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Paleozoic marine sedimentation and associated oolitic iron-rich deposits, Tassilis N'Ajjer and Illizi Basin, Saharan platform, Algeria

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ABSTRACT

The inventory of the Devonian oolitic ironstones of the Central Sahara (Tassilis N'Ajjer and Illizi Basin) indicates the occurrence of 8 beds of ironstones, interbedded with the argillaceous, silty and sandy marine formations of the northern cover of the Precambrian Ahaggar Shield.

Related to major or minor sedimentary regressive cycles, more or less induced by eustatic variations, these ironstones have been accumulated in tranquil conditions, by intrasedimentary concretion. Chamositic and hematitic ooids (more or less goethitic) occur in three petrographic ironstone facies: detrital (FOD), cemented (FOC) and microconglomeratic (FMC). The average chemical composition is: $SiO_2 = 42.59\%$; $Al_2O_3 = 3.74\%$; $Fe_2O_3 = 39.32\%$; $FeO = 0.10\%$; $P_2O_5 = 1.76\%$; Loss on Ignition = 7.05%. An enrichment of some elements, namely Fe_2O_3, MnO, Bi, Cd, Ni, Pb and S, has been recorded.

The iron source seems to have been located in the southern regions of the Nigerian and Congo Shields, whereas the ironstone deposits were accumulated under temperate latitudes (30–60°S) before the breakup of Gondwana.

RÉSUMÉ

L'inventaire des minerais de fer oolithiques dévoniens du Sahara Central (Tassilis N'Ajjer et Bassin d'Illizi) a permis de mettre en évidence 8 niveaux de minerai interstratifiés dans les formations marines argilo-siltogréseuses de la couverture septentrionale du bouclier précambrien de l'Ahaggar.

Liés à des cycles sédimentaires régressifs majeurs ou mineurs, plus ou moins induits par des variations eustatiques, ces minerais de fer oolithiques se sont accumulés en milieu calme, par concrétionnement intrasédimentaire. Les oolithes de chamosite et d'hématite (plus ou moins goethitisées) s'organisent en trois faciès pétrographiques: détritique (FOD), cimenté (FOC) et microconglomératique (FMC).

La composition chimique moyenne est la suivante: $SiO_2 = 42{,}59\%$; $Al_2O_3 = 3{,}74\%$; $Fe_2O_3 = 39{,}32\%$; $FeO = 0{,}10\%$; $P_2O_5 = 1{,}76\%$; Perte au Feu = 7,05%. La comparaison de cette composition moyenne avec celle de la croûte terrestre indique un enrichissement en Fe_2O_3, MnO, Bi, Cd, Ni, Pb et S dans les minerais de fer.

La source du fer est à rechercher plus au sud, dans les boucliers du Nigéria et du Congo, tandis que les dépôts eux-mêmes semblent s'être effectués sous des latitudes tempérées (30–60°S) avant l'éclatement du Gondwana.

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1. Introduction

Within the Devonian shallow marine sediments of the Tassilis N'Ajjer and Illizi Basin (Central Sahara, Algeria; Fig. 1), oolitic ironstones occur, which are components of the North African Paleozoic Oolitic Ironstone Belt. This metallogenic belt extends from Rio De Oro (SOUGY 1964), Morocco (DESTOMBES et al. 1985), Algeria (GUERRAK 1987) to Libya (GOUDARZI 1970; NAKHLA et al. 1978; CHAUVEL & MASSA 1981; VAN HOUTEN & KARAZEK 1981). It contains oolitic ironstones of various age, Ordovician, Silurian and Devonian, with more than 30,000 million tons of ironstone resources (GUERRAK, unpublished data). It can be considered as one of the largest ironstone belt in the world.

A scrutiny of the literature reveals lack of information regarding the occurrence of oolitic ironstones in the northern sedimentary cover of the Precambrian Ahaggar Shield. Previous studies were mostly focused on the general sedimentation and the structure of the region in the context of petroleum exploration. So, the aim of this paper is to bring out the various aspects of the oolitic ironstones of the study area (mineralogy, petrography, geochemistry, sedimentology . . .), and try to explain their occurrence in the light of the paleolatitudinal location of the area, during Devonian times.

2. Structural and sedimentary evolution

Several major sedimentary cycles, induced or disturbed by weak tectonic events, can be identified throughout the Paleozoic deposits, whose characters were inherited from the Pan-African Orogeny (BLACK 1984). A period of pediplanation, lasting about one hundred million years, marked the closing stages of the Pan-African event and resulted in a major widespread unconformity. During the Pan-African movements, there was reactivation of Archean and Proterozoic faults, resulting in the separation of the Central Sahara into NNE–SSW components. After the Pan-African movements the region witnessed alternating periods of transgression and regression, which correspond to twelve Paleozoic sedimentary cycles of various duration (Fig. 2).

Deposition of these cycles was influenced by tectonic movements which have been traditionally assigned to the well known Caledonian (s.l.) and Hercynian (or Variscan) orogenies (FOLLOT 1952; FREULON 1955; BEUF et al. 1971). According to PIQUE (1981), POOLE et al. (1983), LEFORT & VAN DER VOO (1981), LESQUER et al. (1984), FABRE (1985), the tectonic behavior of the Saharan Platform can be summarized as follows.

1. The African Taconic Event, Caradocian in age, was synchronous but different from the Taconic orogeny of Europe and North America. Well-preserved in the Rokelides and more or less in the Mauritanides (LECORCHE 1983), it seems to be the result of collisions between Gondwanan fragments previously detached (Early Ordovician). During the following compressive phase, the N–S trending faults were reactivated, inducing: (i) some small folds related to strike slip movements (CLARACQ et al. 1958; FREULON 1964; BEUF et al. 1971) with the development of a general northwestward slope.

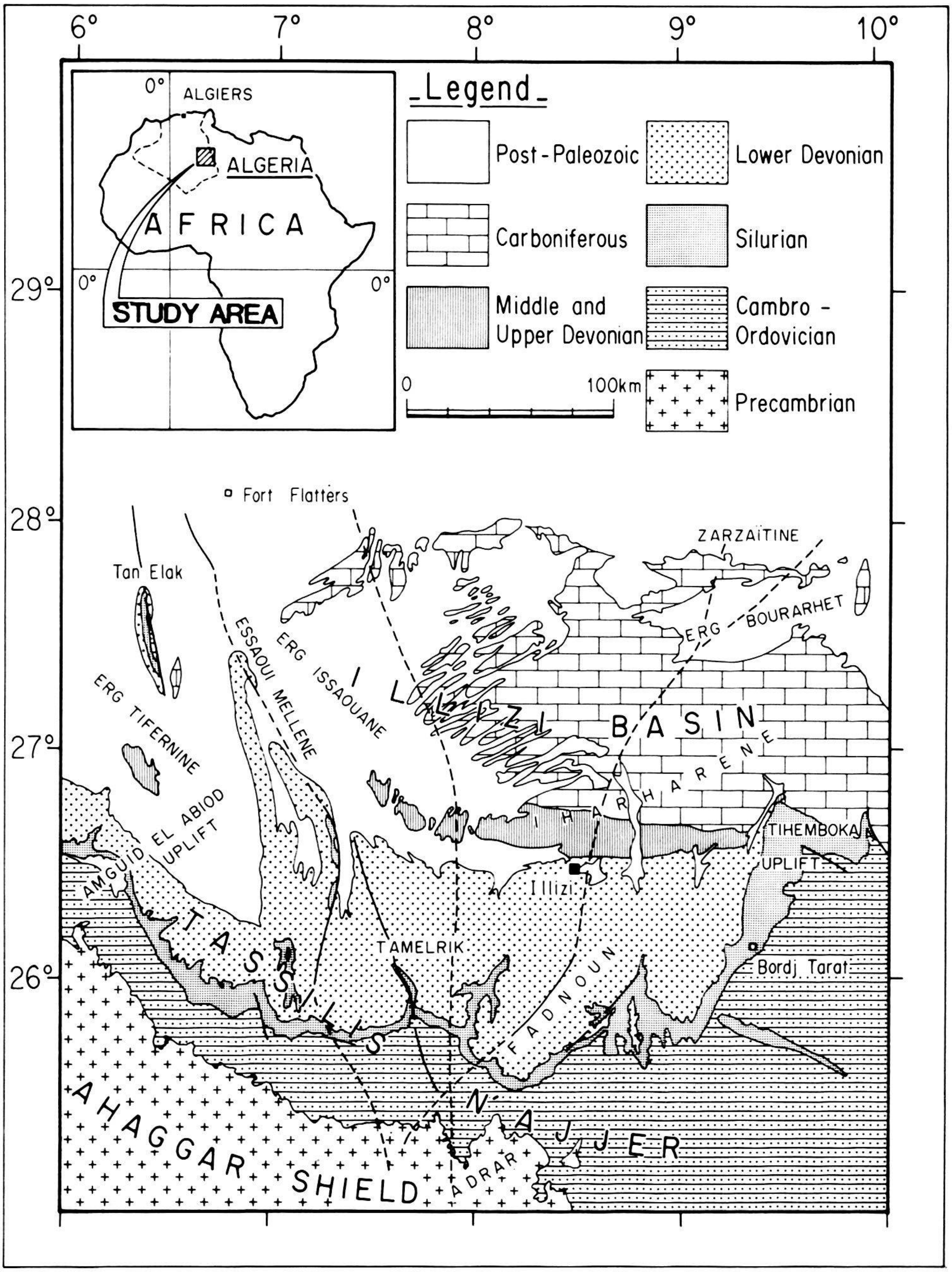


Fig. 1. Geological map of Tassilis N'Ajjer and the Illizi Basin region.

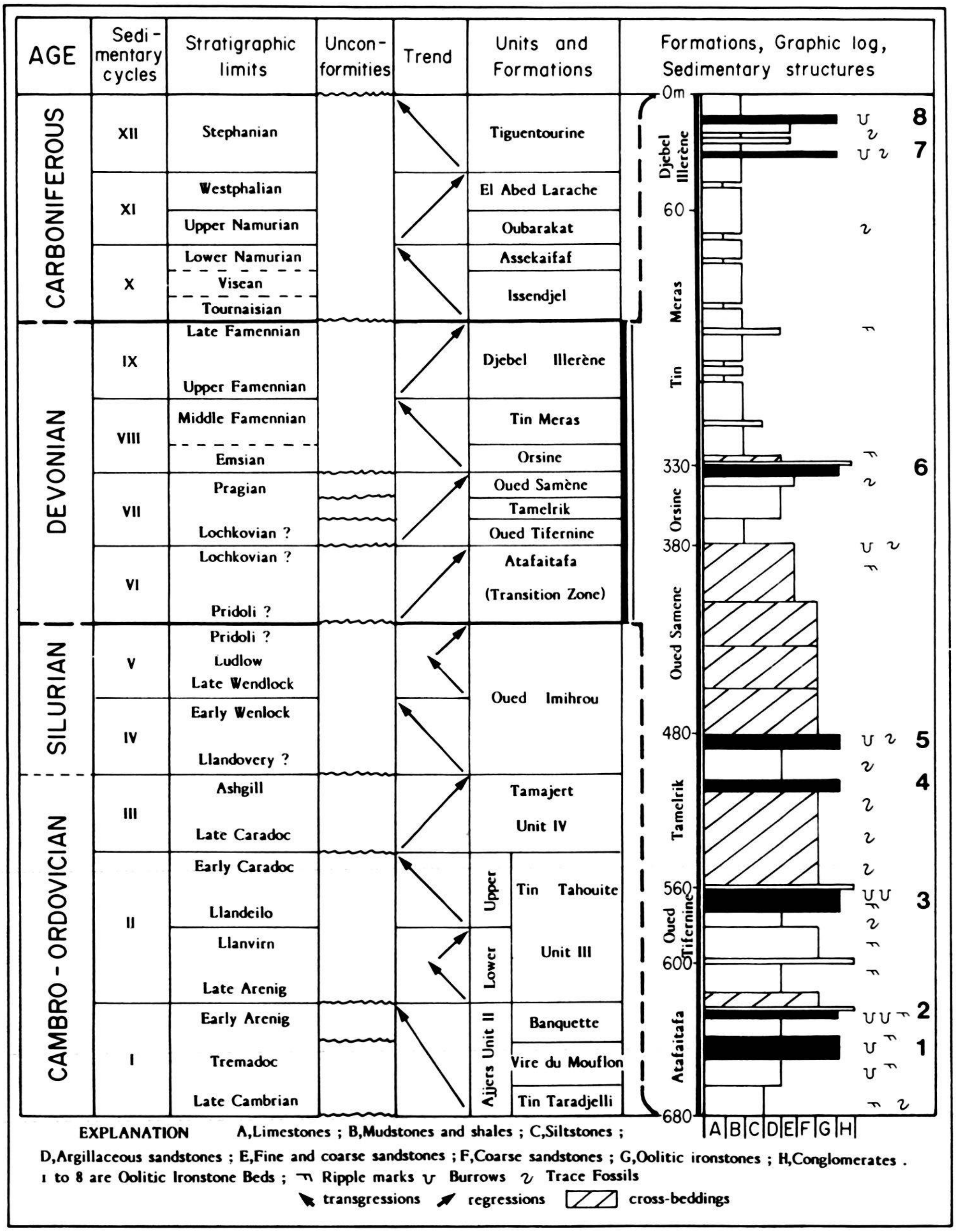


Fig. 2. Paleozoic sedimentary cycles and location of oolitic iron-rich sediments in the study area. Time stratigraphy adopted is mainly after Dubois et al. (1967), Beuf et al. (1971), Attar et al. (1980), Legrand (1985) and unpublished data from petroleum research.

2. During the Late Silurian–Early Devonian period (the time of Iapetus and Theic closures), a new tectonic phase affected the entire Saharan Platform. This pre-Acadian weak phase did not develop an orogen or important folding, but induced the reactivation of previous major submeridian lineaments and secondary faults, and the amplification of the general northwestward slope. At the same time, several positive zones evolved such as the Amguid–El Abiod and Tihemboka uplifts.

3. The final Paleozoic configuration of this area was the result of the last so-called Hercynian (or Variscan) orogeny. Well recorded in the far west (Senegal to Northwest Morocco), this phase also reactivated some faults, particularly those developed in both 40°N and 120°N directions (LATRÈCHE 1982).

The ideas of WENDT (1985) on the origin of the Moroccan Hercynian belt, can be applied to the Central Sahara. The successive desintegration of this part of Gondwana is characterized by distensive phenomena, and therefore induced sedimentary transgressions. On the contrary, the tectonic events reflecting intraplate compressive movements, with a rearrangement of the previously isolated blocks, were characteristic of epeirogenic movements and sedimentary regressions.

3. Distribution of oolitic iron-rich sediments

At the top of the Siluro-Devonian cliffs of Tassilis N'Ajjer, in the canyons that cut across the lower Devonian rocks and in the Upper Devonian Illizi peneplain, eight (8) oolitic iron-rich beds of differing age occur (Fig. 2). These Devonian ironstones, extending over several hundreds of kilometers, are of EXID type (Extensive Ironstone Deposition, GUERRAK 1987). They successively occur within (i) the "Atafaïtafa" Formation (beds 1 and 2), (ii) the "Oued Tifernine" Formation: "Talus à Tigillites" Member (bed 3), (iii) the "Tamelrik" Formation: "Trottoirs" Member (beds 4 and 5), (iv) the "Orsine" Formation (bed 6) and (v) The "Djebel Illerene" Formation (beds 7 and 8).

The lithostratigraphic nomenclature adopted is mainly after DUBOIS et al. 1967; BEUF et al. 1971; LEGRAND 1985. In the following paragraphs a detailed account of the geology of the ironstones is given, based on the original observations made by the author in the field.

Atafaïtafa ironstones

Interbedded with alternating mudstones, siltstones and fine-grained ferruginous sandstones, two ironstone beds occur in the Upper Atafaïtafa Member. Ferruginous ooids occurring in small (centimetric) lenses, are scattered within fine-grained chloritic sandstones. These lenses are oriented parallel to the bedding and exhibit different shapes induced by thickness variations. The thickness of the ironstone – rich beds varies from 1 m (near Imihrou area) to 10 m (near Fadnoun area).

In these sediments, numerous trace fossils, burrows (*Skolithos* ichnogenus), ripple marks, channel filled with microconglomerates and paleosols, a few centimeters thick, can be observed. The mostly angular detrital quartz grains are cemented by a chloritic-ferruginous material. Between the oolitic beds, a succession of alternating fine and medium-grained kaolinitic to chloritic sandstones, separated by hematitic to geothitic layers (a few centimeters thick), occur. Striped hematitites (hematitic crusts) with colloform structures also occur.

Oued Tifernine ironstone

The third oolitic iron-rich bed is continuous throughout the whole Outer Tassilis, disappearing eastward close to the Tihemboka uplift; near the Oued Imihrou area. This bed has a thickness of 5 m.

In the "Oued Karkaï" Member, ooids are scattered within chloritic and quartzitic ferruginous sandstones. Channel and ripple-mark structures dissect the regular laminated layers which contain encrustations of iron oxides and hydroxides, numerous trace fossils (*Cruziana*) and burrows (*Skolithos* ichnogenus).

The vertical and lateral distribution of the ooids is random. The ooids appear to be confined to the fine-grained layers rather than to the coarser ones.

Tamelrik ironstones

Two iron-rich oolitic beds, one at the base (bed 4) and the other at the top (bed 5) of the "Trottoirs" Member are recognized. Bed 4 (5 m thick) is composed of scattered ooids within fine-grained chloritic ferruginous sandstones. The ooids are more abundant within the chlorite dominated layers. Quartz grains are more or less angular, with variations in size. Ripple marks, fossil tracks and *Skolithos* are present, which are sometimes separated by iron crusts.

Orsine ironstone

The Orsine ironstone (represented by Bed 6) constitutes the uppermost part of the Emsian deposits of the Illizi Basin. It is located at the top of a coarsening-upward sequence, overlain by a discontinuous conglomerate. The lithologic succession is green and purple mudstones, finely laminated chloritic sandstones, ooid-bearing chloritic and hematite-rich medium grained sandstones and a scoured surface overlaid by a microconglomerate. The first several meters of the overlying Eifelian transgressive formations (argillaceous-dominated) are constituted by cross-bedded sandstones.

Djebel Illerene ironstones

These Late Famennian ironstone deposits (Beds 7 and 8, Fig. 2) of the Illizi Basin are located at the top of small coarsening-upward sequences which are part of a larger regressive cycle.

Bed 7 (50 cm thick) occurs at the top of monotonous sandstones (30 m thick), whereas Bed 8 (2 meters thick) overlies 4 m of thick chloritic-ferruginous sandstones which are in turn overlain by 3 m thick green mudstones. This oolitic ironstone bed contains numerous *Skolithos* at the top, and hardly recognizable corals are found at the bottom. 10 m of grey mudstone, overlain by 50 cm of a shelly conglomerate rich in fish fragments and phosphatic nodules, constitute the uppermost part of this Famennian Formation.

4. Petrographic analysis of the ironstones

A systematic microscopic analysis (transmitted and reflected light) has been performed on about 100 samples representing the eight oolitic iron-rich beds of the study area.

The study made for each bed brings out some peculiar characters which are related to (i) the mineralogy of ooids, (ii) the mineralogy of the groundmass, (iii) the mineralogy of ooid nuclei, and to (iv) the occurrence of composite ooids, broken ooids, spastoliths, superficial ooids, intraclasts and detrital particles.

The distribution of the different components, particularly of the ooids within the groundmass, allows the recognition of three main types of ironstone facies: (i) a facies with dispersed ooids occurring separately in a quartz-rich chloritic and hematitic groundmass (detrital facies: FOD), (ii) a facies with ooids in a chloritic and hematitic cement and showing pore-filling structures (cemented facies: FOC), and (iii) a microconglomeratic facies (FMC) in which the intraclasts of various nature, sometimes oolitic intraclasts, are dominant. The cement is composed of chlorite or iron oxide. Sometimes a mixed ironstone facies can also occur (FOC/FOD facies). Each one of these ironstone facies can be subdivided into several subfacies based on the mineralogy of ooids (Fig. 3).

The mineralogical composition is very simple and consists of three main types of ferriferous minerals: (i) chamosite which is a trioctahedral 14 Å chlorite (BAYLISS 1975; BAILEY 1980), previously called bavalite (ORCEL 1927; HEY 1954). It is easily detected by X-ray analysis, showing a higher 7 Å peak than 14 Å peak; (ii) hematite, and (iii) goethite.

Non ferriferous minerals such as quartz and calcite (rarely tourmaline) occur within the nucleus of ooids, matrix, and cement. Apatite has not been detected, in spite of the occurrence of phosphorus. The paragenetic sequences are: (i) chamosite–hematite–goethite–quartz (P1), (ii) hematite–goethite–quartz (P2), (iii) chamosite–goethite–quartz (P3), and (iv) goethite–quartz (P4).

The distribution of the ironstone facies and the parageneses is represented in figure 4 which gives an overall view about the relationships between the age of the ironstones, their texture and their composition. As can be seen from the figure, the detrital facies (FOD) appears to be vertically and laterally more widely developed than the cemented facies (FOC). The evolution of this facies has been previously described in other Saharan ironstones (GUERRAK & CHAUVEL 1985; GUERRAK 1987).

Paragenetic variations are strictly related to the oxidation state of the ironstone which can be expressed by the relation chamosite→hematite→goethite. This implies that chamosite is the main primary oolitic mineral, subsequently oxidized to hematite. In spite of the high oxidation state of the ironstones of this area, it is possible to recognize a chamositic skeleton by etching the thin sections of the samples with a solution of "Sodium Dithionite" which eliminates the iron hydroxides.

In the groundmass which is less oxidized than the ooids, the concentration of iron oxides or iron hydroxides is related (i) to the initial porosity and thus to the water

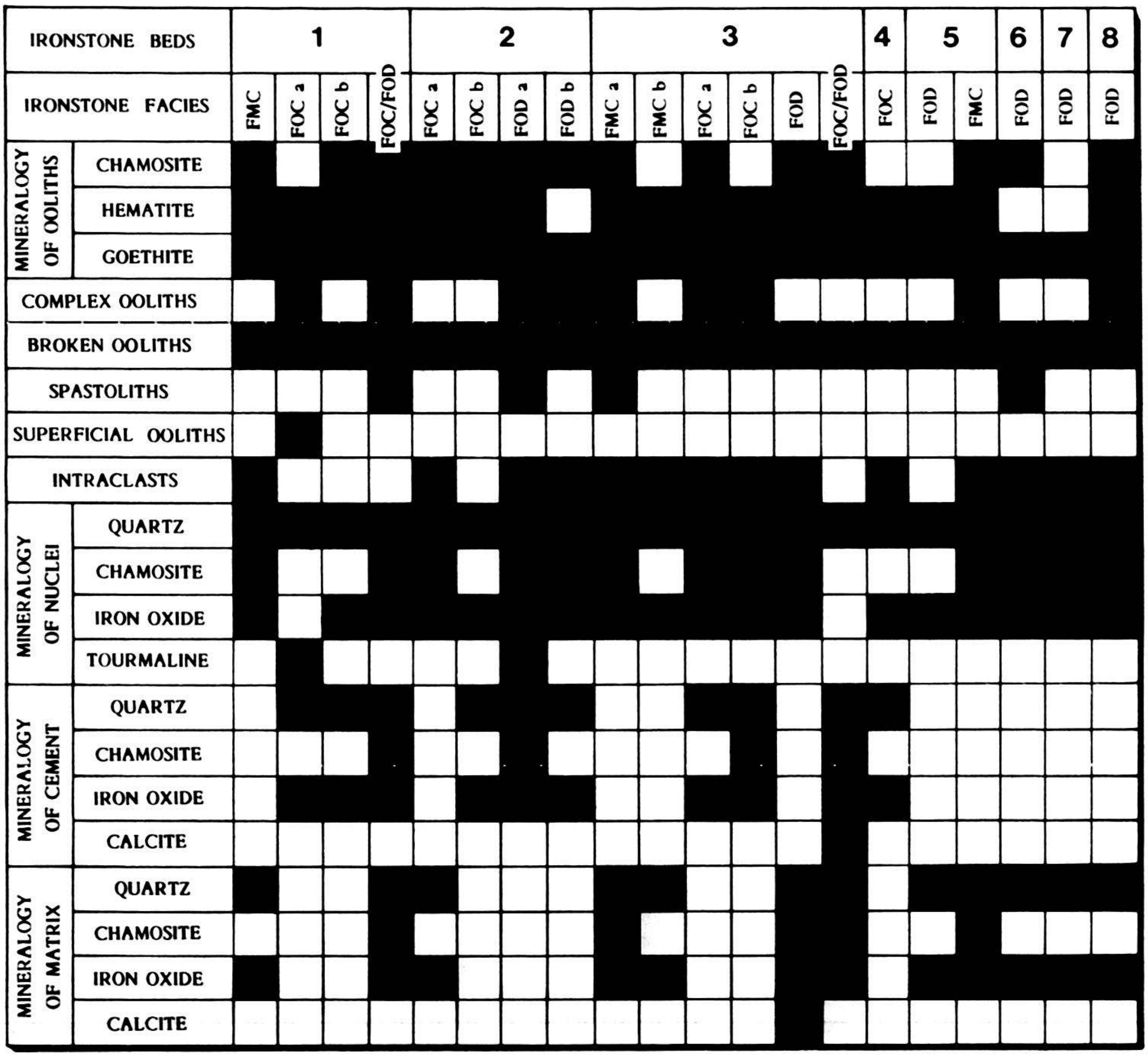


Fig. 3. Mineralogical and petrographical characteristics of the ironstones in Tassilis N'Ajjer and the Illizi Basin.

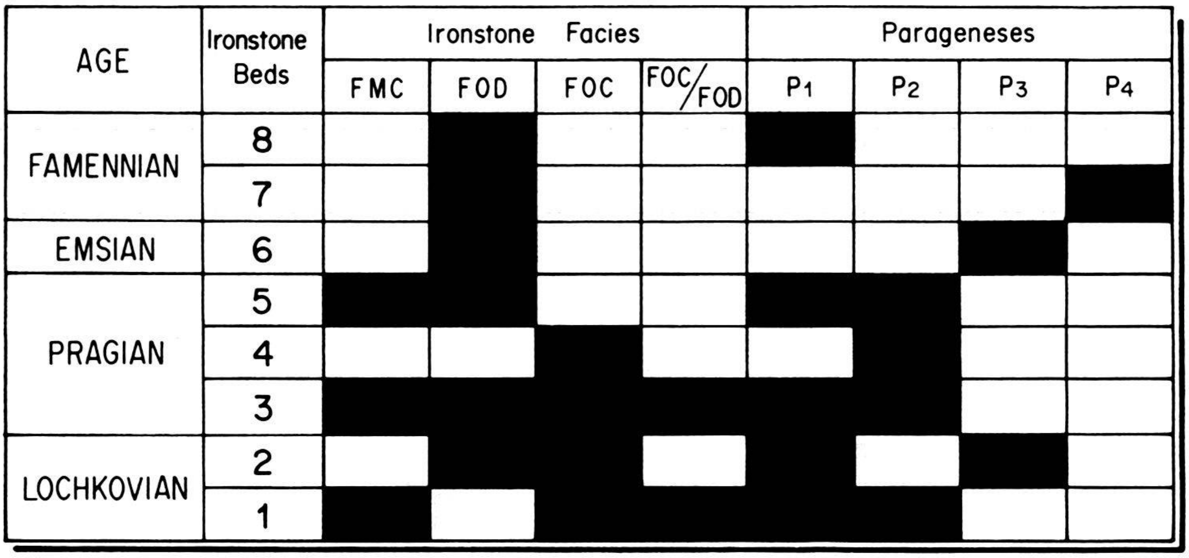


Fig. 4. Distribution of the ironstone beds, of the ironstone facies and of the parageneses, through the geological record, in Tassilis N'Ajjer and Illizi Basin areas.

content during the early diagenesis (MULLER 1967), and (ii) to the weathering processes.

Description of ooid types

Based on morphological features, six main types of ferruginous ooids can be recognized from the different oolitic iron-rich sediments of the Tassilis N'Ajjer and Illizi basin area. There are concentric, eccentric, distorted (spastoliths), superficial, complex, and broken types.

(i) Well rounded (concentric) ooids consist of thin successive cortices, concentrically arranged around a nucleus. These cortices are laminated and composed of alternating oxidized and non-oxidized chamosite, whereas the nuclei are either quartz, tourmaline, or iron oxide grains. The oxidation of the cortices is not regular and thereby the thickness of the layers is variable (Pl. 1: Fig. 1, 2). The size of this type of ooids varies between 100 and 500 µm. These ooids are scarce and occur especially within the detrital facies (FOD).

(ii) Eccentric ooids similar to those described by KNOX (1970) are recognized and constitute at least 75% of the whole ooids. In this type, the cortices were accreted eccentrically around a nucleus of various shapes and mineralogies (Pl. 1: Fig. 3, 4). In most cases, the nucleus is single but in about 5% of the cases, there are several nuclei within one ooid (Pl. 1: Fig. 3, 5). The mineralogies of these multiple nuclei, which occur in all the ironstone facies, are similar or different (quartz, chamosite or hematite). These eccentric ooids are always more or less flattened and range in size from 100 to 700 µm. They can be considered as intermediate between the concentric and the distorted ooids.

(iii) Distorted ooids (spastoliths) have varied shapes from slightly flattened (Pl. 1: Fig. 6; Pl. 2: Fig. 6) to nearly acicular (Pl. 1: Fig. 7, 8). While the size of the ooids varies from 200 µm to 800 µm, the width does not exceed 40 µm. When the nucleus is chamositic it is also affected by the distortion (Pl. 2: Fig. 1, 2, 3).

(iv) Superficial ooids represent the first step in the development of the ooids. The cortex consists of only one irregular layer surrounding a quartz nucleus. Sometimes the layer is discontinuous which is due to incomplete oolitization or abrasion phenomena. The size of these ooids varies from 50 to 100 µm, and is smaller than that of the other ooids.

(v) Composite (multiple or complex) ooids consist of two or more ooids welded together within an oolitic cortex. They commonly occur in the ironstone facies FOD and FMC (Fig. 3). Several types of composite ooids can be recognized, namely, composite ooids with well-laminated cortices (Pl. 2: Fig. 4), composite ooids with irregularly laminated cortices, composite distorted ooids containing either a spastolith or a normal ooid, where as the final cortex is distorted (Pl. 2: Fig. 1, 2). An other variety of composite ooids exists with oolitic intraclasts included in a facies different from the facies in which they were originally developed (Pl. 2: Fig. 6). The size of these ooids ranges from 300 µm to about 1000 µm.

(vi) Broken or fragmentary ooids are very abundant and occur in all the samples. They are often more oxidized than the other ooids and can serve as nuclei for new ooids.

5. Geochemistry of the ironstones

Composition

Analysis of major and trace elements (Bi, Cd, Co, Cr, Cu, F, Ni, Pb, S and Zn) were carried out for 21 representative outcrop samples spanning the study area (Table 1 and Fig. 5). The geochemistry of the ironstones allows the distinction of two main types of ironstones: (i) the first is Fe-poor (total Fe) with Fe_2O_3 content ranging between 20 and 40%, and SiO_2 between 40 and 60%, and characterized by relatively low concentrations of P_2O_5 (0.89% to 1.43%) and CaO (0.47–4.07%). This type is represented by the beds 1, 2, 4, 5 and 8. (ii) The second type is Fe-rich and has an Fe_2O_3 range from 40 to 50%, with SiO_2 values ranging between 10% and 40%. The phosphorus content is more than in the first type with P_2O_5 ranging between 1.79% and 9.6%. The CaO concentration is also high varying from 3.88% to 14.69%. This type is

Table 1: Chemical composition of major and minor elements of the ironstones (beds 1 to 8) in Tassilis N'Ajjer and the Illizi Basin.

%	1	2	3	4	5	6	7	8	Average	Range	
SiO_2	55.28	55.34	34.29	46.85	46.88	15.74	15.65	61.32	42.59	69.91	15.65
Al_2O_3	3.96	5.50	1.79	3.30	6.68	3.33	5.80	1.57	3.74	16.80	0.79
Fe_2O_3	33.32	29.07	49.55	33.80	36.95	49.20	48.21	25.35	39.32	57.95	18.89
FeO	0.22	0.46	0.06	0.04	0.12	0.03	0.03	0.36	0.10	0.61	0.02
MnO	0.18	0.42	0.29	0.39	0.38	0.83	0.47	0.27	0.37	0.94	0.05
MgO	0.13	0.51	0.47	1.14	0.32	2.92	0.10	0.44	0.60	2.92	0.10
CaO	0.72	0.47	3.89	0.48	1.51	10.43	14.69	4.04	3.20	14.69	0.04
Na_2O	0.06	0.05	0.07	0.07	0.05	0.09	0.14	0.05	0.06	0.15	0.02
K_2O	0.09	0.87	0.19	0.22	0.08	0.27	0.13	0.21	0.37	4.18	0.02
TiO_2	0.66	0.25	0.26	0.19	0.16	0.21	0.12	0.51	0.27	0.66	0.12
P_2O_5	1.03	0.94	1.79	0.89	0.91	2.26	9.60	1.43	1.76	9.60	0.68
LOI	4.19	5.90	7.82	10.31	5.76	14.33	4.86	3.97	7.05	14.33	3.96
Total	99.84	99.78	100.47	96.98	99.80	99.64	99.80	99.52	99.44		
ppm											
Bi	200	200	150	200	150	200	200	200	167	200	150
Cd	8	13.33	14	24	12	24	40	8	15	40	8
Co	80	100	95	120	60	120	120	40	89	160	40
Cr	80	80	165	80	140	120	80	160	124	400	40
Cu	40	33.33	22.5	20	20	20	40	20	27	60	20
F	100	150	475	100	250	1200	2600	800	476	2600	100
Ni	200	246.66	590	400	220	700	400	200	407	1400	160
Pb	200	191.66	187.5	200	175	200	200	200	188	200	150
S	1400	2550	5150	288	1650	1600	6000	1600	3505		
Zn	180	321.66	22.5	240	110	350	310	140	245	740	80

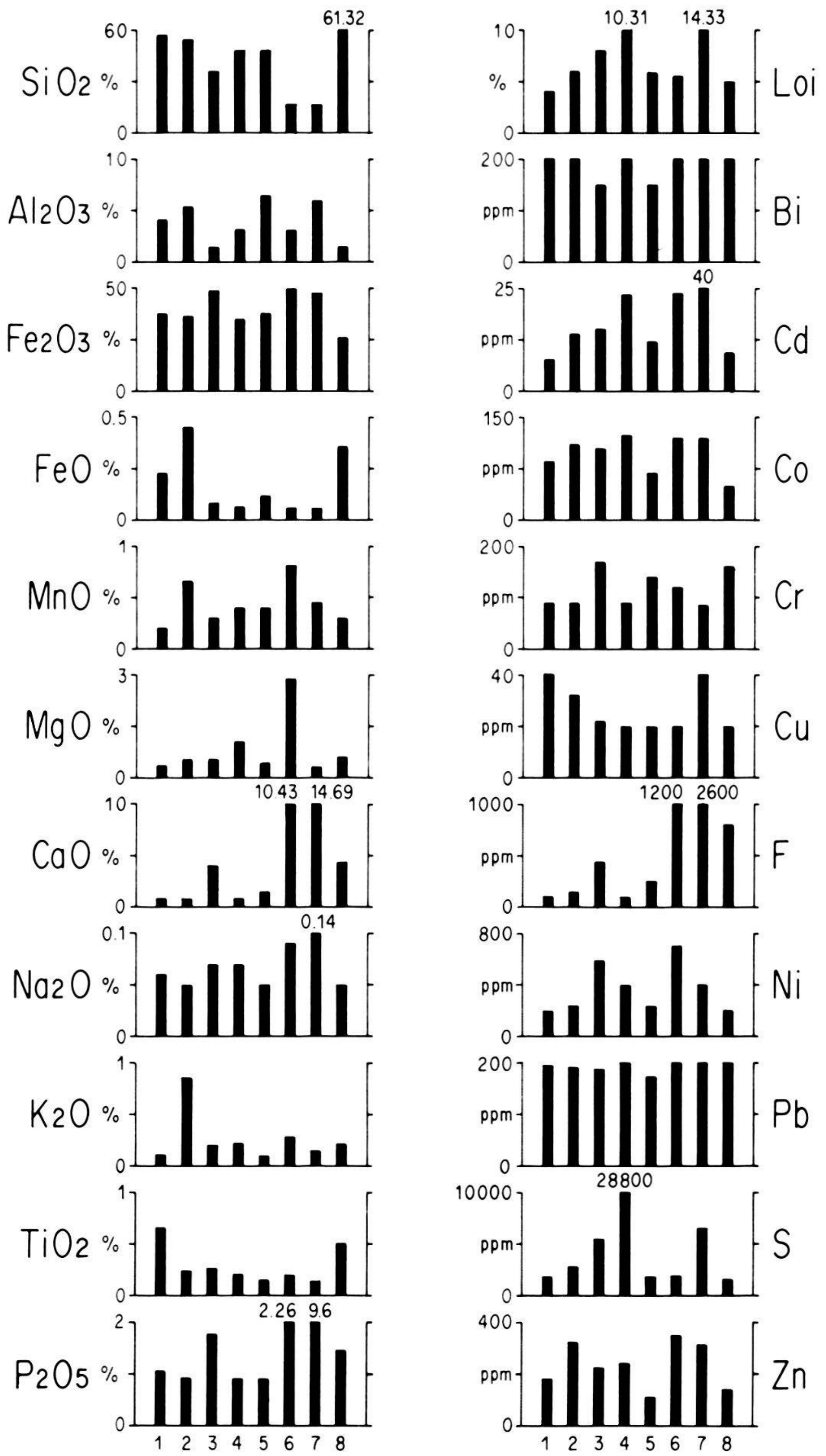


Fig. 5. Chemical composition of the ironstones. Numbers 1 to 8 correspond to the different ironstone beds.

represented by the beds 3, 6, 7. However the FeO content in the two types is low and this may be due to the highly oxidized nature of the ironstones. It may not be incorrect to presume that the bulk chemistry of these ironstones can be related to both primary (diagenesis) and secondary (weathering) processes.

Linear correlations

Linear regression values have been computed for all the analyzed elements, but those with correlation coefficient (r), above 0.5 are taken for interpretation (table 2). (i) SiO_2 has negative correlation with Fe_2O_3, CaO, Na_2O, P_2O_5, F and Ni. This correlation is significant, because the sum of $SiO_2 + Fe_2O_3$ is > 80%. (ii) Positive correlation between Al_2O_3, FeO and Cu are not significant and are probably related to the argillaceous phase of the ironstone. (iii) Positive correlation between FeO/TiO_2 can be explained by the occurrence of Ilmenite microinclusions in iron-rich minerals. However, it seems that the correlation is not due to the presence of detrital rutile; (iv) MgO/Loi (Loss on ignition) correlation can be easily related to chamosite or possible inclusions of dolomite; (v) the strong positive correlations between P_2O_5, CaO and F are linked to the occurrence of apatite which seems to be an occult mineral. The positive correlation between Cd and F can also be explained by the above occurrence of apatite, since Cd has the same ionic radius as that of Ca.

Table 2: Correlation coefficients of the ironstones of Tassilis N'Ajjer and the Illizi Basin.

	Correlation Coefficients (r>0.50)
SiO_2	Fe_2O_3 (- 0.89), CaO (- 0.75), Na_2O (- 0.61), P_2O_5 (- 0.60) F (- 0.62), Ni (- 0.54)
Al_2O_3	FeO (0.57), Cu (0.64)
Fe_2O_3	CaO (0.58), Na_2O (0.57), Ni (0.57)
FeO	TiO_2 (0.69)
MnO	Bi (0.51)
MgO	LOI (0.73), Co (0.55)
CaO	Na_2O (0.64), P_2O_5 (0.86), Ca (0.65), F (0.95)
Na_2O	F (0.58), Ni (0.64), P_2O_5 (0.52)
P_2O_5	Cd (0.73), F (0.93)
LOI	S (0.51)
Cd	F (0.73)
Co	Cu (0.56)

Comparison with crustal abundance

The comparison of the mean chemical composition of the ironstones with the bulk continental crust (Fig. 6) has been made on the basis of H_2O free analyses. The data used in the present paper are those from RONOV & YAROSHEVSKY (1969) for major elements and TAYLOR & MC LENNAN (1985) for trace elements.

(i) The enrichment of Fe_2O_3 is similar to the other Saharan ironstones such as those of the Ougarta ranges in northwestern Sahara (GUERRAK, unpublished data). This enrichment is certainly accentuated by weathering processes. (ii) P_2O_5 commonly occurs within the oolitic ironstones particularly in the Saharan ironstones (GUERRAK & CHAUVEL 1985; GUERRAK 1987). It seems to be related to the occult apatite and the development of organic life. (iii) The increase of Bi has been previously recorded in other Paleozoic ironstones (GUERRAK, unpublished data). This element is rare in crustal rocks (0.3 ppm to 6.5 ppm, WEDEPOHL 1974). The concentration of Bi is probably related to the weathering processes, but further geochemical investigations are necessary to know about its behavior. (v) The Cd enrichment in these ironstones confirms the correlation with Calcium. (vi) The concentrations of Co, Ni and Pb are high as compared with crustal rocks. They seem to be due to the occurrence of these elements in the structure of ferruginous minerals, and may also be related to weathering processes (BESNUS et al. 1969). (vii) Sulfur shows an enrichment which may be due to

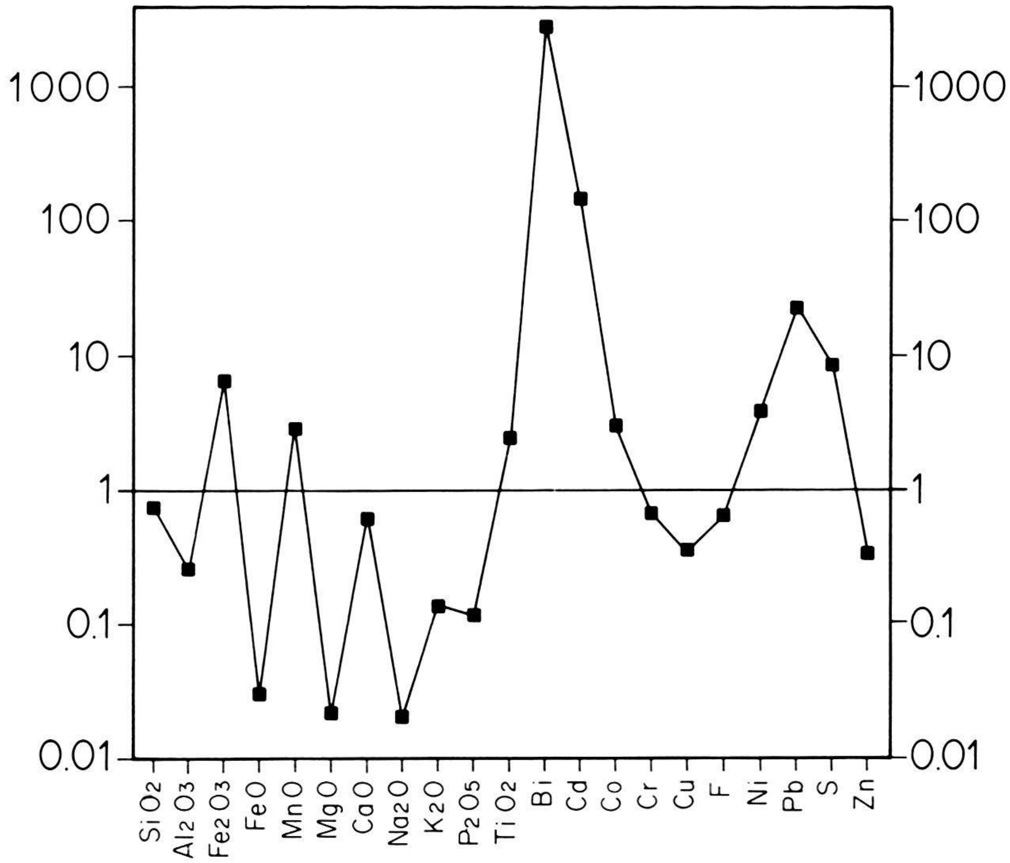


Fig. 6. Chemical composition of the ironstones of the Tassilis N'Ajjer and the Illizi Basin, as compared with crustal abundance (crustal abundance = 1).

the presence of organic matter in shallow environments, such as those rich in *Skolithos* (e.g. the 'Talus à Tigillites' Member or the 'Trottoir' Member). (vi) The impoverishment of some oxides such as SiO_2, Al_2O_3, FeO, MgO and K_2O can be related either to the inverse relationship between iron deposition and detrital sedimentation, and/or to the recent leaching of the rocks. (vii) Low concentrations of copper are similar to the other oolitic ironstones (JAMES 1966). The low values of Zn can be related to the low chamosite concentrations, since this element is often adsorbed on clay minerals, (WEDEPOHL 1974). (vii) The fluorine impoverishment is due to the lower concentration of CaO in these rocks, since there is a positive correlation between the two. (iv) Cr and TiO_2, generally associated with magnetite, are lower when compared to crustal abundance.

6. Discussion and conclusions

The distribution of the oolitic iron-rich sediments within the Paleozoic sediments of the Tassilis N'Ajjer and the Illizi Basin depends on several factors related: (i) to the source of iron, transporation and paleolatitudinal context, (ii) to the oolitization mechanisms and accumulation of the oolitic ironstone, (iii) to the physicochemical conditions of formation of iron-rich minerals, and (iv) to the paleoenvironmental characteristics.

Source, transportation and paleolatitudes

According to previous observations of BEUF et al. (1971) and BIJU-DUVAL et al. (1981), based on paleocurrent analysis connected with iceflow directions, the source of sedimentation in the Paleozoic seems to be located southeast of the study area, in the continental Nigerian and Congo Shields (Fig. 7). The great maturity of the sediments, the occurrence of resistant heavy minerals (rutile and tourmaline) and the sarcity of feldspars and lithic grains, observed in the ironstones and associated sediments, are in conformity with the views expressed by BEUF et al. (1971).

It is sometimes claimed that the source area of some ironstone deposits is necessarily composed of lateritized rocks (SIEHL & THEIN 1978; NAHON et al. 1980 ...). However, the available evidence suggests that the source area need not always be composed of lateritic rocks. Leaching of crystalline rocks (felsic and mafic) mudstones, and ferruginous sandstones on land is adequate to produce a relative concentration of iron in marine sediments.

The anastomosing to meandering rivers flowing to the North during Paleozoic times (BEUF et al. 1971) were probably responsible for the transport of iron over long distances, between 2000 and 3000 km. Though there is a lack of consensus about the transportation of iron (KIMBERLEY 1978, 1981), it is possible to envisage several modes of transportation by rivers. According to CARROLL (1958), and more particularly to GIBBS (1973, 1977), five transport phases of iron exist: (1) crystalline particles, (2) iron hydroxides coatings, (3) solid organic material (4) sorbed material and (5) in solution. The two dominant transporters, crystalline particles and metallic coatings carry about 90% of the iron. While the organic solid phase transport corresponds to about 9% of total iron, the iron phases transported in solution and sorbed on solids, contribute to less than 1%. This does not contradict the conclusions of JAMES (1966) who underlined the very low concentration of iron in solution, and indicated a value of about 10%. In

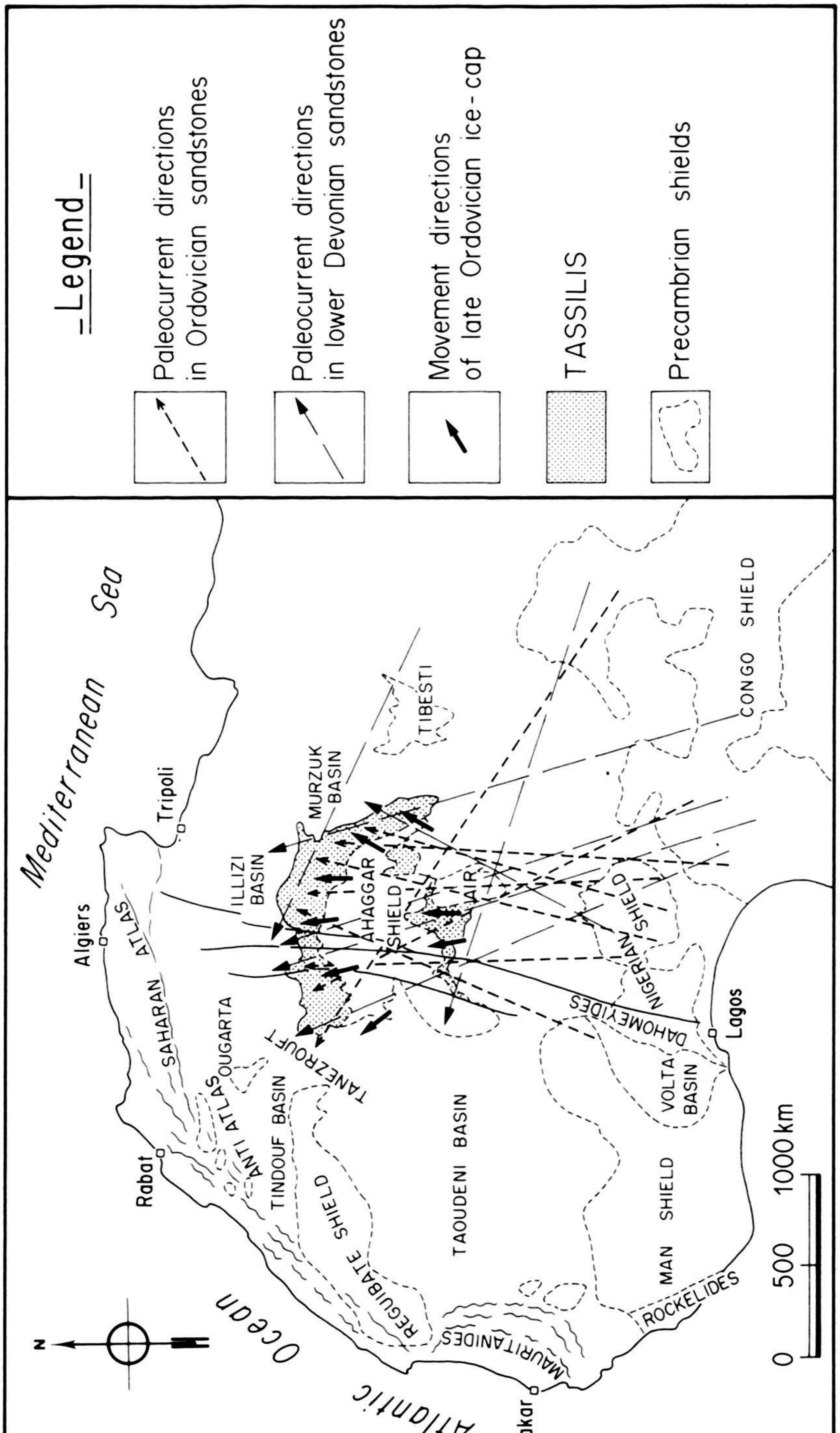


Fig. 7. Reconstruction of supposed iron flow directions and related source (data after BEUF et al. 1971).

the same way, DICKEY (1968), who estimated the precipitation of iron in the Aral Sea (Soviet Central Asia) at 4.7 million tons per year, indicated that only 3% of iron is in solution. This example clearly substantiates the author's view, that there is no necessity to have a lateritic source for ironstone deposition, which is usually advocated in the classical hypotheses.

The nature of ironstone sedimentation presents variations which probably depends on the latitude of the iron source and on the latitude of the site of deposition.

As a matter of fact, the distribution of the ironstones of the study area appears to be strictly related to the paleocontinental location of Gondwana, of which they are an important component of the northern margin.

According to the more common continental drift reconstructions (MOREL & IRVING 1978; SCOTESE et al. 1979; BAMBACH et al. 1980; BOUCOT & GRAY 1983; HARGRAVES & VAN HOUTEN 1985), deposition of the Devonian oolitic iron-rich sediments of the Tassilis N'Ajjer and Illizi Basin may have occurred before drifting of Gondwana. This is in conformity with the conclusions of VAN HOUTEN & BHATTACHARYYA (1982) who presented a time-latitude chart of distribution of Phanerozoic oolitic ironstones with respect to the location of cratonic blocks.

Though there is a lack of consensus for the location of the Devonian pole (HAILWOOD 1974; LIVERMORE et al. 1985; SALMON et al. 1986), the location of the study area seems to have been comprised between latitudes of 30 and 80° S (which corresponds to the present day temperate zone) whereas the source area was situated between 60 to 90° S (in a cold climate). Several authors have presented models indicating that all oolitic ironstones could have been formed under an hot and humid climate.

HECKEL & WITZKE (1979) locate North Africa within temperate latitudes during the Devonian, ranging from relatively cool and humid to warm and locally drier; scattered oolites in the sahara suggesting 'local restriction towards hypersalinity'. They also invoked warm currents flooding Gondwana margins.

WOPFNER & SCHWARZBACH (1976) discussing the climatic controls of ore deposits, stated that 'the most favourable environment for the formation of syngenetic ore deposits exists under warm, humid climates'; 'iron was derived from lateritised land areas'; 'all occurrences of oolitic iron ores . . . are close to or south of the evaporite belts'; '. . . the postulation of a warm, seasonally humid climate for the formation of oolitic iron ores'.

NAHON et al. (1980) proposed a general model of pedogenetic iron ooid formation within ferricrete crusts that are accreted in the Recent, above a variety of sedimentary, metamorphic and igneous rocks, by lateritic weathering in a tropical climate.

Although these models have been applied in the cases of the Minette of Lorraine (Aalenian) or the Clinton ores of North America (Silurian), they cannot be successfully applied in explaining the ironstone occurrences of North Africa: therefore they cannot be considered as universal. It appears that oolitic ironstones may be deposited in temperate as well as in tropical climates.

Oolitization mechanisms and ironstone accumulation

The accumulation of oolitic ironstone and the process of oolitization have long been debated (SORBY 1857), and several hypotheses have been proposed as shown in

Table 3: *Proposed modes of iron concentration and oolitic accretion mechanisms.*

PROPOSED MODE & ADHERENTS
COOLING, DEGASSING AND/OR MIXING WITH SEAWATER OR HYDROTHERMAL SOLUTIONS (Gross, 1965)
DIAGENETIC FERRUGINISATION (Deverin, 1945 ; Jones, 1965)
ELECTROLYTIC AND ACID-BASE PRECIPITATION DURING DISSOLUTION OF ARAGONITIC SEDIMENT BY DESCENDING LEACHATE FROM CLOSELY OVERLYING MUD BEING LEACHED BY ORGANIC ACIDS (Kimberley, 1978 ; 1981)
INCREASE IN ELECTROLYTES BY MIXING OF FRESH WITH MARINE WATER (Alling, 1947 ; Harder, 1964 ; Petranek, 1964)
INORGANIC OXIDATION IN MARINE WATER (Alling, 1947 ; Castano & Garrels, 1950 ; Borchert, 1960 ; James, 1966 ; Curtis & Spears, 1968)
INTRASEDIMENTARY ACCRETION PROCESS DEVELOPED AROUND SCATTERED ELEMENTS WITHIN IRON-RICH MUD (Caillère & Kraut, 1953 ; James, 1955 ; Rohrlich & al., 1969 ; Chauvel, 1971 ; Chauvel & Guerrak, 1986 ; Van Houten & Purucker, 1985 ; Guerrak, 1987)
IRON DERIVED FROM LATERITIC ROCKS. GOETHITIC/HEMATITIC OOIDS ARE FORMED BY ACCRETION AT THE SEDIMENT SURFACE. CHAMOSITE TAKES PLACE AFTER BURIAL INTO REDUCING CONDITIONS BENEATH THE SEAWATER-SEDIMENT INTERFACE AT WATER DEPTHS OF UP TO 100 METERS (Gygi, 1981)
OOIDS PARTICLES ARE OF BIOGENIC ORIGIN AND MAINLY CONSIST IN NUBECULARIA (INCRUSTING FORAMINIFER) (Champetier et al., 1987)
OOIDS PRODUCTION UNDER LATERITIC WEATHERING WHICH DEVELOPED IN SITU AS IRON CRUSTS IN A TROPICAL CLIMATE (Nahon et al., 1980)
OOLITIC ACCRETION DURING LATERITIC WEATHERING, OOIDS BEING FORWARDS TRANSPORTED TO THE SEA AND ACCUMULATED AS A MARINE PLACER (Siehl & Thein, 1978)
OOLITIC ACCRETION GENERATED BY BACTERIA OR MICROBIAL ACTIVITY (Cayeux, 1922 ; Hallimond, 1951 ; Dahanayake & Krumbein, 1986 ; Gehring, 1985, 1986)
PHYSICAL SEPARATION OF IRON-RICH SOLUTIONS AND/OR COLLOIDAL SUSPENSIONS FROM FLUVIAL DETRITUS CAUGHT IN A CLASTIC TRAP (Castano & Garrels, 1950 ; Adeleye, 1973)
SELECTIVE REMOVAL OF NON-FERRIFEROUS MINERALS BY CURRENTS DUE TO DENSITY AND GRAIN-SHAPE DIFFERENCE (Bubenicek, 1968 ; Brookfield, 1971)
SUBAERIAL WEATHERING OF IRON-RICH ROCK, OR LATERITIC PENEPLANATION, FOLLOWED BY MARINE SEDIMENTATION (Erhart, 1955 ; Millot, 1970)

Table 3. The observations of Paleozoic ironstones of Algeria, particularly those of the Tassilis N'Ajjer and Illizi Basin region, suggest a model of intrasedimentary accretion of ooids within very shallow-marine sediments (depth of few meters).

From the petrography and mineralogy of the ironstones of the study area, the following aspects emerge.

1. By leaching with Sodium Dithionite, the internal structure of the ooids can be observed to consist of alternate layers of chamosite and hematite, even when the envelope of the ooids is completely oxidized.

This observation indicates that the iron oxides of the envelopes might have been formed by two processes: either by precipitation of iron oxides within a chamositic framework, or by the oxidation of chamosite.

To establish the more probable mechanism that can be applied to the present case, a detailed chemical analysis (by electron microprobe) accross the cortices was carried out (Fig. 8A) and the following interpretations are made. The envelope is composed of two distinct mineral phases (even if the ooid is greatly oxidized), the dominant mineral being hematite (and more or less goethite), with minor concentrations of chamosite. This is inferred by the variation of FeO total (which corresponds to the ferric iron of hematite and the ferrous iron contained within chamosite) which represents the oxidation state of the ooid.

The analysis of P_2O_5 indicates an inverse relationship with FeO total and therefore a direct relationship with the other elements, constituants of chamosite (Fig. 8C). SiO_2, Al_2O_3 and MgO which are the main components of chamosite have an inverse relationship to iron (Fig. 8B, D, E). On a triangular diagram (SiO_2, Al_2O_3 and MgO present concentrations conform to the composition of classical chamosites (Fig. 8 F). Furthermore, they show linear variations in silica and aluminium which seem to correspond to microcycles in the envelope of the ooid. These microcycles, which are of different thicknesses (about 4 to 24 microns) are apparently the result of episodic accretion of the ooids. During an accretion period there is growth of chamosite (and therefore of SiO_2, Al_2O_3 and MgO, together or separately, this depends on the composition of chamosite) and a decrease of iron oxides. When the ooid accretion ceases (break in oolitization and suspension period), there is increase in iron and impoverishment of SiO_2, Al_2O_3 and MgO.

Thus, the ooids appear to be the result of alternating accretion of chamosite and the transformation (oxidation) of the outer part of each layer during breaks in accretion.

The model is not uniform and presents some gradual variations which seem to be related to the degree of oxidation: under low oxidizing conditions, Al and Mg are more or less leached, whereas under highly oxidizing conditions there is intensive leaching of Al, Mg and Si and concomitant enrichment of Fe (sometimes in Ti).

This example (on an apparent completely oxidized ooid) clearly indicates that a mechanism of centrifugal oxidation of chamosite seems to be more pertinent. In other Saharan deposits (Guerrak 1987 and unpublished data) and in French ooids (Joseph & Beaudoin 1983), this has been observed too.

On the other hand, in the Gara Djebilet deposit for example (Tindouf Basin, Western Sahara) other phenomenon appears sometimes to have operated, causing the development of a very thin layer of apatite, separating each chamositic cortex. This has not been clearly observed in the deposit of the study area.

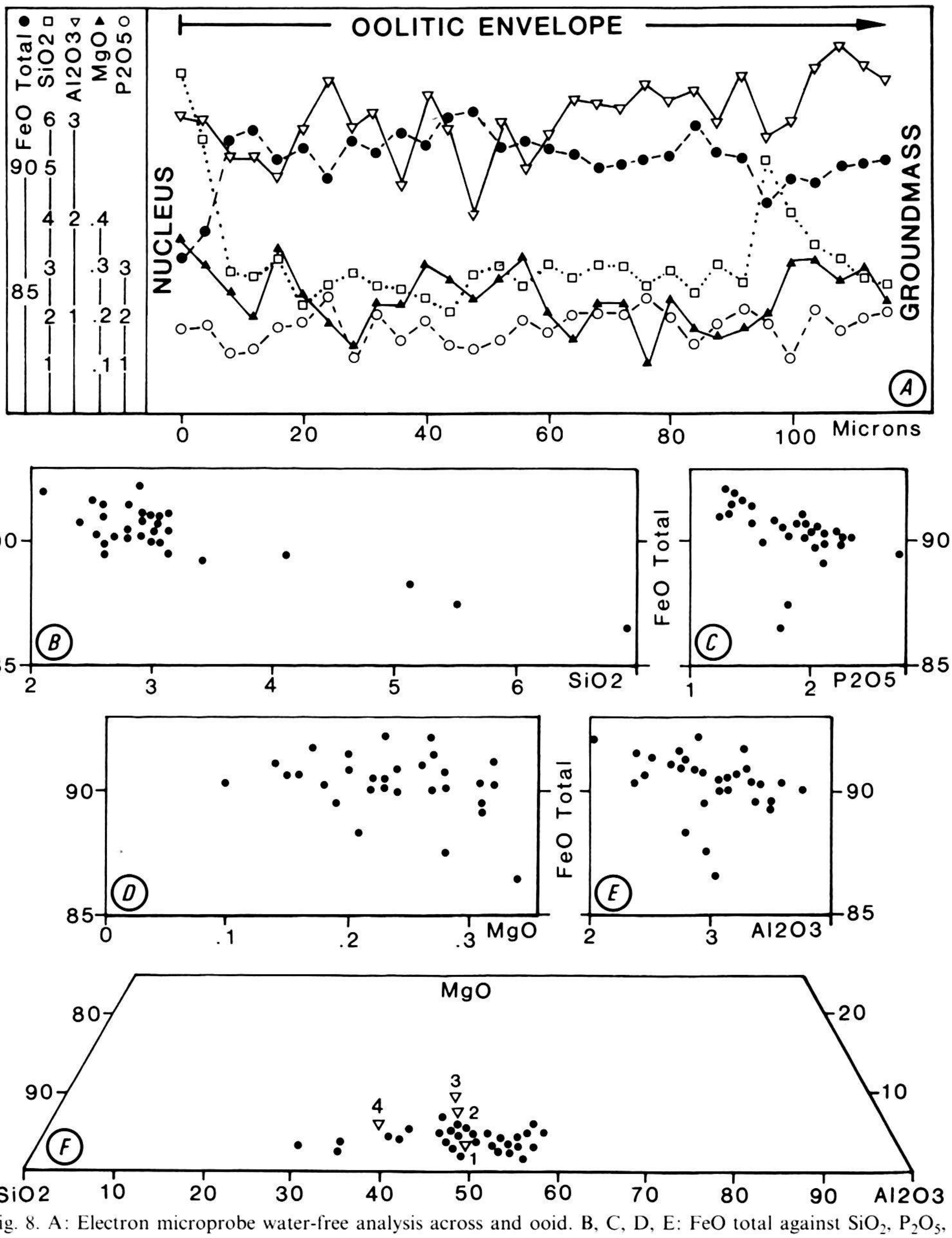


Fig. 8. A: Electron microprobe water-free analysis across and ooid. B, C, D, E: FeO total against SiO_2, P_2O_5, MgO and Al_2O_3 diagrams. F: compositional triangular diagram (SiO_2, MgO, Al_2O_3): reference chamosites are from 1: Wabana (MAYNARD 1986), 2: Japan (IIJIMA & MATSUMOTO 1982), 3: Gara Djebilet and 4: Mecheri Abdelaziz (GUERRAK, unpubl. data). Concentrations are in percentage weight.

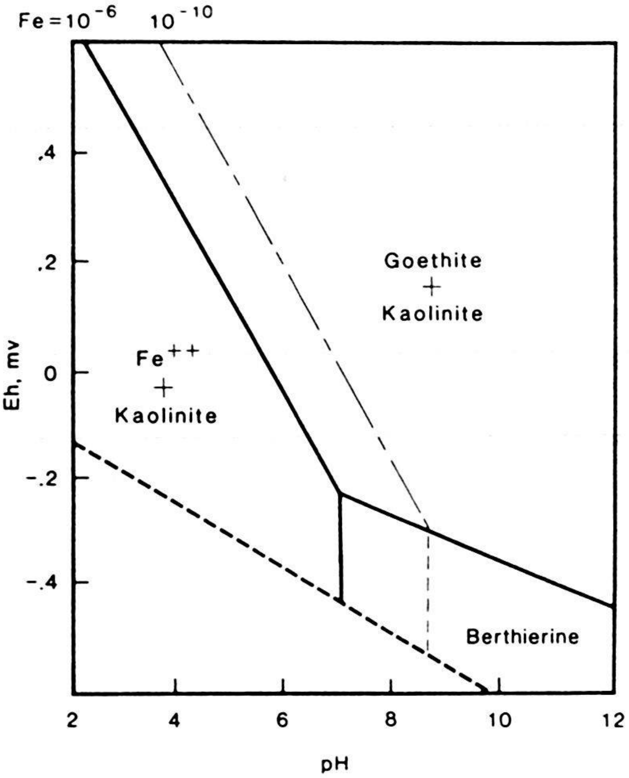
2. The occurrence of re-oolitized ooids (ooids submitted to two steps of coating separated by a step of transport), composite ooids (ooids composed of several ooids), and polynucleatide ooids (ooids composed of several nuclei of the same or different mineralogy) can be considered to be serious arguments against accretion processes in agitated waters. This indicate that the early ooids (superficial ooid or normal ooid, depending on accretion intensity) may have been originally formed in quiet environments. They have been reworked and kept in suspension within agitated waters, acquiring a detrital character, partly oxidized and sometimes broken. During each alternating period of accretion and suspension the ooid recorded the physicochemical characteristics of the sediment and the water.

After the completion of these processes, which explain the centrifugal increase state of oxidation of the ooids, the diagenesis of the groundmass and ooids brought out differences in the oxidation stage.

This proposed mode of oolitization needs not be universal, but seems to be a reasonable interpretation to explain the occurrence of the ferruginous ooids of the Saharan Platform.

Physicochemical conditions of formation of iron-rich minerals

According to the author's model of oolitization, the deposition of the ironstone seems to have primarily developed in tranquil waters. This opinion is in conformity with the observations of VAN HOUTEN & BHATTACHARYYA (1982) in several Phanerozoic ironstones.



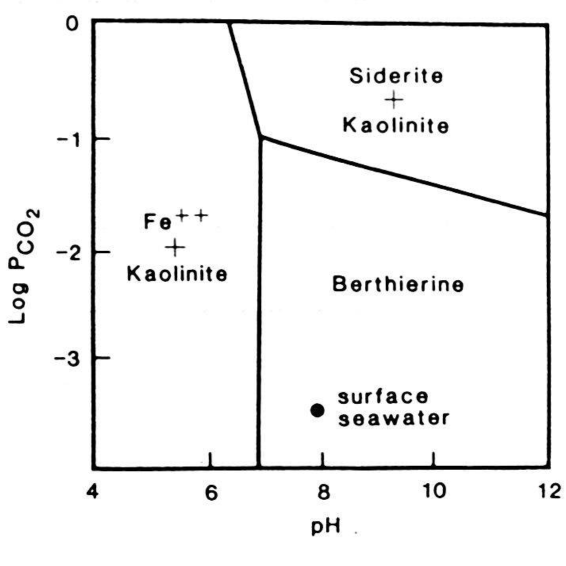


Fig. 9. Eh/pH and Log CO_2 pressure/pH diagrams of iron-rich minerals (after GARRELS & CHRIST 1965, modified by MAYNARD 1986).

The physicochemical conditions of the crystallisation of the iron minerals depend on the pH and Eh of the seawater and the concentration in iron-rich silicates.

The equilibrium diagrams of GARRELS & CHRIST (1965) modified by MAYNARD (1986), indicate the possibility of precipitation of iron-rich koalinite, goethite and berthierine in peculiar conditions of pH, Eh, and Co_2 pressure as indicated in Figure 9.

VAN HOUTEN & PURUCKER (1984) and MAYNARD (1986) suggest that chamosite can be produced by diagenetic transformation of Berthierine; IJIMA & MATSUMUTO (1982) estimating the temperatures of this transformation around 150 °C. On the other hand BHATTACHARYYA (1983) apparently obtained experimentally berthierine from kaolinite in similar conditions to those of early diagenesis. There must be a more or less direct relationship between these minerals, which is as follows:

iron-rich kaolinite→berthierine→chamosite.

However, CURTIS (1985) strongly pointed out that the burial can bring about various diagenetic transformations. Some chlorites can be produced, either by the evolution of smectite-chlorite or vermiculite-chlorite clays, or by direct precipitation.

From the above review, it is very clear that the formation of iron-rich minerals is not a simple phenomenon, but a complex one, involving several complementary processes.

Paleoenvironmental characteristics

The following main features characterize the occurrence of the oolitic ironstones of the study area: (i) the oolitic beds are located near or at the top of coarsening-upward sequences, (ii) they are often capped by continental iron crusts, thus indicating emersion periods, (iii) they are interbedded with formations showing rapid vertical lithologic variations and little lateral evolution, (iv) they are frequently associated with fine-grained sediments rather than the coarser-grained ones (as observed in Ougarta ranges: GUERRAK, unpublished data).

These observations suggest that the Devonian sedimentation can be related to a general epicontinental sedimentation on this northern pericratonic margin of Gondwana.

The model of the ironstone deposition paleoenvironment could have been a very flat widespread coastal clastic platform, containing lagoons and embayments separated from the offshore shelf by sand bars or barrier islands. The generalized section depicted in figure 2, indicates in the Devonian, sedimentary cycles consisting mostly of shallowing-upward sequences. The sequences which are accumulated episodically, are separated by scoured surfaces, mostly marked by iron crusts.

This model can be compared, on another scale, to the punctuated aggradational cycles model developed by GOODWIN & ANDERSON (1985).

Furthermore, the remarkable occurrence of the ironstones within the major or minor regressive cycles of the Devonian, permits the comparison of such oolitic sediments with the Jurassic ironstones of northwestern Europe that were described by HALLAM & BRADSHAW (1979). Thus, it can be concluded that the oolitic ironstones of Tassilis N'Ajjer and the Illizi Basin can be used as indicators of major or minor marine regressions.

On the other hand, the widespread extension of these ironstones of Extensive type (EXID) on several hundreds kilometers can be explained by an eustatic influence. Only such a phenomenon can allow the formation of so constant ironstone deposits: the irregularities in the granulometry and the thickness may be due to epeirogenic movements.

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Plate 1

Fig. 1 Oued Tachet area. Ironstone bed 5. Ironstone facies FMC. Polarizied reflected light. Scale bar: 100 μm. A well rounded ooid is mainly constituted by hematitic layers (He), finely underlined by chamositic laminae (Ch). The irregular originally chamositic nucleus is intirely oxidized in hematite.

Fig. 2 Fadnoun area. Ironstone bed 1. FOC/FOD facies type. Polarized reflected light. Scale bar: 50 μm. Alternating layers of chamosite (Ch) and hematite (He) coat quartz nucleus (Q).

Fig. 3. Fadnoun area. Ironstone bed 1. FOC/FOD facies type. Polarized reflected light. Scale bar is 50 μm. Two flattened ooids (spastoliths): one with a chamositic nucleus surrounded by a large hematitic cortex (He), the other one with a nucleus constituted by two quartz grains.

Fig. 4 Oued Karkaï area. Ironstone bed 3. FMC facies type. Polarized reflected light. Scale bar is 50 μm. An eccentric ooid is constituted by a pure chamositic nucleus (Ch) surrounded by a well marked hematitic ring (He), itself enveloped by fine layers of chamosite (Ch). Acicular hematite crystals (He) are well developed within the cortex.

Fig. 5 Fadnoun area. Ironstone bed 1. FOC/FOD ironstone facies. Polarized reflected light. Scale bar is 50 μm. Twofold quartz grains (Q) serve as nucleus to an hematitic ooid (He). Acicular hematite (He) occurs within the layers.

Fig. 6 Fadnoun area. Ironstone bed 1. FOC/FOD facies. Polarized reflected light. Scale bar is 50 μm. A flattened ooid shows successively four zones: an hematitic rich nucleus (He), a first chamositic cortex (Ch) surrounded by a more oxidized chamositic zone (Ch) itself enveloped by an entirely hematitic outer cortex (He).

Fig. 7 Fadnoun area. Ironstone bed 1. FOC/FOD facies. Polarized reflected light. Scale bar is 100 μm. This spastolith, completely flattened and distorted still shows the interlayered structure of chamosite (grey, Ch) and hematite (white, He).

Fig. 8 Oued Karkai area Ironstone bed 3. FOC facies type. Polarized reflected light. Scale bar is 100 μm. A primarly lenghtened ooid with a broken quartz nucleus (Q) is surrounded by an envelope of pure hematite (He) and by layers of oxidized chamosite (Ch/He).

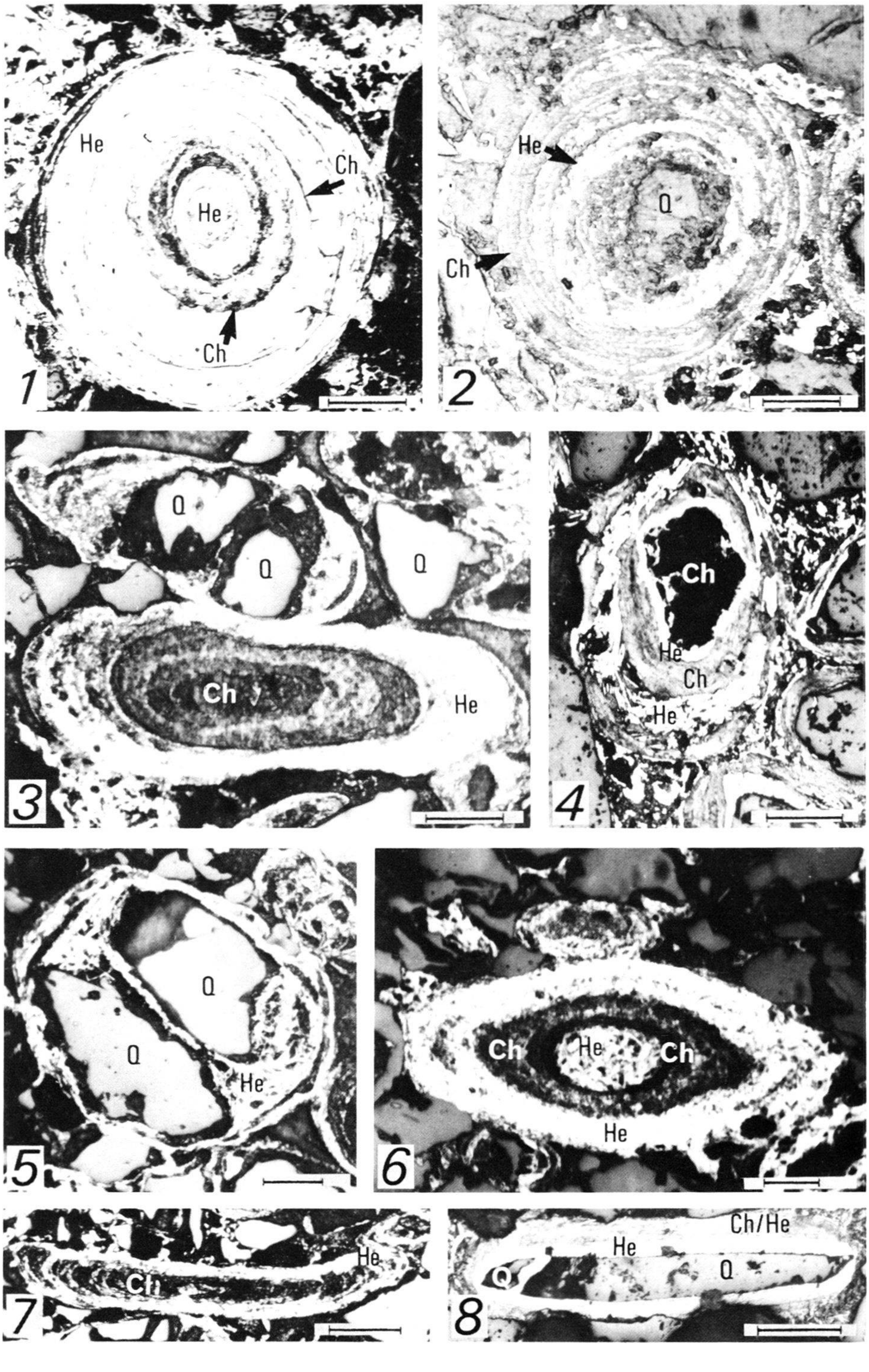

1
He
Ch
He
Ch
2
He
Ch
Q
3
Q
Q
Q
Ch
He
4
Ch
He
Ch
He
5
Q
Q
He
6
Ch
He
Ch
He
7
He
Ch
8
Ch/He
He
Q
Q

Plate 2

Fig. 1 Fadnoun area. Ironstone bed 1. FOC/FOD facies type. Polarizing reflected light. Scale bar is 50 µm. A complex spastolith consists of two distorted ooids coated together by an outer hematitic laminae (He). One ooid presents a quartz nucleus (Q) whereas within the second one, the chamositic nucleus (Ch) develops a "bead structure".

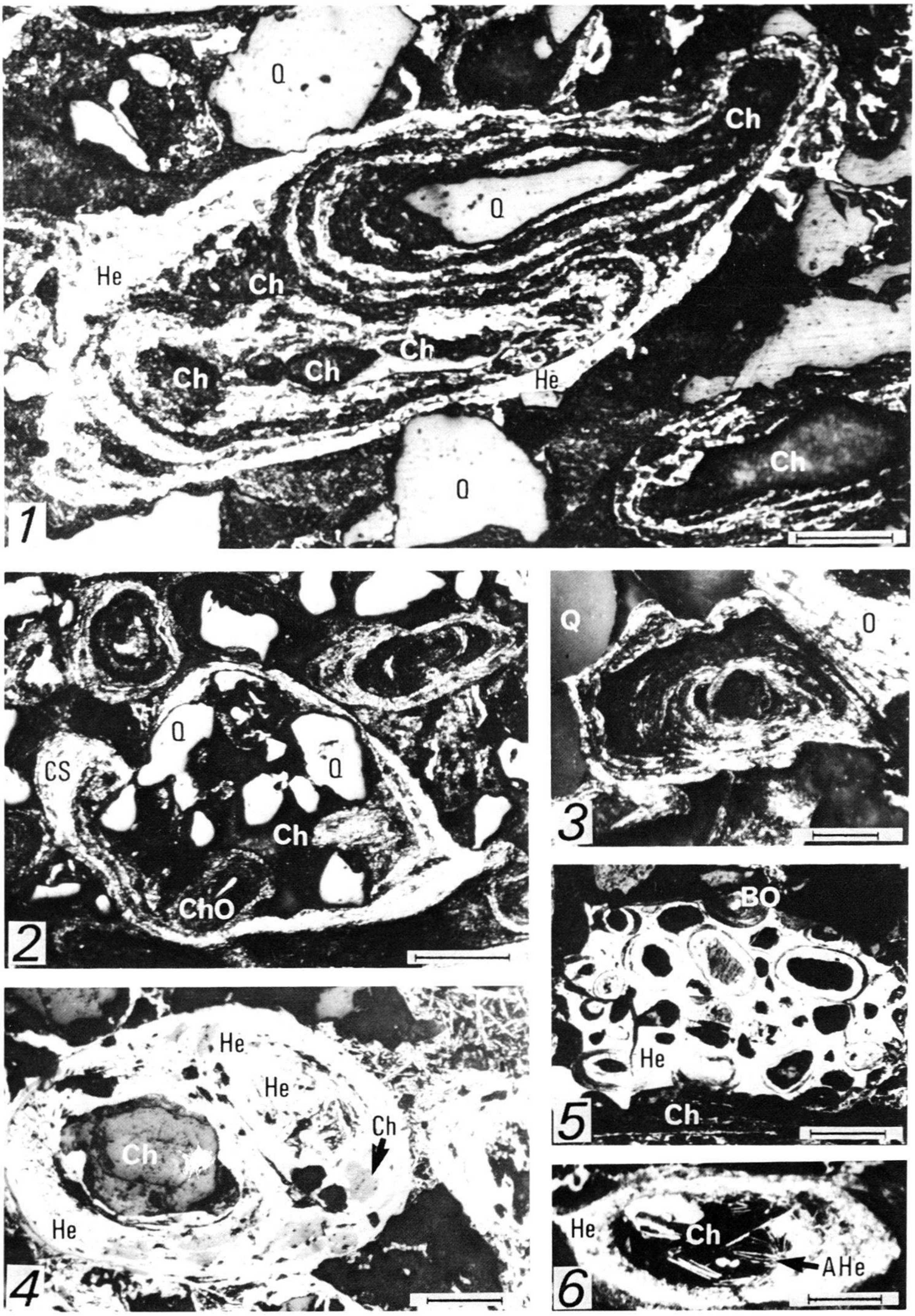
Fig. 2 Fadnoun area. Ironstone bed 1. FOC/FOD facies type. Polarizing reflected light. Scale bar is 150 µm. A complex distorted ooid (Cs) contains non distorted chamositic ooids (ChO) and quartz grains (Q), scattered within a chamositic groundmass (Ch).

Fig. 3 Fadnoun area. Ironstone bed 1. FOC/FOD facies type Oil emersion polarizing reflected light. Scale bar is 50 µm. An hematitic-chamositic spastolith is stretched to the ends and closely pressed against quartz grains (Q) and other ooids (O).

Fig. 4 Oued Karkaï area. Ironstone bed 3. FOC facies type. Polarizing reflected light. Scale bar is 50 µm. A complex ooid consists of two hematitic ooids (He), one with a chamositic nucleus, and the other with an hematitic nucleus (He). Remnants of chamositic layers occur as grey patches (Ch).

Fig. 5 Oued Tachet area. Ironstone bed 5. FMC facies type. Polarized reflected light. Scale bar is 200 µm. An intraclast with hematitic cement (He) is included within a chloritic groundmass (Ch). Some broken ooids (Bo) occur within the intraclast.

Fig. 6 Fadnoun area. Ironstone bed 1. FOC/FOD facies type. Polarized reflected light (oil emersion). Scale bar is 50 µm. This flattened hematitic ooid (He) contains acicular hematite (AHe) occurring within a chamositic nucleus (Ch).


1
Q
Ch
Q
He
Ch
Ch
Ch
Ch
He
Q
Ch
2
Q
CS
Q
Ch
ChO
3
Q
O
4
He
He
Ch
Ch
He
5
BO
He
Ch
6
He
Ch
AHe