Oolitic iron formations : marine or not marine?

Autor(en): **Gygi, Reinhart A.**

Objekttyp: **Article**

Zeitschrift: **Eclogae Geologicae Helvetiae**

Band (Jahr): **74 (1981)**

Heft 1

PDF erstellt am: **21.09.2024**

Persistenter Link: https://doi.org/10.5169/seals-165102

Eclogae geol. Helv.	Vol. 74/1	Pages 233-254	4 figures in the text and 1 table	Basle, March 1981

Oolitic iron formations: marine or not marine?

By Reinhart A. Gygi[1])

ABSTRACT

A model is proposed for oolitic ironstone formation in open, shallow to deeper marine environments. Iron is derived from the land. It is concentrated in the sea by a two-stage process. Iron ooids are formed at the distal fringe of an argillaceous, terrigenous sediment fan in an environment of slow mud deposition below the normal wave base. The model predicts that formation of goethite or limonite ooids side to side with chamosite ooids can take place in normally aerated sea water from moderate depths down to about 100 m. The hitherto unknown origin of iron ooids with alternating crusts of chamosite and goethite/limonite can be understood in terms of this model. No special submarine topography has to be postulated. The origin of oolitic ironstones with a very large areal extent and a diverse ammonite fauna, like those of the Middle and earliest Late Jurassic of east-central France, Switzerland, and southern Germany, can be explained by the model. The model may be applicable to Jurassic iron oolites in northern Germany and southern England too. It is derived from a detailed paleoecological analysis of the iron oolitic Schellenbrücke Bed (Late Jurassic) in northern Switzerland. In this bed, goethite as well as chamosite iron ooids were accreted in an environment supporting abundant and diverse life, in water about 80-100 m deep.

The principal conclusion is that known oolitic ironstones probably have been formed in more than a single mode. There is overwhelming evidence for the marine origin of many European oolitic ironstones, but at the present time, a nonmarine diagenetic mode of formation can not be excluded in some other cases.

ZUSAMMENFASSUNG

Ein Modell für die Eisenoolith-Entstehung in einem offenmarinen Milieu bei geringer bis grösserer Wassertiefe wird vorgeschlagen. Das Eisen stammt von der lateritischen Verwitterung auf dem Land. Seine Konzentration im Meer findet in einem zweistufigen Prozess statt. Die Eisenooide bilden sich unter der normalen Wellenbasis am distalen Rand eines terrigenen, tonhaltigen Schlammfächers bei stark verlangsamter Sedimentation. Chamositooide können sich am gleichen Ort wie Goethit- und Limonitooide in gut belüftetem Wasser bilden, in einem bathymetrischen Intervall von ziemlich seichtem bis etwa 100 m tiefem Wasser. Das Modell gibt auch eine Erklärung für die Entstehung der bis jetzt schwer zu interpretierenden Eisenooide mit alternierenden Rinden aus Chamosit und Goethit/Limonit. Für die Ooidbildung muss keine besondere submarine Topographie postuliert werden. Die Genese von eisenoolithischen Horizonten mit einer sehr grossen horizontalen Ausdehnung und artenreichen Ammonitenfaunen wird verständlich, wie etwa die der Horizonte im Mittleren Jura und im unteren Oberjura des östlichen Mittelfrankreich sowie des Schweizer und des süddeutschen Jura. Das Modell lässt sich möglicherweise in modifizierter Form auch auf die Eisenoolithe des Jura in Norddeutschland und in Südengland anwenden. Es wurde aufgrund einer detaillierten palökologischen Analyse der eisenoolithischen Schellenbrücke-Schicht im Oberen Jura der Nordschweiz entworfen. In dieser Schicht entstanden nebeneinander Goethit- und Chamositooide bei einer Wassertiefe von 80 bis 100 m in einem Milieu, welches einer häufigen und artenreichen Ammonitenfauna günstige Lebensbedingungen bot.

Die wichtigste Schlussfolgerung ist die, dass sich die bis jetzt bekannten Eisenoolithe wahrscheinlich nicht alle auf die gleiche Weise gebildet haben. Bei vielen europäischen Eisenoolithen darf eine marine

[1]) Museum of Natural History, Augustinergasse 2, CH-4051 Basel.

Genese als erwiesen angenommen werden, doch muss gegenwärtig bei gewissen anderen Vorkommen auch mit der Möglichkeit einer nichtmarinen, diagenetischen Entstehung gerechnet werden.

RÉSUMÉ

Un modèle est proposé pour la formation d'oolites ferrugineuses dans un milieu d'une mer ouverte plus ou moins profonde. Le fer est dérivé de l'altération latéritique terrigène. Il est concentré dans la mer par un processus en deux étapes. Les ooides ferrugineuses sont formées au bord distal d'un dépôt de sédiments terrigènes argileux dans un milieu de sédimentation réduite en-dessous de la base normale des vagues. Le modèle prédit que les ooides de goethite ou de limonite peuvent se former au même endroit que des ooides à chamosite dans une mer normalement aérée, dans un intervalle bathymétrique de profondeurs modérées à environ 100 m. L'origine d'ooides avec des croûtes alternantes à chamosite et à goethite/limonite, qui était inconnue jusqu'à présent, devient compréhensible par ce modèle. Il n'est pas nécessaire de postuler une topographie sous-marine particulière. Le modèle fournit une explication de l'origine des oolites ferrugineuses avec une très vaste extension latérale et avec une riche et diverse faune d'ammonites, comme celles du Jurassique moyen et supérieur du centre-est de la France, de la Suisse septentrionale et du sud de l'Allemagne. Il est possible que le modèle soit applicable aux oolites ferrugineuses du nord de l'Allemagne et aussi du sud de l'Angleterre. Le modèle est conçu d'après une analyse détaillée palécologique du dépôt à ooides ferrugineuses de la Schellenbrücke (Jurassique supérieur) au nord de la Suisse. Dans cette couche, l'accrétion d'ooides de goethite et de chamosite a eu lieu à une profondeur d'environ 80–100 m, dans un milieu qui permettait le développement d'une abondante et diverse faune d'ammonites.

La conclusion principale est que les oolites ferrugineuses connues se sont probablement formées de plus d'une seule façon. On peut admettre avec certitude une origine marine pour un nombre important d'oolites ferrugineuses d'Europe, mais à présent on ne peut pas exclure une formation non marine diagénétique dans certains autres cas.

CONTENTS

1. Introduction

The origin of oolitic iron formations is debated to this day. Most English (HALLAM 1966, HALLAM & BRADSHAW 1979) and some German students (ALDINGER 1957*a)* of the European Jurassic envisaged the formation of iron ooids to occur in an open, shallow marine environment. Others pointed out that chamosite is stable only in the absence of free oxygen, and consequently advocated a concretionary origin of chamositic ooids during diagenesis (GEBERT 1964, SCHELLMANN 1969). One shortcoming of the latter hypothesis is obvious. Iron ooids with alternating concentric crusts of goethite and chamosite have been described and figured by several authors. There is no satisfactory explanation how such particles can be formed during diagenesis. Lateritic spherules washed into the sea and being concentrated in nearshore marine waters seem to be another possibility (SIEHL & THEIN 1978). LEMOALLE & DUPONT (1973) presented evidence that goethitic iron ooids are now being accreted in the shallow waters of brackish Lake Chad (Africa). Recently, KIMBERLEY (1979) has proposed a revised version of a model which he apparently thought to be universally valid for the formation of all known oolitic ironstones.

The assertion of KIMBERLEY's model is that in the course of an upward shallowing cycle, aragonitic oolite in an "inland sea" is covered by a sheet of argillaceous, terrigenous mud. Continuing deltaic deposition will eventually raise the sediment surface above sea level, thus allowing fresh water to percolate the argillaceous cover and the underlying oolite. Iron and aluminium are then eluviated from clay minerals, and replace aragonitic ooids completely by chamosite. KIMBERLEY (1979, p. 123) drastically illustrated the old perception that the amount of iron dissolved in seawater is so small that direct precipitation of iron from normal seawater could not lead to the deposition of significant iron orebodies. Several times the volume of the world's oceans would have been necessary to produce the large Pliocene oolitic iron formation of Kerch (USSR, see SOKOLOVA 1964).

BRADSHAW et al. (1980) promptly challenged KIMBERLEY's implicit assertion that all known oolitic iron formations could have been formed only diagenetically on land. In a reply, KIMBERLEY (1980) rejected the arguments of BRADSHAW et al. (1980). At the present time we are not aware of facts proving the model of KIMBERLEY (1979) to be wrong. Therefore, the origin of oolitic ironstone by diagenesis on land must be considered as one possible mode of formation. But, as we shall see, KIMBERLEY's model in its present form is inconclusive. The question now is whether there is sufficient evidence that iron ooids can be formed in the marine environment, as many European workers have maintained. It is the purpose of this paper first to present new evidence that iron ooids *can* be formed in normally aerated seawater, in a depth as great as 100 m. This is concluded from sedimentologic, mineralogic, and paleontologic observations in an iron oolitic horizon of Late Jurassic age in northern Switzerland. Then, the results are combined into a model for iron ooid formation in an open, deeper marine environment. Finally, the applicability of the model to other oolitic iron formations is shortly discussed.

2. The Schellenbrücke Bed

The horizon has only recently been defined and named (GYGI 1977, p. 454). At Herznach, canton Aargau, it is the top bed of a thick iron oolitic succession (Fig. 1). Sedimentation of this facies began with a condensed bed (A 5 of JEANNET 1951) in the subchron of the *Kosmoceras (Zugokosmoceras) enodatum* Subzone, or in late Lower Callovian time (THIERRY 1978), and ended about at the turn between the *Cardioceras (Cardioceras) cordatum* Zone and the *Cardioceras (Subvertebriceras) densiplicatum* Zone (SYKES & SURLYK 1976, p. 423; SYKES & CALLOMON 1979, Fig. 3), or at the end of the Lower Oxfordian (Fig. 2, see also GYGI 1969, p. 101). The lower part of the succession (Lower and Middle Callovian) has been mined for iron ore during the Second World War and some time after. The whole section to the Middle Oxfordian is therefore well-known (JEANNET 1951, GYGI 1977). There are no sediments other than iron oolitic anywhere in the succession. Only the percentage of

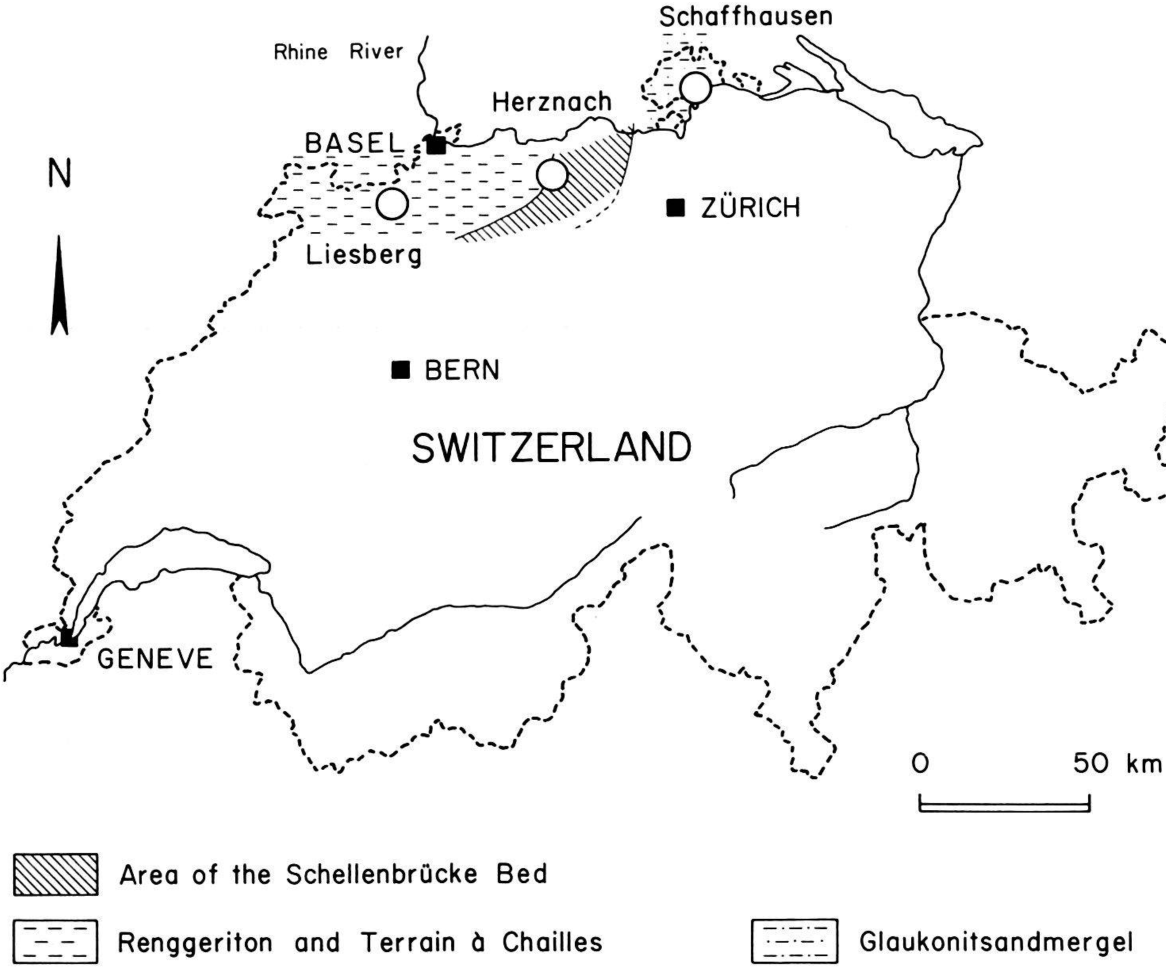

Fig. 1. Geographic range of the Schellenbrücke Bed.

iron ooids in the rocks varied widely with time. Deposition was not continuous. A major hiatus encompassing at least three subzones could be documented by ammonites between the nodular bands of the Scarburgense Subzone (Fig. 2, earliest Oxfordian) and the Schellenbrücke Bed of the Cordatum Subzone, the last subzone of the Lower Oxfordian (MARCHAND & GYGI 1977). In the region of Herznach, iron oolitic sediments have thus been deposited intermittently and at variable rates during a time of about 6 my (VAN HINTE 1976, GYGI & MCDOWELL 1970).

2.1 *Depositional history*

The nodular bed (cf. Fig. 2) of stage *A* in Figure 3 was formed during Scarburgense time, as documented by three *Pavloviceras pavlowi* found in situ in the uppermost nodular horizon (GYGI & MARCHAND 1981). At one place in the Herznach iron mine, the lowest nodular horizon grades into a continuous limestone band. This, and the fact that there is sometimes a lateral transition from nodules to marl, indicates that the nodules have been formed during early diagenesis out of a primary, iron oolitic succession of marl and limey marl, in a process described by HALLAM (1964). This assumption is corroborated by the fact that the abundance of

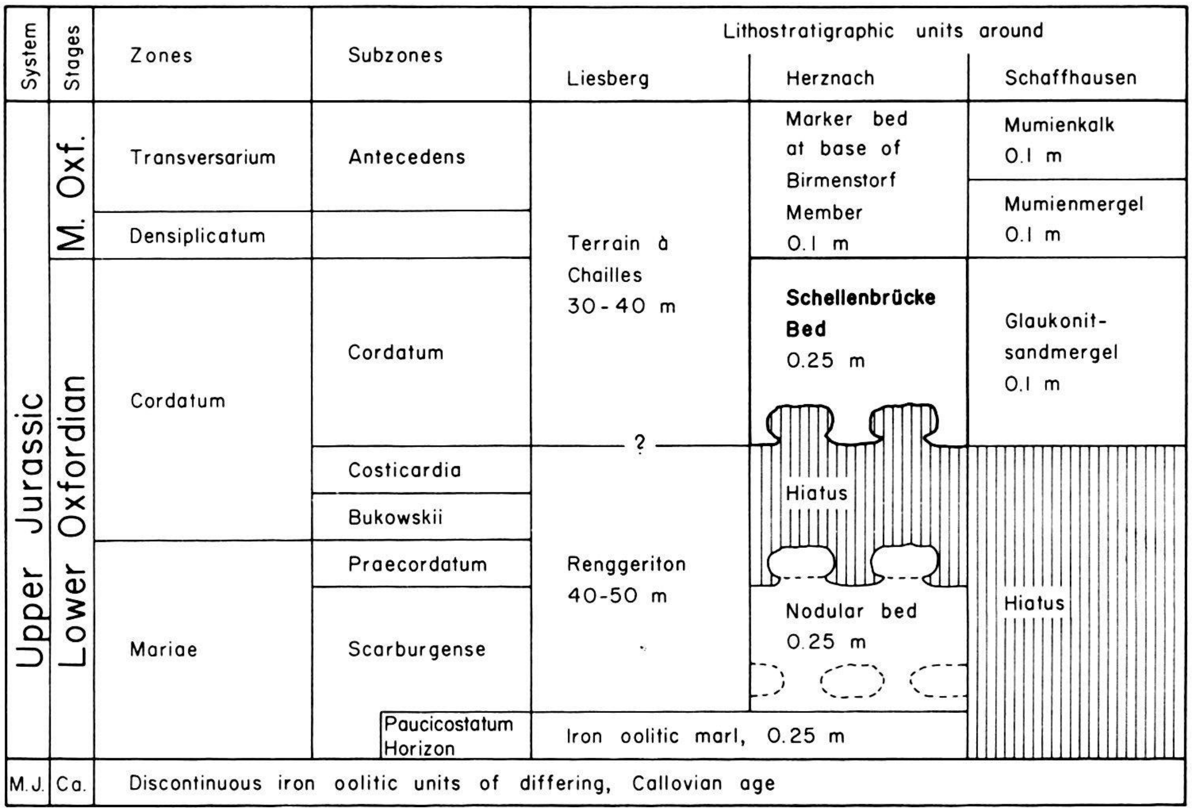

Fig. 2. Chronostratigraphic position and correlation of the Schellenbrücke Bed. Note the great difference of thickness in lithostratigraphic units between Liesberg and Herznach.

iron ooids is typically 5–10% in the nodules, and may exceed 20% in the interstitial marl. The iron ooids in the marl are always oxidized. In the uppermost nodular band as much as 20% of the iron ooids are in part or wholly chamosite. Some of the marl between the upper nodules has been scoured by a strong current prior to the onset of deposition of the Schellenbrücke Bed in Cordatum Subzone time. The bare surfaces of the nodules were then subject to bioerosion. Boring microorganisms, fungi or algae, made the surfaces irregular and rough. This even led to fragmentation of some of the nodules (Fig. 3*B*). The irregular shape and size of the fragments is evidence that they are not mechanically broken and scoured pebbles, as would have been formed if the surface of stage *B* had been emergent in the intertidal zone at any time. In Cordatum Subzone time there was first a short term of sedimentation. A thin deposit of ferruginous lime mud (Fig. 3*C*) made partial to near-complete fossilization of a few *Cardioceras* s.str. possible. Induration of the ammonite casts must have taken place before that of the enveloping sediment. All of this sediment has been whirled up and carried away by a current shortly after its deposition (Fig. 3*D*). Ammonite casts and parts of such (often indistinguishable from lithoclasts) lay bare at the sediment surface for some time. There they became encrusted with limonite. Some of the ammonite casts and "lithoclasts" were then dug into the ferruginous, much older marl under them by burrowing animals (which are not preserved), thus being placed below older ammonites of the Scarburgense Subzone (Fig. 3*E*). The absence of calcareous sediment in the burrows is proof that the lime mud deposited at stage *C* had been entirely removed prior to burrowing.

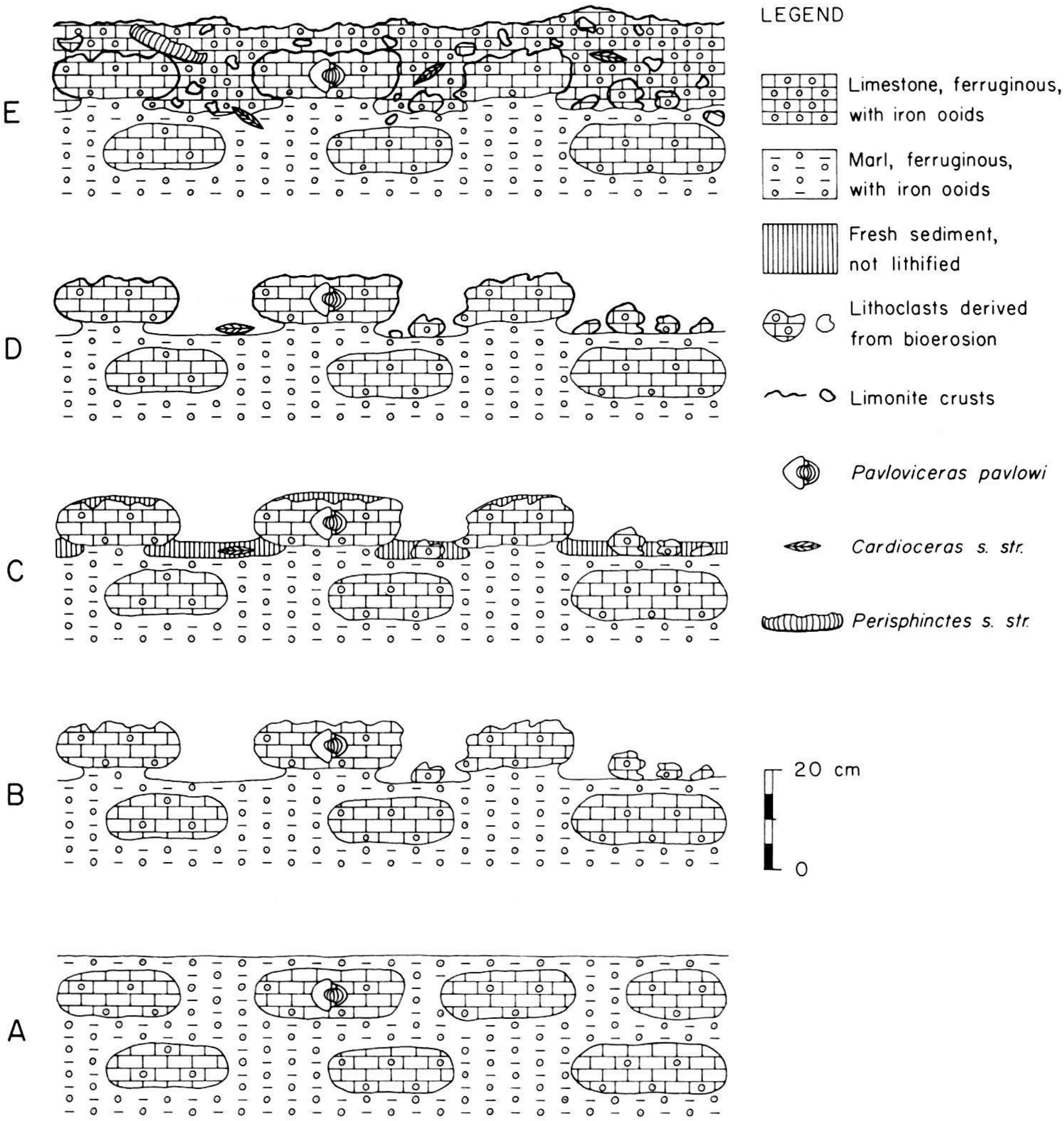

Fig. 3. Depositional history of the Schellenbrücke Bed. Explanation in the text.

Subsequent sedimentation of ferruginous lime mud (Fig. 3*E*) was again rapid and episodic. Two more events of erosion can be recognized to have occurred between times of deposition at stage *E* (GYGI & MARCHAND 1981). Erosion was followed by prolonged intervals when bio- and true lithoclasts were covered by limonite crusts. Thus has been formed a pseudo-conglomerate which resembles the so-called Conglomérat de Bayeux (Bajocian) in northern France, as figured by FÜRSICH (1971, Fig. 1c). The large clasts covered by iron oxide crusts compare well with the "snuff-boxes" as described and figured by GATRALL et al. (1972). Serpulids grew at the underside of undercut nodules, as at stage *B*. Small clasts with a diameter not exceeding 6 or 7 cm may be encrusted all around, indicating that they have been overturned. The small lithoclasts are in fact mostly bioclasts at various stages of incomplete preservation, which may make the recognition of the original fossil (mostly ammonites) impossible. Larger bioclasts of *Perisphinctes*, with a diameter of as much as 20 cm, have been encrusted with limonite on the upper surface only. Some of them were then overturned, and the lower surface, now exposed, was corroded by bioerosion. Thin stromatolitic crusts may occur on such corroded surfaces. Matrix, litho- and bioclasts of the Schellenbrücke Bed have an average content of about 10% iron ooids. Most of these are oxidized near Herznach, but in other sections the majority may consist of pure chamosite (GYGI 1969). Sedimentation of the Schellenbrücke Bed was terminated by the development of a hardground. This was capped by a continuous crust of limonite or chamosite with a shining, verrucose surface. The crust can be followed over a distance of many tens of kilometers (see GYGI 1969, p. 97). There are sections where the Schellenbrücke Bed is reduced to this crust as a result of extreme condensation.

2.2 *Fauna and taphonomy*

In the interior of the Schellenbrücke Bed, a sparse soft ground-burrowing infauna is represented by few echinoids *(Collyrites)* and some bivalves *(Pholadomya)*. The length of *Pholadomya* individuals does not exceed 5 cm. All *Pholadomya* are probably juveniles. The actively suspension-feeding siliceous sponges can only be recognized in thin section. No whole specimens are present. There are but small parts of calcitized skeletons, and these usually grade into the matrix without a well-defined boundary. The parts appear to be the portions of whole, macerated skeletons (ZIEGLER 1964) which happened to be buried and thus preserved by small amounts of sporadically deposited lime mud. Siliceous sponges must then have been

Table: *Faunal assemblage of the Schellenbrücke Bed and of the time-equivalent Glaukonitsandmergel in northeastern Switzerland.*

			Schellenbrücke Bed			Glaukonitsandmergel		
Porifera			P	-		1	0.2%	
Brachiopoda:	Rhynchonellida		15	1.0%		6	1.1%	
	Terebratulacea		13	0.8%		3	0.5%	
Bivalvia			41	2.6%		32	5.7%	
Gastropoda			123	7.9%		9	1.6%	
Cephalopoda:	Nautilaceae		7	0.4%		-	-	
	Ammonoidea	Phylloceratidae	14	0.9%	85.6%	2	0.4%	68.1%
		Oppeliidae	124	7.9%		132	23.4%	
		Perisphinctidae	565	36.2%		162	28.6%	
		Cardioceratidae	568	36.4%		73	12.9%	
		Aspidoceratidae	65	4.2%		16	2.8%	
	Belemnoidea		7	0.4%		102	18.0%	
Echinodermata:	Echinoidea		9	0.6%		5	0.9%	
	Crinoidea: stem fragments		-	-		4	0.7%	
Vertebrata:	individual shark teeth		10	0.7%		18	3.2%	
Total number of specimens:			1561	100%		565	100%	

P: Parts of incompletely fossilized specimens present more abundant in the biocenosis than in the taphocenosis recorded in the Table. Crinoids are passive suspension feeders. They are absent at the excavated sites in the Schellenbrücke Bed. This can hardly be the consequence of unfavourable conditions of fossilization due to prevailing nondeposition. The absence of crinoids may be primary because of bottom currents being normally so slack that only active suspension feeders like sponges or brachiopods could obtain an adequate food supply. Gastropods, mostly pleurotomariids, are the dominant element of the vagile benthos. Cephalopods, most of which are ammonites, are most common by far in the taphocenosis. Boreal Cardioceratidae and tethyan Perisphinctidae make up two large groups of almost equal size (Table). Oppeliidae and Aspidoceratidae are much less abundant, and Phylloceratidae are rare. The predominance of ammonites may have been even greater in the biocenosis than in the taphocenosis. For instance, most brachiopods found in the excavations are well-preserved, whereas ammonites were often so incomplete that they were not collected (GYGI 1977, p.438). The incompleteness of these ammonites was caused by partial fossilization. The calcium carbonate in the ammonite shell is all aragonite. Dissolution of this mineral could begin early in diagenesis, often before the sediment which had got into the shell after the death of the animal was entirely lithified (see below).

2.3 *Facies relationships*

Oxfordian sedimentation in the central Jura mountains began with an iron oolitic marl. This condensed horizon, with a rich, almost pure ammonite fauna including *Cardioceras paucicostatum*, has been formed during the earliest part of the first Oxfordian Subzone, and is only 0.25 m thick (Fig. 2, left side of Fig. 4). Above are 40–50 m of marl-clay grading upwards into marl, the Renggeriton (Fig. 4). In the lowermost 5 m of the Renggeriton are *Scarburgiceras scarburgense* and *Pavloviceras*. These ammonites occur also at Herznach, in the uppermost band of nodules which are incorporated into the Schellenbrücke Bed (Fig. 2 and 3). Above the Lower Oxfordian Renggeriton of the central Jura mountains there is a succession of dark grey marl with bands of calcareous nodules, the so-called Terrain à Chailles (Fig. 4). ROLLIER (1888) has confirmed the conclusion of OPPEL (1856–8) that the thickness

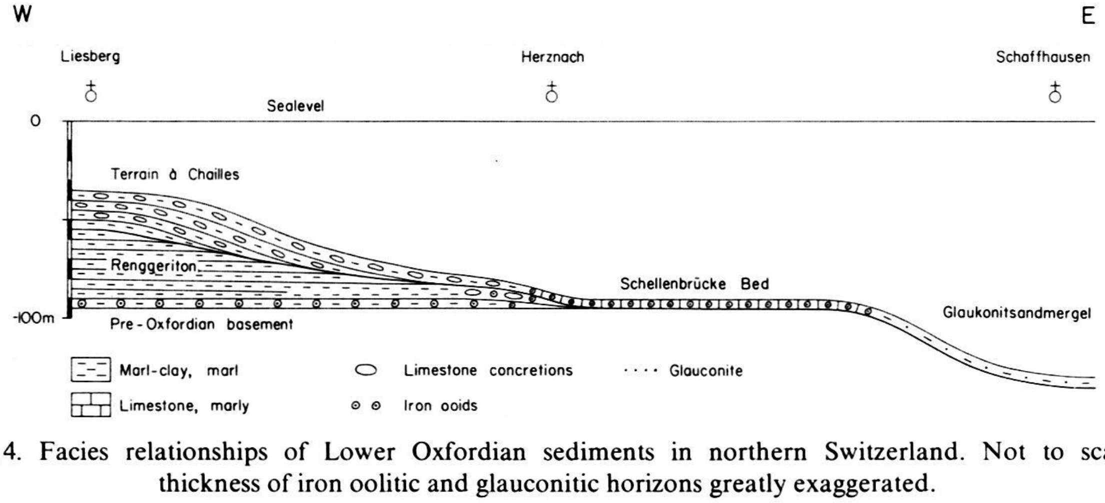

Fig. 4. Facies relationships of Lower Oxfordian sediments in northern Switzerland. Not to scale, thickness of iron oolitic and glauconitic horizons greatly exaggerated.

of the argillaceous Lower Oxfordian sediments is drastically reduced towards the south, where the succession grades laterally into a thin, iron oolitic horizon. It is apparent from an article by KOBY (1899) that only the lower part of the Terrain à Chailles is time-equivalent with the Schellenbrücke Bed proper (Cordatum Subzone, MARCHAND & GYGI 1977). A few ammonites only could recently be collected in situ in the lower Terrain à Chailles. They confirm KOBY's work, but at present only a rough estimate can be made of the portion of the Terrain à Chailles belonging to the Cordatum Subzone. The thickness must be of the order of 10 to 15 m. It decreases to the south and to the east. Where the thickness is reduced to less than about 1 m, the facies becomes iron oolitic (GYGI 1969) and is then called the Schellenbrücke Bed (GYGI 1977). North of the Rhine River, this thin horizon grades into an equally condensed marl-clay (Fig. 1, 2, and 4) with as much as 30% of glauconite (GYGI 1969). This has been named Glaukonitsandmergel by ZEISS (1955). The marl-clay has a dark-green colour. Its upper surface is brown as a result of oxidation. This thin oxidized horizon is time-equivalent with the ferruginous crust on top of the Schellenbrücke Bed. The Glaukonitsandmergel has been dated radiometrically by GYGI & MCDOWELL (1970). The average thickness of the Schellenbrücke Bed as well as the Glaukonitsandmergel is of the order of 10 cm.

2.4 *Environment of deposition*

The sediment of earliest Oxfordian age in northern Switzerland (Paucicostatum Horizon at the base of the Scarburgense Subzone, see MARCHAND in DEBRAND-PASSARD et al. 1978, and this paper, Fig. 2) is condensed. Its maximum thickness amounts to only a few decimeters, and drops to zero east of Herznach (lowermost horizon on the left side of Fig. 4, see also GYGI 1977, Pl. 11, section 2, bed 6). Iron ooids occur in varying amounts in a matrix of argillaceous mud, together with almost pure, diverse cephalopod faunas which are abundant in some places. Sediment as well as fauna suggest deposition below the normal wave base in water moved by weak currents. Adequate aeration and normal salinity of the water may be inferred from the fauna, although it is now apparent that not all cephalopods of the Recent are strictly stenohaline (BRAKONIECKI 1980). The muddy matrix of the sediment indicates that the water directly above was normally quiet. Thus, the minimal deposition rate probably reflects an insignificant influx of terrigenous clay. Seawater must then have been highly transparent, with the photic zone reaching to an exceptional depth. Most of the older sediments in this region were lithified at the onset of Oxfordian deposition. Hard substrate, very clear water, and subsequent sedimentation at an extraordinarily low rate would have been ideal for hermatypic coral growth to start. However, there are no corals at all. If too great a depth alone were the cause for their absence, a minimum depth of about 80 m would have to be inferred because of the probably unusual water transparency (see JAMES & GINSBURG 1979, and BUDD & PERKINS 1980, p. 893). As we have seen, insufficient bottom currents could have been another limiting factor, and the temperature may have been too low. The whole area of what is now northwestern Switzerland was apparently a wide and relatively deep open marine shelf.

At the time the first *Scarburgiceras scarburgense* and *Creniceras renggeri* appeared, sedimentation of bluish grey marl-clay (Renggeriton) began in the central Jura mountains (Fig. 4). The rate of deposition markedly increased, and the argillaceous sediment fan prograded to the south and east. The nodular bed (Fig. 2) of stage *A* in Figure 3 is the condensed, distal facies at the fringe of the fan (see Fig. 4 or Gygi 1977, Pl. 11, section 2, beds 7 and 8). The hiatus between the nodular bed and the Schellenbrücke Bed in the region of Herznach (Fig. 2) points to a regional increase in water depth during the Mariae Chron. Owing to this, the outer limit of sedimentation receded, and a nonsequence developed in the distal part of the Lower Oxfordian sediment lobe (Fig. 4). Ammonites are still fairly abundant in the part of the Terrain à Chailles which is of Cordatum Subzone age. Further up in the succession, the proportion of bivalves increases, indicating that sedimentation outweighed subsidence or eustatic sea level rise at the turn from Lower to Middle Oxfordian. Concomitant with shoaling was a general increase of the carbonate content in the sediment. In the upper (Middle Oxfordian) part of the Terrain à Chailles, the bivalve *Pholadomya* became very abundant, and hermatypic corals made their first, patchy appearance (Koby 1899, p. 211, 217).

In the Schellenbrücke Bed, there is some direct sedimentological evidence for water depth. The demonstrably complete removal of fresh unconsolidated sediment on several occasions could have been accomplished only by an essentially unidirectional current. Stirring up of loose sediment in normally quiet bottom waters may be brought about in moderately deep water by a storm, or in water of any depth by an earthquake-generated tsunami wave (Lucas 1966). Water motion induced by a tsunami may be inferred to be oscillant without a superimposed directional component. There is no indication that the Schellenbrücke Bed was deposited on a surface with an appreciable slope. On a level bottom, sediment resuspended by a tsunami would settle essentially in the same area after water movement had died down. Storms appear to be a better explanation to account for infrequent but strong currents that could sweep away unconsolidated deposit entirely from one place (Fig. 3*D*), and rapidly resediment the same small amounts of deposit elsewhere, as at stage *C* in Figure 3. This sets a maximum limit to the possible depth of water in which the Schellenbrücke Bed has been laid down.

The mode of preservation (taphonomy) of ammonites puts a further constraint on water depth. Broken casts were fractured by differential compaction, for instance when buried above a lithoclast. The fragments are always close together. The fracture edges of the casts are never abraded, but they do sometimes show traces of plastic deformation which took place after the shell was partly or wholly dissolved and prior to complete induration of the cast. This is a strong argument for deposition below the normal wave base, as is the absence of fragments of empty ammonite shells (cf. Ziegler 1962, Fig. 2). The random orientation of ammonite casts, probably due to the action of burrowing animals, points in the same direction. Kuenen (1964, p. 228) stated that storms may displace rocks half a kilogram in weight at depths of over 60 m. The common fossilization of delicate spines in *Cardioceras* is compelling evidence that the sediment enveloping bio- and lithoclasts of the Schellenbrücke Bed was often lithified before a current capable of displacing more than mud-grade sediment intervened.

Overturn of clasts by currents was then unusual, if it happened at all, even though events of erosion must have occurred more often than can be read from the sediment, since loose deposit was not normally present. Well-developed stromatolites (GYGI 1969, Pl. 1, Fig. 2), borings by microorganisms in ammonite shells or in small bioclasts (GYGI 1969, Pl. 2, Fig. 5), and iron oxide crusts in the Schellenbrücke Bed might be interpreted as evidence that the bed got temporarily into an intertidal or even a supratidal environment, for instance by eustatic fluctuations of sea level. Changes in sea level have been invoked to account for parallel observations in a similar sediment (FÜRSICH 1971). Such an assumption is unnecessary, because stromatolites and iron oxide crusts now being formed have been discovered in water more than 100 m deep off Belize by JAMES & GINSBURG (1979). Fossil deep-water stromatolites have been described by PLAYFORD & COCKBAIN (1969) from a beautifully exposed barrier reef in the Devonian of Australia (see PLAYFORD 1980). Borings by microorganisms can no longer be regarded as evidence for shallow water, since SCOFFIN et al. (1980) have observed that the abundance of borings produced by fungi on Rockall Bank actually increases in depths below 150 m where agitation is infrequent. Boring microorganisms in the Schellenbrücke Bed were probably fungi (see below). Taphonomical evidence from the Schellenbrücke Bed indicates that the water depth was greater than 70 m, according to observations in the Recent given by KUENEN (1964, p. 228).

Some information about depth of deposition may be deduced from the minerals chamosite and glauconite. Glauconite is rare in the Schellenbrücke Bed to the point where it can even be absent from some thin sections. In the time-equivalent Glaukonitsandmergel near Schaffhausen there is as much as 30% glauconite which has been identified by several methods (GYGI & MCDOWELL 1970). Chamosite could not be found in that horizon, but varying proportions of the iron ooids in the Schellenbrücke Bed are chamositic. In a borehole, the ferruginous crust capping the Schellenbrücke Bed has been found to consist of chamosite (GYGI 1969, p. 60). Part of the chamosite appears to have been oxidized a short time after its formation, as indicated by chamositic ooids preserved in different stages of pervasive oxidation, in a single thin section. In Recent tropical seas, chamosite-like minerals are being formed in depths ranging from 10 to 150 m (PORRENGA 1967). According to this author, substantial glauconite formation takes place at temperatures lower than 15 °C. This mineral is formed in Indonesia only in water deeper than 55 m (KUENEN, in CLOUD 1955, p. 488). Glauconite was found to be forming off the Niger Delta from a depth of 60 or 70 m, but it becomes abundant only below 125 m (PORRENGA 1967). The paleolatitude of northern Switzerland was somewhat more than 30° N during the Oxfordian (FIRSTBROOK et al. 1979). The tropical zone was wider at that time than now, as indicated by hermatypic corals in the Middle Oxfordian of Greenland (BEAUVAIS 1977) at a paleolatitude of more than 50° N. The Oxfordian sediments of Switzerland were thus deposited in a tropical sea, and the depth of formation intervals for Recent chamosite and glauconite in the tropics may be used as general guidelines. The depth limit below which glauconite formation was prolific, may have been, owing to the relatively high paleolatitude, at a somewhat higher level in the Swiss Oxfordian than it is in Recent equatorial waters, assuming that the temperature gradient with depth was the same in Oxfordian tropical seas as in Recent ones. It is likely that the glauconite-rich Glaukonitsandmergel near Schaffhausen was deposited in deeper water than the Schellenbrücke Bed was, probably below 100 m.

There are marked differences in the ammonite assemblages of the two horizons. 17% of the ammonites in the Glaukonitsandmergel are Cardioceratidae, as are 42% in the Schellenbrücke Bed (Table). DEBRAND-PASSARD & MARCHAND (1979) have found a deposit from the Scarburgense Subzone in northeastern France where the percentage of Cardioceratidae in ammonites is as high as 90. There, the ammonite fauna is associated with abundant bivalves. It is therefore possible that Cardioceratidae preferred relatively shallow water. The Oppeliidae have an opposite tendency. Only 9% of the ammonites in the Schellenbrücke Bed are Oppeliidae, as compared to 32% in the Glaukonitsandmergel (Table). ZIEGLER (1967) defined depth zones for faunal assemblages which are based on reliable field evidence down to a depth of about 200 m. According to this author, the ammonite assemblage of the Schellenbrücke Bed indicates a maximum depth of about 100 m. The much greater abundance of Oppeliidae in the Glaukonitsandmergel, as interpreted according to ZIEGLER (1967), is another, faunal, indication that this horizon was deposited in deeper water than the Schellenbrücke Bed. JAMES & GINSBURG (1979) found iron oxide crusts at depths between about 40 and 150 m. Most of the ammonites in the Glaukonitsandmergel are encrusted with limonite. If it were permissible to apply the observation of JAMES & GINSBURG to the Jurassic, this would mean that the lower limit of the depth of deposition of the Glaukonitsandmergel would be close to 150 m. ZIEGLER (1967) concluded that at about this depth Oppeliidae become more abundant than Perisphinctidae. Oppeliidae are somewhat less common than Perisphinctidae in the Glaukonitsandmergel, which may be interpreted as meaning that the horizon was deposited in water less than 150 m deep.

3. Interpretation and discussion

The depositional history and the facies relationships of the Lower Oxfordian deposits in northern Switzerland can be reconstructed with ammonites at the subzone level. The nodular bed and the Schellenbrücke Bed near Herznach are condensed, relatively deep-water facies at the distal fringe of a thick, argillaceous succession, which filled in, from the northwest, a basin of more or less uniform depth. Condensation appears to be a typical property of many iron oolites (HALLAM 1966, TALBOT 1974). In the Schellenbrücke Bed, condensation has led to an unusual concentration of fossils (MARCHAND & GYGI 1977). KIMBERLEY (1979, p. 120) noted that Phanerozoic iron oolites are "generally highly fossiliferous". On the contrary, Recent aragonitic oolite shoals in the state of formation support only a thin population mainly of echinoids, asteroids, and some minute gastropods. Whole macrofossils in lithified calcareous oolites are rare. KIMBERLEY has never used fossils in order to interpret a paleoenvironment in detail. In addition, KIMBERLEY seems to be unaware of the ample literature on diagenetic textures produced in aragonitic oolites by fresh water in the phreatic zone where he supposes ferruginization to take place. At least some of such textures (see LONGMAN 1980) should be recognizable in iron oolites if these were formed according to KIMBERLEY's model. All Lower

Oxfordian deposits of northern Switzerland are marine, and they are covered by marine sediments.

Sedimentological evidence for the depth of deposition of the Schellenbrücke Bed indicates only a broad outline. The horizon has been formed below the normal wave base, but it has been within reach of storm-generated waves. Chamosite and glauconite provide more detailed information, although many problems of the mode and the environment of formation of these minerals are still unresolved. The virtual absence of glauconite from the Schellenbrücke Bed, when paleolatitude data are taken into account, indicates a water depth of less than 100 m. The taphonomy of ammonites suggests that the water was more than 70 m deep. The depth interval indicated by iron oxide crusts of the kind described by JAMES & GINSBURG (1979) needs confirmation in other warm seas before such crusts can be used with confidence as depth indicators. Deep-water stromatolites have as yet little potential for detailed environmental interpretation. Borings by microorganisms seem to be typical of iron ooids. ELLIS has illustrated and named some from the Frodingham Ironstone as early as 1917, and described them as fungi. Very similar borings in the Schellenbrücke Bed have been interpreted as fungi by GYGI (1969, p.45) as well. Boring fungi in thin section cannot readily be used to estimate water depth (compare BUDD & PERKINS 1980, with SCOFFIN et al. 1980), and they are difficult to distinguish from boring algae. There is no indication to the environmental conditions which caused hard surfaces either to be eroded by boring organisms, or to be covered by ferruginous crusts. The most detailed and reliable paleoenvironmental information can be deduced from the invertebrate fauna. Ammonites are the most useful indicator for deeper water assemblages. The depth zones of ZIEGLER (1967), although defined in Kimmeridgian sediments, can be easily recognized in the Oxfordian of northern Switzerland.

This is not surprising, since NATLAND (1933) for instance found good correlation between depth zones of Recent and Pliocene foraminifera in southern California. The faunal assemblage indicates that the Schellenbrücke Bed has been deposited in water about 80–100 m deep. Three other independent lines of evidence are in accord with this. Therefore it may be concluded that the order of magnitude of the estimate is correct.

The iron ooids in the Schellenbrücke Bed are certainly not lateritic spherules washed into the sea. Neither can they have been formed in nearshore waters and then been transported to a greater depth. Otherwise, iron ooids would occur in the part of the Terrain à Chailles which is the proximal time equivalent of the Schellenbrücke Bed. The ferruginous crust on the surface of the hardground on top of the Schellenbrücke Bed is proof that iron has been concentrated in situ. The sequence ferruginous lime mud/iron ooids/ferruginous crust may reflect different rates of sedimentation in descending order of rate. HESSLAND (1949) was of the opinion that boring and enveloping microorganisms, which he thought to be algae, could form iron ooids by precipitating iron. HESSLAND, however, did not prove that the brown substance in the penetrating filaments was iron oxide. Iron-secreting algae have not as yet been reported from the Recent. It is more likely that the brown substance in the filaments, which probably *is* iron oxide, was precipitated in the empty tubes after the death of the boring organisms (see JAMES & GINSBURG 1979), as micritic carbonate is precipitated in empty boreholes made by algae in Recent aragonitic ooids (see BATHURST 1975). It is evident from GYGI (1969, Pl.2, Fig.5) that perforation of the nucleus of an iron ooid and formation of the cortex are successive, unrelated processes. ALDINGER (1957*a*) and GEBERT (1964) proposed that interstitial water rising from compacting argillaceous sediment could provide iron for iron ooid formation near the sediment/water interface. KIMBERLEY (1979, p. 124) rejected this as being quantitatively insufficient. His argument is based on the iron content of drinkable, subsoil groundwater. This can certainly not be compared with interstitial water of marine sediments. KIMBERLEY's objection is thus untenable. Observations by MCGEARY & DAMUTH (1973) confirmed that precipitation of iron from interstitial water near the surface of an argillaceous sediment can take place, but in the case of the Schellenbrücke Bed this is unlikely. In most of the area where the bed is present there is no argillaceous sediment underneath which might have undergone compaction at the time the Schellenbrücke Bed was being formed. The horizon above the Schellenbrücke Bed is a condensed deposit which is less than 10 cm thick in most sections, even though it encompasses both the Densiplicatum Zone and the Antecedens Subzone of the Transversarium Zone (GYGI & MARCHAND 1981). Condensation and glauconite in this thin horizon are evidence that the sea became deeper during the Densiplicatum and the early Transversarium chron. The horizon represents the initial, "transgressive" part of the next upward shoaling cycle of more than 200 m of fully marine sediments (GYGI 1969). Ferruginization of the Schellenbrücke Bed in terms of KIMBERLEY's model cannot seriously be considered under these circumstances. Neither can ferruginization of the Callovian iron oolitic orebody of Herznach with its almost pure ammonite fauna, because the orebody is overlain by iron oolitic horizons with similar ammonite faunas.

4. Model of marine iron ooid formation

Formation of iron ooids on a relatively deep, open marine shelf, as testified mainly by an abundant and diverse fauna, may have taken place in the following sequence: Ferric iron oxide derived from weathering on land, in a tropical climate with adequate rainfall, coated the surfaces mainly of clay minerals (chiefly kaolinite), but it could also coat other mineral grains from clay to sand size (CARROLL 1958, p.2, cf. ALDINGER 1957*a*, p.8). Rainfall supporting vegetation on neighbouring land is indicated by abundant pieces of fossil wood in the Lower Oxfordian beds 6 and 7*a* of an excavation made near Herznach (GYGI 1977, Pl. 11, section 2). Rivers washed the iron-coated grains into the sea. In early Oxfordian time, the minerals were deposited in northwestern Switzerland in a mainly argillaceous succession about 50 or 60 m thick. Illite is the dominant clay mineral (PFRUNDER & WICKERT 1970). Illite can carry iron, but less than kaolinite. The capacity of the iron transport mechanism as proposed by CARROLL (1958) is thereby restricted, but the dominance of illite and the alleged absence of kaolinite (PFRUNDER & WICKERT 1970) is not an impeding factor, as HALLAM (1966, p. 1309) apparently thought. The observations made by LEMOALLE & DUPONT (1973, p. 175) in Lake Chad are qualitative evidence for this process, and the authors proved by measurements (p. 176) that the iron transportation mechanism as proposed by CARROLL (1958) is effective. After deposition, the *Eh* value of the interstitial water became negative as a result of the decomposition of organic matter (BERNER 1971, p. 119). Ferric iron on the surface of mineral grains was thereby reduced to the ferrous state. This was dissolved in the interstitial water, while compaction of the fresh deposit drove the solution to the sediment surface (ALDINGER 1957*a*, p. 8). The deeper part of the deposit was thus depleted in iron, as has been measured by HINZE & MEISCHNER (1968) in a Recent sediment of the Adriatic Sea. At the arrival at the aerated zone directly below the sediment-water interface, iron was reoxidized and precipitated as amorphous or extremely fine-grained $Fe(OH)_3$. The iron-enriched zone migrated up with the sediment surface in the course of deposition because of smothering, which constantly brought the zone back into a reducing environment. There the iron was again reduced, dissolved and then precipitated near the subsequent sediment surface. No ferruginous horizon has been found to date in the Renggeriton or Terrain à Chailles of northwestern Switzerland except at the upper limit of the Terrain à Chailles in a section near Soulce, canton Jura (GYGI, unpublished data). Therefore, it must have been normal that iron was removed in the course of sedimentation. Unlithified mud of the Schellenbrücke Bed, which was deposited in water possibly as deep as 100 m, could be put in motion by currents induced by waves less than 2 m high, according to the diagram in Figure 5 by KOMAR & MILLER (1975). Since the water over the argillaceous deposits in the northwest was less deep, it may be assumed that storms resuspended the superficial, iron-enriched part of the sediment in the northwest more frequently than that of the Schellenbrücke Bed. If this happened not uncommonly and on a depositional slope, the particles in suspension would tend to be resedimented downslope. The iron hydroxide being the material with the smallest grain size would be the first to be whirled up, and the last to settle during resedimentation. Therefore it would travel the greatest distance. Here is a mechanism of elutriation which might have been the cause of the lateral facies relationship of thick, argillaceous sediment depleted in iron with iron-enriched, condensed deposit at the distal fringe of an argillaceous succession.

Several authors concluded that iron ooids were formed when being rolled by gentle currents at the surface of muddy sediment. The view that mud could be deposited at the same place where iron ooids are accreted by currents seems to be contradictory at first glance. This problem can be solved by assuming an environment where water is normally quiet, and currents occur only occasionally. Times of calm water are a prerequisite for mud deposition. For example, the short quiescence between tidal currents is sufficient. Deposition of mud therefore may occur at any depth from the intertidal zone down. Net mud accumulation is even possible as a matrix between oncoids, with a diameter of up to several centimeters, which are regularly rolled by strong tidal currents (GYGI 1969, p. 112). Net deposition or winnowing of mud in shallow water apparently depends on two factors. One is the rate of mud supply (MCCAVE 1971, p. 93). The second depends on whether or not an unidirectional component is superimposed on oscillant currents (SWIFT 1979). Mud in itself gives no information about depth, or intensity of water movement at the site of deposition. The work of PORRENGA (1967) and KNOX (1970, 1971) indicates that there is a limit of minimum water depth for the formation of chamosite ooids. On an open shelf, episodes of gentle currents recurring at irregular intervals can only be expected below the normal wave base as an unusual condition during rough weather. The maximum depth limit is where episodic currents are never strong enough to move particles the size of iron ooids. It may be concluded from examples given by KUENEN (1964) that this may well be as deep as 100 m. Only ferric iron can be transported in normal seawater. Undeformed goethite or limonite ooids are therefore probably mostly of primary origin. Chamosite ooids are commonly deformed during compaction or by burrowing animals. The internal structures illustrated by KNOX (1970) are evidence that the ooids are formed at the sediment surface, as envisaged by CURTIS & SPEARS (1968). Reduction of ferric iron and crystallization of chamosite must take place after an ooid is buried (see TALBOT 1974) by mud resuspended during ooid accretion. The ooid is thereby placed into the reducing environment beneath the water-sediment interface. Presence of organic matter is a prerequisite for establishment of reducing conditions in interstitial water. A chamosite ooid can be further accreted when a subsequent episodic current has removed the sediment cover. If then the ooid is once more smothered by settling mud after water movement has ceased, the new cortex will again become chamosite. If not, or if the covering sediment sheet is too thin, the iron in the fresh cortex will remain in the ferric state, and the chamositic ooid becomes enveloped by a crust of limonite or goethite, which might give the chamosite in the interior some protection from being oxidized. Repetition of this would lead to the formation of the well-known, but hitherto enigmatic iron ooids with alternating chamositic and goethitic or limonitic crusts (see FINKENWIRTH 1964, p.80, THIENHAUS 1969, Pl.2, Fig.1, GYGI 1969, Pl.3, Fig.8, and FREITAG 1970, Photogr.4, and JAMES & VAN HOUTEN 1979, p.129). Such a mode of ooid accretion seems to be possible only if the net mud sedimentation rate is low, as HALLAM (1966), TALBOT (1974), and JAMES & VAN HOUTEN (1979) have concluded.

5. Discussion of the model

This model gives an explanation for the current-induced accretion of iron ooids on the surface of synsedimentary mud, and for the formation of chamosite ooids in a well-aerated environment supporting abundant and diverse life. The model is compatible with the observation that brown (goethitic or limonitic), undeformed iron ooids can coexist with commonly deformed chamosite ooids, or ooids with alternating crusts of chamosite and goethite/limonite, and chamosite ooids in various stages of oxidation. Close coexistence of different ooid types very probably reflects formation in the same environment (cf. FINKENWIRTH 1964, p.82). The faunas indicate that marine iron ooids may be formed in a relatively broad depth interval from shallow water (TALBOT 1974, p.441) down to about 100 m. It is not necessary to assume a special submarine topography with a clastic trap to account for the observed facies relationships, as HALLAM (1966) has done. The model is in agreement with the fact that oolitic ironstones are typically condensed. Unusually strong storms may be invoked as a cause for the winnowing of mud leading from mudstone to packstone concentration of iron ooids, as is found for instance in the upper half of the Callovian orebody of Herznach. The "transgressive" character of some iron oolitic horizons can be understood in terms of this model.

The Hjulström effect is critical to the model. Unexpectedly high current velocities may be necessary for the resuspension of cohesive mud. However, it is established that mud *has* been resuspended in the Schellenbrücke Bed. Observations are rapidly accumulating that storm-induced mud resuspension and transport can be a large-scale geologic process (SWIFT 1979, CACCHIONE & DRAKE 1979). YOUNG & SOUTHARD (1978) have made in-situ measurements on the bottom of Buzzard Bay, Massachusetts. Another crucial point in the model is the position of the zero *Eh*-level within the sediment. In a mud under normally aerated seawater, the *Eh* of the interstitial water may become negative at a very short distance down from the sediment surface (ZOBELL 1946, Table 6, GINSBURG 1957, p. 88). Precise data are not available due to the inaccuracy of measurements in the field (see FRIEDMAN & SANDERS 1978, p. 137). Personal observations in the intertidal and shallow subtidal zones of the Jadebusen Bay of the southern North Sea and in Bermuda, indicate that the zero *Eh* surface is close to or within millimeters from the sediment surface in mud-grade deposits. A negative *Eh* in a sediment may be assumed from grey hues of varying intensity, and from the odor of hydrogen sulfide. This substance is stable only in the absence of free oxygen. A problem particular to the Schellenbrücke Bed is the rapid lithification of the calcareous mud, as indicated by the preservation of delicate spines in cardioceratid ammonites. The preferential occurrence of marine iron ooids in condensed deposits suggests that iron ooid accretion is a slow process. Did lithification of mud in the Schellenbrücke Bed proceed slow enough to allow for iron ooid formation? We have seen that the matrix of this horizon became lithified at an uneven rate. Some portions must have remained unconsolidated long enough to make iron ooid accretion possible.

Unsolved problems are the selective mode of accretion of iron ooids on the surface of a sediment which, in the case of the Schellenbrücke Bed, has a very different composition than the ooids. We have as yet no satisfactory explanation for the apparently syngenetic impregnation of calcareous ooids with iron hydroxide, as figured by FREITAG (1970, Photogr. 2). The generally good sorting of iron ooids which have been formed in relatively deep water, is not understood. There is no indication of what created the reducing conditions necessary for the formation of the chamositic part of the crust on top of the hardground of the Schellenbrücke Bed, but there is evidence that the crust has not been formed under anoxic water (see above). Another problem which has to be specially discussed in relation to the Schellenbrücke Bed is ooid reworking. The horizon is so thin that reworking of iron ooids from older sediments cannot be ruled out. But this is unlikely, since iron ooids are always present, even where the horizon overlies sediments without iron ooids. Transportation might have been the cause for this, but if currents had displaced ooids over distances of many kilometers they would have winnowed the fine-grained matrix completely in the process. This has never been found. Moreover, the chamositic ooids, which are initially soft and may remain soft until after the muddy matrix is lithified (GYGI 1969, p. 46), would have been destroyed by abrasion in the course of prolonged transportation (cf. ALDINGER 1957*a*, p. 7). It is not known how bioclasts of any size could be overturned in water over 80 m deep. A possible explanation is given by GYGI (1969, p. 105) for another condensed horizon in the Late Jurassic of northern Switzerland. In the Recent, rays and hogfish have been observed to overturn rocks in search of food.

The applicability of this model to other iron oolitic formations cannot be tested in detail, because the lateral facies relationships of most oolitic ironstones are not well-known. The following iron oolitic rocks or ores appear to compare well with the Schellenbrücke Bed by their very wide areal extent and fauna: "Murchisonae Beds" (Upper Aalenian, ALDINGER 1957*b*, Fig. 3), "Humphriesi Bed" (Middle Bajocian, ALDINGER & FRANK 1944, Fig. 4), "Macrocephalus Beds" and "Anceps-Athleta Bed" (Callovian) in northern Switzerland and southern Germany. The Early Jurassic minette ores of Lorraine, France (BUBENICEK 1964), southern Germany (ALDINGER & FRANK 1944, Fig. 2), and northern Germany (HOFFMANN 1969, Fig. 33) have probably been formed in a shallow-water, near-shore environment. Formation of the Late Jurassic iron oolitic ores of northern Germany (SIMON 1969) and southern England must have occurred in a similar, possibly even lagoonal setting (TALBOT 1974, Fig. 5-6). There is no evidence that the iron ooids of these shallow-water formations originated in a mode fundamentally different from the one proposed in the above model. It is evident from the work of HALLIMOND (1925, p. 61) that the Northampton Ironstone is condensed, and that the uppermost bed of the Cleveland Dogger has a wide areal extent (p. 59). The orebodies are associated or even interbedded with argillaceous mud (HALLIMOND 1925, p. 42, TALBOT 1974, Fig. 3). They have accumulated in depressions rather than on submarine highs (SIMON 1969, Fig. 124, cf. HALLAM 1967, Fig. 5). It is possible that iron ooid accretion in a lagoon, or in a closed brackish basin like Lake Chad, can proceed at a high rate (LEMOALLE & DUPONT 1973, p. 176), and without an intervening concentration by compactional eluviation and subsequent elutriation.

6. Conclusions

Iron has been concentrated in the Schellenbrücke Bed either in the fine-grained matrix, in iron ooids, in crusts on litho- and bioclasts, or on the hardground at the upper surface of the bed. The environment of deposition was an open marine shelf with very little sedimentation, supporting an abundant and diverse fauna in water as deep as 100 m. The concentration of iron took place in two stages. Iron was eluviated in the course of compaction from a thick succession of argillaceous sediments in a more proximal position. Iron hydroxide was first concentrated near the slightly sloping sediment-water interface of this deposit (less than one-half degree, GYGI, unpublished data). Repeated whirling up and downslope resedimentation of the uppermost deposit led to further concentration of iron hydroxide by elutriation near the distal fringe of the argillaceous sediment fan. This was the place where ooids with ferric iron were accreted, normally when being rolled by episodic currents. When an ooid was covered by an adequate sheet of settling mud immediately after accretion, the fresh crust was diagenetically transformed into chamosite, provided that the crust had the right chemical composition (see CURTIS & SPEARS 1968), and that the mud contained enough organic matter. Ooids remaining at the surface or being covered by too thin a sediment sheet, or by sediment with too little organic matter, developed crusts of goethite or limonite.

1. There can be no doubt that some Jurassic oolitic iron formations of central and northern Europe originated in shallow to deeper marine waters.

2. Lateritic spherules apparently can be concentrated in shallow marine sediments. Such spherules might be difficult to distinguish from iron ooids of marine origin if the spherules were well-sorted and had distinctly concentric crusts.

3. Careful observations suggest that iron ooids are now being accreted in the shallow and brackish waters of Lake Chad. This is an indication that iron ooid accretion is possible in waters with varying salinities.

4. There is no reason to reject ferruginization of aragonitic oolite by fresh water on land as a possible mechanism for oolitic ironstone formation. But at present it is impossible to judge the validity of KIMBERLEY'S views unless convincing sedimentological and faunal evidence supporting the model is produced. The main conclusion of this paper is that known oolitic ironstones probably have been formed in more than a single mode.

Acknowledgments

The writer is grateful to H.-P. Funk, Zürich, for his valuable comments on the manuscript and helpful suggestions. This paper is based on research supported by the Swiss National Science Foundation grants No. 2.211.69, and No. 2.165-0.78.

REFERENCES

ALDINGER, H. (1957*a*): *Zur Entstehung der Eisenoolithe im Schwäbischen Jura.* - Z. dtsch. geol. Ges. *109*/1, 7-9.

— (1957*b*): *Eisenoolithbildung und rhythmische Schichtung im süddeutschen Jura.* - Geol. Jb. *74*, 87-96.

ALDINGER, H., & FRANK, M. (1944): *Vorkommen und Entstehung der südwestdeutschen jurassischen Eisenerze.* - N. Jb. Mineral. Geol. Paläont. [Abh. B] *88*, 293-336.

BATHURST, R.G.C. (1975): *Carbonate sediments and their diagenesis.* - Dev. Sedimentol. *12* (2nd ed.).

BEAUVAIS, L. (1977): *Une espèce nouvelle de madréporaire dans le Jurassique supérieur du Groenland et de l'Ecosse. Implications paléobiogéographiques.* - Géobios *10*/1, 135-141.

BERNER, R.A. (1971): *Principles of chemical sedimentology.* - McGraw-Hill, New York.

BRADSHAW, M.J., JAMES, S.J., & TURNER, P. (1980): *Origin of oolitic ironstones - discussion.* - J. sediment. Petrol. *50*/1, 295-304.

BRAKONIECKI, T.F. (1980): *Lolliguncula tydeus, a new species of squid (Cephalopoda; Myopsida) from the Pacific coast of Central America.* - Bull. marine Sci. *30*/2, 424-430.

BUBENICEK, L. (1964): *Etude sédimentologique du minerai de fer oolithique de Lorraine.* - Dev. Sedimentol. *2*, 113-122.

BUDD, D.A., & PERKINS, R.D. (1980): *Bathymetric zonation and paleoecological significance of microborings in Puerto Rican shelf and slope sediments.* - J. sediment. Petrol. *50*/3, 881-904.

CACCHIONE, D.A., & DRAKE, D.E. (1979): *Bottom shear stress generated by waves and currents in the Northern Bering Sea.* In: NIO, S.D.: *International meeting on Holocene marine sedimentation in the North Sea basin.* Abstracts (p. 71). - Int. Assoc. Sedimentol., Texel.

CARROLL, D. (1958): *Role of clay minerals in the transportation of iron.* - Geochim. cosmochim. Acta *14*, 1-28.

CLOUD, P.E. (1955): *Physical limits of glauconite formation.* - Bull. amer. Assoc. Petroleum Geol. *39*, 484-492.

CURTIS, C.D., & SPEARS, D.A. (1968): *The formation of sedimentary iron minerals.* - Econ. Geol. *63*, 257-270.

DEBRAND-PASSARD, S., DELANCE, J.-H., LORENZ, J., & MARCHAND, D. (1978): *Le Callovien supérieur et l'Oxfordien inférieur dans les départements du Cher et de la Nièvre. Précisions stratigraphiques, paléogéographiques et paléobiogéographiques.* - Bull. Bur. Rech. géol. min. (2) [Sect. I], *4*, 317-327.

DEBRAND-PASSARD, S., & MARCHAND, D. (1979): *Réflexions sur la répartition des Ammonoïdés dans l'est et le sud du Bassin parisien au Callovien supérieur (zone à Lamberti) et à l'Oxfordien inférieur (zone à Mariae).* - Bull. Bur. Rech. géol. min. (2) [Sect. IV], *2*, 103-112.

Ellis, D. (1917): *On the Jurassic fossil fungus, Phycomycites Frodinghamii (Ellis).* - Geol. Mag. *4*, 102-108.

Finkenwirth, A. (1964): *Das Eisenerz des Lias γ am Kahlberg bei Echte und der Weissjura in Süd-Hannover.* - Beih. geol. Jb. *56.*

Firstbrook, P.L., Funnell, B.M., Hurley, A.M., & Smith, A.G. (1979): *Paleoceanic reconstructions 160-0 Ma.* - Sripps Institution of Oceanography, La Jolla.

Freitag, K.-P. (1970): *Feinstratigraphische und petrographische Untersuchungen im erzführenden Korallenoolith (unter Malm) des westlichen Wesergebirges (Nordwestdeutschland).* - Clausth. H. Lagerstkd. Geochem. miner. Rohst. *9*, 185-214.

Friedman, G.M., & Sanders, J.E. (1978): *Principles of sedimentology.* - Wiley, New York.

Fürsich, F.T. (1971): *Hartgründe und Kondensation im Dogger von Calvados.* - N. Jb. Geol. Paläont. Abh. *138*/3, 313-342.

Gatrall, M., Jenkyns, H.C., & Parsons, C.F. (1972): *Limonitic concretions from the European Jurassic, with particular reference to the "snuff-boxes" of southern England.* - Sedimentology *18*, 79-103.

Gebert, H. (1964): *Mineralogische Untersuchungen zur Genese oolithischer Eisenerze des Dogger epsilon in Baden-Württemberg.* - Arb. geol.-paläont. Inst. tech. Hochsch. Stuttgart [N.F.] *41*, 1-77.

Ginsburg, R.N. (1957): *Early diagenesis and lithification of shallow-water carbonate sediments in South Florida.* In: Le Blanc, R.J., & Breeding, J.G. (Ed.): *Regional aspects of carbonate deposition.* - Spec. Publ. Soc. econ. Paleont. Mineral. *5*, 80-99.

Gygi, R.A. (1969): *Zur Stratigraphie der Oxford-Stufe (oberes Jura-System) der Nordschweiz und des süddeutschen Grenzgebietes.* - Beitr. geol. Karte Schweiz [N.F.] *136.*

— (1977): *Revision der Ammonitengattung Gregoryceras (Aspidoceratidae) aus dem Oxfordian (Oberer Jura) der Nordschweiz und von Süddeutschland. Taxonomie, Phylogenie, Stratigraphie.* - Eclogae geol. Helv. *70*/2, 435-542.

Gygi, R.A., & Marchand, D. (1981): *Les Cardioceratinae (Ammonoidea) du Callovien terminal et de l'Oxfordien inférieur et moyen (Jurassique) de la Suisse septentrionale: taxinomie préliminaire.* - Géobios (in prep.).

Gygi, R.A., & McDowell, F.W. (1970): *Potassium-argon ages of glauconites from a biochronologically dated Upper Jurassic sequence of northern Switzerland.* - Eclogae geol. Helv. *63*/1, 111-118.

Hallam, A. (1964): *Origin of the limestone-shale rhythm in the Blue Lias of England: a composite theory.* - J. Geol. *72*/2, 157-169.

— (1966): *Depositional environment of British Liassic ironstones considered in the context of their facies relationships.* - Nature (London) *209*/5030, 1306-1307.

— (1967): *Sedimentology and palaeogeographic significance of certain red limestones and associated beds in the Lias of the Alpine region.* - Scott. J. Geol. *3*/2, 195-220.

Hallam, A., & Bradshaw, M.J. (1979): *Bituminous shales and oolitic ironstones as indicators of transgressions and regressions.* - J. geol. Soc. (London) *136*, 157-164.

Hallimond, A.F. (1925): *Iron ores: Bedded ores of England and Wales. Petrography and chemistry.* - Mem. geol. Surv. G.B., spec. rep. miner. Resour. G.B. *29*, 1-139.

Hessland, I. (1949): *4. Investigations of the Lower Ordovician of the Siljan District, Sweden. II: Lower Ordovician penetrative and enveloping algae from the Siljan District.* - Bull. geol. Instn. Univ. Uppsala *33*, 409-428.

Hinte, J.E. van (1976): *A Jurassic time scale.* - Bull. amer. Assoc. Petroleum Geol. *60*/4, 489-497.

Hinze, C., & Meischner, D. (1968): *Gibt es rezente Rot-Sedimente in der Adria?* - Marine Geol. *6*/1, 53-71.

Hoffmann, K. (1969): *Paläogeographie der nordwestdeutschen Lias-Eisenerze.* - Beih. geol. Jb. *79*, 104-110.

James, H.E., & van Houten, F.B. (1979): *Miocene goethitic and chamositic oolites, northeastern Colombia.* - Sedimentology *26*/1, 125-133.

James, N.P., & Ginsburg, R.N. (1979): *The seaward margin of Belize barrier and atoll reefs.* - Spec. Publ. int. Assoc. Sedimentol. *3.*

Jeannet, A. (1951): *Stratigraphie und Palaeontologie des oolithischen Eisenerzlagers von Herznach und seiner Umgebung (1. Teil).* - Beitr. Geol. Schweiz, geotech. Ser. *13*/5.

Kimberley, M.M. (1979): *Origin of oolitic iron formations.* - J. sediment. Petrol. *49*/1, 111-132.

— (1980*a*): *Origin of oolitic iron formations - reply.* - J. sediment. Petrol. *50*/1, 299-302.

— (1980*b*): *Origin of oolitic iron formations - reply.* - J. sediment. Petrol. *50*/3, 1003-1004.

KNOX, R.W. O'B. (1970): *Chamosite ooliths from the Winter Gill Ironstone (Jurassic) of Yorkshire, England.* - J. sediment. Petrol. *40*/4, 1216-1225.

— (1971): *An alternative to the "Clastic Trap" interpretation of oolitic ironstone facies.* - Geol. Mag. *108*, 544-545.

KOBY, F. (1899): *Notice stratigraphique sur l'Oxfordien dans la partie septentrionale du Jura bernois.* - Mém. Soc. paléont. suisse *26*, 189-220.

KOMAR, P.D., & MILLER, M.C. (1975): *On the comparison between the threshold of sediment motion under waves and uni-directional currents with a discussion of the practical evaluation of the threshold.* - J. sediment. Petrol. *54*/1, 362-367.

KUENEN, P.H. (1964): *Marine Geology* (5th printing). - Wiley, New York.

LEMOALLE, J., & DUPONT, B. (1973): *Iron-bearing oolites and the present conditions of iron sedimentation in Lake Chad (Africa).* In: AMSTUTZ, G.C., & BERNARD, A.J. (Ed.): *Ores in sediments.* - IUGS [A] *3*, 167-178.

LONGMAN, M.W. (1980): *Carbonate diagenetic textures from nearsurface diagenetic environments.* - Bull. amer. Assoc. Petroleum Geol. *64*/4, 461-487.

LUCAS, G. (1966): *Fonds durcis, lacunes sous-marines, séries condensées et ondes marines séismiques ou "Tsunami".* - C.R. Acad. Sci. Paris *262*, 2141-2144.

MARCHAND, D., & GYGI, R.A. (1977): *L'Oxfordien inférieur d'Herznach (canton d'Argovie, Suisse). Précisions paléontologiques et stratigraphiques.* - C.R. Acad. Sci. (Paris) (D), *285*, 853-856.

MCCAVE, I.N. (1971): *Wave effectiveness at the sea bed and its relationship to bed-forms and deposition of mud.* - J. sediment. Petrol. *41*/1, 89-96.

MCGEARY, D.F.R., & DAMUTH, J.E. (1973): *Postglacial iron-rich crusts in hemipelagic deep-sea sediment.* - Bull. geol. Soc. Amer. *84*/4, 1201-1211.

NATLAND, M.L. (1933): *Depth and temperature distribution of some Recent and fossil Foraminifera in the southern California region.* - Bull. Scripps Instn. Oceanogr., tech. Ser. *3*, 225-230.

OPPEL, A. (1856-8): *Die Juraformation Englands, Frankreichs und des südwestlichen Deutschlands.* - Jh. württemb. natw. Ver. *12-14*, 1-857.

PFRUNDER, V.R., & WICKERT, H. (1970): *Einige Versuche über den Einfluss der chemischen Zusammensetzung und der Mahlung auf die Sinterung von Zement-Rohmehlen.* - Zement - Kalk - Gips *23*/4, 147-152.

PLAYFORD, P.E. (1980): *Devonian "Great Barrier Reef" of Canning Basin, Western Australia.* - Bull. amer. Assoc. Petroleum Geol. *64*/6, 814-840.

PLAYFORD, P.E., & COCKBAIN, A.E. (1969): *Algal stromatolites: Deepwater forms in the Devonian of Western Australia.* - Science *165*/3897, 1008-1010.

PORRENGA, D.H. (1967): *Glauconite and chamosite as depth indicators in the marine environment.* - Marine Geol. *5*, 495-501.

ROLLIER, L. (1888): *Etude stratigraphique sur le Jura bernois. Les faciès du Malm jurassien.* - Arch. Sci. phys. nat. (Genève) *19*, 1-85; Eclogae geol. Helv. *1*, 3-88.

SCHELLMANN, W. (1969): *Die Bildungsbedingungen sedimentärer Chamosit- und Hämatit-Eisenerze am Beispiel der Lagerstätte Echte.* - N. Jb. Mineral. [Abh.] *111*/1, 1-31.

SCOFFIN, T.P., ALEXANDERSSON, E.T., BOWES, G.E., CLOKIE, J.J., FARROW, G.E., & MILLIMAN, J.D. (1980): *Recent, temperate, sub-photic, carbonate sedimentation: Rockall Bank, Northeast Atlantic.* - J. sediment. Petrol. *50*/2, 331-356.

SIEHL, A., & THEIN, J. (1978): *Geochemische Trends in der Minette (Jura, Luxemburg/Lothringen).* - Geol. Rdsch. *67*/3, 1052-1077.

SIMON, P. (1969): *Paläogeographie der nordwestdeutschen Malm-Eisenerze.* - Beih. geol. Jb. *79*, 339-345.

SOKOLOVA, E.I. (1964): *Physicochemical investigations of sedimentary iron and manganese ores and associated rocks.* - Israel Program Sci. Transl. (Jerusalem).

SWIFT, D.J.P. (1979): *Inner shelf sedimentation on wind and tide dominated shelves: The North American Atlantic shelf versus the North Sea.* In: NIO, S.D.: *International meeting on Holocene marine sedimentation in the North Sea basin.* Abstracts (p. 39). - Int. Assoc. Sedimentol., Texel.

SYKES, R.M., & CALLOMON, J.H. (1979): *The Amoeboceras zonation of the boreal Upper Oxfordian.* - Palaeontology *22*/4, 839-903.

SYKES, R.M., & SURLYK, F. (1976): *A revised ammonite zonation of the boreal Oxfordian and its application in northeast Greenland.* - Lethaia *9*, 421-436.

TALBOT, M.R. (1974): *Ironstones in the Upper Oxfordian of southern England.* - Sedimentology *21*, 433-450.

THIENHAUS, R. (1969): *Das Eisenerzlager des Lias γ von Bislich am Niederrhein.* - Beih. geol. Jb. *79*, 93-103.

THIERRY, J. (1978): *Le genre Macrocephalites au Callovien inférieur (Ammonites, Jurassique moyen).* - Mém. géol. Univ. Dijon *4*.

YOUNG, R.N., & SOUTHARD, J.B. (1978): *Erosion of fine-grained marine sediments: Sea-floor and laboratory experiments.* - Bull. geol. Soc. Amer. *89*, 663-672.

ZEISS, A. (1955): *Zur Stratigraphie des Callovien und Unter-Oxfordien bei Blumberg (Südbaden).* - Jh. geol. Landesamt Bad.-Württemb. *1*, 239-266.

ZIEGLER, B. (1962): *Die Ammoniten-Gattung Aulacostephanus im Oberjura (Taxonomie, Stratigraphie, Biologie).* - Palaeontographica (A), *119*, 1-172.

— (1964): *Bewuchs auf Spongien.* - Paläont. Z. *38*, 88-97.

— (1967): *Ammoniten-Ökologie am Beispiel des Oberjura.* - Geol. Rdsch. *56/2*, 439-464.

ZOBELL, C.E. (1946): *Studies on redox potential of marine sediments.* - Bull. amer. Assoc. Petroleum Geol. *30/4*, 477-513.