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The Albian Oceanic Anoxic Events record in Central and Northern Tunisia:

Geochemical data and paleotectonic controls

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Abstract

The Albian organic-rich successions of the lower part of the Fahdene Formation (Albian to Cenomanian, Tunisia) were studied using sedimentology (analysis of carbonate contents and observation of thin sections), bulk organic geochemistry (Rock-Eval pyrolysis), and molecular biomarker distributions. The selected outcrops cover different structural domains from western central Tunisia (Jebel Hamra) to the Diapir zone or the Tunisian Trough (Koudiat Berkouchia, Jebel Ghazouan and Ain Asker from the southwest to the northeast, respectively).

Total Organic Carbon values between 0.17 and 3.43% evidenced the record of the Albian oceanic anoxic events OAE1b and OAE1c in this southern Tethys margin. Rock-Eval data and molecular biomarkers (n-alkanes, steranes and terpanes) revealed thermally mature organic matter (OM) consisting of a mixed Type II/III OM (marine planktonic and continental origins). The abundant tricyclic terpanes, C₃₀ 17 α (H) Diahopanes, and the regular decrease of homohopanes (C₃₁ to C₃₅) in the terpane series, were also indicative of a high contribution of bacterial OM that had been developed under stagnant dysoxic/anoxic environments.

Microscopic thin section observations showed OM-rich micro-laminated mudstone to wackestone microfacies with abundant radiolarian and foraminifera. All of these features point to high surface primary organic productivity (POP) and the installation of an oxygen minimum zone (OMZ) at mid-depth of the Albian Basin. We suggest that the POP was the result of increased rich nutrient surface waters by upwelling and submarine volcanic activity. We integrated organic geochemical analysis with the structural context to explain both the variation in the OM richness and the record of these OAEs. The reactivation of faults limiting tilted blocks during the Aptian-Albian extension associated to the ascension of Triassic evaporites controlled the distribution and the preservation of organic-rich sequences from central-western to northern Tunisia. Therefore, we suggest that, as sea level rose during the Early Albian transgression, the structurally low areas invaded (grabens and inter-diapiric rim-synclines) represent very deep marine basins that lay beneath the OMZ limits. Therefore, the installation of an OMZ only coincided with half-graben structures that were either proximal and/or around the Triassic Diapir horsts. This could be the case of the Jebel Hamra and Koudiat Berkouchia localities during the Early Albian. We further propose that these localities may have recorded the OAE1b in the southern Tethys margin. Furthermore, the oscillations of this OMZ related to fluctuating POP and upwelling nutrient supplies, as well as the continental material supply by intensive weathering and seasonal rainfall, could explain both the facies alternations (marls/marly limestones) and the OM richness fluctuations along each studied section. The OAE1c recorded in the studied sections could result from the expansion of the OMZ all over the deep ramp platform, during the high stand sea level. Towards the extreme northeast part of this basin, the Ain Asker succession spanned the entire Albian stage, belonged to the deeper part of the Tunisian trough, probably below the OMZ limits, and therefore scarcely recorded these events.

Keywords: Albian, Organic matter, Oceanic Anoxic Events, Biomarkers, Tectonics, Tunisia.

1. Introduction

The Albian age was characterized by an exceedingly warm climate, a small equator-to-pole temperature gradient, a lack of ice sheets, a high sea level, and significant tectonic and volcanic activity (Schlanger and Jenkyns, 1976; Jenkyns, 1980; Fenner, 2001; Leckie et al., 2002). Oceanic Anoxic Events (OAEs) occurred during the Albian and caused the expansion of extremely oxygen-poor waters in the oceans (Jenkyns, 2010). The most widely admitted causal factor was an expanded and intensified oxygen minimum zone (OMZ) of the ocean that enhanced the accumulation of marine organic-rich sediment (Jenkyns, 2010). Sea level rising and intense marine productivity are associated with OAEs (Erbacher et al., 1996). These episodic events resulted in the mass extinction of benthic fauna that was probably closely related to anoxia (oxygen-poor conditions) and oxidoreduction conditions of the water column in the ocean. Furthermore, black shales of the Early Albian Paquier event (OAE1b) in the Vocontian Basin (Br  h  ret, 1985) are dominated by isoprenoid biomarkers of archaea as a major constituent of plankton (Kuypers et al., 2002; Tsikos et al., 2004). Black shales formed during OAEs are generally accompanied by an increase in radiolarian silica (Racki and Cordey, 2000; Jenkyns, 2003).

In central and northern Tunisia, the Albian to Cenomanian successions are represented by cyclic alternations of thinly laminated carbonates with marls and shales that form the Fahdene Formation (Buroillet, 1956). Tectonically, the Early Jurassic rifting phase in the southern Tethyan margin (Northern-central Tunisia; Boughdiri et al., 2007) was characterized during this period by the reactivation of the NE-SW to E-W ante-Latest Albian deep-seated structural accidents associated with evaporite Diapir movements and local tilted blocks (Buroillet and Ellouz, 1986; Martinez et al., 1991; Zouari et al., 1999). The resulting basin was filled by organic-rich deposits, in association with a sea level rise and hyper silicious conditions, features

that typify the Tethys Realm during Albian-Cenomanian times (Racki and Cordey, 2000). During the Albian age, the structural context of Northern-central Tunisia was closely related to a general extensional regime that characterized the Tunisian margin. For example, at the Aptian-Albian transition, the tectonic regime was dominated by SW-NE extensive and/or transtensive deformation (Rigane et al., 2010). Thus, the sedimentation of the Fahdene Formation was interpreted to have been controlled by NE-SW, and E-W extensional horst and graben systems associated with synsedimentary normal faults (Rigane et al., 2010).

The reconstruction of the depositional environments by means of planktonic foraminifera (Chihaoui et al., 2010) suggests relatively deep-marine conditions corresponding probably to the outer part of a horst and graben ramp. Geochemical and biostratigraphic investigations showed that the general Albian OAE was recorded in the Albian pelagic sequences of north-western Tunisia (Ben Fadhel et al. 2010), and can be correlated with time-equivalent series of Northern Tethyan margins. Ben Fadhel et al. (2010) also stated that the early Late Albian successions from Northwest Tunisia were characterized by intensive radiolarian productivity as evidenced by silica enrichment induced by hydrothermalism of halokinetic origin.

The present study combines sedimentological analysis (carbonate analysis and microfacies description of thin sections), Rock-Eval pyrolysis (organic matter richness, type and thermal maturity) and molecular biomarker records of samples recovered from the bottom of the Fahdene Formation. Four outcrop sections (Fig. 1) were selected recording different structural domains from: the "graben" zone (Jebel El Hamra: HM) North-west of Kasserine, the diapir zone (Koudiat Berkouchia: BRK and Jebel Ghazouane: GZ), and the Tunisian trough (Ain Asker: ASK). Our aims were to evaluate the paleogeographic and structural controls on Albian OM deposition and therefore the record of the oceanic anoxic events (OAE1b and/or 1c) in this southern Tethys margin.

2. Stratigraphic and structural settings

The Fahdene Formation (Albian-Cenomanian) was defined by Burollet in 1956 at the Oued Bahloul Anticline, Southeast of Makthar in central Tunisia. It consists of very thick series of marine sediments: gray to black marls and shales, with a few beds of limestones or marly limestones interbedded that overlie the limestone dominated Serdj Formation of Aptian age (Burollet, 1956; M'Rabet, 1981; Burollet et al., 1983; M'Rabet et al., 1995). Outside the type locality, the Fahdene Formation overlays the sandstone and clays of the Hameïma Formation (Late Gargasian-Early Clansayesian) in the localities of Slata, Hameïma, Boulahneche, and Djerissa (Zghal, 1994; Zghal et al., 1997; Chihaoui, 2009).

In its type locality (Fig. 1), the Fahdene Formation has been divided (Burollet, 1956) into five members: (1) The Lower Shale Member of Lower Albian age; (2) the Allam Limestone Member of Middle Albian age; (3) the Middle Shale Member of Late Albian age; (4) the Mouelha Limestone Member of Late Albian age and (5) the Upper Shale Member of latest Albian age. The stratigraphy of the Fahdene Formation was revised by Zghal (1994) and divided into only four members. Members 1, 2 and 3 of Zghal (1994) match members 1 to 3 of Burollet (1956), whereas members four and five of Burollet (1956) were grouped together into a fourth member (Zghal, 1994). Nonetheless, the problem of a global definition the Aptian-Albian boundary remains a subject of disagreement (Casey et al., 1998; Kennedy et al., 2000; Hancock, 2001; Mutterlose et al., 2003). Many biostratigraphic studies conducted on the Serdj and Fahdene formations (Bismuth et al., 1981; El Euch, 1993; Zghal et al., 1997; Memmi, 1999; Ben Haj Ali, 2005; Lehmann et al., 2009; Chihaoui et al., 2010; Ben Fadhel et al., 2014) have concluded that the Lower to Middle Albian pelagic successions are diachronic and affected by gaps. For example, El Euch (1993) attributed the series between the Serdj Formation and the Lower shale Member of the Fahdene Formation, in Jebel Hamra, to the Late Aptian-lower Albian.

The distribution of the Cretaceous sedimentary basins, in Tunisia, was controlled by major structural trends (Fig. 1). Northwest-Southeast and Northeast-Southwest trending faults controlled the position and initiation of the sedimentary basins that were cut by North-South and East-West trending transverse faults (Sainfeld 1951; Burollet, 1956; Burollet and Sainfeld, 1956; Chikhaoui, 1988; Chikhaoui et al., 2002). The reactivation of the main faults associated with these structural trends and the Triassic diapirism has resulted in a complex tectonic history. Consequently, the structural evolution of this region has been controlled by successive extensional and compressive tectonic events that also involved halokinesis (Perthuisot 1978; Ben Ayed 1986, 1994; Boukadi 1996; Perthuisot et al., 1998; Kadri and Ben Haj Ali, 1999; Chikhaoui et al., 2002).

The most important episode of Triassic evaporitic materials ascension (Fig. 1) took place during Aptian times (Snoke et al., 1988) and played an important role in the tectono-sedimentary evolution of Tunisia (Boukadi et al., 1992). The horst and grabens created along extensional zones were disturbed by Triassic evaporitic deposits ascension that uplifted salt domes and tilted blocks. This configuration stimulated differential subsidence due to salt withdrawal until Cenomanian times in Central Tunisia (Soyer and Tricart, 1987). Accordingly, the Cretaceous basins in the "Tunisian Trough" matched the Triassic paleo-structures (Bismuth et al., 1982; Boltenhagen, 1985a, b; Bouaziz et al., 2002; Rigane et al., 2005).

The four Fahdene Formation outcrops sampled in this study were chosen from different paleogeographic and structural settings. The Koudiat Berkouchia (BRK) outcrop section belongs to the Mellegue "paleo-graben" (Chikhaoui et al., 1991), bordered by two structural highs: the Koumine to the West and the Nebeur to the East. This region is located at the South-Western extremity of the diapir zone (Perthuisot, 1978; Hatira et al., 2000). It is represented by complex tectonic structures related to the Triassic evaporitic deposits ascension movements and strike-slip faults (Chikhaoui et al., 1991, 2002; Chikhaoui and Turki, 1996; Memmi, 1999) (Fig

1). According to these authors, the structural style complexity was the result of the extensive tectonic movement, which led to the extrusion of Triassic evaporites during the Aptian-Albian period. Consequently, movements of halokinetic and tilted blocks are responsible for the horst and graben structures (Chikhaoui et al., 1991). The Jebel Ghazouane outcrop section (GZ; Fig 1) is located about 26 km North-East from the BRK outcrop. Structurally, this locality belongs to the inter-diapiric zone where thick and complete series of marls were deposited in rim-synclines of the Teboursouk area between the diapirs of Jebel Thibar and Jebel Chehid (Fig. 1) (Perthuisot, 1978; Perthuisot, 1981; Hatira et al., 2000). The Ain Asker (ASK) outcrop section is located in the Northeast part of the Diapir zone (Fig. 1). It belongs to the Eastern part of the subsiding basin of the “Tunisian Trough” (Bolze et al., 1952; Jauzein, 1967; Burollet, 1973). The Jebel El Hamra (HM, Fig 1), an Aptian-Albian age outcrop series, is located in central-Western Tunisia, Northwest of Kasserine and corresponds to the “graben zone”. The Aptian recifal limestone outcrops are weathering resistant structures that dominate the relief in the Tajerouine area. The HM locality is bounded to the East by the NW transverse Graben of Kasserine and Foussana (GKF). The Kasserine region, in the central Tunisia platform (Martinez et al., 1991), is characterized by a major N110° trending fault on the edge of which the grabens of Kasserine and Foussana developed during the Miocene to Quaternary times (Philip et al., 1987). Many authors (Bismuth et al., 1981; Bismuth al., 1982; Bismuth et Mahjoub, 1985; Boltenhagen, 1985a, b; Batik et al. 1987) consider that the fault of Kasserine, like those of Gafsa (GF) to the South, and Kalaat Jerda-Sbiba (K.J-S.F.) to the North, was continually active, in particular during Albian-Aptian times and before the latest Albian transgression. According to Boltenhagen (1985a,b), at the end of the Aptian, the Kasserine Fault (K.F., Fig. 1) corresponded to the Southern limit of a large structural saddle basin. This structural style, which bends to the South, ensures the transition between the platform of central Tunisia and the “Tunisian Trough”, as a stepped substratum with an elevated external ridge. Rigane et al. (2010)

underlined, at Jebel Hamra, the prominent role of the paleo-structures acquired during the evolution of the South-Tethyan basin and their influence of the present-day geometry of the fold belt.

3. Materials and methods

Samples of the Fahdene Formation were recovered from four outcrop sections covering different structural domains in central and northern Tunisia (Fig. 1). From the bottom to the top of each outcrop section samples were selected from non-weathered black bituminous laminated marl and limestone levels.

3.1 Samples

The Fahdene Formation outcropping at BRK (Fig. 2) directly underlies the chaotic Triassic evaporites (gypsum). This formation is represented at the contact of these evaporites by greenish, glauconitic-rich clay. This succession is overlain by alternations of black coloured limestones and nodular marls and yellow to reddish limestone beds. Some limestone beds (e.g. BRK9) show bioturbation and syneresis cracks that represent the lower part of the “Allam Member” defined by Zghal et al. (1997) and Chihaoui et al. (2010).

The Fahdene Formation outcropping at GZ (Fig. 2) is represented by an alternation of limestone and marl levels. Above the Aptian dolostones, the ca. 200m thick clayey limestone and fissile green clays (samples GZ1 to GZ7) have been assigned to the Aptian-Albian age (Fakhraoui et al., 1998). This succession is overlain by alternations of grey marly limestones and limestones that have been attributed to the Latest Albian (samples GZ8 to GZ12) (Fakhraoui et al., 1998).

The Albian Fahdene Formation at HM (Fig. 2) consists of alternations of grey marls, marly limestone and limestone. These alternations have exposed ammonites, belemnite and rudists such as those from samples HM1 to HM17 and from HM24 to HM28.

The sampled Fahdene Formation outcropping at the ASK cross-section (Fig. 2) is represented by ca. 500 m thick of regular alternations of grey marls and marly limestone intercalations and argillaceous limestone. This section covers probably all the Albian to the onset of the Cenomanian. The brown marls from samples ASK16 and ASK17 showed abundant pyrite concretion.

3.2 Carbonate and Organic geochemical analysis

Rock samples were powdered in an agate mortar. Powdered-dry samples were used for Rock-Eval pyrolysis, bitumen extraction and carbonate content analysis. Carbonate content expressed as CaCO₃ (%) was determined by using the Bernard Calcimeter method (Müller and Gastner, 1971; Vatan, 1967). Weighed samples were reacted with 3N HCl. The volume of CO₂ released, after reaction with HCl, was measured and compared with the volume released from pure calcite. Thin sections from indurated limestone samples were prepared for optical microscopic observation and microfacies description.

Bulk organic geochemical parameters were determined using Rock-Eval pyrolysis under standard conditions with a Rock-Eval 6 apparatus (Vinci Technologies; Espitalié et al., 1985; Lafargue et al., 1998; Behar et al., 2001). The Rock-Eval pyrolysis method first consists in heating a small sample (~100 mg) using a programmed temperature in a pyrolysis oven under an inert atmosphere (helium) to release the Pyrolysed Carbon (PC) as hydrocarbonaceous compounds, CO and CO₂. Then, the residual sample is submitted to a second heating phase in a combustion oven, under oxidising conditions to release the Residual Carbon (RC) as CO and CO₂. This method enables quantification of: (i) the free hydrocarbons pre-existing in the rock (S1; mg hydrocarbons (HC).g⁻¹ rock), (ii) the hydrocarbons (S2; mg HC.g⁻¹ rock) that are volatilized during the cracking of OM (kerogen) in the sample and represent the hydrocarbon generative potential, (iii) the CO₂ and CO (S3 as mg of CO and CO₂.g⁻¹ rock) released during pyrolysis at temperatures <390°C, (iv) the temperature of maximum production of

hydrocarbonaceous compounds during pyrolysis at the top of the S2 peak (Tmax), and (v) the Total Organic Carbon (TOC). The type and maturity of organic matter in petroleum source rocks can be characterized from Rock Eval pyrolysis data using the following parameters: (i) the hydrogen index ($HI = S_2 \times 100 / TOC$, mg HC.g⁻¹ TOC), (ii) the oxygen index ($OI = S_3 \times 100 / TOC$, mg CO₂.g⁻¹ TOC), (iii) the production index ($PI = S_1 / [S_1 + S_2]$). PI is used to characterize the evolution level of the organic matter.

The hydrocarbons were then extracted from powdered samples using an organic solvent (CHCl₃, 200 ml/30 g sample). After vacuum filtration and solvent evaporation, we obtained the bitumen or the extractable organic matter (EOM) (Durand, 1993; Hunt, 1995). Then, the bitumen was fractionated using liquid chromatography with a single column packed with activated silica gel, into saturated hydrocarbons (SHC), aromatic hydrocarbons (AHC) and polar compounds (NSO). The column was eluted with hexane to obtain the SHC and with hexane/CHCl₃ (65:35, v/v) to obtain the AHC. The NSO fraction was retained at the top of the column. SHC were analysed using gas chromatography-mass spectrometry (GC-MS) to investigate the distribution of *n*-alkanes, acyclic isoprenoids (pristane and phytane) and cyclic alkanes (steranes and terpanes). GC-MS was performed with an HP 6890 series chromatograph equipped with a CPSIL 5CB fused silica column (25 m * 0.25 mm i.d.) connected to a Hewlett Packard 5973 mass selective detector (MSD) by a 2 m capillary interface at 250°C. After splitless injection for 1 min., the oven temperature was increased from 100°C to 310°C at 3°C min⁻¹. The mass selective detector (MSD) was operated using the following conditions: 220°C; electron energy 70eV; scan range m/z 50–450; 1 s per scan. The carrier gas was He. Peak height and relative intensities were measured from mass chromatograms of characteristic ions for alkanes (m/z 57), steranes (m/z 217) and terpanes (m/z 191).

4. Results

4.1. Rock-Eval pyrolysis

The Rock-Eval analysis results are reported in Table 1 and Figure 3. TOC values in the analysed samples varied widely in the range of 0.01 to 3.43%. The highest TOC values registered were 3.43%, 2.90% and 1.32% in samples BRK5, GZ12 and HM30, respectively. The ASK samples showed a low OM content which did not exceed 0.89% of TOC. The free hydrocarbons S1 were very low (<0.9 mg HC/g Rock) in all the samples analysed. The potential hydrocarbons S2 varied widely, along or between the outcrop sections. The samples BRK12, GZ12 and HM30 registered the highest values of 4.74, 2.79 and 2.3 mg HC.g⁻¹ Rock, respectively. As revealed from the OM contents, the ASK samples had very low S2 values (<0.6 mg HC.g⁻¹Rock). The wide range of variation of the S1 and S2 affected the PI. The PI values varied considerably, between 0.03 and 0.81. The oxygen compounds S3 were very low (<1 mg CO₂g⁻¹ Rock) in all the samples, but three samples (HM12, HM15 and ASK42) registered 2.13, 2.19 and 1.62 mg CO₂g⁻¹ Rock, respectively.

The HI expresses the amount of hydrogen released from OM during pyrolysis. The HI values varied widely and ranged from 3 to 533 mg HC.g⁻¹ TOC (Table 1). The OI represents the oxygen released from the OM during pyrolysis. OI values also varied widely from very low (ca. OI<100 mg CO₂/g TOC) to very high values (OI>> 300 mg CO₂/g TOC). HI and OI values can be used to determine the type of OM and its hydrocarbon proneness in a Pseudo-van Krevelen diagram (Tissot and Welte, 1984; Peters, 1986). Tmax values ranged from 439 to 453°C.

4.2. Carbonate and microfacies analysis

The carbonate content varied widely through samples from ca. 1% to ca. 98% of CaCO₃ (Table 1). These values are consistent with the field lithologies that consist of marls, marl-carbonates and carbonates. However, these values varied irregularly either in each outcrop or between the outcrop sections. Carbonate content in samples BRK varied between ca. 50 and 82% of CaCO₃. The carbonate content of the GZ samples ranged from 0.93% to 89% of CaCO₃.

HM samples registered a CaCO₃ content ranging from 31% to 97%. Finally, the ASK samples contained between 58 and 98% of CaCO₃.

Thin section microscopic observations of representative indurated limestone beds samples from the BRK and GZ outcrops (Fig. 4) exhibited a cyclic alternation of fine dark (organic-rich) and light laminated wackestones (Folk, 1962; Dunham, 1962). These observations revealed abundant radiolarians, calcispheres and other bioclasts. Furthermore, microfractures (stylolites) were filled with brown coloured bitumen.

4.3. Bitumen richness and composition

Nine representative samples from the BRK (samples BRK11, BRK12 and BRK13), GZ (samples GZ8, GZ9 and GZ10) and HM (samples HM15, HM20 and HM30) outcrops were selected for biomarker analysis. The extractable organic matter (EOM) yield ranged from 0.12 to 1.86 mg HC.g⁻¹ rock (Table 2). These values represent ca. 0.5 to 19.2% of the TOC. The SHC in BRK, GZ and HM samples represented 13 to 76% of the EOM. Generally, the hydrocarbon fraction (SHC+AHC) predominated in the EOM of the BRK and GZ samples, whereas NSO compounds represented the majority (NSO>59% EOM) of the EOM of the HM samples.

4.4. The Fahdene Formation molecular signature

4.4.1. n-alkanes and acyclic isoprenoids

The gas chromatography-mass spectrometry (GC-MS) analyses were carried out on the SHC fraction of representative samples from the HM, BRK and GZ cross-sections. The GC-MS chromatograms (Total Ion Current: TIC) showed that most of the samples had *n*-alkane bimodal distributions maximizing in the *n*-C₁₇ to the *n*-C₂₀ and *n*-C₂₄/*n*-C₂₅ range () with a regular decrease towards higher carbon numbers (Fig 4). Acyclic isoprenoids, especially norpristane (*i*-C₁₆), pristane (Pr; *i*-C₁₉) and phytane (Ph; *i*-C₂₀) were relatively abundant in the SHC fraction. Pristane/phytane (Pr/Ph, Table 2) ratio values varied between ca. 1 (sample HM30) to

2.4 (sample BRK12). The Pr/n-C₁₇ ratio values (Table 2) varied from 0.57 to 1.13, whereas the Ph/n-C₁₈ ratio values were very low and did not exceed 0.95 (Table 2). We tentatively identified squalane in high relative proportions in the BRK samples (Fig. 5).

4.4.2. Hopanes and steranes

The representative samples from each outcrop were analyzed by GC-MS for their tricyclic and pentacyclic triterpane (m/z 191) and sterane (m/z 217) distributions (Fig. 6). The tricyclic terpane series (Fig. 6; Table 2) extended from the C23 up to C26. The hopane series extended from the C27 to the C35 and were dominated by 17 α (H),21 β (H) hopane [*abbreviated* C30H; $\alpha\beta$ C30H] and 17 α (H),21 β (H) nor-hopane [*abbreviated* C29H; $\alpha\beta$ C29H]. None of the samples showed any $\beta\beta$ and $\beta\alpha$ hopanes (moretanes). All the samples had a high C30H/C29H ratio (Table 4) ranging from 1.07 to 3.63. The homohopane series decreased regularly from C31 to C35 in all samples (Fig 6). Furthermore, the m/z 191 chromatograms of the BRK12, GZ10, and HM30 samples (Fig. 6) showed high extended tricyclic terpanes from C19 to C29 (Ekweozor and Strausz, 1982), subordinate C24 tetracyclic terpanes, high 18 α (H)-trisorneohopane (C27Ts) compared to 17 α (H)-trisnorhopane (C27Tm) and relatively abundant C29Ts. All the samples exhibited a high peak of the C30 17 α (H)-diahopane. As shown in Table 4 the Ts/Tm ratio ranged from 0.77 (sample HM3) to 3.39 (samples HM30), corresponding to Ts/(Ts+Tm) ratio values ranging from 0.43 to 0.77. The C29Ts/C29H ratio varied between 0.35 (samples HM3) and 1.93 (sample BRK12). All the samples but one (HM3) had a high (>1) C30 17 α (H)-diahopane/C29H ratio ranging from 0.47 to 1.46 and a relatively high Diahopane Index (DI = C30 17 α (H)-diahopane / (C30 17 α (H)-diahopane + C30H) in the range 0.25-0.53 (Table 4).

The sterane (m/z 217) fingerprints of the representative samples from the HM, BRK, and GZ cross-sections are displayed in Fig 6. All the chromatograms show regular and rearranged steranes extending from C27 to C30 (Fig. 6; Table 3). Very low amounts of C30 steranes, much lower than those of the C27 to C29, occurred in all of the samples. The C27

diasteranes were the prominent compounds in all samples compared to the regular C27 to C29 steranes (Fig. 6).

All samples contained small amounts of the $14\alpha(\text{H}),17\alpha(\text{H})\text{-}20\text{RC}28$ sterane ($\alpha\alpha 20\text{R}$ C28) compared to its C27 and C29 homologs. The $14\alpha(\text{H}),17\alpha(\text{H})\text{-}20\text{RC}27$ ($\alpha\alpha 20\text{RC}27$) predominates in the regular sterane series (i.e. $\alpha\alpha 20\text{R}$ C27, C28, and C29 regular steranes) (Fig. 6; Table 3). By contrast, the $14\beta(\text{H}),17\beta(\text{H})\text{-}20\text{R}$ ($\beta\beta 20\text{R}$) and $14\beta(\text{H}),17\beta(\text{H})\text{-}20\text{S}$ ($\beta\beta 20\text{S}$) isomers dominate over the $\alpha\alpha 20\text{R}$ and 20S compounds for the C29 sterane. All samples show high ratios of the $29\alpha\text{S}/29\alpha(\text{S}+\text{R})$, $29\beta\text{S}/29\beta(\text{S}+\text{R})$ and $29\beta(\text{S}+\text{R})/[(29\beta(\text{S}+\text{R})+29\alpha(\text{S}+\text{R}))]$ ranging from 0.45-0.50, 0.44-0.55 and 0.45-0.62, respectively (Table 4). The $27\text{DS}/27\text{D}(\text{S}+\text{R})$ ratio vary from 0.59 to 0.68.

5. Discussion

5.1. The Albian OM sedimentation, richness, origin, and thermal maturity

A source rock can be broadly defined as any fine-grained, organic-rich rock that is able to generate petroleum when it is sufficiently exposed to heat and pressure. Its petroleum-generating potential is directly related to its OM richness and thermal maturity. TOC is an important parameter in the evaluation of petroleum source rocks. Tissot and Welte (1984) and Peters and Casa (1994) proposed a scale for the assessment of the petroleum potential of source rocks, based on TOC, S1 and S2 values. In our samples, the TOC, S1, S2, and S3 parameters varied widely between samples and outcrop localities (Table 1). These variations can be interpreted as dependent on the origin, the preservation and the paleo-environmental conditions in which sediments were deposited. The TOC values showed discrepancies either vertically across one outcrop section or regionally between outcrops. Therefore, the highest TOC values, which range between 0.5 and 3.43%, were registered in the BRK and GZ localities. The TOC values were moderate and did not exceed 0.9% in the HM and ASK outcrops, to the South and to the North-East of the studied area, respectively.

The carbonate content of the studied outcrops (Table 1) varied between ca. 1% and 98% of CaCO₃. These highly variable values are in agreement with the observed rhythmic alternation of limestone, marls and marly limestone. The CaCO₃ and TOC contents did not show a linear correlation in the studied sections. In order to identify the main factor of differences in the relative carbonate and organic matter content in these limestones and marls, we used a TOC-CaCO₃ diagram (Ricken, 1993). The location of the samples on this diagram (Fig. 7) agreed with a moderately bioturbated-dysaerobic deposition environment. Consequently, this variability in TOC contents could be explained by the effect of bioturbation on the sediments allowing their oxygenation and therefore the oxidation of OM.

The microscopic observations of representative limestone bed samples from the BRK and GZ outcrops (Fig. 3) exhibited a cyclic alternation of fine dark (organic-rich) and light laminated wackestone limestone (Folk, 1962; Dunham, 1962). Microscopic observation revealed that the wackestones are radiolarian rich; calcispheres and other bioclasts are also common components. Microfractures (stylolites) are filled with bitumen. These wackestones are indicative of an environment changing from dysoxic to anoxic conditions on the sea floor (Leckie et al., 2002).

The Fahdene Formation outcrops studied show an alternation of marls and limestone beds. These alternations are interpreted as orbitally-forced (Herbert and Fischer, 1986; Sageman et al., 1998; Fiet et al., 2001; Galeotti et al., 2003; Grippo et al., 2004; Tyszka, 2009). Ben Fadhel et al. (2011) grouped these alternations in a package of laminated, grey to black marl and black-coloured limestone beds, which lie within the Mid-Albian *breggiensis* planktonic foraminifer biozone. The black shale layers interbedded within these alternations bearing moderately abundant radiolarian, in the BRK outcrop section, span the entire uppermost part of the *breggiensis* biozone (Ben Fadhel et al., 2011). Arthur et al. (1990) and Bralower et al. (1993) pointed out that the OAE1c straddles the entire *breggiensis* biozone. In our present study and by lithologic correlation with the work of Ben Fadhel et al. (2011), the BRK samples

cover this zone from the upper part of the Allam Member (middle Albian age) to the Mouelha Member (Latest Albian age) (Zghal et al., 1997). Elsewhere, at the HM, GZ and ASK outcrop sections, the sampled successions probably cover the Early to Late Albian. In our study, the part of the HM outcrop between samples HM1 to HM10 is attributed to the Late Aptian-lower Albian age (Buroillet, 1956; El Euch, 1993; Zghal, 1994; Chihaoui, 2009; Jaillard et al., 2013). In this outcrop section, this stratigraphic interval can be considered synonymous of the OAE1b. Accordingly, the organic-rich levels registered in the HM outcrop samples probably correspond to the Early and Late Albian, OAE1b and OAE1c, respectively (Arthur and Silva, 1982; Br  h  ret et al., 1986; Silva et al., 1989; Bralower et al., 1993, 1999; Leckie et al., 2002; Wagner et al., 2008), whereas the organic rich samples of the BRK and GZ outcrops correlate with the Late Albian OAE1c (Pratt and King, 1986; Bralower et al., 1993; Coccioni and Galeotti, 1993; Haig and Lynch, 1993; Erbacher et al., 1996). In the ASK outcrop, which represents the deepest facies of the Albian successions, scarcely recorded these events.

HI and OI are directly related to the atomic hydrogen to carbon ratio (H/C) and oxygen to carbon ratio (O/C), respectively (Espitali   et al., 1985; Lafargue et al., 1998; Behar et al., 2001). Marine organisms and algae, in general, are composed of lipid and protein-rich organic matter with a higher H/C than that of the carbohydrate-rich constituents of land plants. HI values typically range from ~100 mg HC.g⁻¹ TOC to 600 mg HC.g⁻¹ TOC in geological samples. In our study, the HI and OI values (Table 1) showed that the OM in the Fahdene Formation is poorly hydrogenated (HI < 300 mg HC.g⁻¹ TOC) and very variably oxygenated. Only one sample from the GZ locality (GZ9) has preserved a good HI (533 mg HC.g⁻¹ TOC). Obviously, the OI values of the ASK samples and some levels from the HM locality are much higher than those of the other localities (OI > 100 mg CO₂.g⁻¹ TOC). A modified Van Krevelen diagram that plots HI versus OI values (Fig. 8a) reveals that the organic matter preserved in this Formation varies from oil-prone (Type II) kerogen to oil and gas-prone (Type II/III) kerogen with a liquid and

gaseous hydrocarbon production potential. Furthermore, this indicates a poorer residual generation potential of the OM and thus, a higher organic maturity. The higher organic maturity is supported by the T_{max} values ($440^{\circ}\text{C} < T_{max} < 453^{\circ}\text{C}$, Table 1, Fig. 8b) which suggest that the OM of the Fahdene Formation in the studied outcrops has reached the thermal maturity for hydrocarbon generation (oil window, Tissot and Welte, 1978; Durand, 1980; Bordenave, 1993; Vandenbroucke et al., 1993). Other evidence of oil production is the presence of bitumen in the microfractures, as described earlier.

Moreover, most of the samples can be classified as fair to good petroleum source rocks (Fig. 8c). The overall distribution of the generative potential parameters, S1 and S2, and thus the PI, shows a broad range of values (Fig. 8d). These values agree with a thermal maturity corresponding to the oil generation window (Espitalié et al., 1985; Peters, 1986). However, some samples could have undergone either primary migration ($PI < 0.1$) or impregnation by non-indigenous hydrocarbons ($PI > 0.4$).

5.2. Molecular precursors and environmental signature

Distributions of *n*-alkanes in source rocks can be used to indicate the OM source (Tissot et al., 1977; Peters and Moldowan, 1993). C_{15} to C_{35} *n*-alkanes are present in all GC traces (Fig. 5). For the representative samples, the distributions of *n*-alkanes show distributions maximizing first in the $n\text{-}C_{17}$ to $n\text{-}C_{20}$ range and second in the $n\text{-}C_{22}$ to $n\text{-}C_{30}$ range. The second modal distribution of $C_{22}\text{-}C_{30}$ *n*-alkanes shows a slight even-carbon-number preference maximized at the C_{24} or C_{26} homologues. This distribution is generally representative of autochthonous aquatic alga and cyanobacteria (Han and Calvin, 1969; Gelpi et al., 1970; Blumer et al., 1971; Brassell et al., 1978; Chaler et al., 2005). We tentatively identified squalane in high relative proportions in the BRK samples (Fig. 5). This isoprenoid has been identified in very reducing marine conditions and is a marker of methanogenic bacteria with other Archaeobacteria (Holzer et al. 1979; Brassell et al. 1981). Furthermore, relatively intense peaks of the regular isoprenoids

norpristane (*i*-C₁₆), pristane (Pr: *i*-C₁₉) and phytane (Ph: *i*-C₂₀) were found in all samples. These isoprenoids that are primarily derived from the phytol side chain of the chlorophyll molecule (Volkman and Maxwell, 1986) can be used as biomarkers for phytoplankton as well as the oxidoreduction conditions of the environment.

Variations in the redox conditions within the Albian basin are reflected by the fluctuation of isoprenoid concentrations. The Pr/Ph ratio (Table 2) is commonly used to evaluate the redox conditions that prevailed at time of deposition (Didyk et al., 1978). Pr/Ph values ranging from 0.47 to 2.91 in our samples indicate that the organic matter was deposited under a suboxic environment. The isoprenoid/*n*-alkane ratios [Pr/*n*-C₁₇ and Ph/*n*-C₁₈] indicate the source rock facies and the thermal maturity (Hunt, 1995). These ratios decrease with maturation as a result of thermal cracking that produces *n*-alkanes and progressively reduces Pr and Ph (Tissot et al., 1971). On the other hand, both Pr/*n*-C₁₇ and Ph/*n*-C₁₈ ratios increase progressively with *n*-alkane biodegradation (Peters et al., 1999). Our results show that low Pr/*n*-C₁₇ (<1.13) and Pr/*n*-C₁₈ (<1) ratios are consistent with an advanced thermal maturity of the OM at these localities (Fig. 9).

Hopanes derive from prokaryotic organisms (Ourisson et al., 1979) such as bacteria and microbes. The nearly identical terpenoid distribution of samples BRK12, GZ10 and HM3, HM16 and HM30 indicates a common origin. The similarity among the *m/z* 191 fragmentograms is corroborated by the obvious abundance of tricyclic terpanes, C₂₄ tetracyclic terpanes, the high C₂₇Ts and C₂₉Ts, together with the predominance of C₃₀ 17 α (H)-diahopane, C₃₀ Hopane and the regular decrease of the C₃₁ to C₃₅ homohopanes (Fig. 6).

These results suggest marine input of OM deposited under suboxic conditions (Peters and Moldowan, 1991). The high abundance of the C₂₃ associated with the C₂₁ tricyclic terpanes with respect to the other tricyclic homologs (Fig. 6) is also indicative of OM deposited in a carbonate marine environment under reducing conditions (Waples and Machihara, 1991).

The relative abundance of C30 17 α (H)-diahopane has been used for geochemical studies of petroleum source rocks and oils (Farrimond and Teln \ddot{a} s, 1996; Li et al., 2009). Its presence in coals and terrigenous oils has been interpreted as a terrestrial marker (Volkman et al., 1983; Philp and Gilbert, 1986). However, Moldowan et al. (1991) suggested that the C30 17 α (H)-diahopane may be of bacterial origin, with rearrangement occurring during diagenesis by clay-mediated acidic catalysis. Thus, its occurrence in oils or rock extracts may indicate bacterial input to sediments containing clay deposited under oxic or suboxic conditions. Furthermore, the 17 α (H)-diahopane series should be more stable than those of the 17 α (H)-diahopane with increasing thermal maturity (Kolaczowska et al., 1990; Moldowan et al., 1991). Li et al. (2009) stated that the concentrations of 17 α (H)-diahopane in oils or rock extracts is dependent on thermal maturity as well as lithology and environment. These authors defined a Diahopane index (DI) $[C30\ 17\alpha(H)\text{-diahopane}/(C30\ 17\alpha(H)\text{-diahopane} + C30\ 17\alpha(H)\text{-hopane})]$ that increases with increasing burial depth of the source rocks and can be used to indicate the top of the oil generation window. In the present study, the occurrence of C30 17 α (H)-diahopane in the extract of the Albian samples, associated with an elevated H30/H29 ratio (Table 4), tricyclic terpanes, high C27Ts and C29Ts should be consistent with both bacterial origin and thermal maturity of the organic matter deposited with high clastic contents under reducing conditions (Peters et al., 2005). Furthermore, we suggest that the C30 17 α (H)-diahopane could be the typical biological marker of the Albian OM in our study. To our knowledge, this compound has never been detected in other Tunisian petroleum source rocks such as the Cenomanian-Turonian Bahloul Formation (OAE2) (Affouri et al., 2013) or the Ypresian Bou Dabbous Formation (Affouri and Montacer, 2014).

The quantitative thermal maturity of the OM can be evaluated by using the Ts/ Ts+Tm ratio. However, lithology and depositional environments can also affect this ratio (Cornford et al., 1983; Schou et al., 1985; Moldowan et al., 1986). In our samples, where facies are similar,

the relatively high ratio of T_s/T_{s+T_m} (Table 4) associated with high $C_{29}T_s$ is assumed to reflect the advanced thermal maturity of the Albian OM. This interpretation agrees with the other maturity parameters including Rock-Eval pyrolysis (e.g. T_{max} and PI) as well as *n*-alkane distribution and abundant tricyclic terpanes.

The distribution of the C_{27} , C_{28} and C_{29} ($\alpha\alpha 20R$) regular steranes can be used as a facies parameter (Huang and Meinschein 1979; Shanmugam, 1985; Killips and Killips, 2005). Marine steranes are derived from the sterols of eukaryotes such as diatoms, dinoflagellates, zooplankton and higher plants (de Leeuw et al., 1989). In our samples, the regular steranes were dominated by the C_{27} steranes followed by the C_{29} steranes. This feature suggests a mixture of marine and terrestrial OM as revealed by the HI/OI (Fig. 8a) and HI/ T_{max} (Fig. 8b) cross plots.

Two alternative maturity parameters can be calculated from the sterane series (m/z 217; Fig. 6) on the basis of progressive isomerization with increasing temperature: the C_{29} ($14\alpha(H), 17\alpha(H)20S/(20S+20R)$) ratio $[29\alpha S/29\alpha(S+R)]$ and the C_{29} ($14\beta(H), 17\beta(H)(20S+20R)/(14\beta(H), 17\beta(H)(20S+20R)+14\alpha(H)17\alpha(H)(20S+20R))$ $[29\beta(S+R)/((29\beta(S+R)+29\alpha(S+R)))]$ ratio (Seifert and Moldowan, 1986). Both isomerization ratios increase with increasing maturity (Mackenzie et al., 1980). Usually, the $29\alpha S/29\alpha(S+R)$ ratio reaches an equilibrium value of 0.52-0.55 near the peak oil generation. The $29\beta(S+R)/((29\beta(S+R)+29\alpha(S+R))$ ratio is slower to reach an equilibrium value of ca. 0.70 around peak oil generation (Mackenzie et al., 1980; Peters and Moldowan, 1993; Ramón and Dzou, 1999). Peters and Moldowan, (1993) indicated that this ratio is independent of OM source input. In the present study, the analyzed samples display $29\alpha S/29\alpha(S+R)$ sterane ratios between 0.45 and 0.50, i.e. lower than the equilibrium values (0.52-0.55). All analyzed samples, except one (HM3), have $29\beta(S+R)/((29\beta(S+R)+29\alpha(S+R))$ ratios greater than 0.5 but have not reached the equilibrium values (0.67-0.71). Accordingly, these sterane maturity ratios

indicate that the OM of the Albian was at maturity levels corresponding to an early oil generation stage (Waples and Machihara, 1990; Peters and Moldowan, 1993). Furthermore, these samples show a high abundance of rearranged C27 diasteranes relative to regular steranes (Fig. 6). The relative abundance of diasteranes to steranes is commonly used to discriminate carbonate from clay-rich source rocks. Therefore, low diasteranes are consistent with anoxic clay-poor or carbonate-rich source rocks (Clark and Philp, 1989), whereas high proportions are usually associated with source rocks, containing abundant clays (Mello et al., 1988). The transformation of steranes to diasteranes during diagenesis is catalyzed by acidic sites on clays (Kirk and Shaw, 1975; Rubinstein et al., 1975; Sieskind et al., 1979; Grantham and Wakefield, 1988). Moreover, the diasteranes are stable and prevail at higher thermal maturity compared to the regular steranes (Hughes et al., 1985; Goodarzi et al., 1989). In our samples, the relatively high diasterane contents indicate both clay-rich source rock deposited under suboxic conditions and advanced thermal maturity.

5.3. Structural and paleogeographic control on Albian OAEs in Tunisia

Repetitive episodes of OM sedimentation are called oceanic anoxic events (OAE) and are mainly known during the Cretaceous (Schlanger and Jenkyns, 1976), Jurassic (Jenkyns, 1985) and Permian/Trias (Grice et al., 2005). The term "black shale" or "black clay" or "bituminous shale" refers to OM-rich formations. This term does not always reflect reality because the sediments deposited during a given period are not necessarily black, nor always of clay nature (Nzoussi-Mbassani, 2003). Several terms have been proposed to designate these OM-rich sediments. For example, the Lower Toarcian OM-rich clays from the Paris Basin in France are known as "schist cartons" or "bituminous schist" (Goy et al., 1979). In Germany, the sediments are called "Posidonienschiefer" while in England, they are called "Jet-Rock". Despite these different vocabularies, the concept of "black shale" remains to date the most widely used, given its genetic connotation. It should be noted that in the Tethyan domain, black

shales are preserved only in limited sub-basins which favour stagnation and depletion of deep-water oxygen (Baudin and Lachkar 1990; Baudin et al. 1990; Karakitsios, 1995). Several factors may have played a role in creating the specific environmental conditions that favoured the emergence of "OAE" in the Cretaceous. The most notable factor was an extremely hot climate, probably due to high concentrations of atmospheric CO₂ caused by increased volcanism (Arthur et al., 1985; Bice et al., 2006; Sinninghe Damsté et al., 2008; Turgeon and Creaser, 2008). The resulting warmer ocean waters probably resulted in halothermal (hot and salty) oceanic circulation rather than thermohaline circulation (Horne, 1999). As a result, these conditions result in lower deep water ventilation rates as hot waters contain less dissolved oxygen. In addition, the rapid expansion of the ocean floor, underwater volcanism and hydrothermal activity would have had a strong effect on ocean circulation (Hays and Pitman, 1973, Paytan et al., 2004).

The high concentrations of organic carbon in black shales were attributed to both the increased preservation of OM by the anoxic conditions of the water column, probably due to a slow circulation of ocean waters (Bralower and Thierstein 1987, Sinninghe Damsté and Köster, 1998), and increased primary organic productivity (POP) leading to higher rates of OM accumulation and preservation (Schlanger and Jenkyns, 1976). Accordingly, a high POP will also lead to increased oxygen demand for its mineralization, which could exceed the oxidation potential of the ocean and cause anoxia of the water column (Schlanger and Jenkyns, 1976; Kuypers et al. 2002).

The Early Cretaceous sedimentary basin of central Tunisia corresponds to an unstable platform with clastic deposits toward the South and open marine facies to the North (Marie et al., 1984; Dlala, 1999; Rigane et al., 2010). During the Upper Aptian, a major distensive tectonic phase known as the "Aptian Crisis", (Soyer and Tricart, 1987; Rigane et al., 2010) fragmented the northern edge of the Saharan Platform into tilted blocks (Bismuth et al., 1982;

Martinez et al., 1991). These tilted blocks controlled the palaeogeography of central Tunisia during Albian-Cenomanian times. Furthermore, the reactivation of faults limiting tilted blocks during the Aptian-Albian extension was associated with the ascension of Triassic evaporites. This tectono-sedimentary context controlled the distribution and preservation of organic-rich sequences from central-western to northern Tunisia. As concluded by Boltenhagen et al. (1985a, b) the Kasserine Fault corresponds to the southern limit of a large structural saddle basin that ensures the transition between the carbonate platform of central Tunisia (Zebbag Formation) towards the South and the deep basin of the “Tunisian Trough” (Fahdene Formation) towards the North. In this case, the HM locality can be said to represent the southern proximal limit of this basin. Further to the North, at the BRK locality, the Albian series was deposited in a subsiding graben structure (graben of Mellegue). The Triassic evaporite ascension no doubt played a role in this structural configuration and thus the distribution and preservation of OM in the Albian successions. This is the case of the GZ locality, where the Fahdene Formation was deposited in a high edged rim syncline basin positioned between Triassic Diapirs (Jebel Chehid and Jebel Thibar Dipirs). Northeast of the "Tunisian Trough", the ASK locality represent the distal and the deepest part of the Albian Basin. All these structural configurations, from South to the Northeast, have influenced the preservation of OM. Therefore, we suggest that OM richness resulted from the installation of an oxygen minimum zone (OMZ) at mid-depth of this basin, and that this OMZ was induced by a high POP, as revealed by the abundance of radiolarian especially around the Mellegue Graben, the GZ and ASK outcrops (Ben Fadhel et al., 2010, 2014). Therefore, the high OM content in these outcrop sections was promoted by both reduced terrigenous input during transgression phases and the spreading of upwelling-induced OMZ that invaded the upper slope and shelf with rising sea level (Schlanger et al., 1987; Arthur and Sageman, 1994). The structural framework as tilted blocks and halokinesis inherited from the Late Aptian tectonics and subsequent sea level rise

during the early Albian (Weissert and Lini, 1991; Weissert et al., 1998) played a major role in the record of the OAE1b and OAE1c in this southern Tethys margin. These events are comparable to the basal Albian Paquier, in the Vocontian Basin of Southeast France, and the lower Albian Leenhardt, events that have been attributed to elevated primary productivity (Br  h  ret, 1994; Erbacher et al., 1999, 2001). These findings suggest that multiple causes, including primary productivity, sea level changes, climatically driven organic carbon burial and structural framework, contributed to the Cretaceous OAE1b and OAE1c in Tunisia. All of these features are summarized in the schematic model presented in Figure 10. Consequently, we suggest that, when sea level rose during the Early Albian transgression (Weissert and Lini, 1991; Weissert et al., 1998), the invaded structurally low areas (grabens and inter-diapiric rim-synclines) represent very deep marine basins that were below the OMZ limits. Therefore, the installation of an OMZ layer at ocean mid-depth, in relationship with upwelling and high POP, coincided with half-graben high structures as well as around Triassic diapir horsts. An expanded and intensified OMZ is suggested to be the main factor controlling the Early Albian sedimentation in the HM and BRK sections. Both localities probably registered the OAE1b in the southern Tethys margin. The oscillations of the OMZ related to fluctuating POP and upwelling nutrient supplies as well as the continental material supply by intensive weathering and seasonal rainfall (Fenner, 2001) could explain both the facies alternations and the OM richness of the sediments. During the Latest Albian time, the OMZ covered a large domain of this relatively deep-marine outer part of a horst and graben basin (Chihaoui et al., 2010). Consequently, the HM locality coincides with the uppermost limit of the OMZ and therefore registers more fluctuations of OM preservation/non-preservation at this time. During this period, the BRK and ASK localities were in the basin depth corresponding to a position within the OMZ. According to the depositional model suggested, the ASK outcrop stayed in deep water facies of the Tunisian Trough throughout Albian times and within the OMZ.

In all the studied cross-sections, the OM was a mixture of Type II/III of marine planktonic and continental origin. Moreover, the abundant tricyclic terpanes, C30 17 α (H)diahopane and the distribution of homohopane molecules agreed with a strong bacterial activity under a suboxic environment and thus a high bacterial contribution to this OM.

5.4. Petroleum exploration implications

As revealed by the Rock-Eval analysis, the Albian successions of the Lower Fahdene Formation are characterized by fair to good OM richness and petroleum potential. The preserved OM is predominantly marine planktonic Type II oil and gas prone with some contribution of continental Type III gas prone kerogen. The Tmax, PI and molecular maturity parameters (*n*-alkane/isoprenoid ratios, C27Ts, C29Ts, and C29 sterane isomerisation) showed that this OM achieved thermal maturation grades for oil generation. The hydrocarbon exploitation in the Central Atlas of Tunisia started in 1968 by the discoveries of the Douleb, Semmama and Tamesmida oil fields. These oils accumulated in the carbonate reefal facies of the Aptian Serdj Formation reservoir (Hughes and Reed, 1995; Mejri et al., 2006). These authors considered that these oils were sourced from the Lower Fahdene Formation (Albian, Moualha Formation). Belhaj Mohamed et al. (2013) also attributed these oils to the Lower Fahdene Formation. As revealed by Belhaj Mohamed et al. (2013), both crude oil samples and Fahdene source rocks reveal the occurrence of the C30 diahopane in relatively high quantities. In our study, this biomarker was found in all samples. Therefore, we suggest that this compound should be used in oil-source rock correlation as a specific biomarker for oil sourced from the Lower Fahdene source rock. Furthermore, these oil fields are located on horst highs of Jebel Semmama and Jebel Chambi separated by the graben of Foussana-Kasserine (Fig. 1). Taking into account these structures, the best candidate source rock would be the Lower Fahdene Formation similar to that outcropping at the HM locality. As a result, we suggest that the NW-SE subsiding graben of Foussana-Kasserine was the pod of the active source rock for these oils.

Therefore, our results concord with the oil migration pathways suggested by Belhaj Mohamed et al. (2013) and the cross section established by Dhahri et al. (2015; Fig. 3 of these authors). All of these features are synonymous with proven petroleum systems as defined by Magoon and Dow (1994) around the Kasserine area.

6. CONCLUSIONS

The organic geochemical analysis of outcrop samples of the Fahdene Formation (Albian to Cenomanian) from central to northern Tunisia is characterised by both regional and vertical differences. The Rock-Eval pyrolysis results showed that most of the studied samples had a Total Organic Carbon (TOC) content high enough to qualify an OAE. The Hydrogen Index (HI) values (106 to 533 mg HC.g⁻¹ TOC) and Oxygen Index (OI) values (8 to 333 mg CO₂g⁻¹ TOC) are typical of a mixed marine planktonic and terrestrial OM origin (Type II/III OM). This OM is thermally mature, corresponding to the peak of the petroleum generation window in all the studied sections, as revealed by Tmax values between 439 to 453°C.

However, we registered variations in OM contents vertically along the cross-section and between localities. These variations could be due to several factors: oscillations of both oxic/dysoxic and anoxic conditions that prevailed during the deposition of this formation, the continental material supply by intensive weathering and seasonal rainfall as well as the variation in the expansion of the OMZ in relation to the primary organic production and seasonal upwelling. Our results show that the Albian OAEs have been registered in the southern Tethys margin of Tunisia. However, the records of these OAEs are diachronic and could be linked to the local basin structural framework that prevailed during the Late Aptian/Early Albian and Late Albian times. Our integrated structural and geochemical results allow us to identify the OAE1b (Late Aptian-Early Albian) and the OAE1c (Latest Albian) in central to northern Tunisia. The OAE1b record is better preserved on the southern edge of the Tunisian Trough, around the HM locality. This event expresses the expansion of the OMZ during the sea level

rise that invaded the tilted blocks of the HM area. On the other hand, the OAE1b was discreet or absent in the deeper parts of the Tunisian Trough (i.e. ASK locality), in the Triassic diapir horst structures (BRK locality) and in the inter-diapiric basins (GZ locality). These features denoted that the ocean bottom waters were within the OMZ. Despite these aspects, the OAE1c was registered in both the southern edge, at HM, and the central part of the Tunisian Trough, at BRK and GZ. We suggest that the OM richness of the sediments was favoured by the OMZ expansion all over this relatively deep-marine outer part of the basin. Throughout Albian times, the ASK locality registered a deep basin environment within the Tunisian Trough and always within the OMZ. The OMZ was induced by high marine primary organic productivity, as revealed in thin section observations by abundant radiolarian. We propose that increased plankton productivity during the latest Aptian-early Albian OAE1b and the Latest Albian OAE1c resulted from upwelling of nutrient-enriched waters and volcanism. Furthermore, the thin laminations that were observed both at the macroscopic scale (outcrop) and in thin sections suggested anoxic depositional environments.

In our study, the analysis of molecular fossil biomarkers (n-alkanes, steranes and terpanes) showed that the OM in the Albian successions was of marine planktonic and bacterial origin (aquatic algae and cyanobacteria). The molecular distributions of the studied samples showed abundant tricyclic terpanes, C₃₀ 17 α (H) Diahopane, Hopane and the regular decrease of the C₃₁-C₃₅ homohopane series. All of these features corroborated high bacterial activities in the water column and/or the sea floor, under dysoxic/anoxic deposition environments. However, the continental contribution to this OM cannot be excluded as revealed by the subordinate occurrence of tetracyclic terpanes. The molecular thermal maturity parameters showed that this OM is thermally mature with respect to the zone of oil generation.

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FIGURE CAPTIONS

Fig. 1 Locations of the studied outcrop sections on the present-day architecture of the Tunisian Atlas (A), comprising a faulted block and its compartments: the Tunisian Block (TB) is bounded by three mobile zones corresponding to three fault corridors (DZ, NSC and GNTC), each with its own structures, limiting the central block characterized by faults and widely-spaced folds (Rigane and Gourmelen, 2011), (B) Localities cited in the text on the schematic tectonic map of Tunisia (Rigane and Gourmelen, 2011).

Fig. 2 Sampled levels on the lithostratigraphic logs of the studied outcrop sections of the Fahdene Formation (Albian-Cenomanian) in central and northern Tunisia.

Fig. 3 Geochemical logs of TOC, PI and Tmax from Rock-Eval pyrolysis results. TOC, Total Organic Carbon (%); PI, Production Index= $S1/S1+S2$; S1, free hydrocarbons (mg HC.g^{-1} Rock); S2, potential hydrocarbons (mg HC.g^{-1} Rock); Tmax, S2 maximum pyrolysis temperature ($^{\circ}\text{C}$);

Fig. 4 Microscopic observations of representative samples of the Albian series from Koudiat Berkouchia (BRK) and Jebel Ghazouane (GZ) localities. Samples BRK11 and GZ12 show organic-rich wackestone laminations of thin dark layers and bright radiolarian-rich layers (dashed line). R: radiolarians; O: ostracods; Bit.: Bitumen; O: Ostracods; F: Foraminifera; Ta: Traces of algae.

Fig. 5 GC-MS chromatograms (Total Ion Current mode) of aliphatic fractions showing the n-alkanes and acyclic isoprenoids for selected Fahdene Formation samples in central and northern Tunisia. Numbers refer to n-alkanes Carbon number long chain; 17, *n*-C₁₇; 18, *n*-C₁₈; Pr, pristane; Ph, phytane; Sq, Squalane.

Fig. 6 GC-MS chromatograms showing the distribution of terpanes (*m/z* 191) and steranes (*m/z* 217) of representative samples from the Fahdene Formation in central and northern Tunisia (for key see Table 3).

Fig. 7 Location of the studied samples on a Carbonate contents (CaCO_3 (%)) vs. TOC crossplot expressing the abundance of bioturbations and the sea floor bottom water oxygenation.

Fig. 8 Fahdene Formation studied samples organic matter typing, thermal maturity and hydrocarbon potential on Rock-Eval pyrolysis diagrams: HI vs. OI (a), HI vs. Tmax (b), PP vs. TOC and Tmax versus PI. TOC, Total Organic Carbon (% Rock); Tmax, S2 peak maximum pyrolysis temperature ($^{\circ}\text{C}$); HI, Hydrogen Index (mg HC.g^{-1} TOC); OI, Oxygen Index ($\text{mg CO}_2\text{g}^{-1}$ TOC); PP, Petroleum potential yield (S1+S2, mg HC.g^{-1} Rock); PI, Production Index (S1/S1+S2); S1, free hydrocarbons (mg HC.g^{-1} Rock); S2, potential hydrocarbons (mg HC.g^{-1} Rock).

Fig. 9 Location of representative Fahdene Formation samples on a Pr/n-C₁₇ versus Ph/n-C₁₈ cross-plot. Pr/n-C₁₇: pristane/n-C₁₇ peak height ratio; Ph/n-C₁₈: phytane/ n-C₁₈ peak height ratio.

Fig. 10 (A) Palaeogeographic map at ~113Ma (modified after Huber and Leckie (2011) and Sabatino et al., 2015) showing the location of Tunisia (TN) and the Vocontian Basin (VB). (B) Albian Oceanic Anoxic Events (OAE) records on a synthetic model of the deposition of the Lower Fahdene Formation in the studied zone in relationship with oxygen minimum zone (OMZ), upwelling, primary organic productivity (POP) and structural setting.

TABLE CAPTIONS

Table 1 Carbonate contents and Rock-Eval 6 (Espitalié et al., 1985; Lafargue et al., 1998; Behar et al., 2001) analysis of Albian Fahdene Formation samples from central and northern Tunisia.

Table 2 Bulk data for solvent extraction and chromatographic characterization of Fahdene Formation selected samples.

Table 3 Peak assignment for steranes (m/z 217) and triterpanes (m/z 191) in the GC-MS chromatograms of figure 6.

Table 4 Terpane (m/z 191) and sterane (m/z 217) results from GC-MS analysis of the Albian OM in central and northern Tunisia.

Figure 1

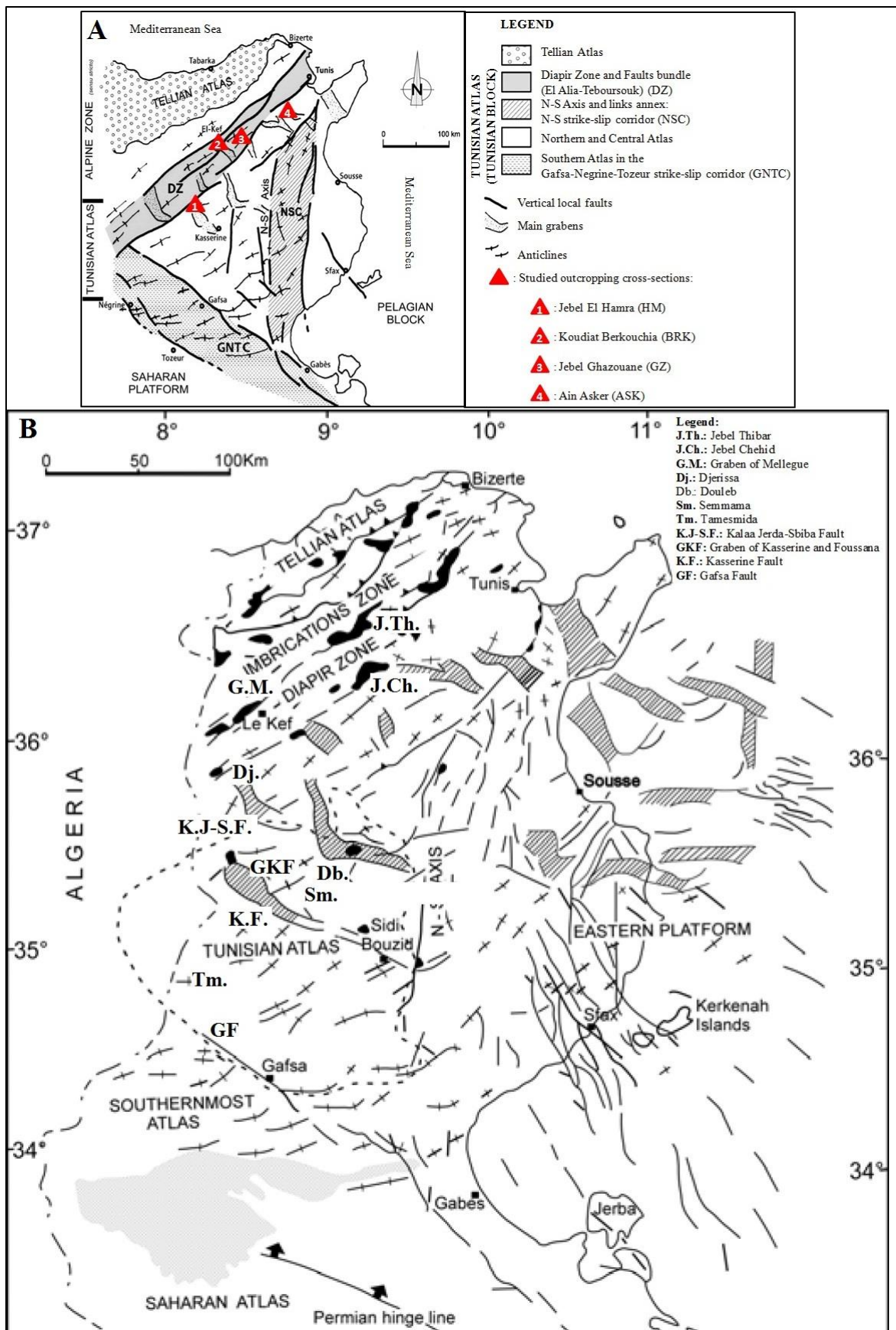


Figure 2

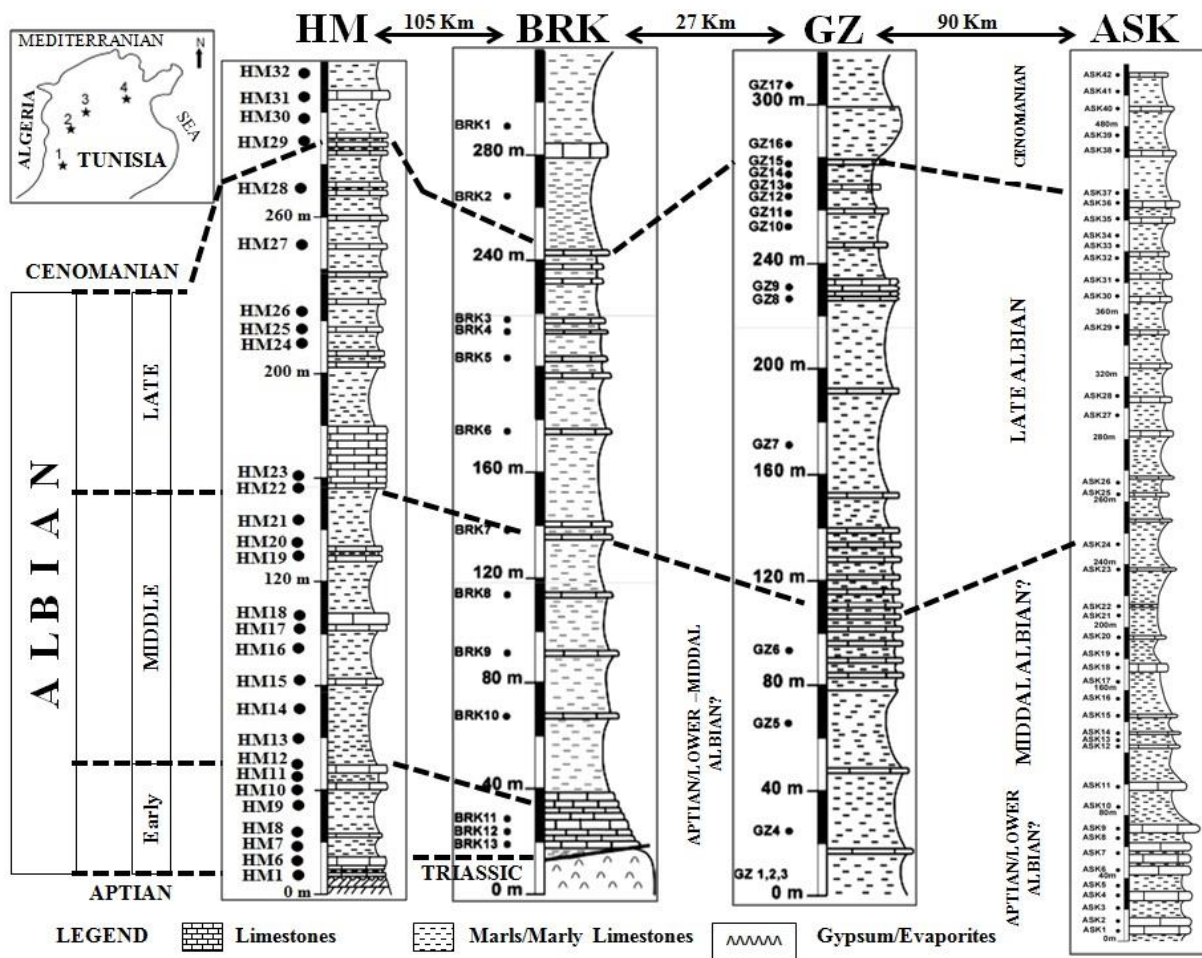


Figure 3

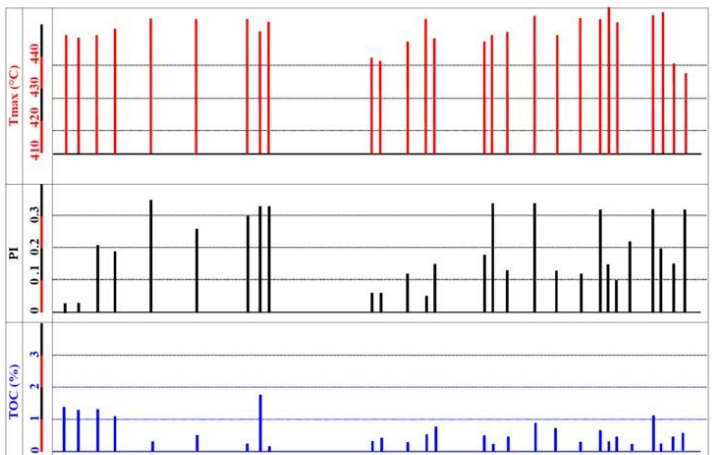
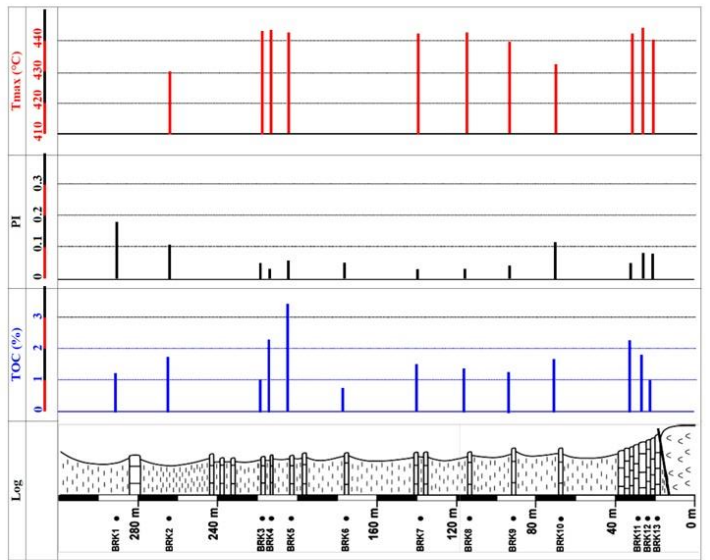
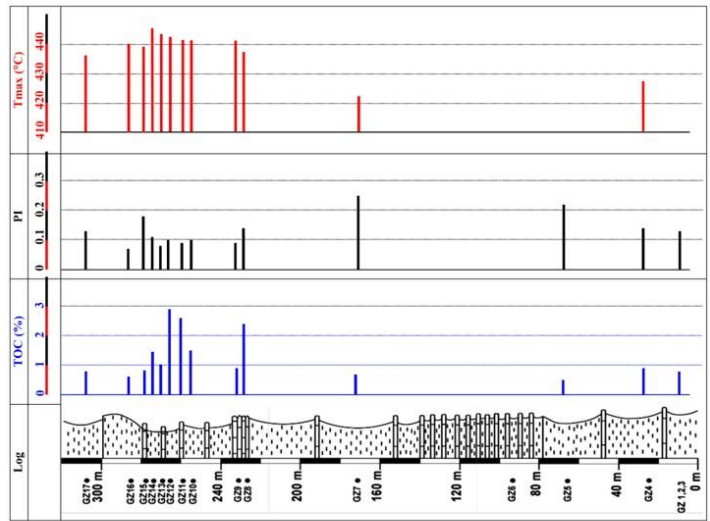


Figure 4

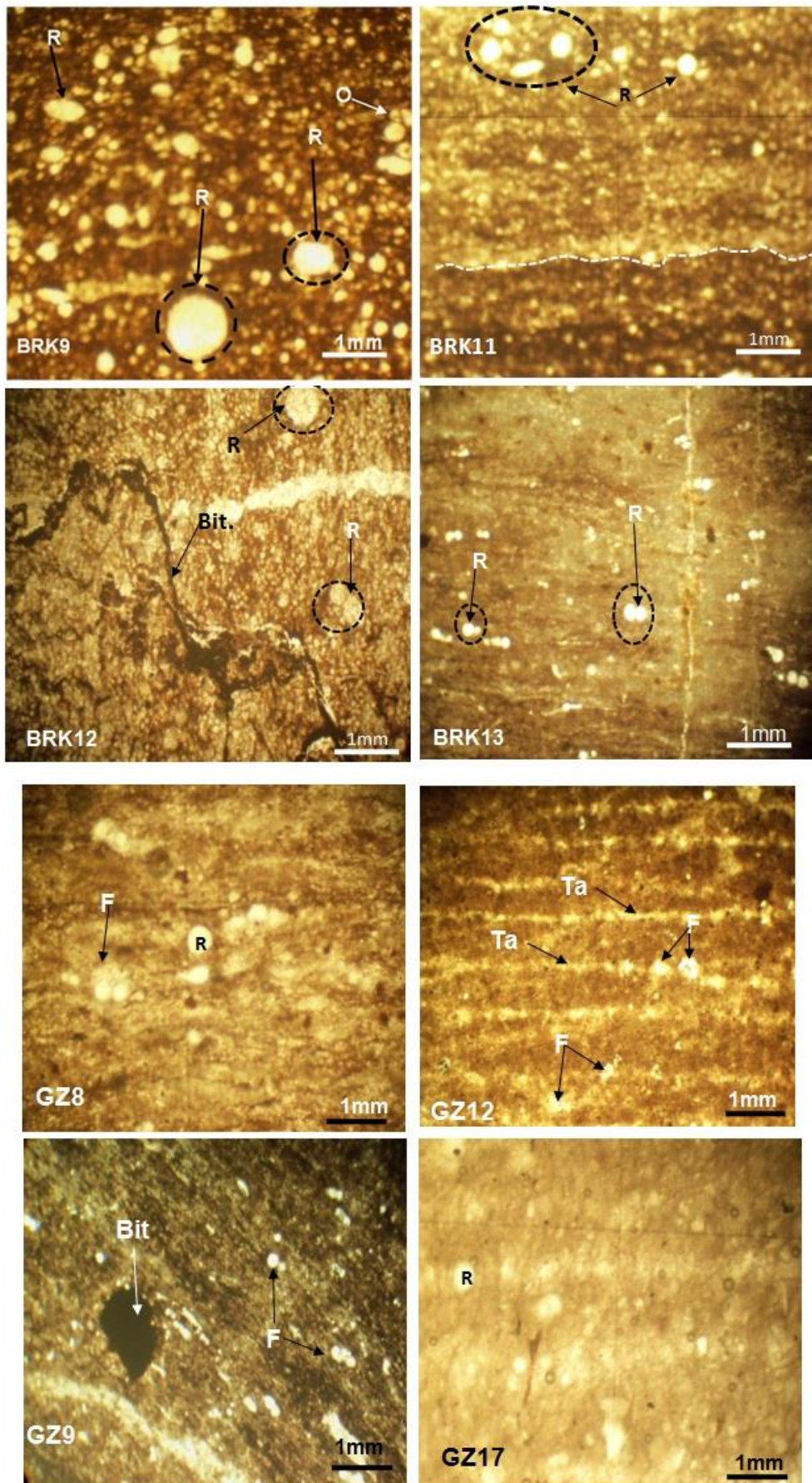
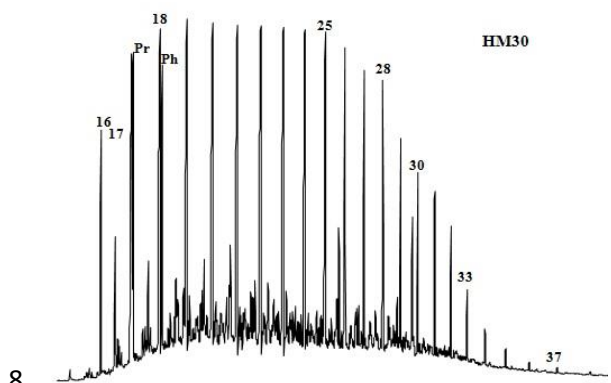
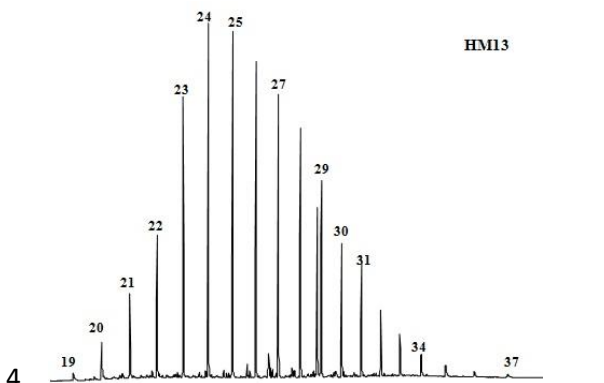
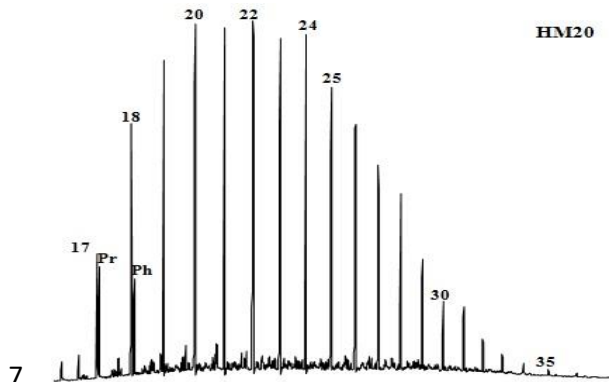
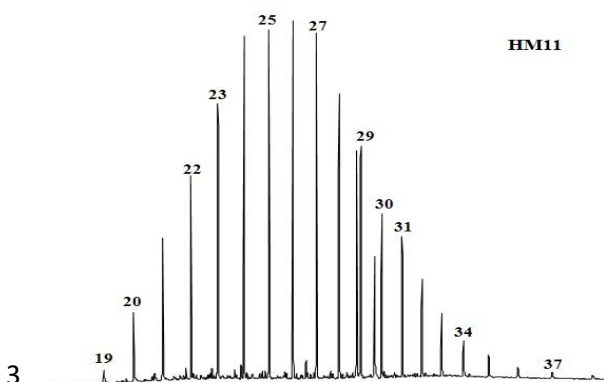
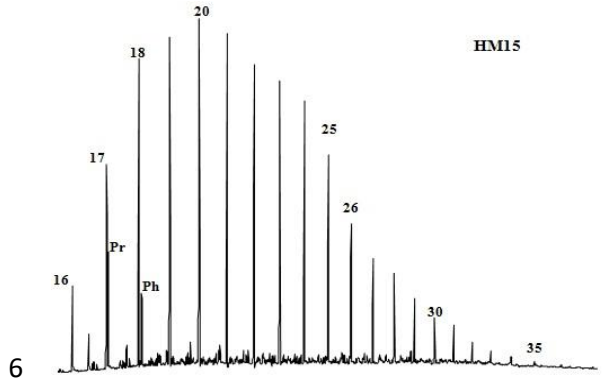
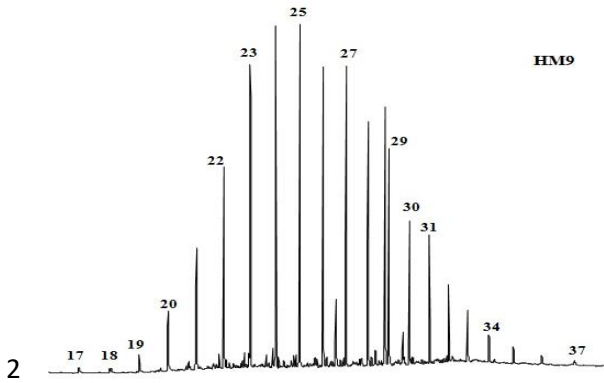
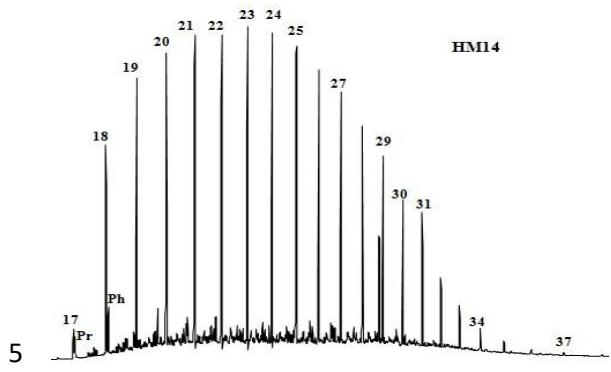
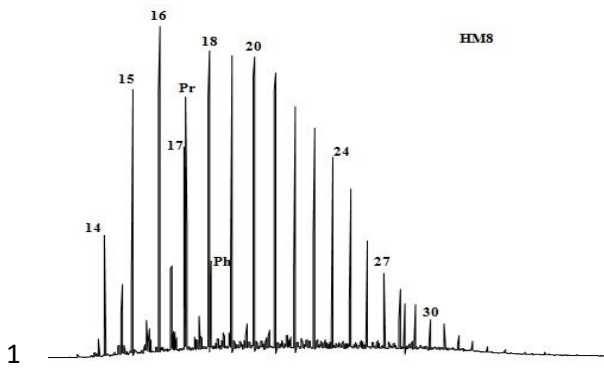
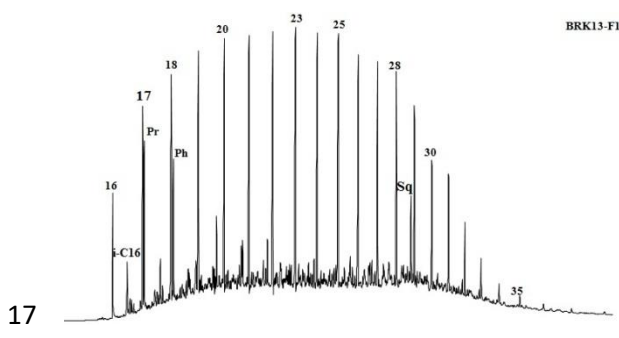
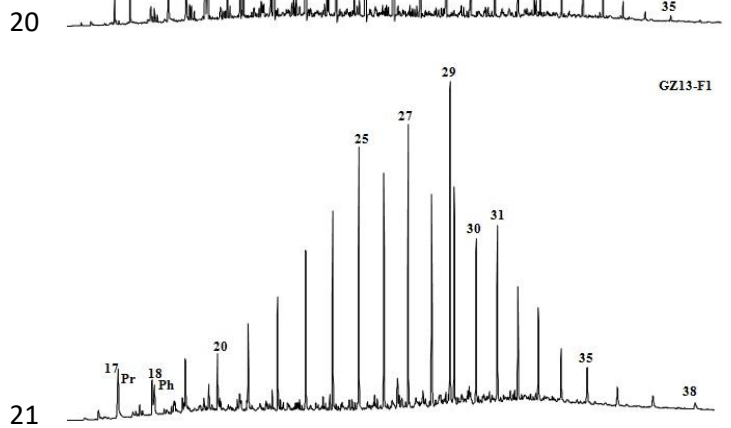
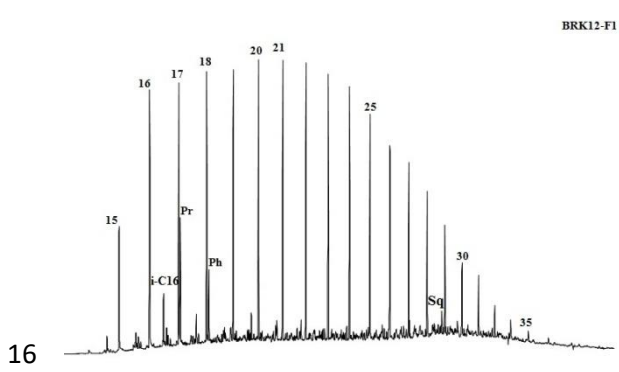
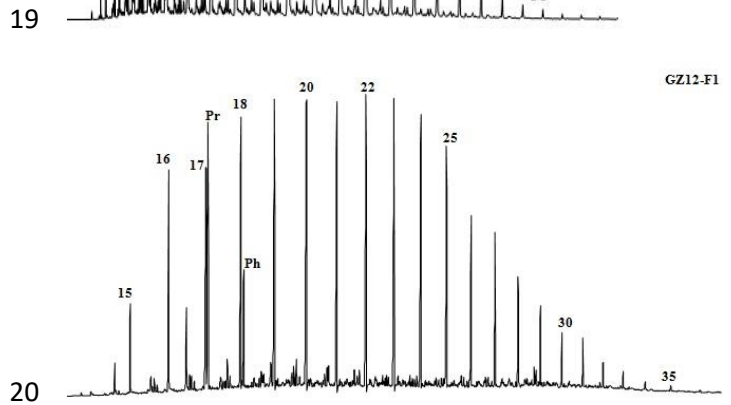
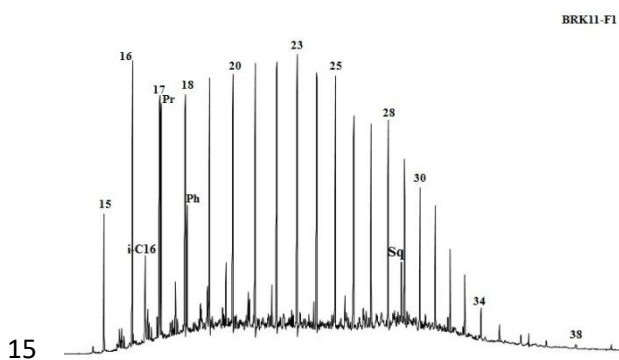
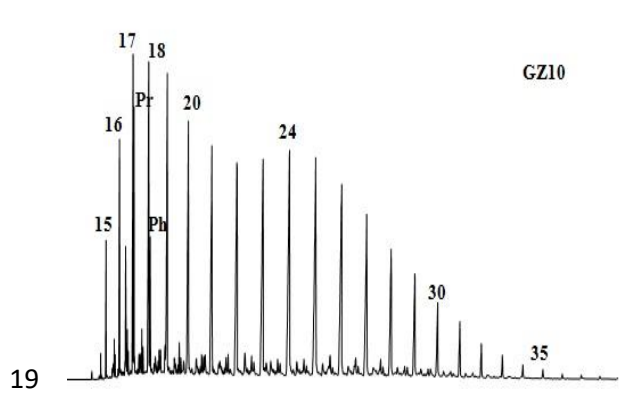
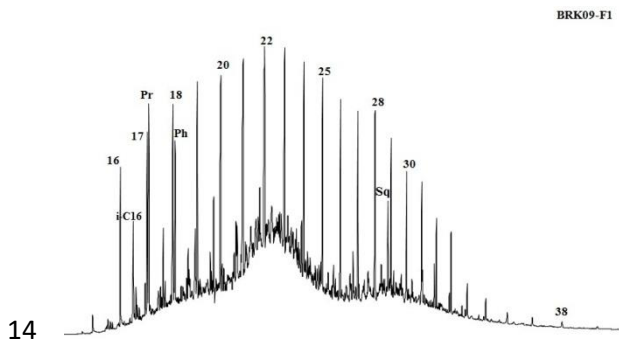
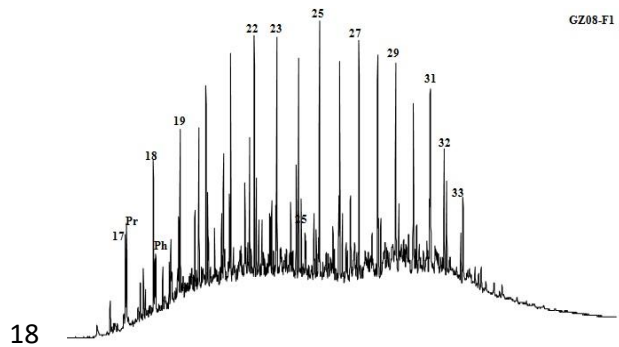
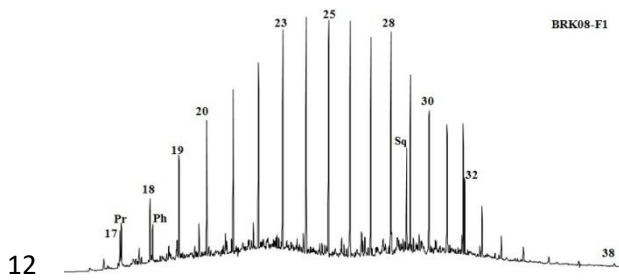


Figure 5



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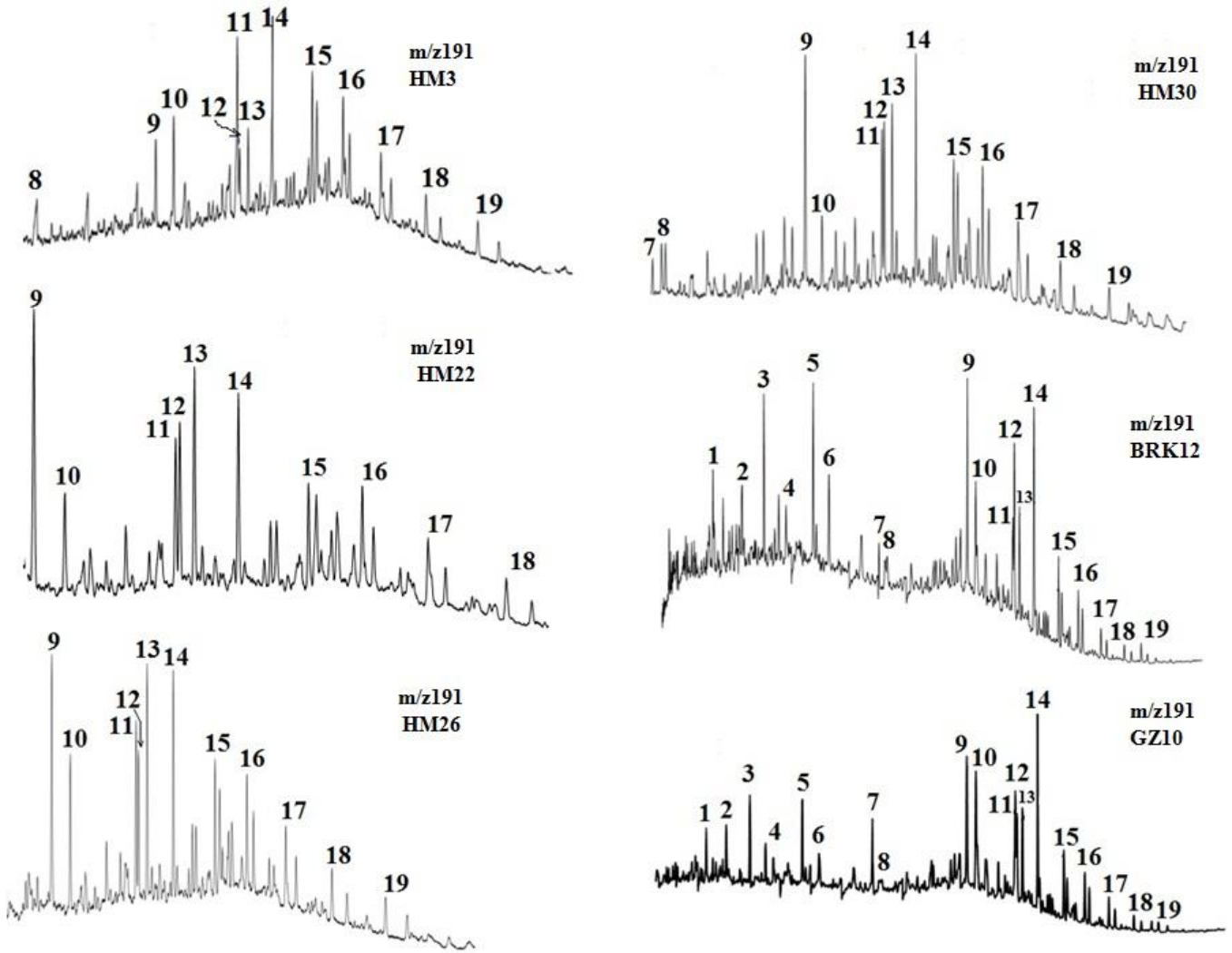
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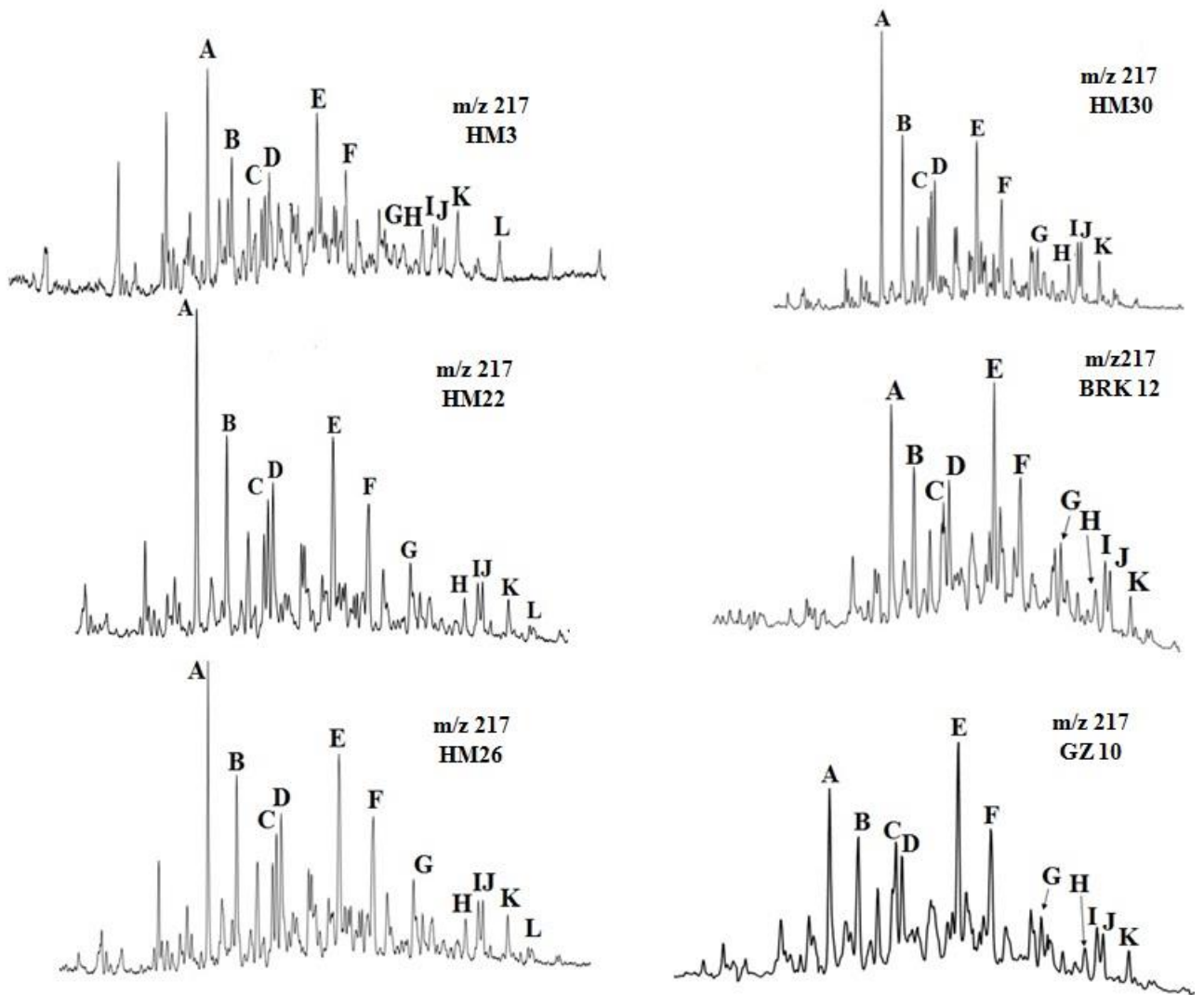
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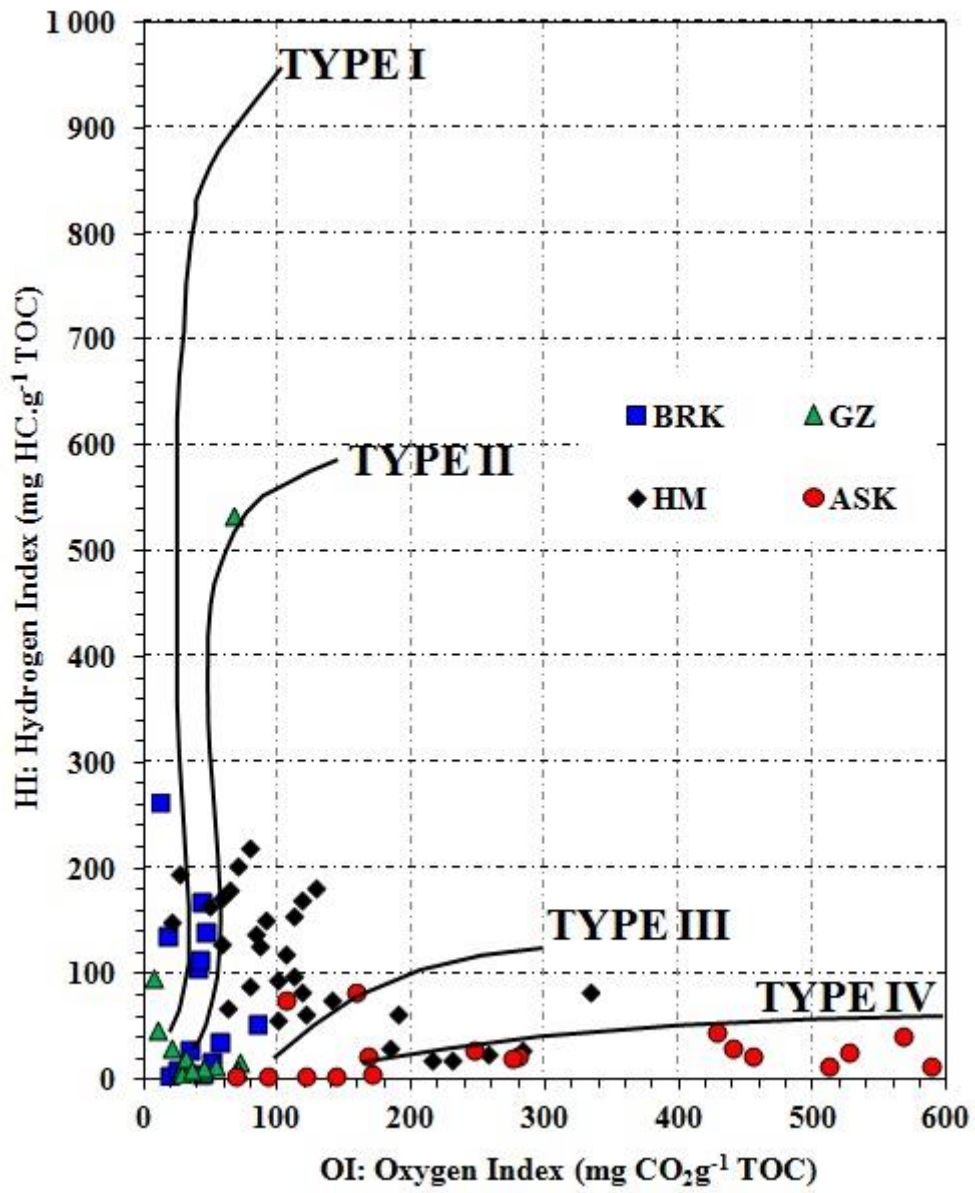
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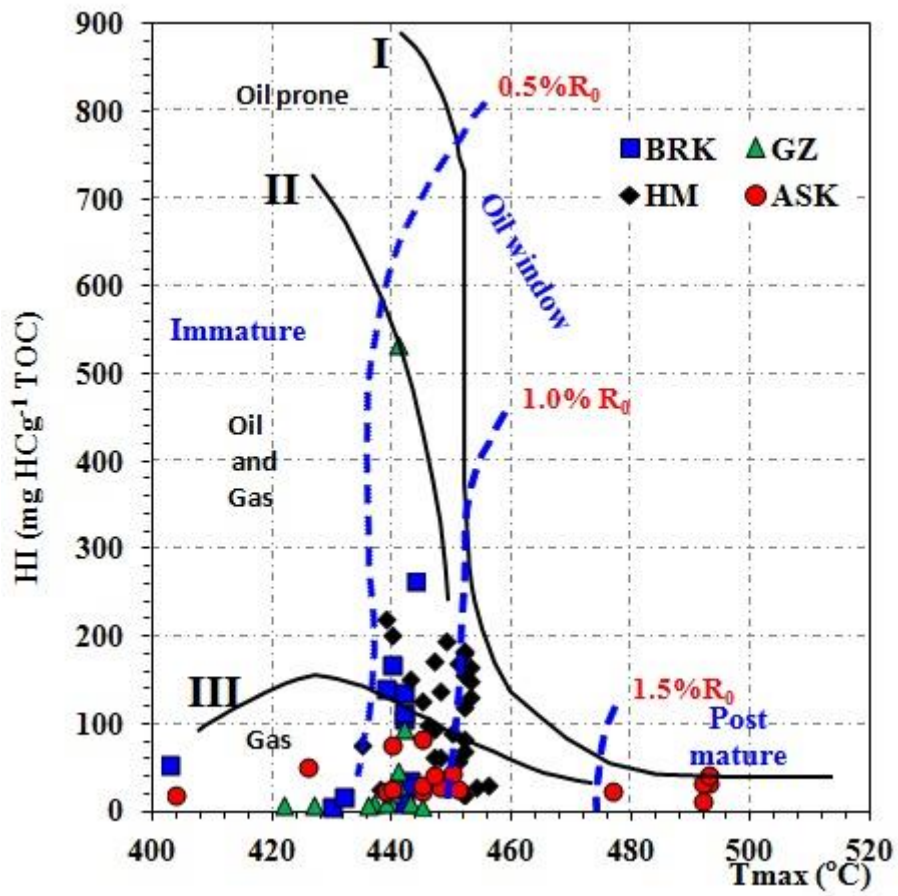
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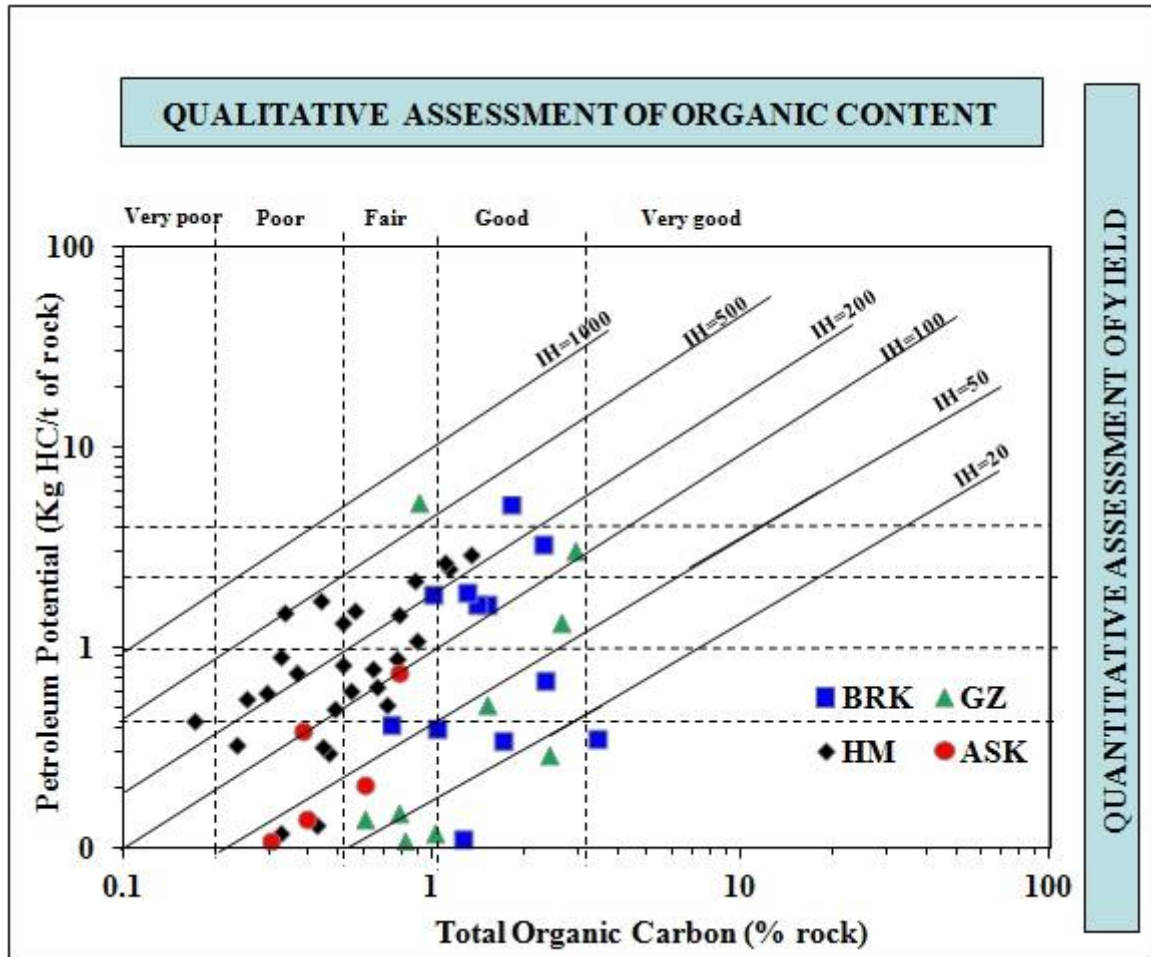
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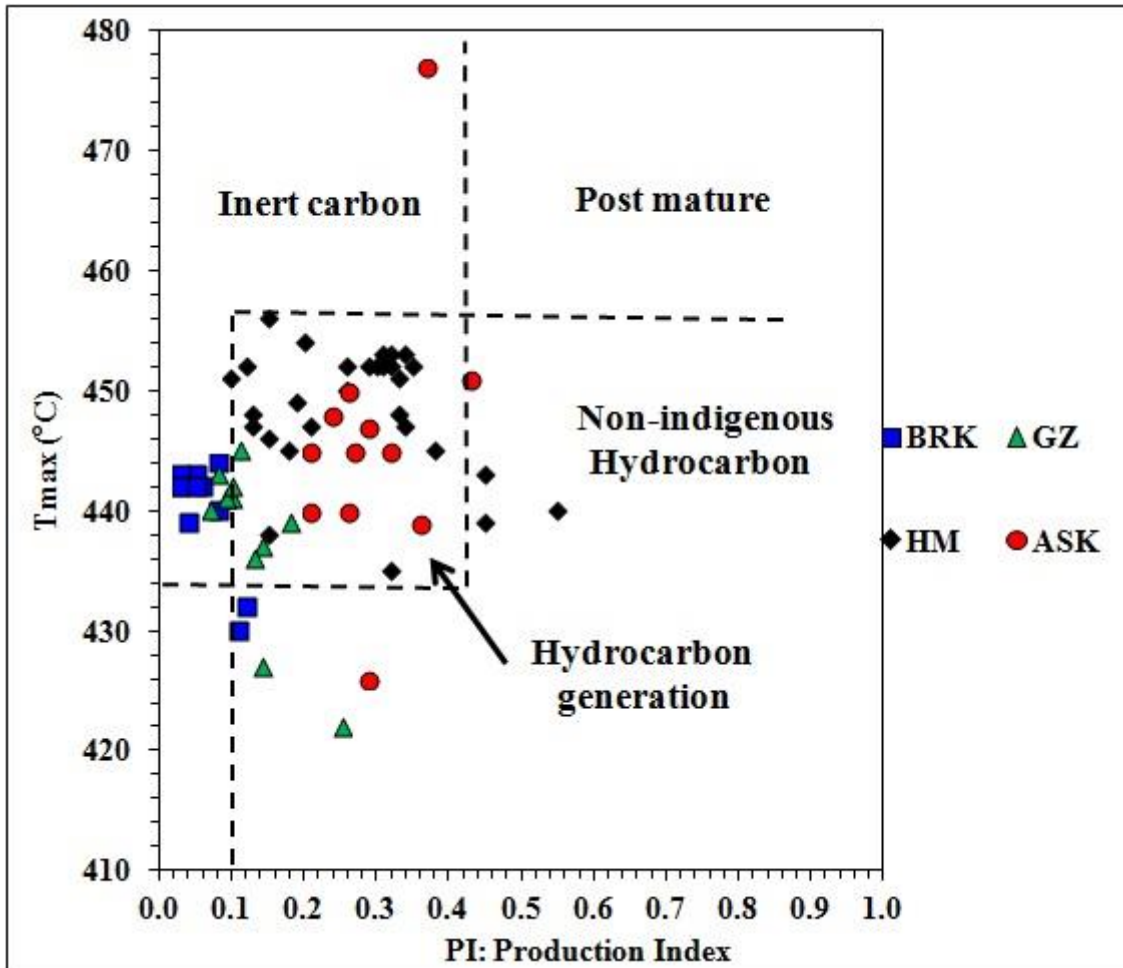
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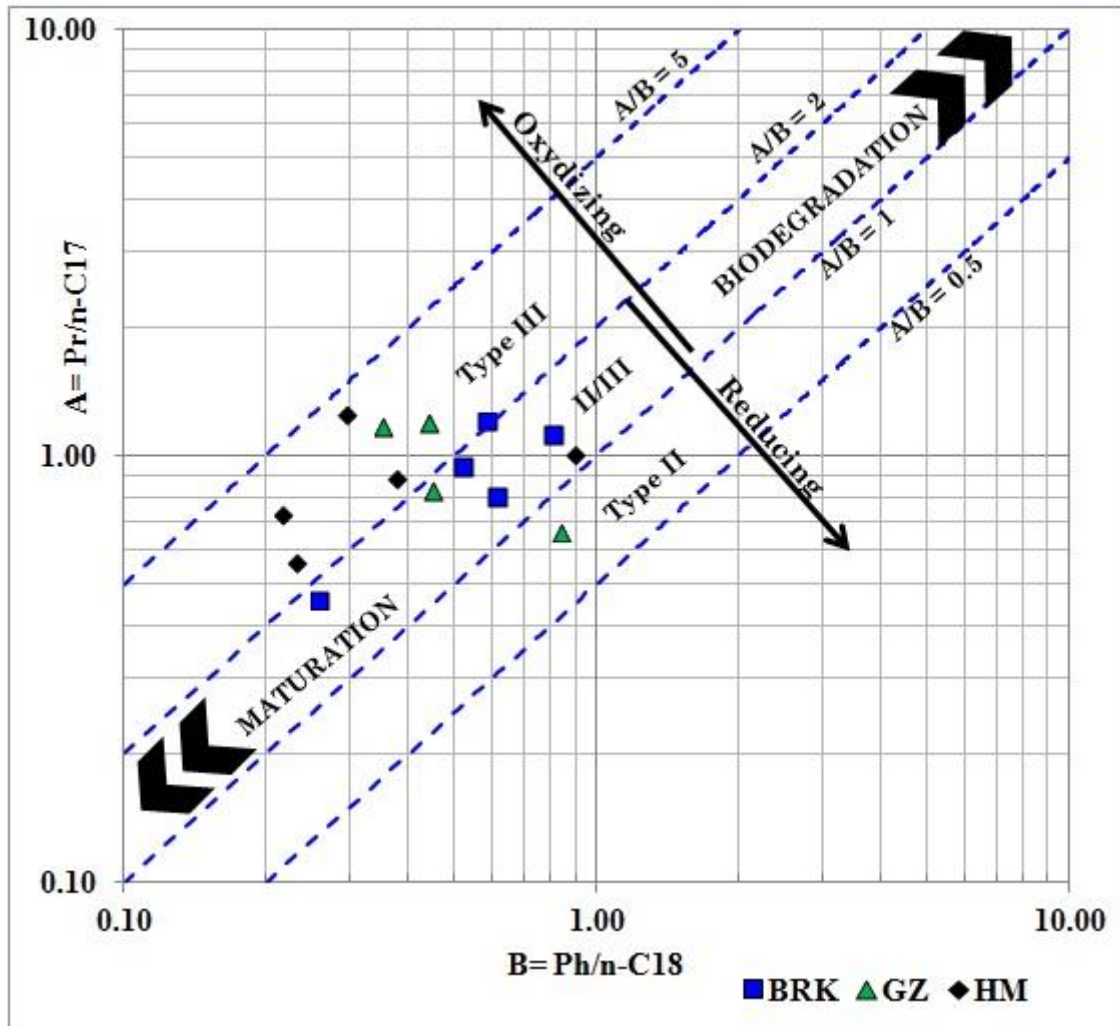
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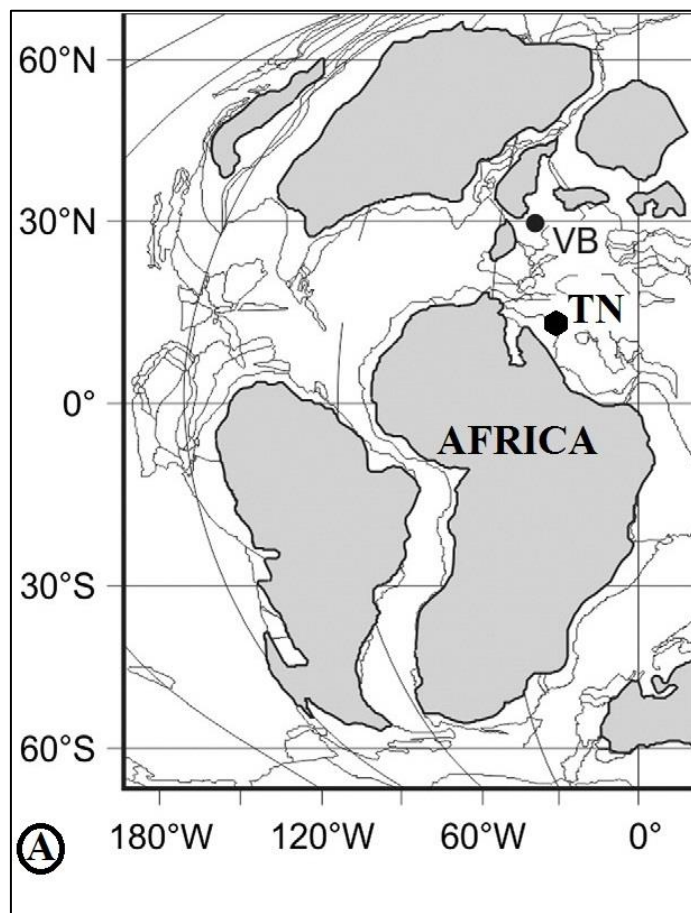
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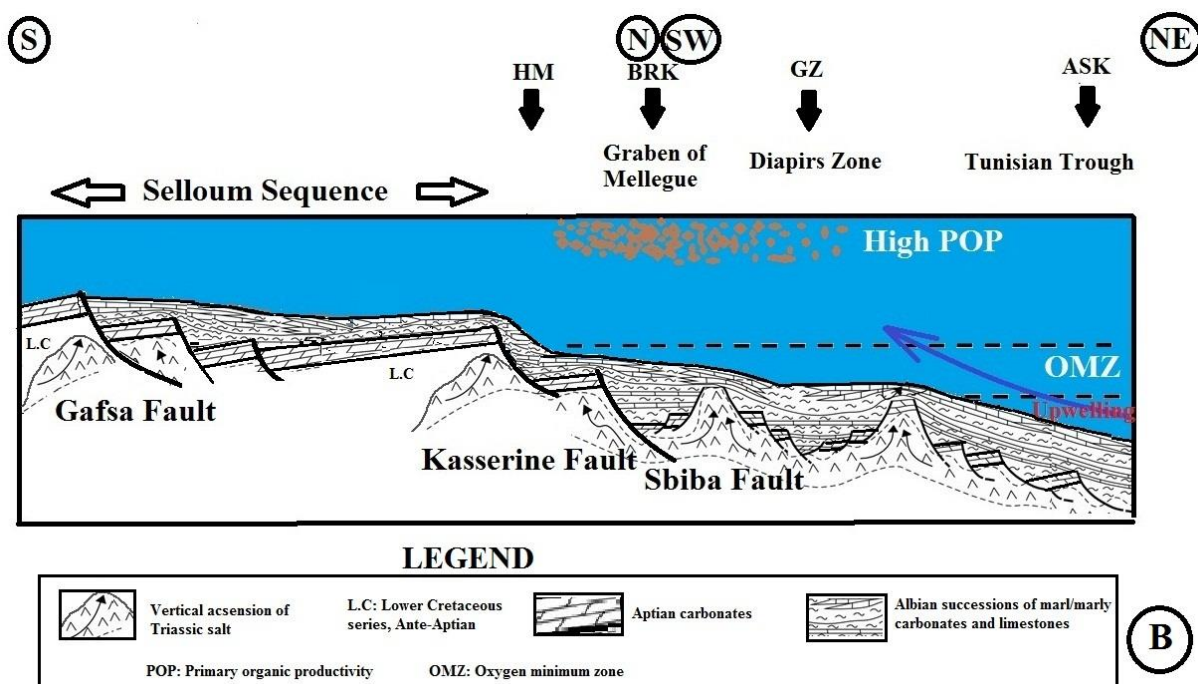
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119 **Figure 10**



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Samples	CaCO ₃ (%)	TOC ^a	S1 ^b	S2 ^c	S3 ^d	PP ^e	PI ^f	HI ^g	OI ^h	Tmax ⁱ
BRK1	61	1.25	0.02	0.09	0.52	0.11	0.18	7	42	361
BRK2	60	1.77	0.01	0.08	0.33	0.09	0.11	5	19	430
BRK3	59	1.04	0.02	0.37	0.58	0.39	0.05	36	56	443
BRK4	55	2.32	0.02	0.67	0.8	0.69	0.03	29	34	443
BRK5	58	3.43	0.02	0.33	0.85	0.35	0.06	10	25	442
BRK6	65	0.74	0.02	0.39	0.62	0.41	0.05	53	84	403
BRK7	50	1.5	0.05	1.59	0.58	1.64	0.03	106	39	442
BRK8	64	1.4	0.05	1.59	0.57	1.64	0.03	114	41	442
BRK9	71	1.3	0.07	1.82	0.6	1.89	0.04	140	46	439
BRK10	82	1.7	0.04	0.3	0.85	0.34	0.12	18	50	432
BRK11	65	2.3	0.16	3.14	0.39	3.3	0.05	137	17	442
BRK12	63	1.8	0.39	4.74	0.21	5.13	0.08	263	12	444
BRK13	77	1.01	0.14	1.71	0.42	1.85	0.08	169	42	440
GZ3	1,3	0.78	0.02	0.13	0.56	0.15	0.13	17	72	390
GZ4	1,4	0.9	0.01	0.06	0.4	0.07	0.14	7	44	427
GZ5	0,93	0.5	0.02	0.07	0.27	0.09	0.22	14	54	353
GZ7	3,27	0.68	0.02	0.06	0.48	0.08	0.25	9	70	422
GZ8	89	2.4	0.04	0.25	0.64	0.29	0.14	10	27	437
GZ9	86	0.9	0.49	4.8	0.6	5.29	0.09	533	67	441
GZ10	45	1.5	0.05	0.47	0.32	0.52	0.1	31	21	441
GZ11	60	2.6	0.12	1.22	0.26	1.34	0.09	47	10	441
GZ12	64	2.9	0.31	2.79	0.23	3.1	0.1	96	8	442
GZ13	75	1.02	0.01	0.11	0.29	0.12	0.08	11	28	443
GZ14	52	1.45	0.01	0.08	0.4	0.09	0.11	6	28	445
GZ15	57	0.82	0.02	0.09	0.37	0.11	0.18	11	45	439
GZ16	54	0.61	0.01	0.13	0.19	0.14	0.07	21	31	440
GZ17	58	0.79	0.01	0.07	0.28	0.08	0.13	9	35	436
HM 1	47	0.54	0.19	0.42	0.76	0.61	0.32	77	140	435
HM 2	68	0.56	0.69	0.86	0.51	1.55	0.45	153	91	443
HM 3	43	0.66	0.18	0.46	0.41	0.64	0.29	70	62	452
HM 4	70	0.77	0.46	1.01	0.45	1.47	0.31	131	58	453
HM 5	54	0.89	0.28	0.81	0.69	1.09	0.26	91	78	450
HM 6	51	0.42	0.02	0.11	1.08	0.13	0.15	26	257	438
HM 7	31	0.24	0.02	0.07	0.68	0.09	0.2	29	283	454
HM 8	74	1.12	0.79	1.69	0.23	2.48	0.32	151	21	453
HM 9	24	0.19	0.01	0.04	0.41	0.05	0.22	21	215	585
HM 10	96	0.46	0.03	0.27	0.46	0.3	0.1	59	100	451
HM 11	61	0.32	0.02	0.1	0.59	0.12	0.15	31	184	456
HM 12	67	0.64	0.25	0.54	2.13	0.79	0.32	84	333	452
HM 13	33	0.3	0.01	0.06	0.69	0.07	0.12	20	230	452
HM 14	45	0.71	0.07	0.45	0.86	0.52	0.13	63	121	447
HM 15	77	0.87	0.74	1.45	0.42	2.19	0.34	167	48	453

HM 16	40	0.44	0.04	0.28	0.83	0.32	0.13	64	189	448
HM 17	95	0.23	0.11	0.22	0.23	0.33	0.34	96	100	447
HM 18	93	0.48	0.09	0.41	0.56	0.5	0.18	85	117	445
HM 19	73	0.76	0.13	0.75	0.85	0.88	0.15	99	112	446
HM 20	82	0.51	0.42	0.93	0.32	1.35	0.31	182	63	452
HM 21	69	0.29	0.23	0.37	0.25	0.6	0.38	128	86	445
HM 22	78	0.43	0.76	0.95	0.34	1.71	0.45	220	79	439
HM 23	91	0.33	0.83	0.67	0.2	1.5	0.55	203	70	440
HM 24	92	0.17	0.14	0.29	0.2	0.43	0.33	171	118	451
HM 25	97	0.36	0.25	0.5	0.3	0.75	0.33	139	83	448
HM 26	72	0.25	0.17	0.39	0.28	0.56	0.3	156	112	452
HM 27	85	0.51	0.21	0.61	0.54	0.82	0.26	120	106	452
HM 28	71	0.32	0.31	0.59	0.41	0.9	0.35	184	128	452
HM 29	83	1.1	0.52	2.17	0.29	2.69	0.19	197	26	449
HM 30	78	1.32	0.6	2.3	0.77	2.9	0.21	174	58	447
ASK1	89	0.3	0.04	0.07	0.84	0.11	0.36	23	280	439
ASK3	81	0.58	0.05	0.02	0.7	0.07	0.71	3	121	299
ASK4	88	0.2	0.03	0.04	0.55	0.07	0.45	20	275	404
ASK5	73	0.78	0.16	0.59	0.82	0.75	0.21	76	105	440
ASK6	87	0.11	0.01	0.03	0.58	0.04	0.26	27	527	440
ASK7	71	0.6	0.07	0.14	1	0.21	0.32	23	167	445
ASK8	81	0.33	0.05	0.02	0.56	0.07	0.69	6	170	310
ASK11	95	0.37	0.04	0.01	0.53	0.05	0.81	3	143	295
ASK12	97	0.08	0.02	0.01	0.47	0.03	0.75	12	588	292
ASK14	93	0.38	0.08	0.31	0.6	0.39	0.21	82	158	445
ASK17	61	0.39	0.03	0.11	0.96	0.14	0.24	28	246	448
ASK18	87	0.13	0.02	0.03	0.59	0.05	0.37	23	454	477
ASK20	89	0.07	0.01	0.01	0.58	0.02	0.51	14	829	317
ASK22	88	0.89	0.05	0.03	0.61	0.08	0.64	3	69	295
ASK23	89	0.6	0.04	0.02	0.55	0.06	0.67	3	92	290
ASK24	58	0.11	0.02	0.03	0.67	0.05	0.43	27	609	451
ASK25	93	0.03	0.01	0.01	0.3	0.02	0.48	33	1000	493
ASK27	92	0.11	0.02	0.05	0.47	0.07	0.26	45	427	450
ASK29	71	0.2	0.02	0.06	0.88	0.08	0.27	30	440	445
ASK31	93	0.02	0.01	0.01	0.32	0.02	0.5	50	1600	287
ASK32	95	0.03	0.01	0.01	0.36	0.02	0.32	33	1200	492
ASK34	96	0.08	0.01	0.01	0.41	0.02	0.34	12	512	492
ASK36	97	0.07	0.01	0.03	0.44	0.04	0.28	43	629	493
ASK38	84	0.12	0.02	0.05	0.68	0.07	0.29	42	567	447
ASK42	98	0.12	0.02	0.06	1.62	0.08	0.29	50	1350	426

124 ^a Total organic carbon (%R); ^b S1 (mg HC.g⁻¹ rock); ^c S2 (mg HC.g⁻¹ rock); ^d S3CO₂ (mg CO₂g⁻¹ rock);
125 ^e Petroleum potential, PP= S1+S2 (mg HC.g⁻¹ rock); ^f Production index, PI=S1/(S1+S2);
126 ^g Hydrogen Index, HI=S2x100/TOC (mg HC.g⁻¹ TOC); ^h Oxygen Index, OI=S3x100/TOC (mg
127 CO₂g⁻¹ TOC); ⁱ S2 maximum peak temperature (°C).

Name	TOC (% R) ^a	EOM (‰ R) ^b	EOM (% TOC)	EOM Composition (%)			GC of SHCs		
				NSO ^c	SHCs ^d	AHCs ^e	Pr/Ph ^f	Pr/ <i>n</i> -C ₁₇ ^g	Ph/ <i>n</i> -C ₁₈ ^h
BRK8	1.4	Nd ⁱ	Nd	Nd	Nd	Nd	1.13	1.21	1.13
BRK9	1.3	Nd	Nd	Nd	Nd	Nd	1.29	1.14	1.29
BRK11	2.3	0.66	2.9	76.4	20	3.6	1.90	0.95	1.90
BRK12	1.8	1.64	9.1	62.5	31.3	6.2	1.76	0.46	1.76
BRK13	1.01	0.62	6.1	45.2	9.6	45.2	1.21	0.81	1.21
GZ8	2.4	0.12	0.5	21.4	64.3	14.3	1.90	1.18	0.35
GZ9	0.9	0.8	8.9	27	56.8	16.2	1.15	1.05	0.94
GZ10	1.5	0.52	3.5	54.2	8.7	37.1	1.94	0.83	0.44
GZ12	2.9	Nd	Nd	Nd	Nd	Nd	2.29	1.20	0.44
GZ13	1.02	Nd	Nd	Nd	Nd	Nd	1.20	0.67	0.83
HM8	0.71	Nd	Nd	Nd	Nd	Nd	2.91	1.25	0.29
HM14	0.87	Nd	Nd	Nd	Nd	Nd	0.47	0.73	0.22
HM15	0.92	1.43	15.5	15	20	65	1.64	0.56	0.23
HM20	0.56	1.07	19.1	17	24	59	1.15	0.88	0.38
HM30	1.2	1.86	15.5	13	21	66	1.09	1.01	0.89

130 ^a Total organic carbon; ^b chloroform extractable OM; ^c nitrogen, sulfur and oxygen heteroatomic

131 compounds; ^d saturated hydrocarbons; ^e aromatic hydrocarbons; ^f Pr/Ph peak height ratio; ^g

132 Pr/*n*-C₁₇ peak height ratio; ^h Ph/*n*-C₁₈ peak height ratio; ⁱ not determined.

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Steranes GC-MS m/z 217	Terpanes GC-MS m/z 191
A – 13β(H),17α(H)-20S Diasterane C27 B – 13β(H),17α(H)-20R Diasterane C27 C - 13β(H),17α(H)Diasterane 20S C28 + 14α(H),17 α (H)-20S Sterane C27 D - 13β(H),17α(H)-20S Diasterane C29 + 14β(H),17β(H)20R sterane C27 E– 14β(H),17β(H)20S sterane C27 + 13β(H),17α(H)20S Diasterane C28 F – 14α(H),17α(H)-20R sterane C27 G - 14α(H),17α(H)20R sterane C28, H - 14α(H), 17α(H)20S sterane C29 I - 14β(H),17β(H)20R sterane C29 J - 14β(H), 17β(H)20S sterane C29 K - 14α(H),17α(H)20R sterane C29 L - 14α(H),17α(H)20R sterane C30	1 –Tricyclic Terpane C19 2–Tricyclic Terpane C20 3 –Tricyclic Terpane C21 4 –Tricyclic Terpane C22 5 –Tricyclic Terpane C23 6 –Tricyclic Terpane C24 7 –Tetracyclic Terpane C24 8 –Tricyclic Terpane C26 (C26T) 9 – 18α(H)-22,29,30-trisnorneohopane (C27Ts) 10 - 17α(H)-22,29,30-trisnorhopane (C27Tm) 11 – 17α(H),21β(H)-norhopane (C29H) 12- 18α(H)-norneohopane (C29Ts) 13- C30 17α(H)-diahopane 14 – 17α(H),21β(H)-hopane (C30H) 15 – 17α(H),21β(H)-C31 homohopane (22S + 22R) 16 – 17α(H),21β(H)-C32 bishomohopane (22S +22R) 17 – 17α(H),21β H)-C33 trishomohopane (22S + 22R) 18 – 17α(H),21β(H)-C34 tetrakishomohopane (22S + 22R) 19 – 17α(H),21β(H)-C35 pentakishomohopane (22S + 22R)

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158 **Table 4**

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TERPANES m/z 191							
Sample	Ts/Tm ^a	Ts/ (Ts+Tm) ^b	C29Ts/ C29H ^c	Diah/ C29H ^d	DI ^e	C30H/ C29H ^f	
160	HM3	0.77	0.43	0.35	0.47	0.30	1.07
161	HM22	2.83	0.74	1.04	1.46	0.53	1.29
	HM26	1.63	0.62	0.84	1.30	0.50	1.30
162	HM30	3.39	0.77	1.08	1.20	0.44	1.53
163	BRK12	1.85	0.65	1.93	1.19	0.25	3.63
164	GZ10	1.11	0.53	0.82	0.88	0.32	1.85

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STERANES m/z 217					
Sample	29 α S/ 29 α (S+R) ^g	29 β S/ 29 β (S+R) ^h	29 β (S+R)/ [(29 β (S+R)+29 α (S+R))] ⁱ	27DS/ 27D(S+R) ^j	
166	HM3	0.47	0.44	0.45	0.68
167	HM22	0.50	0.55	0.55	0.64
168	HM26	0.45	0.50	0.57	0.63
	HM30	0.47	0.49	0.60	0.62
169	BRK12	0.48	0.45	0.62	0.59
	GZ10	0.45	0.45	0.62	0.59

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171 ^a Ts/Tm: 18 α (H)-22,29,30-trisnorneohopane (C27Ts)/ 17 α (H)-22,29,30-trisnorhopane
172 (C27Tm)

173 ^b Ts/Ts+Tm: 18 α (H)-22,29,30-trisnorneohopane (C27Ts)/ [18 α (H)-22,29,30-
174 trisnorneohopane (C27Ts)+17 α (H)-22,29,30-trisnorhopane (C27Tm)]

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176 ^c C29Ts/ C29H : 18 α (H)norneohopane (C29Ts)/17 α (H),21 β (H)-norhopane (C29H)

177 ^d Diah/C29H: C30 17 α (H)-diahopane/ 17 α (H),21 β (H)-norhopane (C29H)

178 ^e DI: Diahopane Index=C30 17 α (H)-diahopane/ [C30 17 α (H)-diahopane+ 17 α (H),21 β (H)-
179 hopane (C30H)]

180 ^f C30H/C29H=17 α (H),21 β (H)-hopane (C30H)/ 17 α (H),21 β (H)-norhopane (C29H)

181 ^g 29 α S/29 α (S+R)= 14 α (H),17 α (H)20S sterane C29/[14 α (H), 17 α (H)20S sterane C29+14 α (H),
182 17 α (H)20R sterane C29]

183 ^h 29 β S/29 β (S+R)= 14 β (H),17 β (H)20S sterane C29/(14 β (H),17 β (H)20(S +R) sterane C29).

184 ⁱ 29 β (S+R)/[29 β (S+R)+29 α (S+R)]= 14 β (H),17 β (H)20(S +R)sterane C29/(14 β (H),
185 17 β (H)20(S+R) sterane C29+14 α (H), 17 α (H)20(S+R) sterane C29)]

186 ^j 27DS/27D(S+R)=13 \square (H),17 α (H)-20SDiasterane C27/[13 \square (H),17 α (H)-20(S+R) Diasterane
187 C27]

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