Cretaceous oceanic anoxic events (OAEs) recorded in the northern margin of Africa as possible oil and gas shale potential in Tunisia: An overview

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1. Introduction

Black shale deposition had been recorded frequently over large domains of the ocean floor throughout Mesozoic time (Schlanger and Jenkyns 1976; Schlanger et al. 1987; Jenkyns 1988; Wignall and Myers 1988; Arthur et al. 1990; Cecca et al. 1994; Wignall 1994; Kuhnt et al. 1997; Barrett 1998; Sageman et al. 1998; Leckie et al. 2002; Kuypers et al. 2002; Luning and Kolonic 2003; Lüning et al. 2004; Soua 2014a, 2014b, 2014c). These fine-grained, organic-rich strata deposited during severe palaeoecological conditions and oxygen-deficient bottom waters are interpreted to be deposited under suboxic to anoxic as well as strongly euxinic conditions (Algeo 2004; Tsikos et al. 2004; Kolonic et al. 2005; Baudin 2005; Ben Fadhel et al. 2011; Soua et al. 2011a; Elkhazri et al. 2013; Soua 2014c).

Typical features of these black shales are: (1) laminated; (2) pyrite-enriched; (3) type II marine organic matter with TOCs ranging between 1% and 20% and HIs ranging between 350 and 850 mg Hc/gTOC; and (4) enriched in trace metals (Algeo and Lyons 2006; Bodin et al. 2006; Scopelliti et al. 2006; Turgeon and Brumsack 2006; McArthur et al. 2008; Soua 2010b; Soua et al. 2011a; Tribovillard et al. 2012; Soua 2013a).

During the past 40 years, thick mid-Cretaceous black shale (Valanginian–Turonian, ca. 140–90 Ma) sections in the Atlantic and Pacific Ocean domains have been discovered through the Deep Sea Drilling Project (DSDP) (Schlanger and Jenkyns 1976) and recently through the Ocean Drilling Program (ODP) (e.g. Baudin 2005; Scopelliti et al. 2006; Tribovillard et al. 2012; Soua and Chihi 2014). These discoveries have been followed by identification of similar deposits through the northwestern Tethyan margin (Figure 1). This has opened up new exploration challenges regarding the deposition of black shale levels and has enhanced the knowledge of their mechanisms and deposition within both the oil industry and academic institutions.

These organic-rich sediments have large economic significance as they include more than 90% of global recoverable hydrocarbon reserves (oil and gas) (e.g. Kolonic 2004; Sorkhabi 2009). In fact, during the Mesozoic more than three major stratigraphic organic-rich intervals have been recorded: (1) the Jurassic (Toarcian and Callovian), ca. 183 and 165 Ma (25% of reserves); (2) the late Cretaceous (Turonian, ca. 140–90 Ma) sections in the Atlantic and Pacific Ocean domains have been discovered through the Deep Sea Drilling Project (DSDP) (Schlanger and Jenkyns 1976) and recently through the Ocean Drilling Program (ODP) (e.g. Baudin 2005; Scopelliti et al. 2006; Scopelliti et al. 2006; Turgeon and Brumsack 2006; McArthur et al. 2008; Soua 2010b; Soua et al. 2011a; Tribovillard et al. 2012; Soua 2013a).
The main Cretaceous Oceanic Anoxic Events (OAEs) discussed in the text and identified worldwide, as well as in Tunisia – a tentative correlation with the events that occurred in Tunisia during the Cretaceous. Timescale, global events, and carbon isotopic data are compiled from Jenkyns (2010) and Martinez (2013).

(Triassic, ca. 229 Ma), the Coniacian–Santonian, ca. 86 Ma (ca. 2% of reserves), and, of less importance, the Valanginian–Barremian period. All these periods are reported as times of oceanic anoxic events (Schlanger and Jenkyns 1976; Pedersen and Calvert 1990; Wignall 1991; Calvert et al. 1996; Nijenhuis et al. 1999; Lüning et al. 2004; Brumsack 2006; Soua and Tribovillard 2007; Jenkyns 2010; Soua 2014a, 2014b, 2014c).

This review paper focuses on the mid-Cretaceous period in Tunisia (Figure 2). For several reasons, this period, especially in several parts of northern Tunisia, is considered as the era which has borne the highest black shale levels. In this contribution, we aim to review (1) the mapping of North African palaeogeographical reconstructions; (2) the existence of geochemical analyses including chemostratigraphic, carbon isotopic, and biostratigraphic (foraminifera and radiolarian) data; and (3) the existence of complete sedimentary records showing laminated lithology and which can demonstrate Milankovitch cyclic characters providing good time control on sedimentation. These data may serve as a correlatable tool for the northern Tethyan domain. A final aim, which can be achieved with this contribution, is generally to define potential Cretaceous oil and gas shale plays in Tunisia.

2. Overview of the geological setting

2.1. Cretaceous North African geological setting for OAEs

The deposition of Cretaceous organic-rich sediments in the northern margin of Africa, namely in Tunisia and Algeria, is strongly controlled by half-graben systems and in some areas by Triassic diapiric extrusion movements, which are related to the global palaeogeographical evolution during the Cretaceous (Figure 3).

The rift-grabens formation is strongly related to the Triassic palaeogeography of Gondwana. In fact, the Triassic period is characterized by several global tectonic events which affected the break-up of the Pangea supercontinent (Ziegler 1988; Veevers 1994, 2004; Withjack et al. 1998; Golonka and Ford 2000; Golonka 2002; Soua 2014a). Generally, within this time interval, the Cimmerian plates drifted northward from Gondwana toward Laurasia, which began in reality since the Carboniferous–earliest Permian, and this is the main cause of the oceanic crust closure of the Palaeotethys (Sengör 1984; Zonenshain et al. 1990; Sengör and Natalin 1996; Golonka 2004; Soua 2014a), and consequently this movement affected the opening of the Neotethys Ocean domain (Burrollet et al. 1978; Jongsma et al. 1985; Turki 1985; Morgan et al. 1998; Guiraud 1998; Touati and Rodgers 1998; Grasso et al. 1999; Soussi 2000; Soua and Tribovillard 2007). These events are supposed to be responsible of the closure of the Palaeotethys and the collision of the drifting Cimmerian Plates with the Chinese blocks (Indosinian orogeny, Golonka 2007).

Alternatively, in Morocco, during the Cretaceous, related synrift sedimentation consisted mainly of deltaic clastics where syndepositional rollover structures are developed, which might also have locally affected the deposition of anoxic intervals over the time interval (e.g. Lüning et al. 2004).
Figure 2. Organic-rich black shale distribution during (a) Early Cretaceous (Valanginian, late Hauterivian, and early Aptian), (b) Albian, and (c) late Cenomanian–early Turonian in North Africa. Data were compiled from Lüning et al. (2004); Zghal and Arnaud-Vanneau (2005); Baudin (2005); Reichelt (2005); Bodin et al. (2006); Chihaoui (2008); Jenkyns (2010); Soua et al. (2011b); Ben Fadhel et al. (2011); Tribovillard et al. (2012).
Figure 3. Palaeogeographic reconstruction map of the western Tethyan domain during the Early Cretaceous (Valanginian, late Hauterivian, early Aptian) and Cenomanian–Turonian transition. Reconstructions were made after Barrier and Vrielynck (2008) and Tribovillard et al. (2012).
Generally, two main Mesozoic rifting events occurred in this northern African margin, namely: (1) the Late Jurassic to early Aptian, associated with E–W-trending half-grabens and volcanism (e.g. in the Pelagian block, North–South axis; Laaridhi-Ouazaa 1994; Lazeez 2004; Mattoussi Kort et al. 2008), which is related essentially to the Neotethyan opening; and (2) from the post-Aptian to early Cenomanian, associated with NW–SE-trending half-grabens related generally to the Sirte rifting phase (Soua et al. 2009).

In northeastern Algeria, this extensional regime also originated from the formation of graben and half-graben systems, which controlled the distribution of carbonate platform and deeper basin facies, with organic-rich sediments being deposited several times during the Cretaceous (Nacer Bey et al. 1995; Nalil et al. 1995; Herkat 1999; Lüning et al. 2004; Chaabane and Salmi-Laouar 2014).

These two recorded extensional phases were interrupted by a supposed regional top-Aptian compressional phase (the Austrian phase), which is responsible for an unconformity detected on seismic profiles and in outcrops (e.g. Ben Ferjani et al. 1990; Grasso et al. 1999). The general graben trends are locally complicated by other faults, which partly dissected the grabens into smaller segments (Ben Ayed and Viguerie 1981; Ben Ayed 1986; Chihi 1995).

This top-Aptian compressional phase also affected the Saharan Platform (e.g. Berkine Basin) and probably resulted from transpressional motion along the N–S-trending Trans-Saharan fracture system (e.g. Yahi 2001).

Commonly, around the Turonian period, a general change from extension to compression direction occurred in North Africa, which was related to the closing of the Neotethys and the onset of North Atlantic rifting (Burollet et al. 1978; Guiraud and Maurin 1991; Grasso et al. 1999; Lüning et al. 2004; Soua et al. 2009). This resulted in the inversion of former rift grabens.

In Algeria, the Mid-Cretaceous extension was most intense in some localities (e.g. Constantine Basin; Herkat and Delfaud 2000), influenced by two principal extensional systems, the NE–SW-trending Aurès–Keftrough to the north and the NW–SE-trending Negrine–Gafsa trough to the south. The former was probably inherited from ‘Tethyan’ rifting, while the latter may correspond to the direction of the Sirte rift system (Soua et al. 2009).

Extensional infra-Cretaceous tectonics were used also to explain half-graben formation (Ensslin 1992; Charrière 1996) segmented into mini-tectonic blocks (Lüning et al. 2004). These latter authors assume that the rift movements were somehow blocked or restrained several times through the Cretaceous, when the anoxic deposition of organic-rich strata occurred.

In northwestern and central Tunisia, as well as in the Pelagian Province, Cretaceous diapiric movements in the Triassic played a major role locally in controlling organic-rich deposition (Perthuisot 1981; Souquet et al. 1997; Vila et al. 1998; Lüning et al. 2004; Soua and Tribovillard 2007). These are characterized by a marked reduction in thickness and partly by the development of detrital horizons (microbreccia) such as in the area of Bargou (Soua 2013b). Complex diapiric vertical extrusions, possibly starting from the Aptian (and even before) through approximately the middle Eocene, were probably continuous but would have increased during periods of tectonic instability. Therefore, earlier diapiric movements and rise-up are superimposed on the extensional features enhancing depocentre individualization in the central parts of rim-synclines.

### 2.2. Sea-level change recording

The Cretaceous OAEs are generally associated with key eustatic transgressions, which are thought to be the most intense flooding events (Haq et al. 1987; Thuroe et al. 1992; Gale et al. 2002; Lüning et al. 2004; Soua and Tribovillard 2007; Karakitsios et al. 2010; Tribovillard et al. 2012). These transgression events were mostly related to specific transgressive surfaces (TSs), which are delineated in the most important eustatic sea-level charts (e.g. Haq et al. 1987; Hardenbol et al. 1998; Haq and Schutter 2008; Snedden and Liu 2010). The maximum flooding surfaces (MFSs) of the third-order cycle are notably diachronous. In fact, the boundaries of the transgressive systems tract (TST) related to these OAEs appears to have a different timing in the Tethyan domain compared with the eustatic sea-level curve of Haq et al. (1987); reported also in Hardenbol et al. (1998).

### 3. Materials and Methods

This study is mainly focused on the assessment and review of the Cretaceous OAEs, from a geological viewpoint, that could constitute extensive oil/gas shale plays in Tunisia (i.e. their distribution, facies, thickness, organic richness, maturity, etc.). The study concerns the northern and central part of the country (Figure 4). Published papers and unpublished petroleum reports are used in this review to evaluate geological and biostratigraphical investigations. In addition, some new data have been used, such as wireline log data (e.g. spectral gamma rays (SGR), uranium, thorium, potassium), from more than 20 wells in Tunisia to estimate organic matter as well as primary productivity and palaeoredox...
conditions. In addition, pyrolysis data and depth intervals and thickness of black shale beds (Valanginian, early Aptian, Albian, and Cenomanian–Turonian) have been used in compilation of the distribution, isopach, maturity, and isobath maps of these units in each sedimentary basin. The positions of wells and some seismic lines helped in generation of large-scale cross sections between the basins (Figure 4).

The biostratigraphic review framework was carried out mainly through palynological analysis investigated earlier by petroleum companies (internal reports) operating in different basins (SEREPT, ENI, Shell, Marathon, Larsen Oil & Gas) as well as by Palaeoservices Ltd for the Lower Cretaceous of the Gulf of Gabes. An ammonite biostratigraphic scheme for the Lower Cretaceous was previously used by Busnardo and Memmi (1972) and
Memmi (1989). Foraminiferal and radiolarian biostratigraphic investigations have also dealt with black shale levels corresponding to the reviewed OAEs, as well as calpionellids and belemnites of the Lower Cretaceous sections (Memmi 1989; Maamouri et al. 1994a, 1994b; Souquet et al. 1997; Rami 1998; Soua et al. 2006; Soua and Tribovillard 2007; Ben Fadhel et al. 2010, 2012; Soua 2011, 2013b; Zaghibb-Turki and Soua 2013). The regional distribution maps of these organic-rich levels were carried out using various well data (depth, thickness, and gamma ray response). Geochemical investigations (total organic carbon (TOC), maturity, molecular geochemistry) for the Cretaceous black shales have recently been published (Montacer et al. 1988; Layeb 1990; Brehm and Trichelli 1991; Saidi 1993; Talbi 1993; Bechtel et al. 1998; Barrett 1998; Lüning et al. 2004; Soua and Tribovillard 2007; Elkhazri et al. 2009), while unpublished TOC petroleum analysis reports were also consulted.

3.1. Wireline SGRs

Well logging tools are generally applied as proxies for many purposes such as mineralogic composition, TOC, and major and trace elemental geochemistry to evaluate productivity and palaeoredox conditions during such anoxic events. Their main parameters can be evaluated from such logging tools.

High total gamma ray (GR) readings are generally an indicator of organic-rich shales, especially within such OAE levels (black shales). The presence of high levels of uranium (U) coeval with low concentrations of potassium (K) and thorium (Th) is suggestive of preserved organic-rich reducing environments. The presence of almost equal K, Th, and U concentrations can suggest that the contribution of K and Th is associated with the clay fraction of the shale, while U is associated with authigenic uranium precipitation under redox conditions. In addition, elevated K and Th concentrations occurring in combination with low U content generally indicate organically leaner shale. High GR readings are generally due to uranium enrichment of black shale, while in the Valanginian–Hauterivian time interval, high thorium concentrations may prevail. Generally, under reducing conditions soluble U$^{4+}$ is converted to insoluble U$^{4+}$ (e.g. Wignall and Myers 1988; Lüning et al. 2005; Tribovillard et al. 2006; Soua 2010b; Soua et al., 2011b). Basically, uranium is removed from seawater and pores into organic-rich sediments (Anderson et al. 1989; Barnes and Cochran 1990; Klinkhammer and Palmer 1991). According to some authors (e.g. Algeo and Maynard 2004; McManus et al. 2005; Tribovillard et al. 2006; Soua et al. 2011b and references therein), U enrichment does not occur in the water column but rather takes place mainly in the sediment, while Zhang et al. (2002) hypothesized that U may occur by the combination of iron and sulphate reduction. The oxidized soluble form of uranium is reduced and fixed in euxinic basin sediments as tetravalent U (Anderson et al. 1989; Arthur and Sageman 1994; Lüning et al. 2000a; Tribovillard et al. 2006; Soua 2012, 2014c).

Several authors (e.g. Zelt 1985; Wignall and Myers 1988; Doveton 1991; Arthur and Sageman 1994; Luning and Kolonic 2003; Soua 2014c) have assumed that in euxinic environments, the U/Th ratio is sensitive to the relative inputs of Th-bearing clay minerals. However, detailed studies by Hurst (1999, and references therein) proved that Th is exclusively concentrated in heavy minerals, which may occur in the clay grade fraction. In addition, Th has a very stable 4$^+$ state that is almost exclusive to heavy minerals such as monazite, apatite, and, in particular, zircon. Confusion has arisen from the fact that Th occurs in close association with clay minerals, but detailed microprobe and SEM studies have shown that it only occurs in heavy minerals possibly of clay–silt grade (see Hurst 1999). In addition, the U/Th ratio, which is considered as a reliable redox index (Riquier et al. 2006; Soua et al., 2011b; Hatch and Leventhal 1992; Jones and Manning 1994), may contradict value distribution between levels of black shale and adjacent sides.

3.2. Time-series analysis

Time-series analysis was performed on the Bahloul Formation (OAE2) by spectral analysis on two opportunistic foraminiferal species using the Blackman and Tukey Method (BTM) (for detailed methodology, see Soua 2010b). This cyclostratigraphic investigation was carried out in order to determine the duration of OAE2 and its related sedimentation rate recorded in the southern Tethyan margin (Tunisia).

3.3. Major and trace elements

Major element concentrations were determined using an inductively coupled plasma atomic emission spectroscopy (ICP-AES) process, and trace elements were determined using inductively coupled plasma mass spectroscopy (ICP-MS) at the University of Lille1 (France) and GeoPack Ltd (Tunisia), both for the Cenomanian–Turonian OAE2 (Soua 2010b, 2011; Soua et al. 2011b). In addition, complete sample preparation and methods have been published by Soua et al. (2011b) and are not presented here since these fall outside the scope of this study.
3.4. Integrated prediction error filter analysis

An integrated prediction error filter analysis (INPEFA) curve was conducted on digital data (LAS files) using CycloLog software, as described by Nio et al. (2005) (see Section 4.2.2). Nio et al. (2005) described the use of an error filter in the linear prediction of the log shape to evaluate the spectral contents. The log attribute is known as the prediction error filter analysis (PEFA) curve (Veeken 2006). This visualizes breaks in continuity of the spectral representation of a log. PEFA can easily be transformed into an INPEFA curve that is more stable and easier to interpret geologically. The log attribute is useful in high-resolution sequence stratigraphy, where climatic cycles are detected in the sedimentary record. As described by Nio et al. (2005), a cumulative negative set of prediction error values may result in a negative INPEFA trend. Negative prediction error values imply overestimation of the SGR values recorded by the filter. Therefore, this represents a segment of the data in which the actual values of the log are less than predicted. Generally speaking, a negative INPEFA trend is related to a regressive trend although its exact significance will be dependent on the geological context. In contrast, a positive trend in the INPEFA of SGR logs may signify a transgressive trend.

3.5. TOC modelling using the Schmoker equation

Generally, in the wells used in this contribution, TOC analysis is not always available. The Schmoker equation (Equation (1)) has been used successfully to provide a very basic estimate of TOC. It was defined and tested by Schmoker and Hester (1983) in the Mississippian and Devonian Bakken Formation black shale. In general, it is calculated as follows:

\[
TOC \text{ (wt\%) = } \left( \frac{154.497}{p} \right) - 57.261, \text{ where } p \text{ is the density (g cm}^{-3})
\]

(1)

4. Oceanic anoxic events during the Cretaceous and organic-rich deposits in Tunisia

Several rifting pulses occurred on the northern margin of Africa, separated in both time and space mainly from the Triassic through the Cretaceous. During oxygen-minimum zone (OMZ) expansion periods, black shale accumulation was recorded through the Mesozoic and notably through the northern margin of Africa. In contrast, paralic to shallow-marine sedimentation occurred along the southern Tunisian margin (e.g. Jeffara Basin, Saharan Platform, Tataouine Basin (Figure 4). Oil-prone and organic-rich strata were deposited in the initial formation of marine sediments of narrow rifts and in parts of the subsiding margin.

Therefore, tectonics played an important role in controlling the distribution of organic matter accumulated on the irregular outer shelf (Lüning et al. 2004), segmented by horsts and grabens, which modelled the central Tunisian domain during these times.

Notably, eight main anoxic events occurred in Tunisia throughout Mesozoic time, which were outlined by black shale layer accumulations generally expressing eccentricity and precession Milankovitch-like cyclicity and formed during third-order transgressive systems tracts. Their main distribution occurred in: (1) the Carnian, (2) the Toarcian, (3) the Callovian, (4) the late Hauterivian, (5) the Bedoulian, (6) the Alban, (7) the Cenomanian–Turonian transition, and (8) with reduced importance in the Coniacian–Santonian transition (Soussi et al. 1992, 1998; M’Rabet et al. 1995; Arfaoui and Montacer 2004; Soua and Trubovillard 2007; Elkhazri et al. 2009; Soua et al. 2011b; Ben Fadhel et al. 2012; Ouadday and Saddem 2012; Soua 2013a; Soua and Chihi 2014; Soua 2014b).

This paper deals with an overview of published Cretaceous (i.e. Valanginian, late Hauterivian, Bedoulian, Albanian, and Cenomanian/Turonian (C/T) anoxic events from Tunisian outcrops and petroleum boreholes using several techniques described in Section 3.

4.1. Valanginian Weissert anoxic event

4.1.1. Distribution and characteristics of the organic-rich Valanginian strata

During the Valanginian (and most of the Early Cretaceous), Tunisia was characterized by an inherited N-dipping slope from the Late Jurassic at the southern Tethyan margin (Soua 2014b), including a variety of facies transects ranging from siliciclastics and locally bioclastic dolomite (Melloussi in central Tunisia), massive carbonate (Douleb-101 in west-central Tunisia) over transitional facies (mixed sandstone, dolomite, and shale), to bathyal (Hamada in northern over east-central Tunisia) (Figure 5). It is noteworthy, as mentioned in Figure 5, that the Berrissien to Hauterivian section is missing over the southern part of the country (Ouaja 2003). Consequently, no equivalent strata have been deposited through this area. In central Tunisia, Trabelsi (1996) reported that the Early Cretaceous sedimentation in the Sidi Aich–Majoura area, limited to the east by the N–S axis, is controlled by horsts and rhomb-graben structures. This configuration is characterized by the reactivation of flower faults in transtensive movements.
The general facies trend was complicated by a complex tectonic palaeorelief configuration, which resulted in dislocation into tilted blocks, individualizing deep half-graben systems toward the north, the northwest, and the west since the late Sinumerian–Carixian interval (Turki 1985; Soussi et al. 1992; Kamoun et al. 1999; Ben Youssef 1999; Bouaziz et al. 2002; Beall et al. 2002; Soussi 2003; Soua 2014b). This configuration resulted in strong lateral thickness and facies variations of the Valanginian strata (Figure 5).

Organic-rich Valanginian strata occur in many places in northeastern and east-central Tunisia (e.g. near the N–S axis, NA-101 well, Figure 5) and are grouped into the Hamada Formation (Figure 6) defined previously by Khessibi (1967), emended later by Peybernès et al. (1994), and grouped earlier in ‘Argiles du Sidi Khelif’...
The distribution of the Hamada Formation in Tunisia is illustrated in Figure 5 (compiled after Castany 1951; Bonnefous and Rakus 1965; Bonnefous 1972; Busnardo and Memmi 1972; Combémorel et al. 1981; M'rabet 1981; Memmi 1989; Ben Ferjani et al. 1990; Soua 2007; Sandman 2008; and our field measurements). The regional distribution of the main Valanginian facies types is also shown in Figure 5, which also includes the distribution of probable organic-rich and -poor Hamada facies trends. The organic-rich Valanginian Hamada exists in four areas, namely the northeast (Jebel Boukornine, Jebel Oust, Oued Guelta in Jebel Zaghouan, and Jebel Beni Kleb sections, as well as in the CB-101 and BAR-1 wells); in the north (Jebel Ammar, Jedeida, and Borj Toumi sections); in the offshore Gulf of Tunis (Raja-1 well); and in central Tunisia (near the N–S axis, NA-101 well). Unfortunately, several neighbouring drilled wells did not reach the Early Cretaceous strata, and therefore much information about the Valanginian anoxic event remains unknown. Rare available well and outcrop data confirm this organic-rich province, although the exact boundaries still remain unclear; especially in the Gulf of Tunis (Figure 4), central Tunisia/Algeria, and the Gulf of Hammamet (Pelagian Sea) due to the non-availability of well data. Geophysical evidence of the Early Cretaceous black shale distribution is mentioned in Figure 7, which

Figure 6. Type stratigraphic section of the Hamada and Seroula Formations logged in Jebel Oust. Lithology, biostratigraphy, sequence stratigraphy, and age constraint are adapted from Khessibi (1967), Maamouri et al. (1994a), Memmi (1989), Souquet et al. (1997), and Martinez (2013).

Figure 7. Seismic line passing near the N–S axis (NOSA) between Jebel Mghilla and Jebel Cherichira (see Figure 4 for location) showing the existence of several half-graben systems to the west of the NOSA and the possibility of the preservation of organic matter and recording of the Faraoni Anoxic Event in this area (modified after Doglioni et al. 1990). (a) Raw seismic line; (b) interpreted seismic line.
shows structuration of half-graben systems and bordering normal faults. This seismic line is located westward, in the southern part of the N–S axis (see Figure 4 for the location of the seismic line and generated cross sections).

A comparison to the palaeogeographic map (Figure 5) suggests that the distribution of the organic-rich Hamada is restricted to intermediate to medium water depths, and may represent the impingement of an OMZ into the northern margin of Africa. The Hamada Formation is generally hundreds of metres thick (80–150 m) in the northeast and the N–S axis areas, with a maximum thickness of about 200 m in the Gulf of Tunis (Soua 2007; Sandman 2008). In Jebel Boukornine (Figure 5), the upper part of the ‘Pelagian Facies’ (= Hamada) Formation (Combémorel et al. 1981) is also represented by basinal black shales interbedded with finely laminated limestones and dark marls (15 to >90 m). Generally, the Hamada Formation consists of a regular alternation of dark-coloured shale and laminated carbonates incorporated within grey to dark grey marls, with mean TOC values of 0.6% and with a single value of up to 1.9% (Raja-1 well) (Figure 5). Unfortunately, no additional organic carbon analyses have been carried out. SGR analysis is available from the BAR-1 well (Figure 8), which provides levels of redox-sensitive uranium, thorium, and potassium, and thereby the redox index U/Th ratio.

Levels of U vary significantly in the BAR-1 depending on the underlying and overlying Hamada Formation beds. The uranium concentration shows slight variation,

![Figure 8](https://example.com/image.png)

**Figure 8.** Spectral gamma rays (SGRs), which provide uranium, potassium (K), and thorium (Th) levels, redox-sensitive U/Th ratio, Th/K ratio, and sequence stratigraphy of the BAR-1 well (see Figure 4 for location). Computed gamma rays are the sum of Th and K.
from 1.51 to 2.76 ppm, in the Hamada Formation (Figure 9, Table 1), which may be considered as a low concentration, but coinciding with the highest values recorded for the U/Th ratio, which ranges between 0.3 and 0.8 ppm.

The average value calculated for the shale U/Th ratio is about 0.225 (Turekian and Wedepohl 1961), used here as a redox index (Riquier et al. 2006; Jones and Manning 1994; Soua et al. 2011b; Table 2). According to Jones and Manning (1994), the U/Th ratio proxy can be subdivided into three intervals. Values lower than 0.75 can be explained by deposition under oxic conditions while sediments with values higher than 1.25 could in turn result from deposition in an anoxic environment (Table 2).

Figure 9 illustrates the trends of the U/Th ratio through the Hamada Formation in the BAR-1 well. Significant increase in the redox-sensitive proxy is observed in the interval 2664–2672 m, where values exceed 0.75 (see Figure 9).

The typically observed ranges of U/Th ratio observed include a pattern with a sine-shape-like sudden increase.

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**Table 1.** Logging characteristics of selected OAEs (Valanginian Weissert, late Hauterivian Faraoni, and early Aptian Selli events).

| Well   | Age       | Value   | SGR   | CGR   | U     | Th    | K     | Density | Schmoker |
|--------|-----------|---------|-------|-------|-------|-------|-------|---------|----------|
| BAR-1  | Valangian | Maximum | 80.15 | 59.51 | 2.76  | 6.56  | 2.15  | 0.78    | 4.63     |
|        |           | Minimum | 40.96 | 28.87 | 1.51  | 2.73  | 1.03  | 0.35    | 1.60     |
|        |           | Average | 55.27 | 39.58 | 1.96  | 4.20  | 1.44  | 0.57    | 2.76     |
|        | Late Hauterivian | Maximum | 122.84 | 120.59 | 3.70  | 18.66 | 4.06  | 3.93    | 4.63     |
|        |           | Minimum | 15.72 | 12.41 | 1.09  | 0.55  | 0.35  | 1.60    | –        |
|        |           | Average | 72.51 | 68.51 | 0.89  | 9.45  | 2.14  | 2.76    | 4.30     |
| EZZ-8  | Late Hauterivian | Maximum | 66.04 | 53.81 | 2.75  | 7.42  | 0.02  | 8.45    | –        |
|        |           | Minimum | 10.67 | 2.87  | 0.50  | −10.00 | 0.00  | 0.40    | –        |
|        |           | Average | 33.51 | 24.18 | 1.40  | 3.06  | 0.01  | 4.43    | –        |
| BAR-1  | Early Aptian | Maximum | 89.08 | 84.13 | 2.76  | 1.14  | 2.74  | 1.41    | 4.62     |
|        |           | Minimum | 23.81 | 18.00 | 3.00  | 76.00 | 66.00 | 0.50    | 1.59     |
|        |           | Average | 47.16 | 40.53 | 1.11  | 5.13  | 1.35  | 0.96    | 2.92     |

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**Table 2.** Redox classification of depositional environments using O2 and U/Th, after Tyson and Pearson (1991), Jones and Manning (1994), and Tribovillard et al. (2006).

| Redox classes  | Oxic | Dysoxic–suboxic | Anoxic | Euxinic |
|----------------|------|-----------------|--------|--------|
| O2 concentration in bottom waters (ml O2/L H2O) | O2 > 2 | 2 > O2 > 0.2 | O2 < 0.2 (a) | O2 = 0 (b) |
| U/Th ratio (c) | U/Th < 0.75 | 0.75 > U/Th > 1.25 | U/Th > 1.25 | – |

Notes: (a) No free H2S in the water column; (b) free H2S present in the water column; (c) Jones and Manning (1994).
in values within the Valanginian upper Hamada interval in the BAR-1 well (Figures 8 and 9).

Using the Schmoker equation, which is based on the density parameter, Figure 9 shows that the estimated TOC confirms an increase in organic richness which may indicate widespread bottom redox conditions in the Valanginian upper Hamada Formation.

In the Gulf of Tunis, well Raja-1 (Supplementary Figure 1; see Figure 5 for location; supplementary figures may be found at http://dx.doi.org/10.1080/00206814.2015.1065516) penetrated a similar stratigraphic section, where it reached 2871 m in depth within the early Valanginian Hamada Formation. This section comprises black shale levels incorporated within grey to dark shale and limestone succession alternating with a few thin turbiditic levels that characterize mostly the overlying Seroula Formation. Similar interpretation of the environment deposition has been described from the type section (Figure 6) located in Jebel Oust (Souquet et al. 1997). Discussion on the relationship between the turbiditic current and black shale accumulation can be raised. This infers the periodicity of the organic matter supply, OMZ expansion, and the widespread nature of redox conditions at that time in northeastern Tunisia.

Average TOC recorded within this level was around 1.28% (Soua 2007), confirming a peak in reducing conditions recorded in the Tethyan realm (Weissert et al. 1998; Price 1999; Melinte and Mutterlose 2001; Erba et al. 2004; Gröcke et al. 2005; Duchamp-Alphonse et al. 2007; McArthur et al. 2008; Gréselle et al. 2011; Martinez 2013).

Within the Cap Bon peninsula, wells CB-1 and CB-101 also penetrated organic-rich black shale levels, which are incorporated within massive shaly succession. The overlying alternation between thin to medium turbidite, shale, and black shale successions (Seroula Formation) produces gas.

The low TOC values characterizing some Valanginian sections (e.g. Raja-1 well) may be explained by the fact that the well did not reach the basal part of the Valanginian black shale interval (typically organically rich in BAR-1; Figure 9), or could be linked to an organically leaner location or even probably deposited underneath the OMZ (Barrett 1998; Lüning et al. 2004; Soua and Trinovillard 2007).

4.1.2. Biostratigraphy and sequence stratigraphy

The upper Hamada Formation was deposited during the Valanginian Anoxic Event (Petransiens Zone), which may have continued through the lower Seroula Formation (Khessibi 1967) in the Verrucosum Zone (Memmi 1989) (Figure 6). This probably reflects the significant input of local upwelling conditions during black shale deposition (e.g. in Jebel Oust). Khessibi (1967) originally proposed a Middle to Upper Jurassic age constraint for Hamada black shales in the Jebel Maiana in Jedeida area, subsequently confirmed by Alouani and Tlig (1991), Alouani (1991), and Chandoul et al. (1993), without recording any biostratigraphic data. This stratigraphic interpretation was later reevaluated by Peybernès et al. (1994) who, by means of biostratigraphy (foraminifera), demonstrated that the Hamada and Seroula Formations in this section span the Valanginian (Figure 6). Macoin (1963), Pini et al. (1971), and Bonnefous (1972) previously provided an Early Cretaceous (Valanginian to Hauterivian) age for the Seroula Formation.

In general, the early Valanginian of the Dorsale area (Jebel Boukornine, Jebel Beni Kleb) is marked by alternating marl, limestone, and laminated black shale beds containing some radiolarians, belemnites, rare calpionellids, scarce benthi foraminifera, and planktonic foraminifera (Busnardo and Memmi 1972; Memmi 1989; Maamouri et al. 1994a).

In western Jebel Zaghouan (Oued Guelta), Combémorel et al. (1981) described these black shales as containing diversified ammonites such as Neocomites spp, Chamalocia aestigmatica, Kiliannella roubaudiana, and Ptychophylloceras semisulcatum, and attributed these to the early Valanginian. So, when referring to these authors, the age of these black shales is now well documented as early Valanginian. Their distribution and extension was revealed to be much more important than what was described previously by Ben Ferjani et al. (1990), since they had already been characterized in the west of the Dorsale (Memmi 1969; Souquet et al. 1997) and more recently in the Jedeida area (Peybernès et al. 1994).

In Jebel Oust, these black shales are very well documented since they pertain to the type section for the Early Cretaceous (Busnardo and Memmi 1972; Combémorel et al. 1981; Souquet et al. 1997; Memmi 1999; Figure 6). They are reported as early Valanginian-aged black shales spanning the Petransiens Zone, which yielded diversified belemnite and ammonite fauna composed generally of Duvalia latalata, Duvalia latazengitana, Pseudobelus bipartitus, Neocmites sp., Calliphylloceras sp., Phylloceras tetys, Eulytoceras aff. subfimbriatum, Bochianites sp., Neolissoceras grasia-num, Olcostephanus astieri, Rogensites atherstoni, Neocmites aff. Paraplesius, and Spiliticus usincertus. The upper part of the Valanginian black shale beds provides Lytoceras sp., Neocomites neocomiens, N. teschensis, N. callidicus, and Oosterella astieri (Memmi 1969; Busnardo and Memmi 1972; Maamouri et al. 1994a). Similar ammonite and belemnite assemblages
have been described recently by Főzy et al. (2010) in the Hungarian Transdanubian range.

The onset Hamada Formation deposition in northeastern Tunisia generally coincides with the base of the early Valanginian Transgressive Systems Tract (TST) (Haq et al. 1987; Vail 1988; Souquet et al. 1997; Hardenbol et al. 1998), corresponding to a major third-order Tethyan eustatic sea level.

A similar sequence of stratigraphic interpretation was presented by Souquet et al. (1997). Particularly in the Jebel Oust section, the sequence boundary (SB) is confused with the transgressive surface (TS) due to the generalized unconformities during the Early Cretaceous (Souquet et al. 1997).

To correlate the TST we used the SGR log of the Raja-1 and BAR-1 wells, respectively (Figures 9 and 10).

The third-order Highstand Systems Tract (HST) overlying the early Valanginian TST in northeastern Tunisia is generally represented by shallower deposits, such as the carbonates, turbidites, and marly limestones of the Seroula Formation and equivalent units (e.g. Khessibi 1967; Peybernès et al. 1994; Souquet et al. 1997). In the area including Jebel Mecella, Jebel Ressas, Hamem Zriba, and Jebel Mdeker, it is important to note that no equivalent Valanginian strata have been deposited since this area was dominated by a reefal complex during Upper Jurassic time (Ressas Formation) (Peybernès 1991). Unfortunately, no data have been provided from the remainder of the offshore Gulf of Tunis, nor the Pelagian Sea or the Mediterranean. Figure 10 shows the N–S seismic line and seismic analysis of the Gulf of Tunis, adjacent to the Dorsale and the northern Tunisian trough; some investigations of the presence of early Cretaceous black shales within this basin have been carried out.

4.2. Late Hauterivian Faraoni Anoxic Event

4.2.1. Distribution and characteristics of organic-rich beds

Regionally, an organically rich deep-marine black shale deposition in the Mediterranean region was recorded near the late Hauterivian interval (ca. 133 Ma). This event, termed the Faraoni Event (Cecca 1994; Cecca et al. 1996; Baudin et al. 1996; Coccioni et al. 1998), preserved a high content of organic matter in pelagic successions such as in Italy, Switzerland, the Vocotian basin (France), and southern Spain (e.g. Baudin 2005; Bodin et al. 2006).

This event lies within the Pseudothurmannia catulloi ammonite subzone, which coincides with the extinction of the Lithraphidites bollii calcareous nanofossil species and records an increase in globular planktonic foraminifera. Carbon-isotope records from the Tethyan and Atlantic sections show a minor positive excursion in the uppermost part of the Hauterivian and lowermost Barremian, suggesting accelerated extraction of organic carbon from the ocean reservoir just after the Faraoni Event. Baudin (2005) attributed such an anoxic event to a continental break-up and basalt flooding event of the Paraná–Etendeka continental basalt province, which occurred at 133 Ma during the earliest Hauterivian (Renne et al. 1996; Courtillot and Renne 2003). This increase in crustal production may have created anoxic conditions in the ocean by driving higher fluxes of nutrients to surface waters and/or decreasing oxygen levels (Jenkyns and Clayton 1997; Sanfourche and Baudin 2001). Organic-rich late Hauterivian strata occur in many places in northern and northeastern Tunisia (e.g. the Gulf of Tunis and Jebel Oust; Figure 11) and are grouped into the M’Cherga Formation, which was defined by Ben Ferjani et al. (1990). The distribution of the lower M’Cherga Formation in Tunisia is illustrated also in Figure 11 (compiled after Busnardo and Memmi 1972; M’rabet 1981; Memmi 1989; Ben Ferjani et al. 1990; Soua 2007; Sandman 2008; and our field measurements). Special attention has been paid to the organic-rich laminated black shale levels included within the Bouhedma Formation in the Gulf of Gabes (KET-1, AJM-1, DJM-1, and EZZ-8 wells) and in the northern Chott Basin (Khenguet Aicha section; Soua et al. 2011a). The regional distribution of the main late Hauterivian facies types is shown in Figure 11, which also includes the distinction of organic-rich lower M’Cherga and intra-Bouhedma facies trends. The organic-rich late Hauterivian exists in several areas, namely the northeast (Jebel Oust, BAR-1, and CB-101 wells), the north (Jebel Rihane, Toukeber), the offshore Gulf of Tunis (Raja-1 well), central Tunisia (between Jebel Chirichira and Jebel Mghilla; Figure 7), the Zaouia area (ZA-1 well), and the southern Gulf of Gabes domain (see Figure 11 for locations). Unfortunately, several neighbouring drilled wells did not reach the early Cretaceous strata, and therefore much information about the late Hauterivian anoxic event drilled is unknown. Rare available well and outcrop data confirm this organic-rich province. Geophysical evidence of the distribution of Early Cretaceous black shale deposits is shown in Figures 7 and 11, which illustrate geoseismic cross sections between Jebel Mghilla and Jebel Chirichira, as well as in the Gulf of Tunis through the Gulf of Hammamet, exhibiting black shale deposition within half-graben systems and bordering normal faults.

The palaeogeographic map of the late Hauterivian (Figure 11) suggests that the distribution of organic-rich lower M’Cherga and intra-Bouhedma horizons is restricted to intermediate to medium and shallow
Figure 10. Regional geoseismic sections in the Gulf of Hammamet (A–A') and Gulf of Tunis (B–B') (see Figure 5 for location).
water depths, respectively, which may represent the impingement of an OMZ onto the northern margin of Africa. The lower M’Cherga Formation is generally tens of metres in thickness (up to 40–60 m, Memmi 1989; M’Rabet et al. 1995) in the Jebel Oust area (northeast Tunisia), with a maximum thickness of about 200 m at the BAR-1 well (Figure 12). In the Gulf of Tunis, the Raja-1 well, the lower part of the M’Cherga Formation (Soua 2007; Sandman 2008) is also represented by basinal black shales interbedded with finely laminated limestone and dark marl beds (70 m). Generally, the lower M’Cherga consists of a regular alternation of dark-coloured shale and laminated carbonate beds included within grey to dark grey marl. In addition, the upper part of the Bouhedma Formation, in the Gulf of Gabes, Jeffara, and Khanguet Aicha (Soua et al. 2011a), shows black shale levels alternating with limestone and marlstone beds with a mean TOC value of 1.82% (Brehm and Trichelli 1991; Figure 12). Unfortunately, no additional organic carbon has been analysed. SGRs are available from the BAR-1 well, which provides uranium, thorium, and potassium.

Figure 11. Palaeogeographic and facies map of the late Hauterivian in Tunisia, as well as the black shale distribution of the Faraoni Anoxic Event (F-OAE).
Figure 12. Late Hauterivian Faraoni Event recognized in some wells in different domains (see Figure 4 for location). The definition of the Faraoni Event is expressed by the redox-sensitive U/Th ratio. Note that in different domains, the U/Th ratio is characterized by two distinct peaks. Analysis of the spectral gamma ray (SGR), uranium, thorium, and U/Th ratio as well as the SGR-INPEFA of (a) well BAR-1, (b) well AJM-1, and (c) well EZZ-8.
Figure 12 illustrates the pattern of the U/Th ratio through the M’Cherga and Bouhedma Formations in the BAR-1 (north-central Tunisia), Ajim-1 (Gulf of Gabes), and EZZ-8 wells (Jeffara basin) (Table 1). A significant increase in the redox-sensitive proxy is observed in the interval comprising 2050–2250 m in the BAR-1 well, where two distinctive peaks are observed. The first peak is at 2060 m, characterized by a U/Th value of about 3.9 (Table 1), which is suggestive of deposition in a severe anoxic environment (Jones and Manning 1994). At approximately 2210 and 2215 m, two significative peaks (3.1 and 2.5, respectively) result from deposition under anoxic conditions. The Faraoni Event is well recorded by the U/Th redox proxy in the BAR-1 well and is thought to be placed at 2060 m. The same pattern was observed in the EZZ-8 well, with significative U/Th values ranging between 2 and 9 and suggestive of a dysoxic/anoxic bottom water environment during deposition in the Bouhedma Formation (Figure 12). Typical observed ranges of U/Th include a pattern with a sine-like shape and sudden increase in values (peaks) within the late Hauterivian Bouhedma and lower M’Cherga intervals in the BAR-1, EZZ-8, and AJM-1 wells (Figure 12). Figure 12 indicates the wide extent of severe bottom redox conditions during the late Hauterivian in the southern Tethyan margin. This was confirmed by the use of the Schmoker ratio, which depends on the density parameter in the BAR-1 well. The Schmoker ratio indicates that the Faraoni interval recorded a TOC up to 18% (wt%).

In the Gulf of Tunis, the Raja-1 well (Figure 11) penetrated a similar stratigraphic section within the late Hauterivian (Soua 2007; Sandman 2008). This section comprises several black shale levels alternating with thick marlstone intervals and limestone succession. The CB-101 well also recorded organic-rich black shale levels in a coeval interval.

The calculated U/Th ratios from different domains (Figure 12) confirm the maximum levels of reducing conditions recorded in the Tethyan realm during the Faraoni Event (Cecca 1994; Cecca et al. 1996; Baudin et al. 1996; Coccioni et al. 1998; Baudin 2005; Bodin et al. 2006; Martinez 2013).

The low TOC values characterizing some late Hauterivian sections can be also explained by the fact that the domain analysed could be linked to an organically leaner location or even probably deposited underneath the OMZ.

4.2.2. Biostratigraphy and sequence stratigraphy

Generally, the Hauterivian–Barremian transition in northern Tunisia is recorded mainly by foraminiferal-rich marls alternating with mudstones, wackestones with abundant ammonites, and belemnites with some black shale levels (Memmi 1969; Busnardo and Memmi 1972; Souquet et al. 1997). In Ajim well (Brehm and Trichelli 1991; Soua 2013a), biostatigraphical constraint for the Bouhedma Formation (2975–3726 m) using dinocyst datums (Figure 13) shows that the first occurrence of Muderongia simplex is recorded at 3250 m (in the lower Bou Hedma Formation), indicating an early Hauterivian age for this level. At around 3000 m, Ctenidodinium elegantum occurs almost simultaneously with Aptea anaphrissa as well as Pseudoceratium pelliferum in the upper part of theFormation sediments, indicating latest Hauterivian-aged assemblage, although A. anaphrissa and M. simplex are markers of the early Hauterivian (Pestchevitskaya 2007). Williams (1974) defined the C. elegantum–P. pelliferum Zone as characterizing the Hauterivian and was redefined later by Bujak and Williams (1978) as the C. elegantum Assemblage Zone characterized by the first occurrence of Subtilisphaera perlicuda, among others. In the Bouhedma successions of the Ajim well, S. perlucida occurs with other Subtilisphaera species in its upper part and in the middle part of the same formation in the KET well, also situated in the Gulf of Gabes. So, when referring to these biostatigraphic data, we may attribute the Bouhedma Formation to the late Hauterivian (Figure 13).

The lower M’Cherga Formation was deposited during the late Hauterivian Anoxic Event (Sayni Zone, Memmi 1969), and may continue through the overlying Balearis Zone (Souquet et al. 1997). Busnardo and Memmi (1972) originally proposed an early to late Hauterivian age for these black shales in the Jebel Oust area, confirmed later by Memmi (1989) and Ben Ferjani et al. (1990). This stratigraphic interpretation was later reevaluated in Souquet et al. (1997), who determined that these shales pertain to the late Hauterivian. These shales yielded diversified ammonites represented generally by P. thetys, P. infundibulum, P. royanum, Eulytoceras sp., N. gracianum, Neocraspedites phillipsi, and abundant belemnites such as Hibolites subfusiformis, Hiobolites sp., and Duvalia dilata (Memmi 1969) belonging to the Hedbergella hauerivica planktonic foraminiferal Zone (Maamouri et al. 1994a). So the age of these black shales is well documented as late Hauterivian.

Generally, the onset of lower M’Cherga deposition in northeastern Tunisia coincides generally with the base of the late Hauterivian Trangressive Systems Tract (TST) (Haq et al. 1987; Vail 1988; Souquet et al. 1997; Hardenbol et al. 1998).
A similar sequence of stratigraphic interpretation was also presented by Souquet et al. (1997). To correlate the TST, we used SGR and the U/Th ratio pattern of the BAR-1, AJM-1, and EZZ-8 wells, respectively (Figure 12).

The third-order Highstand Systems Tract (HST) overlying the late Hauterivian TST along northeastern Tunisia is generally represented by shallower deposits, such as the carbonates, sandstones, and marls of the overlying Barremian beds and equivalent units (e.g. Memmi 1969; Busnardo and Memmi 1972; Mrabet 1981; Maamouri et al. 1994a; Souquet et al. 1997).

4.2.3. Molecular geochemistry
A summarized organic geochemical description of the Hauterivian data from the Gulf of Gabes wells is provided by Brehm and Trichelli (1991) and Soua (2013a).

In the upper part of the Bouhedma Formation in the Gulf of Gabes (Ajim, Ketana, and Midoune wells), an interval described as a brown to black shale and finely crystalline dolostone to lime mudstone was identified as organically rich and is interbedded with a light brown organic-lean dolostone. Only a few samples from this unit have significant TOC contents up to 1.8%. The samples from Ajim have high hydrogen indices (HI, ranging between 650 and 900 mg HC/g TOC), indicating Type I organic matter. A sufficient maturity level of these laminated black shales is confirmed by T\textsubscript{max} values, which are nearly constant between 430°C and 435°C.

The samples taken from this interval are dominated by lower-molecular weight n-alkanes and a low relative abundance of pristine and phytane, suggesting mainly a moderate terrestrial higher plant component supply. The moderate relative abundance of pristane and phytane results, therefore, in low Pr/Ph ratios (0.60–1.20), indicating deposition in an anoxic environment. It suggests also an aquatic depositional environment under reducing bottom conditions (Burgan and Ali 2009). This is concomitant with low levels of C\textsubscript{17}, a terpane distribution, which shows a pronounced C\textsubscript{34} homohopane prominence, a relatively high abundance of gammacerane, and low levels of tricyclic and tetracyclic terpanes. These results confirm the trend of an anoxic event that prevailed during the late Hauterivian and therefore records the Faraoni Event in the southern Tethyan margin.

Figure 13. Biostratigraphy of the late Hauterivian for the Ajim and Kettena wells (Gulf of Gabes).
4.3. Early Aptian Selli Event

4.3.1. Distribution and characteristics of the organic-rich strata

During the early Aptian (Bedoulian), central Tunisia was dominated by shallow-marine sedimentation characterized mostly by dolomitic limestone, bioclastic limestone, shaly limestone, and black shales, which is assumed to have been deposited in a mosaic of middle platform basins (Figure 14) to the north of an Early Cretaceous palaeo-high, the so-called Kairouan Island (M’rabet 1981; Burollet et al. 1983). In the so-called Tunisian trough to the north, basinal sedimentation characterized by shales, shaly limestones, and limestones was deposited (Burollet et al. 1983; Memmi 1989; Ben Ferjani et al. 1990; Saadi et al. 1994; M’Rabet et al. 1995), with some intercalations of turbidites and condensed sections (Saadi et al. 1994). The general facies trend was complicated by a complex tectonic palaeorelief configuration, which resulted in a mosaic of basins and tilted blocks (Figures 7 and 11) (Turki 1985; Bouaziz et al. 2002; Zouaghi et al. 2011). This configuration results in strong lateral thickness and facies variations in the early Aptian strata (Figure 14).

Organic-rich early Aptian strata occur in many places in northern and north-central Tunisia, and are grouped into the Sidi Hamada Formation, which was defined by Tlatli (1981) near Sidi Hamada village situated to the northwest of Jebel Serj and grouped earlier into the Serj Formation by many authors (Burollet 1956; Fournié 1978), or the lower member of the Serj (Ben Ferjani et al. 1990; Zghal 1994; M’Rabet et al. 1995; Bessaies 1998; Benzarti 2002; Zouaghi et al. 2011). It is worth noting that the early Aptian Sidi Hamada Formation should not be confused with the Valanginian Hamada Formation.

This may vary laterally SW to the Bou Laaba dolomite defined by Bismuth et al. (1982) and to the Berrani member, which constitutes the lower part of the Orbata Formation (Ben Youssef et al. 1985).

The regional distribution of the three main early Aptian facies types, modified after Bessaies (1998), is shown in Figure 14. The organic-rich early Aptian exists in three areas, namely northeastern (Rs/Ressas, ms/Mssalla sections) and central Tunisia (Hm/Serj section), offshore northeastern Tunisia (Raja well), and western-central Tunisia (Thala, Khechem El Kelb, and Semmama wells; see Figure 14 for locations). Several neighbouring drilled wells have penetrated black shale horizons or have confirmed the extension of the Lower Cretaceous emerged areas. These data provide new insights into the distribution of the organic-rich provinces of Tlatli (1981), although the exact boundaries remain unclear, especially in the Pelagian Sea, central Tunisia, and Algeria. Figure 15 shows the U/Th ratio from the BAR-1 well plotted against the Schmoker ratio, which indicates

Figure 14. Facies, TOC, and maturity distribution maps of the early Aptian OAE1a (Selli Event) in Tunisia. Data are compiled from M’rabet (1981); Burollet et al. (1983); Heldt et al. (2010); Zouaghi et al. (2011); and Elkhazri et al. (2013). (a) Black shale distribution of the early Aptian Selli event and (b) maturity (Ro%) and organic richness (TOC%) distribution of the early Aptian Selli Event (maturity and TOC values are from Elkhazri et al. 2009, 2013).
values of TOC near to 1%. The U/Th ratio could then illustrate the extension of the organic-rich facies into the Bouarada area. The U/Th ratio ranges between 0.5 and 1.41 (Table 1, Figure 15), indicating bottom water redox conditions. A comparison between the deposition environments from the palaeogeographic map (Figure 14) and the distribution of the organic-rich early Aptian is probably restricted to intermediate to medium water depth and may represent the impingement of an OMZ onto the northern margin of Africa. The Sidi Hamada Formation is generally a few hundred metres thick (100–350 m), with a maximum thickness of about 400 m in the northwestern Tunisian trough. However, the early Aptian black shale package, which corresponds to the Selli Event, does not exceed 20–40 m. In the Jebel Ressas (Rs section) and Msalla (ms section), the upper part of the M’cherga Formation is also represented by hemipelagic black shales interbedded with finely laminated limestone (Souquet et al. 1997; Elkhazi et al. 2009), while Saadi et al. (1994) described a condensed section in the Aptian–Albian section of Jebel Mdeker (see Figure 14 for localization). Generally, the Sidi Hamada consists of two units, the lower generally containing silty bioclastic to peloidal bioclastic wackestone and packstone providing planktonic foraminifera, radiolarians, small benthic foraminifera, ostracods, and sponge spicules with scarce quartz grains. This unit passes laterally toward the southwest to oolitic and bioclastic limestone that alternates with argillaceous limestone, silt, clay, and fine-grained sandstone. The upper unit consists of mudstone to bioclastic wackestone containing planktonic foraminifera, small benthic foraminifera, and ostracods. This unit passes laterally to dolomitic and bioclastic carbonate beds, with rare beds of marl and clayey limestone (Tlatli 1981; Zghal 1994; Benzarti 2002; Elkhazi et al. 2009; Heldt et al. 2010; Zouaghi et al. 2011).

The first unit is characterized by a regular alternation of dark-coloured shales and laminated carbonates and nodular black and greyish limestones and marls with TOC values of up to 1% (Lehmann et al. 2009; Elkhazi et al. 2009; Heldt et al. 2010) (Figure 14). A single 1.99% TOC value was reported by Elkhazi et al. (2009) from the Rs section (Jebel Ressas). The vertical distribution of limestone vs. marl in the different localities depends on the position on the palaeoshelf/slope (Supplementary Figure 2). A general overview of the palaeogeography during the early Aptian Sidi Hamada black shale levels may explain that generally in proximal settings it acquires more carbonate-dominated deposition with abundant bioclasts and becomes dominated by marls in distal settings. Particularly in the Rs/Resas and ms/ Msalla sections (Elkhazi et al. 2009) and the lower 55 m of the Hm/Serj section (Heldt et al. 2010), we note that the TOC values in bioclast and benthic foraminifera-rich calcareous and laminated beds are generally locally organically lean, varying between 0.2% and 0.3%. In the Hm section, the typical range observed includes a pattern with a sineshape and gradual increase and decrease in values. Close inspection of the Sidi Hamada spiky TOC pattern (Supplementary Figure 2), in combination with typical alternations of black shales and organic-rich marls with carbonate-dominated beds, demonstrates the significance of anoxic and dysaerobic environment cycles during deposition of the early Aptian.

The low TOC values characterizing some early Aptian sections can be explained by the absence of the basal part of the Aptian black shale interval (typically organically rich in the Hm section (Supplementary Figure 2) and in the BAR-1 well (Figure 15, Table 1); or by the

Figure 15. Detailed analysis of the uranium and redox-sensitive U/Th ratio, which express the Selli Event (OAE1a) in the BAR-1 well.
In general, the early Aptian of northeastern Tunisia (e.g. the Ressas and Msalla sections; Elkhazri et al. 2009) is marked by alternating marls, limestones, and laminated black shales containing some phosphate and echinoderm debris, radiolarians, ostracoda, and relatively scarce benthic foraminifera. In Jebel Serj (Hm section), Heldt et al. (2010) and Lehmann et al. (2009) described these black shales and attributed them to the Bedoulian. So, when referring to these authors, the age of these black shales is now well documented as early Aptian.

In the northern part of the Gulf of Tunis, these black shales are very well documented, especially in the Raja-1 well. They are reported as early Aptian black shales spanning the *L. cabri* Zone (Soua and Smaoui 2008; Sandman 2008).

The euxinic conditions in the Kasserine area (KEK, SEM, and THA in Supplementary Figure 2; see Figure 14 for location) are clearly diachronous at both the base and the top. We may distinguish a typical facies of black shale and shaly limestone within the Bedoulian. Figure 14 shows that structural interpretation inferred from the isopach map of the Bedoulian (Bessaies 1998) may represent an organic-rich and euxinic sedimentation suggested by the development of restricted circulation within tilted blocks (half-graben systems) and confirmed by the SGR analysis in the same area, which provided the redox-sensitive U/Th ratio 

The onset of oxygen depletion is assumed to have prevailed later in the early Aptian, where the lower unit of the Serj Formation (Sidi Hamada equivalent) is typically missing and decreases in thickness toward the SE (Zouagh et al. 2011). The onset of deposition of the Sidi Hamada Formation in central Tunisia generally coincides with the base of the early Aptian TST of a major third-order Tethyan eustatic sea level (Haq et al. 1987; Vail 1988; Hardenbol et al. 1998; Haq and Schutter 2008).

Particularly in the Hm section, the SB is situated within the underlying succession of the M’Cherga equivalent unit. The TS is confined to the black shale beds and falls with the carbon-isotopic segment C5.

To correlate TST we used the gamma ray profile patterns of the Raja well (Gulf of Tunis) and THA-1, SEM-101, KEK-1, and BAR-1 wells (Supplementary Figure 2).
The third-order HST, which overlies the early Aptian TST, is generally represented by shallower deposits, such as the shallow marine, bioclastic to peloidal wackestones of the lower Serj Formation and equivalent units (e.g., Tlatli 1981; Saadi et al. 1994; Elkhazri et al. 2009; Heldt et al. 2010; Zouaghi et al. 2011).

4.3.3. Palaeoecology

Comprehensive palaeoecological investigations have been carried out on the early Aptian strata by several authors (Ben Youssef 1999; Lazzez and Ben Youssef 2008; Lehmann et al. 2009; Elkhazri et al. 2009; Heldt et al. 2010) using microfacies analysis, planktic and benthic foraminifera, as well as ammonite fauna.

The radiolarian assemblages during OAE1a in Tunisia (as studied by Heldt et al. 2010) are characterized by high abundance, which has been attributed to high fertility and high nutrient concentrations in surface water during deposition of the Sidi Hamada Formation under upwelling settings, which provide nutrients and dissolved silica (Racki and Cordey 2000). Heldt et al. (2010) and Elkhazri et al. (2009) noted that radiolarians co-occur with planktonic foraminifers over most of OAE1a in several sections, suggesting meso- to eutrophic conditions in the upper water column of the study area and possibly higher oxygen levels.

The planktonic foraminiferal assemblages, as studied by Heldt et al. (2010), Elkhazri et al. (2009) and by Ouadday and Saddem (2012), are mainly dominated by globular and primitive forms given mostly by hedbergellids, scarce globigerinelloids, and high-diversity leucoleids. These forms are adapted to meso- to eutrophic conditions in the upper water column and tolerate significant fluctuations in temperature, salinity, nutrients, and oxygen. This group has often been interpreted as an indicator of eutrophic conditions and low-oxygen levels in the upper water column. The OAE1a interval of the Sidi Hamada Formation is also characterized by a low abundance of benthic foraminifera, ostracods, and ammonite, and is possibly related to a dysoxic setting at the bottom–water interface (Lehmann et al. 2009).

4.3.4. Carbon-isotope chemostratigraphy

Only two carbon isotopic studies were scheduled by Heldt et al. (2010) and Lehmann et al. (2009) for the early Aptian OAE1a in the Serj section, and by Elkhazri et al. (2013) for the northeastern Tunisia sections. Recently Godet et al. (2014) published a C-isotopic curve ($\delta^{13}$C) of the dolomitic Barrani member (lower part of the Bouhedma Formation) in the Bir Oum Ali area. The authors did not find a complete recording that would attribute the absence of some characteristics to erosion or non-deposition of sediments.

Figure 16 shows a tentative correlation between the Tunisian OAE1a sub-peaks of the carbon-isotopic excursion and the globally recognized C1–C8 segments from Sierra Madre (Mexico) by Bralower et al. (1999).

![Figure 16](image-url)
Figure 16 is a composite carbon-isotopic and organic carbon section for Tunisia, compiled from different works (Tlatli, 1981; Lehmann et al. 2009; Elkhazi et al. 2009, 2013; Heldt et al. 2010). The $\delta^{13}C_{\text{carb}}$ values of the Hm section range between $-1.2$ and $4\%$, Heldt et al. (2010) divided the $\delta^{13}C_{\text{carb}}$ curve into segments varying from C1 to C8 for chronostratigraphic purposes. In detail, the C1 segment (fluctuating values between 0.1 and 1.7%) is characterized by an increase in $\delta^{13}C_{\text{carb}}$ values, followed by an overall negative shift (C2 segment) and minima characterizing the C3 segment ($-1.2\%$). A significant increase in segment C4 values followed by an interval with more or less stable $\delta^{13}C_{\text{carb}}$ values (C5, 2.4%) is noted. Then, it is followed by a further increase (C6 segment), variable but overall high values (C7 segment, 0.9–3.5%), followed by fluctuating values between 0.81 and 4% (C8 segment). On close inspection, the $\delta^{13}C_{\text{carb}}$ sub-peaks in Tunisia do not correlate with the TOC maxima and confirm the excursion’s global stratigraphic significance. In the Tunisian realm, higher TOC values generally underwent significant variation after the onset of the $\delta^{13}C$ excursion.

4.4. Albian Anoxic Events

4.4.1. Distribution and characteristics of organic-rich strata

During Albian time, Tunisia displayed a structural scheme as still being formed by a N-dipping slope at the southern Tethyan margin, also including a variety of facies transects ranging from continental over shelfal to bathyal (Figure 17).

A well-defined E–W transpressional to compressional event characterized the early Albian, during which several structural features were created (anticlines, folds en echelon, and reverse faults) (Ben Ayed and Viguier 1981; Elkhazi et al. 2009) for the Rs and Hm sections. The organic-rich Sidi Hamada black shales formation is characterized by mixed type II/III kerogene (i.e. planktonic marine type II and ligneous and hemicellulosic continental type III kerogen) with TOC concentrations of up to 1.99%, indicating suitable rock qualities for oil. A sufficient maturity level of these laminated black shales is confirmed by $T_{\text{max}}$ values nearly constant at 438–459°C (Figure 14b). In the Raja Well a single value of 1.2% of TOC has been given with HI equal to 435. In addition, in the Kasserine area (THA and KEK wells), the TOC ranges 0.22–0.87% with HI ranging 434–437°C (Figure 14b).

4.3.5. Molecular geochemistry and deposition environment

Comprehensive organic geochemical description of the early Aptian Sidi Hamada Formation was given by...
Turki 1985; Messaoudi and Hammouda 1994; El Euchi et al. 2002). Several parts displayed erosion (Figure 17), which was very active during this period, sometimes inducing emerged areas characterized by early diagenesis, karstification, cementation, and dolomitization (M’rabet 1981; Zghal 1994; El Euchi et al. 2002; Zaghib-Turki 2003).

In general, during the early to late Albian, four zones of black shale accumulation have been differentiated in Tunisia. From north to south, these are characterized by a subsiding domain with a high accumulation rate. To the northwest, Tandia (2001) described a low subsiding domain which recorded several unconformities. In the centre, a stable domain was characterized by the infill of pre-existing grabens and to the south an emerged area. The panoply of facies during the Albian is displayed north to south as described below.

To the south, the Saharan shield was not an area of non-deposition as suggested by Zghal and Arnaud-Vanneau (2005), Chihaoi (2008), and Ben Fadhel et al. (2011), but rather it was covered during the early Albian transgression by the Foum El Argoub/Chennini transgressive sandstones and the Redouane/Zebbag platforms (Ben Youssef 1999; Ouaja et al. 2004; Soua 2009; Bodin et al. 2010), where the Knemiceras level (Ain El Guettar Formation) represents the equivalent of the mfs of the Allam black shale. Central Tunisia was covered by the carbonate platform of the Zebbag Formation. In the Kasserine area, a reefal facies covered the area called the Selloum Sequence (Figure 17) and was controlled by the Kasserine and Sbiba extensional fault systems (Boltenhagen 1981; Bismuth et al. 1982; Zouaghi et al. 2009). To the north, the basin is characterized by the deposition of the Fahdene Formation, which delineates deepwater sediments (Burollet 1956). In the Nebeur area, several authors (e.g. Salaj 1981; Saidi 1993; Tandia 2001; Ben Fadhel et al. 2012) mentioned the detrital input within early Albian strata of the northern Tunisian basins, which accumulated marls and shaly limestones and black shales (Sousquet et al. 1997) within tilted blocks and the Triassic salt tectonics regime. This configuration results in strong lateral thickness and facies variations of the Albian strata (Figure 18; see Figure 17a for the location).

The regional distribution of the four main Albian facies types, modified after Saidi (1993) and Zghal and Arnaud-Vanneau (2005), is shown in Figure 17, which also includes the distinction of an organic-rich early Albian facies. The organic-rich early Albian (lower Fahdene of Saidi 1993) exists in four areas, namely the north and northwestern onshores and northeast and central Tunisia.

A comparison with the palaeogeographic map (Figure 17) suggests that the distribution of the organic-rich early Albian is restricted to water of intermediate to medium depth and may represent the impingement of an OMZ onto the northern margin of Africa. The lower Albian (lower part of the Fahdene Formation) is generally hundreds of metres thick (120–300 m) with a maximum thickness of about 400 m in the northwestern Tunisian trough (Saidi 1993). Generally, the lower Albian consists of a regular alternation of black shales and laminated carbonates, and nodular black and greyish limestones and marls with TOC values of up to 4% (Saidi 1993; Talbi 1993). The vertical distribution of limestone vs. marl in the different localities depends on the position on the palaeoshelf/slope (Figures 19 and 21).

The Fahdene Formation encompasses the remaining Oceanic Anoxic events of the Albian (i.e. the Paquier, Leenhardt, and Breistroffer levels (Figure 19)).

Lithologically, it can be subdivided into five marker members: first, the lower shale member (early Albian), which generally lies unconformably on the Hamaima Formation (if present) or the underlying strata of the late Aptian Serj Formation. Generally, it consists of green shales interbedded with bioclastic limestones partly dolomitized followed by a conglomeratic bed, which is in turn overlain by a thick shale package with thin limestone intercalations displaying pelecypods, crinoids, belemnites, Ticinella raynaudi, and Ticinella rabet (Rami 1998; Chihaoi 2008; Ben Fadhel et al. 2012). The Allam (source rock, middle Albian) is composed of alternating dark grey- to black-coloured marls and decimetric beds of laminated limestones, which consist of mudstones and wackestones. The uppermost limestone bed exhibits a burrowed surface. The middle shale member is composed of dark grey- to black-coloured marls and shales, intercalated with a few decimetric limestone beds, locally containing ammonites. The uppermost part consists of alternating marls and argillaceous limestones locally exhibiting burrows and belemnites. The marls are rich in planktic foraminifera such as Biticinella breggensi and Ticinella roberti (Rami 1998). The Mouelha (late Albian), considered as potential source rock (TOC up to 2.6 wt %, see Ben Fadhel et al. 2011), is represented by black shales and a limestone unit. At the base of this unit, a conglomeratic limestone with an erosive base is generally present. The overlying series are composed of laminated limestone, which consists of mudstone and wackestone beds, including decimetric organic-rich shale–marl interbeds. The basal shales consist of Rotalipora subticinensis (Rami 1998; Ben Fadhel et al. 2012). The upper shale
Figure 18. Lithology, biostratigraphy, and sequence stratigraphy of the Hameima and Charren sections (Chihaoui 2008), a tentative correlation between the two sections and definition of the Albian anoxic events (Paquier, Leenhardt, Haute Noirs, G7, and G8 levels). (a) Hameima section; (b) Charren section.

Figure 19. Tentative correlation of the Albian events (OAE1b–1d) between the Voccontian basin in France (Reichelt 2005) and TAJEROUNE in Tunisia (Chihaoui 2008) using carbon-isotopic events and a global timescale. The correlation is mainly made to express the Paquier, Leenhardt, Haute Noirs 26, and Breistroffer levels (OAE1b–1d of the Albian) within the Fahdene Formation. NB: The Paquier and Leenhardt levels are located in the lower shale member of the Fahdene Formation, Haute Noirs 26 is located within the Allam member, and the Breistroffer Event (OAE1d) is located within the Mouelha member. (a) Voccontian basin in France; (b) TAJEROUNE in Tunisia; and (c) timescale.
member (latest Albian), consists of relatively thick dark shales and marls admitting several argillaceous limestone beds. This unit is overlain by organic rich shales and marls containing ammonites and abundant planktic foraminifera indicating the latest Albian (Ticinella subti- cinesis). This unit is in turn overlain by early Cenomanian green shales containing Rotaliapor apanin-
ica. Ben Fadhel et al. (2010; 2014) discussed the abundance of radiolaria within early and late Albian successions and demonstrated their relationship with the foraminifera.

The TOC ranges vary significantly in different localities (Saidi 1993; Talbi 1993; Ben Fadhel et al. 2011). Typical ranges observed include mainly patterns with a sine shape-like gradual increase and decrease of values within the Albian interval of the Fahdene Formation. A close inspection of the Albian spiky TOC pattern, in combination with typical alternations of dark grey-to-black coloured shales and organic-rich marls with light beds, highlights the significance of anoxic and dysaerobic environments cycles during the deposition (Supplementary Figure 3).

The low TOC values characterizing some early Albian sections can be explained by the partial absence of the lower shaly member of the Fahdene Formation interval (exhibiting significant organic richness), probably capped by the generalized erosion phase (D8–D9 of Chihaoui 2008; Figure 18), where in many places the Allam member is lying unconformably on Triassic or on late Aptian strata (Chihaoui 2008; Ben Fadhel et al. 2014). TOC values of less than 1% could be linked to an organically leaner location of the logged section; or, probably, they were deposited underneath the OMZ.

Alternatively, Ben Fadhel et al. (2011) suggested that the uppermost Allam black shale unit could possibly correspond to the oceanic anoxic event OAE1b Paquier level, while Chihaoui (2008) using carbon isotopic analysis showed that the latter Paquier level is situated in the Lower Shale Unit of the Fahdene Formation (Figure 19) several metres under the Allam member in the Tajerouine area (see Figure 17 for location). Taking into consideration the position of the first occurrence of P. buxtorfi, Ben Fadhel et al. (2011) correlated the Mouelha black shales with the Breistroffer level (OAE1d) identified in the Vocontian Basin (Erbacher et al. 1996; Bréhéret 1997) and with the Pialli level in the Italian Apennines domain (Coccioni 2001), while the Breistroffer level corresponds to the R. appen-
ninica Zone confined prior to the FO of P. buxtorfi. Figure 20 shows a summarized composite δ13C_carb and TOC for the Tunisian Aptian and Albian anoxic events. The Selli, Paquier, and Breistroffer levels are underlined by grey bands and correlated to the Sierra Madre Section in Mexico (Bralower et al. 1999). The composite δ13C_carb is divided into 15 segments. Segments C1–C8 are the division of Menegatti et al. (1998) and segments C9–C15 are the division of Bralower et al. (1999) in the Sierra Madre (Mexico). Figure 19 shows a tentative correlation of the main anoxic events (OAE1b and OAE1d) using carbon-isotopic events both in Tunisia and worldwide. From this figure, we may conclude that (1) the Paquier and Leenhardt levels are exclusively related to the Lower Shale Unit of the Fahdene Formation; (2) the Allam member displays the same carbon isotopic signature as the ‘Haute Noirs 26’ black shale level of Reichelt (2005) in the Vocontian basin; (3) the Mouelha member is the equivalent of the Breistroffer level; and (4) the D9 and D10 sequence boundaries correspond to the Albian major unconformity (Ei Euchi et al. 2002; Zghal and Arnaud-Vanneau 2005), since several carbon isotopic segments are absent in the Tunisian sections.

4.4.2. Biostratigraphy and sequence stratigraphy

The biostratigraphy of the Albian was studied thoroughly by Ben Youssef et al. (1985), Zghal (1994), Memmi (1999), Rami (1998), and Ben Fadhel et al. (2011, 2012). The Albian organic-rich sediments of Tunisia are confined to Hedbergella planispira and Rotaliapor aticinus. What is interesting is the presence of the typical δ13C_carb isotopic expression of the Paquier and Leenhardt levels corresponding to the H. planispira Zone and H. buldti ammonite Zone (Chihaoui 2008) in the Lower Shale Unit of the Fahdene Formation (Figure 18). The two source rock levels of the Albian, Allam (T. primula Zone) and Mouelha (R. ticinus Zone), revealed good TOC levels and match with the C-isotopic events of the HN 26 and the Breistroffer levels, respectively, of the Albian of the Vocontian basin (Figure 19). Moderate abundant radiolarians have been described by Ben Fadhel et al. (2010; 2012) within the Mouelha black shale, which lies generally within the lower part of the Rotaliapor appenninica Zone. This black shale interval is assigned to the early Albian U.A.10 biochronozone of O’Dogherty (1994), which is equivalent to the Romanus Zone (O’Dogherty and Guex 2002). The same authors indicated that the upper part of the Bticinella breggien-
sis Zone is also characterized by radiolarian-rich black shale layers interbedded within cyclic marl/limestone alternations, which correspond to the middle shale unit and the Mouelha member of the Fahdene Formation. These strata belong to the U.A. 13 and 14 biochronozones. These radiolarian zones indicate early to middle-late Albian.

According to Chihaoui (2008), the Albian Fahdene succession generally corresponds to four depositional
sequences (Figures 18 and 21). The first sequence corresponds to the lower shale member (early Albian). The second sequence starts with a basal conglomeratic level (D9). In general, the lowermost part of the Allam member may still belong to either the lowstand systems tract (LST), such as in the Nebeur area (Ben Fadhel et al. 2012), or the shelf margin wedge (SMW), such as in a proximal setting in the Tajerouine area (Chihaoui 2008), where Burollet (1956) first described the lower part of this member as consisting of calcispheres and echinoderm debris, belemnites, and fragments of ammonites, originating probably from the higher shelf (Supplementary Figure 3).

A burrowed surface with glauconite deposit outlining the base of the Allam member in the Nebeur area has been described by Ben Fadhel et al. (2011), probably corresponding to sediment starvation on the shelf during early sea-level rise.

The remaining Allam deposits represent the TST and are composed entirely of pelagic deposits, platy limestone, and laminated black shales with typically significant TOC values (Figure 19). The HST sediments overlying the Allam TST are represented by the middle shale unit of the Fahdene Formation and are capped by the D10 discontinuity (Figures 20 and 22) where the third sequence in the middle shale unit begins. Starting with the D11 discontinuity level, which is characterized by a conglomeratic bed, an identical sequence stratigraphic interpretation of the Allam member is applicable to the Mouelha member, which exhibits the same lithological and geochemical characteristics. The HST sediments overlying the Mouelha TST are characterized by the Defla shale unit, which displays Cenomanian fauna.

### 4.4.3. Carbon isotope chemostratigraphy

Figure 18 illustrates a tentative correlation of carbon isotopic data from two Albian sections from Northwestern Tunisia, complemented by TOC and foraminiferal, radiolarian, and ammonite biostratigraphic results. As in the Vocontian basin (Reichelt 2005), where 18 segments (Figure 19) have been defined based on subcycle development in the δ¹³C_carb curve, the Tunisian sections seem to have developed the same subcycle characteristics. In the Hameima and Charren sections (Chihaoui 2008; Figure 18) some segments
appear to be absent while further segments are not present. In both sections, the negative peak relative to the Paquier event (A12) occurs in the upper part of the H. planispira zone and slightly underneath the FO of T. primula, and the first positive peak relative to the Leenhardt level (A14) occurs just below the D9 unconformity within the U.A. 9 radiolarian Zone (Figure 18). The Allam member records low carbon isotopic values near to −2 per mille, but generally the highest values are recorded within the middle shale unit, comparable to the boundary between A19/A110 in the Voccotian basin. The Beistroffer Event (OAE1d) occurs in both sections in the Mouelha member (A17). This event occurs slightly after the FO of Rotalipora ticinensis while in the section Planomalina buxtorfi is present, which suggests the existence of diachronism or the
continuity of the D11 unconformity in the second section. Slight discrepancies between the biostratigraphically and isotopically constrained positions of the member partly occur.

As in the Vocontian basin (Reichelt 2005) and in Mexico (Bralower et al. 1999), the positive $\delta^{13}$C$_{\text{carb}}$ sub-peaks in Tunisia do not correlate with the TOC maxima, but Figure 20 illustrates a fair to good correlation between negative sub-peaks with the TOC maxima and confirms the global stratigraphic significance of carbon isotopic excursion. In Tunisia, higher TOC values generally develop almost with the onset of the $\delta^{13}$C$_{\text{carb}}$ negative excursion. The detailed individual shape of both the TOC and $\delta^{13}$C$_{\text{carb}}$ excursions depends generally on the sampling density and, to a lesser extent, the local sediment accumulation rates. This is why the two curves of Ben Fadhel et al. (2011) did not show any excursion for the organic-rich lower Albian strata. If the $\delta^{13}$C$_{\text{carb}}$ curves of Chihaoui (2008) are of high resolution, this may highlight some isotopic excursions which have been assumed to be absent in this review.

4.4.4. Molecular geochemistry

A detailed organic geochemical description of the Albian source rocks pertaining to the lower Fahdene Formation and available data from wells was given by Saidi (1993), Talbi (1993), Saidi and Inoubli (2001), and El Euchi et al. (2002). The organic-rich Albian black shales are characterized by mixed Type II/III (i.e. planktonic marine type II and ligneous and hemicellulosic continental type III kerogen; HI ranging 40–700 mg HC/g TOC), with TOC concentrations of up to 4.5% indicating fair to good source rock qualities for oil and gas. A sufficient maturity level of these laminated black shales is confirmed by $T_{\text{max}}$ values, which vary along a NE-SW trend (Supplementary Figure 3), recording two major intervals of about 430–440°C and 440–460°C (Figure 17). Organic petrological and palynological studies on these sediments indicate that kerogen is dominated by amorphous organic matter (AOM), despite the strong marine influence and the preservation of microspores, which suggests minimum distance and probable autochthonous origin. In addition, terrestrial macerals are described from the lower Fahdene source rock extracts, which include vitrinite, sporinite, and cutinite.

The relatively high Pr/Ph ratio (ranging generally between 1 and 3) and the low tricyclic terpane content shows the presence of oxic/dysoxic cyclicity as well as high thermal temperature, since the Pr/Ph ratio increases with increasing temperature. The moderate to high diasterane content reflects the presence of clay minerals and their ability to catalyse sterane rearrangement reactions. In general, the high Pr/Ph ratio concomitant with low $C_{28}/C_{34}$ homohopane ratio (generally <1) suggests mixed oxic to suboxic cyclicity during deposition with hypersaline-influenced sedimentation conditions. Alternatively, the moderate to high $C_{28}/C_{29}$ sterane ratio (generally present between 0.69
and 0.76) is close to that calculated by Grantham and Wakefield (1988) for Cretaceous rocks. These relatively high values are interpreted as being due to the presence of different phytoplanktons as well as an increased diversity of dinoflagellates, coccolithophores, and diatoms in the Cretaceous relative to older strata. This is in agreement with the marine origin of Albian Fadhene Formation organic-rich beds, since a high contribution of land plant material tends to increase the C_{29} sterane content, and consequently reduce the C_{28}/C_{29} ratio.

### 4.5. Late Cenomanian–Early Turonian Event

#### 4.5.1. Distribution and characteristics of organic-rich strata

During Cenomanian–Turonian times, Tunisia, situated at the southern Tethyan margin, was characterized by a general slope almost dipping north, promoting variable facies ranging from shallow water to lagoonal facies (Keker member of the Zebbag Formation in the Jffara basin and the Saharan platform; Ouaja 2003), shelfal facies (carbonate of the Zebbag Formation, southeastern Tunisia; Burollet 1956), to bathyal facies (black shales and marly limestone of the Bahloul Formation in central and northern Tunisia (Maamouri et al. 1994b; Soua and Tribovillard 2007). The general facies trend is related to a complex tectonic and halokinetic activity resulting in variations in C/T transition facies variability (Bechtel et al. 1998; Lüning et al. 2004; Soua and Tribovillard 2007; Soua et al. 2009; Soua 2011). Organic-rich C/T sediments occur in many sites in northern, central, and offshore Tunisia and are grouped into the Bahloul Formation (Burollet 1956). Figure 21 illustrates the thickness distribution of the anoxic C/T transition facies in Tunisia and Algeria (Soua et al. 2009; Soua 2011; Soua et al. 2011b). The regional distribution of the Tunisian C/T facies, including the distinction of an organic-rich and -lean anoxic facies, is shown in three main provinces (Figure 22). It is mainly represented in Hodna basin, Aurès–Kef trough (eastern Algeria), onshore northwestern Tunisia, the Gulf of Hammamet–Gabès, and locally in the Negrine–Gafsa trough (see Figure 21 for location; Soua et al. 2009). Well and outcrop data confirm these organic-rich provinces, and the exact boundaries have been clearly traced in earlier studies (Layeb 1990; Maamouri et al., 1994b; Saidi et al. 1997; Soua and Tribovillard 2007; Soua et al. 2009, 2011b; Soua 2011; Affouri et al. 2013; Layeb et al. 2013), especially north of the Gulf of Gabès domain (Soua et al. 2009). Dealsing with the Soua and Tribovillard (2007) model, the palaeogeographic setting suggests that distribution of the organic-rich facies represents the impingement of an OMZ onto the southern Tethyan margin and upwelling conditions where the organic-rich facies is associated with siliceous deposition (Soua 2013b). The Bahloul Formation is generally 10–30 m thick, with a maximum thickness of over 50 m recorded in northwestern Tunisia (Figure 21), and it consists generally of shale and limestone alternations. Limestone is dark and thinly laminated wackestone/packstone, whereas the shale interval comprises organic-rich laminated black shales with TOC values up to 13.57% (Bechtel et al. 1998; Soua and Tribovillard 2007). In Zaghouan area (Pont du Fahs section), Salaj (1978) mentioned that these black shale facies may continue up to the early Coniacian age, which pertains to the Ejehaf facies (Salaj 1978; p. 385). The same age-equivalent black shales of La Luna Formation are distributed throughout western Venezuela and eastern Colombia (e.g. Zumberge 1984; Torres et al. 2012; Tribovillard et al. 2012). As a result, it appears that the palaeogeographic distribution, regional lithofacies, and areas of subsidence were dependent on sea-level fluctuation and/or tectonic movements (Soua et al. 2009). Alternatively, during this interval, late Cenomanian transgression took place leading to the filling of palaeostructures and the spreading of open marine sedimentation (Lüning et al. 2004). According to Grasso et al. (1999), an Aptian–Cenomanian rifting phase took place on the Tunisian shelf and extended toward the N–S axis to the south along the Tripolitania–Jarrafa–Sirte basins. Soua et al. (2009) showed that extension and normal faults during the late Cenomanian–early Turonian can be inferred from seismic records in the eastern part of Tunisia. Therefore, no deposition of the anoxic C/T carbonate on uplifts and palaeohighs is recorded. During that time, subsidence led to the accumulation of more than 50 m of laminated black shales in a time span of about 0.4 million years in the Tunisian sections (Soua 2010a).

In summary, this period testifies to the reduction of emerged areas inherited from the Early Cretaceous, and thus we note that: (1) to the west, the subsidence is controlled by a preferential NE–SW-faulting trend (Aurès basin–Kef) where maximum C/T subsidence is recorded; (2) to the southwest, the subsidence is controlled by a preferential E–W-faulting trend; (3) to the east, the subsidence is mostly controlled by NW–SE faults; and (4) towards the central and southern domains there is development of a wide platform with a low subsidence rate (Figure 22).

#### 4.5.2. Biostratigraphy and sequence stratigraphy

The Bahloul Formation generally contains the biomarker of the Rotalipora cushima Zone, the Whiteinella archaeocretacea Zone, and/or not the H. helvetica zone (Soua...
2005; Soua et al. 2011b; Zaghibb-Turki and Soua 2013 and references therein). The last occurrence (LO) of *R. cushmani* is generally found above the base of the Bahloul Formation (Supplementary Figure 4). Soua (2005) subdivided the *W. arachaeocretacea* zone into three subzones, which are the *Globigerinelloides benthensis*, the *Dicarinella hagni*, and *Heterohelix moremani*. This latter subzone starts with the globally acknowledged *Heterohelix* shift. Alternatively, the Bahloul was deposited during the UA18 radiolarian chronobiozone, dominated by nassellarian, whereas the UA21 radiolarian chronobiozone is dominated by spumellarian and radiolarian chronobiozones. The radiolarian assemblages recognized in the Bahloul Formation (Soua et al. 2006; Soua 2013b) contain many species providing sufficient accurate biostratigraphical data such as *Stichomitra*, *Pseudodictyomitra*, and *Rhopalosyringium* genera, suggesting a latest Cenomanian to earliest Turonian age. The upper assemblage (UA21 Zone) seems to be early Turonian in age due to the existence of some ammonite species belonging to *Watencoceras* sp, which indicates the onset of the Turonian (Bengtson 1996). In the Gafsa area (Jebel Chemsi; Soua 2011), despite the absence of *R. cushmani* and *H. helvetica*, the Bahloul foraminiferal assemblages indicate that the upper part of the Bahloul just reached the earliest Turonian, denoting therefore that the Bahloul is probably diachronous (Lüning et al. 2004), at least in its upper part. The cyclostratigraphic constraint complemented by biostratigraphy suggests a duration interval of 380–420 thousand years for the organic-rich facies in several C/T sections from Tunisia (Soua 2010a).

### 4.5.3. Palaeoecology

A detailed palaeoecological study was made by Soua (2005) using planktic foraminifera, and by Soua et al. (2006) and Soua (2013b) using radiolarians.

Spanning the upper part of the *R. cushmani* and *W. arachaeocretacea* zones, two planktonic morphotypes dominate the planktonic foraminiferal assemblages: (1) the cosmopolite biserial *Heterohelix* species reaching 60% of the total planktonic foraminiferal population; and (2) triserial *Guembelitria* species. The relative abundance of *G. cenomana* reaches 50% generally. These small triserial morphotypes thrived in highly adverse environmental conditions, even in the eutrophic stage, and are considered as opportunistic species. The associated *Whiteinella* and *Hedbergella* species are less frequent. The abundance of planktonic foraminiferal species decreases suddenly from 25 species to 5 within the lower part of the Bahloul Formation, coinciding with the global *Heterohelix* shift event. Radiolarian assemblages are characterized by a low diversity assigned to a combination of stress conditions and dissolution during the Bahloul deposition. Therefore, at the maximum increase in primary productivity, coupled with long-term oxygen depletion, the nassellarian/spumullarian ratio shows reversal values at the base of the *Superburn* zone and the nassellarians suffered a severe reduction in species abundance. In contrast, spumellarians became predominant. A similar, negative shift of nassellarians is recorded in Italy and Damerara rise, site 1258 (O’Dogherty 1994; O’Dogherty and Guex 2002; Musavu-Moussavou and Danelian 2006). Palaeoecological data suggest that the C/T deposits in Tunisia were restricted to a bathymetric belt, which corroborates the spread of a small OMZ in north-central Tunisia with local upwelling (where organic-rich facies is associated with siliceous deposition) and locally in the Gafsa Basin. This OMZ is marked by decline in the diversity of planktonic foraminiferal, radiolarian, and nanofossils.

### 4.5.4. Cyclostratigraphy

Detailed cyclostratigraphic investigation for the organic-rich Bahloul Formation has been given by Soua (2010a) and Soua et al. (2010). Using four C/T sections in central Tunisia displaying biotic and lithologic variations, Soua (2010a) applied time-series analysis to the abundance of two opportunistic foraminiferal genera (*Heterohelix* and *Guembelitria* spp.) to calculate and estimate both the average sedimentary rate and duration of OAE2. The spectral analysis applied to fluctuations in *Heterohelix/ Guembelitria* spp. points to the presence of a metre-scale periodicity regarding the two species. The good correlation between the spectral peak ratios and those of the orbital components suggest Milankovitch orbital forcing during deposition in Bahloul. Calculated average sedimentary rates, ranging between 12 and 20 cm per thousand years, are in perfect agreement with the sedimentation rate estimated from Hardenbol et al. (1998) and the relatively similar value (10 cm per thousand years) obtained by Vonhof et al. (1998) for the Mellegue section, while a duration interval of 350–400 thousand years for the Bahloul Formation has been calculated. This interpretation clearly highlights the existence of cyclicity within the Bahloul Formation, which is more complex than is recognizable by visual inspection. This implies that orbital forcing caused the fluctuations in the abundance of *Heterohelix* and *Guembelitria* (Supplementary Figure 4).

From this analysis, it has been recognized as a strong precessional signal comparable to other precessional signals identified all over the world (Kuhnt et al. 1997; Morel 1998; Prokoph et al. 2001; Negri et al. 2003; Meyers et al. 2004; Scopelliti et al. 2006). The palaeoecological preferences of both *Heterohelix* and *Guembelitria* spp. seem to be related to areas of
enhanced surface water fertility and oxygen-minimum zones (e.g. Nederbragt and Fiorentino 1999; Lüning et al. 2004; Soua and Tribovillard 2007). As a result, the long-term fluctuations related to both precession and eccentricity suggest that changes in surface water fertility are linked to the Milankovitch parameters.

4.5.5. Carbon-isotope chemostratigraphy
A threefold subdivision (referred to as peaks I, II, and III) is based on small variations in the C isotopic profiles that characterized the majority of the isotopic curves (δ\(^{13}\)C\(_{org/carb}\)) of all sections studied, as indicated by several authors (Bechtel et al. 1998; Barrett 1998; Nederbragt and Fiorentino 1999; Soua and Tribovillard 2007 and references therein; Supplementary Figure 5). In all sections studied, except for the Gafsa area where no foraminiferal studies were performed, the first isotopic excursion (I) always precedes the specialized keeled rotaliopids (LO). The second isotopic excursion (II) generally occurs above the last occurrence (LO) of R. cushmani and the third (III) just below the appearance of the Quadrum gartneri nanofossil zone marker. The Cenomanian–Turonian boundary usually occurs slightly above the maximum δ\(^{13}\)C values. Soua and Tribovillard (2007) suggested that δ\(^{13}\)C\(_{org}\) seems to be a better correlation tool than δ\(^{13}\)C\(_{carb}\) for the C/T boundary.

4.5.6. Major and trace element chemostratigraphy
Summarized major and trace elements from chemostratigraphy have been studied by Bechtel et al. (1998), and a detailed chemostratigraphic study on the Bahloul has been published by Soua et al. (2011b).

Enrichment in V, Mo, and U within the black shales of the Bahloul Formation indicates that the depositional environment was probably sulphidic (see Brumsack 2006) and depleted in oxygen. The highest enrichment in U suggests bacterial sulphate reduction reactions (Tribovillard et al. 2006). Using U/Th, V/Cr, Ni/Co, and V/(V+Ni) proxies (Supplementary Figure 6), it will be observed that samples from the lower part of the Bahloul show the highest values, suggesting an oxygen-depleted environment.

Alternatively, according to Scopelliti et al. (2006), enrichment in Cr could be indicative of an algal source of organic matter (Supplementary Figure 7). These authors state that organic matter could concentrate Cr during high-redox conditions where reduced Cr\(^{3+}\) availability enabled reaction with organic matter (Hatch and Leventhal 1992). This is in agreement with rock geochemical data deduced from the Bahloul Formation samples (Montacer et al. 1988; Layeb 1990; Saidi and Inoubli 2001).

In addition, enrichment in Ba, Cu, and Ni suggests some increased productivity at the time of the Bahloul deposition (Supplementary Figure 8) inducing severe sulphate-reducing conditions (McManus et al. 1998). The positive excursion of the Ni/Al and Cu/Al ratios indicates episodes of increased OM influx and, therefore, increased primary productivity. This could indicate barite dissolution and Ba remobilization within moderate reducing conditions below the sediment–water interface. This is in agreement with the very low Mn levels, which suggests the absence of suboxic conditions close to the sediment–water interface. In addition, the increased Si content within C/T transition beds outcropping in the Bargou area is probably linked to locally increased biogenic productivity by silica-secret ing organisms (mainly radiolarians) (Supplementary Figure 7B), consistent with coeval enrichment of the productivity proxies (Ni, Cu, and Ba). Siliceous deposits could also be linked to hydrothermal input taking into account Ni and Cu enrichment (see Brumsack 2006). Deposition of the organic-rich Bahloul Formation coinciding with the OMZ is consistent with results from redox-sensitive elements (Supplementary Figure 8). This implies a moderate to high degree of barite saturation. In addition, Ba\(_{\text{xs}}\) data within the Bahloul indicate some preservation of barite in the Bahloul Formation, which suggests the existence of a moderate rate of sulphate reduction at the water–sediment interface. Enrichment in U\(_{\text{auth}}\) may indicate an anoxic tendency at the water–sediment interface during the Bahloul Formation deposition (Supplementary Figure 7B).

4.5.7. Molecular geochemistry
A detailed organic geochemical description of the Bahloul Formation source rock and available data from wells is given by Herbin et al. (1986); Montacer et al. (1988); Farrimond et al. (1990); Layeb (1990); Bechtel et al. (1998); Barrett (1998); Saidi and Inoubli (2001); El Euchi et al. (2002); Lüning et al. (2004); and Soua and Tribovillard (2007) (Table 3). The organic-rich Bahloul black shales are characterized by planktonic marine Type II kerogen (IH 200–700 mg HC/g TOC) with TOC concentrations up to 13.5%, indicating a good source of rock qualities for oil and gas. A sufficient maturity level of these laminated black shales is confirmed by the OM thermal maturity, which presents a rather homogeneous distribution with T\(_{\text{max}}\) ranging between 430°C and 500°C (Figure 22).

The Bahloul displays a kerogen, which is generally dominated by yellow–orange to bright yellow–orange fluorescing marine AOM, where dull yellow fluorescence is assigned to the presence of dinoflagellate cysts. The dominance of dinoflagellate cysts in the palynomorph
Table 3. Sections/wells, TOC (%), and thickness of C/T transition in various domains of Tunisia (after Soua 2011 and references therein). See Figure 4 for location.

| Sections/Wells | TOC content | Thickness | Reference |
|---------------|-------------|-----------|-----------|
| Bir BouEcha   | 2.95        | 25        | Saidi et al. (1997) |
| Toukabber     | 3.5         | 18        | Layeb (1990) |
| J. Jedidi     | 2.65        | 25        | Layeb (1990) |
| O. Khedija    | 2.45        | 40        | Saidi et al. (1997) |
| O. Siliana    | 2.03        | 44        | Layeb (1990) |
| J. Cheid      | 2.84        | 25        | Barrett (1998) |
| J. Boukhil    | 1.8         | 23        | This work |
| J. Kebbouch   | 4.51        | 15        | This work |
| Bougrine      | 2.85        | 55        | Barrett (1998) |
| Ain Zakkar    | 2.45        | 37        | Soua (2011) |
| J. Bellouta   | 2.28        | 39        | Layeb (1990) |
| Skarna        | 2.26        | 40        | Layeb (1990) |
| Agraine       | 2.76        | 35        | Layeb (1990) |
| Guem Halfaya  | 2.91        | 17        | Soua (2011) |
| J. Slaa       | 1.96        | 18        | This work |
| J. Jerissa    | 1.74        | 24        | This work |
| J. Gara       | 1.78        | 18        | This work |
| O. Bahloul    | 2.75        | 35        | Lüning et al. (2004) |
| Ain Jeddour   | 1.11        | 31        | This work |
| DL            | 0.77        | 25        | This work |
| J. Rebeiba    | 1.22        | 24        | This work |
| J. Trossa     | 1.9         | 18        | This work |
| J. Birino     | 1.89        | 10        | This work |
| ELH           | 0.75        | 18        | This work |
| KT            | 1.12        | 24        | This work |
| REF           | 0.37        | 25        | This work |
| CN            | 0.45        | 30        | This work |
| EIM           | 0.75        | 8         | This work |
| ALO           | 1.4         | 41        | This work |
| MGR           | 2.57        | 21        | This work |
| KSF           | 0.4         | 10        | This work |
| MAG           | 0.4         | 15        | This work |
| DEN           | 0.6         | 5         | This work |
| KKN           | 0.4         | 8         | This work |
| EHJ           | 0.8         | 16        | This work |
| GB            | 1.2         | 20        | This work |
| SIT           | 0.8         | 16        | This work |
| MRJ           | 0.8         | 10        | This work |
| RBH           | 0.8         | 15        | This work |
| ASH           | 1.5         | 60        | This work |
| JAW           | 1.6         | 10        | This work |
| NRB           | 0.58        | 20        | This work |
| ALY           | 0.55        | 30        | This work |
| ZHR           | 0.4         | 33        | This work |
| ISIS          | 0.4         | 28        | This work |
| LSW           | 2.9         | 2         | This work |
| BLI           | 1           | 20        | This work |
| J. Berda      | 4           | 20        | Lüning et al. (2004) |

fraction is not surprising considering that these are the most common form of fossilizing phytoplankton in most Mesozoic marine sediments, and can form a high percentage of fossilized organic-walled microplankton (Tyson 1995). It is possible that local currents were responsible for introducing more terrestrial material in this area, perhaps via submarine valleys. The differences in hydrogen index can largely be attributed to differences in preservation, as observed from variations in kerogen fluorescence. The Bahloul typically has an n-alkane distribution from nC11 to nC36. The relative abundance of pristane/phytane (Pr/Ph) ratios ranging from 1.4 to 2.6 may indicate deposition in an anoxic environment. Generally, levels of n-alkane with a maximum around nC16 indicate a dominant algal phytoplankton and/or bacterial input, whereas maxima around nC26–nC37 suggest a stronger terrestrial input possibly generated from a source of plant wax, indicating land plant input and generally a more proximal setting.

Abundant hopanes (total hopanoids/total steroids ratio range 1.1–3.1), and also relatively high levels of bisnorhopane and methylbisnorhopane, are interpreted as a specific bacterial source possibly associated with a highly reducing, sulphur-rich depositional environment or bacterial mats in an upwelling environment under marine anoxic conditions. The relative increase in methylhopanes is more likely to be related to the general increase in hopanoids in these samples, which is a likely result of increased bacterial productivity. It is tentatively suggested that methylhopanes have a specific bacterial source such as cyanobacteria or methyltrophs.

5. Conclusion

Considerable attention has been given to unconventional oil and gas shale in northern Africa, where the most productive Palaeozoic basins are located (e.g. Berkine, Illizi, Kufra). Recently, such attention has been given by some oil companies operating in central and eastern Tunisia, where Cretaceous Bahloul and Fahdene Formations are the principal unconventional shale targets. In this synthesis, Cretaceous OAEs have been reviewed and assessed in the southern Tethyan margin with a special focus on Tunisia. Unfortunately, published research on the Cretaceous OAEs in Tunisia did not exceed the Aptian Selli Event (OAE1a), the Albian events (OAE1b,d), and the Cenomanian–Turonian Bonarelli Event (OAE2). In this review, additional OAEs were potentially identified using the U/Th ratio, which is utilized as a redox-sensitive proxy. These OAEs pertain to the well-known Valanginian Weissert and late Hauterivian Faraoni events. A specially integrated approach was employed in this contribution illustrating a detailed distribution of the organic-rich facies, biostratigraphy, chemostratigraphy, isotopic-carbon stratigraphy, and molecular geochemistry. Other techniques used were spectral gamma rays, which provided levels of uranium, thorium, potassium, and cyclostratigraphy. Such integrated approaches have demonstrated new potential pathways for oil and gas shale unconventional plays. The Valanginian and late Hauterivian (Weissert and Faraoni events) demonstrated new potential pathways for oil and gas unconventional plays. The Valanginian and late Hauterivian (Weissert and Faraoni events) are still under debate, which make it mandatory to instigate integrated studies on their distribution to elucidate the occurrence of global short-term related events.

The patterns of the U/Th ratio have provided considerable new insights concerning the marine Early Cretaceous sediments studied (basinal M’Cherga and
tidal flats of the Bouhedma Formations). Our study focus, particularly on bottom water redox conditions, corroborates data published elsewhere and identified for the first time in Tunisia. Consequently, further extensive investigation is suggested to better assess such oil and gas unconventional plays.

This contribution can therefore help in dealing with global correlations, especially with the well-studied northwestern Tethyan domain, and in opening new discussions regarding the origin and principal palaeoceanographical control of the Cretaceous anoxic events.

Acknowledgements

I wish to thank Dr Moez Ben Fadhel (Faculty of Sciences of Gafsa, Tunisia) who conducted a thorough review of the text and two anonymous reviewers for providing numerous suggestions, which improved the text. The useful constructive reviews and suggestions by the Editor-in-Chief, Robert Stern, are much appreciated. I am grateful for the fruitful discussions with Sabri Arfaoui. I would also like to thank Saudi Aramco and GLTSD for permitting me to publish this paper.

Disclosure statement

No potential conflict of interest was reported by the author.

Funding

I am grateful to Primoil Tunisia, which funded the field trip to Jebel Oust in April 2013.

Supplemental data

Supplemental data for this article can be accessed at 10.1080/00206814.2015.1065516

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