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APPLICATION OF ^{18}O AND ^{13}C STABLE ISOTOPES COMPOSITION OF THE CARBONATE ROCKS OF THE JERIBE FORMATION EASTERN IRAQ; AN APPROACH TO DEFINE THE PALEO TEMPERATURE AND PALEO DEPTH

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ABSTRACT

Stable isotopes $^{18}O/^{16}O$ and $^{13}C/^{12}C$ in the carbonate rocks of the Jeribe Formation are examined here to define the depositional characters in the basin includes paleo temperatures and paleo depth.

The Jeribe Formation comprises a transgressive unit belongs to the Latest Eocene-Recent Megasequence (Ap11). Complete sections of the formation were studied in the core of Himreen structure in eastern part of Iraq. The Himreen structure lies in the outer platform margin of unstable shelf of the Arabian plate. The stable isotope ($^{18}O/^{16}O$ and $^{13}C/^{12}C$) of Jeribe Formation provides the first records of paleotemperature for the Neogene (middle Miocene Transgression) of the Southern Teythes, which indicate cooling during deposition. Microfacies study imply that the skeletal grains are composed of green and red algae, stromatolite and verities of benthic foraminifera. The non-skeletal grains are pellets and intraclasts. Analysis of microfacies suggests lagoon, reef, shoal and open marine environments. The depletion of both ^{18}O and ^{13}C indicated shallowing at the contact with overlying and underlying formations in the shoal and lagoonal facies respectively. While in the middle part in the coral reef facies reveals elevation of the sea surface temperature. These suggest that the enrichment of both ^{18}O and ^{13}C at the cool water temperature associated with upwelling and arise of sea level represented by reef and open marine facies.

Keywords: O and C isotopes; Paleo temperature; Paleo depth; Jeribe Formation

INTRODUCTION

The Miocene Jeribe Formation is distributed from southeastern to northwestern Iraq. It was deposited in two main basins; Mosul and Kirkuk – Dezfol basins. The formation composed basically of dolomite and dolomitic limestone rocks representing a transgressive unit belong to the Latest Eocene-Recent Megasequence (Ap11) in the stratigraphic column of Iraq (Jassim and Goff, 2006). The megasequence (AP11) is subdivided into three sequences of latest Eocene – Oligocene, Early – Middle Miocene and Late Miocene-Recent age. Furthermore, Early – Mid Miocene Sequence is subdivided into two second order sequences; Early Miocene and Mid Miocene sequences. The Jeribe Formation comprises Middle Miocene Subcycle representing a new transgressive stage in the foredeep shelf margin below the evaporate lagoon sediments of Fatha Formation (Jassim and Goff, 2006).

Isotopes geochemistry covered wide applications of the O and C stable isotopes in geology using to define the physical and chemical conditions of the sea waters in the sedimentary basin. C and O isotopes give characteristic geochemical signatures of the sediments as well as the dolomite origin.

This paper aims to estimate the paleo depth and paleo temperatures of the sea waters during the sedimentation of the Jeribe carbonate rocks in the basin. Moreover, constructing relationships between the O and C isotopes composition with the sedimentological properties such as microfacies types and associations. Finally, suggesting an environmental model based on the isotopic results of paleo depth and paleo temperatures relative to the environmental setting, microfacies types and associations.

GEOLOGICAL SETTING

The Jeribe Formation composed of thick massive and bedded dolomite and dolomitic limestones. The Jeribe Formation is exposed in the core of Himreen, Makhool and Sinjar structures in the east, middle and northwestern Iraq, respectively (Fig. 1). Conformable contact was reported with underlying Dhiban Formation and overlying Fatha'h Formation (Jassim and Goff, 2006).

Paleogeography of the shelf margins during the Early to Middle Miocene characterized by development of broad and relatively shallow basins. The axis of the basin is passing parallel to the main trend of the structures, the Jezira Subzone

(submerged Khlesia High) and the Mesopotamian Zone. The successive deposits of the Jeribe Formation have coincided with continental collision between Arabian and Eurasian Plates extending from the Eocene up to the Recent time (Numan, 1997). This phase characterized by shallow epicontinental seas and lagoons. Tectonic classification of Fouad (2012) sited that the deposition of the Jeribe Formation is carried out in two main basins in unstable shelf. It is situated in the outer platform margin within Mesopotamia and western Zagros Fold – Thrust Belt in the Low Folded Zone.The study area is located within the Mesopotamian block (Jassim and Goff, 2006), confining between two transverse fault zone. Sirwan and Takhadide strike slip faults passing under the Kut – Dezfull strike slip Fault in the depocenter of Kirkuk – Dezful basin. The study area is effected by Makhul – Hemrin Fault.

METHODS AND MATERIALS

Two stratigraphic sections of Jeribe Formation are selected from Himreen structure east of Wasit and Dyiala Governorates in eastern part of Iraq. The first (Ja) is located in the SW core flank of Koolic1 anticline. The second (Jb) is located in core of Koolic (3) anticline in Showshareen valley (Fig. 1). The selected sections represent composite ideal column of Kirkuk-Dezfull basin. Forty-two (42) samples are collected represented the whole succession of the Jeribe Formation. Thin-section slides were prepared for petrographic study to identify mineralogical assemblages, textural components (skeletal and non-skeletal grains) followed the procedure listed in Tucker (1988).

XRF geochemical analysis of bulk carbonate samples is enhanced to determine the Ca, Sr and Mn concentrations in the XRF Laboratory, University of Baghdad.

Analysis of carbon and oxygen isotopic were carried out by Mass Spectrometry at Stable Isotopes Laboratories/ UK. Mass spectrometry is a method to obtain the relative numbers of each isotope of an atom in a 10 milligram powdered sample (Ebbing and Gammon 2009). The samples were weighed into clean Exetainer^{TM} tubes and flushed with 99.995% helium. After flushing, phosphoric acid was added to the samples for 3 hours at 90 °C and leave overnight to allow complete conversion of carbonate to CO_2. Reference and control materials were prepared in the same way. The CO_2 gas liberated from samples was analyzed by Continuous Flow-Isotope Ratio Mass Spectrometry (CF-IRMS). Carbon dioxide was sampled from the Exetainer^{TM} tubes into a continuously flowing He stream using a double holed needle. CO_2 gas was resolved on

a packed column gas chromatograph and the resultant peak is carried forward into the ion source of a Europa Scientific 20 – 20 IRMS where it is ionized and accelerated. Gas species of different mass are separated in a magnetic field and simultaneously measured using a Faraday cup collector array to measure the isotopomers of CO_2 at m/z 44, 45, and 46. Isotopic results are reported in per mil deviation from the V-PDB (Vienna Pee Dee Belemnite) using the standard delta notation.

The reference material used during analysis of samples is IA-R022 (Iso-Analytical working standard calcium carbonate, $\delta^{13}C_{V\text{-}PDB}$ = -28.63‰ and $\delta^{18}O_{V\text{-}PDB}$ = -22.69‰). IA-R022, NBS-18 (carbonatite, $\delta^{13}C_{V\text{-}PDB}$ = -5.01‰ and $\delta^{18}O_{V\text{-}PDB}$ = -23.20‰) and IA-R066 (chalk, $\delta^{13}C_{V\text{-}PDB}$ = +2.33‰ and $\delta^{18}O_{V\text{-}PDB}$ = -1.52‰) were run as quality control check samples.

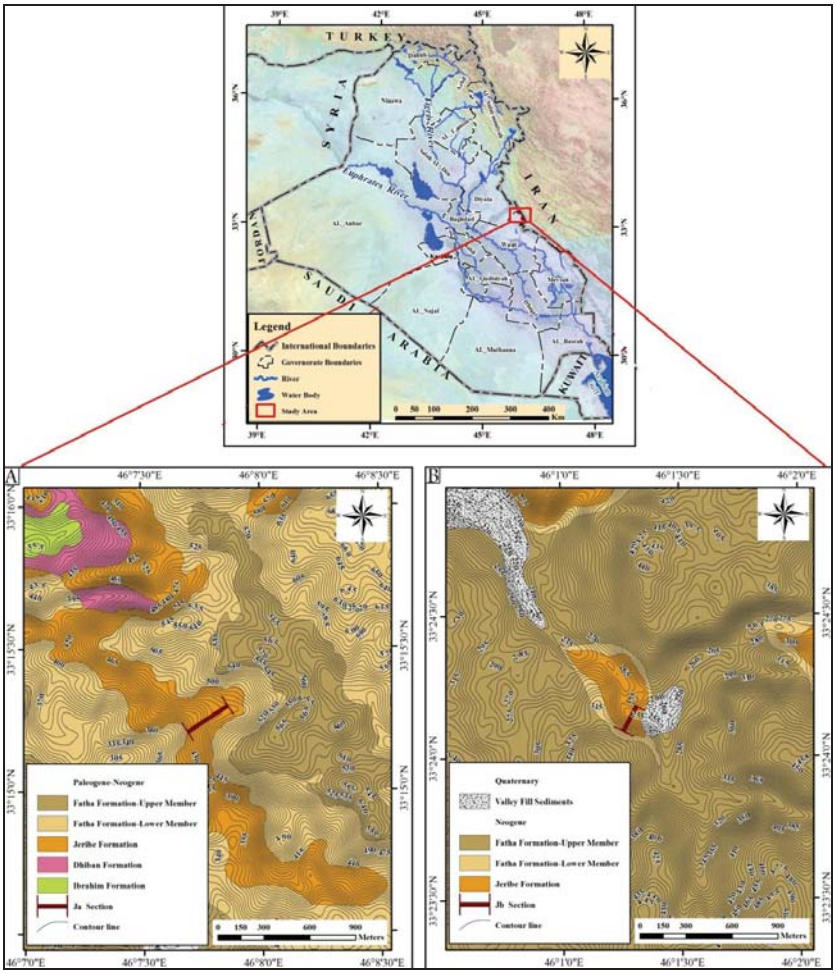

Fig. 1: Geological and location maps of the studied areas show the selected stratigraphic sections (red lines) (A) Ja, and (B) Jb

SEDIMENTOLOGY AND MICROFACIES ANALYSIS

Six microfacies types are identified in the carbonate rocks of the Jeribe Formation, these are; framestone, boundstone, grainstone, packstone, wackestone and mudstone (Table 1). These microfacies are grouped in four facies associations comprised depositional environment and paleoecology conditions (Flügel, 2010).

According to petrographic analysis, the carbonate rocks are composed of skeletal and non-skeletal grains. The identification of the carbonate components is based on Scholle and Ulmer-Scholle (2003) and Flügel (2010). The microfacies classifications are followed Dunham (1962), Embry and Kloven (1971) and James (1984) (Fig. 2 and 3).

Table 1: Microfacies types and the depositional environment of the Jeribe Formation

Main Facies	Characteristics	Classification	Depositional environment
Framestone	Domal Stromatolite, Densely laminated	Stromatolite Framestone	Reef
Boundstone/ Grainstone	Red algal and stromatolite	Coralline red algae Boundstone/ Grainstone	Reef
Boundstone	coral with less algae stromatolite, and benthic foraminifera	Coral Boundstone	Reef
Grainstone	Red algal with stromatolite, and benthic foraminifera	Red algal grainstone	Shoal
Packstone	Benthic foraminifera, shell fragments and pellets	Bioclasts algae dolomicrite packstone	Open marine
		Red algae packstone	Back reef
		Peloidal dolomicrite packstone	Shoal
Wackestone	Algae and less assemblage of benthic	Bioclasts dolomicrite wackestone	Open marine
		Algae dolomicrite mudstone-wackestone	Lagoon
		Pelletal Algae dolomicrite wackestone	Lagoon
Mudstone	Molluscs shell and algae	Pelletal Algae Mudstone	Lagoon
		lamination mudstone	Lagoon

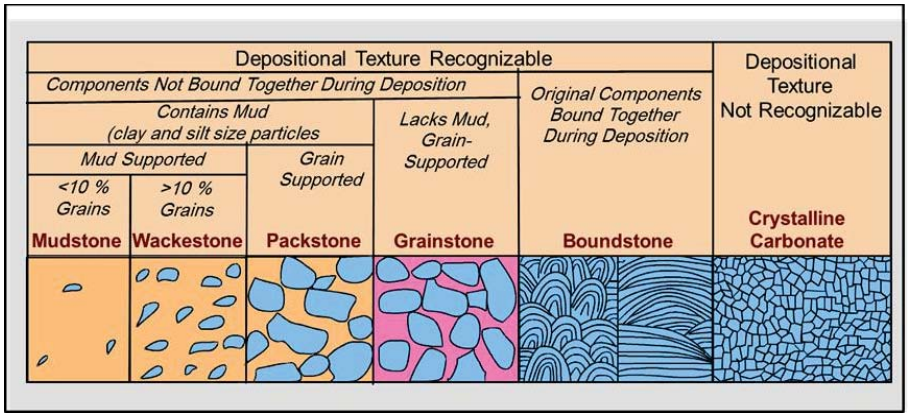

Fig. 2: Classification of Dunham (1962)

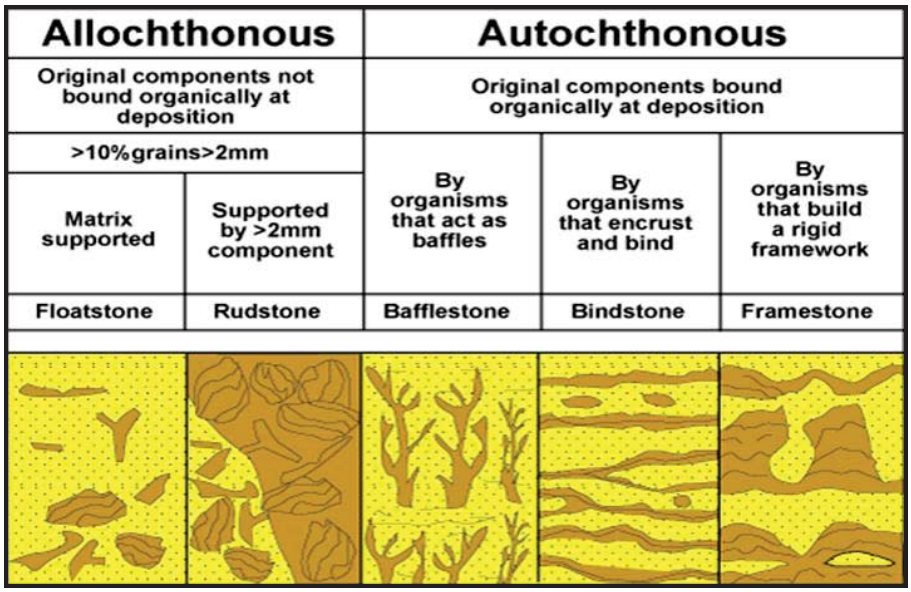

Fig. 3: Classification of reef after Embry and Kloven (1971) and James (1984)

The skeletal grains composed of green and red algae, stromatolite and verities of benthic foraminifera. The non-skeletal grains are composed of pellets and intraclasts (Fig. 4). Facies associations reveal four depositional settings in shallow marine environment, these are; lagoon, reef, shoal and open marine. The lower part of the Jeribe Formation composed of mudstone to wackestone microfacies. In the middle part of reef setting composed of binding coral and red algae interbedded with packstone to grainstone microfacies. The upper part is characterized by domal stromatolite. The identified facies associations are discussed in the followings:

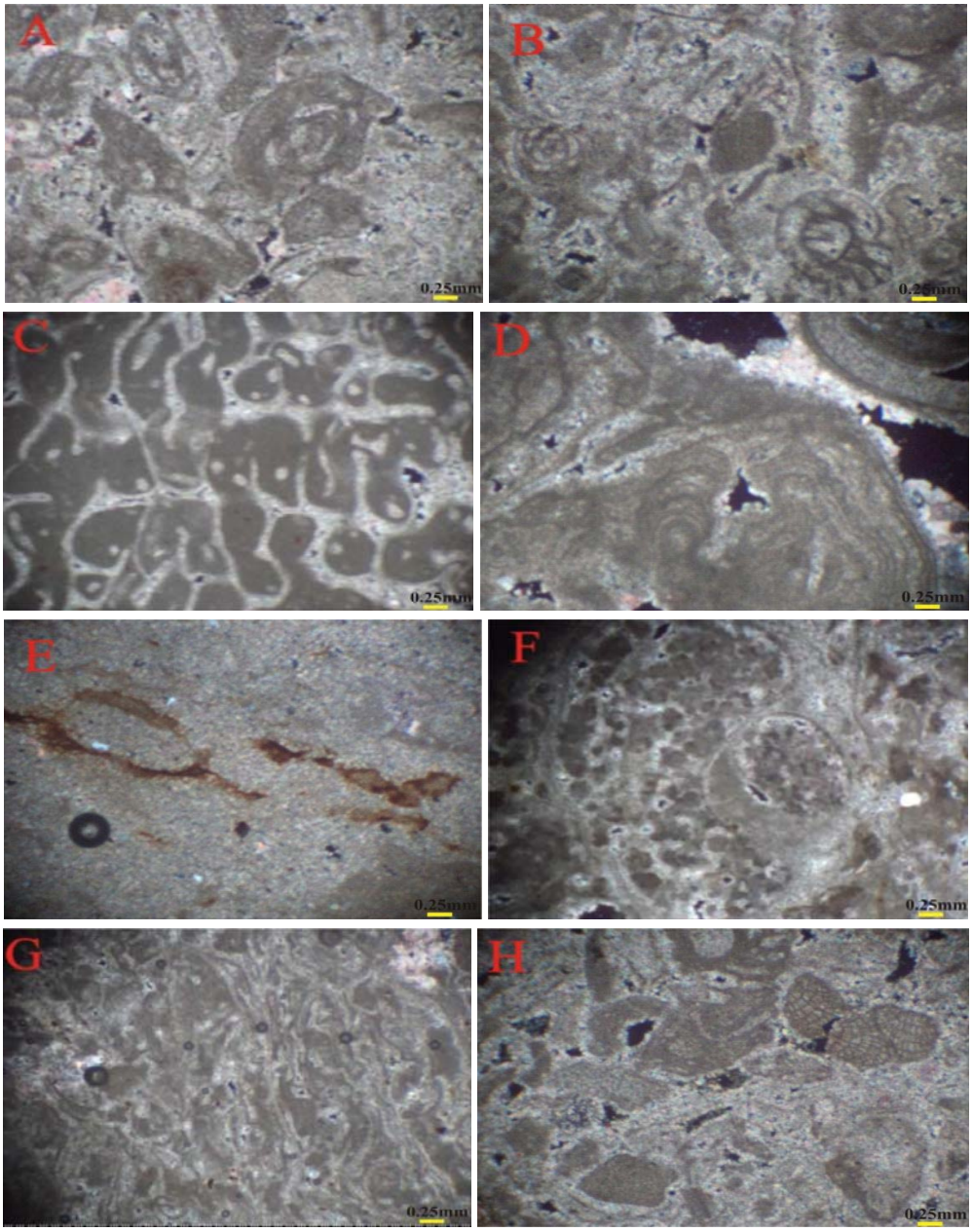

Fig. 4: Photomicrographs of different microfacies types in the Jeribe Formation (x40X)

- **A.** Wackestone microfacies identified in the lower part of Jb.
- **B.** Wackestone/ packstone microfacies identified in lower part of Jb.
- **C.** Coral boundstone microfacies in the middle part of Ja.
- **D.** Stromatolite framestone microfacies in the upper part of Ja.
- **E.** Laminated mudstone/ wackestone microfacies in the lower part of Ja.
- **F.** Pelletal packstone microfacies in the lower part of Ja.
- **G.** Grainstone microfacies in the middle part of Jb.
- **H.** Reworking wackestone microfacies in the middle part of Jb.

– **FA1; shallow marine/ lagoon:** FA1 is characterized by algal dolomicrite muddy wackestone, pelletal algal dolomicrite wackestone and bioclasts dolomicrite wackestone microfacies. FA1 composed mainly of algae, mollusca shells and subangular fine quartz grains. Pelletal algal dolomicrite mudstone is characterized by laminated algae with some pellets in micritic groundmass. Petrographically, FA1 consists of medium crystalline dolomite with very rare biomoldic fossils. Micritization, dissolution and dolomitization are the main effective diagenetic processes (Fig. 4A and E).

– **FA2; Inner shelf:** FA2 composed of pelletal packstone abundant in foraminifera, mollusca and laminated algae, which are supported by micritic mud. While bioclasts grainstone microfacies contains red alga of Archaeolithothamnium sp., and predominance of milioliods idntified in the middle part of the formation. The major allochem are pellets and benthic foraminifera. The bioclastic dolomicrite grainstone is less common microfacies in the middle part. FA2 composed of very fine crystalline dolomite mixed with calcite cement (Fig. 4F).

– **FA3; shallow, open-marine:** FA3 is characterized by wackestone and packstone microfacies. These are composed of very fine crystalline dolomite with calcite cement mostly of secondary origin. This microfacies is dominated by benthonic foraminifera of soritidae, alveolinidae and miliolids species with red algal type. FA3 is characterized by bioclastic dolomicrite packstone, red algae packstone and peliodal packstone, which are repeated at various levels through the studied sections and in the upper contact with Fatha'h Formation, (Fig. 4B and H).

– **FA4; Shallow/ reef-back reef:** FA4 is characterized by framestone and boundstone microfacies recognized in the middle part of the formation. It is subdivided into coral framestone (the framework composed of red calcareous coralline algae), and coralline red algal boundstone (composed of coral, stromatolite and benthic foraminifera). The stromatolite framestone microfacies is characterized by laminated domal stromatolite and existed in upper part of Ja section. It is associated with coral and laminated algal mat in micritic matrix (Fig. 4D and G).

Oxygen and Carbon Isotopes Composition

The most abundant stable isotopes of oxygen are ^{18}O and ^{16}O and carbon is ^{13}C and ^{12}C. The concentration of ^{18}O and ^{13}C in a sample is conventionally estimated as per mil. $\delta^{18}O$ and $\delta^{13}C$ are reported between the isotope ratios in the sample and those in the international Pee Dee Belemnite (PDB) standard which, by definition has $\delta^{18}O$ and $\delta^{13}C$ values of 0‰ (Hudson, 1977).

Composition of oxygen isotope values in the carbonate rocks of the Jeribe Formation show significant variation ranging from (-1.15 – 2.86) ‰. The majority of samples are enriched by ^{18}O, which are approximately the same values recorded by Veizer and Hoefs (1976) for the Tertiary period. While carbon isotope composition shows significant variation ranging from -1.14–2.42‰, which are approximately the same values recoded by Veizer and Hoefs (1976), for Tertiary period. These are compared with compiled data in an attempt to verify to previously reported measurements of pre-Quaternary rocks. He was measured both O and C isotopes in both limestone and dolomite rocks (Fig. 5) and (Table 2).

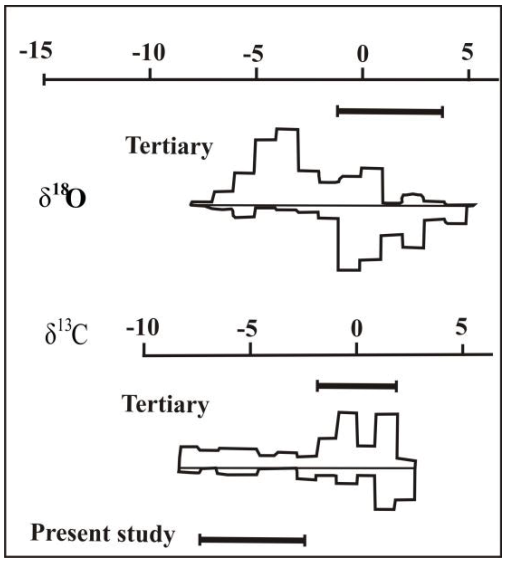

Fig. 5: $\delta^{18}O$ of the Jeribe Formation compared with the Tertiary period after Veizer and Hoefs (1976)

Table 2: Analytical results of $\delta^{18}O$ and $\delta^{13}C$ in the carbonate rocks of the Jeribe Formation

Sample	Result d-$^{13}C_{V\text{-}PDB}$	Result d-$^{18}O_{V\text{-}PDB}$	Sample	Result d-$^{13}C_{V\text{-}PDB}$	Result d-$^{18}O_{V\text{-}PDB}$
Identification	(‰)	(‰)	Identification	(‰)	(‰)
JA3	1.30	1.23	JB9	2.74	2.67
JA9	0.71	1.18	JB11	-0.45	1.49
JA13	-0.14	1.17	JB15	2.14	1.50
JA17	1.01	1.21	JB17	2.26	2.66
JA21	1.80	1.47	JB21	2.37	2.78
JA23	-0.87	0.12	JB27	2.14	1.75
JA25	1.07	1.62	JB31	1.65	0.99
JA31	1.59	2.80	JB37	1.31	1.22
JA40	-1.07	-0.97	JB39	0.05	0.45
JB7	2.44	2.89	JB42	-1.14	-1.15

$\delta^{18}O$, $\delta^{13}C$ and Mn/Sr Ratio

Bivariate plots involving $\delta^{18}O$ and $\delta^{13}C$ are common and convenient way of distinguishing the depositional and/ or diagenetic (paleo) environments responsible for carbonate formation (Nelson and Smith, 1996). $\delta^{18}O$ and $\delta^{13}C$ cross- plot was prepared by Hudson (1977), who distinguished a number of characteristic isotope fields for carbonate having different origins. Many workers have been followed, adapted and extended such as $\delta^{18}O - \delta^{13}C$ plots (Bathurst, 1980; Choquette and James, 1987; Moor, 1989; and Morse and Makenzie, 1990). A modified version of Hudson's plots reproduced by Nelson and Smith (1996).

The plot data of $\delta^{18}O$ and $\delta^{13}C$ from the Jeribe Formation display unaltered samples and have primary signature of isotopes (Fig. 6). Mn/Sr ratios are estimated in the same samples as the stable isotope measurements. Kaufman and Knoll (1995) observed that both limestones and dolostones with Mn/Sr <l0 commonly retain near-primary $\delta^{13}C$, while in the Jeribe Formation Mn/Sr range from 0.19 to 0.64 with average 0.3 (Table 3). Ratio and plotted $\delta^{18}O$ and $\delta^{13}C$ infer primary proxy for the isotope composition.

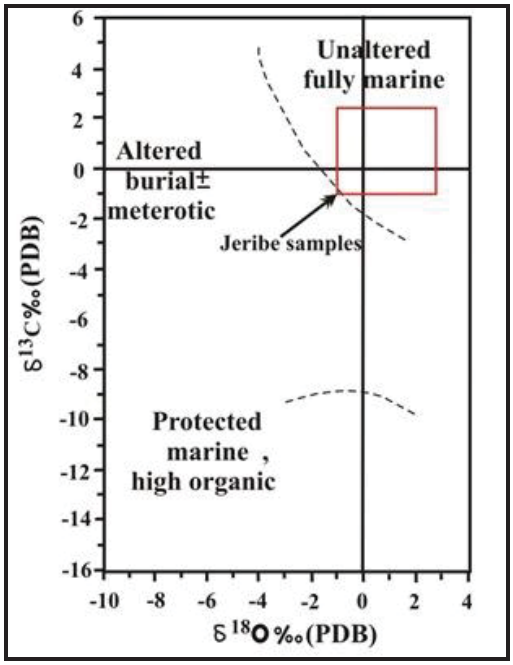

Fig. 6: Shows $\delta^{18}O$ and $\delta^{13}C$ plot (Nelson and Smith, 1996)

Table 3: Geochemical parameters and estimated paleo temperature

Sample no.	Temp. C	$\delta^{18}O$ vsmow	$\Delta\delta^{18}O$	Mn ppm	Sr ppm	Mn/Sr
JA3	21.6	2.1	-0.87	36.55	139.18	0.64
JA9	14.5	2.1	-0.92	38.26	197.99	0.57
JA13	11.4	2.1	-0.93	26.87	213.51	0.55
JA17	12.3	2.1	-0.89	39.80	193.21	0.39
JA21	9.4	2.4	-0.92	27.10	118.29	0.44
JA23	5.8	1.04	0.92	40.19	288.34	0.23
JA25	6.0	2.5	0.88	26.9	104.26	0.24
JA31	10.3	3.8	1.00	29.66	129.88	0.51
JA40	10.3	-0.07	0.90	50.49	281.54	0.25
JB42	6.0	2.7	0.16	96.58	149.85	0.25
JB39	5.3	1.9	0.77	71.02	123.82	0.34
JB37	11.4	2.1	-0.61	61.65	110.85	0.26
JB31	11.6	1.3	0.20	45.30	113.56	0.19
JB27	11.6	-0.2	2.86	60.02	136.39	0.12
JB21	11.5	3.6	-0.89	31.9	136.22	0.20
JB17	10.4	2.4	-0.65	30.20	123.54	0.22
JB15	15.9	2.4	-1.41	49.87	95.97	0.13
JB11	9.8	3.6	-2.38	34.07	131.99	0.25
JB9	5.5	3.7	-3.25	25.32	100.96	0.22
JB7	20.7	2.7	-3.85	27.64	81.26	0.09

$\delta^{18}O$ as Paleo Temperature Indicator

The $^{18}O/^{16}O$ ratio is measured as $\delta\ ^{18}O$ in the waters, which is mostly controlled by temperature, where salinity can have an impact as well (Wanamaker *et al*., 2007). A relationship was established between water temperature and the $^{18}O/^{16}O$ ratio, which can be used to reconstruct the sea-surface temperatures and seasonal variability (Yan *et al*., 2013). Emiliani and Epstein (1953) were firstly developed equation for calcite paleo temperature that used to calculate the "isotope temperature" of calcite formation by providing $\delta^{18}O$ value of the water from which the calcite precipitated. The equation is:

$$\mathbf{T = 16.5 - 4.3\ \delta + 0.14\ \delta 2}$$

where T is the temperature in °C (based on a least-squares fit for a range of temperature values between 9 °C and 29 °C, with a standard deviation of ± 0.6 °C, and δ is $\delta^{18}O$ for a calcium carbonate sample. This equation has since undergone many revisions by Rye and Sommer (1980). The results of the studies by Epstein *et al*. (1951) and Emiliani and Epstein (1953) have implications that are still relevant today in archaeology.

The estimated paleo temperature of the carbonate rocks of the Jeribe Formation are ranged from (5.3 – 15.9) and few samples show high temperature (Table 3).

Paleo Depth and Pale Climate

$\delta^{18}O$ composition of Miocene seawater given as calculated $\delta^{18}O_{sw}$ values by the following equation:

$$\delta^{18}Ovsmow = 1.03092\ \delta^{18}Ovpdb + 30.92$$

Where $\delta^{18}Ovsmow$ is Vienna Standard Mean Ocean Water, $\delta^{18}Ovpdb$ is Vienna Pee Dee Belemnite (VPDB), **1.03092** and **30.92** are constant. The $\delta^{18}O$ of seawater can vary with time due to several processes, which influence the $\delta^{18}O$ of the global ocean, and the local $d\delta^{18}O$ of seawater. The $\delta^{18}O$ of the global ocean is primarily influenced by the changes in amount of stored water as ice on land, which influences the $\delta^{18}O$ of the global ocean on the timescale (Shackleton, 1967), and by the changes in temperature-dependent isotopic exchange with the oceanic crust (Gregory and Taylor, 1981; Muehlenbachs and Clayton, 1976).

The dominant factor influenced the $\delta^{18}O$ of the global ocean is related to paleo climatic changes. The enrichment of ^{18}O in sea water suggest glacial period at the middle Miocene.

During glacial periods, low $\delta^{18}O$ in waters is stored in the ice sheets, and the mean $\delta^{18}O$ value of the world oceans is relatively high (Zachos *et al*., 2011), which mean enrichment of ^{18}O in the sea water during glacial period. Ando *et al*. (2010) used $\Delta\delta^{18}O$ for determination the paleo depth of water by the equation:

$$\Delta\delta^{18}O = \delta^{18}O_{Calcite} - \delta^{18}O_{SMOW}$$

The estimated paleo-depth of the sediments in the Jeribe Formation suggests thermocline and shallowing upwards represented by mixed layer within water column except sample Jb 11, which shows depletion in ^{13}C suggest deeper depth. While Jb 27 shows enrichment in ^{13}C reflect minimal deeper (Fig. 7 and Table 3).

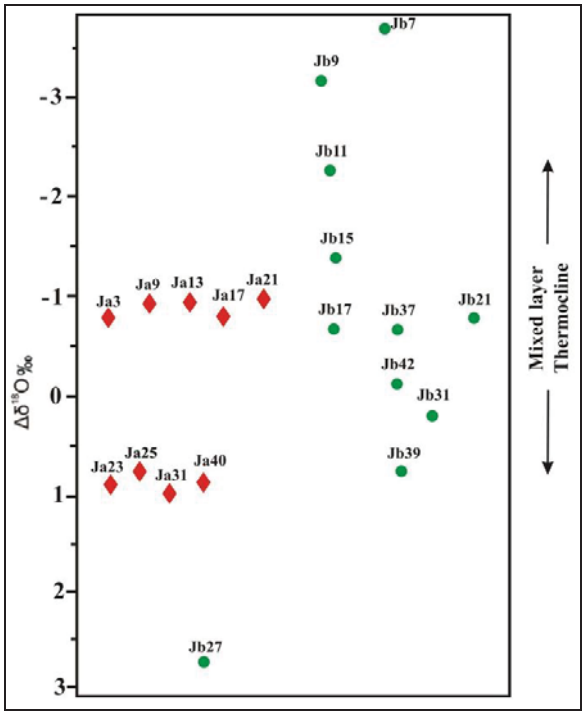

Fig. 7: Shows $\Delta\delta^{18}O$ of determine paleo depth in the water column, (after Ando *et al*., 2010)

DISSUASION AND INTERPRETATION

The C and O isotopic ($\delta^{13}C$ and $\delta^{18}O$) compositions used as main geochemical tools to reconstruct the paleoecology of carbonate minerals, water depth and presence/ absence of algal photo symbionts compositions (Edgar *et al*., 2015). Hoefs (2015) pointed that well preserved textures and trace element contents have recorded the primary oxygen isotope composition and can be used to deduce the past ocean composition. The Mn/Sr ratio of less than (2) indicates no significant influence of diagenetic process (Jacobsen and Kaufman, 1999; Marquillas *et al*., 2007; Nagarajan *et al*,. 2008; and Kano *et al*., 2007). The estimated low Mn/Sr ratios (0.09 – 0.64) (Table 3) in the carbonate rocks indicate less effects during diagenesis and have primary isotope signature. Ratios and plotted $\delta^{18}O$ and $\delta^{13}C$ together with well-preserved texture infer primary proxy for the isotope composition.

The $\delta^{18}O$ values of carbonate rocks primarily depends on the $\delta^{18}O$ of the seawater ($\delta^{18}O$ sw) and the temperature of the depositional environment. While carbon isotope data linked to marine bicarbonate and biological kinetic isotope fractionation, which is not sensitive to temperature changes as in $\delta^{18}O$. It indicates the magnitude and the time

of upwelling events (Jones *et al*., 2012). The **$\delta^{18}O$** of seawater can be vary with time due to several processes, which influence the **$\delta^{18}O$** of the global ocean related to paleo climate. Since the fractionation is based on temperature by evaporation of lighter ^{16}O to cause enrichment of ^{18}O in times of cold sea water. Isotopic results of the Jeribe Formation show that the seawater is relatively enriched in ^{18}O, which is most probably marked cold climate e.g. glacial period event. (Fig. 8).

Most of $\delta\ ^{18}O$ and $\delta\ ^{13}C$ of the Jeribe samples are clustered at small positive values with extension of small negative values for both $\delta\ ^{18}O$ and $\delta\ ^{13}C$. The group of samples of small positive $\delta\ ^{18}O$ and $\delta\ ^{13}C$, suggest mainly precipitated of carbonate close to isotopic equilibrium with ambient shelf water (Roa and Nelson, 1992). The $\delta^{18}O$ isotope is increasing with depth according to decreases in temperature, in which oxygen isotope fractionation between ambient seawater and foraminiferal calcite during calcification is strongly depends on temperature (Bemis and Howard, 1998; Emiliani and Epstein (1953); and Pearson, 2012). The C and O isotopic variations are demonstrating facies controlled (Veizer and Hoefs, 1976). Three samples show variation in $\delta^{18}O$ and $\delta^{13}C$ composition because non-equilibrium precipitation owing to very shallow water, where the environment influences by freshwater runoff, warm temperatures and algal activity. The curve of C and O isotopes shows three negative excursions in the lower part of lagoon environment at the contact with underlying Dihban Formation. The lower part is characterized by abundant calcareous foraminifers, *mollusca* and calcareous algae. The other two samples of the middle part show negative value of both ^{18}O and $\delta\ ^{13}C$ which are associated with coral reef environment, which is virtually originally calcitic. The samples of the upper part of the formation represent shoal environment, which is characterized by calcareous peloids suggest short term shallowing event in Middle Miocene basin. The most negative values of oxygen isotope are observed in the lower and upper parts of the formation, suggest non-steady state in sedimentary basin, most probably due to tectonic effect (Fig. 9).

The interpretation of paleo-depth of the Jeribe isotopic is based on uses of **$\Delta\delta^{18}O$** fractionation plots after Ando *et al*. (2010), which is inaccurate due to rejection of $\delta^{13}C$ results. Edgar *et al*. (2015) was mentioned to use $\delta\ ^{13}C$ aid to determine comparatively paleo depth. Their utility arises because ^{12}C is preferentially utilized by photo symbionts and phytoplankton, leaving the foraminiferal microenvironment and ambient seawater

enriched in ^{13}C. Below the photic zone, the ^{12}C of the water column increases relative to surface waters as a function of reduce photosynthetic activity and remineralization of ^{12}C enriched-organic matter. This is leading to decrease $\delta^{13}C$ values in the test of microfossils (many foraminifera are lacking photo symbionts) (Figure 10).

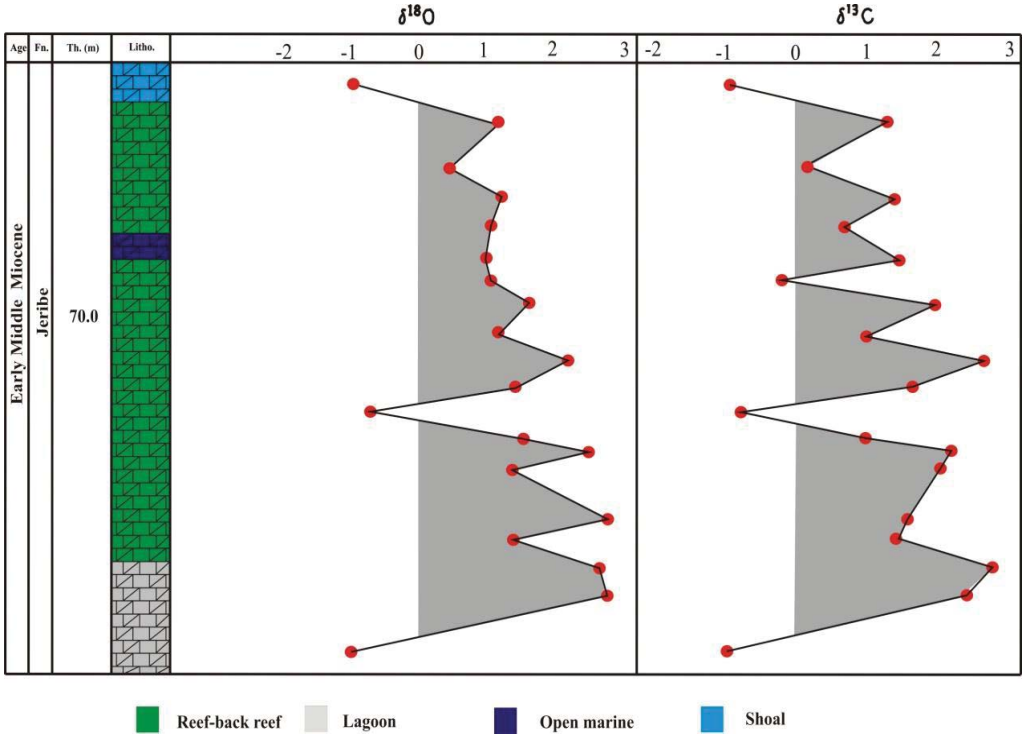

Fig. 8: Shows distribution of δ ^{13}C and δ ^{18}O through the stratigraphic log of the studied sections of the Jeribe Formation and their temperature

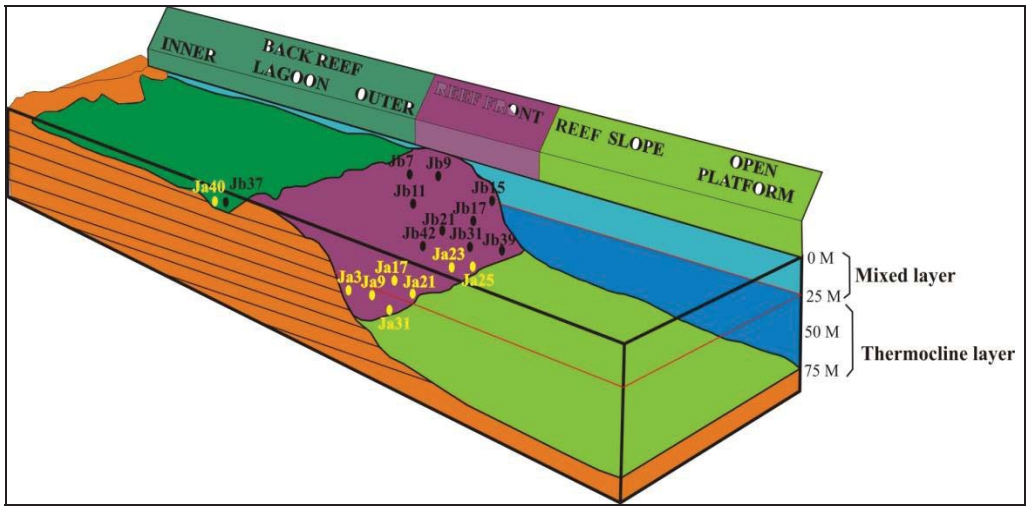

Fig. 9: Sketch diagram shows paleo depth and environment subdivisions of Jeribe Formation in Miocene Basin

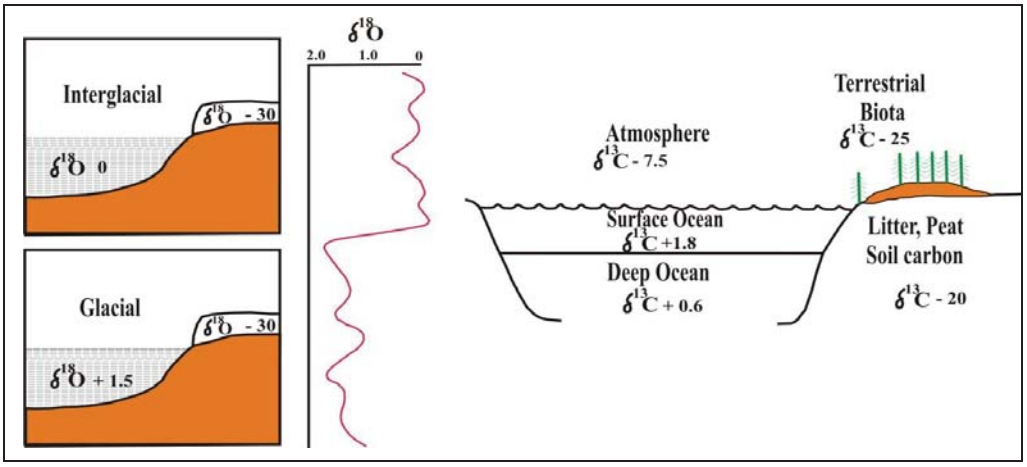

Fig. 10: ^{13}C composition from ^{13}C cycle and ^{18}O composition in glacial and interglacial event, after (Zachos *et al.*, 2011)

Nelson and Smith (1996) suggest that the ^{13}C isotope is enriched in shallow water condition due to concentration of bicarbonate more than deep water. They pointed that increases of $\delta^{13}C$ is accompanied with sea-level rise, while decreases in $\delta^{13}C$ is accompanied to sea-level fall.

The carbonate rocks of the Jeribe Formation show enrichment in $\delta^{18}O$ and $\delta^{13}C$, which suggests upwelling and cool temperate environment represented by open marine and back reef-reef facies. These facies are characterized by coralline red algae composed essentially of High-Mg calcite, while phylloid algae are composed of High-Mg calcite. Some of benthic foraminifera are common to have both High-Mg and Low-Mg calcite. The depletion in $\delta^{18}O$ and $\delta^{13}C$ of Jeribe carbonate rocks (distinct negative oxygen isotope values) are commonly exposed to meteoric water before they have reached mineralogical stability with relatively high temperature and atmospheric CO_2 exchange. This suggests mixed layers as a paleo-depth in water column (Figs. 8 and 9).

CONCLUSIONS

The variable composition of $\delta^{13}C$ and $\delta^{18}O$ of the Jeribe carbonate rocks is controlled by sea level oscillation and facies types. Depletion of both $\delta^{13}C$ and $\delta^{18}O$ isotopes are associated with drop of sea level and restricted to shallow facies. While enrichment of both $\delta^{13}C$ and $\delta^{18}O$ isotopes is associated with sea level rise. The Middle Miocene is a relatively steady period of cool water, which is indicated by enrichment of stable isotopes. Furthermore, implies that transgression and deepening of water depth was took place at amidst of mixed layer and thermocline zone.

REFRENCES

Ando, A., Huber, B.T. and MacLeod, K.G., 2010. Depth-habitat reorganization of planktonic foraminifera across the Albian/Cenomanian boundary, Paleobiology, 36: 357 – 373.

Bathurst, R.G.C., 1980. Deep crustal diagenesis in limestones: Revista del fnstituto de Investaciones Geologicas, Universit of Barcelona, 89 – 100.

Bemis, E.B. and Howard, J.S., 1998. Reevaluation of the oxygen isotopic composition of planktonic foraminifera: Experimental results and revised paleotemperature equations. Paleoceanography, 13 (2): 150 – 160.

Choquette, P.W. and James, N.P., 1987. Diagenesis 12: Diagenesis in limestones - 3. The deep burial environment: Geoscience Canada, 14 (I): 3 – 35.

Dunham, R.J., 1962. Classification of carbonate rocks according to depositional texture. AAPG, 1: 108 – 121.

Epstein, S., Buchsbaum, R., Lowenstam, H.A. C Urey, H., 1951. Revised Carbonate-Water Isotopic Temperature Scale, GSA Bulletin, 64 (11): 1315 – 1326.

Emiliani, C. and Epstein, S., 1953. Temperature variations in the lower Pleistocene of Southern California. J. Geo!., 61: 171 – 181.

Ebbing, D.D. and Gammon, S.D., 2009. General Chemistry, 10th Edition, ISBN 10.

Embry III, A.F. and Kloven, J.E., 1971. Alate Devonianreef tracton northeastern BankIsland, NWT, Bulletin of Canadian Petroleum Geology, 19: 730 – 781.

Edgar, K.M., Anagnostou, E.N., Pearson P. and Foster, G.L., 2015. Assessing the impact of diagenesis on δ11B, δ13C, δ18O, Sr/Ca and B/Ca values in fossil planktic foraminiferal calcite, Geochimica et Cosmochimica Acta, ELSEVER, Volume 166: 189 – 209.

Fouad, S.F.A., 2012. Western Zagros Fold – Thrust Belt, Part I: The Low Folded Zone. Iraq bull. Geol. Min, Special Issue (5): 39 – 62.

Flügel, E., 2010. Microfacies of Carbonate Rocks, Analysis, Interpretation and Application. Springer-Verlag, Berlin, 976pp.

Gregory, R.T. and Taylor, H.P., 1981. An oxygen isotopic profile in a section of cretaceaous oceanic crust, Samali ophiolite: Evidence for d18O buffering of the oceans by deep (>5 Km) sea waterhydrothermal circulation at midocean ridges. Journal of Geophysical Research, 86: 2737 – 2755.

Hudson, J.D., 1977. Stable isotopes and limestone lithification. Journal of the Geological Society of London 133: 637 660pp.

James, N.P., 1984. Shallowing-upward sequences in carbonates. In: Walker RG (ed) Facies models. Geosci Can Reprint Ser 1: 213 – 228.

Jacobsen, S. and Kaufman, A., 1999. The Sr, C and O isotopic evolution of Neoproterozoic seawater. Chemical Geology 161: 37 – 57.

Jones, J.R., Mulville, J.A., McGill, R.A.R. and Evershed, R.P., 2012. Palaeoenvironmental modelling of $\delta^{13}C$ and $\delta^{15}N$ values in the north Atlantic Islands: understanding past marine resource use, Rapid Commun. Mass Spectrom, 26: 2399 – 2406.

Jassim, S.Z. and Goff, J.C., 2006. Geology of Iraq. Dolin, Prague and Moravian Museum, Berno, 177 – 184.

Kaufman, A.J. and Knoll, A.H., 1995. Neoproterozoic variations in the carbon isotopic composition of seawater: Stratigraphic and biogeochemical implications. Precambrian Research, 73: 27 – 49.

Kano, A., Kakizaki, Y., Takashima, C., Wei Wang, and Ryo Matsumoto, 2007. Facies and depositional environment of the uppermost Jurassic stromatoporoid biostromes in the Zagros Mountains of Iran, Journal of the Geological Society of Sweden, 129 Issue (2): 107 – 112.

Marquillas, R.A., Sabino, I., Sial, A.N., Del Papa, C., Ferreira, V. and Matthews, S., 2007. Carbon and oxygen isotopes of Maastrichtian-Danian shallow marine carbonates: Yacoraite Formation, northwestern Argentina. Rio de Janeiro, Journal of South American Earth Sciences, 23 (4): 304 – 320.

Muehlenbachs, K. and Clayton, R.N., 1976. Oxygen isotope composition of the oceanic crust and its bearing on seawater. Journal of Geophysical Research, 81: 4365 – 4369.

Nagarajan, R., Sial, A.N., Armstrong-Altrin, J.S., Madhavaraju, J. and Nagendra, R., 2008. Carbon and Oxygen isotope geochemistry of Neoproterozoic limestones of the Shahabad Formation, Bhimabasin, Karnataka, Southern India. Rev Mex Cienc Geol 25 (2): 225 – 235.

Numan, N.M.S., 1997. A plate tectonic scenario for the Phanerozoic succession in Iraq, Iraq Geological Journal, 30 (2): 85 – 110.

Rye, D.M. and Sommer II, M.A., 1980. Reconstructing paleotemperature and paleo- salinity regimes with oxygen isotopes. In: D.C. Rhoads and R.A. Lutz (Editors), Skeletal Growth of Aquatic Organisms. Plenum Press, New York, N.Y., 169 – 202.

Shackleton, N.J., 1967. Oxygen isotope analyses and Pleistocene temperatures reassessed. Nature, 215: 15 – 17.

Scholle, P.A. and Ulmer-Scholle, D.S., 2003. A Color Guide to the Petrography of Carbonate Rocks: Grains, Textures, Porosity, Diagenesis. AAPG Memoir 77, 459pp.

Veizer, J. and Hoefs, J., 1976. Nature of O18-O16 and C13-C12 Secular Trends in Sedimentary Carbonate Rocks. Geochimicaet Cosmochimica Acta, 40 (11): 1,387 – 1,395. doi: 10.1016/0016-7037(76)90129-0

Wanamaker, A.D., Kreutz, Karl, J., Borns, Harold, W., Introne, Douglas, S., Feindel, Scott; Rawson, Paul, D., and Barber, Bruce, J., 2007. Experimental determination of salinity, temperature, growth, and metabolic effects on shell isotope chemistry of Mytilus edulis collected from Maine and Greenland. Paleoceanography 22 (2): PA2217

Morse, J.W. and Mackenzie, F.T., 1990. Geochemistry of Sedimentary Carbonates, Elsevier, New York, 707pp.

Moore, C.H., 1989. Carbonate diagenesis and porosity. Developments in sedimentology 46. Amsterdam, Elsevier.

Zachos, J.C., Pagani, M., Sloan, L., Thomas, E. and Billups, K., 2011. Trends, rhythms and aberrations in global climate 65ma to present, Science, 292: 686 – 693.

Hoefs, J., 2015. Stable isotopes geochemistry, 7th edition, ISBN 978-3-319-19716-6

Rao, C. and Nelson, C.S., 1992. Oxygen and carbon isotope fields for temperate shelf carbonates from Tasmania and New Zealand . Marine Geology, 103 Issue (1 – 3): 273 – 286.

Tucker, M.E., 1988. Techniques in Sedimentology, Oxford, London, 394pp.

Nelson, C.S. and Smith, A.M., 1996. Stable oxygen and carbon isotope compositional fields for skeletal and diagenetic components in New Zealand Cenozoic nontropical carbonate sediments and limestones: a synthesis and review. New Zealand Journal of Geology and Geophysics, 39: 93 – 107.

Pearson, N.P., 2012. Oxygen isotopes in foraminifera: Overview and historical review. Paleontological Society Papers 18, 1 – 38.

Yan, X., Fang, Z.T., Hong, F.Z. and Yang, W., 2013. Large magnesium isotope fractionation in peridotite xenoliths from eastern North China craton: Product of melt–rock interaction, Geochimica et Cosmochimica Acta, 115: 241 – 261