

High-resolution biostratigraphy and chemostratigraphy of the Cenomanian/Turonian boundary event in the Vocontian Basin, southeast France

Danièle Grosheny^{a,*}, Bernard Beaudoin^b, Laurence Morel^b, Delphine Desmares^{a,b}

^a *Université Louis Pasteur, Strasbourg 1, EOST, UMR7517, 1 rue Blessig, F67084 Strasbourg Cedex, France*

^b *Ecole des Mines de Paris, CGES Sédimentologie, 35 rue St Honoré F77305 Fontainebleau Cedex, France*

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Abstract

The Cenomanian/Turonian boundary black shale has been traced in southeastern French subalpine ranges in several sections correlated over tens of kilometres, both on a bed-by-bed basis and with the control of $\delta^{13}\text{C}$ isotope curves. Correlations show a thickness increase slopeward and a slight increase in the time span of black shale deposition basinward. They also show that the Vergons section cannot be used as a local reference section for Ocean Anoxic Event 2 (OAE2) because it has probably been affected by synsedimentary sliding. Finally they show that the vertical range of the *Whiteinella archaeocretacea* Zone changes considerably over short distances. This could be easily understood in widely separated basins but is harder to explain at such a local scale. In addition, using results obtained in the US Western Interior Basin, we question the validity of this Partial Range Zone.

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1. Introduction

Since the late 1970s, Oceanic Anoxic Event 2 (OAE2) (Schlanger and Jenkyns, 1976; Arthur and Schlanger, 1979; Jenkyns, 1980; Jenkyns, 1997) has drawn continuous attention because of its short duration and because of the superimposition of three crises during the event: a sedimentary one with the wide scale occurrence of black shales (Schlanger and Jenkyns, 1976; Arthur and Schlanger, 1979; Hart and Bigg, 1981; Arthur and Premoli Silva, 1982; Amédéo and Robaszynski, 1993; Arthur and Sageman, 1994); a planktic biologic one with the temporary disappearance of keeled planktic foraminifera in the world's oceans (Eicher, 1969; Eicher and Worstell, 1970; Sigal, 1977; Hart and Bigg, 1981; Caron and

Homewood, 1982; Eicher and Diner, 1985; Leckie, 1985; Hart and Ball, 1986; Hart, 1996; Lamolda et al., 1997; Grosheny and Malartre, 1997; Keller et al., 2001; Caron et al., in press); and a geochemical one with a positive $\delta^{13}\text{C}$ excursion, also recorded in many basins both on the carbonate and organic-matter fractions (Pratt and Threlkeld, 1984; Pratt, 1985; Hilbrecht and Hoefs, 1986; Schlanger et al., 1987; Jarvis et al., 1988; Jenkyns et al., 1994; Accarie et al., 1996; Luderer and Kuhnt, 1997; Morel, 1998; Paul et al., 1999). A number of sections have been studied worldwide, some of them eventually becoming reference sections, such as the Eastbourne section in the Anglo-Paris Basin (Jarvis et al., 1988; Gale et al., 1993; Hart et al., 1993; Paul et al., 1999), the Pueblo section in the US Western Interior (Pratt and Threlkeld, 1984; Pratt, 1985; Eicher and Diner, 1985; Leckie, 1985; Pratt et al., 1993; Leckie et al., 1998; Morel, 1998), the Gubbio section in the northern Tethys (Arthur and Premoli Silva, 1982; Corfield et al., 1991; Premoli Silva and Sliter, 1995; Luciani

* Corresponding author.

E-mail address: grosheny@illite.u-strasbg.fr (D. Grosheny).

and Cobianchi, 1999), the Wunstorf section in boreal German basins (Weiss, 1982; Ernst et al., 1984; Hilbrecht, 1986), and the Tarfaya section on the Atlantic margin (Kuhnt et al., 1990; Luderer and Kuhnt, 1997). The event has been also recognized in the deep ocean at many DSDP and ODP sites. Regional syntheses have been attempted like that of Lüning et al. (2004) for the North African craton. Basinal deposits, in both the deep ocean and the deepest parts of epicratonic basins, usually include one or several black shale layers close to the Cenomanian/Turonian boundary, but shallower deposits may be devoid of them. These nevertheless show the $\delta^{13}\text{C}$ anomaly (work in progress in Saharan and Middle East carbonate sections devoid of black shales), which appears to be the best way of correlating sections worldwide in different depositional environments.

The event is also recorded in the French subalpine basin where the “Thomel level” (Crumière, 1990; Crumière et al., 1990) of the Vergons section has been termed as a local equivalent of the “Bonarelli level” (Arthur and Premoli-Silva, 1982) of the Gubbio section in Italy. The Vocontian Basin is interesting because many sections can be studied and correlated on a bed by bed basis in both basinal and slope deposits. In this paper we deal only with deep basinal sections that include the black shales.

The Cenomanian/Turonian boundary is difficult to define purely from a palaeontological viewpoint. Using ammonites, the boundary lies between the last occurrence of *Neocardioceras juddii* and the first occurrence of *Watinoceras coloradoense* or *W. devonense* (Bengston, 1996). In the thin Pueblo section (US Western Interior) the boundary is between two successive limestone beds, 84 and 86 (numbering system of Cobban and Scott, 1972; Kennedy et al., 1999, 2000). The transitional interval is as thin as an interbed, which is among the reasons why this section was proposed for the boundary reference section during the Second International Symposium on Cretaceous Stage Boundaries (Bengston, 1996). In other basins, the transitional interval is usually thicker owing to either the lack of ammonites or to a greater accumulation rate. With planktic foraminifera, the boundary is more difficult to define. The *W. archaeocretacea* Partial Range Zone (or “zone à grosses globigérines” of Sigal, 1977) has been defined (Robaszynski et al., 1979; Robaszynski and Caron, 1995) between the last occurrence of the upper Cenomanian index species *Rotalipora cushmani* and the first occurrence of the lower Turonian index species *Helvetoglobotruncana helvetica*, within which is the Cenomanian/Turonian stage boundary. This *W. archaeocretacea* Zone has been proven recently to be extremely reduced in the Pueblo section (Beaudoin et al., 1995; Morel, 1998; Desmares et al., 2004). Beaudoin et al. (1996) have shown that in the lower Cenomanian of Italy, the vertical range of planktic foraminifera may vary markedly between closely spaced sections within the same basin. We have used the closely spaced basinal sections of the Vergons area similarly to investigate the subalpine basin at the Cenomanian/Turonian boundary.

It has been shown that several sections in distant basins (Eastbourne, Wunstorf, Tarfaya) show the same three-peaked

$\delta^{13}\text{C}$ positive excursion pattern as that defined in the Pueblo section (Pratt and Threlkeld, 1984). We have used this particular pattern to correlate our sections in the subalpine basin. High-resolution correlations based on both bed bundles and $\delta^{13}\text{C}$ curves thus enabled us to check the vertical ranges of planktic foraminifera against anoxic layers, as demonstrated in this paper. We also show that the Vergons section proposed by Crumière et al. (1991) as the local reference section is unsuitable because it is affected by synsedimentary sliding.

2. Geological setting

The study area is located in the Castellane tectonic arc area of the southern subalpine ranges in southeast France (Fig. 1A). The northern part of the arc belongs palaeogeographically to the eastern reaches of a relatively deep basin known as the Vocontian Trough. The southern part of the arc corresponds to the outer part of the Provence Platform situated further to the south. The slope area in between was relatively stable throughout the Cretaceous Period (see facies maps in Debrand-Passart et al., 1984). The geographic extension of the black shale basin in southern subalpine ranges is shown on Fig. 1B. The Cenomanian basinal facies comprise alternating mudstones and marlstones, limestone beds occurring as bundles, 10–20 m thick, separated by marly intervals equally as thick or sometimes thicker. The Turonian succession comprises stacked limestone beds forming a distinct cliff. Between the last definite upper Cenomanian bundle of beds and the base of the “Turonian” limestones, lies an interval of alternating marlstones and limestone beds, 10–15 m thick, comprising several black shale layers (Crumière, 1990). The mudstone facies is not fully pelagic as seen in the southern Tethyan counterpart, i.e., in the Gubbio section in northern Italy. Large amounts of bioclastic silt and sponge spicules seen in thin section indicate a hemipelagic setting and thus a probable moderate depositional depth (a few hundred metres) in the black shale area. To the south, in a slope setting, thicknesses increase dramatically by a factor 10–15 depending on location. At the southern limit of the study area (Fig. 1B), the facies become shallower with evenly laminated, sandy tempestites layers in the lower Cenomanian (Conard-Noireau, 1987), occurrences of oysters in upper Cenomanian mudstones, and bundles of calcarenites with hummocky cross-stratification in the lower Turonian, all indicative of an outer platform depositional setting.

3. Material and methods

Six sections (Fig. 1A) spread over 100 km² (Ondres, Pont d’Issole, Lambruisse, Font de Poule, Sérec, Vergons) north of Castellane have been studied for lithology, micropalaeontology (benthic and planktic foraminifera) and geochemistry. Some data have already been published (Tronchetti and Grosheny, 1991; Grosheny et al., 1992; Grosheny and Tronchetti, 1993; Grosheny and Malartre, 1997; Morel, 1998; Grosheny, 2002). The Ondres section is described here for the first time. Three of the sections are illustrated in this

paper (Ondres, Pont d'Issole and Vergons), the Vergons section being located at the base of slope, and the Ondres section being the most distal (Fig. 1).

The Pont d'Issole section has been sampled over 25 m across the black shale boundary layer. Fifty-two samples were taken for foraminiferal analysis. Carbon isotope geochemistry has been carried out on 49 samples. In the Vergons section 120 samples were taken over a thickness of 36 m in order to study faunal changes within minor lithologic variations. Isotope analyses have been carried out on 20 samples. In the Ondres section, 38 samples were studied for micropalaeontology only.

The foraminiferal analysis was carried out both on washed samples (light-grey marlstones and dark shales) and in thin sections (limestone beds) in order to eliminate sampling bias. Marly samples were put into a mixture of water and a small amount of 110 vol. hydrogen peroxide for 12 h, then washed using sieves of 1 mm and 63 μm mesh. Residues were dried at 40 °C. Foraminifera were identified and counted until a total of 250 specimens was reached, if possible. The material is well-preserved, preventing any problems with identification. Determinations were made using the systematics of Loeblich and Tappan (1988). The list of species found is given

in the Appendix. The planktic zonal schemes used are those of Robaszynski et al. (1979), Caron (1985) and Robaszynski and Caron (1995).

Counts were made from washed residues only, thin sections being used only for more accurately determining the first and/or last appearances of individual taxa. Several parameters were calculated for planktic foraminifers: abundance, species richness, relative abundance of planktics (% planktics / % planktics + % benthics, noted as planktic % on figures) and relative abundance of keeled planktics vs. total planktics (noted as keeled %). These parameters are widely used in recent deposits to define K and *r* strategies (MacArthur and Wilson, 1967) but must be used with caution in ancient deposits, according to Gibson (1989). In periods of K selection the reproduction system of evolved forms allows a great diversification of morphotypes, with the occurrence of large, keeled, deep-water dwelling forms (*Rotalipora*, *Dicarinella*, *Margino-truncana*, *Helvetoglobotruncana*) and large trochospiral forms. When environmental conditions are more extreme, an *r* strategy takes place. This is represented by low specific diversity, a great number of individuals, and the disappearance of keeled forms, these being replaced by small, globular, trochospiral, biserial or planispiral forms (*Hedbergella*, *Whiteinella*, *Heterohelix*, *Globigerinelloides*). Any increase in these parameters is interpreted as indicating open, well-oxygenated deep waters, allowing a K-strategy. Conversely, any decrease means the establishment of an *r* strategy, coincidental with poorly-oxygenated deep waters.

Carbon isotope analyses were made on whole rock samples. Organic matter was eliminated by putting samples in a solution of 5% sodium hypochlorite at 5% for 12 h. The $^{13}\text{C}/^{12}\text{C}$ ratio was determined using a Fison VG Optima mass spectrometer, the carbonate being attacked with phosphoric acid. Results are expressed in permil deviation vs. PDB standard reference. Any error measurement is lower than 0.06‰.

4. Results

Lithologic, isotopic and gross biostratigraphic data obtained in the Ondres, Pont d'Issole and Vergons sections are illustrated on Fig. 2.

4.1. Lithology

Morel (1998) defined three lithological units (M1–3, Fig. 2) in the interval encompassing the OAE2. These units comprise laminated black shale layers, grey bioturbated marlstones and light grey bioturbated limestone beds.

At Pont d'Issole, the M1 unit, which is 8 m thick, comprises a recessive shaly interval (5.5 m thick) at the base and of a bundle of limestone beds (2.5 m) at the top. The lower part consists of dark laminated shales rich in radiolarians but planktic foraminifera are scarce. The overlying bed bundle comprises bioturbated wackestones and marlstones rich in planktic and benthic foraminifera, radiolarians and inoceramid debris, with minor amounts of sponge spicules, echinoid debris, sharks' teeth and ostracods. The M2 unit, also 8 m

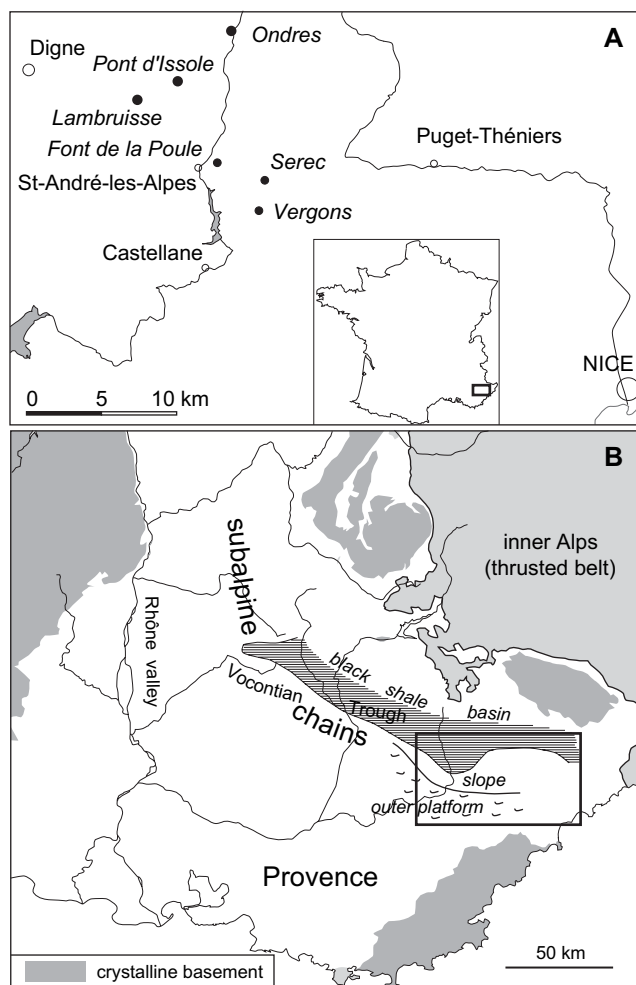


Fig. 1. A, location of the sections studied and/or cited in the study area. B, location of the black shale basin in southern French subalpine ranges.

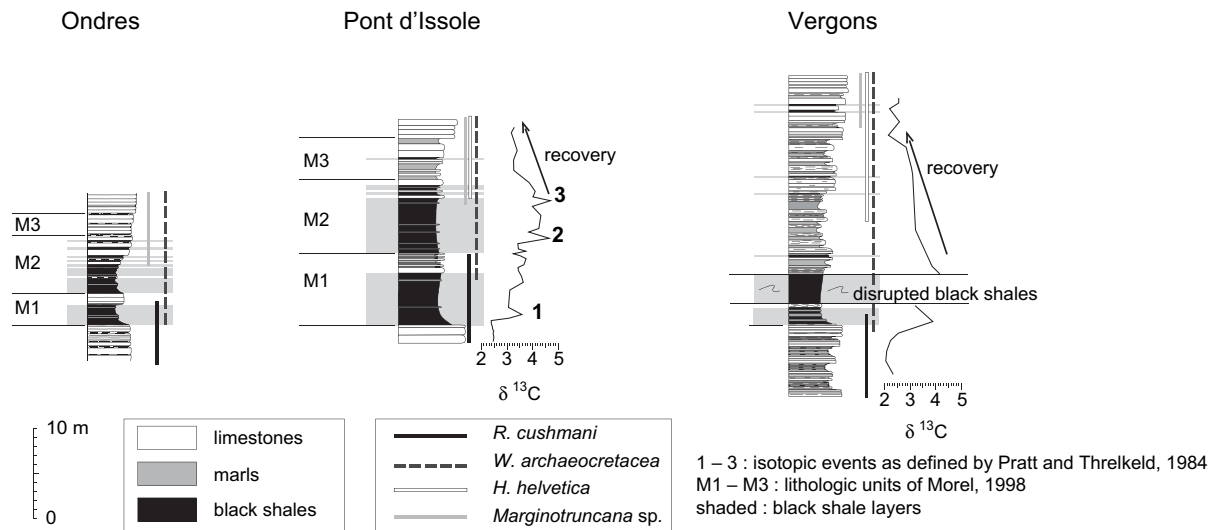


Fig. 2. Stratigraphic range of the index foraminiferid taxa and position of the $\delta^{13}\text{C}$ isotope anomaly versus the black shale interval in the sections studied.

thick, is very similar to the underlying unit in both facies and microfossil content. Coal debris occurs in some horizons. Unit M3 is thinner (5 m) with a thin recessive interval at the base. Limestone beds are wackestones that are very rich in planktic foraminifera and poor in other organic remains (benthic foraminifera, radiolarians, ostracods, inoceramid debris, sharks' teeth and bryozoan debris). The interbeds are marlstones.

These three units have been traced in other sections, including the unfigured Lambruisse and Font de la Poule locations (Fig. 1A), and are easily recognisable in the new Ondres section (Fig. 2), where their thicknesses are reduced. The first occurrence of lower Turonian *Marginotruncana* at the top of the M2 unit in both the Ondres and Pont d'Issole sections supports the correlation, as does the last occurrence of *R. cushmani* within the bed bundle at the top of the M1 unit. If the correlations drawn are correct, slight differences exist between the sections. For instance, the bed bundle at the top of the M1 unit at Pont d'Issole equates with a single composite bed at Ondres, the interbeds present at Pont d'Issole being reduced to thin seams at Ondres. However, comparison with the Vergons area (Sérec section, Fig. 1) shows greater changes in the succession, even if they are not readily apparent. At first sight, the small bundle of beds in the middle of the recessive interval at Vergons, immediately overlying the disrupted black shales (Fig. 2), could have been correlated with the bundle of beds topping the M1 unit in the Pont d'Issole section. This is apparently supported by the first occurrence of lower Turonian planktic foraminifera just above, i.e., in what could have been interpreted as the M2 unit. However, tracing the beds in intermediate sections such as at Sérec suggests there is a hiatus in the Sérec and Vergons sections and that the M1 unit is lacking (Morel, 1998).

4.2. Carbon isotopes

The Pont d'Issole section shows a $\delta^{13}\text{C}$ curve (Morel, 1998) in which the three isotope events of Pratt and Threlkeld (1984)

can be clearly recognised (Fig. 2). In the US Western Interior, isotope event 1 occurs within the upper part of the *R. cushmani* Zone as at Pont d'Issole and event 3 is within the base of the Turonian as in southeast France. The isotope curve obtained in the Vergons section is atypical (Fig. 2) and supports the idea of a hiatus in this section. A first event occurs at the base of the black shale interval, as in the Pont d'Issole section, but Morel (1998) considered this to be equivalent to event 2 of Pratt and Threlkeld (1984). No analyses have been made in the disrupted interval at Vergons, but what occurs above it is similar to the trend seen in the Pont d'Issole section, i.e., the "recovery" to normal values of the isotope ratio, or the end of the $\delta^{13}\text{C}$ excursion. All this suggests that part of the M1 and/or M2 units of Pont d'Issole are absent at Vergons, either because of non-deposition or for some other reason.

4.3. Foraminifera

4.3.1. Planktic foraminifera

Twenty-nine species of planktic foraminifera were recognised during the course of this study. Their vertical distribution is shown in Figs. 3 and 4. These are based on observations on assemblages from marly interbeds (washed samples) rather than from limestone beds (thin sections).

Biostratigraphy. Three major assemblages can be defined, with minor differences between sections. The first assemblage is diverse and recorded from the upper Cenomanian bundle of limestone beds underlying the black shale interval to the top of the M1 lithologic unit. It comprises *Heterohelix globulosa* (Ehrenberg), *H. moremani* (Cushman), *Globigerinelloides ultramicra* (Subbotina), *Schackoina cenomana* (Schacko) *Hedbergella simplex* (Morrow), *H. delrioensis* (Carsey), *H. planispira* (Tappan), *H. hoelzli* (Hagn and Zeil), *Whiteinella brittonensis* (Loeblich and Tappan), *W. paradubia* (Sigal), *W. baltica* (Douglas and Rankin), *W. aprica* (Loeblich and Tappan), *W. archaeocretacea* (Pessagno), *Praeglobotruncana*

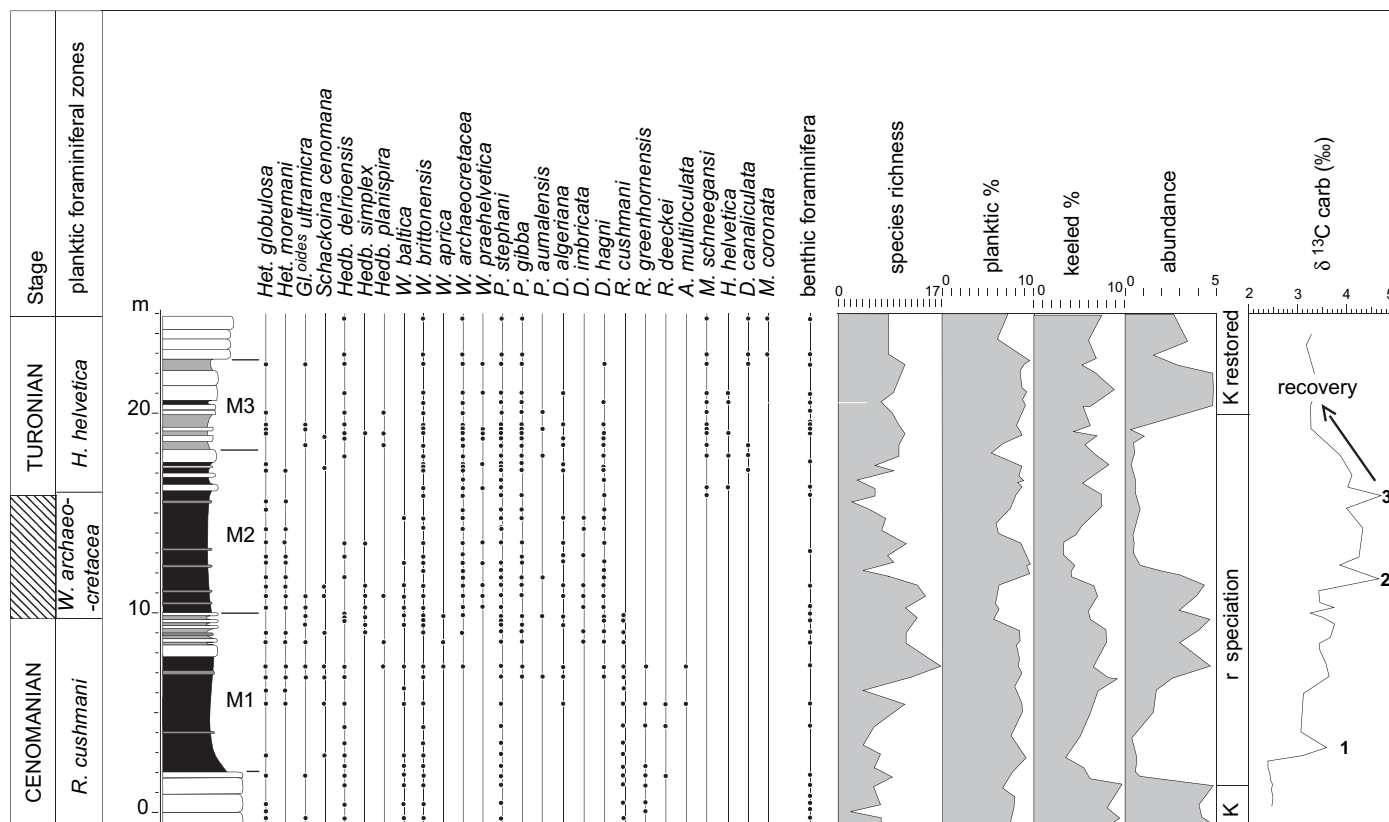


Fig. 3. Biostratigraphic and palaeoecologic data in the Pont d'Issole section.

stephani (Gandolfi), *P. gibba* (Klauss), *P. aumalensis* (Sigal), *Dicarinella algeriana* (Caron), *D. hagni* (Scheibnerova), *D. imbricata* (Mornod), *Rotalipora cushmani* (Morrow), *R. deeckei* (Franke), *R. greenhornensis* (Morrow), and *Anaticinella multiloculata* (Morrow).

W. archaeoeretacea occurs higher in all sections. *H. planispira* is found only at Pont d'Issole and *H. hoelzli* only at Vergons. The last occurrences of *R. deeckei*, *R. greenhornensis*, *A. multiloculata* and *R. cushmani* are found in this stratigraphic order. They disappear earlier at Vergons (at the base of the black shale interval) than at Pont d'Issole and Ondres (at the top of the M1 unit). This has to be discussed further because of the hiatus problem suggested above.

This first assemblage is characteristic of the upper part of the *R. cushmani* Total Range Zone owing to the presence of *R. deeckei*, i.e., the upper Cenomanian. The second is found in the lower, shaly part of unit M2. It is less diverse with reduced recovery. However planktic foraminifera are present throughout the black shales. *Rotalipora* and *Anaticinella* are absent. *W. praehelvetica* (Trujillo) and *P. elata* (Lamolda) are also present. This assemblage is characteristic of the *W. archaeoeretacea* Partial Range Zone. It should be noted that the interval of disrupted black shales seen at Vergons still contains species of the first assemblage, especially *Rotalipora*. This observation does not support the absence of the M1 unit at Vergons, as interpreted by Morel (1998) but it does support the idea that the first isotope event recorded in this section is event 1 instead of event 2.

The third assemblage is first encountered at the base of the upper calcareous part of the M2 unit and continues throughout the M3 unit. Five new taxa appear: *H. helvetica* (Bolli), followed by *Marginotruncana schneegansi* (Sigal), *M. coronata* (Bolli), *M. renzi* (Gandolfi) and *D. canaliculata* (Reuss). The first occurrence of *H. helvetica* is also at the base of the bundle of beds marking the top of the M2 unit at Pont d'Issole and Ondres. At Vergons, this first occurrence is within the M3 unit. The first occurrence of marginotruncanids is within the M2 unit at Ondres, at the top of M2 at Pont d'Issole, and within M3 at Vergons. The last occurrence of *W. baltica* also seems to be diachronous (see discussion below), i.e., it occurs earlier at Pont d'Issole than seen at Vergons. It is clear that the third assemblage belongs to the *H. helvetica* Total Range Zone; it equates with an early Turonian age, but not the earliest.

From the above discussion, the Cenomanian/Turonian boundary should be placed within the lower, shaly part of the M2 unit. The position of the boundary is now better defined than in earlier publications (Crumi re, 1990; Tronchetti and Grosheny, 1991; Crumi re et al., 1991; Grosheny et al., 1992; Grosheny and Tronchetti, 1993; Grosheny and Malartre, 1997).

Palaeoecology. Our discussion below is based on observations made from washed residues only. Parameter curves in Figs. 3 and 4 are thus constructed from fewer data points, corresponding to samples rich enough in foraminifera to allow

reliable counts (at least 250 individuals). Within the limestone bundle below the black shale interval, washing was carried out on samples from thin marly seams between the limestone beds.

Classical specimens of *Rotalipora* are found together with “atypical” forms of *R. cushmani*, as well as non-keeled forms with secondary sutural apertures on the umbilical face (globular morphotype) at both Pont d’Issole and Vergons (Figs. 3, 4). Atypical forms of *R. cushmani* show some morphological variation from forms smaller than the norm to morphotypes with fewer chambers or with a poorly developed keel and often a more pronounced dorsal convexity. Globular forms with secondary sutural apertures belong to the genus *Anaticinella* as defined by Eicher (1972). The morphotype observed in the subalpine basin has a greater number of chambers (6.5–9) in the last whorl. It is therefore attributed to *Anaticinella multiloculata* (Morrow). At both Pont d’Issole and Vergons, *A. multiloculata* is found over a few metres of section only but not in the same lithostratigraphic unit, namely below the black shale interval at Vergons and within the lower part of the M1 unit at Pont d’Issole. This species is so abundant in some horizons at Vergons that it may outnumber keeled forms (*Rotalipora*). In both sections, the disappearance of *A. multiloculata* coincides with that of *Rotalipora greenhornensis*. *R. cushmani* is present a little higher (through a few more metres) but its frequency drops dramatically.

Atypical forms of *R. cushmani* have been reported in other basins: the US Western Interior (Desmares et al., 2003); the lower Saxony basin in Germany (Grosheny, 2002); and the Moroccan basins of Agadir (Jati et al., 2003) and Tarfaya (Kuhnt et al., 1997).

Although long considered an upper Cenomanian, North American endemic form (Eicher, 1972; Leckie, 1985), *A. multiloculata* has now been reported from Eastbourne in England (Keller et al., 2001) and from Wadi Bahloul in Tunisia (Caron et al., in press). In the US Western Interior, the appearance of *Anaticinella* is interpreted to be a case of neoteny sensu Gould (1977) (Desmares et al., 2003). The persistence of juvenile characters and the loss of the peripheral keel in adult specimens apparently enabled the genus to survive global anoxia by staying in shallow waters throughout its life cycle (Desmares et al., 2003). However this selective advantage was insufficient to prevent extinction during times of expanded anoxia in marine waters as they finally disappeared at the same time as *R. greenhornensis* from both the US Western Interior and the French subalpine basin.

Atypical forms of *R. cushmani* found in the US Western Interior have been interpreted as intermediate between *R. cushmani* and *Anaticinella planoconvexa* (Longoria). *A. planoconvexa* would have evolved from *R. cushmani* in the same way as *A. multiloculata* was derived from *R. greenhornensis* (Leckie, 1985; Desmares et al., 2003). In the French subalpine basin these atypical forms do not reach the evolutionary stage of losing the keel as seen in *A. planoconvexa*.

The occurrence of *A. multiloculata* in several distant basins suggests that its appearance, development and disappearance could have been the result of an adaptation to similar

environmental changes in each of these locations. The problem is that in the French subalpine basin this “*Anaticinella* event” does not occur at the same time in nearby sections: e.g., it occurs below the black shale interval at Vergons and within the black shales of the M1 unit at Pont d’Issole.

The analysis of palaeoecologic parameters such as abundance, species richness, and planktic and keeled percentages in the subalpine basin shows homogeneous results. A number of changes may be defined. They are clear only in the Pont d’Issole section (Fig. 3) because of the hiatus problem in the Vergons section (Fig. 4). The cause and the position of this hiatus is discussed below.

Within the upper Cenomanian bundle of beds situated below the black shale interval, at both Vergons and Pont d’Issole, the four parameters show high values, except for species richness which is lower at Pont d’Issole (between 9 and 7) than at Vergons (between 16 and 7). Planktic percentages are between 60 and 90% and keeled forms are greater than 80% at Vergons and Pont d’Issole. Abundance is high. The dominance of trochospiral keeled morphotypes (*Rotalipora*, *Dicarinella*, *Praeglobotruncana*) adapted to well-oxygenated deep waters is indicative of a K adaptive strategy (MacArthur and Wilson, 1967), which is the rule within the whole *R. cushmani* Zone.

At the base of the black shales of the M1 unit, a rapid and dramatic drop of parameter values is recorded in both the Vergons and the Pont d’Issole sections. Planktic foraminifera are scarce but do not entirely disappear in black, shaly horizons. They are more abundant when the carbonate content of the marls increases. The keeled per cent index drops to values as low as 30%. Species richness drops to 3, but the planktic per cent index remains relatively high (>50%). This interval is characterised by the dominance of globular forms, often very small (*Whiteinella*, *Hedbergella*, *Heterohelix*), dwelling in shallow waters and unaffected by the anoxia or hypoxia in deeper waters. This change corresponds to the setting of an r adaptive strategy.

Normal values indicating the return to a K strategy resume definitively within the M3 unit, i.e., at the base of the lower Turonian limestones. In between, in the black shale interval, within the M1, M2 and the base of the M3 units, fluctuations occur. These roughly follow lithology in the Vergons section (Fig. 4) but not at Pont d’Issole (Fig. 3). For example, in the upper part of the M1 unit, parameter values are again almost as high as those occurring below the black shale interval, meaning a return to quasi-normal oceanographic conditions. Values remain high within the lowest part of the M2 unit in spite of the return to black shale facies. The next sharp drop in abundance and keeled percentages occurs within the black shales, not at the base of the M1 unit. Abundance stays low in the upper, more calcareous, part of unit M2 in spite of the fact that interbeds are no longer true black shales. Higher, within the M2 and M3 units, the keeled per cent index more closely follows lithology, but abundance remains low. So, if there is really an approximate correspondence between lithology and fluctuations in the parameters at the scale of the whole black shale interval, the link is not clear in the detail.

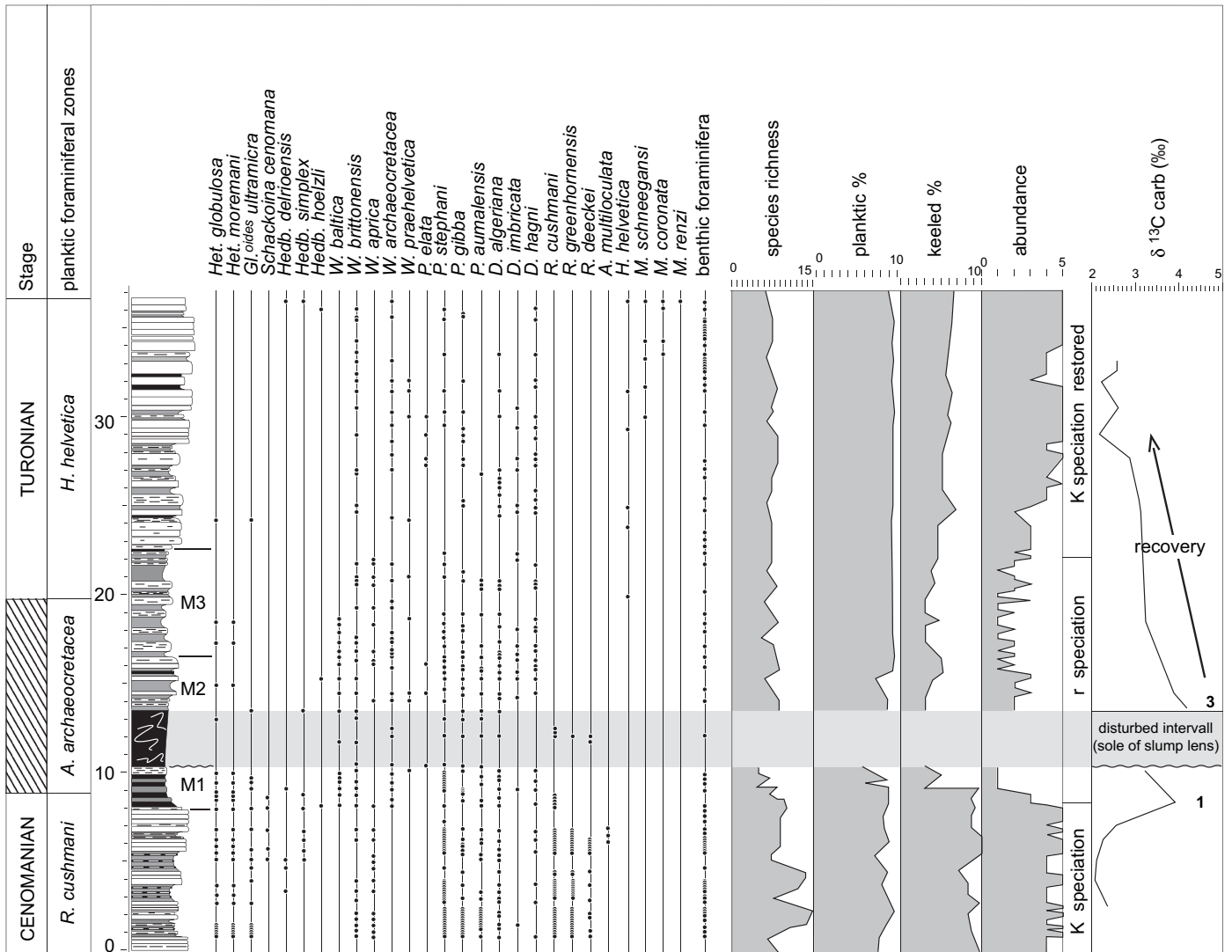


Fig. 4. Biostratigraphic and palaeoecologic data in the Vergons section.

The temporary return to quasi-normal conditions within the black shale interval at Pont d'Issole is not recorded at Vergons, again supporting the idea of a hiatus in the latter section.

4.3.2. Benthic foraminifera

Qualitative and quantitative analysis of benthic foraminifera has also been carried out on washed residues. Species richness is globally high since 70 species have been found in the Vergons and Pont d'Issole sections (Tronchetti and Grosheny, 1991; Grosheny and Tronchetti, 1993; Grosheny and Malartre, 1997; Grosheny, 2002). Species of the sub-order Textulariina are less abundant (up to 20% in some samples) than Lagenina and Rotaliina (up to 98% in some samples). Among the latter, Nodosariidae, Anomalinidae and Turrinellidae are dominant, indicating a base of slope depositional environment (Koutsoukos and Hart, 1990; Holbourn et al., 2001).

In marl seams within the bundle of beds below the black shale interval, benthic foraminifera are abundant, well-diversified and of normal size. Nodosariidae are dominant (up to 34%), rich in species (11) and morphotypes. All this indicates

normal oxygenated bottom conditions. Within the black shale interval, there is a marked decrease in abundance, species richness and size of individuals. There are also strong variations between the shaly layers and grey marls. Even within black shale layers variations may be strong, some horizons being barren of benthic species, others bearing almost normal assemblages. In the latter kind of black shale facies Textulariinae may make up to 20% of the assemblage and Gavelinellidae (up to 35%) are dominant over Nodosariidae (up to 25%) among the Lagenina. In the disrupted interval of the Vergons section (Fig. 5), benthics are scarce both in species richness and number of individuals.

These fluctuations probably reflect changes in bottom-water oxygenation, with alternating periods of true anoxia and periods of hypoxia, the latter allowing the re-establishment of a depleted microbenthos. The species found in these hypoxic levels are also found in the normal assemblages of the bed bundle below the black shale interval, suggesting that they are opportunistic species able to resist poor bottom conditions occurring during the black shale interval (Jarvis et al.,

1988; Koutsoukos et al., 1990). Among these opportunistic species, *Lenticulina* seem really adapted to poorly oxygenated environments (Honigstein et al., 1986). The genus *Praebulimina* is also often found elsewhere in hypoxic levels rich in organic matter or in black shale layers (van der Zwann et al., 1986; Verhallen, 1986).

Grey marly interbeds occur more frequently than black shales in the upper part of the black shale interval, and assemblages are very similar to those found at the base of the sections, with Nodosariidae becoming dominant again (up to 32%) together with Gavelinidae and Turriliniidae. However, in the few black laminated horizons still occurring high in the sections (Figs. 3, 4), benthics may be totally absent, as seen below in the main black shales, indicating that the return to normal oxygenated conditions in the lower Turonian was interrupted by short pulses of total anoxia.

5. Discussion

5.1. Correlations

There is no problem in correlating the Pont d'Issole and Ondres sections (Fig. 5). The vertical ranges of planktic foraminiferid index species supports the correlations between lithological units M1 to M3, the base of true Turonian deposits being defined either by the first occurrence of marginotruncanids or by the index species *H. helvetica*. The last occurrence of *R. cushmani* is at the base (Ondres) or the top (Pont d'Issole) of the upper, calcareous part of the M1 unit. This result suggests that bundles of beds like those occurring at the top of the M units could be synchronous over large distances within the basin with only minor changes. This is no real surprise as such bed-by-bed correlations over long distance (about 100 km) have already been made by several workers in the Vocontian Trough series at different stratigraphic levels, such as in the Valanginian (Cotillon et al., 1980), upper Jurassic and Aptian (Levert and Ferry, 1988).

The problem is the correlation of the Pont d'Issole and Vergons sections. According to the unpublished research of Morel (1998), the M1 unit should be totally absent at Vergons. So the first isotope excursion in this section was correlated to isotope event 2 at Pont d'Issole. However, the obviously truncated isotope curve at Vergons suggests that the bed bundle in the middle of the recessive interval cannot be the local equivalent of the bed bundle at the top of the M1 unit in the Pont d'Issole and Ondres sections. Because this bed bundle at Vergons is within the "recovery" interval of the isotope curve, it should be correlated to the M3 unit instead of the M2 unit of Pont d'Issole (Fig. 5). The top of the upper Cenomanian bundle of beds at Vergons does not show any particular surface, either omission or mechanical truncation, as there should have been in the correlations proposed by Morel (1998). It has been shown (Ferry and Flandrin, 1979) that in the limestone-marl alternations making up the Vocontian series, truncation surfaces due to large slides on palaeoslopes could be identified at outcrop because they are iron-encrusted by recent meteoric waters. No such surfaces are found at the equivalent level in the Vergons section. It

is thought most likely that the Vergons series has been truncated by a slide plane at the base of the disrupted shaly interval (Fig. 5). This disrupted interval is interpreted as being the sole of a large, thick syndepositional slide on the palaeoslope. The upper part of the rotational lens is not deformed except for the few metres at its base. Such large, gravity-induced rotational lenses are common features on basin margins. In the case of Vergons, the slide would have mixed at its base the shales of the M1 and M2 units, which would explain the occurrence of *R. cushmani* (Fig. 4) at a higher stratigraphic level (Fig. 5). There is no evidence of repeated reworking, such as bedded turbidites. There is only one deformed layer, interpreted as the result of a single event. If the above is correct, the first isotope excursion at Vergons is event 1, not event 2. The mid black shale bed bundle of Pont d'Issole is thus thought to top lap the slide surface slopeward (Fig. 5). So, whatever the way correlations are made, it is now clear that the Vergons section once used by Cru-mière et al. (1991) to define the Thomel level is incomplete and can no longer be used as a reference section for the Vocontian Trough.

5.2. Facies and thickness changes

If the above interpretation is correct, it is easier to understand that thicknesses of units M1 to M3 may regularly increase slopeward in spite of a slight decrease in the thickness of the Cenomanian/Turonian boundary recessive interval at Vergons. It is simply obscured by truncation at this location. The thickness of the M3 unit increases regularly from Ondres to Vergons (Fig. 5), as would the underlying M units, had the slump not been present. This is still visible in the increase in thickness of the calcareous top of the M1 unit, which comprises a single compounded bed at Ondres, but expands into a bundle of beds at Pont d'Issole.

Within this interpretation we can also see that black shale layers do have a regular spatial evolution from Ondres to Vergons. Their first appearance is approximately synchronous, but they last longer at Ondres (into the M3 unit) than at Vergons where they almost disappear at the top of unit M2. In detail, all interbeds of the upper part of unit M2 are black shales at Ondres, whereas there is only a thinly laminated horizon at Vergons within this interval (Fig. 5). These spatial changes are compatible with the palaeogeographic position of the sections (Fig. 1B). The black shale layers of Ondres appear to pinch out onto the slope and/or to change into regular bioturbated marls. Work in progress in slope sections further to the south supports this interpretation.

5.3. Biostratigraphy

Many samples were studied in the Vergons section in order to eliminate any bias and the last occurrence of *R. cushmani* is still a lot earlier at Vergons, at the base of M1 instead of the top of M2 as at Pont d'Issole (Fig. 5). This is supported by the fact that in both sections the last occurrences of *R. deeckeii*, *R. greenhornensis* and *R. cushmani* are in this normal stratigraphical order (Figs. 3, 4). In the Vergons section the

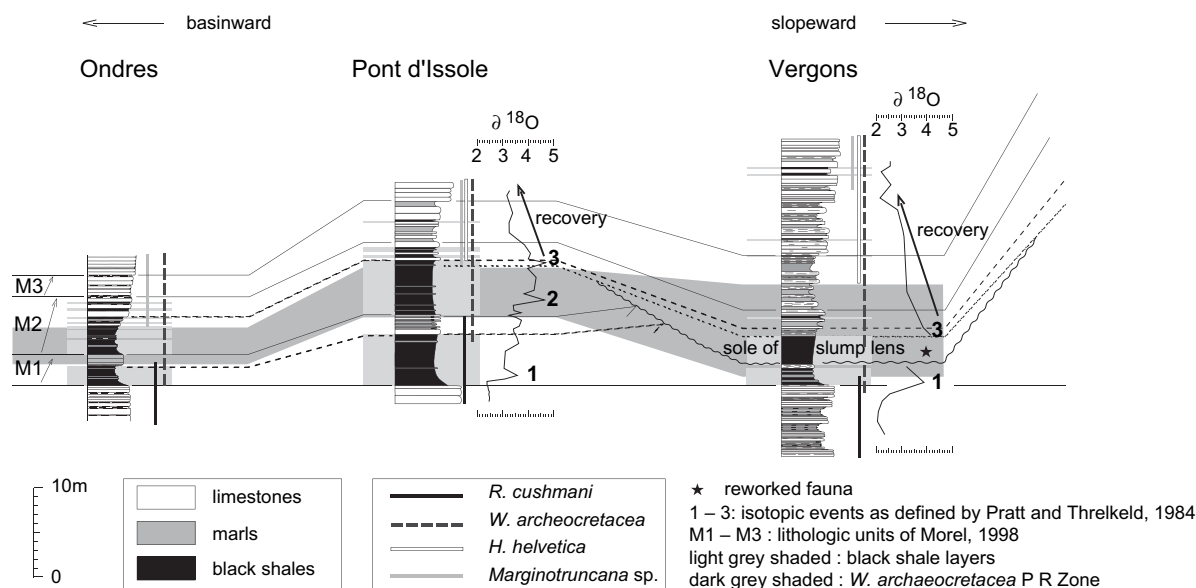


Fig. 5. Proposed correlations between the sections showing the changes in the stratigraphic range of the *W. archaeoeretacea* Zone.

occurrence of *R. deeckeri*, *R. greenhornensis* and *R. cushmani* within the disrupted black shale interval is due to the reworking of the M1 unit in which the three species occur nearby (Fig. 3). The base of the Turonian is better defined here by the first occurrence of marginotruncanids (Fig. 5). The first occurrence of the lower Turonian index species *H. helvetica* is strongly diachronous instead.

Hence, from available biostratigraphic and isotopic data, and if the above lithologic interpretation is correct, the definition of the *W. archaeoeretacea* Partial Range Zone appears to change from section to section (Fig. 5). It roughly follows the lower part of the M2 unit at Ondres and Pont d'Issole, but crosses lithologic correlations sloeward at Vergons where *R. cushmani* disappears below the sole of the slump, i.e., within the base of the M1 unit, as discussed above. We think it is unlikely that the bed bundles may be so diachronous over such short distances. We favour diminishing depositional depth, changing life conditions in base of slope environments, or some other cause(s) for explaining the expansion of the *W. archaeoeretacea* Zone to the south.

It has been shown that the duration of the *W. archaeoeretacea* Zone can be strongly reduced, almost to nothing in the US Western Interior (Desmares et al., 2003, 2004). We show here that its duration may change over short distances within the same basin.

The absolute duration of the *W. archaeoeretacea* Zone has been estimated to be between 700 and 200 ka, using cyclostratigraphy (Herber and Fischer, 1986; Robaszynski et al., 1993; Kuhnt et al., 1997; Morel, 1998; Caron et al., 1999; Prokoph et al., 2001). At Pont d'Issole and Ondres it covers only two thirds of the M2 unit (which is also a marl-limestone cycle). At Vergons, the truncation by the slump being taken into account as discussed above, it covers at least two cycles, from the upper half of M1 to the lower half of M3 (Fig. 5). As a result, should the M1–M3 units (or cycles) have the same duration, the time span of the *W. archaeoeretacea* Zone would be about twice as long at Vergons.

Our study has, therefore, yielded two major results. The first is the reduction of the *W. archaeoeretacea* Zone in the French subalpine basin, when compared to earlier works that were based only on the Vergons section (Crumière, 1990; Grosheny et al., 1992; Morel, 1998). The second is the strong variation in the amount of the time encompassed by this zone over short distances in a seemingly uniform depositional environment.

6. Conclusions

The Cenomanian/Turonian boundary black shales have been correlated over several sections at the base of the northern Provence platform slope in the French subalpine basin. The local $\delta^{13}C$ isotope anomaly shows three events (1, 2, 3 or A, B, C) defined in other basins by others workers. The Cenomanian/Turonian boundary spans roughly the duration of the black shale anomaly. The setting of black shale deposition is well defined and equates with profound changes in both benthic and planktic foraminiferid assemblages that may reflect changes in the oxygenation of the bottom waters. In detail, the boundary event is complicated, with intervening periods of re-oxygenation. The return to fully normal conditions of deposition is gradual. Correlation of bed bundles shows that black shale deposition ceases later in the deeper sections than in the Vergons section, which is situated at the base of a palaeoslope. Correlations also show that the Vergons section can no longer be used as a local reference for the Cenomanian/Turonian boundary because it is probably truncated by a large slump or syndepositional, rotational lens. The Pont d'Issole section is better as a reference, because it is both complete and more expanded than other distal sections situated further north in the basin.

Detailed study of planktic foraminiferids shows that the duration of the *W. archaeoeretacea* Partial Range Zone varies widely in closely spaced sections, either because the last

occurrence of *R. cushmani* or the first occurrences of lower Turonian forms (*H. helvetica* or marginotruncanids depending on the section) are highly variable, although the suspected slump truncation within the Vergons section remains a hypothesis, and its role in expanding the Partial Range Zone there may be questionable. If our correlations are correct, it is difficult to explain this discrepancy as being due to sampling bias, because the number of samples studied is high and all lithologies have been examined (limestone beds as well as shaly or marly interbeds). As with results obtained in the US Western Interior, we question the validity of this zone, which can be reduced locally to almost nothing. The role of anoxia in the disappearance of keeled forms is not clear. It might be better understood by studying sections (yet to be found) corresponding to depositional environments both deep enough to sustain the life cycle of keeled planktics and shallow enough to have escaped lower water column anoxia.

Finally we show that the local duration of the *W. archaeocretacea* Zone has to be reduced by half when compared to earlier results from this basin.

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Appendix

List of species and genera mentioned in text with author attributions and dates

- Anaticinella multiloculata* (Morrow, 1934)
Dicarinella algeriana (Caron, 1966)
Dicarinella canaliculata (Reuss, 1854)
Dicarinella hagni (Scheibnerova, 1962)
Dicarinella imbricata (Mornod, 1949)
Globigerinelloides ultramicra (Subbotina, 1949)
Hedbergella delrioensis (Carsey, 1926)
Hedbergella hoelzli (Hagn and Zeil, 1954)
Hedbergella planispira (Tappan, 1940)
Hedbergella simplex (Morrow, 1934)
Helvetoglobotruncana helvetica (Bolli, 1945)
Heterohelix globulosa (Ehrenberg, 1840)
Heterohelix moremani (Cushman, 1938)
Marginotruncana coronata (Bolli, 1945)
Marginotruncana pseudolinneiana (Pessagno, 1967)
Marginotruncana renzi (Gandolfi, 1942)
Marginotruncana schneegansi (Sigal, 1952)
Praeglobotruncana aumalensis (Sigal, 1952)
Praeglobotruncana elata (Lamolda, 1977)
Praeglobotruncana gibba (Klauss, 1940)
Praeglobotruncana stephani (Gandolfi, 1942)
Rotalipora cushmani (Morrow, 1934)
Rotalipora deeckeii (Franke, 1925)
Rotalipora greenhornensis (Morrow, 1934)
Schackoina cenomana (Schacko, 1897)
Whiteinella aprica (Loeblich and Tappan, 1961)
Whiteinella archaeocretacea (Pessagno, 1967)
Whiteinella baltica (Douglas and Rankin, 1969)
Whiteinella brittonensis (Loeblich and Tappan, 1961)
Whiteinella paradubia (Sigal, 1952)
Whiteinella praealvetica (Trujillo, 1960)