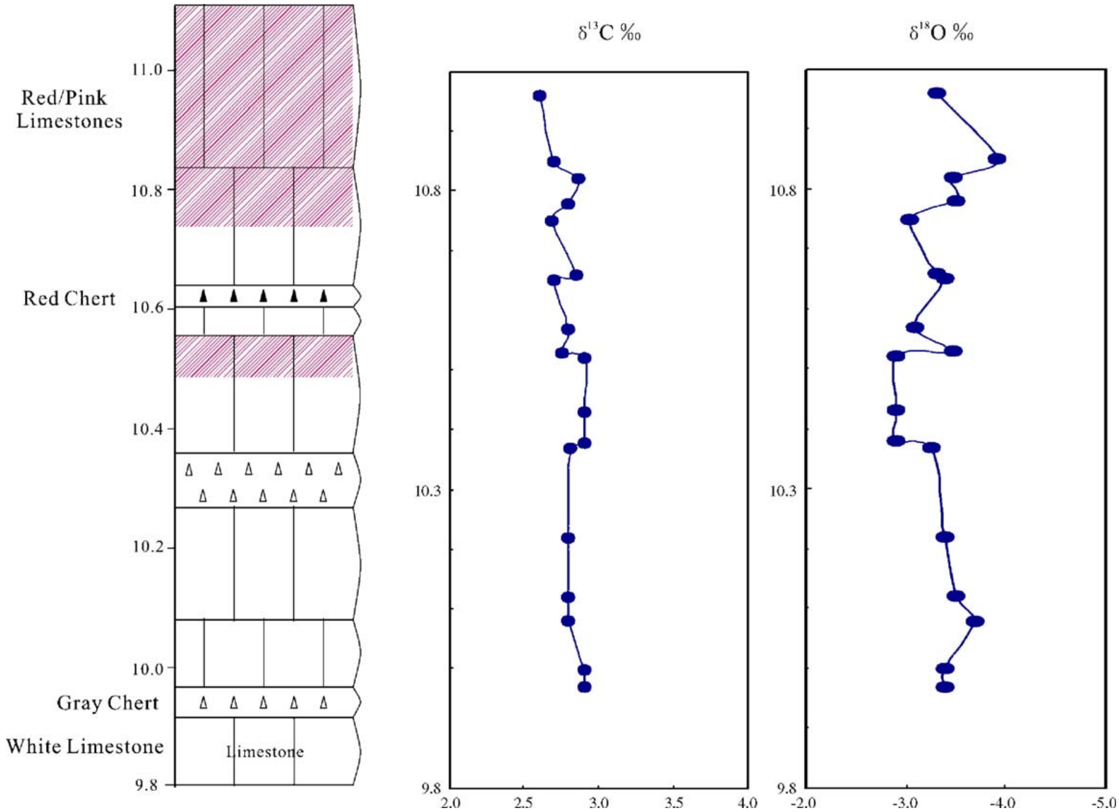


Fig. 6. Carbon and oxygen isotope records from the transition from Scaglia Bianca to Scaglia Rossa in the Vispi Quarry section. Note that there is no obvious isotopic change for the occurrences of the red beds.

due to a higher content of organic matter in the carbonates, which would indicate that the organic matter oxidation was a very slow process?

5. Mid-Cretaceous ORB's correlation in Tethys and Atlantic

The broad, regional search for mid-Cretaceous ORBs reveal that they are not uncommon constituents of Cretaceous sedimentary strata deposited along the northern margin of the Tethys. Their occurrences have been documented not only in the Umbria–Marche Basin, central Italy, but also in the Southern Alps, the Austrian Alps, the Czech Republic, in Slovakia, the Polish Carpathians, the northern Caucasus, the central North Atlantic, in northeastern England, as well as in the Zanskar Himalayas (see below) (Fig. 8).

In the Southern Alps, in the Belluno Basin, the Cismon core variegated pinkish grey, light reddish grey and dark reddish brown limestones alternate with reddish brown marlstones in Unit 2, 4.87 m thick (Erba et al., 1999). Planktonic foraminifera date this unit as middle Albian and extends from the *T. primula* zone to the lower part of the *Biticinella breggiensis* zone (Erba et al., 1999). Reddish brown chert nodules are also intercalated in this unit (Erba and Larson, 1998; Erba et al., 1999) and form a regional marker bed. The biostratigraphic evidence shows that this unit could be synchronous with the ORB5 in the Umbria–Marche Basin (Fig. 8).

In the Northern Calcareous Alps, Austria, three mid-Cretaceous ORBs were found. The lowest one occurs only in some of the thrust sheets (e.g., Frankenfels nappe of Weyer area, Austria) (Wagreich, 2002), where it is represented by the *Hedbergella* limestone. It is a 50-cm to several-meter-thick, condensed, reddish colored limestone with plentiful planktonic foraminifera indicating a late Aptian age (Wagreich, 2002). This red limestone overlies the shallow-water carbonates of the Lower Cretaceous Schrambach Formation and underlies shales of the Tannheim marlstones. In the dark grey shales and shaly marls of the Tannheim Formation, a zone of red shales occurs. The age of this younger reddish colored zone is latest Aptian (Wagreich, 2005). These two red beds could be synchronous with ORB1 and ORB2 in the Umbria–Marche

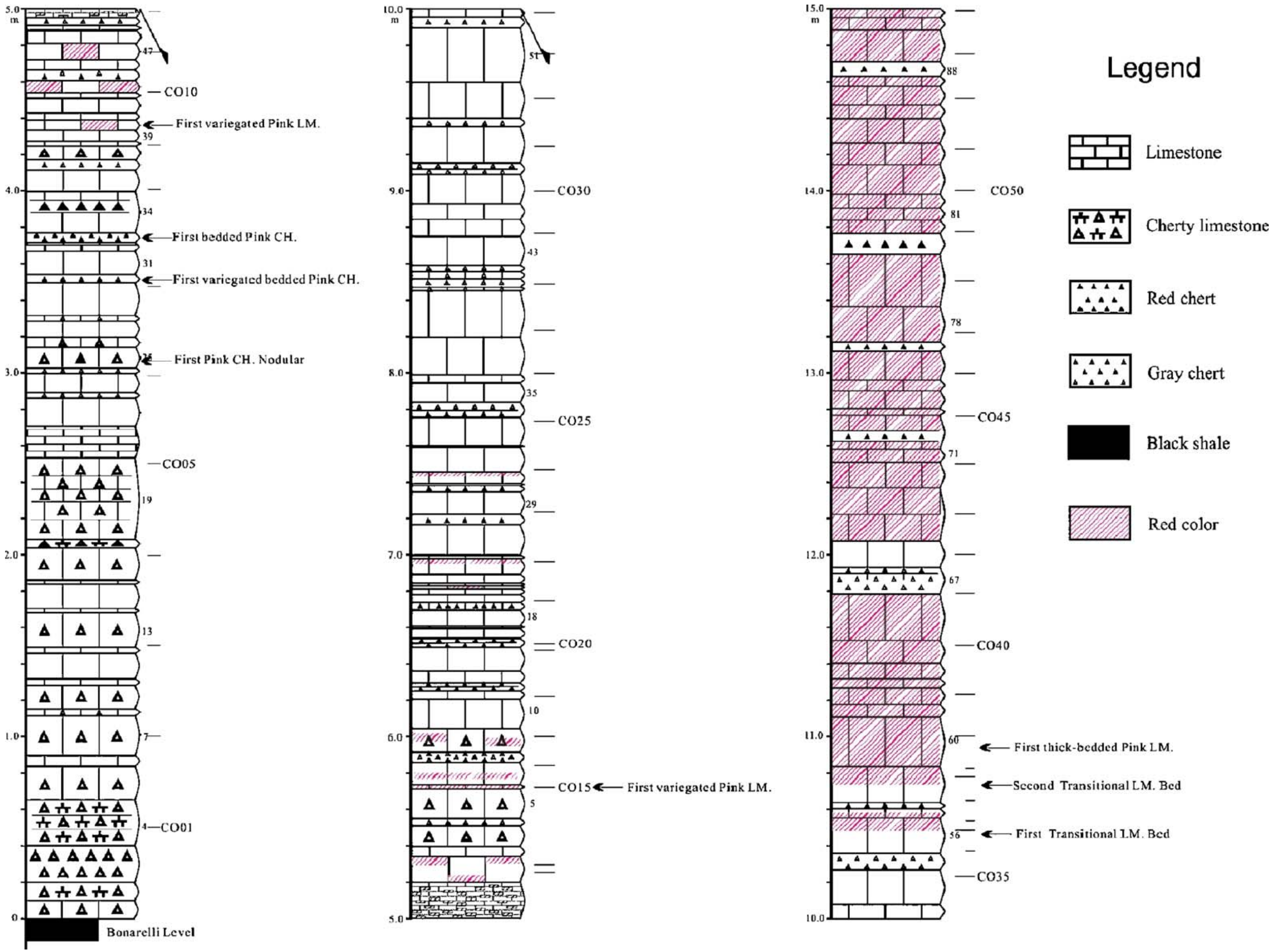


Fig. 7. Detailed lithogical log of the Vispi Quarry section from the top of the Bonarelli level to the lower part of the Scaglia Rossa Formation to demonstrate redox changes in this 15-m-thick interval. Note the occurrence of a reddish color both in the limestones and in the cherts. Symbols as in Fig. 2, except for color. Red cross-hatching denotes reddish coloration of beds.

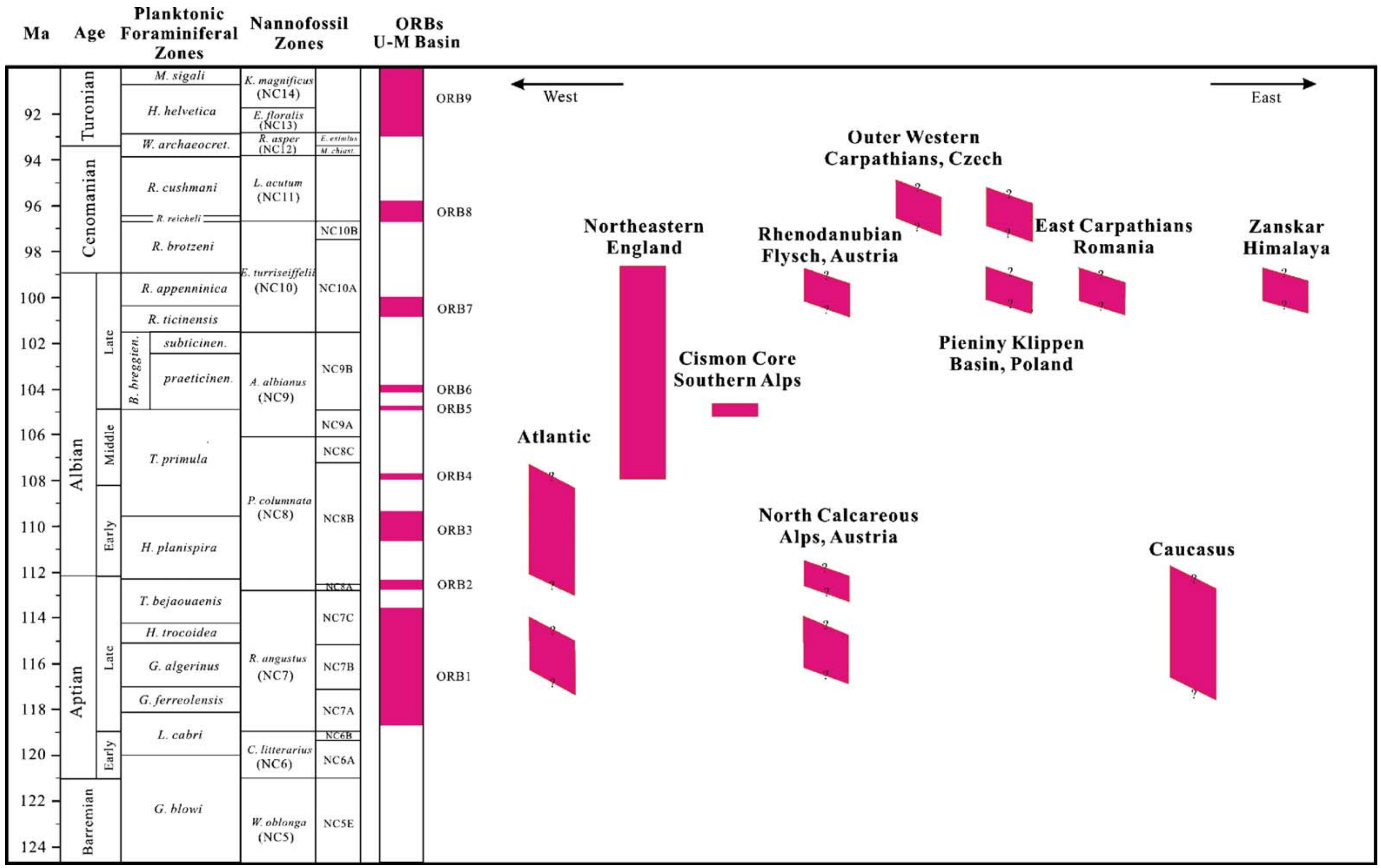


Fig. 8. Chronostratigraphic distributions of the mid-Cretaceous oceanic red beds in Tethys and Atlantic. The mid-Cretaceous integrated calcareous plankton biostratigraphy is after Leckie et al. (2002).

Basin (Fig. 8). Another level with reddish marly shale beds (up to 30 cm) is intercalated in a sandstone turbidite sequence in the "Untere Bunte Schiefer" of the Rhenodanubian Flysch in Austria. The latter red beds have been dated of latest Albian to early Cenomanian age, assigned to the dinoflagellate *Litosphaeridium siphoniphorum* zone, foraminiferal topmost *Rotalipora apenninica* zone and the *Rotalipora globotruncanoides* zone, and nannofossil standard zone CC9/UC0 (Wagreich, 2005) and, therefore, may correlate with ORB7 in the Umbria–Marche Basin (Fig. 8).

ORBs are also present in the Upper Cretaceous Silesian Unit of the Outer Western Carpathians, Czech Republic. In the 95-m-thick Mazak Formation, zones of ORBs are intercalated with grayish and greenish shales and sandstone turbidite beds. Dinoflagellate cysts indicate middle to late Cenomanian age for the Mazak Formation (Skupien and Vasicek, 2003). In the Polish part of the Pieniny Klippen Basin, lowest mid-Cretaceous variegated facies were found in the Chmielowa Formation, which has a variable thickness from 0.5 m to 10 m (Bak, 2002) and is comprised of red and variegated marly limestones and marls. These red sediments are overlain by an upper Albian anoxic black shale facies in the *Rotalipora ticinensis* zone (Bak, 2002). Pelagic marls and limestones in the Chmielova Formation contain abundant planktonic foraminifers of the genus *Hedbergella*, rare benthic foraminifers, calcified radiolaria and tiny fragments of bivalves. The red beds were placed in the *Ticinella roberti* zone by Bak (2002) and, thus, may correspond to the ORB7 level in the Umbria–Marche Basin (Fig. 8). In the Pieniny Klippen Basin in Poland, another unit of ORBs is the Skalski Marl Member (Bak, 1998). Alternating variegated (green and pink) marls, marly shales and marly limestones 2–10 m thick enclose a few thin black shale layers (Bak, 1998). Foraminifera place this ORB in the *Rotalipora reicheli* to *R. cushmani* zones (middle–upper Cenomanian), which is approximately synchronous with ORB8 in the Umbria–Marche Basin (Fig. 8).

In the Romanian East Carpathian, the lower member of red shales of the Dumbravioara Formation crop out in the Outer Dacides and the Inner Moldavides units (Melinte et al., 2004). This red bed unit is tens of meters thick and spans the uppermost Albian to lowermost Cenomanian foraminiferal *Rotalipora appenninica* zone, which could be synchronous with ORB7 in the Umbria–Marche Basin (Fig. 8).

On the western slope of the Northern Caucasus, which is characterized by Cretaceous deep-water flysch facies, the lower part of the Medoveevskaya Formation contains Aptian red marls (Egoyan, 1986; Natalia Tur, personal communication, 2003). In sections in the Matzestin area (Abkhazia), the 2-m-thick red bed in the lower part of the Medoveevskaya Formation overlies Barremian limestones with an erosional contact and belongs to the *Acanthoplites nolani* zone—the lower zone of the upper Aptian (Egoyan, 1986). The upper Aptian dark-pink and grayish red marls were also found in the south slope of the Great Caucasus (Georgian block and Azerbaijan block) (Egoyan, 1986). Despite imprecise biostratigraphic data, it appears that these ORBs could also be synchronous with ORB1/ORB2 in the Umbria–Marche Basin (Fig. 8).

ORBs are also found in the central North Atlantic basin. In the Cape Verde Basin in the eastern central North Atlantic, upper Aptian–lower Albian strata at Site 367 (Core 24; Jansa et al., 1978) and at ODP Sites 1049 and 603 in the western North Atlantic (Norris et al., 1998) change from poorly oxidized sediments to highly oxidized sediments. At Site 367, coring was not continuous, so the contact and composition of underlying sediments is unknown. However, from drilling penetration data, we can safely assume that the "red bed" is embedded in a black shale facies, similar to those overlying this bed. At Site 1049 on Blake Nose off Florida, a unit up to 38 m thick is dominated by decimetre-scale rhythmic alternation of greenish and reddish clayey, nannofossil chalk and nannofossil clay and was dated by nannofossils as late Aptian to middle Albian in age (Bellier et al., 2000). The variegated unit is disconformably overlain by Campanian nannofossil ooze. The base this unit grades into light-colored chalk of Aptian age (Norris et al., 1998). At Site 603 east of Cape Hatteras near the edge of Hatteras Abyssal plain, two zones of reddish colored claystones are intercalated in mid-Cretaceous greenish and black carbonaceous claystones (van Hinte et al., 1987). One is in Albian strata (Subunit 4B) and the other is Aptian age (subunit 4D). The latter is underlain by greenish gray and black claystones. Because no high precision biostratigraphic data are available from the sites drilled in the North Atlantic, we can only speculate about their comparative positions to the Umbria–Marche Basin ORBs scheme (Fig. 8). The red claystone unit in Site 367 was placed in the nannofossil *Parhabdolithus angustus* zone (Cepek, 1978) and is therefore synchronous with ORB1 in the Umbria–Marche Basin. The lower reddish claystone at Site 603 is probably also correlative with ORB1, and the younger (Subunit 4B) may be equivalent to one of the horizons between ORB2 and ORB4 (Fig. 8). The red beds unit in Site 1049 could have similar correlation. Therefore, the late Aptian–early Albian red beds in the southern central North Atlantic are part of an ocean-wide event, because their occurrence extends into the western "Mediterranean" Tethys. Importantly, these events could also be evidence that not all ORBs have the same origin. The late Aptian–early Albian red bed event at Site 367 has been interpreted to be a result of both tectonics and ocean circulation by Jansa (1999), when for a brief period during the opening of the equatorial seaway, high-salinity, oxygen-rich South Atlantic surface waters entered the southern North Atlantic, where they were incorporated into North Atlantic bottom waters.

Mid-Cretaceous red-colored chalk (so-called Red Chalk) strata crop out in northeastern England at Speeton (Mitchell, 1995), South Ferriby (Gaunt et al., 1992), and Hunstanton (Owen, 1995) areas. The Red Chalk Formation is 24 m thick at Speeton, about 2 m thick at South Ferriby and about 10 m thick at Buckton and is composed of a condensed series of brick-red marls and pink coccolithic nodular limestones (Mitchell, 1995). They contain abundant fossils, such as belemnites, brachiopods and foraminifera giving an early middle Albian–earliest Cenomanian age at Speeton (Mitchell, 1995). It appears that the Red Chalk Formation could be synchronous with the development of ORB4-ORB7 in the Umbria–Marche Basin (Fig. 8).

Upper Albian burrowed reddish to light grey and greenish-grey marly limestones in the Fatu La Formation crop out in the Nerak section, Zangla Unit, Zanskar Himalayas (Premoli Silva et al., 1991). As reported by Premoli Silva et al. (1991), this reddish bed sequence is over 100 m thick and rich in planktonic foraminifers, which confine the red beds to the foraminiferal late Albian *R. appenninica* zone. Therefore, these red beds would be synchronous with ORB7 in the Umbria–Marche Basin (Fig. 8).

6. Discussion—implications of ORBs for mid-Cretaceous paleoceanography and climate

The mid-Cretaceous ORBs exposed in several areas of southeastern and central Europe were originally deposited near the northern margin of the Tethys, where deposition occurred mostly above the carbonate compensation depth (CCD). It resulted in the preservation of calcareous microfauna, allowing detailed biostratigraphic zonation and a broader regional correlation of ORBs (Fig. 8). The duration of individual ORBs was highly variable (Scott et al., 2005). In the Umbria–Marche Basin, where the biostratigraphic data are most detailed, the shortest event lasted ~0.13 my (ORB4), and the longest was 4.54 my in duration (ORB1) (Table 1, Fig. 9). We have not found any systematic relation between the duration time and the stratigraphic position of these beds, which may be indicative of the strong influence by local conditions. However, the broad regional extent of some of these events favors a more general cause. The highest number (five) of ORBs events was noted in the Albian, with two in the Aptian and one in the Cenomanian (Fig. 9). How significant is it that the number of ORBs is similar to the number of OAEs is not clear yet, because there is no visible association between the OAE and ORB horizons, which do not directly overlie each other, but are separated by a variable thickness of light grey-colored beds. But, because most of them somehow alternate (see Fig. 2), it suggests an as-yet unrecognized, deeper, more general paleoceanographic cause.

The color change of the deep-sea pelagic sediments reflects changes in redox conditions in the benthic boundary layer, including the upper several to tens of centimeters of sediments beneath the sediment–water interface. In very simplified terms, the process of ocean bottom oxidation is the result of changes in the dissolved oxygen/organic matter ratio in oceanic waters. However, both of these factors are themselves complex, because the presence of organic matter is dependent on the bioproductivity, which in turn is governed by the presence of nutrients in the ocean. Therefore, the process is influenced by surface ocean dynamics, which is affected by climate changes and by the input of nutrients from the continents; therefore, the system also responds to sea level changes and, thus, indirectly affected by plate tectonics. The distribution of oxygen in the ocean is also complex and governed by many concurrent processes, such as its liberation by photosynthesis in the euphotic zone, transport as a result of water movement, exchange with the atmosphere, climate change and consumption by the oxidation processes taking place in the entire water column and in the sediments. As many of these processes can occur concurrently, it remains very difficult to establish with any certainty which if any of these processes represents the main triggering mechanism for the redox changes seen in deposited sediments, and secondly, it is very probable that similarly appearing ORBs may have different origins in different areas and/or at different times.

Below, we discuss in more detail several different processes that may have influenced the origin of the mid-Cretaceous ORBs.

1) According to Betts and Holland (1991), atmospheric O_2 cycles through the biosphere once in approximately 5000 years. A very small fraction (~0.2%) of the organic matter generated by photosynthesis

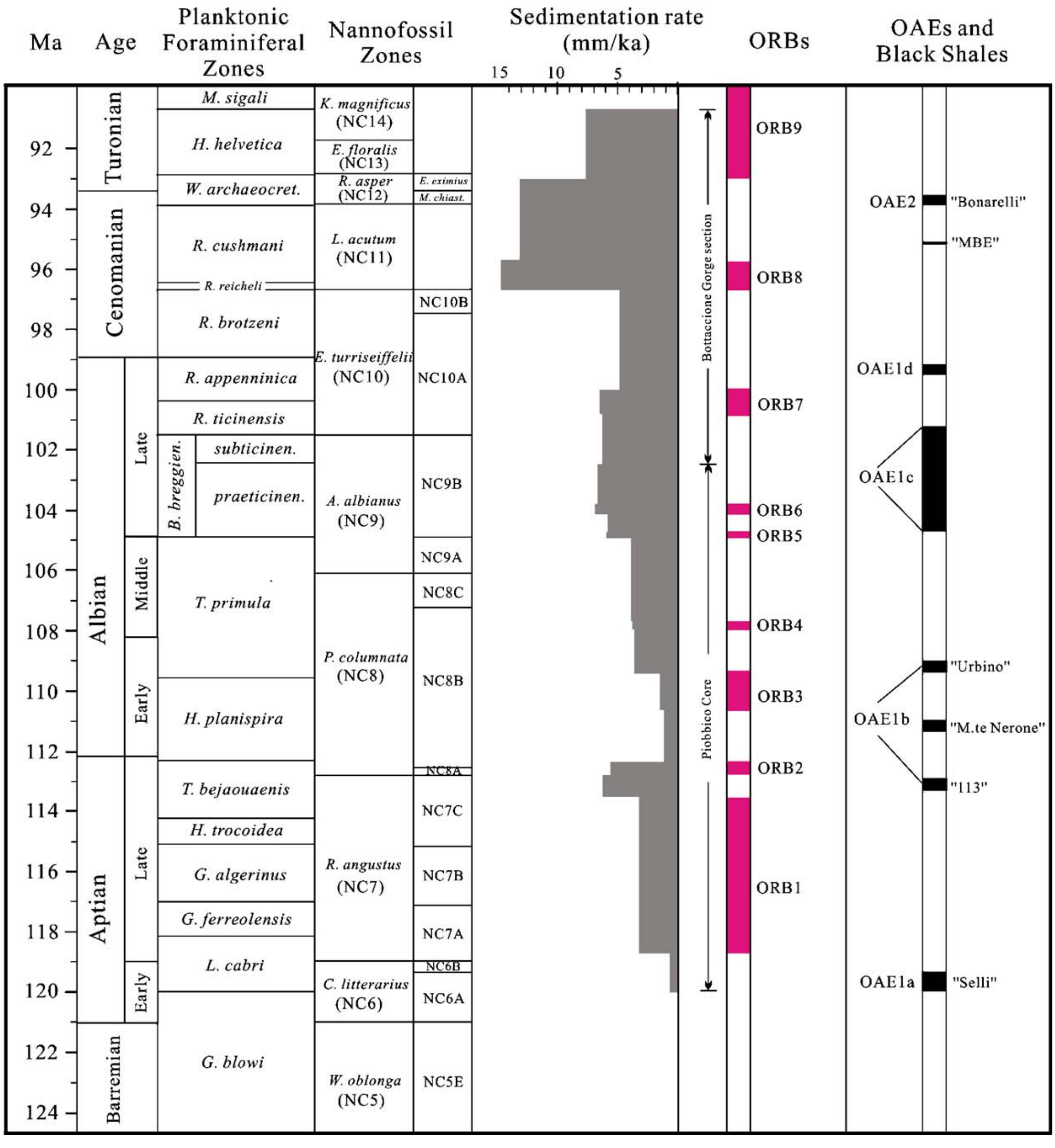


Fig. 9. The schematic summary of upper Barremian to Turonian stratigraphy in the Umbria–Marche Basin, the position of oceanic red beds (ORBs) and ocean anoxic events (OAEs). The mid-Cretaceous integrated calcareous plankton biostratigraphy is after Leckie et al. (2002); sedimentation rates are calculated from combined Piobbico core and the Bottaccione Gorge section.

escapes the oxidative destruction and is buried in marine sediments. According to these authors, the rate at which the molecular oxygen is liberated during production of this quantity of organic matter is sufficient to double the present amount of O_2 in the atmosphere in ca. 3 my. Therefore, could the increased burial of organic carbon in OAEs influence atmosphere composition? A similar mechanism was implied by Arthur et al. (1988), when they proposed that a change from mid-Cretaceous organic carbon-enriched shales to Late Cretaceous red beds was the result of "CO_2 drawdown." According to the latter authors, at the rate of burial of excess C_{org} that occurred at the Cenomanian/Turonian boundary

(Bonarelli level), it would take only 10,000–20,000 years to strip the atmosphere of CO_2. However, it has been suggested that the p_{CO_2} of mid-Cretaceous atmosphere may have been 4–8 times higher than at the present (Bice and Norris, 2002). Therefore, if for modelling p_{CO_2} was eight times greater than at present, then the burial of C_{org}. would be still sufficient to strip the atmosphere of CO_2 in 120,000-year episode. Such a CO_2 drawdown would add atmospheric O_2 and increase the oxidizing capacity of the oceanic deep waters (Arthur et al., 1988), if accompanied by a concurrent cooling period. It is conceivable that an increased burial of atmospheric CO_2, as C_{org} in marine sediments, could lead to a global cooling and production of colder, more oxygenated bottom waters (Voigt and Wiese, 2000). Arthur et al. (1988) implied that this mechanism was perhaps part of the explanation for the widespread onset of multicoloured, oxidized pelagic clay deposition at the deeper Atlantic sea floor during the post-early Turonian, after prolonged deposition of dark-coloured, C_{org}-rich sedimentary deposits. Can the alternation of OAEs and ORBs in the mid-Cretaceous be the result of such a triggering mechanism with the interval of light gray sediments separating OAEs and ORBs representing a delayed response of the system to a change in atmospheric oxygen?

2) An earlier hypothesis proposed by Arthur (1979) for the origin of Late Cretaceous ORBs in the North Atlantic suggested that a large reservoir of chemically reduced metallic ions existed in the anoxic pore water of black shales; these ions were mobilized and diffused upward and precipitated rapidly in the oxidized zone of slowly accumulating Upper Cretaceous deep-sea clays. As our observations of the mid-Cretaceous strata in the Umbria–Marche Basin show, the organic carbon-enriched beds are not directly overlain by red beds, as occurs in places in the North Atlantic, but are separated by various gray and light gray-colored sediments. Therefore, it is more than unlikely that there is a direct geochemical relationship between these two types of sediments as suggested by Arthur (1979).
3) Modern deep-sea red beds are deposited under low-productivity gyres in the Pacific Ocean, where the sedimentation rate is less than 1 mm/ka (Glasby, 1991). A similar origin can be applied to Late Cretaceous ORBs in the North Atlantic, which were deposited below the CCD at similar very low sedimentation rates of several mm/ka (Jansa et al., 1979). Muller and Mangini (1980) pointed out that for Pacific deep-sea sediments, sedimentation rates of <1cm/ka allow the sediment column to remain oxic for a sufficient length of time to completely oxidize organic matter. Modern deep-sea sediments in the southern Nares abyssal plain (water depth 5770 m) and the Madeira abyssal plain (water depth 5390 m) in the North Atlantic were deposited at higher sedimentation rates of ~2 cm/ka, and yet the sediment column remains oxidizing for a sufficient length of time so that most of the organic matter is remineralized, which in turn aids $CaCO_3$ dissolution and the diagenetic environment is oxic, with deposited sediment acquiring its red color due to the presence of ferric oxides (Cranston and Buckley, 1990). As we presented above, ORBs in the mid-Cretaceous of the Umbria–Marche Basin were deposited at rates ranging from 1.8 mm/ka for the ORB3 to 14.7 mm/ka for the ORB8 (Fig. 9), therefore, in an oxidizing envelop for modern, red deep-sea sediments. The sedimentation rate of the gray beds intercalated with red beds have apparently similar sedimentation rates that vary from 1.2 mm/ka for the gray beds between Selli Level and ORB1 to 13.4 mm/ka for the white beds between Bonarelli Level and ORB8 (Fig. 9). Thus, even though the theory of bottom sediment oxidation due to very low sedimentation rate could be applied to the origin of mid-Cretaceous ORBs in the Umbria–Marche Basin, oceanographic conditions were much more complex, as indicated by rates of intercalated gray beds. However, the latter may represent turbidites or periods of increased organic matter burial, which is not supported by Arthur and Fischer (1977) and Alvarez and Montanari (1988).
4) Fischer (2002) suggested that ORBs in the Umbria–Marche Basin may be the result of periodic, intensive downwelling. That such process operated during mid-Cretaceous in parts of the Tethys has been demonstrated by ODP drilling of the Blake–Bahama slope (Norris et al., 1998), where in the Aptian to Cenomanian sedimentary succession several zones of reddish colored sediments occur. Therefore, downwelling can be applied to areas near or on the continental margins, thus, to be of local extent. Lack of indicators of the margin, or close margin proximity of the mid-Cretaceous sedimentary deposits of the Umbria–Marche Basin, does not provide convincing support for the application of such theory, even though it is possible that such a “shallow basin” could have had different ocean circulation and the water mass structure from the deep oceanic basin.
5) If we accept as valid theories of increased bioproductivity to be the main cause of OAEs (e.g., Weis-

sert, 1991; Erba, 1994; Hochuli et al., 1999; Kuypers et al., 2002), then productivity-driven changes in the organic carbon flux to the bottom could explain the red/gray/green/black aspect of color variation (Dean et al., 1984; Wortmann et al., 1999). According to such theories, delivery to the bottom would decrease during intervals of low productivity, C_{org}, resulting in reduced oxygen demand, establishment of oxic bottom conditions and deposition of greenish, or reddish colored sediments. In contrast, during periods of higher productivity, high consumption of oxygen by the breakdown of organic matter would lead to the depletion of dissolved oxygen and development of dysoxic, or anoxic bottom conditions, a reducing diagenetic environment and sediment colors from green to dark grey and black. Despite the fact that only cursory stable isotope data were published for the Scaglia Bianca of the Umbria–Marche Basin (Jenkyns et al., 1994; Stoll and Schrag, 2000; Fig. 5), the shift to lower $\delta^{13}C$ during ORB1, ORB4 and ORB8 may suggest decreased productivity during these events. Several different processes such as decreased bioproductivity, climate and/or sea-level change, change in runoff, upwelling, or downwelling (Fisher and Arthur, 2002) could result in changes in redox conditions at the ocean floor. However, we lack detailed geochemical studies of ORBs to assess if changes in the bioproductivity occurred during their development.

6) The oxygen isotopic record indicates a prolonged interval of global warming from the Albian to the Turonian (Huber et al., 1995, 2002; Bralower et al., 2002). A large positive excursion of 1.5‰ in the lower Cenomanian correlates with the lower part of the ORB8 (Fig. 5). If the oxygen isotope variations do reflect primary climatic variations, then it would suggest a brief cooling period during the early Cenomanian, which may have resulted in increased production of oxygen-rich bottom waters. Alternatively, this oxygen isotope change could be caused by differential alteration of the sediments during diagenesis. There are no detailed oxygen isotope data for the other ORBs. Negative excursions in $\delta^{13}C$ associated with some ORBs (Fig. 6) point to periods of decelerated carbon cycling, low p_{CO_2} and perhaps cooler climate. Weissert and Lini (1991) used C-isotopes to trace changes of paleoclimate during the Aptian. They concluded that cooling events occurred during the late Aptian (*Globigerinelloides algerianus* zone) and the early Albian (*Ticinella bejaousensis* zone). However, as we show in Fig. 2, ORB1 were deposited during warm, cool and again warm climate periods and its deposition terminated during an onset of the next cool period.
7) Another mechanism to explore is that an inflow of colder and, thus, more oxygenated bottom waters could lead to changes in redox conditions on the ocean floor. The Early to mid-Cretaceous was a period of rapid sea level fluctuations and of tectonic upheaval (Ziegler, 1988; Larson, 1991; Ricou, 1996), leading to dramatic paleogeographic changes. Such changes emanated from the northward movement of Africa, counterclockwise rotation of Asia, northward motion of the Indian plate and opening of the seaway between South America and Africa at the equatorial region. All these represent important mechanisms for major changes in the ocean circulation patterns and water properties (Poulsen et al., 1998) and therefore predict a dynamic Cretaceous climate and marine environment. Tectonic changes with an effect on the ocean basins topography and physiogeography could have been one of major contributors to changes in the ocean bottom circulation patterns and therefore contributing element to the supply of colder, more oxygenated deep waters to some of the western Tethys subbasins. Thus, the origin of some of ORBs in the mid-Cretaceous could have been driven by changes in bottom ocean circulation and not by changes in bioproductivity, and thus, they may be an indirect, regional, or global response to changes in plate tectonics. Support for this conclusion is that not all ORBs are synchronous (see Fig. 8) and that not all are present in the areas we studied.

We consider it very important that some of the mid-Cretaceous ORBs, now exposed in Alpine and Carpathians terrains, are synchronous with organic carbon-enriched "black shales" deposited in the central North Atlantic (Hatteras Formation, Jansa et al., 1979). Therefore, bottom water chemistry, or water structure, such as vertical and horizontal distribution of waters with higher content of dissolved oxygen in the western Tethys had to be different from the eastern Tethys. Could this indicate different chemical fractionation of oceanic waters between individual subbasins, or on the larger scale between the western and eastern Tethys in the Cretaceous global ocean system? The periods of deep sea oxic environments were relatively brief during the mid-Cretaceous, as represented by ORBs 1–8, but it became a dominant state of the central north Atlantic basin during the Late Cretaceous (Jansa et al., 1979). During the Turonian–early Eocene interval, noncalcareous deep-sea clays were deposited below the CCD in the central North Atlantic basin and in some subbasins along western Tethys margin. At the same time in the shallower, above CCD oceanic setting of the western Tethys, pink and reddish colored pelagic carbonate and marl facies similar to Scaglia Rossa in the Umbria–Marche Basin were also deposited in an oxic environment (Hu et al., 2005). This indicates that oxic bottom conditions expanded from deep into intermediate oceanic waters during this time interval.

Our study does not provide a simple mechanism for deposition of ORBs occurring in the mid-Cretaceous of the Umbria–Marche Basin in Italy. The study provided evidence that several factors may have acted in a concert leading to periodic development of oxic bottom conditions in the Mediterranean Tethys. We conclude that it could have been tectonic changes that allowed periodic inflow of colder, more oxygen-rich bottom waters into subbasins of the circum-Mediterranean as suggested by Premoli Silva et al. (1989), perhaps with the content of dissolved oxygen in water increased during cooler climate periods. Low sedimentation rates allowed oxygen from the overlying seawater to diffuse downward, oxidizing the organic matter and precipitating iron oxides, which give the red color to the ORBs. More sophisticated geochemical studies are needed to reveal processes that led to periodic changes in the redox conditions at the ocean floor and resulted in the development of the mid-Cretaceous ORBs.

7. Conclusion

Study of the mid-Cretaceous sedimentary sequences in the Umbria–Marche Basin, Italy, demonstrates that aside from organic carbon-enriched beds, there occur several zones of reddish and pinkish colored sedimentary beds (ORBs), which mostly lack any accumulation of organic carbon. The presence of such beds indicates periodic development of oxic depositional conditions at the ocean bottom. The occurrence of such beds has been mostly overlooked in an effort to develop a general model of a greenhouse climate, and a deep dysoxic/anoxic ocean environment for the mid-Cretaceous. Such ORBs are exposed in the Gubbio–Piobbico area in Aptian to Cenomanian strata, defined by the Selli Level (OAE1a) at the base and by the Bonarelli Level (OAE2) at the top. In this sedimentary sequence, eight horizons with pink or reddish colored oceanic red beds were identified and numbered from the base to the top-ORB1 to ORB8. Two of these horizons occur in the Aptian, five in the Albian and one in the Cenomanian (Fig. 1). It appears that the number of ORBs and OAEs in the Umbria–Marche Basin is not only similar, but that they mostly reoccur alternately. Their stratigraphic position varies, as does the duration of individual events. The longest duration of 4.54 my we calculated for ORB1, which has a thickness of 17.51 m in the Piobbico core and spans the *G. ferreolensis* zone to the *T. bejaouaensis* zone. The shortest time duration of 0.13 my is ORB4, which in the Monte Petrano section is represented by a 0.6-m-long interval and occurs in the *Ticinella primula* zone. The mid-Cretaceous ORBs occur not only in the Umbria–Marche Basin, central Italy, but are widely distributed, with known occurrences in the Alps, Carpathians, northern Caucasus, central North Atlantic, northeastern England, and in western Himalayas. It is important to note that not all are present at all studied areas of western Tethys (Fig. 8). But their broad regional distribution is evidence that at least some represent hemispherical paleoceanographic events, while the others could be results of local paleoceanographic conditions.

The oxygen isotope records from the Cretaceous strata of the Gubbio–Piobbico area have been interpreted as indicating a warming trend from the Albian to Turonian (Stoll and Schrag, 2000). Such records show a lack of correlation between occurrence of ORBs and oxygen isotope shifts, except for positive excursion of $\delta^{18}O$ at the Middle Cenomanian, which correlates with ORB8 and may indicate a brief period of cooling. Several studies (see Introduction) indicate the existence of cool periods during the Early Cretaceous. The ORB1 in the Piobbico area which in part seems to correspond to a lowest value in $\delta^{13}C$ after post-Selli positive excursion (Erbacher and Thurow, 1997; Weissert et al., 1998), extends from warm to cool and through the next warm period, with the deposition of red beds terminated at the onset of the next early Albian cool period (Weissert and Lini, 1991). If carbon isotope data are interpreted correctly, then there is no undisputable relation between climate change and ORBs deposition, even though colder climate should theoretically result in a production of bottom waters richer in dissolved oxygen. Another broad negative shift in carbon isotope occurs in the late Albian and correlates with ORB7 and may indicate that some of the ORBs could be associated with decreased bioproductivity; however, higher resolution geochemical data is needed to support this interpretation.

It is widely assumed and applied to a score of paleoceanographic/paleoclimate models that dysoxic to anoxic bottom conditions were the common state of the mid-Cretaceous ocean. The occurrence of ORBs in the mid-Cretaceous marine strata document that this was not the case and that the ocean system was much more dynamic and environmentally less stable. If, in the background of these changes were changes in the paleoclimate, then the presence of ORBs in so-called mid-Cretaceous "greenhouse" period would suggest the existence of brief periods of a probably cooler climate, similar to the Early Cretaceous (Weissert and Lini, 1991; Hochuli et al., 1999; Weissert and Erba, 2004). Alternatively, development of ORBs could be triggered by changes in the deep water circulation, which to resolve would need much more detailed reconstructions of the ocean bottom topography and paleogeograpahy than currently available.

Acknowledgements

We thank Francesco Baroncini for the valuable assistance during field studies. We acknowledge information provided by Krzysztof Bak (Poland), Michael Wagreich (Austria), Nataliya Tur (Russia).

We are thankful to Franc Thomas (Canada) and R. W. Scott (USA) for correcting the English text and to reviewers of the previous draft of this study—I. Premoli Silva, R. Coccioni, and S. Galeotti and E. Erba for their helpful comments and constructive suggestions. We are thankful to Prof. Chengshan Wang for his encouragement and support. This study was in part supported by a postdoctoral fellowship at Universita' Politecnica delle Marche, Italy and at the Laboratory of Marine Geology, Tongji University; by the MOST 973 Program (2006CB701402) and NSFC Project (40332020) to X. Hu, and by grants from the Ministry of Foreign Affairs, Italy to M. Sarti. This is a contribution to UNESCO International Geosciences Project 494 and 463.

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