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Eustatic sea level changes of the Oxfordian (Late Jurassic) and their effect documented in sediments and fossil assemblages of an epicontinental sea

By Reinhart A. Gygi[1])

ABSTRACT

The eustatic sea level history of the Oxfordian was reconstructed from a comparative study of subsidence and of depositional sequences, and from a paleobathymetric analysis of microfacies and of fossil assemblages in a well-known platform to basin transition of northern Switzerland. A strong pulse of eustatic sea level rise occurred late in the Early Oxfordian, and another in the early Middle Oxfordian. The vertical crowding of peritidal facies in proximal environments and strong progradation into the basin in the later part of the Middle Oxfordian suggest that sea level rose less at this time. There is no direct evidence for a sea level fall. Sea level rises in the Late Oxfordian were greater again. The eustatic sea level changes of the Oxfordian resulted in a net rise which may have been as much as 25 or 30 m.

The maximum rate of supply of siliciclastic mud and silt from land occurred in the late Middle Oxfordian. At this time, a carbonate platform was formed. Tidal currents and storms transported a large part of the fine-grained terrigenous sediment across the platform and into the basin. The major part of basement subsidence was caused by sediment loading. The abundance and the composition of the ammonite fauna changes with paleodepth but is independent of lithofacies. The composition of the fossil assemblages changes rapidly and thoroughly with paleodepth down to at least 150 m. Therefore, the macrofauna provides detailed information about the water depth.

RÉSUMÉ

L'histoire des variations eustatiques du niveau de la mer à l'Oxfordien a été reconstruite à l'aide d'une étude comparative de la subsidence et des séquences sédimentaires, et d'une analyse paléobathymétrique des microfaciès et des assemblages de fossiles dans le cadre d'une transition plate-forme/bassin de la Suisse septentrionale. Le niveau de la mer montait rapidement vers la fin de l'Oxfordien Précoce, puis dans la première partie de l'Oxfordien Moyen. Dans la dernière partie de l'Oxfordien Moyen, des horizons du milieu péritidal se succèdent fréquemment dans le domaine proximal et on observe une progradation rapide dans le bassin. Cela indique que le niveau de la mer a moins monté pendant ce temps; un abaissement ne peut pas être prouvé directement. Le niveau de la mer recommençait à monter à l'Oxfordien Tardif. Les variations eustatiques du niveau de la mer pendant l'Oxfordien entier résultaient d'une montée nette qui est estimée entre 25 et 30 m.

L'apport de vases terrigènes et de sables à grain fin du nord atteignait un maximum dans la dernière partie de l'Oxfordien Moyen. C'est à cet âge que se constituait une plate-forme carbonatée. Des courants de marée et des tempêtes ont transporté une grande partie du sédiment terrigène à grain fin surtout siliciclastique au travers de la plate-forme dans le bassin. La majeure partie de la subsidence du socle a été provoquée par la surcharge des sédiments. La fréquence et la composition des faunes d'ammonites changent avec la profondeur de l'eau indépendamment des lithofaciès. La composition des assemblages fossiles change rapidement et fondamentalement à une paléo-profondeur d'au moins 150 m. La macrofaune fournit donc des informations détaillées sur la profondeur de l'eau.

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ZUSAMMENFASSUNG

Die eustatischen Meeresspiegelschwankungen des Oxfordian wurden rekonstruiert mit Hilfe einer vergleichenden Untersuchung der Subsidenz, der Ablagerungssequenzen sowie durch die paläobathymetrische Analyse der Mikrofazies und der Fossilvergesellschaftungen in einem Plattform/Becken-Übergang der Nordschweiz, welcher die ehemaligen Formationen «Rauracien» und «Argovien» einschliesst. Am Ende des Frühen Oxfordian fand ein starker und schneller Anstieg des Meeresspiegels statt, und dann nochmals im frühen Mittel-Oxfordian. Das vertikal gedrängte Vorkommen von peritidalen Fazies im proximalen Bereich und starke Progradation ins Becken im späten Mittel-Oxfordian bedeuten, dass der Meeresspiegel in jener Zeit weniger anstieg. Ein Absinken des Meeresspiegels konnte nicht direkt nachgewiesen werden. Im Späten Oxfordian begann der Meeresspiegel wieder stark anzusteigen. Die eustatischen Meeresspiegelschwankungen führten im Oxford-Alter zu einem Netto-Anstieg, welcher auf 25–30 m geschätzt wird.

Im späten Mittel-Oxfordian erreichte die Zufuhr von vorwiegend siliziklastischem Schlamm und Silt vom Land ein Maximum. Zur gleichen Zeit entstand eine Karbonatplattform. Gezeitenströmungen und Stürme verfrachteten einen grossen Teil des feinkörnigen terrigenen Sediments über die Plattform hinweg ins Becken. Der grössere Teil der Subsidenz wurde durch Sediment-Auflast verursacht. Die Häufigkeit und die Zusammensetzung der Ammonitenfauna änderten sich mit zunehmender Wassertiefe unabhängig von der Lithofazies. Die Zusammensetzung der Fossilvergesellschaftungen ändert sich mit zunehmender Paläo-Wassertiefe schnell und gründlich. Die Makrofauna liefert deshalb detaillierte Informationen über die Wassertiefe.

CONTENTS

1. Introduction

Improved correlation of the Late Jurassic platform to basin transition in northern Switzerland (GYGI & PERSOZ 1986) provided an opportunity to reconstruct the depositional and sea level history of the transition. This paper is complementary to the work of GYGI & PERSOZ (1986) and gives additional explanations of Plate 1 in that publication. The purpose of this paper is to interpret and discuss published work on the Late Jurassic of northern Switzerland in the light of recent research carried out elsewhere, mainly in the field of seismic stratigraphy. Conclusions taken from earlier papers are normally cited here without further explanation. Some hitherto unpublished data are inserted into the text where necessary.

Estimates of the water depth are made by the paleoecologic methods given by GYGI & PERSOZ (1986). Paleobathymetric conclusions made from fossil assemblages are based on more than 10,000 macrofossil specimens collected from in situ. Qualitative evidence of eustatic sea level changes was derived from analysis of lateral and vertical facies changes, or from the geometry of lithologic units. Approximate figures of subsidence and of relative sea level rise were calculated from the depositional history. Eustatic sea level rises

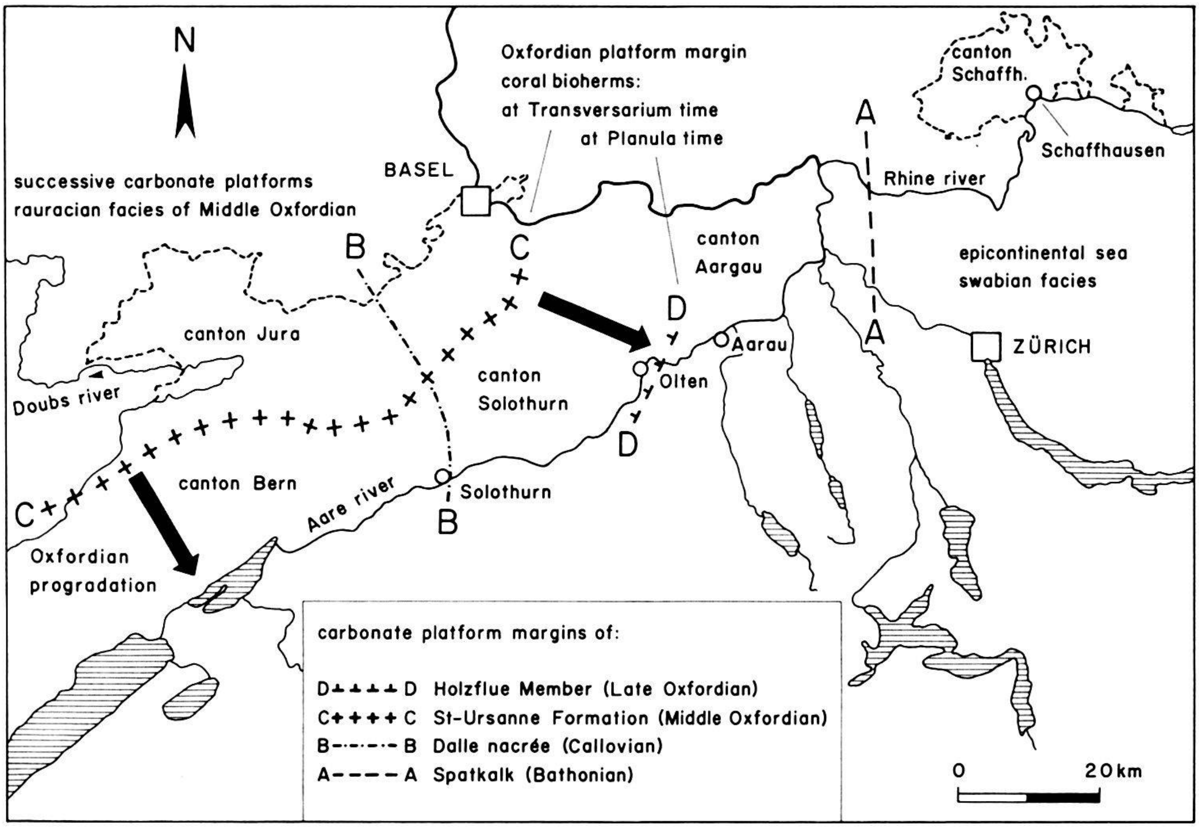

Fig. 1. Boundaries at maximum extension of successive carbonate platforms prograding from the northwest in Middle and in Late Jurassic time (cf. Fig. 2).

were estimated by comparison of a vertical facies change with sedimentation in the Recent. The paleobathymetric intervals of characteristic fossil assemblages of the early Kimmeridgian were established on a previously calculated depositional profile. The paleobathymetry concluded from fossil assemblages of the Early Oxfordian was checked against the depth intervals established of the Kimmeridgian fossil assemblages. Then, the effect of eustatic sea level changes on sedimentation is discussed. A summary of the seaward advance of the coral bioherm facies in Oxfordian time is given in the Figures 1 and 2. Many conclusions are made within the text. Only the main conclusions are summarized at the end of the paper.

The platform to basin transition considered is in an epicontinental sea. The basin is therefore referred to with quotation marks in the following text. Stratigraphic names are used according to Table 2 in Gygi & Persoz (1986). Geographic names in the text not included in Figure 1 are to be found in Figure 1 by Gygi & Persoz (1986).

2. Evidence of sea level changes from the depositional history

2.1 *Middle Jurassic: The drowning of carbonate platforms*

The bathymetric profile at the beginning of the Oxfordian was influenced by the marginal parts of several successive Middle Jurassic carbonate platforms (Burgundy Platform of Purser 1979) which extended from the northwest into our region. Maximum progradation occurred late in the early Bathonian when the thin bioclastic, cross-bedded

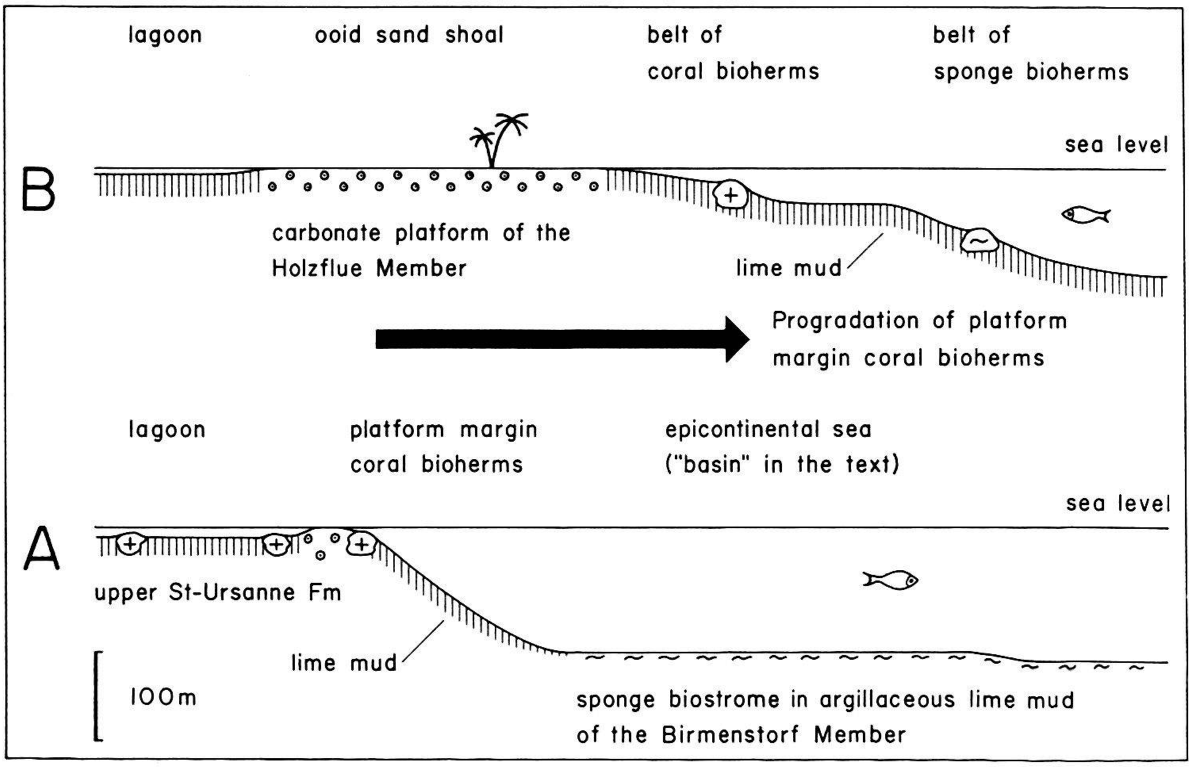

Fig. 2. Progradation of the marginal coral bioherm belt in successive carbonate platforms of the Oxfordian. A: Middle Oxfordian, late Transversarium Chron. B: Late Oxfordian, Planula Chron (Gygi & Persoz 1986, Table 2, and Pl. 1A). Bathymetry is approximate, see discussion in the text. No horizontal scale is given, because the platform slope and the amount of progradation vary along depositional strike.

calcarenite of the Spatkalk reached the eastern part of canton Aargau (Fig. 1). The Spatkalk was named by Moesch (1867, p. 93) and biochronologically dated by Hahn (1966). This bioclastic limestone member has a substantial iron content. The Spatkalk is apparently a ferriferous facies belt about 20 km wide that was marginal to a platform with more or less pure carbonates in the interior. Little or no sediment was deposited in eastern canton Aargau from the middle Bathonian to the Early Oxfordian (Gygi & Marchand 1982, Fig. 2). The next, important carbonate platform was the bioclastic calcarenite of the Dalle nacrée Member which was deposited in the later part of the early Callovian (Bayer et al. 1983, Fig. 4). The margin of this platform crosses the transect of Plate 1A by Gygi & Persoz (1986) obliquely near Solothurn. The margin of the Dalle nacrée platform is displaced far in the proximal direction with respect to the Spatkalk platform margin (Fig. 1). Vertical transition from the intertidal or shallow subtidal environment of the Spatkalk to deeper marine starved basin conditions occurred near Auenstein in canton Aargau at the end of the early Bathonian (earlier than the Subcontractus Subchron, see Gygi & Marchand 1982, Fig. 2), and late in the early Callovian (earlier than the Enodatum Subchron) further northwest (Bayer et al. 1983, Fig. 4). The vertical facies transitions resulted from sea level changes and from very marked changes from high rates of deposition to near-nondeposition.

In northern Switzerland, the late Middle Jurassic (middle and late Callovian) was a time of very slow or intermittent sedimentation or of non-deposition. Where sediments occur, they are thin sheets or lenses of iron-oolitic marl or limestone. No sediments are known of the early Lamberti Chron, the last chron of the Middle Jurassic. Sedimentation recommenced at a slow rate in the Lamberti Subchron just before the beginning of the Late Jurassic, when a thin sheet of iron-oolitic marl or limestone was laid down in northwestern Switzerland (GYGI & MARCHAND 1982, Fig. 2 and 3). This bed is continuous at least from near Courgenay to Péry (GYGI & PERSOZ 1986, Pl. 1A). From there in the distal direction, the sediment is only known as isolated patches or lenses.

The paleoenvironment of this type of thin and widespread oolitic ironstone with a groundmass of mud and a cephalopod-dominated macrofauna was investigated by GYGI (1981). The ferriferous ooids of the Early Oxfordian were accreted at the sediment surface of argillaceous or carbonate mud when the net rate of mud sedimentation was very slow. They were formed at a depth of as much as 100 m. A water depth of about 100 m is apparently the lower limit to iron ooid accretion in an epicontinental sea.

2.2 Late Jurassic: the bathymetric profile at the beginning of the Oxfordian

Sediments of the earliest Oxfordian of northern Switzerland are, where present, invariably thin, iron-oolitic sheets or lenses, with either an argillaceous or a carbonate-rich mud matrix. The macrofauna is dominated by cephalopods. The uniformity of the facies and of the fauna at the transition from the Middle to the Late Jurassic is an indication that the seafloor topography was subdued at that time. Neither sills nor troughs can be discerned. The sediments were laid down below normal wave base (Fig. 3A).

In northern Switzerland, Bathonian and early Callovian deposits are mainly of three successive carbonate platforms, which in the south and in the east grade abruptly into thinner marl laid down in deeper water. Water depth at the beginning of the Oxfordian must therefore have been least in the northwest and greatest near Schaffhausen. The transition from the Middle to the Late Jurassic is not documented by sediments in canton Schaffhausen. However, a few kilometers across the border in southern Germany, a section northwest of Blumberg (no 87 in GYGI 1969, p. 51), and another in the former open-pit iron mine at the foot of Stoberg hill northeast of Blumberg (ZEISS 1955, Fig. 31), indicate that there, the vertical change from iron-oolitic to glauconitic sedimentation probably coincided with the Callovian/Oxfordian boundary. According to GYGI (1981, p. 244), this means that near Blumberg, the water depth increased from less than about 100 m in the latest Callovian to more than 100 m in the Oxfordian. The probable water depth near Schaffhausen at the beginning of the Oxfordian is assumed from this to be approximately 100 m (Fig. 3A).

The water depth at this time in the northwest was less than 100 m, as indicated by the iron ooids in an argillaceous mud matrix in the thin and widespread iron-oolitic horizon at the base of the Renggeri Member (GYGI & PERSOZ 1986). B. ZIEGLER (1967, 1971) demonstrated that the abundance and composition of the cephalopod assemblage present in a macrofauna is indicative of depth. It is apparent from a comparison of macrofossil assemblages from different depths, based on more than 10,000 specimens collected from in situ by R. and S. Gygi, that from a paleodepth of about 80 m downward, cephalopods comprise more than 80% of the macrofauna (see Fig. 6B). The macrofauna of the iron-oolitic horizon at the base of the Renggeri Member is 84% cephalopods near Péry:

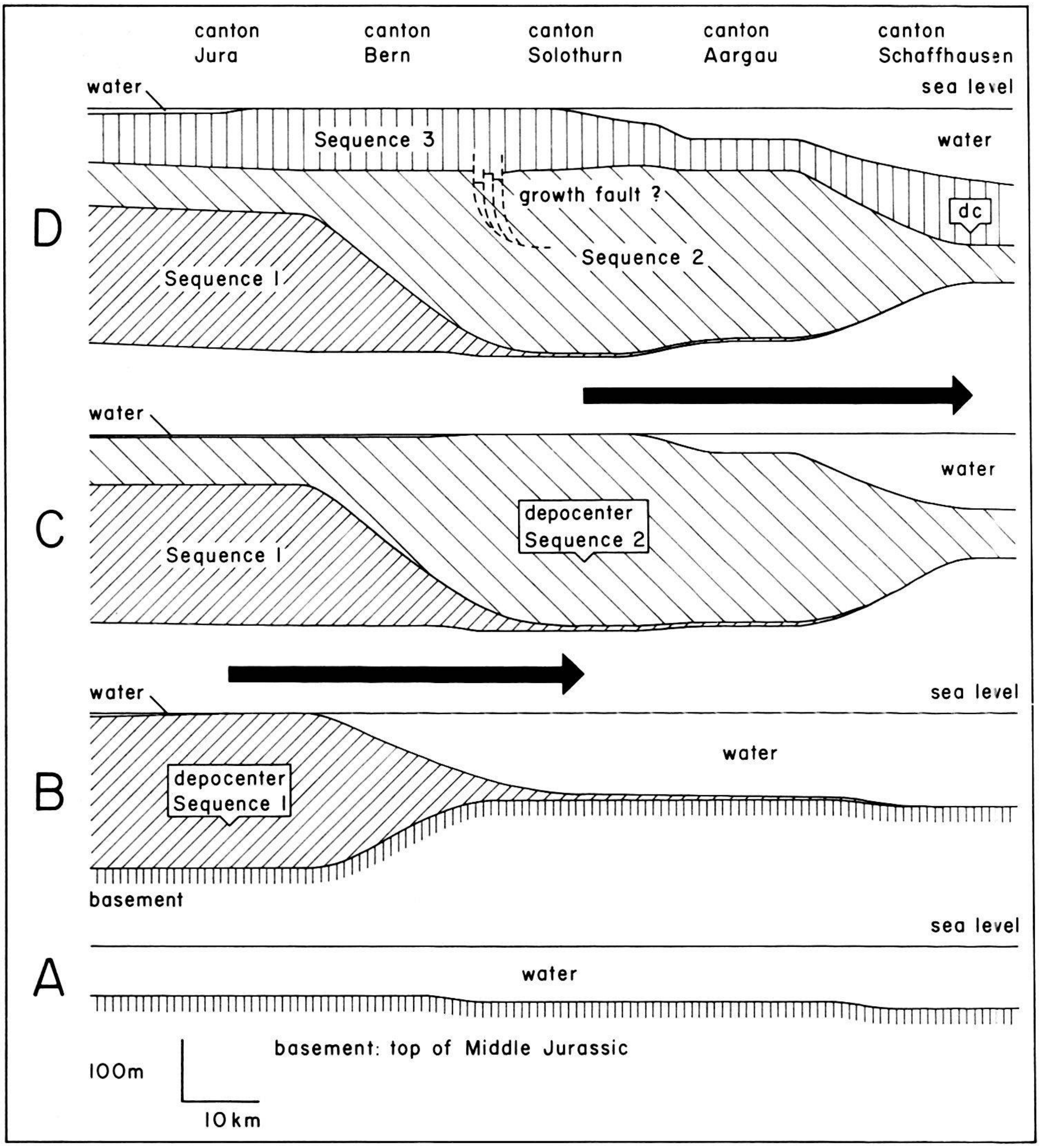

Fig. 3. Differential basement subsidence under shifting depocenters. A: Bathymetric profile at the beginning of the Oxfordian. B: End of Transversarium Chron. C: End of Bimammatum Subchron. D: End of Planula Chron. Sequences according to GYGI & PERSOZ (1986). The great thickness anomalies in the Günsberg Member, which are probably related to growth faults, are not shown (compare GYGI 1969, Fig. 9, with WEIMER & DAVIS 1977, Fig. 9). Compaction is taken into account in this schematic representation.

68% ammonites and 16% belemnites, with 16% other groups, mainly infaunal echinoids. GYGI (1981, p. 245) deduced mainly from sedimentologic evidence that the depth of deposition of the similar thin Schellenbrücke Bed was between 80 and 100 m. In this sediment, there are ferriferous ooids in a carbonate mud matrix, and cephalopods are 86% of the macrofauna (GYGI 1981, p. 239). A minimum water depth of 80 m in the northwest at the beginning of the Oxfordian is also suggested when the thicknesses, facies, and subsidence history of sediments from the middle Bathonian to the end of the Callovian are compared between the easternmost canton Aargau and the northwest by the methods used below for the Oxfordian.

2.3 The lower Oxfordian argillaceous mud bank in northwestern Switzerland

The Renggeri Member, the Terrain à Chailles Member, and the Liesberg Member formed a submarine bank of mainly siliciclastic mud. The mud of the three members accumulated a positive physiographic structure, because the net rate of sedimentation was greater than that of compaction and basement subsidence. Gradual shoaling of the seafloor in the course of sedimentation is documented by the macrofauna. Ammonites predominate from the base of the Renggeri Member as far up as the middle Terrain à Chailles Member. Bivalves and brachiopods are most abundant in the upper Terrain à Chailles Member. The base of the Liesberg Member is characterized by the sudden and massive appearance of dish-shaped hermatypic corals. Dish-shaped coral colonies are a substantial part of the rock volume of the Liesberg Member. In Recent reefs, dish-shaped coral colonies are dominant in the deep fore-reef at a depth as great as 100 m or more, or in shaded areas higher up. They thus indicate primarily a low average illumination, and only indirectly water depth. Their Late Jurassic counterparts are therefore not necessarily indicative of a great water depth. The first corals of the Liesberg Member became established at a depth of probably less than 20 m (GYGI & PERSOZ 1986) while siliciclastic mud was deposited at a relatively high rate. The thickness of the Liesberg Member is 25 m

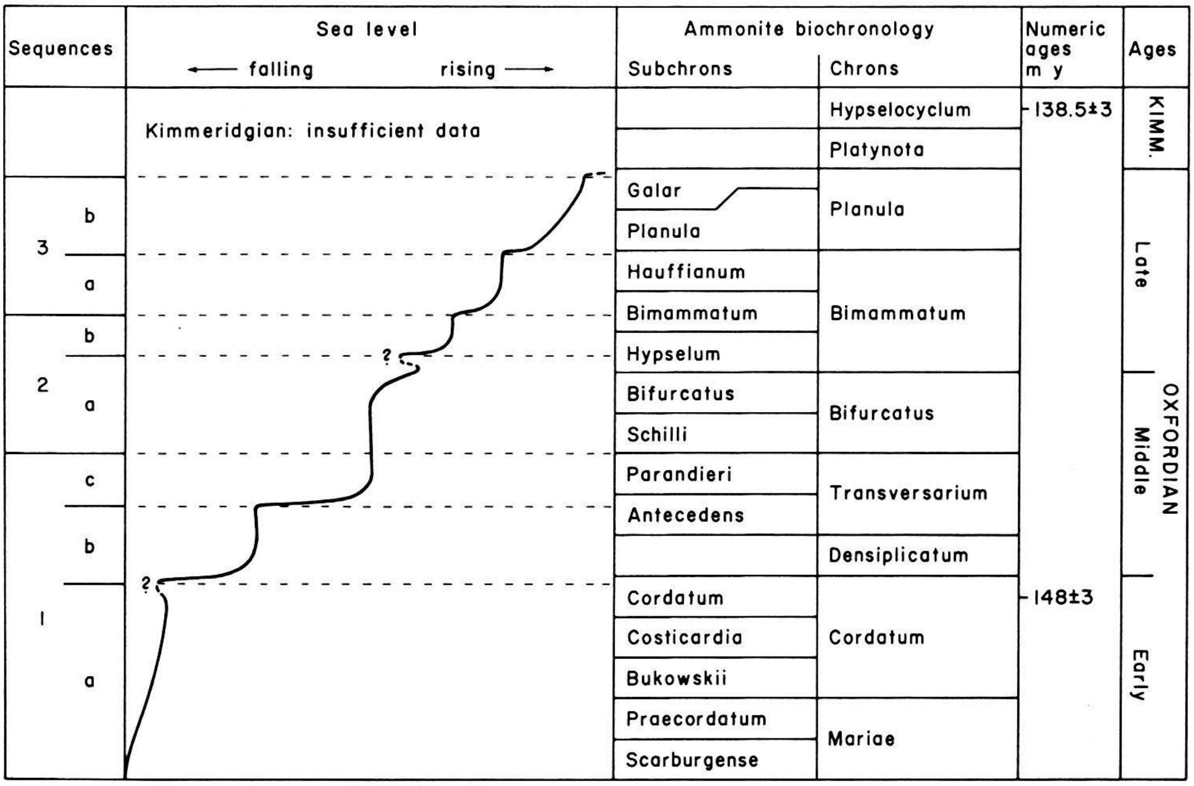

Fig. 4. Semiquantitative curve of eustatic sea level changes in the Oxfordian Age (see estimate in the text). Numeric ages are based on glauconites measured by GYGI & MCDOWELL (1970). For later refinements in ammonite biochronology: see text and GYGI & PERSOZ (1986, Table 3).

near Liesberg. This was laid down in a fraction of the Antecedens Subchron (Gygi & Persoz 1986, Table 2). The average time span of an Oxfordian subchron is of the order of 670,000 years (see Fig. 4 for numerical ages). Therefore, the mean turbidity of the water may have been substantial. Turbidity of the water does not inhibit hermatypic coral growth even when it is permanent (Zlatarski & Martinez 1982, p. 395), but it reduces light penetration and thus the depth range in which hermatypic corals may live (Umbgrove 1947, p. 772). The first coral bioherms of the St-Ursanne Formation developed above the margin of the mud bank in Antecedens time (Middle Oxfordian, Gygi & Persoz 1986, Pl. 1A). The carbonate platform of the St-Ursanne Formation was shaped by the configuration of the siliciclastic mud bank below (see Contini 1976, Fig. 3, for a paleogeographic map of the southern part of the mud bank).

2.4 Qualitative evidence of discrete eustatic sea level changes

Through much of Early Oxfordian time, the ratio of sediment supply to relative sea level rise was such that the margin of the mud bank receded (Gygi & Persoz 1986, Pl. 1A). The net sedimentation rate beyond the distal fringe of the bank was initially zero in most places (Gygi & Marchand 1982, Fig. 4). This tendency was reversed in the Cordatum Subchron when, possibly because of a relative fall in sea level (Fig. 4), mud spilled down the slope of the bank and reached the "basin". This is when the decimeter-thick iron-oolitic Schellenbrücke Bed and the glauconitic Glaukonitsandmergel Bed were laid down.

2.4.1 Evidence from the ammonite fauna

The iron-oolitic Schellenbrücke Bed was deposited at a water depth of 80 to 100 m. This conclusion of Gygi (1981) was based on evidence of the sedimentology, the mineral composition, the taphonomy of ammonites, and of the abundance and the composition of the ammonite fauna in the bed. The lateral facies boundary between the Schellenbrücke Bed and the coeval Glaukonitsandmergel Bed is thought to be at a paleodepth of about 100 m, at the lower limit of iron ooid formation. A vertical change from formation of ferriferous ooids to glauconite generation occurred in canton Schaffhausen prior to the Cordatum Subchron of the Early Oxfordian. The same vertical facies change took place in canton Aargau while the condensed bed at the base of the Birmenstorf Member was sedimented (Gygi 1969, Pl. 17, section 32, horizon 5, or section 60, horizon 4). Formation of this bed in canton Aargau began in the Densiplicatum Chron of the Middle Oxfordian (Gygi & Marchand 1982, Fig. 2). Consequently, glauconite formation began first in canton Schaffhausen, and later expanded into canton Aargau. In the Cordatum Subchron, the water depth was greater than 100 m in canton Schaffhausen, and less than 100 m in canton Aargau (Gygi 1981, p. 244). The difference in water depth persisted in the earlier Middle Oxfordian because of very low sedimentation rates in both regions (see Gygi & Marchand 1982, Fig. 2). There is reason to assume that the vertical facies change took place both in canton Schaffhausen and in canton Aargau when the water depth increased to more than 100 m as a consequence of a relative sea level rise. If the rise was eustatic, it would provide for an explanation why the vertical facies transition took place first in the somewhat deeper water of canton Schaffhausen, then in the shallower water of canton Aargau (cf. GYGI & PERSOZ, Pl. 1A, base of sequence 1 in the area between Aarau and Schaffhausen).

9% of the ammonites in the iron-oolitic Schellenbrücke Bed are Oppeliidae (Table 1). Oppeliidae are 34% of the ammonites in the time-equivalent, glauconite-rich Glaukonitsandmergel Bed of canton Schaffhausen. According to B. ZIEGLER (1967, 1971), the percentage of Oppeliidae in an ammonite fauna is sensitive to paleodepth, and it increases with paleodepth (see chapter 3 on fossil assemblages below). In canton Schaffhausen, the percentage of oppeliid ammonites increases vertically from 34 in the Glaukonitsandmergel Bed (Cordatum Subzone) through the Mumienkalk Bed to 49 in the thin glauconitic marl of the Parandieri Subzone above the Mumienkalk Bed (Table 1). In canton Aargau, the percentage of oppeliid ammonites increases vertically from 9% in the Schellenbrücke Bed of the Cordatum Subzone to 51% in the upper Birmenstorf Member of the Parandieri Subzone (Table 1).

The relative sea level rise leading to the vertical change in the ammonite fauna in both areas began late in the Cordatum Subchron. The water was at that time deeper than 100 m

Table 1: *Lateral and vertical variation in the composition of the ammonite fauna of successive lithostratigraphic units.*

	Canton Aargau		**Canton Schaffhausen**	
	Birmenstorf Member, normal facies with glauconite, Parandieri Subzone Excavations RG 60, 210, 225, 230, 276		Unnamed glauconitic marl, ca 0.2m thick, above Mumienkalk, Parandieri Subzone, Excavations RG 81, 207, 212	
	Individuals	Percentages	Individuals	Percentages
Phylloceratidae	15	0.8	-	-
Oppeliidae	912	**51.0**	86	**49.1**
Perisphinctidae	780	44.0	85	48.6
Cardioceratidae	15	0.8	1	0.6
Aspidoceratidae	60	3.4	3	1.7
	1782	100	175	100
			Mumienkalk, with glauconite Late Antecedens Subzone Excavations RG 80, 81, 88, 207, 212	
			Individuals	Percentages
Phylloceratidae	Insufficient data		6	0.4
Oppeliidae			728	**45.0**
Perisphinctidae			850	52.5
Cardioceratidae			13	0.8
Aspidoceratidae			22	1.3
			1619	100
	Schellenbrücke Bed, iron oolitic Cordatum Subzone, Excavation RG 208		Glaukonitsandmergel, Cordatum Subzone Excavations RG 81, 207, 212	
	Individuals	Percentages	Individuals	Percentages
Phylloceratidae	14	1.0	2	0.5
Oppeliidae	124	**9.3**	132	**34.3**
Perisphinctidae	565	42.3	162	42.1
Cardioceratidae	568	42.5	73	18.9
Aspidoceratidae	65	4.9	16	4.2
	1336	100	385	100

in canton Schaffhausen, and shallower than 100 m in canton Aargau. The composition of the ammonite fauna changed with increasing water depth independently from the lithofacies (Table 1). It is evident from ammonites that the relative sea level rise occurred at the same time in areas tens of kilometers apart, and that the rate of sedimentation varied between slow and non-deposition in that time. Early Middle Oxfordian sedimentation was so insignificant in canton Aargau and in canton Schaffhausen that it did not influence subsidence of the basement.

It was inferred above that the difference in water depth between canton Jura in the northwest and canton Schaffhausen was of the order of 20 m at the beginning of the Oxfordian (Fig. 3A). Gygi (1981, Fig. 4) estimated the paleobathymetric difference between the Schellenbrücke Bed in canton Aargau and the Glaukonitsandmergel Bed in canton Schaffhausen, both of the Early Oxfordian Cordatum Subchron, at several tens of meters. Indeed, there is a marked difference in the percentage of the depth-sensitive oppeliid ammonites between the Schellenbrücke Bed and the coeval Glaukonitsandmergel Bed (Table 1). It can be seen from Table 1 that later, the difference in the oppeliid percentage almost disappeared in the upper Birmenstorf Member of canton Aargau and in its time equivalent in canton Schaffhausen of the Parandieri Subchron (see Fig. 4 for scale of biochronologic units, and Table 2 in Gygi & Persoz 1986, for lithostratigraphic units).

Provided the proportion of the oppeliid percentages between the Schellenbrücke Bed of canton Aargau and the Glaukonitsandmergel Bed of canton Schaffhausen changed in younger sediments, because the difference in water depth changed with time, then sedimentation cannot be the direct cause, since the measured thickness of sediments from the time interval of the Cordatum Chron to the end of the Transversarium Chron is less than 5 m in canton Aargau, and less than 0.5 m in canton Schaffhausen (Gygi 1977, Pl. 11). At the same time, sediments with an average compacted thickness of 185 m were laid down in the northwest (sequence 1, Fig. 3B). It could be argued that the high rate of deposition and the strong basement subsidence under the sediment load in the northwest (Fig. 3B) caused the basement some distance out in the neighbouring "basin" to form a broad swell in a compensational movement. The uplift would have occurred in parts of canton Solothurn and of canton Aargau. The rate of uplift in the "basin" would have been greatest in the Cordatum Chron. The uplift in the "basin" would have died down at the end of the Transversarium Chron when the sedimentation rate in the shallow-water realm of the northwest was sharply reduced once the sediment surface in that area reached the intertidal zone. All this is improbable, because it may be assumed by comparison with the Recent that the isostatic adjustment of the basement was prompt with respect to the observed sedimentation rates (see below). In fact, ammonite assemblages changed with time, so that the paleoenvironmental evidence of a given ammonite percentage must be evaluated separately for any interval of time (see chapter 3).

2.4.2 Evidence from sediments and coral bioherms

Evidence of a rapid and substantial relative sea level rise beginning in the late Antecedens Subchron comes from the lower boundary of subsequence 1c near St-Ursanne where calcareous oolite passes vertically into coral bioherms (see Gygi & Persoz 1986, Pl. 1A). The rate of relative sea level rise became greater than that of oolite sedimentation at this location. In the meantime, the area of ooid sand generation shifted

Fig. 5. Coral bioherm (lagoon reef) in the St-Ursanne Formation northeast of St-Ursanne. 1 = oolitic limestone with a slight content of oncoids of the lower St-Ursanne Formation, upper part of subsequence 1b. 2 = porous and friable lime mudstone, covered by forest, with conspicuous lagoon reef of the upper St-Ursanne Formation, subsequence 1c. Vertical extension of reef is about 18 m. 3 = peritidal, bedded micritic limestone of the Vorbourg Member, sequence 2. For lithostratigraphic units see GYGI & PERSOZ (1986, Table 2).

to the platform margin (see GYGI & PERSOZ 1986, Pl. 1A). Near St-Ursanne, the water grew deeper until it was too deep for the formation of ooid grains. A lagoon with a normal marine salinity was formed. Coral patch reefs developed in the lagoon as soon as the water circulation was sufficient (Fig. 5). The base of subsequence 1c cannot be discerned at the platform margin. Carbonate precipitation is at a maximum near the margin of a platform. The area of ooid sand generation shifted from near St-Ursanne in subsequence 1b to the platform margin beyond Glovelier in subsequence 1c (GYGI & PERSOZ 1986, Pl. 1A). There, the rate of ooid sand formation was sufficient to keep pace with the relative sea level rise. The lagoon became thus rimmed by ooid sand shoals and by isolated coral bioherms. The bioherms of the platform margin were growing either within the marginal part of the oolite shoals, possibly in deep tidal channels as is now the case near the Joulters Cays north of Andros in the Bahamas, or somewhat off the platform margin in deeper water.

2.4.3 Sea level changes in France and in southern England

The event of rapid relative sea level rise in the late Antecedens Subchron near St-Ursanne is not merely local. ENAY (1966, p. 284) drew attention to the "transgression argovienne" above an interregional unconformity in southeastern France. Normal sedimentation of the Birmenstorf Member began in the southern part of the French Jura Mountains at the same time as in canton Aargau. TALBOT (1973, p. 308–9) interpreted the unconformity in southeastern France (discordance antéargovienne of ENAY 1966, Fig. 76) as being a horizon of marine transgression, and he invoked a eustatic rise of sea level to be the cause instead of Enay's differential subsidence. Talbot correlated the transgression with the base of his Osmington Oolite Group in southern England (TALBOT 1973, Table 2).

An unconformity of late Antecedens age was recognized in southern England below the Coral Rag, the Wheatley Limestone, and the Oakley Beds (ARKELL 1947, p. 99). TALBOT (1973, p. 308) concluded that this represents a transgression caused by a eustatic sea level rise. Additional evidence of this transgression in southern England is to be found in R. C. L. WILSON (1968, Fig. 7B). Talbot divided the Oxfordian sediments of southern England into cycles. He thought that the boundary between his cycles 1 and 2 was within the Vertebrale Subzone, which is the equivalent of our Densiplicatum Zone. It is possible that this boundary is in fact near the top of the Cordatum Subzone, because, according to WRIGHT in COPE et al. (1980, Fig. 11a), the Nothe Grit is of the Cordatum Subzone. If this is correct, the boundary between the cycles 1 and 2 of Talbot would be the same age as the boundary between subsequences 1a and 1b of GYGI & PERSOZ (1986), and the cycles 1, 2, and 3 of Talbot would be the equivalent of our subsequences 1a, 1b, and 1c.

2.4.4 Timing of Early and Middle Oxfordian eustatic sea level rises

The latest sediment deposited at the platform margin of the St-Ursanne Formation was oolite formed in the intertidal or in the shallow subtidal zone. The fauna is dominated by hermatypic corals. The depth of deposition of the coeval Birmenstorf Member in the "basin" was apparently greater than 100 m (Fig. 2A). We conclude this mainly from the muddy sediments, the macrofauna which is dominated by siliceous sponges and ammonites, from the composition of the ammonite fauna (see Table 1), and from the vertical facies transition from the iron-oolitic sediments of the Early Oxfordian to the glauconitic marl and limestones of the Birmenstorf Member (Middle Oxfordian) at the toe of the slope facies near Günsberg (GYGI & PERSOZ 1986, Pl. 1A). The vertical facies change at the base of the Birmenstorf Member can be followed from eastern canton Aargau through canton Solothurn to Günsberg (section RG 14 in Pl. 18 of GYGI 1969). The youngest iron ooids of northern Switzerland occur in the basal, condensed bed of the Birmenstorf Member.

The vertical change from ferriferous ooid to glauconite facies took place prior to the Cordatum Subchron in canton Schaffhausen, and early in the Densiplicatum Chron in canton Aargau (GYGI 1977, Pl. 11). The vertical transition from calcareous ooid sand generation to coral reef growth near St-Ursanne occurred late in the Antecedens Subchron. It is by comparison with southeastern France and with southern England that the latter vertical facies change can be related to a eustatic sea level rise. Coincidence of particular pulses of endogenic subsidence in regions so far apart is highly improbable. Eustatic sea level rises are the most probable cause for the earlier vertical facies changes as well. Therefore, eustatic sea level rises are likely in the Early Oxfordian (before the Cordatum Subchron), at the turn from the Early to the Middle Oxfordian (Cordatum/Densiplicatum Chron), and in the late Antecedens Subchron (Fig. 4).

2.4.5 *Regional unconformities*

Given a constant rate of sediment supply, a rapid eustatic sea level rise will push back the sediment source and reduce sedimentation to a lower rate in the proximal part of the "basin", or to zero further in the distal direction (see below). A hiatus will thereby develop. BOLLIGER & BURRI (1970, Fig. 30) were of the opinion that tectonics governed sedimentation in the region from the latest Callovian to the Middle Oxfordian, and they interpreted a hiatus to be evidence of emergence. The hiatus of the early Lamberti Chron below the iron-oolitic marl at the base of the Renggeri Member (GYGI & PERSOZ 1986, Table 2) and the hiatus below the base of the Pichoux limestone (GYGI & PERSOZ 1986, Pl. 1A, and Table 2) certainly did not develop because sediments below the unconformities were emergent at any time. The basement in the northwest moved mainly because of loading with the accumulating lower Oxfordian argillaceous mud bank and the carbonate platform of the St-Ursanne Formation. The subsidence induced by these sediments can be calculated to be about 115 m (Fig. 3B). The hiatuses below and above the Schellenbrücke Bed can be related to eustatic sea level rises in the Early Oxfordian (Fig. 4). The interregional hiatus of essentially the Densiplicatum Chron above the Schellenbrücke Bed and the condensed bed at the base of the Birmenstorf Member (Densiplicatum Chron and Antecedens Subchron) are probably the equivalent of the condensed section J3.1 of VAIL et al. (1984, Table 1).

2.4.6 *Later sea level changes*

The rapid progradation of the carbonate platform of the Günsberg Member in Bifurcatus time was probably caused by a stillstand or by a minor relative fall of sea level. The vertical crowding of peritidal deposits in the distal part of the Natica Member (subsequence 2a of GYGI & PERSOZ 1986, Pl. 1A) supports this conclusion. Progradation of the Günsberg Member was interrupted by a minor and probably gradual sea level rise which began late in the Bifurcatus Chron (Fig. 4). Thin coal seams and other supratidal facies at the top of subsequence 2a suggest that sea level possibly fell in Hypselum time.

The thin but widespread marine sediments with hermatypic corals, echinoids, and brachiopods of the uppermost Natica Member (lower part of subsequence 2b, GYGI & PERSOZ 1986, Pl. 1A) are most probably the effect of a small-scale but rapid eustatic sea level rise in the Hypselum Subchron (Fig. 4). This caused the marine facies to advance at least 35 km landward to Bressaucourt near Porrentruy (GYGI & PERSOZ 1986, Pl. 1A) across a coastal plain with a minimal slope and relief. The transgression reduced the rate of deposition significantly in the marginal part of the "basin". The result was an above-average concentration of bivalves and brachiopods in a thin marker bed which was observed between Olten and Aarau by GYGI (1969, Pl. 18, section no 21, Pl. 19, upper section). A presumed stillstand of sea level caused the oolite shoals of the Steinibach Beds to prograde more than 20 km seaward in the latest part of the Hypselum Subchron. Coral bioherms developed near Olten at this time, and the vast, shallow subtidal algal biostrome of the oncolitic Hauptmumienbank Member was formed in the platform interior.

Another relative sea level rise is recorded by the following vertical facies changes: The proximal part of the Hauptmumienbank Member is an almost unfossiliferous carbonate mud deposit from a shallow, restricted lagoon. The fauna of the Humeralis marl above indicates an environment with a normal or a near-normal marine water circulation (see above). The Humeralis marl grades laterally and in the distal direction into the oolite of the Oolithe rousse. This oolite is in most sections above oncolite of the Hauptmumienbank Member. The oolite is therefore transgressive. Formation of the oolite shoals of the Oolithe rousse and of the lagoon of the Humeralis marl was brought about by a relative sea level rise by which the belt of ooid sand generation was pushed back landward. This rise was a major event since it had a noticeable effect on sedimentation from the shallow water realm down to at least the marginal part of the "basin". The fauna of the Geissberg Member of canton Aargau is composed of mostly infaunal bivalves. Partly or entirely fossilized siliceous sponges, glauconite, and corroded bedding planes in the Crenularis Member above indicate a reduced rate of sedimentation. The rather abundant ammonites and the more or less complete fossils of siliceous sponges (see below) in the Crenularis Member in eastern canton Aargau suggest that the member was deposited in deeper water than the underlying Geissberg Member. The Crenularis Member and the upper part of the Hornbuck Member of the Bimammatum Subchron (GYGI & PERSOZ 1986, Table 2, and Pl. 1A) are probably equivalent to the global mid-Late Oxfordian unconformity of VAIL et al. (1984, Table 1).

The relative sea level rise documented by the Late Oxfordian Knollen Beds (Table 2 and Pl. 1A in GYGI & PERSOZ 1986) in the "basin" was certainly of less importance than that of the Crenularis Member. Ammonites are rare in the lower Letzi Member of Mellikon, but they are fairly common near the top of the member. This is an indication that sea level was rising faster than the sediment surface was raised by deposition, and that the water depth increased in the "basin" in the later Planula Chron. The same relative sea level rise caused the margin of the carbonate platform of the Holzflue Member to recede (GYGI & PERSOZ 1986, Pl. 1A).

2.5 The bathymetric profile at the end of the Oxfordian

2.5.1 Outline of the profile

The most prominent feature of the bathymetric profile at the end of the Oxfordian was the vast oolite shoal of the Verena Member. The calcareous ooid sand of the member was formed in the intertidal and in the shallow subtidal zone. The ooid sand graded on the proximal side into carbonate mud which was laid down in a shallow lagoon (GYGI & PERSOZ 1986, Pl. 1A). No fringe of coral bioherms is known from the basinward margin of the younger part of the shoal.

Paleodepth near Schaffhausen is estimated by interpreting the composition of the macrofauna and the sedimentology of the lower part of the Schwarzbach Member (equivalent to the Baden Member, see GYGI 1969, p. 104, Table 9). A macrofauna of 817 specimens was recovered from the lower Schwarzbach Member in the unpublished excavation RG 239 at the Oxfordian/Kimmeridgian boundary near Schaffhausen. The excavation is at the same place as section RG 83 in GYGI (1969, Pl. 16). Horizons from the uppermost Planula Zone to the lowermost Hypselocyclum Zone were worked. Most of the fauna is from the early Kimmeridgian Platynota Zone (see Fig. 4). 81% of the fauna are ammonites (Fig. 6B), and 25% of the ammonites are Oppeliidae (Fig. 6A). The ammonite fauna at the Oxfordian/Kimmeridgian boundary is then similar to that in the Glaukonitsandmergel Bed of the Early Oxfordian (Table 1). The abundance and the composition of the ammonite fauna and the mud grade of the sediment of the Oxfordian/Kimmeridgian transition beds suggest that the depth of deposition was about the same as that of the Glaukonitsandmergel Bed. This is to say that the water was deeper than 100 m at the end of the Oxfordian near Schaffhausen, or about as deep as in the Cordatum Subchron before normal Oxfordian sedimentation began.

2.5.2 Differential basement subsidence under shifting depocenters

It appears in the cross section A of Plate 1 in GYGI & PERSOZ (1986) that the water was 200 m deep at the end of the Oxfordian near Schaffhausen instead of somewhat more than 100 m as concluded above. The cross section A of this plate was drawn as if basement subsidence was the same everywhere in the region considered, or else it would have been necessary to split the section up into time slices. The section was drawn by summing up and by averaging the measured compacted thicknesses of Oxfordian sediments above the bathymetric profile as it probably was at the beginning of the Oxfordian. This was done in order to make the cross section comparable with published stratigraphic sections.

Analysis of the depositional history of Oxfordian sediments gave qualitative evidence that basement subsidence varied widely in time and space as a consequence of the shifting of depocenters, and ultimately of sea level changes (Fig. 3). A quantification of this was necessary in Plate 1A by GYGI & PERSOZ (1986) in order to correct the base of the upper cross section B. Paleodepth, endogenic basement subsidence, load-induced subsidence, and eustatic sea level changes cannot be calculated accurately. On the other hand, the available data are thought to be detailed enough to justify a calculation of approximate figures. The calculation includes the bathymetric interpretations made by GYGI (1981), which are based on several, independent lines of evidence. The most uncertain figures are the initial water depths, the averaged thicknesses, and the thickness reduction of sediments under compaction. The latter had to be calculated according to general empiric data in the literature. Most of the grainstones (mainly oolitic and bioclastic) were observed in thin section to be compacted. Compaction by dissolution of carbonate at grain contacts under pressure was not taken into account, because the amount of compaction can only be estimated from thin sections, and because compaction was found to be variable within individual grainstone units. Compaction of grainstones commenced probably in post-Oxfordian time. CZERNIAKOWSKI et al. (1984) concluded that pressure solution at grain contacts is insignificant at a burial depth of less than 300 m. In the following calculations, carbonate mud is assumed to be compacted about as much as argillaceous mud, as was concluded by TERZAGHI (1940, p. 89). Another assumption is that the rate of endogenic basement subsidence (caused by forces other than loading of the lithosphere with sediments or water) was constant and uniform in the whole region. Uniform subsidence of northern Switzerland in the Late Jurassic was concluded by BÜCHI et al. (1965, p. 33, and Fig. 16).

The essential aspects of basement subsidence history of northern Switzerland can be outlined in the following way: At the beginning of the Oxfordian, the initial water depth in the northwest was about 80 m (see above). By the end of the Transversarium Chron, the northwestern part of the "basin" was filled with sediments to the intertidal zone as indicated by sedimentary structures and fossils in the lowermost Vorbourg Member (GYGI & PERSOZ 1986).

1) Initial water depth in northwestern Switzerland at the beginning of the Oxfordian ca 80 m
2) Water depth at the end of the Transversarium Chron 0 m
3) Measured and averaged thicknesses of lower Oxfordian marls and of St-Ursanne Formation combined (sequence 1 of GYGI & PERSOZ 1986, Pl. 1A) 185 m
4) Original (decompacted) thickness of sediments mentioned above: Biolithite and grainstones with an aggregate thickness of about 45 m were not compacted in Oxfordian time. 140 m of argillaceous and carbonate mudstones with an average post-Oxfordian overburden of about 500 m had an original thickness of 205 m, as read from the nomogram by PERRIER & QUIBLIER (1974, Fig. 11). Total original thickness of sequence 1 in the northwest: 205 + 45 m 250 m
5) Apparent basement subsidence (lithosphere subsidence under sediment load plus relative sea level rise, composed of eustatic sea level rise and endogenic subsidence) in the northwest from the beginning of the Oxfordian to the end of the Transversarium Chron: 4) minus 1) minus 2) 170 m

The initial water depth near Schaffhausen at the beginning of the Oxfordian was approximately 100 m (see above). On the hypothetical assumption that the apparent basement subsidence was equal in the whole region, the water depth near Schaffhausen at the end of the Transversarium Chron would have been the 100 m of initial water depth plus 170 m of apparent subsidence, amounting to 270 m. The composition of the macrofauna of the late Transversarium Chron near Schaffhausen (Table 1) suggests, according to B. ZIEGLER (1967, 1971), that the water depth was about 150 m (Fig. 2A). If the depth is rated at 150 m based on the macrofauna and on additional evidence given below, then the basement would have subsided near Schaffhausen about 120 m less than in canton Jura in the northwest (Fig. 3B):

6) Initial water depth near Schaffhausen at the beginning of the Oxfordian ca 100 m
7) Hypothetical water depth near Schaffhausen at the end of the Transversarium Chron, when a uniform apparent basement subsidence in all northern Switzerland is assumed: 5) plus 6) 270 m
8) Probable water depth near Schaffhausen at the end of the Transversarium Chron ca 150 m
9) Difference in basement subsidence: difference between 7) and 8), ca 120 m

10) Increment in water depth (relative sea level rise) from the beginning of the Oxfordian to the end of the Transversarium Chron near Schaffhausen: difference between 8) and 6), ca 50 m

A relative sea level rise of 100 m is likely for the whole Oxfordian Age (see 16) below). The time span from the beginning of the Oxfordian to the end of the Transversarium Chron is somewhat more than half the Oxfordian Age (cf. Fig. 4), if the duration of the subchrons is assumed to be equal. The increase in water depth by about 50 m near Schaffhausen between the beginning of the Oxfordian and the end of the Transversarium Chron occurred while less than 1 m (original thickness) of sediment was deposited. Basement subsidence under the load of sediments was then negligible. Endogenic subsidence driven by forces other than sediment loading must have been one, but not the only cause for the increment in water depth, because a comparison with coeval sedimentation in France and England produced evidence that two events of eustatic sea level rise were also a factor (see above). It was calculated above that by the end of the Transversarium Chron, the basement in the northwest subsided about 120 m more than near Schaffhausen. When endogenic subsidence is assumed to be uniform, then the additional 120 m of subsidence restricted to a distinct area must be the result of loading of the lithosphere with sediments. The lithosphere subsidence caused by the load of sediments was then about two-thirds the compacted sediment thickness of sequence 1, as stated in 3). At the same time, endogenic subsidence amounted to only part of 50 m. It is apparent from this that sediment loading was by far the most important single factor in basement subsidence in our region in Oxfordian time, however uncertain some of the calculated figures above may be.

It may be concluded from the work of SAURAMO in Scandinavia (in PRESS & SIEVER 1982, Fig. 18–30) that in our region, isostatic adjustment of the lithosphere caused by loading with sediments or water as compared with the calculated sedimentation rates was so swift that the lithosphere was at no time significantly off the isostatic equilibrium. Sedimentology and the composition of the macrofauna in the thin "basinal" deposits of sequence 1 from canton Solothurn to canton Schaffhausen indicate that strong subsidence caused by sediment loading in the northwest did not extend beyond the area where thick sediments accumulated (Fig. 3B).

The thicknesses of the thin, distal horizons of sequence 1 had to be greatly exaggerated in Plate 1A by GYGI & PERSOZ (1986). The aggregate thickness, ranging from a compacted 6 m in canton Solothurn to 0.5 m in canton Schaffhausen, is represented to scale in GYGI (1977, Pl. 11).

Restriction of sediment load-induced basement subsidence to the area of heavy sedimentation indicates that the adjustment of the lithosphere was local isostatic in the sense of STECKLER & WATTS (1978, p. 1). Local isostatic adjustment is only possible when the lithosphere is fractured at close intervals. There is evidence from recent deep drilling in northern Switzerland that the basement is fractured, and that some of the fractures were active in Mesozoic time (MÜLLER et al. 1984, p. 112).

The depocenter of sedimentation shifted from the northwest to a belt running across canton Solothurn and canton Aargau when sequence 2 was deposited in Bifurcatus and in early Bimammatum time (Fig. 3C). The basinal depocenter of sequence 3 is in canton Schaffhausen (Fig. 3D, or GYGI & PERSOZ 1986, Pl. 1A). Oxfordian basement subsidence related to loading with sediments then varied widely in time and space in this region.

2.5.3 *Water depth at the end of the Oxfordian near Schaffhausen*

The water depth near Schaffhausen at the end of the Oxfordian may be calculated if the relative sea level rise during the whole Oxfordian Age in northern Switzerland can be established. The relative sea level rise (composed of endogenic basement subsidence, eustatic sea level rise, and eventual basement subsidence under additional water load) can be calculated from the depositional history in the northwest, where Oxfordian sedimentation began at a water depth of about 80 m (see above), and ended in the intertidal zone. Basement subsidence caused by loading of the lithosphere with the total of Oxfordian sediments in the northwest is assumed to be about two-thirds the averaged compacted thickness of the sediments, as was concluded above for sequence 1.

11) Initial water depth in the northwest at the beginning of the Oxfordian ca — 80 m
12) Water depth at the end of the Oxfordian — 0 m
13) Average compacted thickness of Oxfordian sediments in the northwest ca — 310 m
14) Original (decompacted) thickness: Biolithites and grainstones with an aggregate thickness of about 110 m were not compacted in Oxfordian time. 200 m of siliciclastic and of carbonate mudstones with an average post-Oxfordian overburden of about 500 m had an original thickness of 270 m. Total original thickness: 110 + 270 m — 380 m
15) Basement subsidence under Oxfordian sediment load: About two-thirds the compacted sediment thickness of 13), cf. 3) and 9), or ca — 200 m
16) Relative sea level rise in northern Switzerland in the Oxfordian Age: 14) minus 11) minus 12) minus 15) — 100 m

The water depth near Schaffhausen at the end of the Oxfordian must then have been about 120 m:

6) Initial water depth at the beginning of the Oxfordian near Schaffhausen — 100 m
17) Thickness of compacted Oxfordian sediments near Schaffhausen ca — 120 m
18) Original (decompacted) thickness of these sediments (post-Oxfordian overburden ca 300 m, see GYGI & MCDOWELL 1970, p. 114) — 160 m
19) Subsidence of basement under the load of Oxfordian sediments: by about two-thirds the compacted sediment thickness (see above), or ca — 80 m
16) Relative sea level rise in the Oxfordian in northern Switzerland — 100 m
20) Water depth near Schaffhausen at the end of the Oxfordian: 6) plus 19) plus 16) minus 18) — 120 m

The base of the upper cross section B of Plate 1 in GYGI & PERSOZ (1986) was corrected accordingly.

2.6 *Estimate of the individual eustatic sea level rises of the Oxfordian*

There is ample qualitative evidence (HALLAM 1978) and wide agreement that a net eustatic sea level rise occurred in the Oxfordian. The main results of the semiquantitative calculations above are that endogenic subsidence of the lithosphere in the region was much less than basement subsidence under the load of sediments, and that the relative sea level rise in the Oxfordian Age in northern Switzerland was about 100 m. This figure was obtained from different and independent evidence. The individual steps in the Oxfordian eustatic sea level rise are represented in Figure 4. The semiquantitative eustatic sea level curve in Figure 4 was arrived at by the following qualitative, then quantitative reasoning:

2.6.1 *Summary of the qualitative evidence*

Some eustatic sea level rise must have occurred in the Early Oxfordian prior to the Cordatum Subchron, because the margin of the argillaceous mud bank of the Renggeri Member did not prograde. The sea level rise late in the Cordatum Subchron, at the turn from the Early to the Middle Oxfordian, was a major eustatic event, because a sequence boundary caused by the rise was recognized both in southern England and in northern Switzerland. The effect of the sea level rise of the Antecedens Subchron was described in northern Switzerland and in adjacent France, in England, and it can be observed in the southern part of extra-carpathian Poland. A eustatic sea level rise appears to be the only possible explanation for the coincidence of a major event in regions so far apart. The sea level rise of the Bifurcatus Chron indicated in Figure 4 is probably the cause for the "discontinuité no. 19" of GABILLY et al. (1985, p. 396). The rapid but small-scale sea level rise of the Hypselum Subchron was not reported from outside Switzerland. The eustatic sea level rise concluded in northern Switzerland of the Bimammatum Subchron was not discerned by TALBOT (1973) in southern England, but, as far as can be judged from Figure 12 by VAIL Et al. (1984), these authors considered the rise to be a major global event. Recommencement of sea level rise at the beginning of the Planula Subchron caused the formation of the slightly condensed Knollen Bed of GYGI (1969, p. 72) and of the Ammonitenlager of KOERNER (1963, Fig. 73) in adjacent southern Germany.

2.6.2. *Quantitative evidence*

The quantitative approach to the individual eustatic sea level rises of the Oxfordian is not made in stratigraphic order. Observations in northwestern Switzerland (see above) suggest that the rise in the Antecedens Subchron was the greatest and the one best suited for a quantitative consideration. Therefore, this eustatic sea level rise is estimated first. Earlier and later rises will be estimated by comparison with the rise of the Antecedens Subchron, as far as this is possible.

The rise in the Antecedens Subchron increased the water depth above the earlier part of the St-Ursanne carbonate platform sufficiently to make lush growth of coral reefs possible on top of an oolite shoal near St-Ursanne (Fig. 5), and further in the platform interior near Boncourt (GYGI & PERSOZ 1986, Pl. 1A). It is concluded from a comparison with Recent reefs in the lagoon of the Bermuda atoll (see GARRETT Et al. 1971, Fig. 4), that vigorous reef growth as occurred near St-Ursanne far in the platform interior required a water depth of at least 10 m to provide for adequate food supply and constant salinity. When it is assumed that the increment in water depth was 10 m near St-Ursanne in the late Antecedens Subchron, then the eustatic component of it was only a part of 10 m. A eustatic sea level rise induces the basement to subside by about 40% the rise under the additional water load (P. A. ZIEGLER 1982, p. 106), when the rate of sedimentation and of endogenic subsidence are neglected. Under these conditions, the eustatic sea level rise of the Antecedens Subchron would have been about 7 m. This estimate is regarded to be a minimum, because the increment in water depth near St-Ursanne was probably at least 10 m.

An earlier eustatic sea level rise occurred late in the Cordatum Subchron (Fig. 4). This rise initiated subsequence 1b of GYGI & PERSOZ (1986). Subsequence 1b is probably the equivalent of Talbot's cycle 2 (see above). TALBOT (1973, p. 313) estimated the eustatic sea level rise at the beginning of his cycle 2 at 10.5 m. This figure must be revised, because neither compaction nor loading with water were taken into account. When this is considered, the data presented by TALBOT (1973) indicate that the eustatic sea level rise of the late Cordatum Subchron was of the same order of magnitude as the rise in the Antecedens Subchron. The eustatic sea level rise that occurred in the Early Oxfordian prior to the Cordatum Subchron must have been gradual and cannot be estimated, but it had a noticeable effect on basinal sedimentation: the margin of the accumulating argillaceous mud bank of the Renggeri Member was kept stable or even caused to recede by the rise. There is no evidence of significant sea level falls in the first half of the Oxfordian. The net eustatic sea level rise from the beginning of the Oxfordian to the Parandieri Subchron is therefore estimated at no less than 15 m.

Eustatic sea level rise is most probably the process that stopped progradation of the Günsberg carbonate platform in the Bifurcatus Chron (subsequence 2a in Plate 1A by GYGI & PERSOZ 1986). A few meters of eustatic sea level rise in the Hypselum Subchron were sufficient to produce the observed effects on sedimentation in the platform facies of subsequence 2b of GYGI & PERSOZ (1986, Pl. 1A). A minimum of 5 m of eustatic sea level rise is likely for sequence 2 in Figure 4. The rise in the Bimammatum Subchron that initiated sequence 3 influenced sedimentation not only on the platform, but in the shallow part of the "basin" as well. Nevertheless, the amount of the rise cannot be estimated directly. The eustatic sea level rise that caused the slight condensation of the Knollen Beds of northern Switzerland (base of subsequence 3b) or of the Ammonitenlager in adjacent Germany just before the beginning of the Planula Subchron had a lesser effect on sedimentation in the shallow part of the "basin" than the rise of the Bimammatum Subchron. But there is reason to assume that sea level continued to rise eustatically throughout the Planula and the Galar Subchrons, and that the amount of this gradual rise exceeded the rise of the Bimammatum Subchron (see Fig. 4).

A substantial eustatic sea level rise during deposition of sequence 3 must be concluded from the geometry and from the sedimentology of the Holzflue Member (see GYGI & PERSOZ 1986, Pl. 1A), and from the shifting of the margin of the carbonate platform during deposition of sequence 3. At this time, the platform margin receded in the proximal direction (Fig. 3C and D). Sedimentation of the Holzflue Member began above the oolite shoal of the Steinibach Beds. This was possible only if relative sea level rise occurred. It can be read from sections measured by GYGI (1969, Pl. 18) that the Holzflue Member between Olten and Balsthal is a wedge. The mineralostratigraphic correlations I, J, and K made by GYGI & PERSOZ (1986) suggest that the difference in thickness within the member developed mainly in the Panula and in the Galar Subchrons. At the end of the Oxfordian, the upper surface of the wedge was a slope. This is indicated by the oolite facies near Balsthal and by the carbonate mudstone facies near Olten. The cause for the development of such a slope above the flat surface of a carbonate platform is most probably a eustatic sea level rise which was gradual and of a considerable amount. Development of small coral bioherms far in the interior of the Verena oolite shoal, above oncolitic horizons from a lagoonal environment (GYGI & PERSOZ 1986, Pl. 1A), is another indication of substantial eustatic sea level rise near the end of the Oxfordian.

It is evident from correlation J of GYGI & PERSOZ (1986, Pl. 1A) that the onset of sedimentation of the oolitic Verena Member almost coincided with the beginning of the Planula Subchron. Ooid sand formation began in the shallow subtidal or in the lower intertidal zone, this is to say at a negligible water depth. Deposition of the Verena Member ended in the intertidal zone (GYGI & PERSOZ 1986). The mean thickness of the Verena Member above correlation J is close to 40 m (GYGI & PERSOZ 1986, Pl. 1A). This thickness is the effect of endogenic subsidence, compaction of mud of the Effingen Member below, and of eustatically rising sea level. When it is postulated that endogenic subsidence was constant in the Oxfordian, and that ammonite subchrons were of equal duration, then sedimentation of the Verena oolite member of subsequence 3b can be directly compared with sedimentation of the upper part of the St-Ursanne Formation of subsequence 1c. Subsequence 1c, with a mean thickness of 35 m of ooid sand, coral reefs, and inter-reef sediment, was deposited in the course of somewhat more than one subchron (Fig. 4). Subsequence 3b, with a mean thickness of close to 40 m of ooid sand, was laid down in two subchrons. Endogenic subsidence must therefore have had a larger share in creating room for sedimentation of subsequence 3b than in subsequence 1c. Initiation and the thickness of subsequence 1c were mainly the effect of about 7 m of eustatic sea level rise. Bearing endogenic subsidence in mind, and that the Galar Subchron possibly represents less time than the Planula Subchron (Fig. 4), then a eustatic sea level rise of about 5 m must be assumed for the time of the Planula and of the Galar Subchrons. This is a minimum, because post-Oxfordian compaction of the ooid sand of the Verena Member was not quantified and taken into account.

2.6.3 Total of eustatic sea level rises in the Oxfordian

The semi-quantitative Oxfordian seal level curve of Figure 4 was obtained from a comparison of the eustatic sea level rise of the Antecedens Subchron with earlier and later rises. The 7 m of eustatic sea level rise in the Antecedens Subchron is only a coarse estimate. The error of this estimate influences all other estimates. If the figures of eustatic sea level rise given above are summed up in spite of their inaccuracy, and if the possibility of minor eustatic sea level falls in the Cordatum Subchron and in the Hypselum Subchron is taken into account, then it may be concluded that the eustatic sea level changes in the Oxfordian Age resulted in a net rise which is estimated at 25 m to 30 m. No indications of substantial eustatic sea level falls in Oxfordian time were found as were suggested by HALLAM (1978, Fig. 11).

2.6.4 Remarks on the numeric time scale of Figure 4

Radiometric ages of 145 and 146 ± 3 m.y. were obtained by GYGI & MCDOWELL (1970, Table 1) from authigenic glauconites of the Glaukonitsandmergel Bed (see GYGI & MARCHAND 1982, Fig. 3) of the Early Oxfordian. The average age of the three samples measured of this bed was later recalculated to be 148 ± 3 m.y. (ODIN 1982, p. 818, NDS 141). H. Fischer of the Federal Institute of Technology at Zürich, Switzerland, measured the glauconite sample RG 81/11 from Gächlingen and the samples RG 207/14a and RG

212/5 from Siblingen, which were collected from the Glaukonitsandmergel Bed in 1970 and 1971. H. Fischer (oral communication) confirmed the recalculated ages published by ODIN (1982). In 1970, there was some uncertainty in the correlation of the radiometric ages with the biochronologic ammonite scale, because at that time, few diagnostic ammonites were available. Numerous and well-preserved ammonites could be collected from the Glaukonitsandmergel Bed in systematic excavations made since 1970. Cardioceratids from these excavations, diagnostic of ammonite zones and subzones from the latest Callovian to the Middle Oxfordian, are figured in GYGI & MARCHAND (1982). The specimens figured from the Glaukonitsandmergel Bed indicate that this bed was deposited in the Cordatum Subchron alone (see Fig. 4, this paper).

Two glauconite samples from the lower Baden Member of the early Kimmeridgian were radiometrically dated at 134 and 137 ± 3 m. y., respectively, by GYGI & MCDOWELL (1970, Table 1). The mean recalculated age is 138.5 ± 3 m. y. (ODIN 1982, p. 819). Several hundred ammonites collected since 1970 from the lower Baden Member of Mellikon (see section no 70 in GYGI 1969, Pl. 17) indicate that the radiometrically dated glauconites of the early Kimmeridgian are of the Hypselocyclum Chron (Fig. 4).

The accuracy of the apparent radiometric ages given above cannot be judged, because we know of no comparable work on sediments the same age. Many authors feel that potassium-argon ages of intrusive rocks are more reliable than those of glauconites. However, the numeric ages of intrusive rocks cannot normally be related directly and accurately to the biochronologic time scale. The Fanos granite in northern Greece mapped by MERCIER and dated by BORSI et al. (1966) at 150 m.y. may be an exception. BORSI et al. (1966, p. 285) had reason to assume that the ophiolite complex intruded by the granite was emplaced in the early Tithonian. The Fanos granite is slightly younger, not of Kimmeridgian age as stated by ARMSTRONG (1978, p. 85, items 480 and 481). This incongruence is only nomenclatorial: the Kimmeridgian in the sense of ARKELL as adopted by ARMSTRONG (1978) includes the early part of the Tithonian of recent authors (or of the Portlandian as conceived by French authors). If the apparent age of the Fanos granite proved to be accurate, and if the correlation of this age with the early Tithonian as concluded by BORSI et al. (1966) could be ascertained, then the radiometric ages given in Figure 4 (this paper) would be far too young. However, according to BORSI et al. (1966, p. 282), the western contact of the granite is a tectonic thrust plane. No primary contact between the Fanos granite and biochronologically dated sediments was found. Under the circumstances, the Oxfordian and the Kimmeridgian ages proposed by ODIN (1982) are to be preferred.

3. Fossil assemblages and water depth

The water depth probably never exceeded 150 m in the Oxfordian or in the early Kimmeridgian in northern Switzerland. The sediments of these ages are therefore all from more or less shallow water. There is an obvious relation between depth and the composition of the macrofauna even though the differences in water depth were slight (Fig. 6B).

3.1 Corals

Colonial corals occurred only in shallow water at the margin or in the interior of carbonate platforms. They took part in accumulating bioherms. Coral bioherms vary from mud mounds with less than 10% of the volume being corals at the margin of the St-Ursanne carbonate platform at Sornetan to what were probably reefs with as much as 50% corals in the platform interior at St-Ursanne. The coral bioherms that could be discerned in the Oxfordian and in the Kimmeridgian are all isolated buildups (Fig. 5). The lateral diameter of the bioherms is between less than 10 m and several tens of meters. The tallest bioherms do not exceed 30 m from base to top. The slope from the margin of the bioherms or reefs to the surface of the adjacent flat-lying sediment was gentle at every growth stage (PÜMPIN 1965, Fig. 14). The steep sides of the lagoon reef represented in Figure 5 are not primary. They are the effect of differential resistance to weathering of the non-porous reef core and the porous, friable inter-reef sediment. Sponge bioherms weather out similarly with steep or vertical sides. Coral bioherms resembled sponge bioherms in growth form and in size. The principal difference between the two bioherm types is that sponge bioherms grew at a water depth which was substantially greater than that of coral bioherms (Fig. 2B). The coral bioherms of the Günsberg Member typically started in soft argillaceous mud and were covered by ooid sand. The bioherms of this member may be laterally very close together, and some are stacked on top of each other (GYGI & PERSOZ 1986, Pl. 1A). Coalescence of individual bioherms into a barrier reef was not observed. Ammonites are always rare in the coral bioherm environment.

3.2 Bivalves

Bivalves are dominant off the bank margin from a depth of about 20 to less than 50 m. They are the most common fossils in Kimmeridgian sediments near Olten which were deposited at a depth of about 30 m (Fig. 6B, cf. GYGI & PERSOZ 1986, Pl. 1B). The very abundant bivalves of the Banné marl lived in a lagoonal environment with argillaceous mud sedimentation from nutrient-rich water. The Banné marl contrasts with the very pure lagoonal carbonate-mud deposits of the upper St-Ursanne Formation which were sedimented at a similar depth. In the carbonate mud, bivalves were rare probably because of normally clear water with a low nutrient content.

3.3 Sponges

Sponges secreting skeletal elements of opaline silica occur in the Recent from the coral reef environment to great depth (see RÜTZLER 1978, and RÜTZLER & MACINTYRE 1978, in GYGI & PERSOZ 1986). In the biocoenosis of the Oxfordian, sponges must have been common from a depth of 20 m or less. Their state of preservation varies greatly with paleodepth. Only spiculae of siliceous sponges, replaced by calcite, are to be found with hermatypic corals (GYGI 1969, Pl. 7, Fig. 29 and 28). Complete fossils of siliceous sponges were formed from a minimum depth of more than 50 m, as for instance in the Wettingen Member of Villigen (GYGI & PERSOZ 1986, Pl. 1B). Even in sediments from relatively deep water like the Birmenstorf Member (Fig. 2A), it appears that in most cases, only irregular lumps (tuberoids) of varying shape and size survived of this quantitatively important fossil group. These tuberoids were formed by partial fossilization of siliceous sponges and adhering algal crusts, without intervention of mechanical breakdown. When the rate of sedimentation was very low, at a depth where sponge fossilization was otherwise common, complete fossils of sponges were not formed: few and incomplete remains of

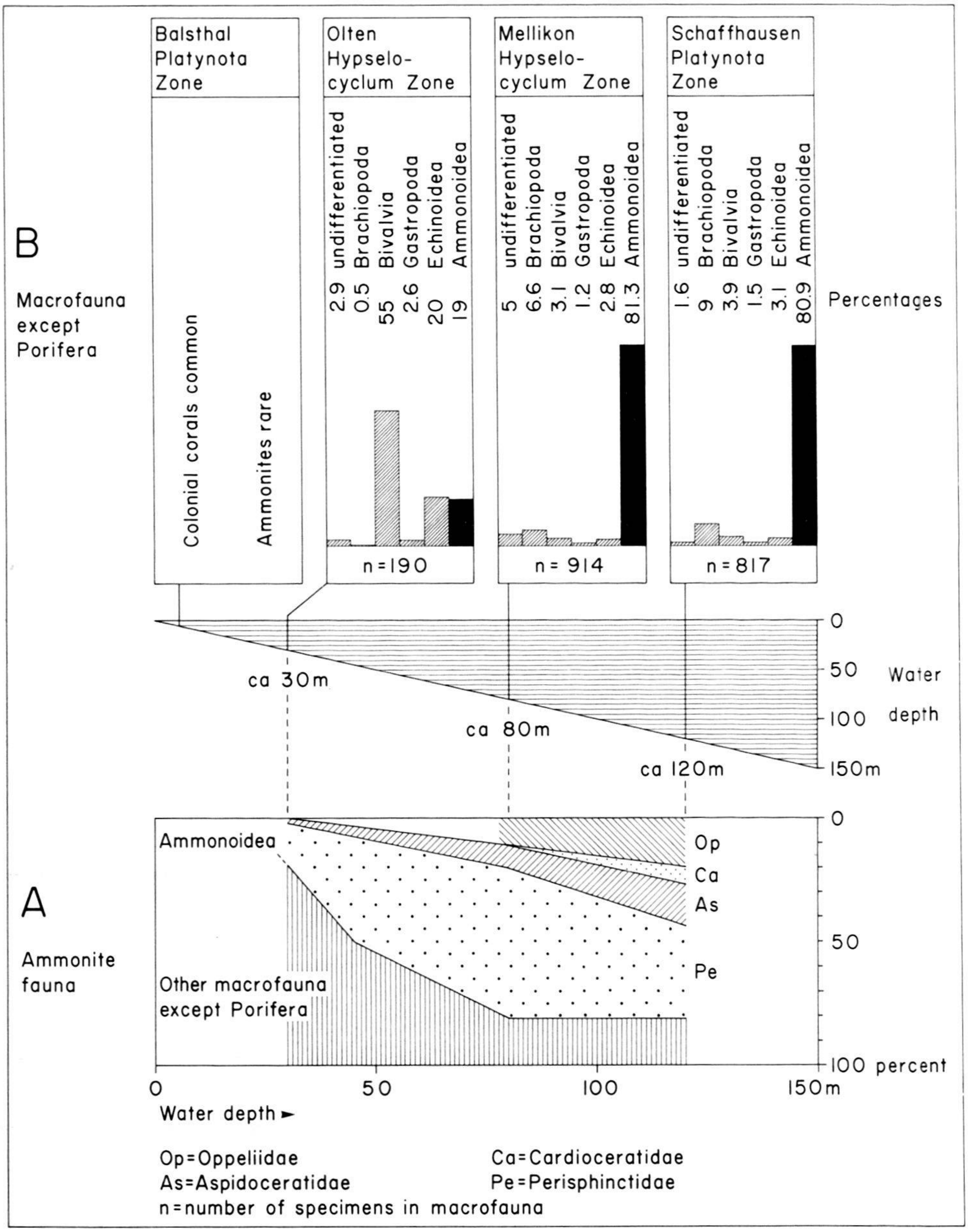

Fig. 6. A: Variation in the abundance and in the composition of the ammonite fauna of the early Kimmeridgian with increasing water depth. B: Variation in the composition of the whole macrofauna of the early Kimmeridgian with increasing depth. Porifera, a quantitatively important element of the macrofauna in water deeper than 100 m, are not represented for reasons given in the text.

sponges were found in the Schellenbrücke Bed of the Early Oxfordian, where entire and well-preserved fossils of ammonites are abundant (GYGI 1981, p. 239). Siliceous sponges can therefore not be used in a quantitative interpretation of a taphocoenosis. Sponge bioherms probably developed from a minimum depth of about 80 m (Crenularis Member and Knollen Beds of Mellikon, GYGI & PERSOZ 1986, Pl. 1A). Where tuberoids or whole fossils of sponges are abundant, there are many ammonites.

3.4 Ammonites

Ammonites are the dominant element of the macrofauna in sediments deposited from a certain minimum depth down. Ammonites are 19% the macrofauna in Kimmeridgian sediments of the Hypselocyclum Zone near Olten (Fig. 6B), which were deposited at a depth of about 30 m, to judge from GYGI & PERSOZ (1986, Pl. 1B). 89% of these ammonites are perisphinctids. Ammonites of the Kimmeridgian Hypselocyclum Zone make up 81% of the macrofauna of the glauconitic Baden Member of Mellikon (Fig. 6B), which was deposited at a depth of about 80 m (GYGI & PERSOZ 1986, Pl. 1B). Perisphinctids are 75% of this ammonite fauna (Fig. 6A). There is a rich macrofauna of mainly the Platynota Zone in the lower Schwarzbach Member near Schaffhausen. This member is the same age as the Baden Member, but it was sedimented in water about 120 m deep (Plate 1B in GYGI & PERSOZ 1986). 81% of this macrofauna are ammonites. 46% of the ammonites are perisphinctids, and 25% are oppeliids (Fig. 6A). Consequently, the percentage of ammonites in the macrofauna grows with increasing depth, and there is a concomitant change in the composition of the ammonite fauna, as was concluded before by B. ZIEGLER (1967, 1971).

The depth at which cephalopods begin to predominate in the macrofauna is between 30 and 80 m. This can be bracketed further by calculating the depth of deposition of the Crenularis Member (Late Oxfordian) as well as of the Baden Member (early Kimmeridgian, see GYGI & PERSOZ 1986) in the area Aarau–Auenstein in canton Aargau, and by comparing the macrofauna of the two members. In the Crenularis Member (Bimammatum Subchron) at the base of sequence 3, the macrofauna is mostly bivalves. The depth of deposition of the member was calculated from the:

21) Initial water depth at the beginning of the Oxfordian between Aarau and Auenstein (see Fig. 3A) ca 90 m
22) Compacted thickness of Oxfordian sediments below the Crenularis Member (from GYGI & PERSOZ 1986, Pl. 1A) 220 m
23) Load-induced subsidence of pre-Oxfordian basement: by two-thirds the compacted sediment thickness of 22), or ca 150 m
24) Relative sea level rise from the beginning of the Oxfordian until the Bimammatum Subchron, from 16) and Figure 4, more than 70 m
25) Original or decompacted thickness of 22), as read from PERRIER & QUIBLIER (1974, Fig. 11) 280 m
26) Depth of deposition of lower Crenularis Member: 21) plus 23) plus 24) minus 25), equivalent to more than 30 m

The water depth at the end of the Oxfordian in the area Aarau–Auenstein can be calculated in the same way as above to be 50 m. In the Kimmeridgian Baden Member directly above the Oxfordian, ammonites prevail near Möriken, a village 4 km east of Auenstein. The Baden Member of Möriken is assumed to have been laid down at a water depth greater than 50 m, because the rate of sedimentation was low in the area, like near Villigen and Mellikon (compare GYGI 1969, Pl. 17, sections no 62 and 70, with GYGI & PERSOZ 1986, Table 2), and because the transgression which occurred between Olten and Balsthal in the early Kimmeridgian (see below, and GYGI & PERSOZ 1986, Pl. 1B) suggests that a rapid eustatic sea level rise occurred at this time. The water depth therefore increased in the area from the Late Oxfordian through the early Kimmeridgian, and cephalopods began to prevail when the water became more than 50 m deep in the early Kimmeridgian.

The depth at which ammonites became the dominant element of the macrofauna may have been slightly less than 50 m. This is suggested by the fossil-rich horizon with mostly ammonites in the distal part of the Terrain à Chailles Member (GYGI & PERSOZ 1986, Table 2). The horizon coincides with biostratigraphic correlation II as represented in Plate 1A by GYGI & PERSOZ (1986). The fossil-bed is somewhat more than halfway up a shallowing-upward succession, the deposition of which began at a depth of about 80 m (Fig. 3A) and ended in the shallow subtidal zone (at the level of the mineralostratigraphic correlation A in Plate 1A of GYGI & PERSOZ 1986). The depth of deposition of the proximal part of the fossil-bed is estimated at more than 40 m, because sedimentation of the bed began with the eustatic sea level rise of the late Cordatum Subchron (Fig. 4), which increased the water depth by an estimated total of nearly 10 m.

The percentage of ammonites in the early Kimmeridgian macrofauna increases to a water depth of about 80 m, then it seems to remain more or less constant beyond this depth (Fig. 6). On the other hand, the composition of the ammonite fauna continues to change when the water depth increases beyond 80 m: the percentage of opeliids grows with increasing water depth, whereas the perisphinctid percentage further decreases in deeper water (Fig. 6A). This may be concluded from the early Kimmeridgian Baden Member of canton Aargau and from the coeval Schwarzbach Member of canton Schaffhausen, where the difference in paleodepth is much greater than between the Early Oxfordian Schellenbrücke Bed and the Glaukonitsandmergel Bed.

To judge from the changes in the ammonite fauna with paleodepth in the early Kimmeridgian sediments, the vertical increase of oppeliid percentages from the Early Oxfordian Schellenbrücke Bed upward, and upward from the coeval Glaukonitsandmergel Bed (Table 1), reflects deepening of the water in a time of strongly reduced sedimentation in a starved basin. Accordingly, one would expect the perisphinctid percentage of the Schellenbrücke Bed to decrease vertically to the Birmenstorf Member which was deposited in deeper water. In fact, perisphinctids in the Birmenstorf Member were found to be slightly more abundant than in the Schellenbrücke Bed (Table 1). It is evident from the early Kimmeridgian sediments that perisphinctids of this age preferred relatively shallow water as compared with other ammonites (Fig. 6A). Cardioceratid ammonites of the Early Oxfordian preferred a water depth interval which is similar to that of early Kimmeridgian perisphinctids (GYGI & MARCHAND 1982, p. 531). In the Schellenbrücke Bed, cardioceratids and perisphinctids are almost equally abundant (Table 1). In the Birmenstorf Member, cardioceratids are rare mainly for reasons other than water depth: *Cardioceras* died out at the end of the Antecedens Subchron, and the earliest *Amoeboceras* descendants were rare in the Parandieri Subchron of the Transversarium Chron (cf.

"*Cardioceras* gap" of MALINOWSKA 1980). The general rarity of cardioceratids in the Transversarium Chron as caused by evolution was further enhanced in our region by a probable warming of the climate, which caused the boreal cardioceratids to retreat to the north since the beginning of the Antecedens Subchron. The aggregate cardioceratid and perisphinctid percentage in the Birmenstorf Member is much less than in the Schellenbrücke Bed below (Table 1). The slightly greater percentage of perisphinctids in the Birmenstorf Member with respect to the Schellenbrücke Bed may be a consequence of the near-disappearance of the ecologically similar cardioceratids in the Birmenstorf Member, on the assumption that perisphinctid ecology did not substantially change from the Early Oxfordian to the early Kimmeridgian. We know of no indications of such a change.

3.5 Conclusions about fossil assemblages

The bathymetric profile at the end of the Oxfordian being established, it is possible to define the paleodepth intervals of early Kimmeridgian fossil assemblages with greater accuracy than could be attained previously. Hermatypic corals were very abundant in shallow water to a depth of about 20 m. From 20 to more than 40 m, bivalves were most common. From a paleodepth of somewhat less than 50 m, ammonites are dominant in fossil assemblages. The composition of the whole macrofauna provides detailed paleobathymetric information to a depth of about 50 m. Greater depth may be concluded from the composition of the ammonite fauna (Fig. 6). Criteria for this must be elaborated separately for any interval of time (see discussion of Schellenbrücke Bed and Birmenstorf Member). Shallowing-upward of a sedimentary succession can also be well documented by fossil assemblages, as for instance in sequence 1 in the lower part of the Oxfordian in northwestern Switzerland. Macrofossil assemblages are important, because they are the only source of detailed paleobathymetric information in the deep subtidal zone down to at least 150 m, where sediments are almost exclusively mud-grade and normally include little evidence of water depth. Our correlation of water depth intervals with particular macrofossil assemblages is remarkably close to the results arrived at by B. ZIEGLER (1967, 1971) on independent lines of evidence.

Time correlations in the Late Jurassic formations of Central Europe are best made with perisphinctids where they occur, because this ammonite group alone ranged from the deeper marine ammonite facies to the shallow-water coral-oolite facies. Perisphinctids are the only ammonites which were found to date between coral bioherms or in calcareous oolite of the Late Jurassic in northern Switzerland.

4. The influence of changes in sea level and climate on sedimentation

4.1 Sea level changes and depositional sequences

The cephalopod-dominated macrofauna and the iron ooids in the thin ferruginous marl-clay at the base of the Renggeri Member suggest that sedimentation of sequence 1 began in a pre-existing "basin" that was between 80 and 100 m deep. The margin of the growing mud bank did not prograde in the course of shoaling except in the Cordatum Subchron. Development of the carbonate platform of the St-Ursanne Formation began when accumulation of the argillaceous mud bank had raised the sediment surface to within about 10 m of mean sea level (see Gygi & Persoz 1986). The St-Ursanne Formation is the depositional continuation in carbonate facies of the underlying argillaceous mud bank, with a transitional boundary between marly sediments below and pure limestones above. The margin of the early part of the carbonate platform of the St-Ursanne Formation therefore coincides with the upper margin of the argillaceous mud bank below (see above). The margin of the shallow-water realm remained more or less stationary until most of the carbonate platform of the St-Ursanne Formation was deposited. No sediments from the supratidal environment were found within the St-Ursanne Formation. There was little or no sedimentation, and the water depth increased in the "basin" when the argillaceous mud bank and the carbonate platform of sequence 1 were deposited. The water depth in the "basin" increased until the late Transversarium Chron. Sequence 1 was most probably sedimented in a time of eustatically rising sea level even though the sequence records shallowing-upward where it is complete (compare Fig. 3A with B).

Regression prevailed when sequence 2 was deposited. A stillstand or a fall of sea level and a strong supply of siliciclastic mud and silt from land led to widespread development of peritidal facies in the Natica Member and to rapid progradation of the Günsberg Member. There is no simple relation between the rate of sediment supply and regression. The continuing strong supply of terrigenous sediment and more or less constant sea level in the Bifurcatus Chron caused the peritidal facies of the Natica Member to advance seaward beyond the margin of the St-Ursanne carbonate platform below. If sea level did fall in the Hypselum Subchron, then part of the Effingen Member must be reworked sediment of the earlier Oxfordian. It is apparent from the ammonite succession and from the mineralostratigraphic correlations E and F in the Effingen Member, and from the stenohaline hermatypic corals and echinoderms in the uppermost Natica Member above peritidal facies (see Gygi & Persoz 1986, Pl. 1A), that regression of the upper Natica Member reversed to transgression in the Hypselum Subchron (base of subsequence 2b) in spite of a continuing strong supply of terrigenous sediment. This transgression was very probably the effect of a small-scale but rapid eustatic sea level rise. Transgression soon turned back to regression in the late Hypselum Subchron when the margin of the Steinibach oolite shoal (upper part of subsequence 2b) prograded more than 20 km in less than one ammonite subchron. This progradation began when the rate of supply of terrigenous sediment dropped to a low value.

Sequence 3 was sedimented in a time of rising sea level and of strongly reduced terrigenous sediment supply. Formation of the carbonate shoal of the Holzflue/Verena Member and of the shallow basin to the northwest of the shoal initiated a paleogeographic pattern which lasted until the end of the Late Jurassic. Continuing sea level rise at the beginning of the Kimmeridgian caused hermatypic corals to retreat from near Olten to Balsthal, where small coral bioherms developed when sequence 4 was sedimented (Gygi & Persoz 1986, Pl. 1B). This documents transgression. On the other hand, the carbonate mud flat of the lowermost Reuchenette Formation near Moutier was probably accessible to large land animals (Gygi & Persoz 1986). The latter is a case of regression in a time of eustatically rising sea level, brought about by a high rate of more or less autochthonous production of carbonate sediment.

There is qualitative evidence that eustatic sea level changes were the principal cause of transgression and regression in the area. Quantification of small-scale sea level changes is uncertain in a regional study. We were only able to give a gross minimum estimate of the net eustatic sea level rise in Oxfordian time.

It may be concluded from subsequence 2b and from the mineralostratigraphic correlations H and I that transgression and regression over distances of several tens of kilometers alternated in a fraction of an ammonite subchron over a wide coastal plain (GYGI & PERSOZ 1986, Table 2 and Pl. 1A). A transgression pushes the source of terrigenous sediment back with respect to a given area on the adjacent "basin" floor. The effect of this can be expected to be in most cases a reduced sedimentation rate (starving) or non-deposition in the "basin" (see above).

The subsequence 1a of GYGI & PERSOZ (1986, Pl. 1A) is an example of this. Deposition of the subsequence ended in the Cordatum Subchron. This is documented by a great number of ammonites represented as biostratigraphic correlation II in Plate 1A by GYGI & PERSOZ (1986). Termination of subsequence 1a was the effect of a change from constant or slightly falling sea level to a rapid and strong eustatic sea level rise (Fig. 4). The upper boundary of subsequence 1a cannot be discerned in the Terrain à Chailles Member in neither of the two clay pits near Liesberg, because in the proximal part of the Terrain à Chailles Member, normal sedimentation kept pace with the sea level rise. In the distal part of the Terrain à Chailles Member near Bärschwil, Vellerat, or Sornetan, the sea level rise brought about slight condensation as indicated by unusually abundant macrofossils at the boundary between subsequence 1a and 1b (GYGI & PERSOZ 1986, Table 2 and Pl. 1A). Further out in the "basin", a hiatus developed above the upper boundary of subsequence 1a (GYGI & MARCHAND 1982, Fig. 4, or GYGI & PERSOZ 1986, Table 2).

According to the ammonites of the Cordatum Subchron in the uppermost part of subsequence 1a (biostratigraphic correlation II), the hiatus above subsequence 1a in the "basin" was initiated synchroneously relative to the biochronologic ammonite scale, from south of Sornetan to Siblingen. This is equivalent to a distance of about 80 km when projected on a line of maximum facies change perpendicular to the depositional strike (GYGI & PERSOZ 1986, Pl. 1A).

The mineralostratigraphic correlations B and I (GYGI & PERSOZ 1986, Pl. 1A) are parallel and close to the upper boundary of subsequence 1b and of sequence 2, respectively. Correlation C is subparallel to the upper boundary of sequence 1 because of differential rates of sedimentation: beginning regression reduced the sedimentation rate on the carbonate platform, while deposition on the slope and in the proximal part of the "basin" increased. Correlation L is subparallel to the upper boundary of sequence 3 due to rising sea level, which led to normal sedimentation on the carbonate platform in lowermost sequence 4, and to starving on the lower slope (GYGI & PERSOZ 1986, Pl. 1B).

The average time required for the transport of clay minerals from the source (primary or secondary) to the area of ultimate deposition is thought to be negligible as compared with the calculated sedimentation rates. If this is so, then sequence or subsequence boundaries are quasi-isochronous, provided they can be precisely discerned and correlated. Detailed time-stratigraphic comparison of our sequence boundaries with those differentiated by VAIL et al. (1984, Fig. 2) is not possible, because the position of Vail's sequence boundaries in relation to the ammonite scale is not known.

4.2 Variation in the siliciclastic sediment supply and in paleobathymetry

Siliciclastic mud and silt apparently came mainly from the Rhenish Massif in the north (GYGI & PERSOZ 1986). In northern Switzerland, the sedimentation rate of terrigenous sediment varied with time. A maximum occurred in the Early Oxfordian, and another in the late Middle Oxfordian. Minima were in the Parandieri Subchron, in the early Bimammatum Subchron, and in the Planula Chron (see GYGI & PERSOZ 1986, Table 2 and Pl. 1A). The change in the sedimentation rate of mainly siliciclastic matter is well documented by the biostromes of siliceous sponges of the lowermost Effingen Member in the "basin" of canton Solothurn and of canton Aargau. When progradation began early in the Bifurcatus Chron, because of an increased rate of terrigenous sediment supply, the biostromes were suffocated as soon as the belt of heavy argillaceous sedimentation reached them (see GYGI 1969).

The carbonate content has a general tendency to increase from the base to the top of a depositional sequence. This was observed in most of the subsequences described by GYGI & PERSOZ (1986). However, there are significant differences between individual subsequences in a given environment. In the proximal shallow-water realm, the lower part of subsequence 1c is a very pure limestone laid down after a major eustatic sea level rise (this paper, Fig. 4), whereas the lower part of subsequence 3a is a marl in the same environment (GYGI & PERSOZ 1986, Pl. 1A). The eustatic sea level rises that initiated the subsequences 1c and 3a were of comparable size, and they led to a similar increment in water depth. Therefore, the difference in the content of siliciclastic matter between the two subsequences in the same environment cannot be caused by a difference in the amount of the two eustatic sea level changes. The unequal content of siliciclastic matter in the two subsequences is an indication that the rate of supply of siliciclastic sediment was not controlled by eustasy alone. The climate was probably another important factor (see below).

Ferriferous ooids and glauconite in the sediments investigated by GYGI & PERSOZ (1986) indicate a particularly low rate of deposition. The vertical transition from the iron-oolitic sediments of the Callovian to the glauconitic facies of the Oxfordian in canton Schaffhausen, or from the iron-oolitic Schellenbrücke Bed to the glauconitic Birmenstorf Member in canton Aargau, documents deepening of the water beyond 100 m (see above). The bathymetric change cannot be the only cause for this facies transition, because the sediments of the earliest Kimmeridgian are glauconitic both in canton Schaffhausen and in canton Aargau, notwithstanding that the depth of deposition was about 120 m near Schaffhausen, around 80 m at Mellikon in canton Aargau, and even less further west (Fig. 6).

4.3 Siliciclastic sediment supply and carbonate production

The small carbonate platform of the lower Günsberg Member developed when the supply of siliciclastic mud and silt from land was at a maximum. A large amount of the fine-grained terrigenous sediment must have bypassed the ooid sand shoals and the coral bioherms in suspension, either regularly through tidal channels, or occasionally as a sheet during storms. The only traces left of this within the platform are thin marly intercalations between ooid sand bodies and pockets of marl within coral bioherms. The net sedimentation rate in the realm of the Natica Member was low in spite of the ample supply of terrigenous sediment, and in spite of the high carbonate production rate on the platform at the same time. Most of the terrigenous sediment and some of the fine-grained carbonates produced on the shoal of the Günsberg Member were swept into the "basin" and accumulated the sigmoid, mixed siliciclastic-carbonate sediment bodies of the Effingen Member. A marked lateral gradient of the carbonate content exists between the lower Günsberg Member and the proximal part of the coeval Natica Member, or the Effingen Member, respectively. The corresponding lateral gradient of the carbonate content between the lagoon, the platform margin, and the "basin" is minimal in subsequence 3b. This is another indication that the supply of terrigenous sediment varied at the source, as caused by variations in the climate.

The carbonate content of the three Oxfordian sequences generally increases upward, but there is great variation in this. The upper, mostly "pure" carbonate part is least in sequence 2 of canton Solothurn, and it is greatest in sequence 3. The low carbonate content in the lower part of sequence 1 is rather homogenous as compared with the lower part of sequence 2 in canton Aargau where the carbonate content is highly variable (GYGI 1969, Pl. 17, section RG 37).

4.4 Sediment transport

The sedimentology and the very gentle slope of detritus aprons issuing from bioherms near St-Ursanne and near Péry (Fig. 7) indicate that sediment transport by storms may have been quantitatively important in shallow water. The depositional dip of these beds is of the order of 5° (PÜMPIN 1965, Fig. 14). BRENNER et al. (1985, p. 369) concluded that in a comparable Oxfordian shallow-water environment, storms may have transported at least as much sediment as tidal currents. Storm layers (tempestites) from a paleodepth approaching 100 m were found in the Schellenbrücke Bed (GYGI 1981, p. 242) and in the Lamberti Bed below (GYGI & MARCHAND 1982, p. 525). There is reason to assume that episodic transport of mud-grade sediment by storms may have been quantitatively important even in the more distal part of the "basin".

GYGI (1969, p. 107) presented evidence from micropaleontology that the water above the surface of some sediments of the Effingen Member in the "basin" near Auenstein was at times stagnant and anoxic, and that at other times, the water had a low oxygen content. Therefore, the mixed siliciclastic-carbonate mud of the Effingen Member cannot have been dispersed on the "basin" floor by continuous currents, or else all the sediments of the Effingen Member would have been deposited in well-aerated water. More or less continuous sedimentation of homogenous mud out of suspension from slowly moving water can only have been important in the marginal part of the "basin" where dense, sediment-laden water with free oxygen descended the slope. Under normal conditions, most or all of this oxygen was consumed in the stagnant bottom water of the "basin". A mechanism to transport fine-grained sediment on a large scale across the floor of the epicontinental basin of the Effingen Member appears to be currents driven by exceptionally strong storms occurring at long intervals. Such storms would not only have stirred up and transported fine-grained sediment. They must have mixed anoxic bottom water with aerated water from near the surface. The result was the slightly oxygenated water above some of the sediments as was concluded from micropaleontologic evidence. The free oxygen mixed to the bottom water during a storm would be partially or totally consumed

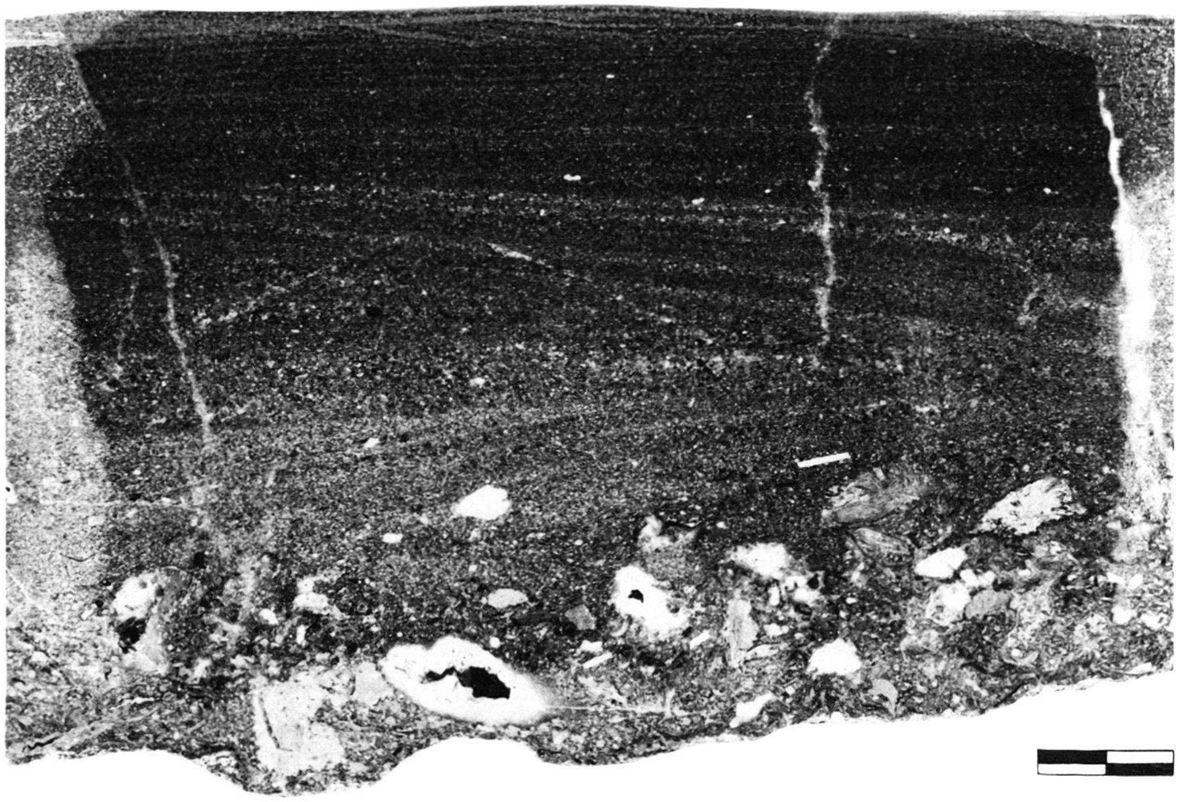

Fig. 7. Storm deposit (tempestite) adjacent to a coral bioherm. Coarse-grained coral detritus with whole brachiopods at the base of the bed. Above: fine-grained and cross-laminated, bioclastic pelmicrite to pelsparite with mostly silt-grade detrital quartz. Uppermost Effingen Member, bed no 160e of section RG 307, La Charuque quarry, Péry BE (see GYGI & PERSOZ 1986, Fig. 1).

in the quiet interval before the next storm strong enough to affect the bottom occurred. There is growing evidence from the Recent that currents driven by storms are the principal mechanism of sediment transport in deeper water of continental shelves (see for instance VINCENT et al. 1981, or DRAKE et al. 1985).

Thin debris flows are common mainly in the upper Effingen Member (GYGI 1969, Pl. 4, Fig. 13, and GYGI & PERSOZ 1986, Fig. 3) and in the Geissberg Member even though the average depositional slope was minimal. We know as yet of no means to discriminate with certainty between distal turbidites derived from debris flows and flat laminated storm deposits in the Effingen Member. Some of the quartz-rich, laminated pelsparites of the Effingen Member, that BOLLIGER & BURRI (1970) regarded to be eolian deposits, are certainly tempestites, because the abundance and the thickness of these horizons rapidly increase upward concomitant with shoaling in the uppermost Effingen Member at Péry.

4.5 Climate

A diverse assemblage of hermatypic corals indicates that the Late Jurassic climate of the region was at least subtropically warm. The presence of thin coal seams, limnic ostracods, and of characean oogonia in marls of the Natica Member is evidence that there was enough rainfall to produce swamps and freshwater pools in the supratidal environment. Limnic ostracods are also known from a younger horizon in the lower Reuchenette Formation near Solothurn (see GYGI & PERSOZ 1986). Evaporite deposits are primarily absent from the marginal marine facies of the Natica Member or of the Reuchenette Formation. Relatively abundant kaolinite in Oxfordian sediments suggests that there were seasons with substantial rainfall supporting a vegetation in the source area not far to the north. Rainfall must have been sufficient to support a fairly thick growth of land plants in northern Switzerland, because the remains of a specimen of *Cetiosauriscus greppini*, a large, plant-eating dinosaur weighing several tons, were found in a tidal carbonate mudflat paleoenvironment of the lowermost Reuchenette Formation near Moutier (see GYGI & PERSOZ 1986). It is likely that this land animal was fossilized in the vicinity of its habitat.

Mud cracks in micritic limestone beds of the Natica Member reach to a depth of as much as 20 cm. Pseudomorphs of calcite after evaporite minerals occur in the proximal facies of the Hauptmumienbank limestone Member. Clusters of such pseudomorphs are rather common in the Verena limestone Member of the Balsthal Formation (GYGI & PERSOZ 1986). A solution breccia found in the Verena Member of section RG 403, horizon 4, near Vermes, is strictly local. These are indications that there were dry seasons with more evaporation than rainfall. The climate of the area was therefore seasonal, at least from the Middle Oxfordian.

The carbonate platform of the lower Günsberg Member grades both landward and towards the "basin" into marl. Instead, the proximal as well as the distal time equivalents of the St-Ursanne carbonate platform and of the carbonate platform of the Balsthal Formation are limestone. This may mean that there were long-term climatic changes in the Oxfordian. The climate could have been relatively wet from the beginning of the Oxfordian to the early Transversarium Chron, then may have become drier to the end of the Transversarium Chron when the St-Ursanne Formation was deposited. Another comparatively wet period would have provided for the ample supply of terrigenous sediment of the Natica and mainly of the Effingen Member. The insignificant content of terrigenous sediment in subsequence 3b would be an indication of a final dry period in the Oxfordian. Nevertheless, rainy seasons must have existed even in such relatively dry periods, since cycadophytes were growing on adjacent land when the St-Ursanne Formation was sedimented (PÜMPIN 1965, Fig. 21), and because a land vegetation sufficient to support large animals must be postulated for the early Kimmeridgian when the almost pure limestone Reuchenette Formation was sedimented. The climate of the Oxfordian and of the early Kimmeridgian was therefore seasonally wet in Central Europe, and the transition to the arid belt of the lower latitudes was somewhat further to the south than indicated in Figure 7 by HALLAM (1985).

4.6 Shallow- versus deep-seated origin of subsidence

The margin of the St-Ursanne carbonate platform subsided during deposition (GYGI & PERSOZ 1986, Pl. 1A). The total thickness of the carbonate rocks of the platform varies between 35 m in the interior and as much as 105 m near the margin (not shown in the plate mentioned above). There is also great variation in the thickness of the carbonate platforms of the Günsberg and of the Holzflue Members. The greatest thicknesses occur in both members above the proximal end of the clinoforms of the Effingen Member directly below. We therefore assume that the great thickness changes over short distances in the carbonate platforms as reported by GYGI (1969, Fig. 8–11) are related to processes in the sediment of the Effingen Member below, which was then being compacted. Small growth faults are the most probable process, because there are no areally restricted thickness minima in the carbonate platforms as would be there if clay diapirism was the cause of thickness variation. The slope facies of the Effingen Member below the Günsberg Member is a wedge, the thickness of which increases rapidly in the distal direction. The total amount of thickness reduction by compaction grows with the thickness of the wedge. This is why the base of the carbonate platform of the Günsberg Member subsided increasingly and the thickness of the platform generally augmented with progradation (GYGI & PERSOZ 1986, Pl. 1A).

Loading of the lithosphere with sediments was by far the most important single factor in basement subsidence of the region in Oxfordian time. The lithosphere subsided differentially under the shifting depocenters of sediments. The isostatic adjustment took place either along deep fractures spaced at close intervals, or by flexure of the crust (see WATTS & RYAN 1976, p. 42). A depression would have developed at the toe of the slope facies of sequence 1 in the Solothurn Jura at the latest by the Transversarium Chron if the adjustment of the crust was flexural. There is no evidence of this.

5. Conclusions

Some of the Oxfordian sea level changes were most probably eustatically controlled. There is evidence that in fact all sea level changes concluded above from sedimentologic observations, fossil assemblages, calculations of subsidence, or from the geometry of sediment bodies were eustatic. Sea level normally rose in a short time, then remained more or less stable for a comparatively long time. However, the substantial sea level rise in the latest part of the Oxfordian was apparently gradual. The eustatic sea level changes of the Oxfordian Age resulted in a net rise which is estimated at 25 to 30 m.

Sedimentation was strongly influenced by the eustatic sea level changes. Some of the boundaries of sequences or of subsequences as differentiated by GYGI & PERSOZ (1986) therefore can be used in correlation. Eustatic sea level changes of only a few meters had in some cases a noticeable effect on sedimentation and biocoenosis to a depth of more than 100 m. Carbonate production in shallow water continued when siliciclastic sediment was brought into the environment at a high rate. The varying ratio between terrigenous and carbonate sediment supply was probably influenced by both eustasy and changes in the climate.

Lithosphere subsidence of northern Switzerland occurred in the Late Jurassic mainly under the weight of sediments. The majority of basement subsidence was the effect, not the cause of sedimentation. Differences in time and space in the loading of the lithosphere with sediments led to differential deep-seated subsidence. Some small and isolated parts of carbonate platforms subsided more than the platform as a whole. This may be the result of growth faults.

The climate was at least subtropically warm and moderately seasonal. Long periods of relatively humid climate alternated with comparatively arid periods. When the climate was humid, then terrigenous sediment was supplied at a high rate. In times of increasingly arid climate, the influx of terrigenous sediment diminished. Carbonate sedimentation then prevailed, and in the Late Oxfordian, even small amounts of evaporites appeared. In the relatively humid Bifurcatus Chron, rainfall became so scarce in the dry seasons that prism cracks developed at the surface of mud-grade sediments of the peritidal zone. On the other hand, in relatively arid climatic periods, precipitation was sufficient for land plants to survive in an adequately abundant growth to support large plant-eating animals.

The marine macrofauna changed rapidly with paleodepth. Cephalopods were rare in very shallow water where hermatypic corals were prolific. Cephalopods began to predominate in the macrofauna at a depth of about 50 m. Cephalopods increased to more than 80% the macrofauna at a paleodepth of about 80 m, then apparently did not become much more abundant with further increasing depth. Greater depth to at least 150 m can be concluded from the composition of the ammonite fauna. The ammonites lived apparently in close relation to benthic communities. Macrofossil assemblages provide detailed paleobathymetric information down to at least 150 m, where most sediments are mud-grade and normally include little evidence of water depth.

An unsolved problem is why all sediments of the Early Oxfordian and a portion of the Middle Oxfordian sediments wedge out completely at the toe of clinothems near Péry. There is no direct connection between the lower end of the clinothems and the coeval starved basin deposits (see GYGI & PERSOZ 1986, Table 2 and Pl. 1A). A similar problem is the marked thinning of the early Kimmeridgian Baden Member within a clinothem (GYGI & PERSOZ 1986, Pl. 1B). It is not known how the submarine terrace of the Villigen Formation or of the Gerstenhübel Beds developed in the deep subtidal zone between Aarau and Villigen (see GYGI & PERSOZ 1986, Pl. 1A).

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