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Mineralostratigraphy, litho- and biostratigraphy combined in correlation of the Oxfordian (Late Jurassic) formations of the Swiss Jura range

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ABSTRACT

Oxfordian sedimentation began in northern Switzerland in the cephalopod facies of a moderately deep epicontinental sea adjacent to the Tethys. Argillaceous mud supplied from the northwest accumulated a submarine bank until the Middle Oxfordian. A carbonate platform with coral bioherms evolved in Middle Oxfordian time above this bank in the northwest. Later, the carbonate platform facies prograded 40 km basinward in about 4 m.y., mainly because of a second phase of strong supply of argillaceous mud. It is the Middle Oxfordian carbonate platform and coeval basinal deposits that inspired GRESSLY (1838) to develop his theory of facies.

Newly-measured sections are presented in a composite cross section along a line of maximum facies change. Sedimentologic data and ammonites collected from in situ revealed that the sediments in cephalopod facies of the Middle and of the Late Oxfordian formed gently sloping progradational bodies (clinothems of RICH 1951), sigmoid in cross section, which are to a large extent juxtaposed to the earlier Oxfordian sediment stack. Cephalopods are so rare in the shallow water facies that detailed correlation based on ammonites between shallow-water and basinal facies was not possible.

Unlike ammonites, detrital clay minerals are ubiquitous. Clay minerals and detrital quartz were supplied from the north. Changes in the source area caused short-term vertical variation in the clay mineral assemblages. The kaolinite content was found to be little influenced by facies, lithology, or diagenesis. Distinct highs or lows of the kaolinite content can be correlated from section to section, from terrestrial to basinal facies. Stratigraphic correlations based on kaolinite are near-isochronous. Mineralostratigraphic correlations were calibrated with the biochronologic ammonite scale by clay mineral analysis through part of the Oxfordian and the early Kimmeridgian cephalopod facies. The resolution in correlation by vertical changes of the kaolinite content is of the order of one ammonite subzone. Problems of correlation between the basin and the shallow-water realm could be solved, and correlations within the shallow-water facies were refined by this method.

Mineralostratigraphic correlations and some ammonites are proof that the shallow-water St-Ursanne Formation is time-equivalent to the basinal Birmenstorf Member, and that the Natica Member is equivalent to the Effingen Member as BOLLIGER & BURRI (1970) alleged. The Hauptmumienbank Member is the same age as the Geissberg Member. The boundary between the Court Formation and the Reuchenette Formation almost coincides with the Oxfordian/Kimmeridgian boundary.

RÉSUMÉ

Au nord de la Suisse, la sédimentation oxfordienne débute par des faciès à céphalopodes dans un bassin peu profond (80 à 100 m), dans une mer épicontinentale adjacente à la Téthys. Les premiers sédiments oxfordiens apportés du nord-ouest comblaient presque la partie marginale du bassin. A l'Oxfordien moyen une plate-forme carbonatée avec biohermes de coraux s'individualise au-dessus du haut-fond argileux, puis migre de 40 km vers le sud-est en près de 4 m.a. suite à une seconde phase de sédimentation de vase terrigène. Le contraste lithologique

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entre la plate-forme carbonatée et les faciès vaseux du bassin a conduit GRESSLY (1838) à formuler le concept de faciès.

De nouveaux levés, ordonnés selon une direction perpendiculaire aux lignes d'isofaciès, ont permis de construire une coupe palinspastique (pl. 1). Grâce aux ammonites et aux données sédimentologiques on peut démontrer que les éléments sédimentaires, dans les faciès à céphalopodes de l'Oxfordien moyen et supérieur, présentent une structure sigmoïdale avec faible pente en direction distale. Cette structure résulte dans une large mesure d'une adaptation à la morphologie du bassin de l'Oxfordien inférieur. Les céphalopodes sont si rares dans les faciès d'eau peu profonde qu'il est impossible de corréler précisément avec des ammonites entre bassin et plate-forme.

Par contre, les minéraux argileux sont omniprésents. Les minéraux argileux et le quartz détritiques ont leur source au nord. Dans ces zones d'alimentation des modifications d'origine incertaine (climatique, tectonique, morphologique) entraînent des variations verticales de teneurs au sein des minéraux argileux, lesquels semblent peu influencés par les milieux de dépôt, la lithologie et la diagenèse. Grâce à d'importantes variations verticales, les minima et maxima de teneur en kaolinite peuvent être corrélés de coupe en coupe depuis les domaines terrestres à ceux du bassin. Des arguments sédimentologiques démontrent que les corrélations basées sur la distribution des teneurs en kaolinite sont quasi isochrones. Dans les faciès à céphalopodes, les corrélations minéralostratigraphiques ont pu être testées et calibrées par l'échelle biochronologique des ammonites, ceci à l'Oxfordien et au Kimméridgien inférieur. Le domaine de résolution de ces corrélations minéralostratigraphiques est de l'ordre d'une sous-zone d'ammonites. Grâce à cette méthode il a été possible de résoudre la question débattue des corrélations entre les faciès de plate-forme et ceux du bassin, et d'affiner la stratigraphie des faciès de faible profondeur.

Les corrélations minéralostratigraphiques et quelques ammonites démontrent que la Formation de St-Ursanne et le membre des Couches à Natica déposés en milieux peu profonds sont des équivalents-temps du Membre de Birmenstorf, respectivement de celui des Couches d'Effingen déposé en milieux bassin. Ces corrélations confirment en gros celles proposées par BOLLIGER & BURRI (1970). Les Membres de l'Hauptmumienbank et de Geissberg sont de même âge. La limite entre la Formation de Court et celle de Reuchenette est en quasi coïncidence avec la limite Oxfordien/Kimméridgien.

ZUSAMMENFASSUNG

In der Nordschweiz begann die Sedimentation des Oxfordian in der Cephalopodenfazies eines mässig tiefen, an die Tethys grenzenden epikontinentalen Meeres. Von Nordwesten her geschütteter, vorwiegend toniger Schlamm füllte den proximalen Teil des Beckens weitgehend auf. Im Mittel-Oxfordian entstand über der tonigen Schlammbank eine Karbonat-Plattform mit Korallenbiohermen, welche infolge einer zweiten Phase starker terrigener Schlammzufuhr in etwa 4 Millionen Jahren 40 km beckenwärts progradierte. Aufgrund dieser Karbonat-Plattform und der entsprechenden Beckensedimente entwickelte GRESSLY (1838) seine Fazieslehre.

Neu aufgenommene Detailprofile wurden in einem dem maximalen Faziesgefälle entlang verlaufenden Sammelprofil dargestellt. Sedimentologische Daten und horizontierte Ammoniten ergaben, dass die Sedimente des Mittleren und Späten Oxfordian in Cephalopodenfazies als parallele Gürtel mit sigmoidem Querschnitt den älteren Oxford-Sedimenten grösstenteils seitlich angelagert sind. Cephalopoden sind in der Seichtwasserfazies so selten, dass detaillierte Korrelationen zwischen Becken- und Seichtwasserfazies mit Ammoniten nicht möglich sind.

Im Gegensatz zu den Ammoniten sind detritische Tonmineralien ubiquitär. Tonmineralien und detritischer Quarz stammen aus dem Norden. Veränderungen im Liefergebiet verursachten kurzfristige vertikale Wechsel in den Tonmineralspektren. Die Kaolinitgehalte wurden weder von der Fazies, der Lithologie noch von der Diagenese erheblich beeinflusst. Bestimmte Kaolinit-Maxima oder -Minima können von Profil zu Profil korreliert werden, und zwar von der terrestrischen bis zur tiefermarinen Beckenfazies. Stratigraphische Korrelationen mittels Kaolinit sind nahezu isochron. Die mineralstratigraphischen Korrelationen wurden an der biochronologischen Ammoniten-Zonierung geeicht, mittels tonmineralogischer Analyse durch einen Teil des Oxfordian und das untere Kimmeridgian in Cephalopodenfazies hindurch. Korrelationen aufgrund von vertikalen Schwankungen des Kaolinitgehaltes haben ein Auflösungsvermögen von etwa einer Ammoniten-Subzone. Damit wurden Korrelationen von der Becken- zur Seichtwasserfazies möglich, und innerhalb der Seichtwasserfazies ergab sich eine verfeinerte Korrelation.

Die mineralstratigraphischen Korrelationen und mehrere Ammoniten beweisen, dass die St-Ursanne-Formation der Seichtwasserfazies gleich alt ist wie die Birmenstorfer Schichten im Becken und dass die Natica-Schichten den Effinger Schichten entsprechen, wie BOLLIGER & BURRI (1970) behaupteten. Die Hauptmumienbank ist gleich alt wie die Geissberg-Schichten. Die Grenze zwischen der Court- und der Reuchenette-Formation fällt beinahe mit der Oxfordian/Kimmeridgian-Grenze zusammen.

CONTENTS

1. Introduction

1.1 Previous work

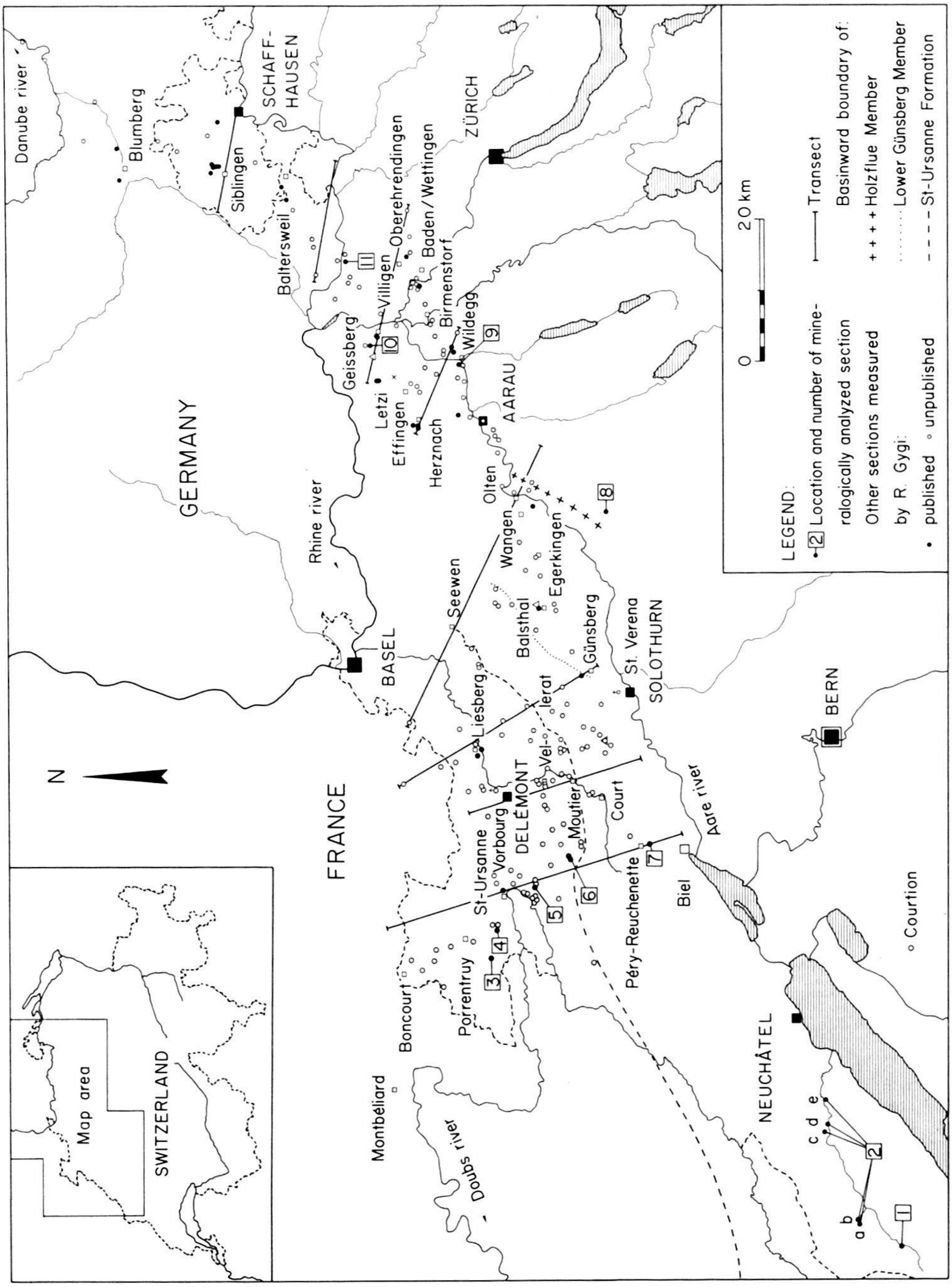
Stratigraphic work in the Swiss Jura range began before 1820. MERIAN (1821) correlated the Early and Middle Oxfordian marl-clays (Renggeri Member and Terrain à Chailles Member) of northwestern Switzerland with the Effingen Member in canton Aargau, and the coral limestones of the St-Ursanne Formation in the "basinal" realm of canton Aargau (see Pl. 1A). GRESSLY (1838–41) carried out extensive geological mapping in canton Solothurn and in adjacent areas. Close observation of the coral bioherms and the coeval fine-grained sediments of the St-Ursanne Formation at La Caquerelle near St-Ursanne inspired this author to introduce the concept of facies into the scientific literature. Ironically, the great effort in mapping, practical use of the new stratigraphic method, and ample fossil collecting did not allow this distinguished geologist to arrive, in his own opinion, at a satisfactory time correlation between the deposits from shallow water and those of deeper marine origin. The progress of paleontology as pioneered by OPPEL (1856–8, 1862–3) in the Oxfordian of this region led to an important revision of Merian's correlation. On the evidence of ammonites, OPPEL (1857, p. 626) recognized that the thick Renggeri Member and part of the Terrain à Chailles Member in the northwest thin out to the southeast and grade into what is now called the Schellenbrücke Bed (Table 2 and Pl. 1A).This is a ferruginous marly limestone with iron ooids. The thickness of the bed is normally less than 10 cm. ROLLIER (1888, p. 87) correlated the Liesberg Member with the Birmenstorf Member, and the St-Ursanne Formation with the Effingen and Geissberg Members. He had no ammonites from the platform deposits to support his assertion. Later, ROLLIER (1911, Fig. 54) reaffirmed his view, and it remained unchallenged until 1967, when BOLLIGER & BURRI proposed another correlation which was based on the distribution of detrital quartz. The significance of the important paleontological work by de Loriol in relation to Rollier's correlation was not appreciated by most stratigraphic workers, perhaps because de Loriol paid so little attention to stratigraphy. In a short review, ARKELL (1956, p. 95–96) threw some light on the stratigraphic implications of the ammonites published by de Loriol from the Oxfordian in northwestern Switzerland.

M. A. ZIEGLER (1962) studied the platform to basin transition of what is now the St-Ursanne Formation (Table 2). He did not question Rollier's correlation of this formation with most of the Wildegg Formation (Table 2). Ziegler presented evidence for the wide areal extent of a supratidal horizon with characeans and limnic ostracods in the Natica Member above the St-Ursanne Formation (see OERTLI & ZIEGLER 1958). Thin coal seams in about this level have been reported before by HEER (1865, p. 125) and by KEMMERLING (1911, p. 22, see also LAUBSCHER & PFIRTER 1984, p. 208). BOLLIGER & BURRI (1970) figured peritidal stromatolites from the Natica Member. ROLLIER (1898, p. 58) reported limestone bands in the Natica Member with sufficient fine-grained detrital quartz that they could be worked to grindstones near Damvant, west of Porrentruy. M. A. ZIEGLER (1962, p. 26, 42) concluded from the good sorting of the quartz that it might be of eolian origin. This view was taken up by BOLLIGER & BURRI (1967). They considered the vertical variation of the quartz content to be a valuable means of correlation between the platform and the "basin". In doing so, they arrived at a correlation of the Natica Member with the Effingen Member, and at a correlation of the St-Ursanne Formation with the Birmenstorf Member alone. The latter correlation has since been corroborated on the strength of ammonites as found in situ by PÜMPIN (1965, Pl. 1A, section 2) in the upper St-Ursanne Formation at St-Ursanne. These ammonites were identified by R. Enay and R. Gygi to be from the late Antecedens Subchron (see BAYER et al. 1983, p. 128).

The correlations by M. A. ZIEGLER (1962) and by BOLLIGER & BURRI (1967, 1970) are conflicting. Moreover, there are some shortcomings in the argumentation by Bolliger and Burri relating to the alleged eolian origin of the detrital quartz and the correlations based on it (see below). The following problems remained unresolved in 1970:

Fig. 1. Location of measured and mineralogically analyzed sections. The township name is followed by the abbreviated name of the canton (NE = Neuchâtel, JU = Jura, BE = Bern, LU = Luzern, AG = Aargau), then by the locality name. Coordinates are after the Swiss National Map. First set of coordinates: starting point of section. Second set: end of section. The number of sections measured by R. Gygi is preceded by RG.

1: La Côte-aux-Fées NE, Noirvaux, 529.450/190.520, 529.525/189.855.
2a: St-Sulpice NE, Haut de Tour II, 532.620/196.160, 532.230/196.060.
2b: St-Sulpice NE, Haut de Tour I, 532.985/196.440, 532.890/196.360.
2c: Noiraigue, NE, well near Clusette, 544.912/200.859.
2d: Noiraigue NE, well near Clusette, 546.779/200.787.
2e: Boudry NE, Combe Garot, 551.240/201.430, 551.960/201.240.
3: Bressaucourt JU, water well, RG 359, 569.980/248.490.
4: Courgenay JU, Chemin paulin, RG 350, 573.850/247.100, 573.950/247.430.
5: Glovelier JU, road near Foradrai, RG 323, 579.760/242.000, 580.090/242.070.
6: Sornetan BE, Gorges du Pichoux, RG 314, 584.150/237.230, 584.170/237.220; RG 315, 584.070/237.170, 584.030/236.720.
7: Péry BE, La Charuque and La Reuchenette quarries, RG 307, 585.600/225.250, 585.850/226.400.
8: Pfaffnau LU, gas well Pfaffnau 1, 632.708/231.789.
9: Auenstein and Veltheim AG, Unteregg and Jakobsberg quarries, RG 226, 653.950/252.800, 654.000/252.420, RG 37, 653.900/252.400, 653.800/252.050.
10: Villigen AG, Gabechopf quarry, RG 294, 656.550/264.900, 656.570/264.850.
11: Mellikon AG, new quarry, RG 70, 668.100/268.650, 668.300/268.600.


LEGEND:
2 Location and number of mineralogically analyzed section
Other sections measured by R. Gygi:
• published ∘ unpublished
Transect
Basinward boundary of:
+ + + + Holzflue Member
Lower Günsberg Member
St-Ursanne Formation
0 20 km
N
GERMANY
FRANCE
Danube river
Rhine river
Aare river
Doubs river
SCHAFF-HAUSEN
ZÜRICH
BASEL
AARAU
SOLOTHURN
BERN
DELÉMONT
NEUCHÂTEL
Blumberg
Siblingen
Baltersweil
Oberehrendingen
Baden / Wettingen
Villigen
Birmenstorf
Geissberg
Wildegg
Letzi
Effingen
Herznach
Olten
Wangen
Egerkingen
Seewen
Balsthal
Günsberg
St. Verena
Liesberg
Vel-lerat
Court
Moutier
Vorbourg
St-Ursanne
Péry-Reuchenette
Biel
Porrentruy
Boncourt
Montbéliard
Courtion
Map area
SWITZERLAND

— What is the time equivalent of the Liesberg Member and of the St-Ursanne Formation in the "basin"?
— Can the Hauptmumienbank marker bed of P. A. ZIEGLER (1956) be related to any unit in the "basin"?
— How can the boundary between the Court Formation and the Reuchenette Formation be defined, and what is the age of this boundary?

GYGI (1969) worked out Oxfordian and Early Kimmeridgian lithostratigraphy and ammonite biostratigraphy mainly in the "basinal" facies of the cantons Aargau and Schaffhausen. GYGI & MARCHAND (1982) figured the cardioceratid ammonites which are relevant for biochronology from the end of the Middle Jurassic Epoch to the Middle Oxfordian Age. In these papers and in the one by BAYER et al., the base of the Oxfordian Stage in northern Switzerland was defined lithostratigraphically and biostratigraphically. Major problems of correlation were solved from the base of the Oxfordian upward to the St-Ursanne Formation. The sedimentology and the paleoenvironment of the Early Oxfordian Schellenbrücke Bed was investigated by GYGI (1981). An outline of the depositional history of the Oxfordian in northern Switzerland is to be found in BAYER et al. (1983). The radiometric ages obtained by GYGI & MCDOWELL (1970) of the Glaukonitsandmergel (Early Oxfordian) and of the lower Baden Member (Early Kimmeridgian) were republished, with slight revisions, by ODIN (1982): 148 m.y. and 139.5 m.y., respectively.

In 1979, R. Gygi began measuring sections and collecting ammonites in the part of northwestern Switzerland previously worked on by P. A. ZIEGLER (1956), M. A. ZIEGLER (1962), and BOLLIGER & BURRI (1967, 1970), in order to re-evaluate the controversial results of these authors. This work is not yet completed: some sections from south of the Laufen Basin to the southern boundary of the Jura range have yet to be measured or supplemented. The aim was to subdivide the successions into sequences, to calibrate the sequences with the previously established ammonite biochronology and with numerical ages, and to use sequence boundaries for correlation in successions without ammonites. It soon became apparent that in some sections, the base of sequence 3 as conceived in Plate 1 of this paper could not be discerned with certainty in the field. The boundary between the Balsthal Formation and the Reuchenette Formation could be well recognized and correlated in the field between Courgenay and Egerkingen (see this paper, Fig. 1 and Pl. 1A), but there were no ammonites to test the correlation in detail.

All fossils, including ammonites, are more or less restricted to certain environments. Ammonites are rare or absent in sediments from very shallow water (B. ZIEGLER 1971, p. 35). Detrital clay minerals have the advantage that they are ubiquitous. They are present even in winnowed deposits like carbonate oolite. PERSOZ & REMANE (1976, p. 35) and PERSOZ (1982, p. 15) concluded that the vertical variation in the distribution of clay minerals, and particularly the variation in kaolinite, could be used for stratigraphic correlation from terrestrial to deeper marine environments.

1.2 Purpose of this paper

It is the purpose of this paper to investigate the feasibility of detailed correlations based on clay minerals, and to find what degree of resolution can be obtained by mineralostratigraphic correlations in an area of limited geographic extent. The Oxfordian of the Swiss Jura range was chosen, because on the one hand, these sediments have been investigated at intervals since the beginnings of stratigraphy in this country, and because, on the other hand, no consensus on correlation has been reached to this day in spite of the large amount of work done. We wish to stress that the results presented in this paper are preliminary. Most of the sections measured by R. Gygi are unpublished, and so are most of the ammonites collected by R. and S. Gygi since 1962, on which biostratigraphy and most of the regional correlations of the Oxfordian as presented in this paper are based.

2. Methods

The biochronologic framework for the mineralostratigraphic correlations was provided by sampling three sections for clay mineral analysis in the "basinal" cephalopod facies of canton Aargau (no 9, 10, and 11 in Pl. 1). Ammonite biochronology is well known in the rhodano-swabian "basin" (ENAY 1966) which was a shallow sea on the southern continental shelf of Jurassic Europe. The deeper marine Oxfordian and Kimmeridgian sediments discussed in this paper were laid down in part of this epicontinental sea, which was less than 200 m deep. Therefore, the term basin is referred to with quotation marks in this paper. The other sections sampled for clay minerals include paleoenvironments from the slope, the margin, and the inner part of carbonate platforms, as well as environments of the intertidal and the supratidal zones.

2.1 Lithostratigraphy

The palinspastic cross section of Plate 1 is assembled from sections measured by R. Gygi (Fig. 1). The sections are spaced as closely as possible in order to facilitate correlation. They are arranged in transects perpendicular to depositional strike, or lines of equivalent facies, respectively. Thus, the cross section of Plate 1 is drawn along a line of maximum facies change. One difficulty in preparing the plate was that the slope of the Pichoux limestone between the St-Ursanne Formation and the Birmenstorf Member (see Table 2) varies from transect to transect. Another difficulty is the change of thickness within the St-Ursanne Formation which is from 35 m in the platform interior to more than 100 m at the platform margin. Even greater thickness variance over short distances occurs within the Natica Member at the transition to the Günsberg Member. Thicknesses had to be averaged in order to make Plate 1 interpretable.

When measuring the sections, every recognizable horizon down to a thickness of a few centimeters was sampled individually. Thick-bedded or massive limestones were sampled at intervals of 1 m or less. Some of the thick-bedded or massive shallow-water limestones in the sections done by GYGI (1969) were sampled at larger intervals. Only the limestone samples processed to thin sections or polished slabs, or marls to be washed, are kept in the Museum of Natural History, Basel. Plate 1 is based on more than 2150 thin sections and more than 560 polished slabs from 182 measured sections. The sections were subdivided into shallowing-upward sequences. Rock names for primary mixtures of clay and lime mud are from PETTIJOHN (1957, Fig. 99).

2.2 *Biostratigraphy*

Ammonites are the fossils most suitable for establishing a biochronology in the "basinal" deposits considered. The ammonites relevant to biochronology and regional correlation as indicated in Plate 1 are mostly from approximately 5000 specimens prepared out of the 8000 collected by R. and S. Gygi since 1962. These were either taken from systematic excavations, or collected from in situ in measured sections. Many an important ammonite included in Plate 1 has been given on loan or as a gift to the Museum of Natural History Basel. Ammonites cited from this museum have an individual number preceded by J or Gy. Paleontologic procedures are described in GYGI (1977). Ammonite zonation is discussed by GYGI & MARCHAND (1982).

2.3 *Mineralostratigraphy*

The methods of mineralogical analysis and principally those of phyllite minerals discussed in PERSOZ & REMANE (1976) and PERSOZ (1982) are based on the classical works by BROWN (1961) and BRINDLEY & BROWN (1980). The quartz content is measured by X-ray diffractometry with an external standard. It is expressed relative to the total rock. The phyllite percentages are measured on granulometric fractions of less than two microns on oriented samples. The percentage is calculated relative to the sum of all phyllites present. These are measured by the height of the peak, without any corrections at 17 Å (smectites), between 11 and 17 Å (mixed layers), at 10 Å (illite-micas), and at about 7 and 3.5 Å (kaolinite and chlorite).

The estimation of the kaolinite/chlorite ratio is sometimes difficult on the basis of the two peaks at 3.5 Å. In this paper, this uncertainty is without great consequence on the kaolinite content due to the very low values of chlorite percentages, which are generally less than 10%.

The mixed-layer minerals are generally interstratifications of illite and smectites. These most often form a broad peak on diffractograms. Sometimes one or two discrete lines are present (Fig. 6 and 7). It was impossible to analyze this complex system in routine work.

The term illite-micas was chosen due to the difficulty in discriminating between true micas and illites which are clay grade micas (BRINDLEY & BROWN 1980, and KÜBLER 1984).

The crystallinity index of illite-micas (Fig. 8) is defined by the width of the peak at half of its height, at 10 Å on oriented samples as treated with ethylene glycol (IAGI). This procedure is necessary because of the presence of swelling mixed layers which artificially increase the width of the peak. IAGI is not equivalent to the crystallinity index of KÜBLER (1964, 1966) which is measured in the same manner, but on air-dried samples. IAGI is generally lower than the KÜBLER index except at the base of the anchizone and in the epizone where it approaches the KÜBLER index. In the diagenetic zone, values less than $0.42° 2\theta$ (Neuchâtel calibration), which is for the KÜBLER index the limit between diagenesis and anchimetamorphism, indicate a probable metamorphic or anchimetamorphic origin of illite.

The crystallinity index of kaolinite (IANK) is measured on air-dried samples on the (001) peak at about 7.1 Å. This is possible because of the very low content of chlorite whose (002) peak is also at 7 Å. There is no statistical correlation between IANK and chlorite content except in the Effingen Member of the Reuchenette and Aargau sections (Fig. 9), where the correlation coefficient is significant at 95% of confidence level, but not at 99%. In this case, the chlorite/kaolinite ratio is higher than in all the other members. Another statistical control shows that the variation of the height of the (001) peak of kaolinite has no influence on the IANK values.

Mineralogical analysis was carried out on more than 650 samples from 9 composed sections as measured by GYGI (1969), BOLLIGER & BURRI (1970), and completed by us in the upper part, as well as from sections by BIELER (1972), KETTIGER (1981), PERSOZ & REMANE (1973), and from a section near Courgenay as well as one near Bressaucourt as measured by Gygi (unpublished). The spacing of the samples is indicated in Figures 9 and 10.

2.4 Paleoecology, correlation, and depositional history

The paleobathymetry and paleogeography of northern Switzerland in Early Oxfordian time was worked out by GYGI (1981). Observations on sedimentology, on assemblages of the macrofauna, on the taphonomy of ammonites, and on the regional distribution of the minerals glauconite and chamosite/berthierine led to the conclusion that iron ooids could be accreted on the surface of marine mud at a depth of as much as 100 m. The thin, widespread, and mostly continuous iron-oolitic horizons at the base of the Oxfordian of northern Switzerland are sediments from relatively deep water. The deep water origin of oncoids several centimeters across in a thin glauconitic bed of the Middle Oxfordian in canton Schaffhausen was established by the same methods (GYGI et al. 1979, see Fig. 14).

Shallowing-upward sequences are defined by the type of sediment as identified in thin section or on polished slabs, and by the assemblages of fossils. Broad estimates of paleodepth were made from faunal assemblages according to B. ZIEGLER (1971). Minimum depth may be deduced from the taphonomy of ammonites in some cases. Argillaceous or carbonate mudstone with a macrofauna dominated by ammonites is taken as an indication of a water depth greater than about 50 m (GYGI 1981, p. 245). Biostromes and bioherms of colonial corals were formed in the shallow subtidal or in the lower intertidal zone. Calcarenite with preserved cross-bedding is a sediment from the intertidal and the shallow subtidal zones. Prism-cracked mudstone in itself is not conclusive evidence for the supratidal environment (see HARDIE 1977, Fig. 28). An environment of the upper intertidal or of the supratidal zone must be assumed when there are stromatolites with crinkled fenestrate laminations, and when stromatolites can be observed to pass laterally into mudstone with pebbles. A limestone with rounded or angular pebbles and cobbles of carbonate rock above a hummocky erosion surface of oolitic limestone is considered to be evidence of the supratidal zone.

Correlation was made with ammonites when possible. Lithologic marker beds were used when ammonite zonation was uncertain (see GYGI 1969, Fig. 5). Sequence boundaries as for instance the one between sequence 3 (Court Formation) and 4 (Reuchenette Formation, see Plate 1) were also used for correlation. Part of the correlations depend on the vertical variation in the kaolinite content alone.

3. Results

3.1 Formations, members, and beds

Most of the lithostratigraphic units mentioned below have recently been either revised or given a new name and definition. We use these names provisionally without revision as far as possible. A general revision of lithostratigraphic units will be made elsewhere. Lithostratigraphic names and abbreviations are listed in Table 1. The geometry of the units is represented in Plate 1. Ammonites critical of the age are mentioned with the lithostratigraphic units. The age of these units is represented graphically in Table 2.

Table 1: *Names and abbreviations of lithostratigraphic units represented in Figures 9 through 15 and in Plate 1. All units were recently defined except ON, OR, and SPO.*

BAD:	Baden Member	MUM:	Mumienmergel
BAN:	Banné marl	NAT:	Natica Member
BIR:	Birmenstorf Member	ON:	Oolithe nuciforme
CRE:	Crenularis Member	OR:	Oolithe rousse
EFF:	Effingen Member	PIC:	Pichoux limestone
GEI:	Geissberg Member	REN:	Renggeri Member
GER:	Gerstenhübel Beds	REU:	Reuchenette Formation
GSM:	Glaukonitsandmergel	RKK:	Rauracien-Korallenkalk
GÜN:	Günsberg Member	SCH:	Schellenbrücke Bed
HMB:	Hauptmumienbank	SPO:	Spongitien
HOB:	Hornbuck Member	STE:	Steinibach Beds
HOL:	Holzflue Member	SUF:	St-Ursanne Formation
HRA:	Hautes Roches Algenkalke	SWB:	Schwarzbach Member
HUM:	Humeralis marl/limestone	TAC:	Terrain à Chailles Member
KKS:	Kreidige Kalke von St-Ursanne	TO:	Tiergarten-Oolith
KNO:	Knollen Beds	VER:	Verena (Ste-Vérène) Member
KÜS:	Küssaburg Member	VOR:	Vorbourg Member
LET:	Letzi Member	WAN:	Wangen Member
LIE:	Liesberg Member	WAT:	Wangental Member
MKK:	Moutier Korallenkalk	WET:	Wettingen Member
MUK:	Mumienkalk		

3.1.1 Early to Middle Oxfordian marls

The Oxfordian Stage of northwestern Switzerland begins with a succession of marls. The maximum thickness of the marls exceeds 100 m. This unnamed formation thins or wedges out to the south and to the east. It is made up of four units (from base to top):

1. At the base is a thin (0.2 to 0.3 m), ferruginous marl-clay with ooids of iron hydroxide and an abundant macrofauna of mostly ammonites. This is the uppermost part of the Anceps–Athleta Beds auctorum. The ammonite taxon *Quenstedtoceras lamberti* (J. Sowerby) is common in the lower part of the unit in the clay pit of Andil southwest of Liesberg and in the cement quarry of La Charuque south of Péry (see Fig. 1). These ammonites are from the Lamberti Subzone. This is the last subzone of the Callovian Stage or of the Middle Jurassic. Ammonites of the earliest subzone of the Oxfordian are less abundant. *Cardioceras (Scarburgiceras) leachi* (J. Sowerby) J 30709 of the Scarburgense Subzone was found in the upper part of the ferruginous marl-clay in the cement quarry near Péry. The boundary between the Middle and the Upper Jurassic is therefore within the ferruginous marl-clay near Péry as well as elsewhere (see Table 2). This oldest lithostratigraphic unit of the Oxfordian wedges out not far from Péry. Isolated lenses of it are near Herznach and near Gansingen (Gygi & Marchand 1982, Fig. 2 and 3).

2. The Renggeri Member (P. Choffat 1878, p. 35) is a homogenous, blue-grey marl-clay several tens of meters thick. *Cardioceras (Scarburgiceras) scarburgense* (Young & Bird) was found in the lowest part of the unit near Liesberg (J 28155) and near Péry (J 30934). The youngest ammonites from the Renggeri Member are *Cardioceras (Cardioceras) costicardia* S. Buckman. Crushed specimens of these were found in situ in the upper part of the

Renggeri Member in the clay pit of Andil near Liesberg. Good representatives of the taxon are so far known from the drift only.

3. The Terrain à Chailles Member (THURMANN 1830, p. 23) is a grey marl with bands of ellipsoidal limestone concretions and with occasional continuous bands of marly limestone. The normal thickness is between 40 and 50 m. *Cardioceras (Cardioceras) persecans* (S. S. BUCKMAN) was found in situ (GYGI & MARCHAND 1982, Pl. 4, Fig. 5) in a fossil-rich bed which occurs about in the middle Terrain à Chailles, in the distal part of the member (Table 2 and Pl. 1A). The ammonite is common in the Cordatum Subzone (see Table 3). The youngest ammonite known from the Terrain à Chailles Member is *Glochiceras (Glochiceras) subclausum* (OPPEL) J 30932 from a landslide near Montfaucon as figured by DE LORIOL (1901, Pl. 1, Fig. 6). The vertical range of this ammonite taxon begins about in the middle of the Antecedens Subzone and ends in the upper Parandieri Subzone (Table 3).

4. The Liesberg Member (ROLLIER 1888, p. 71) is at the top of the marl formation. The member resembles the Terrain à Chailles in that it is a grey marl with bands of limestone concretions. The difference is that there is generally less interspace between the nodule bands than in the Terrain à Chailles, and that the nodules have an irregular shape. Fossils characteristically have white spots of chert on the surface and in the interior. Dish-shaped hermatypic corals are very common. We have seen no ammonites in the Liesberg Member.

All of the four units wedge out in the distal direction (Pl. 1A). Thin and incomplete time equivalents reappear some distance out in the "basin" (Table 2). Detailed sections of the thin "basinal" sediments of the Early and the Middle Oxfordian as well as part of the ammonite fauna from these horizons were figured by GYGI (1966, 1969, 1977), GYGI et al. (1979), and by GYGI & MARCHAND (1982).

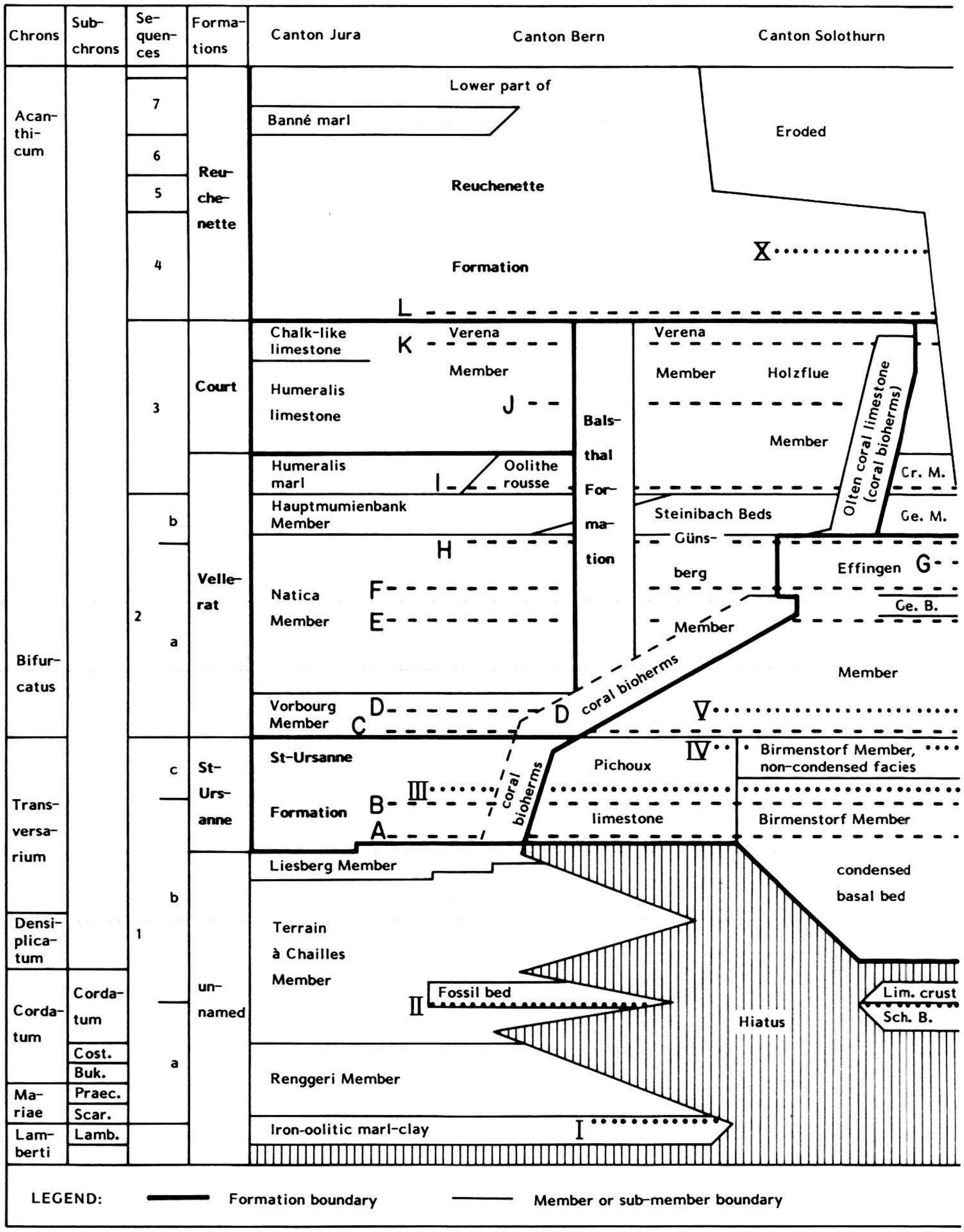
A thin bed of iron-oolitic marl or marly limestone with an ammonite-dominated macrofauna forms the base of the Oxfordian Stage in most of the region (Pl. 1A). It is apparent from ammonites that the ferriferous bed is not of uniform age. To the north of Péry (section no 7 in Pl. 1), the upper (Oxfordian) part of the iron-oolitic horizon is of the early Scarburgense Subchron (Table 2). The Oxfordian part of the bed has a maximum thickness of 0.3 m. East of Péry, time equivalents of this sediment occur only as isolated lenses. In eastern canton Solothurn and in most of canton Aargau, the iron-oolitic Schellenbrücke Bed (GYGI 1977, p. 454) is at the base of the Oxfordian (Pl. 1A). This substantially younger sediment is of the Cordatum Subchron. It is the distal time equivalent of the lower part of the marly Terrain à Chailles Member (Table 2). The Schellenbrücke Bed may include lenses of ferriferous oolite of earlier Oxfordian or even of Callovian age, as for instance near Gansingen (GYGI 1966, p. 938). The Schellenbrücke Bed rests in most places directly on horizons of the Callovian or of the Bathonian (GYGI & MARCHAND 1982, Fig. 2). The thickness of the iron-oolitic horizons had to be greatly exaggerated in Plate 1A.

The Terrain à Chailles Member and the Liesberg Member do not extend as far as Péry. The very thin time equivalent of the middle and upper Terrain à Chailles Member and of the Liesberg Member in the "basin" is the lowermost part of the condensed bed at the base of the Birmenstorf Member (Table 2). The maximum total thickness of this condensed bed is less than 10 cm.

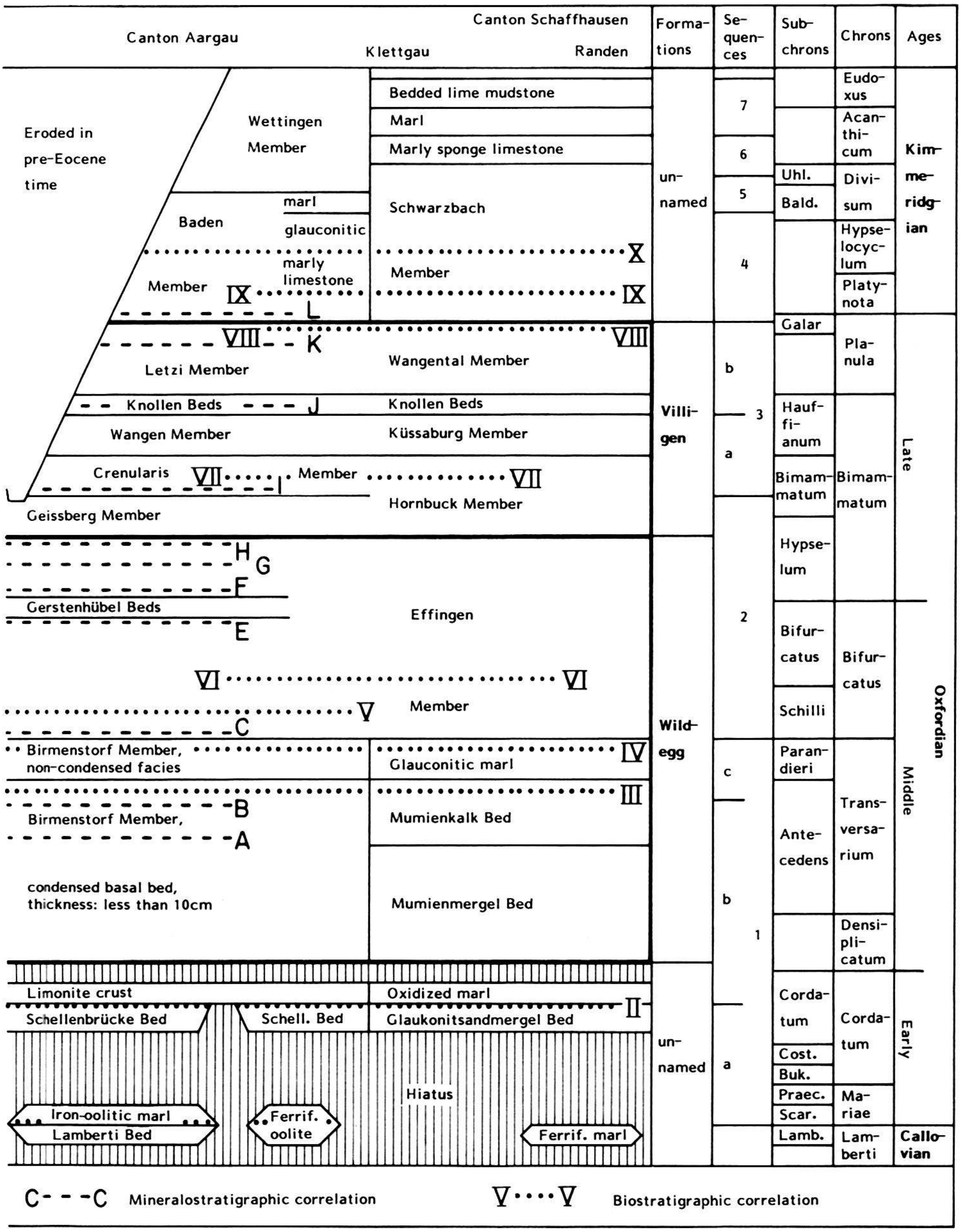
3.1.2 St-Ursanne Formation (BOLLIGER & BURRI 1970, p. 69)

The St-Ursanne Formation as defined by BOLLIGER & BURRI (1970) is the Rauracien of GRESSLY (1864, p. 96) with the exception of the Liesberg Member and an unnamed coral limestone member at the base, and of the Vorbourg Member at the top. The St-Ursanne Formation is a carbonate platform deposit. It is restricted in northwestern Switzerland to where the marl formation of the Early and Middle Oxfordian has a

Table 2: *Time-stratigraphic position of Oxfordian and early Kimmeridgian lithostratigraphic units and hiatuses, with mineralostratigraphic correlations A to L and biostratigraphic correlations I to X (cf. Pl. 1). Well-defined biostrati-*



graphic correlations are the effect of low sedimentation rates. The vertical extent of the lithostratigraphic and biostratigraphic units in this table is unrelated to the time represented by the units. Bold lines are formation boundaries.



minimum thickness of about 100 m. As GRESSLY (1864, p. 100) remarked, the formation is composed of a complicated array of limestone facies (see our Pl. 1A). J. B. GREPPIN (1870) and BOLLIGER & BURRI (1970, Fig. 37) attempted to subdivide the formation into members, but the result was misleading. The thickness of the St-Ursanne Formation varies between 35 m near Kleinlützel (4 km NNW of Liesberg) and about 105 m near Vellerat (see Fig. 1). The basinward boundary of the formation is where hermatypic corals disappear. We draw the base above the uppermost continuous marl band of the Liesberg Member. The upper boundary is where massive, often very porous lime mudstones or oolites become a well-bedded and low-porosity lime mudstone. This is often a vertical transition, and then the formation cannot be precisely delimited at the top. GRESSLY (1864) included the Vorbourg Member in his Rauracien and so avoided this difficulty.

The only ammonites from in situ and from a known stratigraphic position in the St-Ursanne Formation are the *Perisphinctes (Dichotomosphinctes) dobrogensis* SIMIONESCU J 23073 and the *Perisphinctes (Perisphinctes) alatus* ENAY J 23074. The specimens were found by PÜMPIN (1965, Pl. 1, section no 2). The vertical range of the first taxon is within the Antecedens Subzone. The second taxon ranges in southeastern France from the Antecedens Subzone into the Parandieri Subzone. Most of the St-Ursanne Formation was therefore deposited in the later part of the Antecedens Subchron. Only the uppermost part of the formation is of the Parandieri Subchron (Pl. 1A and Table 2, cf. BAYER et al. 1983, Fig. 2).

3.1.3 Pichoux limestone (BOLLIGER & BURRI 1970, p. 71)

The Pichoux limestone is a succession of well-bedded lime mudstones transitional between the carbonate platform deposits of the St-Ursanne Formation with corals and the Birmenstorf Member (see 3.1.8) with abundant ammonites and siliceous sponges in the "basin" (Pl. 1A). A marly intercalation with a maximum thickness of several meters divides the lower part of the Pichoux limestone from the upper part (unit 2 in Fig. 6). There are several sections like no 6 in Plate 1A (Gorges du Pichoux) where the marginal part of the upper St-Ursanne Formation rests on Pichoux limestone. H. and A. Zbinden found the *Perisphinctes (Dichotomosphinctes) antecedens* SALFELD J 27994 in the quarry of La Charuque near Péry 2.5 m above the base of the Pichoux limestone.

3.1.4 Vellerat Formation (BOLLIGER & BURRI 1970, p. 71)

The Vellerat Formation is a succession of limestones and marls from the shallow subtidal to the supratidal zones. As a consequence, both vertical and lateral facies changes are very pronounced. The formation is above the platform interior facies of the St-Ursanne Formation. The average thickness of the Vellerat Formation is 55 m. The lower part of the Vellerat Formation passes above the margin of the St-Ursanne Formation laterally into the coral biolithites and calcarenites of the Günsberg Member (Pl. 1A). The upper part of the Vellerat Formation becomes mostly limestone further in the distal direction. The thickness of the formation increases substantially with the facies transition (Pl. 1A). Both the lower and the upper boundary of the formation may be a transition.

Four members can be discerned, from base to top: 1) The Vorbourg limestone Member, 2) the marl-limestone succession of the Natica Member, 3) the Hauptmumienbank limestone Member, and 4) the Humeralis marl/Oolithe rousse couplet. A single ammonite was found in situ in the Vellerat Formation, in a small outcrop of the Natica Member transitional to the Günsberg Member near Seewen (14 km SSE of Basel, BITTERLI-BRUNNER et al., in prep.). It is the *Perisphinctes (Perisphinctes) panthieri* ENAY J 27257 of the Bifurcatus Zone. This means that part of the Natica Member, the Günsberg Member, and of the Effingen Member have the same age (Table 2).

Vorbourg Member (M. A. ZIEGLER 1962, p. 21)

The name was first proposed by E. GREPPIN (1893, p. 16) for the lower part of what is now the Reuchenette Formation (sequences 4 to 6 in our Pl. 1B). Greppin's name was not taken up by later authors. M. A. ZIEGLER (1962, p. 21) introduced the name once again, but with a different meaning and without mention of Greppin. The Vorbourg limestone of M. A. ZIEGLER (1962, p. 22) includes in the type section the uppermost massive part of the St-Ursanne Formation. H. FISCHER (1965, p. 17) drew attention to the fact that the Vorbourg limestone of M. A. ZIEGLER also includes part of the Natica Member. Modern authors use the name Vorbourg Member for the well-bedded succession of mostly micritic, "pure" limestones between the massive St-Ursanne Formation below and the marly Natica Member above.

Fenestrate stromatolites and oncolitic horizons may occur at any level from the base to the top of the Member. Prism-cracked beds and tidal channels were only found from the middle part to the top. The thickness of the Vorbourg Member increases from the margin of the carbonate platform of the St-Ursanne Formation below towards the platform interior. The maximum thickness of the Vorbourg Member is about 14 m. The member grades laterally and in the distal direction into bioclastic and oolitic limestone with coral bioherms (Pl. 1A). Small, cross-bedded oolitic sand bars exist in the platform interior near Courtemaîche south of Boncourt (see Fig. 1).

Natica Member

An adequate definition of the Natica Member does not exist. However, the name is well established in the regional geologic literature. Most authors used the notion as illustrated by P. A. ZIEGLER (1956, Fig. 14). The Natica Member encompasses the highly variable strata between the top of the Vorbourg Member below and the base of the Hauptmumienbank Member above.

Marl predominates in the Natica Member above the platform interior facies of the St-Ursanne Formation. There, the member weathers out as a terrace when strata are horizontal, or as a depression when strata are tilted. Fine-grained detrital quartz may be very abundant in marl and limestone of the Natica Member (ROLLIER 1898, p. 58). The average thickness of the Natica Member is about 35 m. The upper Natica Member progrades over the small carbonate platform of the Günsberg Member (Pl. 1A).

Intertidal to supratidal facies of marl and limestone apparently form a continuous horizon in the upper part of the member. M. A. ZIEGLER (1962) found marls with abundant limnic ostracods and characean oogonia in the Natica Member. We found a thin coal seam in an excavation on the roadside in the southern part of the gorge of Moutier, 12.1 m below the base of the Hauptmumienbank Member. The maximum thickness of the seam is only 3 cm. However, this must be more than a local occurrence since coal lenses in about the same stratigraphic position were reported by KEMMERLING (1911, p. 22) from the opposite side of the gorge, and by LAUBSCHER & PFIRTER (1984, p. 208) from nearby Roches. The thin coal seam reported by HEER (1865, p. 125) from near Pfeffingen (10 km south of Basel) has probably about the same age. The upper Natica Member includes stromatolites of lime mudstone, some of them prism-cracked (Fig. 4), and black pebble conglomerates (Fig. 5). Small tidal channels are cut in pelletoid wackestones. A hardground, sometimes developed on a planed erosion surface with large boreholes and ostreids (Fig. 5), can be observed on the surface of the peritidal limestones of the upper Natica Member in many sections.

Above the peritidal horizons and below the base of the Hauptmumienbank Member is a succession of marls and limestones with a fauna indicative of seawater with a normal salinity. Local coral biostromes and oolite are laterally replaced by marl (Pl. 1A). The best-exposed coral biostrome is in the Gorges de Court west of Moutier (horizon 15 in section 49 by P. A. ZIEGLER 1956, p. 93). BOLLIGER & BURRI (1970, p. 72) gave it the ambiguous name Moutier-Korallenkalk. The marine sediments of the upper Natica Member have an average thickness of less than 10 m. They are the lower part of a small-order shallowing-upward sequence (no 2b in Pl. 1A).

Hauptmumienbank Member (P. A. ZIEGLER 1956, p. 42)

Hauptmumienbank means main mummy bed. STEINMANN (1880, p. 152) called an oncoid a mummy when the core is a recognizable shell. The core of Oxfordian shallow-water oncoids in our region is typically a small nerineid gastropod or an ostreid bivalve shell. The oncoid diameter in sediments of the tidal or of the shallow subtidal zone is less than 6 cm. GASCHE (1956) recognized that the crusts of oncoids in the Oxfordian horizons as studied by P. A. ZIEGLER (1956) were formed by algae. The name Hauptmumienbank implies that there may be more than one oncolitic horizon in a given section. The Oxfordian Hauptmumienbank has by far the greatest geographical range of all oncolitic horizons in northern Switzerland. It is an excellent marker bed for mapping at the scale of 1:25,000. GRESSLY (1864, p. 99) and all mapping geologists after him were aware of this.

We use the name Hauptmumienbank provisionally for the whole limestone succession between the marls and the marly limestones of the Natica Member below and the Humeralis marl or Oolithe rousse above, even where only part of the succession contains oncoids. This limestone member is normally less than 10 m thick. It was a thin but widespread carbonate platform. The succession is the upper, "pure" limestone part of the shallowing-upward subsequence 2b in Plate 1A. The limestone succession of the platform interior weathers out as an escarpment when beds are horizontal, or as a prominent ridge when the succession is tilted. The upper boundary of the Hauptmumienbank may locally be a planed erosion surface, for instance at Roches north of Moutier, on the road from Hautes Roches to the ancient farm Le Trondai.

The facies of the member in the platform interior is thick-bedded lime mudstone with few and small oncoids in the upper part of the succession. Pseudomorphs of calcite after acicular crystals of probably calcium sulfate are common in the core of these oncoids. Fossils other than oncoids are rare. Further in the distal direction, oncoids become common from the base to the top of the member. These oncoids have typically a diameter of 1 to 3 cm and from 20 to 40% of the rock by volume. They float in a matrix of biomicrite. Towards the platform margin, the member becomes oolitic first at the base, then further up until it is all oolite (Table 2). The oolitic transitional facies of the Hauptmumienbank is exposed for instance in the Gorges de Court, where BOLLIGER & BURRI (1970, Pl. 12, Fig. 1–2, and Pl. 16, section 4, sample 1301) have confounded the Hauptmumienbank Member with their "Hautes Roches Algenkalk".

We now infer that the equivalent of the Hauptmumienbank in Mt. Weissenstein north of Solothurn and in Mt. Harzer north of Welschenrohr is a white oolite with a low restistance to weathering. It forms a marked depression on the flanks of these mountains. Therefore, this soft oolite has been interpreted as Verena oolite by BUXTORF (1907, p. 54), P. A. ZIEGLER (1956, section 52, horizon 23, p. 96), and by GYGI (1969, Pl. 19, section 2). It is probable that the oôlite passes laterally and in the distal direction into the Steinibach Beds as named and defined by GYGI (1969, p. 85).

The value of the Hauptmumienbank as a marker bed has been questioned mainly because there are several superimposed oncolitic horizons in the Oxfordian of northwestern Switzerland. The oldest oncolite appears in the lower St-Ursanne Formation (PÜMPIN 1965, Fig. 6–7). Another, local, horizon is in the uppermost St-Ursanne Formation, for instance near Leymen, France, 10 km southwest of Basel. There are local oncolitic horizons in any level of the Vorbourg Member and of the Natica Member. Above the Hauptmumienbank are oncolitic horizons in different levels of the Humeralis limestone. These are the "Hautes-Roches-Algenkalk" of BOLLIGER & BURRI (1970, p. 74), or "akzessorische Mumienbänke" of P. A. ZIEGLER (1956, p. 42). Ziegler's name is better since these local oncolitic beds can be followed at best over a distance of a few kilometers. They are indeed accessory. We use the name by Bolliger and Burri provisionally even though it is misleading, because it is this name that was used by later authors. In the Verena Member, oncoids occur only as scattered particles.

Most of the subordinate oncolitic horizons mentioned above were previously observed by H. FISCHER (1965, p. 22) in the region southwest of Basel. The Hauptmumienbank oncolite is unique in northern Switzerland because it can be followed over a distance of at least 100 km from the village of Blauen 12 km SSW of Basel as far as canton Neuchâtel. This is not surprising, since similar widespread oncolites were found in the Kimmeridgian of southeastern France (ENAY 1966) and in the Kimmeridgian of southern Poland (KUTEK 1968).

Humeralis marl and Oolithe rousse

The Humeralis marl and its lateral equivalent, the Oolithe rousse, are the uppermost components of the Vellerat Formation. The Humeralis marl grades upwards into the marly, micritic limestones of the platform interior facies of the Court Formation. Laterally, the marl passes into the Oolithe rousse and progrades over it (Pl. 1A). The lateral boundary between the Oolithe rousse and the Balsthal Formation is not yet exactly known.

3.1.5 Court Formation (BOLLIGER & BURRI 1970, p. 73)

In the platform interior, the Court Formation begins with marly, micritic limestones. They are well-bedded and become increasingly pure towards the top. This is the so-called Humeralis limestone. The uppermost part of the Court Formation in the platform interior is a massive, almost white limestone which has normally a high porosity and a low resistance to weathering (Pl. 1A, Table 2). The average thickness of the massive limestone is about 15 m. TSCHOPP (1960, p. 9) and others have called this limestone "Bank A". In the distal direction, the lower Court Formation may become a well-bedded, low-porosity lime mudstone, an oncolite, an almost white, dedolomitized oolite with ill-defined bedding planes at large intervals (facies of the Verena Member, see below), or a brownish oolitic packstone with some oncoids. Brown bands of coarsely crystalline dolomite or dedolomite are uncommon, and so are thin intercalations of marl. The "Hautes-Roches-Algenkalk" oncolite of BOLLIGER & BURRI (1970, p. 74) is not a mappable unit. The Verena Member will be described with the Balsthal Formation.

One large unidentified perisphinctid ammonite has been found in the lowermost Court Formation several years ago at Dittingen BE, in one of the Schachlete quarries north of Laufen. At present, this formation cannot be directly related to any ammonite zone.

3.1.6 Balsthal Formation (GYGI 1969, p. 83)

A monotonous limestone succession of oolite with few intercalations of marl, marly limestone, oolite with hermatypic corals, or oncolitic limestone, forms the upper part of the Oxfordian Stage in a belt about 20 km wide (Pl. 1A). The thickness of the oolitic succession is normally in excess of 100 m where the lower and the middle Günsberg Member is present. Beyond the basinward margin of the small carbonate platform of the lower and middle Günsberg Member, the thickness of the Steinibach Beds and the Holzflue Member combined may increase to more than 100 m. This abnormally great thickness has been found in a section west of Balsthal as measured by GYGI (1969, Pl. 19, section 15). The exceptional thickness has been omitted from Plate 1A because it is probably restricted to a belt only a few kilometers wide. The Balsthal Formation can be divided into four units:

Günsberg Member (GYGI 1969, p. 83)

A limestone belt 10 to 20 km wide with coral bioherms mainly at the base and oolite above separates the marly Natica Member from the coeval part of the marly Effingen Member (Pl. 1A). GYGI (1969, p. 83) proposed to name this almost pure limestone succession Günsberg Member (Table 2). He included the oolitic Steinibach Beds, which replace the Hauptmumienbank distally, as the uppermost part in the Günsberg Member. The lower and the middle

Günsberg Member are laterally replaced east of Günsberg by the Effingen Member. The "type" section of the Günsberg Member is within the area of this facies change.

Steinibach Beds (GYGI 1969, p. 85)

The Steinibach Beds are an oosparitic calcarenite, usually with inclined bedding. The thickness is between 6 and 15 m. The unit is in most sections above marl of the Effingen Member with or without hermatypic corals. There are neither ammonites nor mineralostratigraphic data from the Steinibach Beds proper. We infer from the mineralostratigraphic correlations H and I in the section of Péry (Fig. 10) that the Steinibach Beds are the distal equivalent of the Hauptmumienbank Member. An unusual transition facies between the oncolitic Hauptmumienbank Member and the oolitic Steinibach Beds exists near Péry on the western wall of the cluse of Rondchâtel, where the unit is an oolitic wackestone to packstone with peloids and small oncoids at the top. The bedding is inclined, with a depositional dip of the foresets of about 20° towards the WNW. Calcarenites with inclined bedding have normally a grainstone texture. The thickness of the unit in the section of Péry is 5.7 m, with the base being 232 m above the Callovian/Oxfordian boundary (GYGI 1982, Fig. 6).

"Hautes-Roches-Algenkalk" and Verena Member

Above the Steinibach Beds is a succession of oolitic or micritic limestone with oncolitic intercalations. Most of the accessory oncolites are, unlike the light Hauptmumienbank, from brown to petroleum green or grey. This is what BOLLIGER & BURRI (1970) called Balmberg-Oolith and Hautes-Roches-Algenkalke. The Verena Member above is a yellowish-white, massive limestone which forms as a rule the highest crest where Oxfordian and Kimmeridgian limestones are steeply dipping or vertical. The dominant facies of the Verena Member is oolite with mostly micritized ooids and some oncoids (GYGI 1969, Pl. 13, Fig. 47). The rock has a complicated diagenetic history of dolomitization, replacement probably by calcium sulfate in small patches, and dedolomitization of the whole rock except occasional small relicts of anhedral dolomite (GYGI 1969, p. 78). The primary oolitic texture was blurred in the process. The micritized ooids and small oncoids are well visible only on clean, weathered surfaces, but they are difficult to discern on a freshly broken surface. Large pockets of the rock as much as 10 or 20 m across may be porous and weather out as hollows and caves, as for instance at the St. Verena chapel, the type locality of the Verena Member near Solothurn. Such pockets occur in any level of the Verena Member which has an average thickness of about 45 m. The misinterpretation of the Verena Member by GYGI (1969, p. 86) is caused by the erroneous correlation with a local facies of white and porous oolite which is probably time-equivalent with the Hauptmumienbank (see above). Unaltered, cross-bedded oolite is uncommon in the Verena Member. It is less resistant to weathering than the massive, dedolomitized facies. Both facies occur side by side on the eastern wall of the Gorges de Moutier where the massive facies weathers out like bioherms.

Holzflue Member (GYGI 1969, p. 86)

The Verena Member was interpreted by most authors as above (see BOLLIGER & BURRI 1970, p. 74). It forms a belt which is palinspastically about 40 km wide (Pl. 1A). The member thus defined has the disadvantage that the Verena facies may appear already in the heterogenous complex of oolitic, micritic, or oncolitic limestones below the Verena Member, whereas there is almost pure micrite also in the Verena Member itself (GYGI 1969, Pl. 19, section 15, TSCHUMI 1983, and MARTIN 1984). The base of the Verena Member cannot be satisfactorily defined, at least not east of the meridian of Solothurn. East of Günsberg, more or less pure lime mudstone intercalations become more and more numerous in the Verena Member until the whole unit is mudstone west of Olten. GYGI (1969, p. 86) proposed the name Holzflue Member for the whole limestone complex from the top of what he then regarded to be the Hauptmumienbank to the base of the Reuchenette Formation in order to provide for a convenient mapping unit (Table 2). The Verena Member grades also towards the platform interior into atypical, bedded lime mudstone (Pl. 1A). The uppermost 12 to 15 m of these mudstones of the platform interior are massive and chalk-like ("Bank A" of TSCHOPP 1960, and others, see above). There is an uneven erosion surface about 18 m below the top of the Verena Member in the limestone quarry of La Reuchenette near Péry (section 7 in Pl. 1A). Above are angular and rounded blackened lithoclasts as much as 10 cm across.

Few ammonites have been found in situ in the Balsthal Formation. *Perisphinctes (Dichotomoceras) bifurcatus* (QUENSTEDT) J 30935 from the platform margin facies of the lower Günsberg Member of Günsberg (GYGI 1969, p. 99) and an *Orthosphinctes (Lithacosphinctes) evolutus* (QUENSTEDT) J 30530 found about 10 m above the base of the Reuchenette Formation near Balsthal by P. Tschumi and B. Martin indicate that the Balsthal Formation begins in the Bifurcatus Zone and ends at about the boundary between the Oxfordian and the Kimmeridgian Stages.

3.1.7. Reuchenette Formation (THALMANN 1966, p. 32)

The Reuchenette Formation was defined by THALMANN (1966, p. 36) in the limestone quarry of La Reuchenette near Péry BE. There, it is a monotonous succession of bedded limestones with few and thin intercalations of marl. Lime mudstone is the dominant facies in the area considered, but peloidic wacke- to grainstones and some oolitic horizons are also major components of the formation. Coral biostromes are uncommon. The best section of the formation is in the Gorges du Pichoux near Sornetan (section 6 in Pl. 1B). The average total thickness is about 140 m. Only the lower part of the formation is represented in plate 1B. The mean thickness of this now unnamed member (the former Couches du Vorbourg of E. GREPPIN 1893, p. 16) between the Verena Member below and the Banné marl (Table 2) above is 45 m. The Banné marl is named after the hill Le Banné 1 km south of Porrentruy (Fig. 1). The Banné marl either wedges out or grades into limestone to the south (Pl. 1B).

The base of the Reuchenette Formation as defined by THALMANN (1966, Fig. 5) at La Reuchenette is a horizon with blackened lithoclasts on top of an uneven erosion surface. Above is a massive limestone 18 m thick. The lower 8 m of this massive unit are Verena facies with an oncolitic bed. The upper 10 m are mostly mudstone with occasional patches of oolitic wackestone. Above this massive limestone begins a succession of well-bedded mudstones and peloidic wacke- to grainstones with two bands of fenestrate stromatolites (GYGI 1982, Fig. 6). The boundary between massive and well- and normally thick-bedded limestone is conspicuous, and it can be observed in all sections of the lowermost Reuchenette Formation, whereas the horizon with blackened lithoclasts is restricted to only a part of the quarry of La Reuchenette. This is why we draw the line between the Court or Balsthal Formation and the Reuchenette Formation at the base of the well-bedded limestone succession. Fenestrate stromatolites occur in many sections in the lowermost Reuchenette Formation a few meters above the base. Stromatolites with very well-visible prism cracks are 2 m below the Banné marl in a section about 1 km southwest of Glovelier along the road to St-Brais (5 km SSE from St-Ursanne, see Fig. 1). Neither stromatolites nor glauconitic horizons nor coral biostromes can be used to correlate sections of the Reuchenette Formation in detail. Nor can be marls like the Banné marl, because the Reuchenette Formation is a pure limestone succession between Sornetan and Péry, with only some very thin marl seams. Three parallel facies belts can be discerned in the lowermost member of the Reuchenette Formation in the area considered here.

The first facies belt is cut across between Glovelier and the region of Porrentruy in the Ajoie (Pl. 1B). The macrofauna of this facies belt is from marginal marine to marine with mostly bivalves and gastropods. The faunal diversity is low, but the number of individuals per species may be unusually great in the Banné marl and in a horizon about 15 m below. Coral biostromes are few and areally restricted. The coral biostrome of the Banné hill, now well exposed 25 m above the floor of the quarry west of La Rasse between Porrentruy and Fontenais, was figured by GRESSLY (1840, Pl. 9), and correlated with the Oxfordian St-Ursanne Formation of Pont d'Able.

Cephalopods like the large nautiloid *Paracenoceras giganteum* (D'ORBIGNY) J 22778 from the slightly glauconitic horizon below the Banné marl (Pl. 1B) are rare. Ammonites are very rare. *Aspidoceras* cf. *acanthicum* J 30714 was taken by H. and A. Zbinden from a block which fell presumably from a marly limestone 1.5 m below the Banné marl. *Alveosepta* are the most conspicuous foraminifers.

The second facies belt, between Sornetan and Péry, is made up almost exclusively of limestone. There are generally very few macrofossils in the lowermost 20 m of the formation. Fenestrate stromatolites and tidal channels a few meters above the base of the Reuchenette Formation indicate a tidal flat environment. Two large tidal channels are visible for instance in the small gorge southeast of the St. Verena chapel near Solothurn (Fig. 1). The lime mud was firm enough to support large land animals like the dinosaur *Cetiosauriscus greppini* (VON HUENE) which has been found in a small quarry about 1 km northeast of Moutier (J. B. GREPPIN 1870, Pl. 1). A coral biostrome appears near Seehof BE (northwest of Welschenrohr) in about the same level above the formation base as the coral biostromes of Porrentruy and Glovelier (Pl. 1B). Marly horizons reappear about 35 m above the base of the Reuchenette Formation in the ancient quarries northeast of Solothurn. Hundreds of turtles were found in these quarries, mainly in one distinct, marly horizon (LANG & RÜTIMEYER 1867, p. 3 and 12). Limnic ostracods, nerineid gastropds, and *Pseudocyclammina (= Alveosepta?)* are in the same marly limestone as the turtles (THALMANN 1966, p. 105). One or two meters above the horizon with turtles (as inferred by THALMANN 1966, p. 104), or about 40 m above the base of the Reuchenette Formation, is the horizon of an *Aulacostephanus (Pararasenia) quenstedti* DURAND. The vertical range of this taxon is from the Acanthicum Zone to the Eudoxus Zone (B. ZIEGLER 1962, p. 130). This is evidence that the uppermost marly horizon in the quarries near Solothurn is about time-equivalent with the Banné marl, and that the lowermost member of the Reuchenette Formation below the Banné marl becomes thinner towards the "basin". The thickness of the Reuchenette Formation near Solothurn would then be only somewhat more than 40 m instead of the 188 m as stated by BUXTORF (1907, p. 58).

A third, fully marine facies belt of the lower Reuchenette Formation crops out between Balsthal and Olten. TSCHUMI (1983) and MARTIN (1984) have found hermatypic corals and an *Orthosphinctes (Lithacosphinctes) evolutus* (QUENSTEDT) J 30530 just above the base of the formation near Balsthal (Pl. 1B). The ammonite taxon appears in southeastern France in the Galar Subzone of the Oxfordian and continues into the Platynota Zone of the Kimmeridgian (ATROPS 1982, p. 131, 323). In northern Switzerland, the taxon has only been found in the Platynota Zone near Schaffhausen so far. The hermatypic corals therefore appeared near Balsthal about at the turn from the Oxfordian to the Kimmeridgian. Small coral bioherms are visible directly below Alt Falkenstein castle at Balsthal. Ammonites become fairly abundant from Oberbuchsiten to the east. Almost all of them are from the Hypselocyclum Zone (GYGI 1969, Pl. 18, section 21). The transition of the Reuchenette Formation to the "basinal" facies is not known, since the whole of the formation as well as part of the Upper Oxfordian strata below were eroded between Aarau and Wildegg in pre-Eocene time (Table 2).

3.1.8 Wildegg Formation (GYGI 1969, p. 64)

The Wildegg Formation is the time equivalent in "basin" facies of the upper Terrain à Chailles, the Liesberg Member, the St-Ursanne Formation, and of the Natica Member (Table 2). The Wildegg Formation begins above a hiatus (Table 2). The formation normally rests on the thin, iron-oolitic Schellenbrücke Bed of the Early Oxfordian. This marker bed thins out to zero near Wildegg on both sides of the Aare river. The Middle Oxfordian Wildegg Formation rests near Auenstein (section 9 in Pl. 1A) and near Holderbank on a thin iron-oolitic horizon with *Tulites (Rugiferites) polypleurus* (S. BUCKMAN) of Middle Bathonian age. The important hiatus in between is marked by an uneven limonite crust with a shining surface and a thickness of about 1 mm (Table 2). The age of the crust could be determined to be of the late Cordatum Subchron by an Oxfordian ammonite which was found in a pocket under the crust (GYGI & MARCHAND 1982, Fig. 2, section 39, and Pl. 9, Fig. 1). GYGI (1969, Pl. 2, Fig. 4) regarded the thin iron-oolitic bed of the Middle Bathonian near Auenstein, then known of a borehole only, to be the equivalent of the Oxfordian Schellenbrücke Bed. The oldest bed of the Wildegg Formation in canton Aargau, the condensed bed at the base of the Birmenstorf Member (Table 2), is only about 5 cm thick. Ammonites of the Densiplicatum Chron and of the Antecedens

Subchron were found in this horizon (GYGI & MARCHAND 1982, Pl. 12, Fig. 2 and 3). The horizon is the time equivalent of the upper Terrain à Chailles Member, of the Liesberg Member, and part of the St-Ursanne Formation.

GYGI (1969, p. 64) included in his Wildegg Formation the Birmenstorf Member at the base, the Effingen Member in the middle, and the Geissberg Member at the top. He interpreted the Geissberg Member to be a regional calcareous facies of the uppermost marly Effingen Member. We now conclude from the mineralostratigraphic correlations H and I (Pl. 1A) that this is not so: the Geissberg Member is time-equivalent with the Steinibach Beds at Péry (Pl. 1A). Where the Upper Oxfordian limestones form a cliff in canton Aargau, the Geissberg Member is the lowest part of it. Therefore, we now include the Geissberg Member into the Villigen Formation (Table 2) in order to make the formation convenient to map.

Birmenstorf Member (MOESCH 1863, p. 160)

The upper, non-condensed part of the Birmenstorf Member is the thin, "starved basin" equivalent of the upper St-Ursanne Formation. The member is a biostrome of siliceous sponges with abundant ammonites. Small sponge bioherms are exceptional (GYGI 1982, p. 25 and Fig. 6). All sponges are calcified. They are in a matrix of siliciclastic-dominated or lime mudstone. Beds of marl alternate with limestone bands (GYGI 1969, Pl. 17, section 60). The average thickness of the member is about 5 m. The thickness is greatest at the transition to the Pichoux limestone. From there, the thickness decreases to less than 1 m in eastern canton Aargau and in canton Schaffhausen (GYGI 1977, Table 2 and Pl. 11). The thickness of the Birmenstorf Member had to be exaggerated in Plate 1A. Details of the sedimentology, paleoecology, and ammonite zonation of the Birmenstorf Member are given by GYGI (1969, 1977) and GYGI & MARCHAND (1982).

Sedimentation of the Birmenstorf Member began in the Densiplicatum Chron. The Densiplicatum Zone was redefined by GYGI & MARCHAND (1982, p. 534–5). The end of deposition of the Birmenstorf Member coincided with the end of the Transversarium Chron (GYGI 1977, p. 511). The upper boundary of the Birmenstorf Member is a transition (see discussion in GYGI 1969, p. 65). The abundant and diverse ammonite fauna of the Birmenstorf Member in northern Switzerland was used by Oppel to define the Transversarium Zone (see GYGI & MARCHAND 1982, p. 534, for an emendation of the original definition of the zone).

Effingen Member (MOESCH 1857, p. 55)

Most of the Effingen Member is composed of clinothems (RICH 1951), or what TURCOTTE & KENYON (1984, Fig. 1) called sigmoidal progradational clinoforms. We know from ammonites that in the section of Péry (no 7 in Pl. 1A), the entire Effingen Member is of Bifurcatus age (BAYER et al. 1983, p. 129), whereas east of Auenstein, probably less than the lower half of the member is of the same age. The upper part of the Effingen Member in the Auenstein section has been deposited in the Hypselum Subchron (see below). The Effingen Member is mostly made up of marl with a wide range in carbonate content (GYGI 1969, Pl. 17, section 37). Successions of marly to "pure" limestone are intercalated in the marl. None of these limestone successions except the Gerstenhübel Beds can be followed laterally far enough that it could be used for correlation.

There is a low-angle intraformational truncation surface in the upper Effingen Member in the quarry of Jakobsberg near Auenstein (Fig. 2). The bed above the surface is a debris flow deposit 20 cm thick. This is bed 98 of section 37 in GYGI (1969, Pl. 17). Another such bed is horizon 102 in the same section with large, plastically deformed clasts of laminated, silty pelsparite (Fig. 3). We infer that the thin beds with parallel laminae or small ripples of peloidal quartz silt as figured by GYGI (1969, Pl. 4, Fig. 12) are the distal turbidite facies of the small debris flows mentioned above. MEYER (1984) concluded for one of these beds with abundant asteroids (starfish) that it had a multiphase origin and that the rippled parts of the bed were tempestites. We have doubts whether sediment redeposition by storms in relatively deep water can be rapid enough to cause the death of a whole population of asteroids by smothering as has been observed by Meyer. There is reason to assume that the water depth at the site was greater than 100 m (see below). This bed with asteroids is located west of Güntberg (Fig. 1), about 6 m above the base of the Effingen Member. It is at the toe of a clinothem in the slope facies of the lowermost Effingen Member (Table 2 and Pl. 1A).

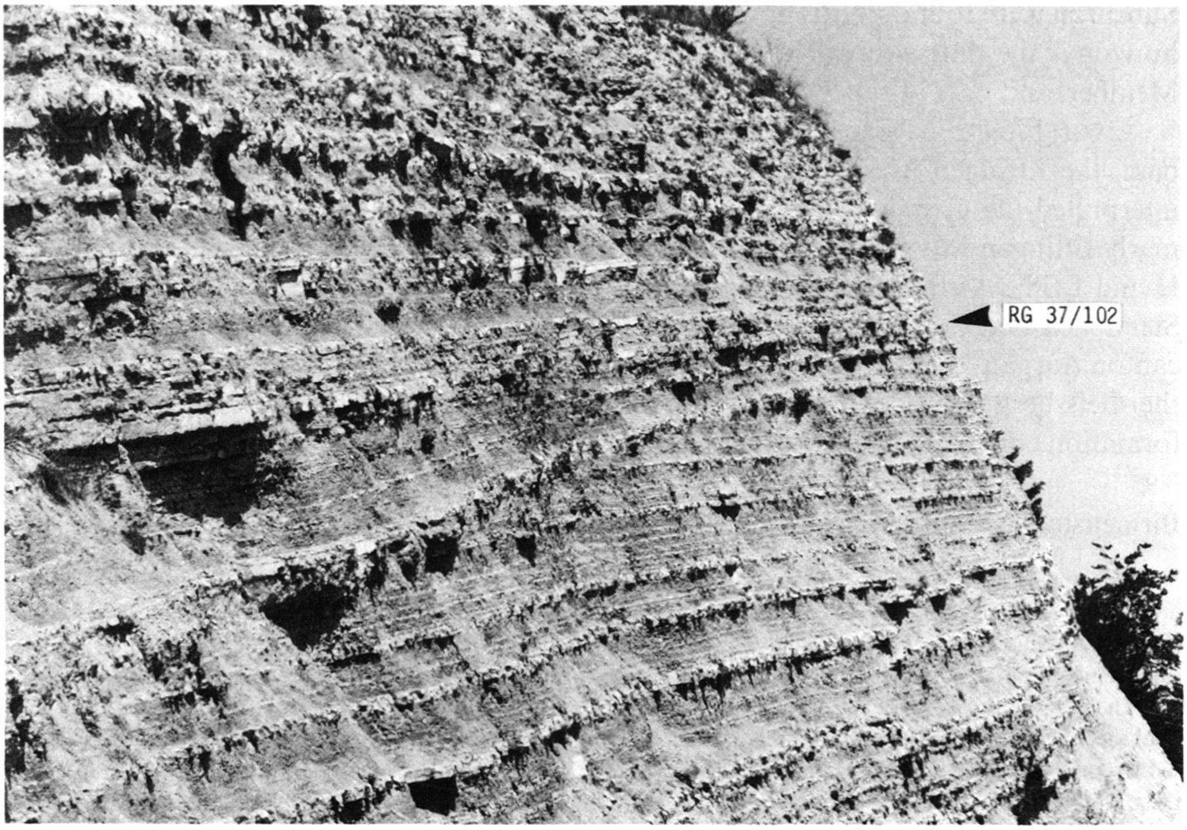


Fig. 2. Intraformational truncation surface in the upper Effingen Member of the Wildegg Formation, below bed no 102 of section 37 in GYGI (1969, Pl. 17). Jakobsberg quarry near Auenstein AG.

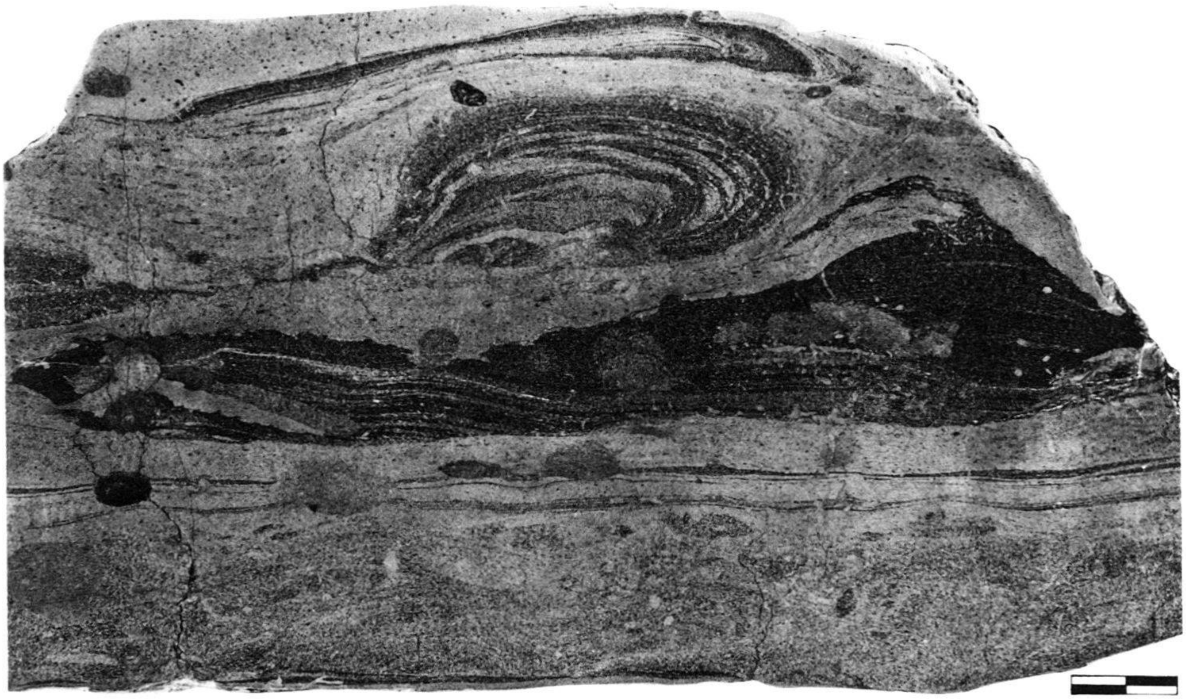


Fig. 3. Submarine debris flow deposit, possibly set off by a growth fault. Polished slab of marl-limestone, bed no 102 of section RG 37, upper Effingen Member, above the truncation surface represented in Figure 2. Some of the large, cross-laminated pelsparite clasts were plastically deformed, then burrowed. The light laminae have a high content of silt-grade detrital quartz. Bar is 2 cm.

Gerstenhübel Beds (GYGI 1969, p. 66)

The Gerstenhübel Beds are a succession of "pure" limestone beds about 12 m thick. This can be mapped in most of the Aargau Jura (GYGI 1969, Pl. 19). Sedimentary structures in the individual limestone horizons of the Gerstenhübel Beds in the Auenstein and Holderbank sections are much like those as figured by GYGI (1969, Pl. 4, Fig. 4) from another bed in the Effingen Member with plastically deformed mud clasts. The Gerstenhübel Beds of the Auenstein and Holderbank quarries are now thought to be debris flow deposits in the sense of MIDDLETON & HAMPTON (1973, p. 20). The average depositional slope of the Effingen Member adjacent to the carbonate platform of the lower and middle Günsberg Member must have been less than 1°, to judge from Plate 1A. The slope after compaction and load-induced basement subsidence was even less.

Ammonites are fairly abundant in the lowermost 15 to 20 m of the Effingen Member. They are from the Schilli Subzone and from the Bifurcatus Zone s. str. above (Pl. 1A). Further above, ammonites are rare. *Amoeboceras* Gy 2403 was found by J. Haller and R. Trümpy in the Gerstenhübel Beds near Mönthal in canton Aargau. The specimen is adult and near-complete. The ribs, tubercles, and the keel are like on inner whorls of *Amoeboceras serratum* (J. SOWERBY), but the diameter of our specimen is only 19 mm as compared with the more than 100 mm of the neotype selected by SYKES & CALLOMON (1979, Pl. 117, Fig. 1). Provided we rightly compare *Amoeboceras* Gy 2403 and other specimens from northern Switzerland with *Amoeboceras serratum*, then the age of the Swiss material is either of the late Bifurcatus Chron or of the early Hypselum Subchron (see Table 3).

3.1.9 Villigen Formation (GYGI 1969, p. 68)

The Villigen Formation is a well-bedded succession of mostly micritic, "pure" limestones with few and thin marly intercalations in the lower part (GYGI 1969, Pl. 17, section 62). The ammonite succession of the formation is discussed by GYGI (1969, p. 98–99). There are five parts.

Geissberg Member (MOESCH 1857, p. 57)

The Villigen Formation begins with the Geissberg Member. The base of this member is a transition. The boundary is where the marls and marly limestones of the upper Effingen Member become sufficiently calcareous to form a cliff. The Geissberg Member is a succession of mostly thick-bedded calcilutites with marly intercalations. The fauna is dominated by bivalves. Among the few ammonites are large perisphinctids (*"Decipia", or Lithacosphinctes*, respectively). No ammonites diagnostic of a subzone have yet been found.

Crenularis Member (MOESCH 1863, p. 157)

The Crenularis Member is made up of bedded, biomicritic limestone. The normal thickness is less than 5 m. Glauconite, corroded bedding planes, and an abundant macrofauna are indications of a reduced rate of sedimentation. Siliceous sponges are common. They form major bioherms near Mellikon (Pl. 1A) and near Baden. Several specimens of the ammonite *Epipeltoceras bimammatum* (QUENSTEDT), the index taxon of the Bimammatum Zone, were found in the member along with several large *Ringsteadia*. The Crenularis Member grades laterally and in the distal direction into the upper part of the Hornbuck Member (Pl. 1A). The mineralostratigraphic correlation I is in the lowermost Crenularis Member.

Wangen Member (MOESCH 1867, p. 162)

The micritic, well-bedded limestone of the Wangen Member has normally a low porosity and a thickness between 5 and 10 m. Near Villigen, the upper part of the member is a white, porous, and very pure limestone. Ammonites are few in the Wangen Member. The member grades laterally into the much thicker Küssaburg Member of canton Schaffhausen, where several specimens of *Taramelliceras litocerum* (OPPEL) have been found (Pl. 1A).

Knollen Beds (MOESCH 1863, p. 163).

The Knollen Beds are normally a micritic limestone only 1 or 2 dm thick. The bedding planes are often marly or corroded and encrusted with iron hydroxide. Abundant siliceous sponges, brachiopods, bilvalves, and some glauconite indicate a low rate of deposition. Siliceous sponges form bioherms several meters high in the quarry near Mellikon. The Knollen Beds are a glauconitic marker horizon which can be followed over a distance of more than 120 km from east of Olten through canton Aargau and canton Schaffhausen into the Swabian Alb of southern Germany. The mineralostratigraphic correlation J coincides with this lithologic marker horizon.

Letzi Member (MOESCH 1863, p. 165)

Most of the Letzi Member is well-bedded and "pure" micritic limestone. The member grades laterally into the Wangental Member of canton Schaffhausen. There is glauconite in the uppermost Letzi Member near Mellikon and in the uppermost Wangental Member near Schaffhausen (GYGI 1969, Pl. 16 and 17). Ammonites are somewhat better represented in the Letzi Member than in the Wangen Member. They are of the Planula Chron (GYGI 1969, Fig. 5).

Idoceras occur only in the uppermost part of the Planula Zone near Mellikon (GYGI 1969, Fig. 5). *Sutneria galar* (OPPEL) is present in the uppermost few meters of the Letzi Member near Mellikon, and it is abundant in the uppermost Wangental Member near Schaffhausen. Calcareous nannoplankton is very rare or absent in the lower Villigen Formation of canton Aargau. Some coccoliths have been found in the Letzi Member (GYGI 1969, Fig. 1). The relative rarity of nannoplankton in the Villigen Formation is not due to recrystallization of the fine-grained limestones. It is probably a primary feature.

3.1.10 Baden and Wettingen Members

There is no formation name for the strata above the Letzi Member.

Baden Member (MOESCH 1863, p. 165)

The Baden Member does not crop out any more at the type locality, the southern end of the road tunnel in the centre of Baden. 550 m to the west, the lower part of the member is a succession of marls and marly limestones (GYGI 1969, Pl. 17, section 47). Marl predominates at the base, and limestone at the top. Most of the limestone beds are glauconitic and have a rich macrofauna which is dominated by ammonites. The thickness of the incomplete section of the member at Baden is 2.5 m. Complete sections are near Villigen and Mellikon (GYGI 1969, Pl. 17, sections 62 and 70). There, the lower part of the member is a massive, marly limestone with glauconite, somewhat more than 1 m thick. Above is a soft, yellow-brown marl with almost no ammonites. There is a small sponge bioherm in this bed in the large quarry west of Mellikon. The ammonites in the lower, glauconitic part of the member are from the Platynota, the Hypselocyclum, and from the Divisum Zone (Table 2). A marked facies change takes place in the Baden Member about 5 km east of Baden, and about 3 km east of Mellikon. The thickness increases, and the marly limestone grades into marl with some intercalations of marly limestone. East of the facies boundary only the lowermost 1.5 to 2.5 m of the member contain beds of "pure" limestone or marly limestone as for instance near Schaffhausen and at the east end of Mt. Lägern. Details of the facies change are not known because of insufficient outcrops. The exact position of the mineralostratigraphic correlation M with respect to the biochronologic ammonite scale is not known.

Wettingen Member (MOESCH 1867, p. 193)

There is no suitable section of the Wettingen Member near Wettingen. SENFTLEBEN (1923, p. 51) found that the lowermost 10 m of the member near Wettingen are a bedded micritic limestone indistinguishable from the Villigen Formation. Above are at least 20 m of massive limestone with siliceous sponges. The top of the massive unit was eroded everywhere in canton Aargau in pre-Eocene time. West of Villigen, the Wettingen Member begins with three beds of light micritic limestone with *Idoceras balderum* (OPPEL) (GYGI 1969, Pl. 17, section 62). Above is a massive limestone with siliceous sponges like at Wettingen. In the large quarry near Mellikon, there are up to 10 m of bedded micritic limestone of the lower Wettingen Member with *Idoceras balderum* (OPPEL) near the base. The Wettingen Member then begins in the Divisum Zone (Table 2). This is confirmed by several *Aspidoceras uhlandi* (OPPEL) which were found in this succession during quarrying operations at Mellikon. However, the exact horizon of the specimens (now in the Museum of Natural History Basel) is not known. The limestone succession of the lowermost Wettingen Member at Mellikon thins out or grades into marl to the east, since F. J. and L. WÜRTENBERGER (1866, p. 14) have found *Aspidoceras uhlandi* near Baltersweil (about halfway between Mellikon and Schaffhausen, see Fig. 1) near the top of a succession of marl and marly limestone with a thickness we estimate at about 20 m (Pl. 1B). The same marly succession is about 18 m thick at the eastern end of Mt. Lägern near Regensberg ("eigentliche Badener Mergel" of NOTZ 1924, section at the end of his Ph. D. thesis). This marly succession is the time equivalent of the Baden Member and of the lowermost Wettingen Member, and of the "Weissjura gamma" in southern Germany. The boundary between the facies of the Aargau Jura and the facies of the Swabian Alb of southern Germany runs in a N–S direction from Geisslingen north of the Rhine river through the middle of Mt. Lägern east of Baden.

In the Swabian facies realm, in a quarry at the eastern end of Mt. Lägern near Regensberg, a sponge bioherm about 10 m thick is above the marly lower Kimmeridgian succession. At Baltersweil north of the Rhine (Fig. 1), the corresponding unit is a bedded, marly limestone about 9 m thick, with siliceous sponges in the upper part. This is unit c in section 3 by F. J. and L. WÜRTENBERGER (1866, p. 14). No diagnostic ammonites have yet been found in this unit, neither at Baltersweil nor near Regensberg. Above is a blue-grey marl with a mean thickness of 3–4 m. We have found an adult and near-complete *Aspidoceras acanthicum* (OPPEL) in this unit near Regensberg. The specimen is now in the Paleontologic Museum of the University of Zürich. This means that the marl is probably the time equivalent of the Banné marl near Courgenay, and that it corresponds to the Weissjura delta 2 in southern Germany. The bedded micritic limestone unit above the marl with a thickness of about 10 m and with *Aulacostephanus eudoxus* (D'ORBIGNY) found near Regensberg is the highest unit represented in Plate 1B. This is Weissjura delta 3 or what is called Quaderkalk near Schaffhausen. This unit is the upper part of sequence 7 in Plate 1B.

3.2 Main types of facies

3.2.1 Iron-oolitic marls and limestones

A marl-clay with up to 10% of limonitic iron ooids is at the base of the Oxfordian Stage under the Renggeri Member in the northwest (Pl. 1A, Table 2). The best outcrop is in the clay pit of Andil near Liesberg (section RG 280, horizon 6, see GYGI 1982, Fig. 2). There, the thickness of the horizon is only 0.3 m. This is typical of this peculiar type of iron-oolitic formation. The diameter of the iron ooids is between 0.5 and 1.5 mm. More than 80% of the macrofauna are ammonites and belemnites. The low concentration of iron ooids in a mud-grade matrix, the abundance, and the composition of the fauna suggest that this sediment was laid down at a slow rate in an open marine environment in water several tens of meters deep, at the distal fringe of a thick argillaceous succession. A model of this mode of formation was proposed by GYGI (1981). According to Walthers's law, the iron-oolitic marl-clay at the base of the Oxfordian in the northwest grades upward into the Renggeri Member (Pl. 1A).

A marly limestone with a low content of iron ooids and with a similar, abundant fauna is at the base of the Oxfordian in canton Aargau. This is the Schellenbrücke Bed (GYGI 1977, p. 454). Fossils and lithoclasts up to several centimeters across may be encrusted with limonite with a glossy surface. Again, more than 80% of the fauna are cephalopods (GYGI 1981, Table 1). There are traces of siliceous sponges. Sponges may have been abundant in the biocoenosis, but they are not preserved for lack of smothering sediment. The multiphase mode of formation of the sediment was interpreted and discussed by GYGI (1981), and by GYGI & MARCHAND (1982). The Schellenbrücke Bed was formed in an open marine environment, at a depth of about 80 to 100 m.

Iron-oolitic deposits of this kind may be exceptional in comparison with other iron-oolitic deposits of the whole sedimentary record. However, they are not rare in Central Europe, and they typically occur as thin, widespread marker horizons.

3.2.2 Marls (primary mixtures of argillaceous and carbonate mud)

Renggeri Member

The vertical transition from the dark grey-brown, iron-oolitic marl-clay at the base of the Renggeri Member to the blue-grey Renggeri Member proper marks an increase of the originally slow sedimentation rate to a normal value. According to an analysis by PFRUNDER & WICKERT (1970) and the classification by PETTIJOHN (1957, p. 410), the Renggeri Member is a clayey marl with 32% carbonate. The sediment is massive, with no apparent bedding or lamination. The macrofauna is dominated by cephalopods. Small ammonites, or the innermost whorls of larger specimens, are preserved as iron sulfide casts. Benthic, sessile organisms are few. This means that the Renggeri Member was sedimented in a shallow basin with very slack bottom currents, but with oxygenated water above the sediment/water interface. Only the interstitial water of the sediment was rather strongly reducing. There must have been a sharp gradient of the E_H value just below the sediment surface due to the small particle size of clay minerals.

Effingen Member

Marls predominate in the Effingen Member. They are indistinguishable from the Renggeri Member in a perfectly fresh outcrop, but they have a greater carbonate content, and the fauna is much less abundant. The minimum carbonate percentage in this member is 37 at Auenstein (GYGI 1969, Pl. 17, section RG 37, horizon 9), and the carbonate content varies widely to "pure" limestone concentration within this section and others. The Effingen Member as a whole can easily be distinguished from the Renggeri Member by the well-bedded, more or less pure limestone intercalations where the two members occur in the same succession. Bands of carbonate nodules are rare in the Effingen Member. Siliceous sponges and ammonites are fairly abundant in the lowermost 20 to 30 m of the Effingen Member. Deposition of this member must have begun in water which was deeper than 100 m. Considerable depth of the water is suggested by the presence of some nannoplankton (GYGI 1969, Pl. 5, Fig. 17). Further up in the succession, there is no macrobenthos except very few brachiopods and bivalves. Foraminifers and ostracods are rare and stunted (OERTLI 1959, GYGI 1969). Bottom water was stagnant and reducing at times (GYGI 1969, p. 107), but apparently not for long enough to allow organic matter to accumulate a major content (GYGI 1969, p. 21). The water was mildly oxidizing and moving in a slack current at other times. The latter is documented by the dwarf microfossils and by bioturbated horizons (GYGI 1969, Pl. 4, Fig. 12). Ammonites are very rare. Water became shallower during deposition of the Effingen Member (GYGI 1969, p. 107). The existence of a depositional slope is indicated by a truncation surface (Fig. 2), debris flow sediments (Fig. 3), and thin, fine-grained turbidites. This belongs to the standard facies belt 3 of J. L. WILSON (1975, p. 351).

Terrain à Chailles Member

The Terrain à Chailles Member above the Renggeri Member is mostly a blue-grey marl with a somewhat higher carbonate content than the Renggeri Member. Here, continuous bands of limestone like the ones in the Effingen Member are exceptional. The carbonate is usually concentrated in flattened nodules with a diameter from one to several decimeters. These nodules occur in well-defined horizons. Some of them are slightly chertified in the interior. Ammonites, bivalves, and brachipods are only found in the nodules. Fossils are not abundant except in a somewhat condensed horizon in the middle Terrain à Chailles Member (Pl. 1A). The marl in the Liesberg Member is similar. Nothing is known about the carbonate content, but horizons with carbonate nodules succeed each other at shorter intervals in this member than in the Terrain à Chailles Member, and the nodules have an irregular shape with a rough surface. The nodules of the Liesberg Member contain a small amount of chert in the matrix as well as in the fossils. The rich fauna of the Liesberg Member is dominated by hermatypic corals. These and the associated fauna occur in the marl matrix and in the nodules. The high ratio between ammonites and other organisms in the Renggeri Member decreases sharply in the Terrain à Chailles Member. This and the advent of hermatypic corals in the Liesberg Member indicates a general trend of shallowing of the water from the beginning of deposition of the Renggeri Member.

Natica Member

The marls in the Natica Member are mostly grey, but reddish to greenish varieties are not uncommon. Some of these marls have a high content of detrital quartz, and there is some feldspar. The coal seams as reported by HEER (1865) and KEMMERLING (1911), and limnic characeans as figured by OERTLI & ZIEGLER (1958) occur in some of these marls. In other marly intercalations in the Natica Member there are mass occurrences of algal nodules or nerineid gastropods from the shallow subtidal or intertidal zone. During deposition of the lower Natica Member, the environment apparently was marginal marine. It was mostly supratidal with widespread coastal swamps and freshwater ponds when the upper part of the member was laid down, then turned to open marine at last. The latter is indicated by the occurrence of the giant nautiloid *Paracenoceras giganteum* (D'ORBIGNY) J 30716 just below the Hauptmumienbank near Grandval in section RG414, horizon 38, east of Moutier (unpublished), or by hermatypic corals as far in the platform interior as Bressaucourt southwest of Porrentruy (Pl. 1A), as reported by SCHNEIDER (1960, p. 7), and seen again by R. Gygi.

3.2.3 Fine-grained carbonate mudstones (calcilutites)

"Basinal" Calcilutites

Fine-grained carbonate mudstones or calcilutites are common in the "basinal" facies. They make up most of the Villigen Formation (Table 2). The mudstones are well-bedded, and they may be almost pure micrites (GYGI 1969). Porosity is normally low. Relatively high-porosity mudstones are common only in the upper St-Ursanne Formation. Others are in the Balsthal Formation and in the Villigen Formation (GYGI 1969, Pl. 19). High porosity only occurs when the carbonate content approaches 100%. Lateral changes from high to low-porosity mudstone over a distance of only 10 to 20 m have recently been observed in the now vast underground limestone quarry at the St-Ursanne railway station. The reason for this is not known. Bedding of "basinal" calcilutites is parallel where it is not influenced by sponge bioherms. There are horizons in the Effingen Member and in the Geissberg Member which appear to be the deposits of sediment gravity flows (GYGI 1969, Pl. 4, Fig. 13). Occasional relicts of flat lamination indicate that some of the mudstones of the Villigen Formation of canton Schaffhausen are fine-grained, distal turbidites, or tempestites (for instance horizon 129, Knollen Beds, of section RG 82 in GYGI 1969, Pl. 16). Presence of micritized carbonate ooids in the Knollen Beds near Beggingen northwest of Schaffhausen, in the "basin" facies with ammonites, can only be explained by the action of sediment gravity flows or strong tempests (see GYGI 1969, p. 58). The fauna of the "basinal" calcilutites is usually scarce. There are mainly ammonites and some bivalves and brachiopods.

Shallow-water Calcilutites

On the platform, the well-bedded micritic limestone of the Vorbourg Member may at first glance have a very similar aspect. However, bedding in this member is usually significantly thicker, and there are prism-cracked bedding planes and angular or subangular lithoclasts of varying size. The lithoclasts may be dark-grey to black. Marine bivalves and mainly nerineid gastropods, the latter locally with an oncolitic crust, or foraminifers of the genus *Alveosepta* and the hydrozoan *Cladocoropsis mirabilis* FELIX are evidence that the environment of deposition of part of the carbonate mudstones of the Vorbourg Member was the shallow subtidal or the intertidal zone (see TURNŠEK et al. 1981). Mud-cracked surfaces and black pebbles are indicative of the intertidal or of the supratidal zone (M. A. ZIEGLER 1962, Pl. 1, Fig. 3). Well-bedded, mainly micritic limestones with *Alveosepta* and occasional lithoclasts (storm layers) replace the Holzflue Member laterally in the proximal direction north of St-Ursanne.

Fenestrate carbonate mudstones

Beds of micrite with fenestrae (millimeter-scale pores filled with a calcite mosaic or mud, named by TEBBUTT et al. 1965), laminated or unlaminated, are to be found mainly in the Natica Member, but also in the lower Reuchenette Formation. They are briefly referred to below (stromatolites).

3.2.4 Sponge limestones and marls

Limestones and marls with abundant and well-preserved siliceous sponges, either in biostromes or in bioherms, are only known from deeper water (GYGI 1969, Pl. 2, Fig. 6, Pl. 6, Fig. 23). The sponge skeleton is, when preserved, replaced by calcite. Siliceous sponges thrived when the rate of deposition was below average. The siliceous sponge biostromes of the Birmenstorf and of the Baden Members both contain authigenic glauconite. This and the rich ammonite assemblages associated with the sponges are indications of a reduced rate of deposition, and of a water depth of 100 m or more (GYGI et al. 1979, p. 946).

In the Terrain à Chailles Member, primarily siliceous megasclere spicules of sponges are replaced by calcite as well. Such spicules even occur in coral limestones (see GYGI 1969, Pl. 7, Fig. 29). The silica in the partly chertified carbonate nodules and in the fossils of the uppermost Terrain à Chailles Member and of the Liesberg Member, and in the fossils of the argillaceous limestone of the lowermost St-Ursanne Formation, is probably derived from the decay of sponges which lived in the same environment as the chertified organisms. This may be inferred since sponges are quantitatively important in Recent coral reefs (RÜTZLER 1978). Calcareous sponges (class Calcarea) make up only a small percentage of the sponge biomass in Recent reefs. Members of all other sponge classes produce spicules of opaline silica (RÜTZLER & MACINTYRE 1978, p. 147). These authors have found sponge spicules to be the main component of particulate silica in perireefal sediments within the barrier reef tract of Belize in the Caribbean.

3.2.5 Coral limestones and marls

Small solitary corals were found in the "basinal" Schellenbrücke Bed, in the Crenularis Member (GYGI 1969, Pl. 6, Fig. 23), and in the Baden Member (Table 2). They belong to unidentified, probably ahermatypic deep-water species. When going up a shallowing-upward sequence, the first hermatypic coral colonies to appear are flattened like those in the biostrome of the Liesberg Member. The matrix of the Liesberg Member is a grey marl. The flattened coral colonies may make up as much as 30 to 40% of the rock. The same colonies, but in a lime mudstone matrix, occur in the St-Ursanne Formation to the south of Basel, or in the Günsberg Member east of Moutier (Fig. 1). The latter are figured in M. A. ZIEGLER (1962, Pl. 2, Fig. 7). Flat coral colonies are an adaptation (genetic or environmental) to low average illumination (see GRAUS & MACINTYRE 1976). Further up a shallowing-upward sequence, colonies become thicker, then dome-shaped to massive or branching. It is in the latter level that most bioherms begin to develop. There is as yet no method to establish with some degree of accuracy the depth intervals of these ecologic zones. In clear water, substantial growth of hermatypic corals starts at a depth of about 80 m (JAMES & GINSBURG 1979, p. 43). However, we have reason to assume that growth of dish-shaped hermatypic corals in the Liesberg Member at Liesberg commenced in water no deeper than about 20 m, as a consequence of significant light absorption by a high average amount of suspended matter in the water. ZLATARSKI (in ZLATARSKI & MARTINEZ 1982, p. 395) found that hermatypic corals can build reefs in permanently turbid water with a visibility as low as 20 to 30 cm.

Thick, flat coral colonies occur to the top of the Liesberg Member in the quarry of Chestel near Liesbergmüli at Liesberg. There, the first massive coral colonies appear in marl 4.4 m below the top of the Liesberg Member. When the influx of argillaceous mud ceased, deposition of 1.5 m of arenitic, bioclastic packstone with oncoids of the St-Ursanne Formation set in and was followed by 4.2 m of bioclastic-oncolitic grainstone with some massive coral colonies not in life position. The latter sediment is from very shallow water. This is evidence that growth of dish-shaped coral colonies may have continued in this section until the water was less than 10 m deep.

We conclude from this that depth intervals as defined by the growth form of hermatypic coral colonies may be telescoped to a fraction of the value as observed in clear water if light penetration is often or even permanently reduced by a substantial amount of mud-size siliciclastic or calcareous particles in suspension.

Coral Bioherms

Development of coral bioherms started in Middle Oxfordian time at the margin of the platform of the lower St-Ursanne Formation. The oldest bioherms had a vertical extension of less than 10 m. Later bioherms did not exceed 30 m of vertical extension. This is at variance with PÜMPIN (1965, p. 859) who claimed to have found bioherms as high as 70 m near Glovelier. New masurements of these bioherms gave a maximum vertical extension of about 30 m. Pümpin failed to take the steep tectonic dip of the strata at this location into account, which can be measured in the well-bedded Vorbourg Member overhead. The elevation of the bioherms near St-Ursanne above the surrounding seafloor was slight at all times (PÜMPIN 1965, Fig. 14). This appears to be a general feature of Oxfordian coral bioherms in northern Switzerland. We have as yet no evidence that a barrier reef evolved at the margin of the St-Ursanne Formation. Only individual bioherms with more or less lateral and vertical interspace were found.

The coral bioherms at the base of the Günsberg Member are normally less than 10 m thick and less than 20 m wide. Some were growing side-to-side and on top of each other like those for instance at the foot of the cliff west of Mt. Hasenmatt (Fig. 1, triangle northwest of Solothurn). The best outcrop of such a bioherm is at Péry in the quarry of La Charuque (section RG 307, horizon 160, see Fig. 6 in GYGI 1982). Bioherms in the Balsthal Formation and biostromes in the Reuchenette Formation are few and isolated except near Olten (Pl. 1B).

Maximum development of coral bioherms is in shallow water at the windward margin of a platform where the food and oxygen supply by currents is optimal. In our region, it is only in the lagoonal environment of the St-Ursanne Formation that an indication of the direction of surface water paleocurrents could be found. PÜMPIN (1965, p. 843) inferred that the predominant direction of currents was from the SSW near St-Ursanne, at a paleolatitude of about 35° N (SMITH & BRIDEN 1977, or FIRSTBROOK et al. 1979). AGER (1975, Fig. 3) assumed that currents in the western Tethys came from the south during the Kimmeridgian. However, the pattern of surficial currents within the epicontinental archipelago of West-Central Europe in the Late Jurassic is mostly unknown.

3.2.6 Bioclastic and oolitic limestones

Bioclastic limestones, mainly arenitic and with isolated coral colonies, are common in the lower St-Ursanne Formation (Pl. 1A). They form a belt between the coral bioherms near the platform margin and the interior, where they grade into oolitic, then oncolitic limestones. Oolite occurs in the lower and in the upper St-Ursanne Formation behind the platform margin in bands several kilometers wide. The older band crops out near Liesberg and near St-Ursanne. The younger oolite belt of the St-Ursanne Formation is the "Tiergarten oolite" of BOLLIGER & BURRI (1970). Oolite makes up the bulk of the Balsthal Formation (Pl. 1A). In the Verena Member, the ooids are largely micritized and often dedolomitized. Many ooids of this member are indented by pseudomorphs of calcite after clusters of probably sulfate crystals (see above). Microfacies of bioclastic limestones transitional to oolite, and of oolites from the Balsthal Formation have been figured by GYGI (1969, Pl. 8, Fig. 31–33, Pl. 9, Fig. 34 and 36, Pl. 12, Fig. 46, and Pl. 13, Fig. 47 and 50). Today, ooid sand is forming in water less than 5 m deep where there are strong tidal currents. Recent, active ooid shoals comparable to those in the Upper Jurassic of northern Switzerland are to the north of Andros Island, Bahamas, some distance from the reef tract at the platform margin towards the bank interior (see map by PURDY 1963, or GEBELEIN in WIEDENMAYER 1978, Fig. 1).

The Oolithe rousse is a facies which deserves special attention. The name refers to the thin red-brown crust on the surface of the ooids which gives the weathered rock its characteristic colour. Maximum thickness is about 6 m only. The unit is above the oncolitic Hauptmumienbank Member. Therefore, it is transgressive. The proximal time equivalent is the Humeralis marl. The oolite grades from wackestone to grainstone texture. Grainstones are locally cross-bedded. Their late diagenetic calcite B cement has normally a low iron content. When it is ferroan, the iron probably came from the coeval proximal Humeralis marl which was then in the process of compaction (see OLDERSHAW & SCOFFIN 1967, and GYGI 1981). The groundmass of argillaceous wackestones and packstones is often made up mostly of dedolomitized rhombs. The surface of the rhombs and ooids is coated with a brown pigment (limonite). Clay minerals are probably the source of the iron (see CARROLL 1958, and MCHARGUE & PRICE 1982). LAUBSCHER (1963, p. 10) referred to probable equivalents of the Oolithe rousse as horizons with iron ooids. There are similar oolites in the Natica Member.

3.2.7 *Oncolitic limestones or oncolites*

Oncolite is a rock with a major content of oncoids (HEIM 1916, p. 542). According to the original definition by HEIM (1916, p. 566), an oncoid is an individual particle. The diameter of Oxfordian oncoids varies between less than 2 mm and more than 30 cm (GYGI et al. 1979, p. 943). HEIM (1916) already noticed that discrimination between small oncoids and (micritized) ooids may be difficult. At least the majority of the large, concentrically structured Oxfordian oncoids was hard at the time of deposition since many of them are bored by *Lithodomus* (GYGI 1969, p. 38, and Pl. 10, Fig. 37, see also KUTEK & RADWANSKY 1965, Pl. 3, Fig. 5). These oncoids were formed by a biocoenosis of lime-accreting organisms other than red algae. Jurassic oncoids are unlike calcified algal nodules in the Recent marine environment. Recent marine calcified algal nodules (rhodolites) are formed by Rhodophyta (red algae).

All Oxfordian oncolites of the platform facies in northern Switzerland were formed in the shallow subtidal or in the intertidal zone. Oncoids are best developed and most common when they are in a matrix of primarily lime mud (GYGI 1969, p. 112). A mud matrix indicates that the environment was sufficiently protected for mud-size particles to accumulate. On the other hand, the concentric crusts of the large oncoids are evidence that the nodules were rolled by strong currents quite often or even at regular intervals. Settled mud must have been stirred up in the process. The net mud deposition rate was relatively slow. Otherwise, the algae in the nodules would have been suffocated. DAHANAYAKE (1978, p. 314) came to the same conclusion for the very similar Kimmeridgian oncolites of southeastern France. Comparable nodules now exist in limnic environments only. This has been stressed by MONTY (1974, p. 612). A mass occurrence of living, calcified algal nodules growing on or within unconsolidated lime mud and silt in a tropical tidal environment of the Recent has been found by POLLOCK (1928). However, the resemblance with Jurassic paleoenvironments is only superficial since the author stated (p. 27) that the nodules were accreted by a red alga (see also BOSELLINI & GINSBURG 1971).

Oxfordian oncoids from the shallow-water environment of northwestern Switzerland have a maximum diameter of about 6 cm. Oncoids with a nerineid gastropod in the core were probably formed in the shallow subtidal or in the lower intertidal zone. Pseudomorphs of calcite after probably calcium sulfate in the core of oncoids in an otherwise unfossiliferous rock are indicative of the upper intertidal zone, where the nodules were subject to prolonged periods of desiccation. No ammonites have been reported from shallow-water oncolites so far.

Oncoids with glauconite and *Cycloserpula,* with a diameter of 1 to 2 cm, were found in the section RG 28, horizon no 47 (lowermost Kimmeridgian) near Schönenwerd (GYGI 1969, p. 8). This horizon with glauconite-impregnated oncoids reappears in the quarry south of Löchli near Däniken SO (see Pl. 1B). Large oncoids with glauconite and limonite were formed in water at least 100 m deep in the Middle Oxfordian Mumienmergel and in the Mumienkalk near Schaffhausen, in a matrix of argillaceous mud or lime mud, respectively (Pl. 1A). The diameter of the largest, discoidal oncoids is in excess of 30 cm. The thin, glauconitic Mumienmergel and the Mumienkalk are time-equivalent with the lowermost Birmenstorf Member in canton Aargau (Table 2). The oncoids of the Mumienmergel and of the Mumienkalk are associated with a rich fauna which is dominated by cephalopods (mainly ammonites). This is evidence of an open marine environment in relatively deep water. Figures of these oncoids, a table of the macrofaunal assemblage, and a discussion of the paleoenvironment of the Mumienkalk are to be found in GYGI et al. (1979).

3.2.8 Stromatolites

Stromatolites were long taken as unequivocal evidence of a shallow subtidal to a supratidal environment. Only a few authors have recognized fossil stromatolites from a relatively deep marine environment. Stromatolites may only be interpreted as being

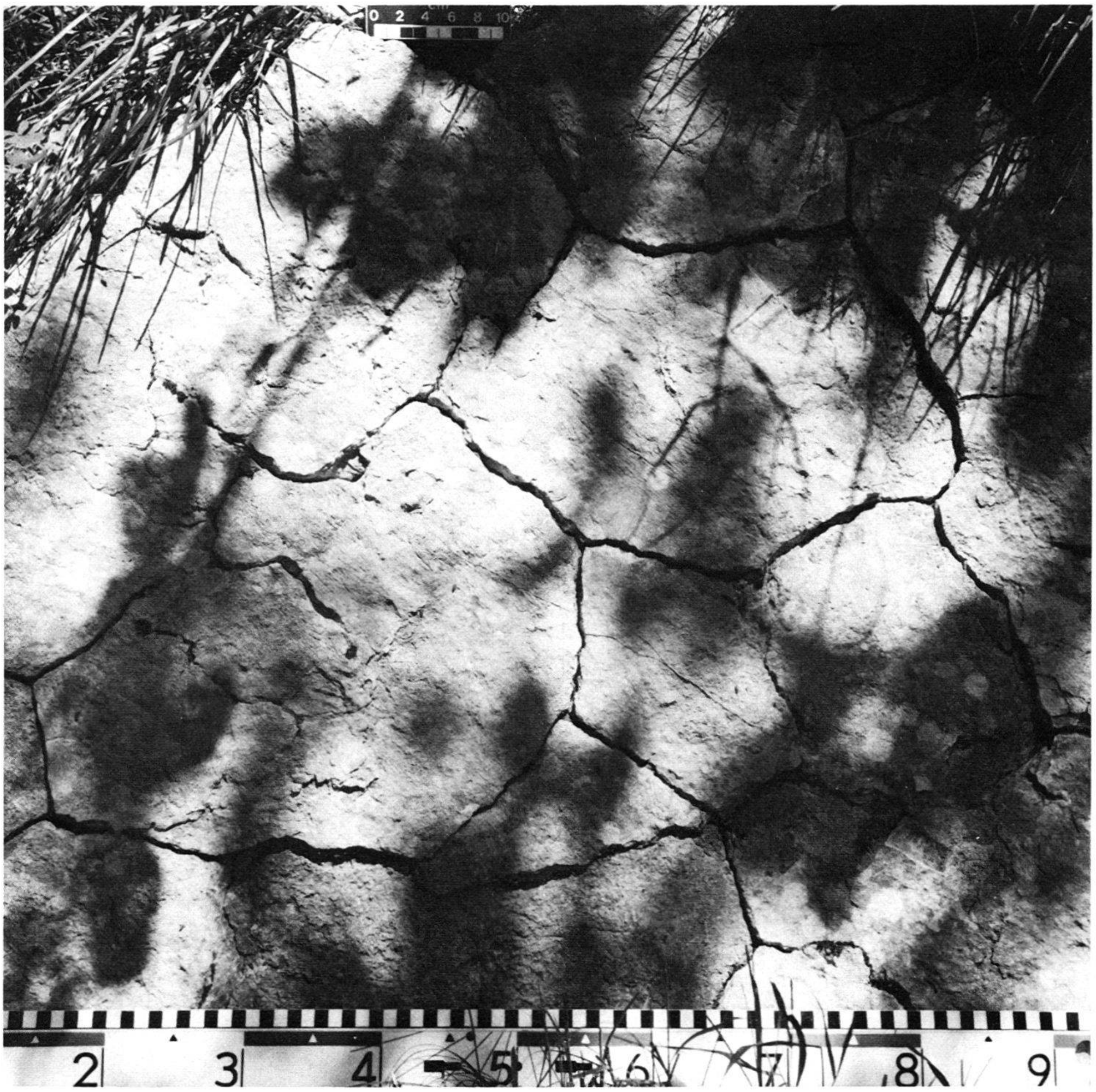


Fig. 4. Upper surface of a prism-cracked lime mudstone bed, with diffuse flat lamination directly below the surface, and with large, laminated and desiccation-cracked angular clasts in the interior. Peritidal deposit of the upper Natica Member, top of subsequence 2a (see Pl. 1A). Bed no 42 of section RG 417 near Crémines BE (see text). Scale is decimeters.

formed in the upper intertidal or in the supratidal zone if they are intersected by prism and sheet cracks (A. G. FISCHER 1964, Fig. 8), or if there are fenestrae (TEBBUTT et al. 1965), or both. Prism-cracked supratidal stromatolites (Fig. 4) may disintegrate into lithoclasts as figured by M. A. ZIEGLER (1962, Pl. 1, Fig. 4) from the Vorbourg Member, or from the Natica Member by BOLLIGER & BURRI (1970, Pl. 2, Fig. 1). More examples of lithoclasts from stromatolites have since been found in new sections measured by R. Gygi in the Natica Member (Fig. 5). The prism-cracked stromatolite surface in our Figure 4 is bed 11 of section 56 by P. A. ZIEGLER (1956, p. 98) near Crémines. This particular

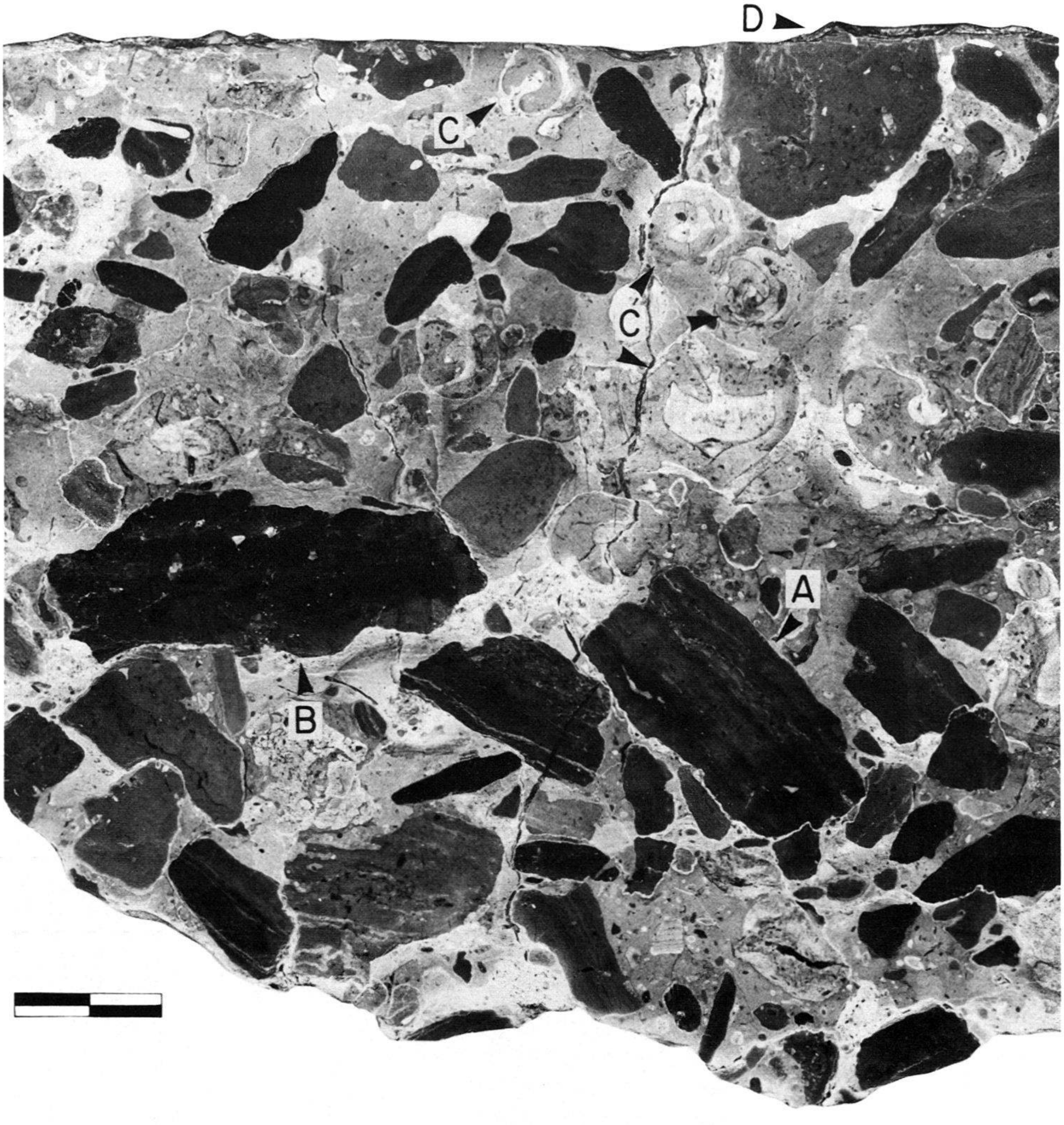


Fig. 5. "Black pebble conglomerate", polished slab of limestone. Peritidal deposit of the upper Natica Member, top of subsequence 2a (cf. Pl. 1A), same unit as in Figure 4. Bed no 49 of section RG 406 near Vermes JU. The angular clasts, some with crincled lamination (A), and some conglomeratic (B), are preserved at different stages of blackening. The clasts and nerineid gastropods (C) are embedded in a lime mud matrix. The top of the bed is a planed erosion surface. This is a bored hardground encrusted with ostreid bivalves (D). Bar is 2 cm.

stromatolitic unit reappears near Grandval in the upper part of horizon 3 of section 8b by M. A. ZIEGLER (1962). It seems to be continuous over a distance of several kilometers on the south slope of Mt. Raimeux east of Moutier, and it reappears on the north slope of the mountain near Vermes JU (Fig. 5). It is the uppermost unit of subsequence 2a (Pl. 1A).

Deepwater stromatolites such as the one figured by GYGI (1969, Pl. 1, Fig. 2) and re-interpreted by GYGI (1981) only occur as local patches. There are neither desiccation cracks nor fenestrae within them. Such deepwater stromatolites are associated with a rich macrofauna of mainly cephalopods. The paleoenvironment of stromatolites as formed in a water depth of 80 to 100 m was discussed by GYGI (1981), and by GYGI & MARCHAND (1982).

3.3 Lateral facies transitions

The main problem of Oxfordian stratigraphy in northern Switzerland since Gressly's time was the following: How does the facies of the platform carbonates of the St-Ursanne Formation in the northwest change towards the "basin" in the southeast? GRESSLY (1864, p. 103) recognized that at least part of the Pichoux limestone was to correlate with the Birmenstorf Member. ROLLIER (1911, Fig. 54) thought that the St-Ursanne Formation graded into the whole of the thicker Wildegg Formation, including the Geissberg Member. A facies change of the kind envisaged by Rollier is well exposed in the Helvetic nappes of the Swiss Alps. There, the Schrattenkalk platform carbonates pass into thicker, deeper water marls and limestones in the southeast (HEIM 1916, Fig. 105). BOLLIGER & BURRI (1970) felt they had presented conclusive evidence for the transition of the St-Ursanne Formation to only the Birmenstorf Member in the "basin". The facies transition, according to these authors, would then be a small-scale counterpart of for instance the transition of the thick Dachstein platform carbonates to the thin, deepwater Hallstatt Limestone, which is a starved basin deposit of the alpine Triassic (A. G. FISCHER 1964, Fig. 2).

Rollier and others pointed out several localities where the lateral facies transition from the St-Ursanne Formation to the "basinal" time-equivalent deposits could be directly observed in the field. We have found only one outcrop where this facies change can be seen. As far as we know, this has not been described or even mentioned before. The outcrop is in the Combe des Geais north of Grandval, on the south slope of Mt. Raimeux east of Moutier (Fig. 1). The relative position of the cirque of the Combe des Geais is slightly basinward from section 6 in the Gorges du Pichoux near Sornetan (Pl. 1A). The north rim of the cirque is formed by a vertical cliff which is nearly 80 m high on the west side (below point 1065.3 m of the Swiss Federal Map 1:25,000). The cliff can be divided into six units (Fig. 6 and 8). Units 1 to 3 are calcilutites of the Pichoux limestone. Unit 2 is a marly limestone grading into marl towards the south (Pl. 1A). Rare ammonites (fragments of perisphinctids) occur near the top of unit 2. Isolated hermatypic coral colonies and solitary corals appear in unit 3. They mark just about the basinward boundary of the St-Ursanne Formation. Units 4 and 5 are an almost massive limestone on the west flank of the cirque (Fig. 6). Unit 4 grades into a marly limestone towards the southeast where it is capped by the marl of unit 5 which is sufficiently argillaceous to act as an aquifer confining bed. Unit 5 develops a terrace when it becomes increasingly marly on the east side of the Combe des Geais (Fig. 7, compare with Fig. 8 for details). There are low-relief coral bioherms in unit 6, with bioclastic limestone in between, on the north side of the cirque. On the north side, where one can climb through the wall, bioherms are built by massive, thickly-spaced coral colonies in a bioclastic matrix. On the east side, the bioherms of unit 6 weather out as isolated knobs (Fig. 8). These bioherms are built by massive coral colonies in a lime mudstone matrix. Only 250 m to the southwest from the rock represented in Figure 8, or less than 100 m basinward, bioherms are replaced by a biostrome of thickly-spaced flat coral colonies in a lime mudstone matrix (M. A. Ziegler 1962, Pl. 2, Fig. 7). This is where M. A. Ziegler (1962) has measured his section 8a.

There is a small amount of detrital quartz in units 4 to 6. This is an indication that the units belong to an early stage in the evolution of the Günsberg Member (Pl. 1A). The lower and middle Günsberg Member is time-equivalent with the argillaceous Natica Member of the platform interior (Table 2). The Natica Member becomes increasingly calcareous towards the platform margin where it grades laterally into the Günsberg Member. The Günsberg Member is the almost pure carbonate facies of the platform margin. Unit 5 changes between the western face of the Combe des Geais and the eastern side from a massive limestone to a soft grey marl of the slope facies (of the Effingen Member). Rollier (1901) mapped units 1 to 3 as "Argovo-Rauracien", this is to say as a facies transitional between the St-Ursanne Formation and the Wildegg Formation. He included units 4 to 6 in his "Séquanien" which is now the Günsberg Member of the Balsthal Formation.

A minor lateral facies transition from limestone to marl is visible in the lowermost St-Ursanne Formation near Bärschwil (southeast of Liesberg, see Fig. 1). In a landslide southeast of Bärschwil, near the farm Vögeli, the lowermost St-Ursanne Formation

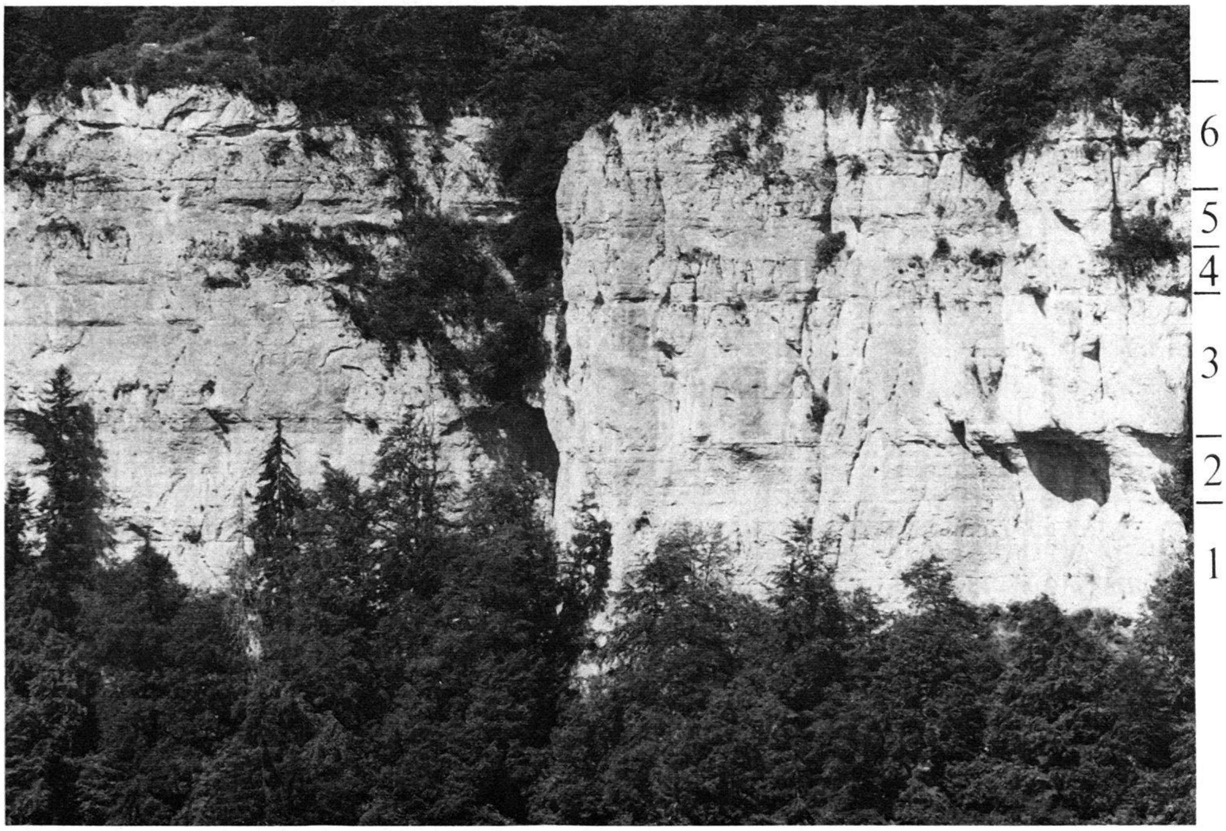


Fig. 6. Combe des Geais north of Grandval (east of Moutier, cf. Fig. 1), west wall of the cirque. Units 1 to 3 = Pichoux limestone, 4 to 6 = Günsberg Member. Explanation in the text.

Fig. 7. Combe des Geais, north wall (left) and east wall (right). The forested bottom of the cirque is above marl of the Early and Middle Oxfordian.

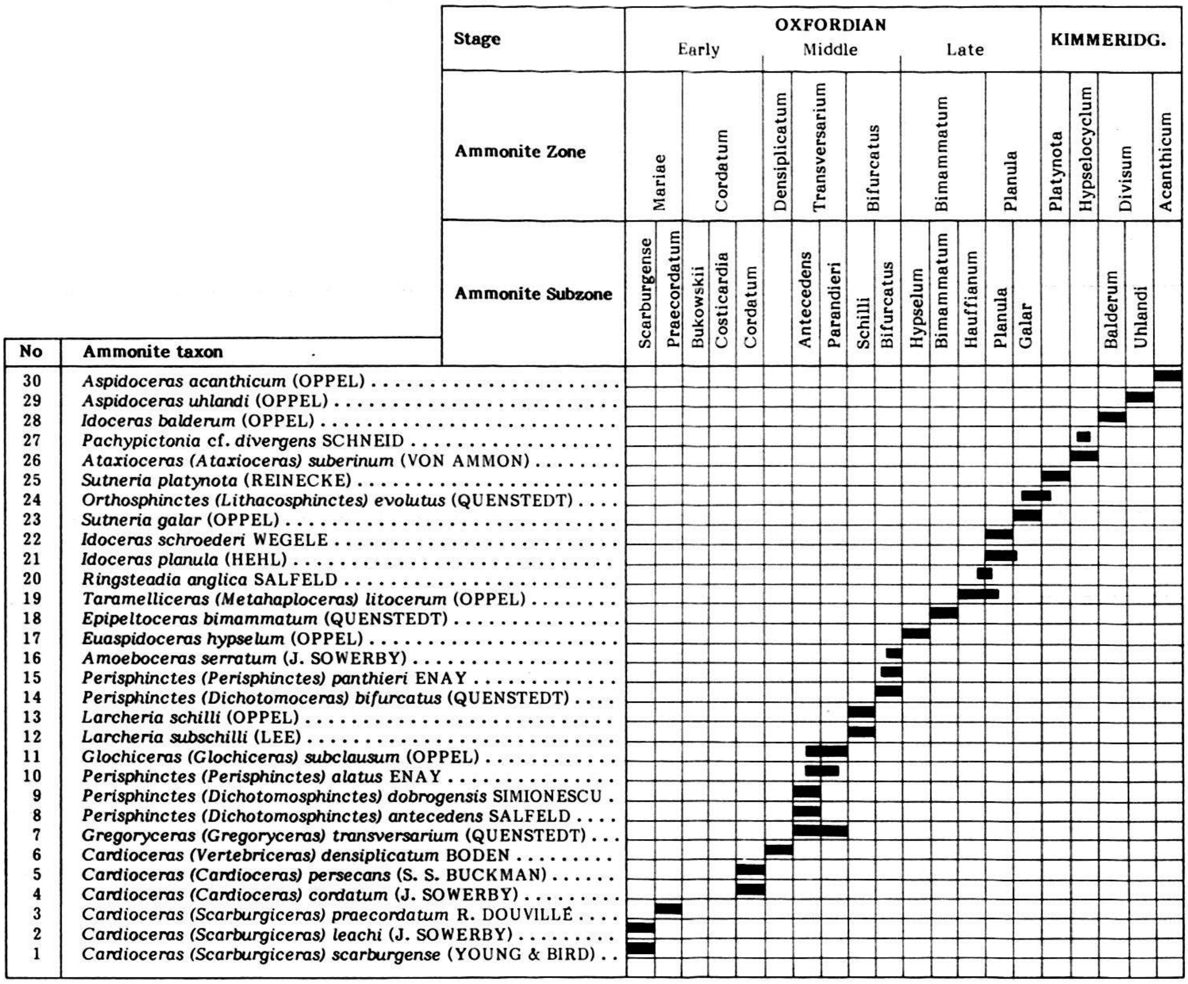
Fig. 8. Combe des Geais, detail of the east wall. Coral bioherms are in units 4 and 6. Bioherms of unit 6, weathered out as isolated knobs, protrude from the forest. Unit 5 is mostly marl (cf. Fig. 6 and text).

grades laterally into the Liesberg Member towards the "basin" in the southeast. In the landslide called Gschlief north of Günsberg, northeast of Solothurn, the lower Günsberg Member passes laterally into the Effingen Member (GYGI 1969, Fig. 3).

3.4 The ammonite succession

There is a complete ammonite succession in the Oxfordian and in the lower Kimmeridgian of northern Switzerland. The subdivision of the Oxfordian into ammonite zones and subzones is given in Table 3. We use the zones and subzones which can be identified by their index taxon in this country. All the Early Oxfordian ammonite zones and subzones are represented in the Renggeri Member and in the Terrain à Chailles. Zonation of the Middle and of the Late Oxfordian is documented in the Wildegg and in the Villigen Formation. The abundance of ammonites varies widely. Ammonites are very abundant for instance in the condensed iron-oolitic horizons at the base of the Oxfordian, but they are rare or even absent in parts of the upper Effingen Member in canton Aargau. The ammonite specimens most important to the biochronologic calibration of Oxfordian strata in northern Switzerland and the specimens important to correlation within our

Table 3: *Vertical range of ammonite taxa represented in Plate 1.*



region are indicated in Plate 1A. Most of the ammonites cited below have an individual number (see "Methods").

3.4.1 *Early Oxfordian*

The Early Oxfordian *Cardioceras (Scarburgiceras)* cf. *scarburgense* J 30717 and *Cardioceras (Scarburgiceras) leachi* (J. Sowerby) J 30709, both of the Scarburgense Subzone, were found in the quarry of La Charuque near Péry in the iron-oolitic marl-clay at the base of the Renggeri Member. This is horizon 20 of section RG 307 (Gygi 1982, Fig. 6). *Cardioceras leachi* Gy 1017 from the Schellenbrücke Bed near Gansingen in canton Aargau was figured by Gygi & Marchand (1982, Pl. 2, Fig. 1). Specimens of *Cardioceras scarburgense* close to the type of the taxon occur in the Renggeri Member proper near Liesberg about two meters above the base of the member. The well-preserved *Cardioceras (Scarburgiceras) praecordatum* R. Douvillé J 30949 was taken by R. Himmler from in situ about 27 m above the base of the Renggeri Member near Liesberg. We do not have *Cardioceras (Scarburgiceras) bukowskii* Maire and *Cardioceras (Cardioceras) costicardia* S. Buckman from in situ, but the taxa do occur in the Renggeri Member since we have found specimens in the drift. Hence, the ammonite succession in the Renggeri Member must be complete. *Cardioceras (Cardioceras) cordatum* (J. Sowerby) is very rare. The only specimen J 23027 in our collection from the Glaukonitsandmergel in an excavation near Siblingen, canton Schaffhausen (Pl. 1A), is figured in Gygi & Marchand (1982, Pl. 10, Fig. 1). The coeval taxon *Cardioceras (Cardioceras) persecans* (S. S. Buckman) is common and can be used for biostratigraphic correlation of the Cordatum Subzone instead of the subzone index (see Table 2). Gygi & Marchand (1982) figured specimens from the lower Terrain à Chailles Member of Gempen, canton Solothurn, from the Schellenbrücke Bed of Herznach, canton Aargau, and from the Glaukonitsandmergel Bed of Siblingen, canton Schaffhausen. Specimens of the taxon were also found near Bärschwil, Vellerat, and Sornetan.

3.4.2 *Middle Oxfordian*

The Densiplicatum Zone is equivalent to the Vertebrale Subzone of Great Britain. The Middle Oxfordian *Cardioceras (Vertebriceras) densiplicatum* Boden occurs in the Mumienmergel Bed near Siblingen. Two specimens are figured in Gygi & Marchand (1982, Pl. 11, Fig. 5 and 6, see also Fig. 2, section RG 212). The oldest *Gregoryceras (Gregoryceras) transversarium* (Quenstedt) we know is from the Mumienmergel Bed near Blumberg (see Fig. 1). This and other specimens from the Birmenstorf Member in canton Aargau are figured in Gygi (1977). A *Perisphinctes (Dichotomosphinctes) antecedens* Salfeld from Siblingen, very similar of the type of the taxon, is figured in Gygi (1966, Pl. 3, Fig. 2). Several more specimens have since been found in systematic excavations near Siblingen and Herznach (Gygi 1977). H. and A. Zbinden have taken *Perisphinctes antecedens* J 27994 from about 2.5 m above the base of the Pichoux limestone in the quarry of La Charuque near Péry (Pl. 1A). *Perisphinctes (Dichotomosphinctes) dobrogensis* Simionescu Gy 1824 as figured by Gygi (1966, Pl. 2, Fig. 2) is from the Mumienmergel Bed in an excavation near Siblingen. Another specimen J 23167 from the lowermost Birmenstorf Member was excavated near Herznach (Gygi 1977, p. 442). The taxon has the same vertical range as *Perisphinctes antecedens* (Table 3). *Perisphinctes (Dichotomosphinctes) dobrogensis* is important for regional correlation because two good representatives are known from the chalk-like limestone of the upper St-Ursanne Formation in the underground limestone quarry near the railway station of St-Ursanne. The best specimen from St-Ursanne is no 10327 kept in the Natur-Museum Solothurn. The cast J 30706 of it is in the Museum of Natural History Basel. V. Pümpin found *Perisphinctes (Perisphinctes) alatus* ENAY J 23074 at the same locality and in about the same level (Pl. 1A).

Glochiceras (Glochiceras) subclausum (OPPEL) Gy 1076 as figured by GYGI (1966, Pl. 4, Fig. 3) is a common taxon in the Birmenstorf Member of canton Aargau. We know from systematic excavations in canton Schaffhausen that *Glochiceras subclausum* makes its first appearance in the middle Antecedens Subzone (GYGI 1977). DE LORIOL (1901, Pl. 1, Fig. 6) figured the specimen J 30932 from a landslide near Montfaucon. The adhering matrix indicates that it is from the Terrain à Chailles Member. Sedimentation of this member then continued in the later part of the Antecedens Subchron. *Larcheria schilli* (OPPEL) appears after *Gregoryceras transversarium* and *Gregoryceras riazi* have disappeared. This statement is based on more than 1800 ammonites which were taken from four excavations and from several measured sections in the Birmenstorf Member in canton Aargau (GYGI et al. 1979, Fig. 3, see also GYGI 1977, p. 511). It is in agreement with what CARIOU (1966, p. 49) and ENAY (1966, p. 142) have found. *Larcheria schilli* (OPPEL) J 23539 is from horizon 47 in the lowermost Effingen Member of section RG 226 near Auenstein (GYGI 1973, Fig. 3). The taxon was cited by error from the Birmenstorf Member in that paper. Another specimen, J 27792, is from horizon 10 of the excavation RG 51 in the former cementstone quarry near Oberehrendingen (GYGI 1977, p. 446 and Pl. 11). *Larcheria subschilli* LEE J 30684 was taken by B. Hostettler from horizon 99 of section RG 307 in the quarry of La Charuque near Péry, about 24 m above the base of the Effingen Member.

Perisphinctes (Dichotomoceras) rotoides RONCHADZE occurs in the Schilli Subzone. It is one of the earliest representatives of its subgenus. We have four specimens, among which J 27971 is the best one, from horizon 46 of section RG 276 near Holderbank, canton Aargau (see Fig. 3 in GYGI et al. 1979). *Perisphinctes rotoides* of recent authors apparently are mostly *Dichotomosphinctes* from the Antecedens Subzone. *Perisphinctes (Dichotomoceras) falculae* RONCHADZE J 23737, *Perisphinctes (Dichotomoceras) stenocycloides* SIEMIRADZKI J 23701, and *Perisphinctes (Dichotomoceras) bifurcatoids* ENAY of the early Bifurcatus Chron are in horizon 50 of section RG 276 near Holderbank. *Perisphinctes (Dichotomoceras) bifurcatus* (QUENSTEDT) J 23543 is from the somewhat younger horizon 55 in section RG 226 near Auenstein (GYGI 1973, Fig. 3). Another *Perisphinctes bifurcatus*, J 30935, was found by ENAY (1966, p. 274) in the Günsberg Member near Günsberg (see GYGI 1969, Pl. 18, section RG 14, probably horizon no 137).

3.4.3 Late Oxfordian

Euaspidoceras hypselum (OPPEL) J 27259 (no 17 in Pl. 1A) was found by D. Krüger in a fallen block in the quarry of Jakobsberg near Auenstein (section RG 37, or no 9 in Pl. 1A). To judge of the site where the ammonite was found, and of the lithology of the material adhering to the specimen, the ammonite can only be from the limestone succession 55–80 of section RG 37 in GYGI (1969, Pl. 17). The limestone of this succession has about as high a carbonate content as the Gerstenhübel Beds. While the Gerstenhübel Beds are debris flow deposits with plastically deformed mud clasts, the limestone beds of the succession 55–80 are homogenous mudstone.

Only the Bimammatum and the Galar Subzones of the Late Oxfordian are well documented by abundant ammonite faunas. *Epipeltoceras bimammatum* (QUENSTEDT) Ve. S. 6686 as indicated in Plate 1A from near Auenstein was found in the Crenularis Member in a quarry east of the village (section RG 36, horizon 31, unpublished, for exact location see GYGI 1969, p. 8). The specimen is kept at the Federal Institute of Technology, Zürich. *Epipeltoceras bimammatum* Gy 1699 is from horizon 42 of the unpublished section RG 74 near Geisslingen, about 5 km across the Rhine river from Mellikon (Pl. 1A). The specimen Gy 1733 is from the upper Hornbuck Member near Erzingen (GYGI 1969, Fig. 2). GYGI (1969) found that in northern Switzerland, there is a gap between the last *Epipeltoceras bimammatum* and the first *Idoceras planula*. *Taramelliceras litocerum* (OPPEL) occurs in this gap, in the Küssaburg Member (GYGI 1969, Pl. 16, sections no 77 and 82).

The earliest *Ringsteadia* of northern Switzerland appear at about the same time as *Epipeltoceras bimammatum* (GYGI 1969, Fig. 2). The adult *Ringsteadia* sp. Gy 1343, complete with most of the peristome at a diameter of 550 mm, was found in horizon 6 of section RG 70 near Mellikon (GYGI 1969, Pl. 17). The inner whorls of the specimen resemble *Ringsteadia pseudoyo* SALFELD. A near-complete adult of *Ringsteadia anglica* SALFELD from the Knollen Beds (middle Villigen Formation) near Gosheim (southern Germany) is in the collection of the Federal Institute of Technology Zürich (without number, see GYGI 1969, p. 101). *Idoceras planula* (HEHL), *Idoceras schroederi* WEGELE, and *Idoceras laxevolutum* (FONTANNES) occur in the Letzi Member and in the Wangental Member of the upper Villigen Formation (GYGI 1969, Pl. 16 and 17).

The boundary between the Oxfordian and the Kimmeridgian Stages is conventionally drawn between the last *Sutneria galar* and the first *Sutneria platynota*. The interval between the two taxa is only about 10 cm in the excavation RG 239 near Schaffhausen (unpublished, made in 1974 at the same locality as section RG 83 in GYGI 1969, Pl. 16). The succession in the excavation is normal. There is no indication of a hiatus. Thus, the Oxfordian/Kimmeridgian boundary can be biostratigraphically indicated with unusual precision at this locality.

3.4.4 Kimmeridgian

Sutneria platynota (REINECKE) is well-represented in the excavation RG 239 near Schaffhausen (see above). There is also one specimen (Gy 1494) from near Mellikon (GYGI 1969, Pl. 17, section RG 70, horizon 120). Well over 200 *Ataxioceras* were studied from horizon 124 of section RG 70 near Mellikon. Among these is *Ataxioceras (Ataxioceras) suberinum* (VON AMMON) Gy 1548. This taxon from the Hypselocyclum Zone is easy to identify, and it is the most widespread in northern Switzerland: the adult and complete specimen J 25921 was taken from a fallen block in the cement quarry near Olten (GYGI 1969, Pl. 18, section RG 21, probably from horizon 57), and another is from the glauconitic horizon in the old quarry of Oberbuchsiten near Egerkingen (Fig. 1). The original of the latter ammonite (without number) is kept in the Natur-Museum Olten, and the cast J 30719 of it is in the Museum of Natural History, Basel.

Pachypictonia cf. *divergens* J 26468 listed in Table 3 is also from horizon 124 of section 70 near Mellikon. This and other Pachypictoniae from the same locality are probably from the Hypselocyclum Zone like the Pachypictoniae described by Schneid from southern Germany. We cannot be more specific because the exact level of individual ammonites found within horizon 124 was not recorded. Therefore it is not possible to discriminate subzone or even zone boundaries within this glauconitic limestone. CONTINI & HANTZPERGUE (1973) have figured a specimen they identified as *Pachypictonia indicatoria* from the equivalent of the Banné marl near Montbéliard (Fig. 1). Judging from their figures, we believe that the specimen is a *Lithacosphinctes* comparable with forms which range in northern Switzerland from the Bimammatum Zone to at least the Divisum Zone. The age of the Banné marl is very probably of the later Acanthicum Chron (see above). *Idoceras balderum* (OPPEL) first appears in horizon 124 of section 70 near Mellikon (J 24356) and continues into the lowermost Wettingen Member (GYGI 1969, Pl. 17, sections 62 and 70), where *Aspidoceras uhlandi* (OPPEL) is not rare (Table 2 and Pl. 1B). *Aspidoceras acanthicum* (OPPEL) wa taken from in situ near Regensberg (10 km east of Baden) from the uppermost marly member represented in Plate 1B (in a projected position about halfway between Mellikon and Siblingen). This specimen is now at the Institute of Paleontology of the University of Zürich. *Aspidoceras* cf. *acanthicum* was found in the limestone just below the Banné marl near Courgenay (see above and Pl. 1B).

3.5 The nature and distribution of the siliciclastic minerals

3.5.1 Distribution of phyllites

The distribution profiles of the argillaceous minerals are presented in Figures 9, 10, and 11. In the Oxfordian succession, the arithmetical mean of the different phyllites (Table 4) is in accord with the observations by PERSOZ (1982).

Illite-micas, kaolinite and mixed-layers are the principal constituents. In general, there is very little chlorite. In the platform realm (Courgenay to Reuchenette sections, Fig. 9) occurrences of smectites are few and in no regular pattern. Smectites are more common in the "basin" domain (Pfaffnau–Aargau) where generally the content increases mainly at the expense of kaolinite, but also of the mixed-layers and sometimes even of illite.

The greatest variations are in the kaolinite content. This mineral correlates negatively with illite-micas, with mixed-layers, the two other most abundant minerals, and often

Table 4: *Mean percentages of clay minerals in the Oxfordian.*

	Mean	Extreme values
Illite-micas	51	21 - 87
Kaolinite	22	0 - 78
Mixed layers	19	0 - 49
Chlorite	5	0 - 17
Smectite	3	0 - 63

with chlorites (Fig. 9, see for example: St. Ursanne Formation at Bressaucourt and Pichoux, Court Formation at Pichoux, Hauptmumienbank at Reuchenette, "Sequanien" at Pfaffnau, Villigen Formation in canton Aargau). With increasing kaolinite content, we observe a simultaneous decrease in the abundance of the three other phyllites. In the kind of clay mineral analysis we are using, where the total content of phyllites is 100%, the increase of one phyllite could be artificially due to the decrease of another, the former being in reality constant. However, in our case it would be difficult to explain a simultaneous decrease of illite-micas, mixed-layers and chlorite, which could cause an apparent maximum of kaolinite. For this reason we believe that most variations of kaolinite are probably real and not due to other phyllites. This seems not to be true in three cases (Fig. 9: Birmenstorf Member at Reuchenette, upper part of "Argovien" at Pfaffnau and upper part of Villigen Formation in canton Aargau). There, the decrease of kaolinite content could be apparently explained with a "dilution" process by a great increase of smectite. But even in this case we should observe a few percent of kaolinite. However, the total disappearance of kaolinite renders this interpretation unrealistic.

The scarcity of smectites and the abundance of mixed-layers is a prominent feature of this association as compared with other Jurassic rocks in northern Switzerland. The mixed-layers have a variable rate of interstratification between illite and smectite and possibly chlorite (Fig. 13 and 14a). They are irregular. These observations can be made in nearly all of the Oxfordian rocks of the Jura Mountains and in the inframolassic domain of Switzerland (Persoz 1982). Bausch (1980) has made a study of argillaceous associations of the European Upper Jurassic, based on 300 samples from 27 sections. He concluded that the absence of smectite and the abundance of mixed-layers is a typical feature of the deposits in the epicontinental seas around the hercynian massives of France and Germany. We can confirm this for Oxfordian time, but this is no longer the case in the later stages of the Upper Jurassic where smectites are more and more abundant up to the Cretaceous where they may locally attain great importance (Persoz & Kübler 1968, Persoz et al. 1979).

3.5.2 Distribution of kaolinite and correlations

Correlations between sections are made by a sequential analysis of the morphology of the kaolinite distribution profile (Fig. 10 and 11). Evidently this is possible only in an area where the lithostratigraphic framework has been clarified previously.

In this respect, it is necessary to distinguish two parts in the rock sequence: a) the Upper Oxfordian and Lower Kimmeridgian, and b) the Middle and Lower Oxfordian. In the Upper Oxfordian it is often possible to correlate each peak of the distribution profiles, for example peaks 6 to 8 in Figure 10 between Pichoux and Aargau, or peaks 6a to 10 between Noirvaux and Areuse (Fig. 11). In addition, refined analysis of the morphology of the peaks sometimes corroborates the proposed correlations. This is the case between Pichoux and Aargau. In the Pichoux section, the kaolinite content increases sharply at the top of the Natica Member, while the same is observed in canton Aargau at the top of the Effingen Member (peak 6, Fig. 10). This sharp rise is followed by a near-regular slope between peaks 6 and 7 (Fig. 10 and 11). The abrupt diminution of the kaolinite content on the upper side of peak 10 (Fig. 10 and 11) is also conspicuous in each section. In the inner

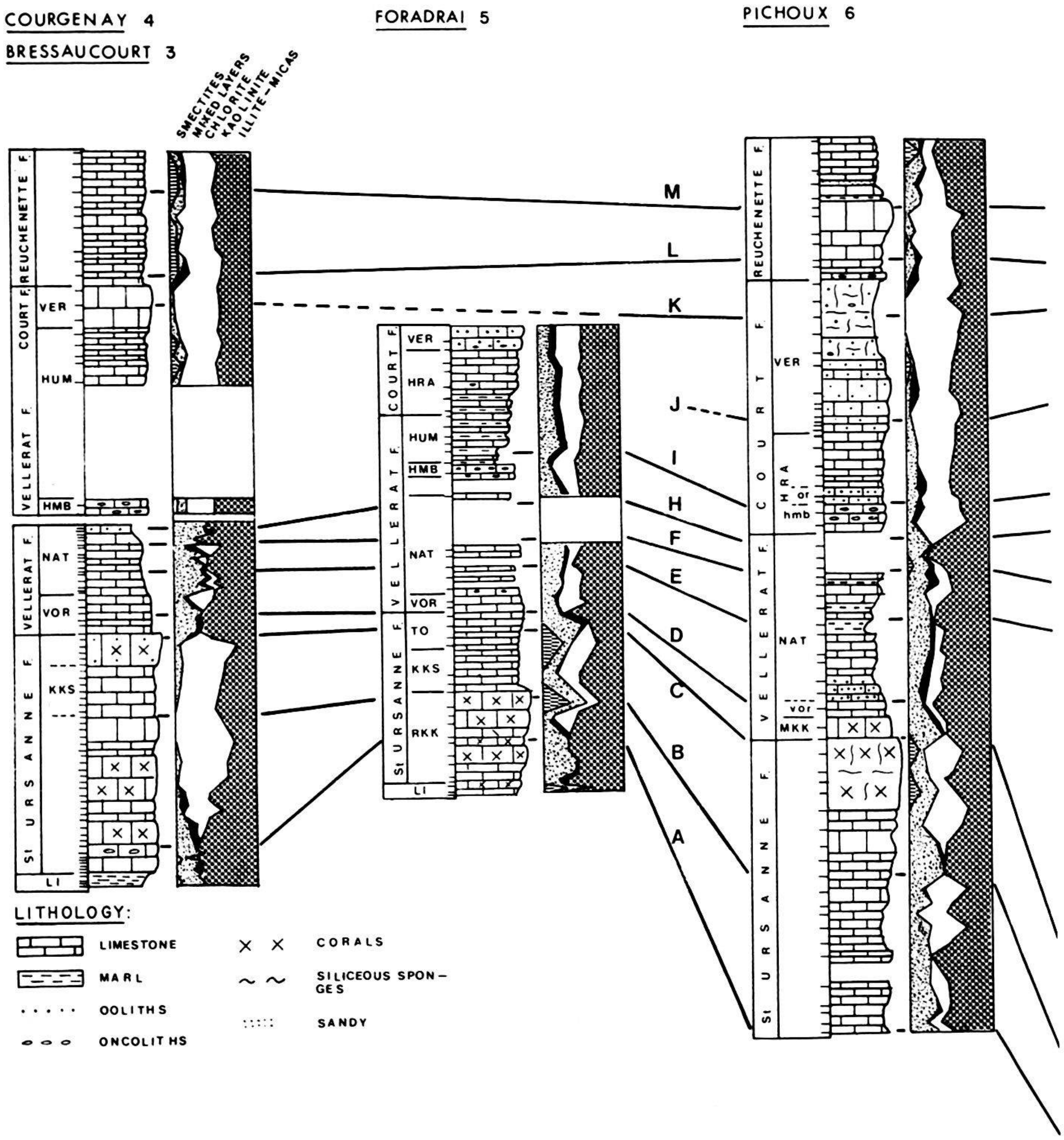
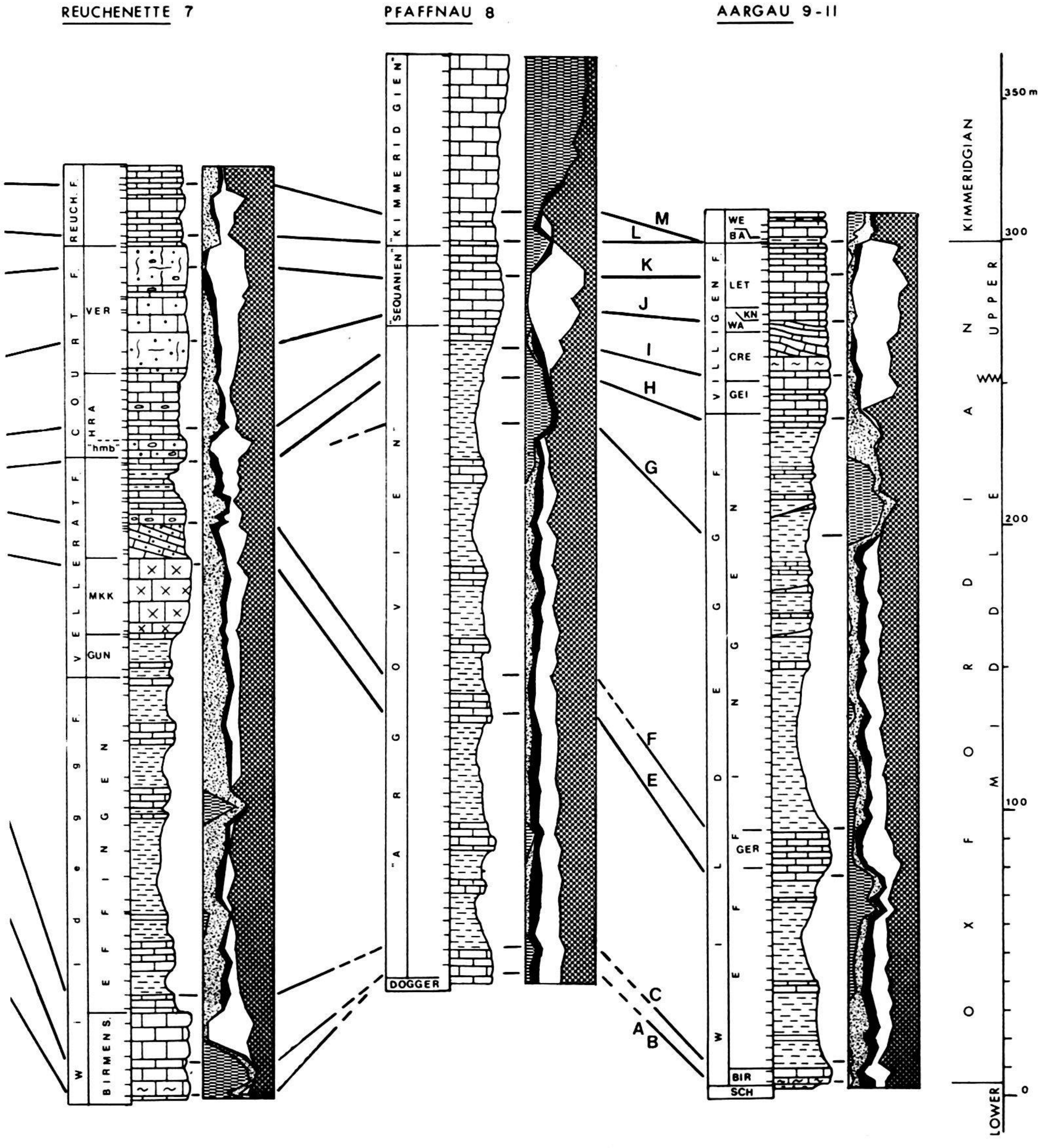


Fig. 9. Lithologic sections with clay mineral distribution, from the carbonate platforms in canton Jura to the deeper marine realm in canton Aargau. The section "Aargau" is assembled of sections no 9 through 11 in Plate 1. Names of formations and members are by THALMANN (1966, p. 32), GYGI (1969, p. 104), and BOLLIGER & BURRI (1970, p. 70).

part of the platform (Foradrai, Courgenay) there is less variation in the kaolinite content and correlations require more interpretation.

In the Middle and Lower Oxfordian (below the Hauptmumienbank and the Geissberg Members), there is more than one possible solution in correlating the kaolinite distribution profiles (Fig. 10). The proposed correlations are constrained by the discovery of several ammonites (Pl. 1A). In the proposed model, the three peaks at Bressaucourt in the St-Ursanne Formation are reduced to two at Foradrai, two at Pichoux and Reuchenette, and one in the Aargau section. The unequal homogeneity in the succession of peaks in the different sections is interpreted as being caused by too large intervals between



samples. In fact, the correlations A to C reflect very well the thinning out of the Pichoux limestone towards the "basin" (Pl. 1A).

The peak 4 (Fig. 10) between correlations D and E, very distinct at Bressaucourt and Foradrai, more attenuated at Pichoux, is no longer apparent further in the "basin" where it disappears by an increasing content of the other phyllites.

In Figure 11 it is possible to extend the model of correlation to the western part of the Jura Mountains, to the section of Areuse (PERSOZ 1982) and Noirvaux (KETTIGER 1981). These two sections are in a position, with respect to the isopic contours, comparable with Reuchenette (Fig. 1) at a distance of about 40 km. In the two sections, the profiles of

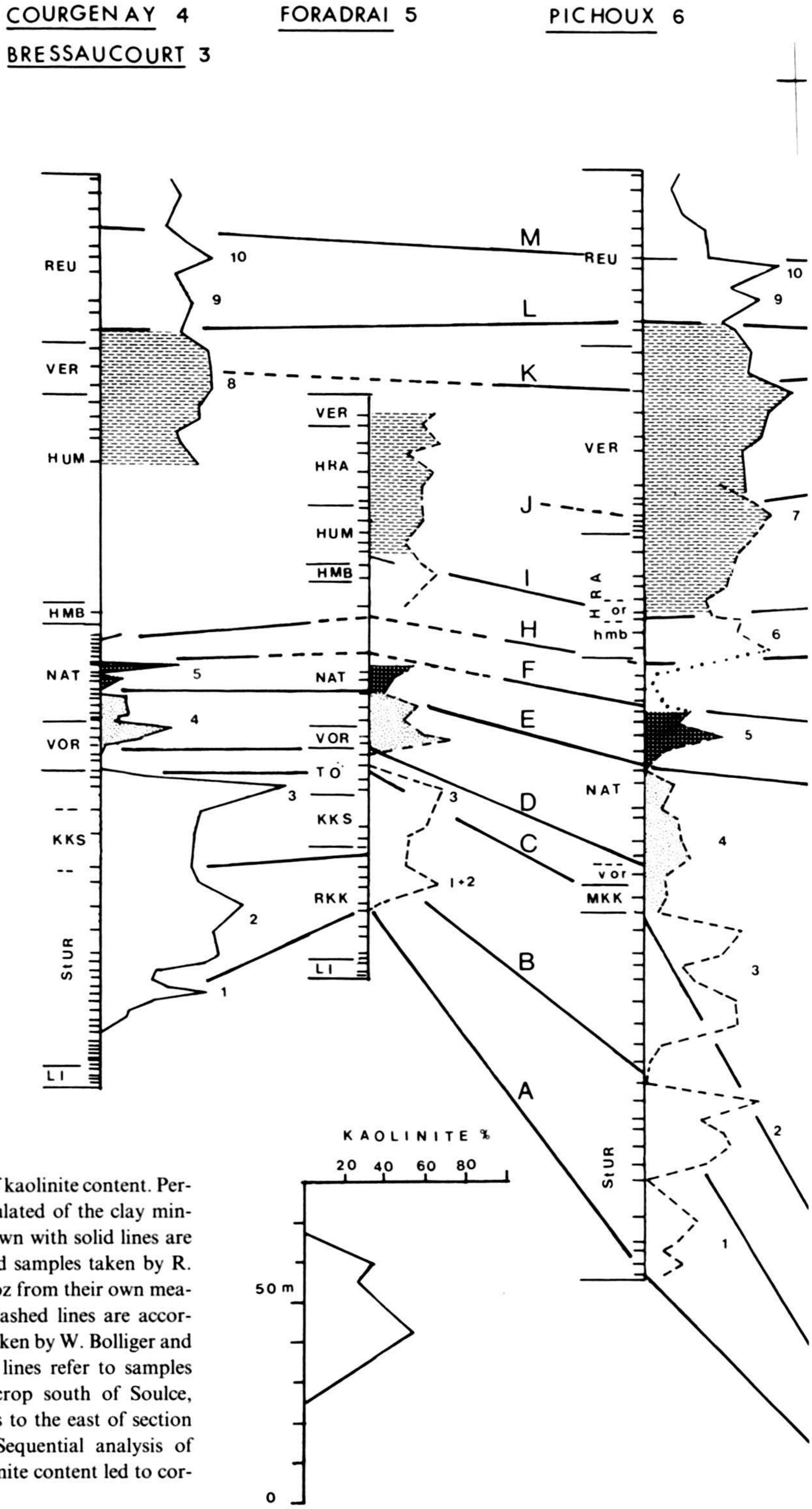
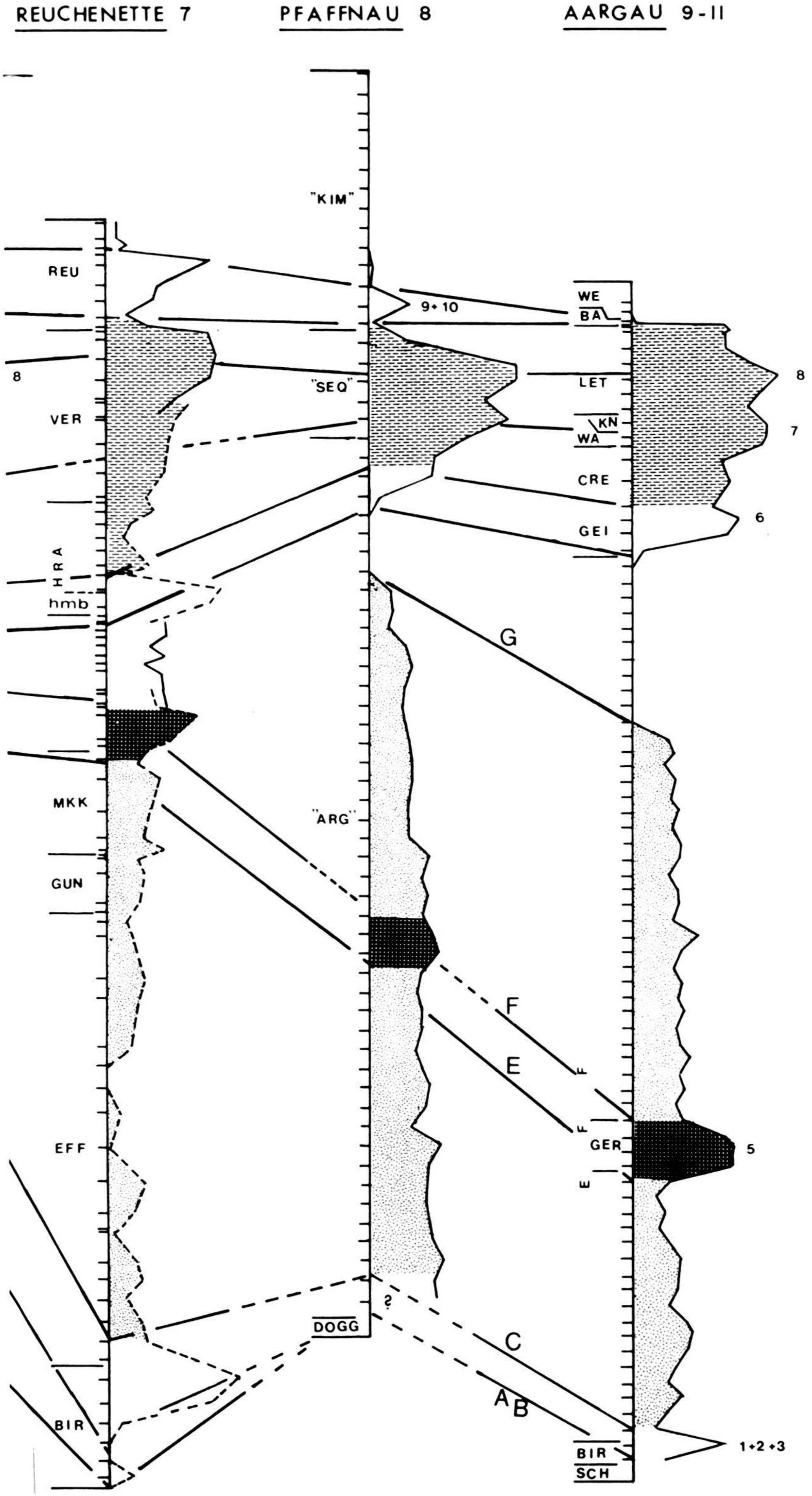


Fig. 10. Profiles of kaolinite content. Percentages are calculated of the clay minerals. Profiles drawn with solid lines are based on analyzed samples taken by R. Gygi and F. Persoz from their own measured sections. Dashed lines are according to samples taken by W. Bolliger and P. Burri. Dotted lines refer to samples taken at an outcrop south of Soulce, several kilometers to the east of section no 6, Pichoux. Sequential analysis of variation in kaolinite content led to correlations A to M.


REUCHENETTE 7
PFAFFNAU 8
AARGAU 9-11
"KIM"
REU
9+10
WE
BA
8
"SEQ"
LET
8
VER
KN
WA
7
CRE
GEI
6
HRA
hmb
G
MKK
"ARG"
GUN
F
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GER
5
EFF
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DOGG
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A B
BIR
BIR
SCH
1+2+3

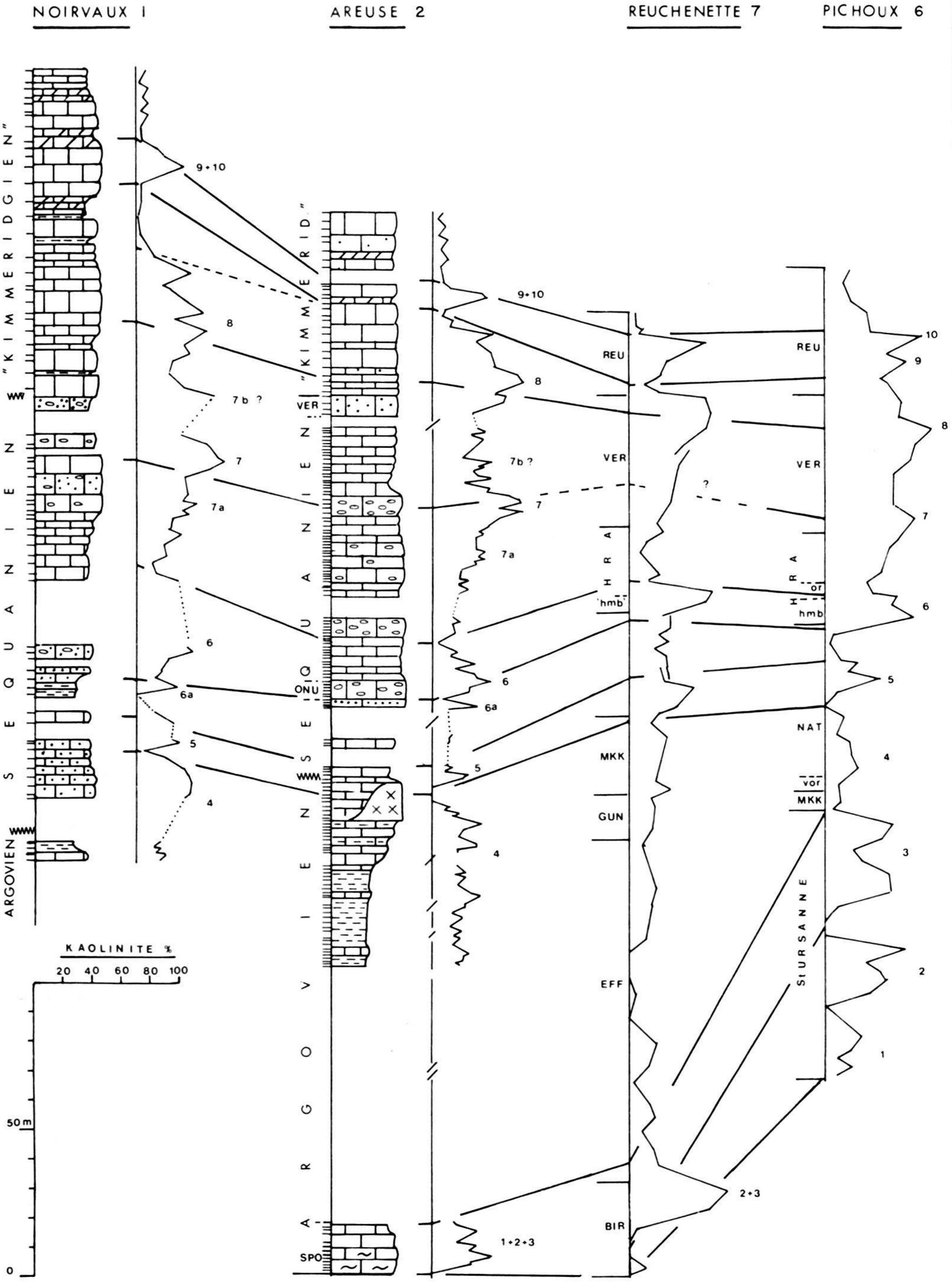

Fig. 11. Lithologic sections with profiles of kaolinite content of canton Neuchâtel compared with sections no 6, Pichoux, and 7, Reuchenette, of Figure 10. Section no 1 of Noirvaux is taken from KETTIGER (1981). Section no 2 of Areuse is assembled from several sections as represented in Figure 1 (see PERSOZ & REMANE 1973).

kaolinite distribution compare well with the profiles in Figure 10. The sample spacing in Areuse section being dense, this is an indication that the profiles in Figure 10 are reliable.

The correspondence between the correlations in this study with those by PERSOZ (1982), which cover a much larger area between the Lakes of Constance (Bodensee) and Annecy, are the following: correlation I = M3, J = M5, K = M6, L = M8. It appears then that the general distribution as represented in PERSOZ (1982) remains more or less constant over a distance of more than 200 km to Annecy.

3.5.3 *Kaolinite and lithology*

In "pure" limestones, kaolinite is commonly enriched mainly at the expense of illite-micas in relation to marls. The carbonate content of marls varies widely around a mean of about 60% carbonate. Nevertheless, there are many exceptions to this. For example, the marls of the Humeralis Member at Foradrai, and the marls of the Natica Member at Reuchenette contain nearly as much kaolinite as the limestones. On the other hand, in many calcareous parts of the sections, there is little or no kaolinite. Even in the "pure" limestones of the central carbonate platform, the kaolinite content can vary from 0 to 90%. For example, there is no kaolinite in the lowermost part of the St-Ursanne Formation at Bressaucourt and Foradrai. At Pichoux, there is much kaolinite in the coral limestone of the St-Ursanne Formation. This constrasts with the Moutier Korallenkalk above, where the kaolinite content is strongly reduced (Fig. 9). The same observation can be made in oolitic, oncolitic and "pure" micritic limestones. The kaolinite content is remarkably constant in the Verena Member which is micritic near Courgenay, mostly oolitic in the Pichoux and Reuchenette sections and pure micrite in the Aargau section (Fig. 9, Pl. 1A). Therefore, there is no clear-cut relation between lithology and kaolinite content. However, there is a tendency of kaolinite content being elevated in limestones.

3.5.4 *Kaolinite and depositional environments*

In the proposed model of correlation, the profiles of kaolinite variation can be traced from backreef to the reef, slope, and "basin" environments (Pl. 1).

In most cases the concentrations of kaolinite have no regular gradient. In two cases, the kaolinite content decreases towards the "basin": 1) Peaks 9 and 10 (lower Kimmeridgian) between Pichoux and Aargau, and 2) peak 4 in the Middle Oxfordian between Foradrai and Reuchenette, which are in the same transect (Fig. 10). In the wells of Courtion (Fig. 1) and Berlingen (PERSOZ 1982, Fig. 1 and 5), which are in a more "basinal" position in respect to Reuchenette and Aargau in Middle Oxfordian time, the mean kaolonite content diminished from about 15 to 10% as compared with a corresponding increase of mainly illite-micas. This observation confirms the negative gradient of the kaolinite content from the platform to the "basin". Yet this pattern is overprinted by another one: the kaolinite content is at a maximum in the Middle Oxfordian in the area Aarau–Pfaffnau, and it decreases both to the southwest (Courtion) and to the northeast (Berlingen).

Along a given anomaly, the general aspect of diffractograms is very similar. A good example is peak 6 of kaolinite in Figure 10 from the Hauptmumienbank and Geissberg Members, where the only differences are due to quartz and goethite contents and a

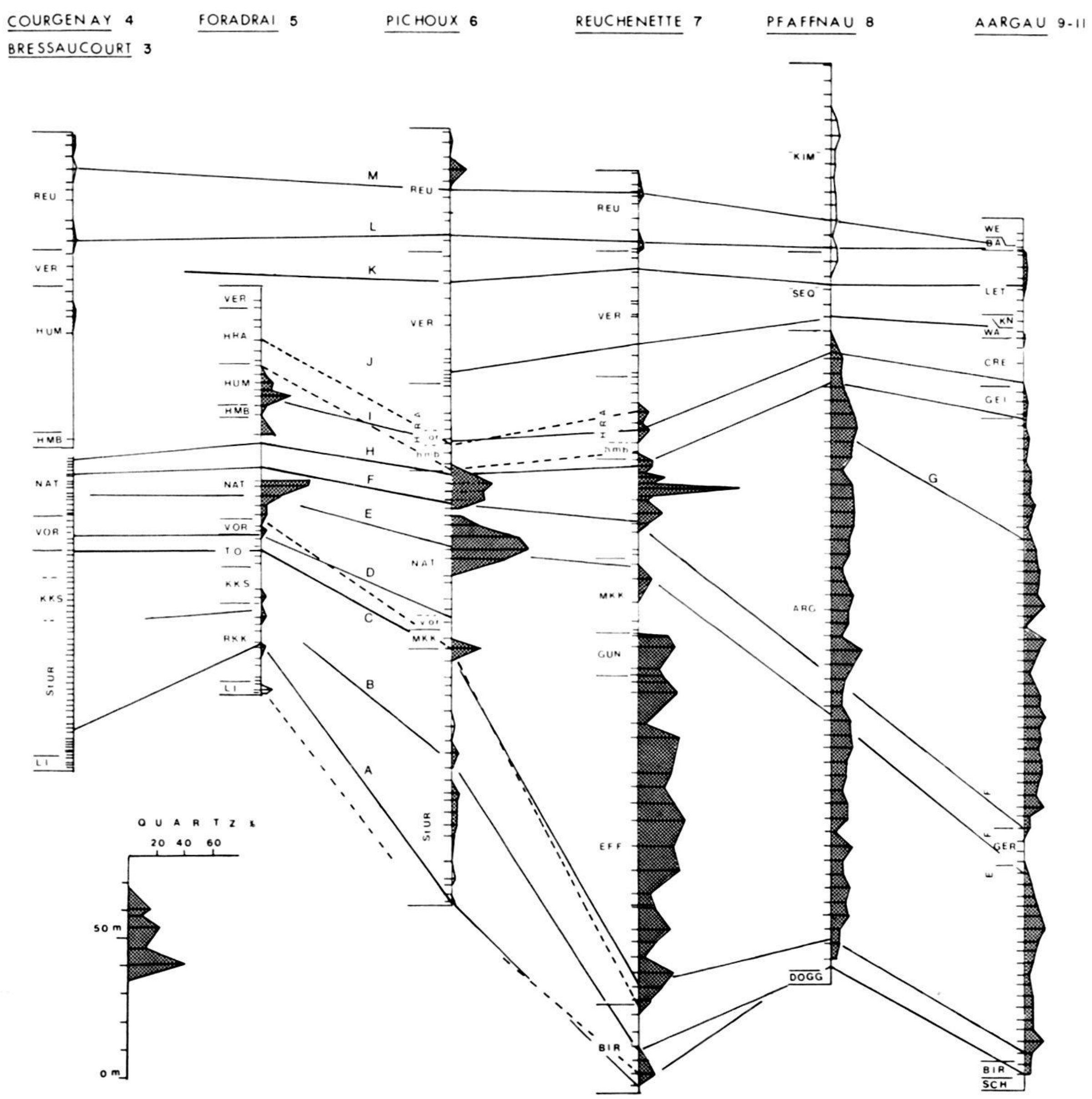

Fig. 12. Vertical variation in quartz content, measured with x-ray diffractometry of whole rock samples. Correlations with solid lines are those based on kaolinite represented in Figure 10. Dashed lines are correlations made by Bolliger & Burri (1967) according to the vertical variation in quartz content they obtained from thin section evaluation.

tendency to form more discrete swelling mixed-layers in the "basin" realm (Fig. 13). Between the Verena and Letzi Members (peak 8, Fig. 10, and Fig. 14a, spec. 14218, 119) of Reuchenette and Aargau, the diffractograms are equally very similar with again a tendency to form more discrete mixed-layers. On the other hand, the diffractograms are often different in lithologic units of different age, for example between the lower part of the Reuchenette Formation in the section of Reuchenette (peaks 9 and 10 in Fig. 10) and the Baden and Wettingen Members in the Aargau section (Fig. 14a, spec. 14224, 14227, 126, 135). However, there are many exceptions to this similarity of diffractograms in coeval beds.

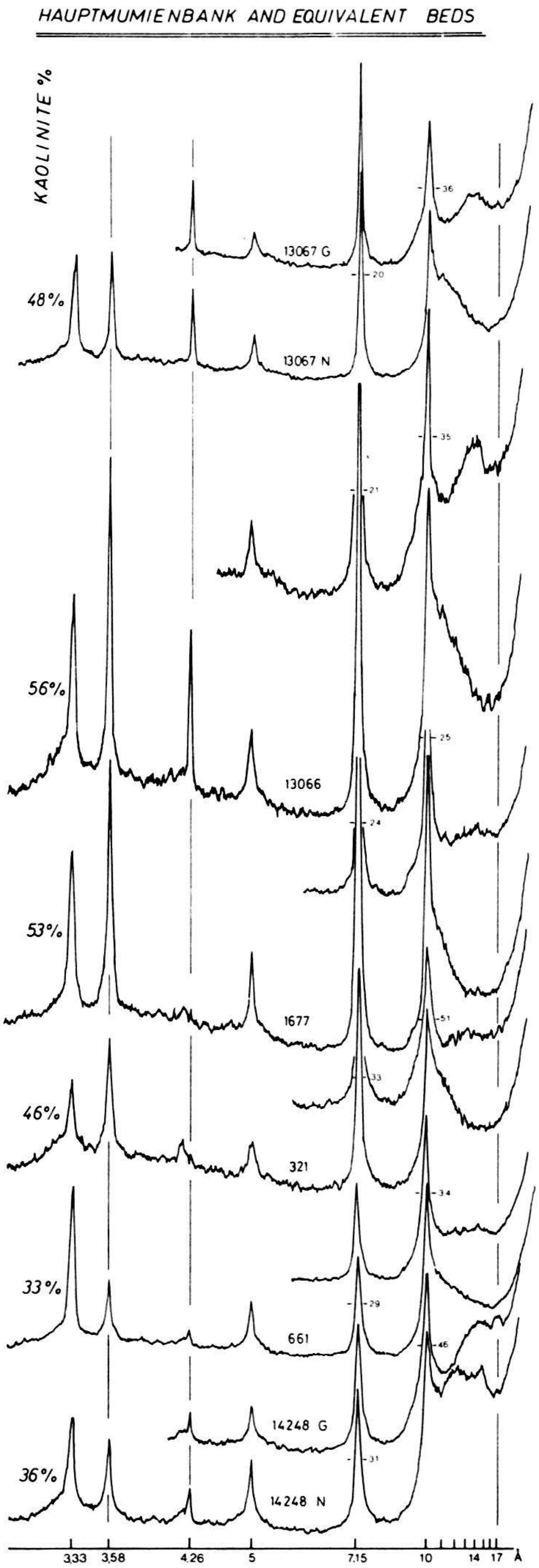

Fig. 13. Diffractograms of oriented samples, air dried (N), and treated with ethylene glycol (G). Source: Cu Kα, Philips apparatus, specifications: see PERSOZ (1982). Samples are from Hauptmumienbank and Geissberg Members: 14248 Courgenay, 661 Foradrai, 321 Pichoux, 1677 Reuchenette, 13066-67 Aargau.

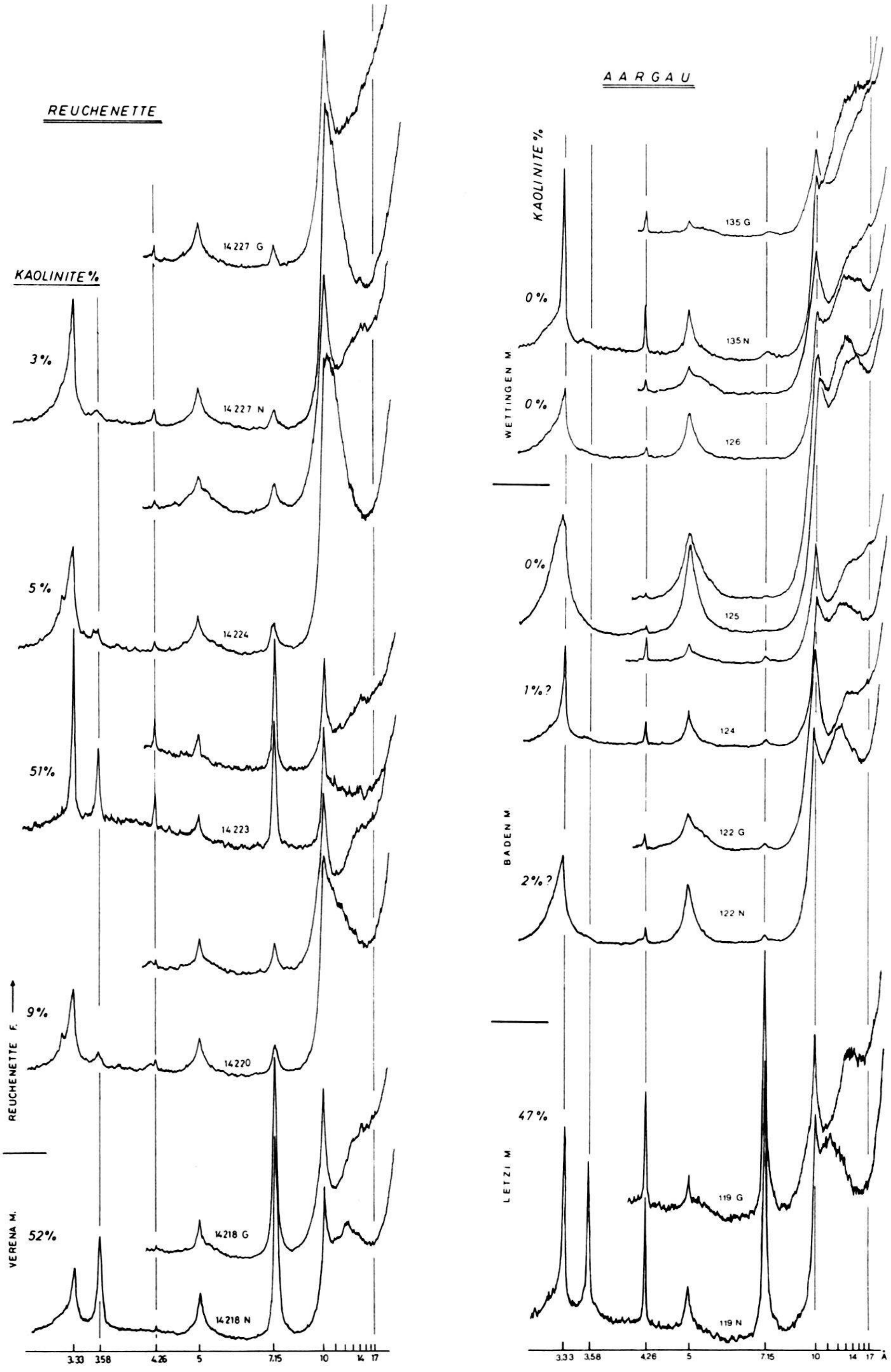
REUCHENETTE
KAOLINITE %
3 %
5 %
51%
9 %
52%
14227 G
14227 N
14224
14223
14220
14218 G
14218 N
REUCHENETTE F.
VERENA M.
3.33
3.58
4.26
5
7.15
10
14
17
AARGAU
KAOLINITE %
WETTINGEN M
0 %
0 %
135 G
135 N
126
0%
1 %?
125
124
BADEN M
2 %?
122 G
122 N
LETZI M
47 %
119 G
119 N
3.33
3.58
4.26
5
7.15
10
14
17
Å

Fig 14a

3.5.5 Crystallinity of illite-micas and kaolinite

In Figure 15, the crystallinity index of illite-micas (IAGI) is compared with that of kaolinite (IANK) in the different environments from the platform to the "basin" for the principal units correlated. A survey of Figure 15 demonstrates

a) a general positive correlation of IANK and IAGI, except in some cases (correlation 1, 2, 3, 8, 9),

b) for the platform, the correlations are better than on the slope and in the "basin" realm;

c) 41% of all measured IAGI values are lower than 0.42°2θ which is, for the KÜBLER index, the limit between the diagenetic and the anchimetamorphic zones;

d) the illite-micas with IAGI lower than 0.42°2θ are generally accompanied by kaolinites with equally low values of IANK (75% of all samples with IAGI less than 0.42°2θ have values of IANK lower than 0.3°2θ).

The interpretation of these facts is not simple, since the necessary parallel observations with the electron microscope have not been carried out. It is known that the sharpness of a peak in a diffractogram is an expression of at least two parameters (GUINIER 1964, BRINDLEY & BROWN 1980): 1) stacking defects like interstratifications, incomplete substitution, or presence of foreign cations for instance, and 2) particle size. The last is only effective on the sharpness of a peak for particle sizes less than about 0.1 micron (BRINDLEY & BROWN 1980). The crystallinity index grows with an increasing abundance of the defects, and with diminishing particle sizes. KÜBLER (1964, 1966) has shown that the crystallinity index of illites is a good indicator of diagenesis and of very low grade metamorphism. Thus, low values probably less than 0.42°2θ on the IAGI scale, define inherited anchi- and/or epimetamorphic micas (PERSOZ 1982). Therefore, it appears that about half of the analyzed specimens are derived from metamorphic terranes, either directly or with one or more intermediate stages of sojourn in sediments.

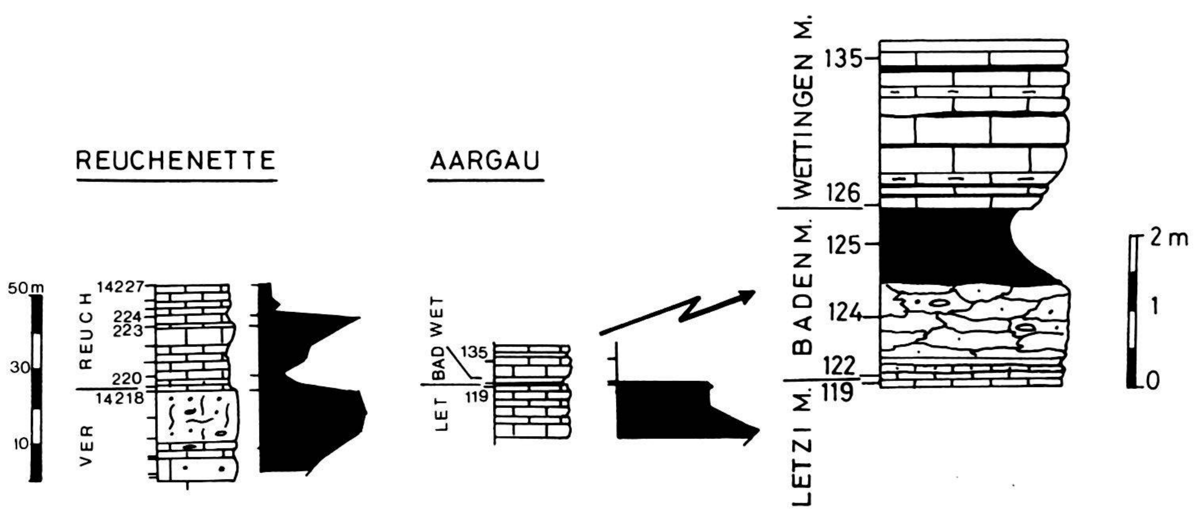

Fig. 14. Diffractograms of oriented samples from the Oxfordian–Kimmeridgian transition compared between section no 7, Reuchenette, and 11, Mellikon (Aargau). To the right of lithologic sections in Fig. 14b are the profiles of kaolinite content, and to the left are the numbers of analyzed samples. The upper part of section 11, Mellikon, is enlarged 20 times (cf. GYGI 1969, Pl. 17, section no 70).

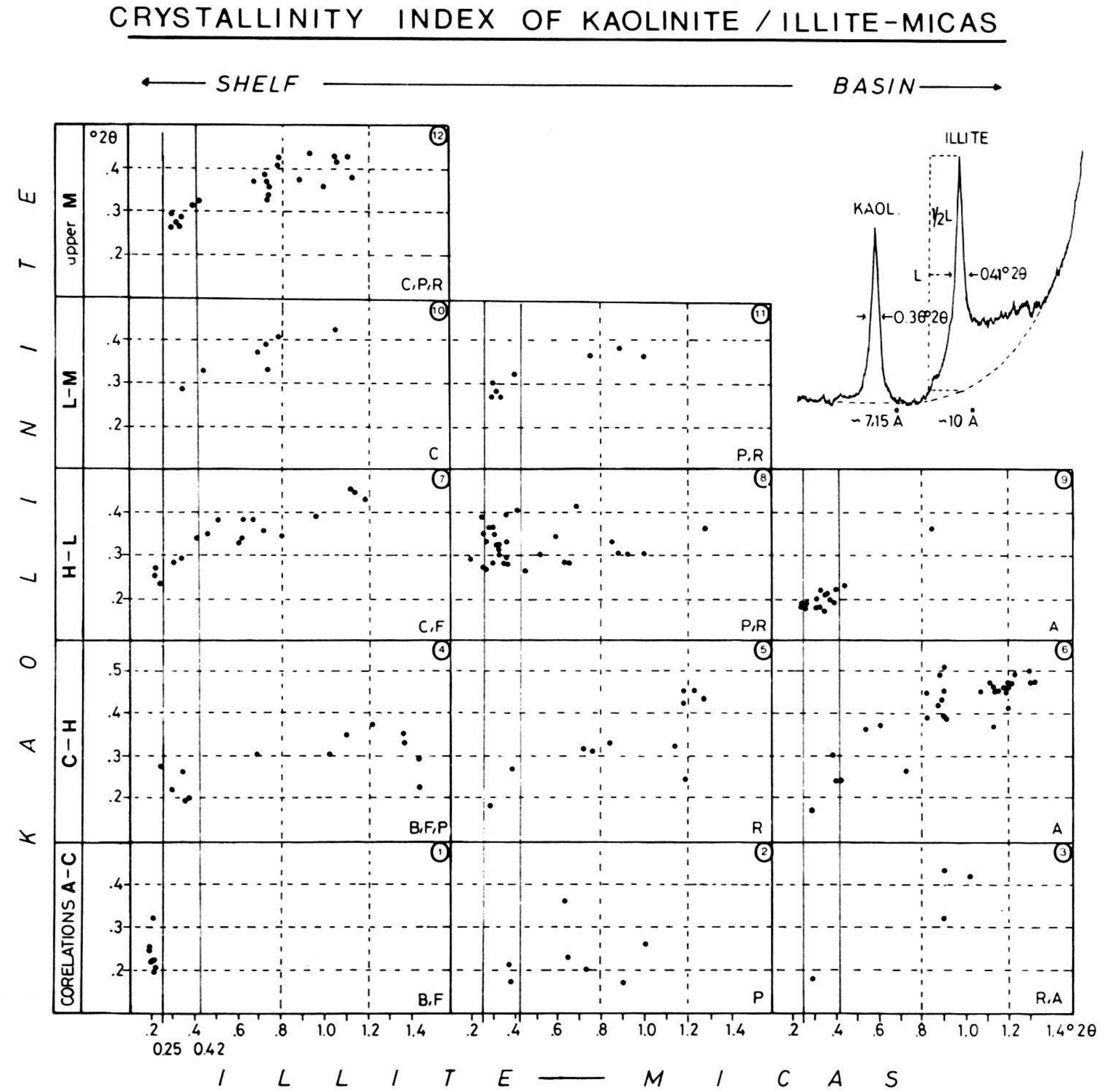

Fig. 15. Platform to basin variation in the crystallinity index, measured at half the elevation of peaks, (001) of illite at about 10 Å, and (002) of kaolinite at 7.15 Å. The correlation fields, numbered from 1 to 12, are arranged in vertically successive blocks, which are related to the stratigraphic correlations A to M in the left column (cf. Fig. 9). Letters down right in individual cases are abbreviated names of sampled sections: B for Bressaucourt, C for Courgenay, etc.

For kaolinite, the crystallinity index on the (001) peak reflects essentially the same parameters (THIRY 1982), but it is not related to anchi- or epimetamorphism, because kaolinite is not stable in metamorphic rocks. The association of well-crystallized illites and well-crystallized kaolinite in one sample cannot be explained either by erosion and mixing processes in the source area, or by diagenetic processes in sediments which were never at a burial depth in excess of about 1 km (PERSOZ 1982). One possible explanation of this association of kaolinite and illite with correlated degrees of crystallinity may be segregation of particle size in the course of hydrodynamic processes. This explanation is supported by the fact that IANK/IAGI correlations are generally better on the platform where wave turbulence is an important hydrodynamic factor as compared with the "basin". If this hypothesis is correct, we see that in each domain and even in each facies there are all values of the crystallinity index, and that any regular gradient from platform to "basin" does not exist (Fig. 15). The only exception is unit C to H in Figure 15 (correlations 4 to 6 where IAGI and IANK grow from the platform interior to the "basin"). However, this is uncertain, because chlorite influences the crystallinity of kaolinite when it is abundant, and chlorite is more abundant in the "basin" as compared with the platform.

3.5.6 Distribution of quartz

BOLLIGER & BURRI (1967) interpreted the quartz distribution as a means for stratigraphic correlation. They found three maxima of quartz content. A small one is at the base of the Middle Oxfordian, the main maximum is in the Natica and in the Effingen Members, and a third, small one is in the lower part of the Court Formation. The quartz content was estimated in thin sections and in insoluble residues, or in washed samples.

We re-examined this distribution by x-ray analysis (Fig. 12). The feldspars were not quantified. This is without consequence since feldspars are never more than a few percent in relation to quartz. We did not find all the maxima of BOLLIGER & BURRI (1967), possibly because our analyzed samples were too widely-spaced. Our distribution is at variance with the one found by BOLLIGER & BURRI (1967) in relevant details. With mineralostratigraphic data alone, it would have been difficult to rediscover their correlations. Between Pichoux and Reuchenette (Fig. 12), their correlations fit well with ours which are based on kaolinite. There is less congruence between Foradrai and Pichoux. But on the whole, our correlations are in agreement with theirs.

4. Interpretation and discussion

4.1 Provenance and variation in siliciclastic minerals

4.1.1 Provenance of kaolinite

1. The classical works by GRIM (1953, 1958), KELLER (1956, 1970), MILLOT (1952, 1964), and WEAVER (1958, 1960) gave ample evidence for clay mineral associations to be modified at each step of their transport, sedimentation, and diagenesis. In each environment, the inherited associations can be preserved (héritage of MILLOT 1964) or modified, either by transformation of one or several clay minerals, or by neoformation.

In carbonate environments with normal salinities, most students do not believe in the possibility of neoformation of kaolinite (FÜCHTBAUER & MÜLLER 1963, MILLOT 1964, and WEAVER 1967). The principal reason is the high activity of Ca and Mg and the high p_H (HEGELSON et al. 1969, KELLER 1970, and LIPPMANN 1979). In the late diagenesis and during the early steps of burial diagenesis, kaolinite is most often preserved in carbonate rocks with a normal geochemical character (PERSOZ 1982). Preservation of kaolinite is often explained by the low permeability of carbonate sediments and therefore very low activity of water at the level of microenvironments (KÜBLER 1966, PERSOZ 1982). KELLER

(1970) claims that dissolution of kaolinite should be possible in marls with a high carbonate content, but he thinks that this process is subordinate. Therefore, kaolinite is mainly an inherited mineral.

2. We infer that in Oxfordian time, the source rocks were the emergent Hercynian massifs and their sedimentary cover. The presence of a great quantity of well crystallized illite-micas in the argillaceous association, which can only be derived from crystalline rocks or from reworked sedimentary rocks in which the original crystallinity was preserved, is evidence for this provenance.

The location of the source rocks in the Rhenish and Bohemian Hercynian massifs is supported by two arguments: a) the distribution of kaolinite itself and b) the grain size of quartz.

a) The kaolinite content diminishes from the platform to the "basin" in the Middle Oxfordian (maximum 4 in Fig. 10), in the lower Kimmeridgian (Reuchenette Formation), and in the Effingen Member from Reuchenette to Courtion and from Aargau to Berlingen. In the other part of the analyzed stratigraphic record we do not see any tendency.

The negative gradient in the kaolinite content was often explained as a polarity vector in the detritic supply. This interpretation was supported by PARHAM (1966) in old sedimentary sequences and by GRIFFIN & GOLDBERG (1963), BISCAYE (1965), and PORRENGA (1966) in Recent sedimentation.

The mechanisms most favoured by these authors are: 1) particle size segregation, and 2) flocculation. In the first case, clay minerals of smaller size (smectite, illite, mixed-layers) migrate further than the relatively coarse-grained kaolinite (PARHAM 1966, GIBBS 1974). In the second case, kaolinite settles faster than the smectites due to the differential rate of flocculation (HAHN & STUMM 1970). GRIM et al. (1949) explained this gradient by dissolution of kaolinite with neoformation of illite during their transport in marine environments. THIRY (1982) has shown by analysis of the crystallinity that the finest-grained kaolinite crystals travel the greatest distance, possibly because they have a greater electrical charge than the larger crystals. It was not possible for us to demonstrate a distinct gradient of crystallinity neither for kaolinite nor for illite. This can be caused by too large sample spacing. We have to bear in mind that individual beds cannot be traced from platform to "basin".

In the Middle Oxfordian, the lower content of kaolinite in the Effingen marls could be caused by diagenetic processes where 2:1 phyllites and chlorite are aggraded in respect to kaolinite. Therefore, the only case of an unambiguous kaolinite gradient is in the "pure" limestone of the lower Reuchenette Formation.

b) Another argument in favour of a N–S polarity of the detritic supply was presented by BOLLIGER & BURRI (1967, 1970). They have shown the mean grain size of quartz to diminish by about 20% (from 40 to 25–30 microns at the base and from 45–50 to 40 microns at the top of the Natica and the Effingen Member, respectively) over the distance of 60 km from platform to "basin". In addition, the quartz content decreases towards the "basin" (Fig. 12).

3. BOLLIGER & BURRI (1970) thought the detrital quartz to be of eolian provenance. We do not agree with this for the following reasons: The rate of sedimentation of quartz in the Effingen Member is of the order of 0.02 mm/year. This figure is five times greater than Recent eolian sedimentation in proximity of endorheic basins with scanty vegetation

(BUCHER & LUCAS 1984) and 10 to 40 times greater than the rate measured in the North Atlantic (WINDOM & CHAMBERLAIN 1978).

In Oxfordian time, northern Switzerland was in the tropics or in the subtropics (HALLAM 1975). The vegetation was probably thicker than in an endorheic environment (see GYGI 1986), and the share of eolian quartz in the total quartz sedimentation was probably no more than 20%. So small a proportion cannot cause the strong variation in quartz content as was found by BOLLIGER & BURRI (1967) and used for their stratigraphic correlations.

4.1.2 Variation in kaolinite content

It is possible to follow each maximum and minimum of kaolinite content through different lithologies and depositional environments, which are often very different. This is a strong argument for a detrital provenance for the kaolinite.

This means that the observed variation in the kaolinite pattern in the sediment reflects the variation in the detrital supply. It also means that the original clay mineral associations were not much affected by the depositional environment and diagenesis (hydrodynamism of water, low variation in salinity, processes of oxidation or reduction at the sediment-water interface, etc.).

However, the influence of diagenesis and the depositional environment should not be disregarded. Several facts and observations are not explained by a model of pure detrital clay mineral sedimentation, for example: the absence of a kaolinite gradient in the Vellerat Formation, the preferential enrichment of kaolinite in limestone in relation to marl, the apparently erratic distribution pattern of the rare smectites, and the good correlation between kaolinite and illite-mica crystallinity.

Recently, several authors (VIEBAN 1983, DARSAC 1983, and ADATTE & RUMLEY 1984), stressed the importance of the depositional environment as defined by a correlation of microfacies classes with clay mineral assemblages. In these publications, the "roentgenofacies" (morphology of diffractograms) are classified into groups of similar appearance, which are then compared with the classes of microfacies. This leads to a characterization of each depositional environment by a mineral facies. Unfortunately, these authors failed to make a systematic comparison of the vertical variation in the clay mineral content between sections. Therefore, the influence of the detrital mineral source and environmental factors as sediment transport, differential settling, or maybe diagenesis, cannot be discriminated in their depositional environments. In fact, their correlation of mineral assemblages ("roentgenofacies") with distinct depositional environments is not unequivocal.

In these models referring to the Urgonian (VIEBAN 1983) and the Valanginian (DARSAC 1983) in the transitional domain between the Jura Mountains and the French Subalpine Chains, kaolinite is the main phyllite of the platform interior; a mixture of kaolinite, smectites and mixed-layers is characteristic of the slope, and smectites predominate in the basinal facies.

In the Valanginian of the Central Jura Mountains (ADATTE & RUMLEY 1984), individual depositional environments are characterized by varying proportions of micas, mixed-layers, and smectites, whereas there is no significant change in kaolinite content.

This is in good agreement with the conclusions by PARHAM (1966), but it cannot be adopted without reserve in the Oxfordian. The reason for this is maybe the absence of smectites in the original detrital supply (see below).

BAUSCH (1980), in his paper on the nature of clay mineral associations in the European Upper Jurassic, explains the variation in kaolinite by changes in climate. This explanation is possible, but if we consider the time interval of an Oxfordian kaolinite sequence (from one maximum to the next, see our Fig. 10) we see that it is of the order of one-fourth to one ammonite subzone. This means that several mechanisms can be invoked, for example: reactivation of tectonics in local domains, changes in climate and fluctuations of the detrital supply by modifications in the course of rivers (GRIFFIN 1962), etc.

4.1.3 The smectites

The scarcity of smectites and the relative abundance of mixed-layers are a fundamental feature of the clay mineral associations of the Oxfordian of the Swiss, French, and the Swabian Jura, and, after BAUSCH (1980), of all Oxfordian sediments which were deposited around the Hercynian massifs of Central Europe outside the Alpine realm. This scarcity can be explained in two ways: a) the absence of smectite is an original feature of the sediment supply; b) it is due to the transformation of smectites into mixed-layers by diagenesis (PERRY & HOWER 1970, EBERL & HOWER 1977, and KÜBLER 1984).

Two arguments are in favor of the first possibility:

1. Most of the analyzed Oxfordian sediments (this report, BAUSCH 1980, and PERSOZ 1982) apparently were never buried deeper than 1000–1500 m. Most authors believe that the stability field of smectites is within this burial depth interval (see KÜBLER 1984).

2. The low content of smectites is about the same in the Oxfordian of the Jura Mountains as in the deeper-buried inframolassic domain of Switzerland. In both areas, the Middle Jurassic below the Oxfordian and the upper Kimmeridgian and the Tithonian above contain a great deal of smectites in marl as well as in limestone. Therefore, it is unlikely that transformation of smectites into mixed-layers occurred in Oxfordian strata (PERSOZ 1982).

There are two important modes of smectite formation (MILLOT 1964, TARDY et al. 1973): in flat continental basins with poor drainage and by weathering of volcanics (essentially mafic). The Oxfordian contrasts from the Middle Jurassic and from the Kimmeridgian by its low content of smectites. This difference could be caused by changes in the source area, but little is known of the confines and of the geomorphology of land masses contributing sediment to our area. The low smectite content of the Oxfordian may reflect a slight volcanic activity in Oxfordian time. In fact, HALLAM (1975) and P. A. ZIEGLER (1982) report very few centres of volcanic activity in Oxfordian sediments of Central Europe. The second interpretation requires corroboration by further evidence.

4.2 Correlation

4.2.1 Early Oxfordian

The boundary between the Middle and the Late Jurassic can be established with ammonites in a succession without apparent hiatuses near Liesberg, Péry, and Herznach. The biostratigraphic correlation I (Table 2, Plate 1A) is based on cardioceratids of the earliest Oxfordian figured by GYGI & MARCHAND (1982). The ferruginous marl with ferriferous ooids at the base of the Renggeri Member wedges out near Péry. The Oxfordian part of the iron-oolitic marl with ammonites reappears with bands of limestone nodules near Herznach, and as small lenses within the Schellenbrücke Bed near Gansingen. A small lens of Lamberti age was found at the base of the Glaukonitsandmergel Bed near Siblingen (Table 2). The Cordatum Subzone is well documented in a single limestone nodule band in the lower Terrain à Chailles Member near Bärschwil, Vellerat, and Sornetan (Pl. 1A). The subzone is very well represented in the Schellenbrücke Bed mainly near Herznach, and in the Glaukonitsandmergel Bed near Siblingen and Gächlingen. The biostratigraphic correlation II in Table 2 and Plate 1A is based on the *Cardioceras (Cardioceras) persecans* S. S. BUCKMAN figured by GYGI & MARCHAND (1982). Non-deposition in the "basin" at the end of the Cordatum Chron was preceded by the formation of a crust of iron hydroxide with a shining surface and with intercalated phosphatic laminae above the Schellenbrücke Bed. The uppermost centimeter or so of the dark-grey Glaukonitsandmergel Bed in canton Schaffhausen was oxidized to a rusty brown at the same time. These oxidized surfaces and the slightly condensed fossiliferous horizon in the distal part of the lower Terrain à Chailles Member are a stratigraphic datum plane.

4.2.2 Middle Oxfordian

The mineralostratigraphic correlations A and B are in the lower half of the St-Ursanne Formation. They coalesce east of Péry because of condensation (Fig. 10 and Pl. 1A). The correlations A and B are both in the later part of the Antecedens Subchron, because *Glochiceras subclausum* occurs below correlation A in the uppermost Terrain à Chailles Member (no 11 in Pl. 1A), and because *Perisphinctes (Dichotomosphinctes) dobrogensis* was found above correlation B (Pl. 1A). According to GYGI (1977, Table 1), the earliest *Glochiceras subclausum* appear in the middle Antecedens Subchron (see our Table 3). The biostratigraphic correlation III is after *Perisphinctes (Dichotomosphinctes) dobrogensis* SIMIONESCU from St-Ursanne (see above), Herznach (GYGI 1977, p. 442), and Siblingen (GYGI 1966, Pl. 2, Fig. 2). Correlation IV represents the rich ammonite fauna of the Parandieri Subchron in the upper part of the Birmenstorf Member proper and in coeval horizons, as figured by GYGI (1966, 1977), and GYGI et al. (1979). The non-condensed facies of the Birmenstorf Member proper in canton Aargau is therefore the time-equivalent of only the uppermost part of the St-Ursanne Formation. The relative age of the mineralostratigraphic correlation C can be established by a comparison between section RG 226 near Auenstein (GYGI 1973, Fig. 3) and RG 276 near Holderbank (GYGI et al. 1979, Fig. 3). Two errors must be rectified first: the uppermost bed of the Birmenstorf Member near Holderbank, according to GYGI (1969, p. 66), must be no 31. *Larcheria schilli* J 23534, as indicated by GYGI (1973, Fig. 3) to be from the Birmenstorf Member, was found by W. Ryf in the drift below section RG 226 near Auenstein. GYGI (1973) attributed the ammonite to a bed of the Birmenstorf Member by comparison of the adhering rock with horizons in the section. Subsequent collecting from section RG 226 produced *Larcheria schilli* J 23539 from bed 47 in the lower Effingen Member, 5.5 m above the top of the Birmenstorf Member. Mineralostratigraphic correlation C is then between the last *Gregoryceras* and the first *Larcheria*, very near or at the boundary between the Transversarium and the Bifurcatus Zone. Biostratigraphic correlation V is based on *Larcheria subschilli* (LEE) from Péry (see above) and on *Larcheria schilli* (OPPEL) from Auenstein and Holderbank (GYGI et al. 1979, Fig. 3), as well as from Oberehrendingen (GYGI 1977, p. 446). The mineralostratigraphic correlation D cannot be calibrated biostratigraphically. The biostratigraphic correlation VI is between horizons with *Perisphinctes (Dichotomoceras) bifurcatus* (QUENSTEDT) in section RG 226 near Auenstein and section RG 278 near Blumberg.

Provided our comparison of *Amoeboceras* Gy 2403 (no 16 in Pl. 1A) and several more specimens with *Amoeboceras serratum* (J. SOWERBY) is correct, then the mineralostratigraphic correlation F would be in the later part of the Bifurcatus Chron or in the earliest Hypselum Subchron.

4.2.3 *Late Oxfordian*

The mineralostratigraphic correlation G is not far above *Euaspidoceras hypselum* J 27259, indicated as number 17 in Plate 1A, the index of the Hypselum Subzone. There are no ammonites to calibrate correlation H. Correlation I is in canton Aargau within the glauconitic marker bed of the Crenularis Member, in which the index ammonite of the Bimammatum Subzone was found (no 18 in Pl. 1A). The mineralostratigraphic correlation I is close to or coincides with the biostratigraphic correlation VII. The mineralostratigraphic correlations H and I are the only means by which we can say that the Steinibach Beds and the Hauptmumienbank Member are probably time-equivalent to the Geissberg Member (Pl. 1A). The mineralostratigraphic correlation J is in the Knollen Beds (Table 2). The Knollen Beds are just below the earliest *Idoceras schroederi* WEGELE in canton Schaffhausen (GYGI 1969, Pl. 16, section 82). To judge from correlation J, what was called Hautes-Roches-Algenkalk or Balmberg-Oolith by BOLLIGER & BURRI (1970) is about time-equivalent with the Wangen Member of canton Aargau. Correlation K is in the later part of the Planula Chron.

4.2.4 *Kimmeridgian*

Correlation L is drawn along a minimum of kaolinite content (Fig. 10). The sharp decline from 47 to about 2% kaolinite between horizons 119 and 122 of section RG 70 near Mellikon (Fig. 14, see enlarged inset at the bottom, and GYGI 1969, pl. 17, for ammonites) probably begins at the Oxfordian/Kimmeridgian boundary. We do not know exactly, because the Platynota Zone is very thin in this section: horizon 120 is only 15 cm thick, but it contains both *Sutneria platynota* and *Ataxioceras*. The latter genus appears in the latest Platynota Chron. *Orthosphinctes (Lithacosphinctes) evolutus* J 30530 from near Balsthal (Table 3 and Pl. 1B) is evidence that correlation L is within the Platynota Zone. The base of the Reuchenette Formation must then be very close to the Oxfordian/Kimmeridgian boundary. There is no indication that the lower boundary of the formation is heterochronous. The mineralostratigraphic correlation M is not calibrated biostratigraphically, because it cannot be differentiated in the section RG 70 of Mellikon (no 11 in Pl. 1B). Correlation M was therefore omitted from Table 2.

4.3 Depositional sequences

4.3.1 Oxfordian

The Oxfordian Stage of northern Switzerland can be divided into three shoaling-upward sequences beginning with marl and ending with limestone. The marl to limestone proportion is very unequal in the individual sequences, and so is the degree of shoaling. Our sequences are not genetic in the sense of VAIL et al. (1984).

Sequence 1

Sequence 1 has a mean thickness of about 185 m in the northwest where it is complete. The sequence can be subdivided into three subsequences (Pl. 1A, left side).

Subsequence 1a can only be discerned from where the early Oxfordian marls begin to thin out towards the "basin" (Pl. 1A). There, the upper subsequence boundary in the middle Terrain à Chailles Member is marked by a band of limestone nodules with a rather abundant macrofauna of mostly ammonites (Table 2). This boundary becomes imperceptible in the proximal direction to the northwest. In the "basin", the upper boundary turns into a hardground with a ferruginous crust on top of the thin Schellenbrücke Bed. The upper boundary of subsequence 1a in the terrain à Chailles Member developed before deposition raised the sediment surface into shallow water. This can be deduced from the composition of the fauna in the fossil bed at the upper boundary of subsequence 1a (Table 2): ammonites are still fairly abundant, but hermatypic corals are not yet present. Subsequence 1a is incomplete in the "basin" (Pl. 1A, right side). There, it is represented almost exclusively by the Schellenbrücke Bed with a thickness of 10 to 20 cm, and by the time-equivalent Glaukonitsandmergel Bed (GYGI 1981, Fig. 4). The Schellenbrücke Bed and the Glaukonitsandmergel Bed were formed in the Cordatum Subchron only (GYGI & MARCHAND 1982, Fig. 3). The two beds are above a major hiatus (Table 2).

Subsequence 1b encompasses the upper part of the Terrain à Chailles Member, the Liesberg Member, and the lower part of the St-Ursanne Formation. The upper boundary is best exposed in the quarry near the railway station of St-Ursanne, where coral reefs are above a thick unit of slightly oncolitic oolite of the uppermost subsequence 1b. The vertical transition from oolite of subsequence 1b to coral biolithite of subsequence 1c documents a rapid increase of the water depth (BAYER et al. 1983, p. 135).

The lower, marly part of subsequence 1b wedges out on the slope towards the "basin". A hiatus thus developed below the Pichoux limestone. Subsequence 1b is represented in the "basin" of canton Aargau by a condensed glauconitic bed less than 10 cm thick at the base of the Birmenstorf Member (Table 2). The equivalent of subsequence 1b in canton Schaffhausen is the oxidized marl above the Glaukonitsandmergel Bed and the thin marl to limestone sequence of the Mumienmergel Bed and the lower half of the Mumienkalk Bed (GYGI 1977, p. 455, and Pl. 11).

Subsequence 1c is all limestone on the platform. A rapid relative sea level rise which brought subsequence 1b to an end created a lagoon over the platform interior (see GYGI 1986). Water circulation in the lagoon became later sufficient to support luxuriant coral patch reefs near St-Ursanne (PÜMPIN 1965) and further to the northwest near Boncourt (rocks of Les Cantons 1 km north of Buix, cf. LINIGER 1970, p. 7, see our Pl. 1A). The coral bioherms of the platform interior are apparently restricted to subsequence 1c. The lower subsequence boundary, which is so conspicuous near St-Ursanne by the interruption of oolite sand generation, is difficult to recognize or imperceptible near the platform margin where the relative sea level rise was compensated for by rapid formation of oolite sand. The boundary is again well visible in the slope facies where subsequence 1c begins with a few meters of blue-grey marl (Pl. 1A). Subsequence 1c in the "basin" is mainly the siliceous sponge biostrome or "normal" facies of the Birmenstorf Member (GYGI & MARCHAND 1982, Fig. 2 and 3).

The upper boundary of sequence 1 is difficult to define. Shallowing on the platform by inter-reef sedimentation proceeded to the point where reefs were killed off when water circulation became insufficient. Coral reefs were then smothered by lime mud. The sequence boundary is arbitrarily drawn at the base of the Vorbourg Member where the first marly seams and in some places the earliest fenestrate stromatolites occur. It is apparent from mineralostratigraphic correlation C that this boundary is imperceptible where the platform margin calcarenite facies of the Vorbourg Member is indistinguishable from the St-Ursanne Formation below. This is the case near Glovelier (Pl. 1A) and probably near Liesberg and elsewhere. The upper boundary of the sequence is also indistinct in the "basin", where the Birmenstorf Member is difficult to delimit from the Effingen Member (see above).

Sequence 2

Sequence 2 appears to be a simple shallowing-upward succession in the Solothurn Jura. Most of sequence 2 is juxtaposed to sequence 1 by progradation (Pl. 1A). Correlation problems in the past arose mainly from this situation. Sequence 2 of the Solothurn Jura resembles the lower part of sequence 1 in the northwest so much that sequences 2 and 3 of the Solothurn Jura were regarded for several decades in the last century to be time-equivalent to sequence 1. This view persisted even though the superposition of sequences 1 and 2 is very well exposed in a natural outcrop on the eastern wall of the Gorges de Court southwest of Moutier. The position of this section in our Plate 1A would be slightly basinward from Sornetan (Pichoux). GRESSLY (1864, p. 103) knew the outcrop and was aware that at least part of the Pichoux limestone of this section was to be correlated with the Birmenstorf Member, but in spite of this he did not question the correlations of MERIAN (1821) which were outdated at that time by the work of OPPEL (1857, p. 626).

Sedimentation of sequence 2 began in canton Aargau in water deep enough for an abundant and diverse ammonite fauna to develop. In the shallow-water realm, the environment became marginal marine when the Vorbourg Member was deposited. There are little more siliciclastic admixtures in this limestone than in the St-Ursanne Formation below. The vertical change to predominantly argillaceous sedimentation of the Natica Member was gradual. On the bank, subsequence 2a could as well be assigned as an end-member to sequence 1. Only subsequence 2b began with a recognizable event: a rapid small-scale sea level rise brought about a transgression of open marine marl over peritidal sediments (GYGI 1986). The base of subsequence 2b is clearly defined on the muddy bank, but it is indiscernible in the "basin" where the increment in water depth was too small to change the total depth significantly. A thin fossil-rich bed with mainly bivalves probably marks the base of subsequence 2b between Aarau and Olten (GYGI 1969, Pl. 18, section no 21, bed 11, and Pl. 19).

Oolite shoals replace marl laterally in the lower part of subsequence 2b at the platform margin and in the platform interior near Liesberg. The upper part of subsequence 2b is limestone. This is the widespread oncolitic Hauptmumienbank of the platform interior. The unit becomes laterally more and more oolitic from the base upward towards the platform margin, and it eventually passes into the oolitic Steinibach Beds (schematically represented in Table 2). Subsequence 2b is a thin, but distinct shallowing-upward succession. However, the lower and the upper boundary of the subsequence may be difficult to see or indiscernible near the basinward margin of the Günsberg carbonate platform where there is at least one section with an uninterrupted oolite succession from the lower Günsberg Member to the Holzflue Member.

Sediments of sequence 2 are, in the area above the St-Ursanne carbonate platform, either from the shallow subtidal zone, from the tidal zone, or they are of supratidal origin. Widespread supratidal, paralic, and limnic sediments are at the top of subsequence 2a (OERTLI & ZIEGLER 1958). A planed erosion surface marks the upper boundary in many places (Fig. 5). The last sediments of sequence 2 are the oncolitic Hauptmumienbank Member, the oolitic Steinibach Beds, and the micritic Geissberg Member with an almost pure bivalve fauna.

Sequence 3

Sequence 3 begins in the shallow water realm with a marl with brachiopods, ostreid bivalves, and echinoderms (Humeralis marl). The presence of echinoderms indicates that this vertical change was brought about by a small-scale but rapid relative sea level rise which restored near-normal marine water circulation. The marl grades in the distal direction into and progrades over the Oolithe Rousse.

In the deeper water of canton Aargau, the base of sequence 3 is formed by the thin limestone of the Crenularis Member. The member is only a few meters thick. This, the corroded bedding planes, authigenic glauconite, and a rich fauna of suspension-feeding siliceous sponges in the northeastern part of canton Aargau indicate a reduced rate of deposition. The ammonites returned to the environment with the siliceous sponges. This indicates deepening of the water, probably caused by a relative sea level rise. Above are the regularly bedded, micritic limestones of the Wangen Member and the Letzi Member. The two members are separated by the thin Knollen Beds. The Knollen Beds are the lowest part of subsequence 3b (Pl. 1A, right side). Sporadic glauconite, corroded bedding planes, and occasional small sponge bioherms characterize this widespread marker horizon as another starved interval caused by a relative sea level rise. The sea level rise slowed down the sedimentation rate of the Knollen Beds in the "basin". We conclude from the mineralostratigraphic correlation J that this sea level rise caused sedimentation on the platform to change from patchy deposition of oolite, oncolite, or lime mud to widespread generation and sedimentation of ooid sand. The Verena Member in the sense of BOLLIGER & BURRI (1970) is therefore about the same age as the Knollen Beds and the Letzi Member together.

The upper boundary of sequence 3 on the platform is where the first distinct bedding plane appears above the pure and massive oolitic limestones of the Verena Member and its mudstone equivalent of the platform interior. In the quarry east of Le Banné hill south of Porrentruy, the boundary is a seam of soft marl only a few millimeters thick above a ferruginous crust. The boundary is also clear-cut in the "basin" near Villigen and Baden. There it is between "pure", micritic limestone of the Letzi Member below and marly limestone with glauconite above. Further east, the boundary becomes blurred when the uppermost meter or so of the Letzi Member turns marly and glauconitic near Mellikon (GYGI 1969, Pl. 17).

The lower boundaries of subsequences 1c and 2b, of sequence 3, and of subsequence 3b are the effect of a reduction of the sedimentation rate caused by rapid relative sea level rises (GYGI 1986). The boundaries can therefore be regarded to be close to isochronous. They are stratigraphic datum levels. The mineralostratigraphic correlations B, I, and J are each parallel and close to one of the boundaries. These correlations must then be quasi-isochronous. The mineralostratigraphic correlation L runs near-parallel and close to the base of sequence 4. This sequence boundary almost coincides with te Oxfordian/Kimmeridgian boundary which is defined by ammonites with unusual accuracy (see above). We conclude that correlation L is also close to isochronous. This means that changes in the source area affected clay mineral assemblages of northern Switzerland almost simultaneously as compared with the average sedimentation rate.

The boundary between subsequences 2a and 2b can be discerned in the field only in the shallow water facies, whereas the boundary between subsequences 3a and 3b can be seen only in the "basin". It is because of this situation that it was not possible with previous methods to decide where the upper boundary of sequence 2 was in the shallow water realm. It is only with mineralostratigraphy that the upper boundary of sequence 2 could be correlated between the basinal and the shallow water facies.

4.3.2 Kimmeridgian

The depositional sequences of the early Kimmeridgian sediments are not easy to recognize. They will be discussed elsewhere, so a few and in part speculative remarks will have to do here. Deposition of the Kimmeridgian Stage began near Schaffhausen with the small marl-limestone succession of the Platynota Zone. This has a thickness of between 1 and 2 m only (GYGI 1969, Pl. 16). Above is a grey marl with a marly limestone on top. These two subsequences are provisionally grouped into sequence 4 with an aggregate thickness of about 10 m (Pl. 1B). They grade laterally into the glauconite-rich marly limestone of the lower Baden Member in the southwest. The marl of the upper Baden Member of Mellikon and the 10 m of bedded limestone of the lower Wettingen Member above are sequence 5. The limestone part of this sequence thins out to the northeast, and it becomes marly. *Aspidoceras uhlandi* from the limestone (no 29 in Pl. 1B) is proof that sequence 5 was formed in the Divisum Chron. The sequence is in the upper part of what was called Weissjura gamma in southern Germany. The thinned-out marly limestone of the upper part of sequence 5 in canton Schaffhausen is very probably equivalent to the

Balderum-Bänke of GEYER & GWINNER (1962, Fig. 22). The lower boundary of sequence 5 on the platform may be marked by biostromes of hermatypic corals near Porrentruy and elsewhere (see Pl. 1B). The probable upper boundary is marked by horizons with well-developed prism cracks near Glovelier (Pl. 1B). However, neither horizons with corals nor with prism cracks can be correlated in the lower Reuchenette Formation over greater distances.

Sequence 6 of the "basin" is a lithologically well-defined succession of marl to marly limestone with siliceous sponges. The total thickness is about 12 m, but there are as yet no diagnostic ammonites. The marl may be equivalent to a very fossiliferous marly limestone with some glauconite between Glovelier and Porrentruy. The marl in the lower part of sequence 7 in the "basin" can be assigned to Weissjura delta 2 on the strength of *Aspidoceras acanthicum* (see above). The same ammonite taxon very probably occurs in the limestone directly below the Banné marl. This is evidence that the Banné marl in the lower part of sequence 7 (Plate 1B, left side) is equivalent to Weissjura delta 2 in southern Germany.

4.4 Comparison with adjacent France

Mineralostratigraphy as calibrated with the biochronologic ammonite zonation and combined with sequential analysis is a tool to establish time-stratigraphic datum planes in shallow-water facies with few or without ammonites. Our correlations go as far northwest as Courgenay and Bressaucourt near Porrentruy. The lithostratigraphic units differentiated in the Ajoie region near Porrentruy can easily be recognized in the area around Montbéliard across the French border.

In adjacent France, there is great uncertainty about the age of lithostratigraphic units, from the equivalents of the Terrain à Chailles Member upward to the equivalents of the lower Reuchenette Formation. For instance, the equivalents of the Vorbourg Member are included in the "Rauracien" of the Middle Oxfordian in the Note explicative of the Carte géologique 1:50,000 Damprichard XXXVI-23 by GOGUEL (1965). On sheet Montbéliard 1:50,000 XXXV-22 by KERRIEN (1973), the Vorbourg Member is called Calcaire à Natices, and this is assigned an early Kimmeridgian age.

A generalized stratigraphic column is included in the sheet Montbéliard where the lithostratigraphic units can be unambiguously identified. The St-Ursanne Formation, which is now known to be of the Transversarium Chron (Middle Oxfordian), has the symbols j6 and j7a in this column and is interpreted to be partly of Late Oxfordian and partly of early Kimmeridgian age. The friable white limestone of the latest Oxfordian as indicated on the left side of our Plate 1A has the symbol j7d on sheet Montbéliard. The upper boundary of this unit is indicated to be the boundary between the lower and the upper Kimmeridgian. According to our correlations, this boundary at Montbéliard is between the Oxfordian and the Kimmeridgian Stages.

5. Conclusions

As a result of measuring sections and of collecting a great number of ammonites from in situ in recent years, it is possible to demonstrate that the Oxfordian and the Kimmeridgian Stages of northern Switzerland include a complete succession of thick, non-condensed sediments. These sediments can be subdivided into depositional sequences. All ammonite zones and subzones currently used in Central Europe can be recognized from the Late Callovian Lamberti Zone to the middle Kimmeridgian Acanthicum Zone. Thick sediments in cephalopod facies from the base of the Oxfordian to the Middle Oxfordian Antecedens Subzone (sequence 1) are restricted to the northwest (canton Bern and canton Jura). Non-condensed sediments in cephalopod facies from the upper boundary of the Antecedens Subzone to the Bimammatum Subzone (mostly of sequence 2) occur in canton Aargau. Late Oxfordian sediments with cephalopods of sequence 3 occur in canton Schaffhausen. Most of the sediments in cephalopod facies of the Middle and Late Oxfordian are gently sloping, progradational sigmoid bodies or clinothems (RICH 1951), which are to a large extent juxtaposed to each other in parallel belts. In western canton Schaffhausen, the ammonite succession is complete and non-condensed from the Bimammatum Zone across the Oxfordian/Kimmeridgian boundary to the middle Kimmeridgian Acanthicum Zone. We conclude from ammonites and from other observations (see GYGI 1986) that sequence boundaries are quasi-isochronous stratigraphic datum levels.

Shallow-water facies with hermatypic corals and oolite first developed in the Antecedens Subchron in the northwest. In Oxfordian time, there was an average advance of the coral bioherm facies of 40 km seaward, from the Antecedens Subchron to the Planula Chron or in about 4 m.y. Sedimentation of deposits from very shallow water with few ammonites or from supratidal environments without ammonites continued in the northwest to the middle Kimmeridgian. Previous attempts at correlating between the shallow-water or supratidal facies and the deeper marine cephalopod facies led to controversial results.

Our approach at correlation between different facies is based on the fact that clay minerals (also: phyllites, or phylisites) are ubiquitous in the Oxfordian and in the early Kimmeridgian limestones and marls analyzed. The non-carbonate clay fraction of these rocks consists mainly of illite-micas, kaolinite, mixed-layers, and of some chlorite and smectite. The Oxfordian and the early Kimmeridgian clay mineral assemblages differ from those of the Middle Jurassic and from those of the late Kimmeridgian or the Tithonian mainly in that smectites are normally scarce or absent and that mixed-layers are abundant. There is a short-term variation in the clay mineral assemblages of the Oxfordian and of the early Kimmeridgian. In particular, the major vertical changes in the kaolinite content can be traced from section to section. Distinct highs or lows of kaolinite change little from deposits of the supratidal realm to litoral calcarenites and to mud-grade sediments of the "basin". The absence of a distinct correlation in a given horizon kaolinite content and the depositional environment or lithology is evidence that neither the depositional environment nor diagenesis influenced the kaolinite content substantially. The observation that the vertical variation in the kaolinite content is remarkably constant through different depositional environments and lithologies is evidence that this variation reflects changes in the source area. This made regional stratigraphic correlations based on kaolinite possible. The kaolinite content increases in some cases in the proximal direction. This indicates that the clay mineral assemblages were influenced in the course of sediment transport by differential settling velocities according to the grain or floccule size. The increase of the kaolinite content and the growth of the maximum grain size of detrital quartz in the proximal direction suggest that the source of siliciclastic sediment was in the north.

We chose to discriminate 13 prominent vertical changes in the kaolinite content and lettered them from A to M. Correlation C is subparallel to the upper boundary of sequence 1. Correlation I is very close to the upper boundary of sequence 2, and correlation L runs almost parallel with and close to the base of sequence 4. Since sequence boundaries may be regarded to be isochronous datum levels, we conclude that changes in the source area influenced clay mineral assemblages of northern Switzerland almost simultaneously as compared with the average sedimentation rate, and that our mineralostratigraphic correlations are near-isochronous. The mineralostratigraphic correlations were tied in with the biochronologic ammonite scale by analysis of the clay minerals of the Oxfordian and of the lower Kimmeridgian in cephalopod facies of canton Aargau. The resolution of the mineralostratigraphic correlations is of the order of one ammonite subchron.

The mineralostratigraphic correlations A to C confirmed that the St-Ursanne Formation is time-equivalent to the Birmenstorf Member as was concluded before on the strength of ammonites. The Natica Member is indeed coeval with the Effingen Member just as Bolliger and Burri inferred. The Hauptmumienbank Member is the same age as the Steinibach Beds, and these beds are, according to the mineralostratigraphic correlation I, time-equivalent to the Geissberg Member. Mineralostratigraphic correlation is the only means by which the position of the upper boundary of sequence 2 could be recognized in the shallow water realm. Subdivision of sequence 2 is possible only in the shallow water realm, whereas subdivision of sequence 3 can be done only in the "basin". Correlation L suggests that the boundary between the Balsthal Formation and the Reuchenette Formation almost coincides with the Oxfordian/Kimmeridgian boundary.

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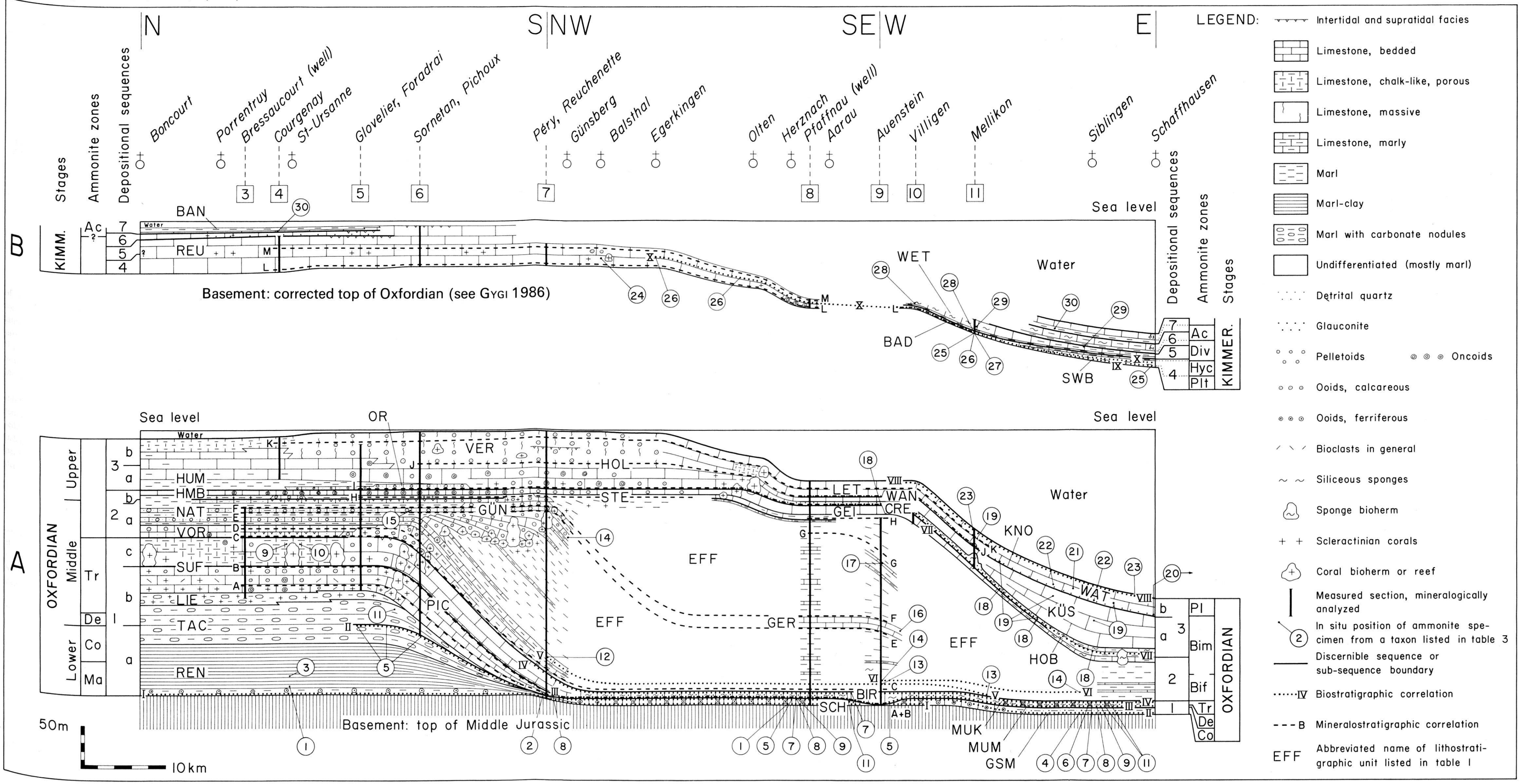
Plate 1

Geometry and time correlation of Oxfordian and early Kimmeridgian units in northern Switzerland: palinspastic cross sections 120 km perpendicular to depositional strike, assembled from the transects given in Figure 1. The measured thicknesses are averaged. Thicknesses of the thin, iron-oolitic and glauconitic beds at the base of the Oxfordian had to be greatly exaggerated. Bathymetry is discussed by GYGI (1986). The time-stratigraphic position of lithostratigraphic units and hiatuses is represented in Table 2. Names and abbreviations of lithostratigraphic units are listed in Table 1. For geographic position of measured sections see Figure 1.


N
S NW
SE W
E
Stages
Ammonite zones
Depositional sequences
Boncourt
Porrentruy
Bressaucourt (well)
Courgenay
St-Ursanne
Glovelier, Foradrai
Sornetan, Pichoux
Péry, Reuchenette
Günsberg
Balsthal
Egerkingen
Olten
Herznach
Pfaffnau (well)
Aarau
Auenstein
Villigen
Mellikon
Siblingen
Schaffhausen
Sea level
Water
Basement: corrected top of Oxfordian (see GYGI 1986)
Basement: top of Middle Jurassic
KIMM.
KIMMER.
OXFORDIAN
Upper
Middle
Lower
Ac
Div
Hyc
Plt
Pl
Bim
Bif
Tr
De
Co
Ma
BAN
REU
WET
BAD
SWB
OR
VER
HOL
HUM
HMB
STE
NAT
GÜN
VOR
SUF
LIE
TAC
PIC
REN
EFF
GER
LET
WAN
CRE
GEI
BIR
SCH
KNO
WAT
KÜS
HOB
MUK
MUM
GSM
50m
10km
LEGEND:
Intertidal and supratidal facies
Limestone, bedded
Limestone, chalk-like, porous
Limestone, massive
Limestone, marly
Marl
Marl-clay
Marl with carbonate nodules
Undifferentiated (mostly marl)
Detrital quartz
Glauconite
Pelletoids
Oncoids
Ooids, calcareous
Ooids, ferriferous
Bioclasts in general
Siliceous sponges
Sponge bioherm
Scleractinian corals
Coral bioherm or reef
Measured section, mineralogically analyzed
In situ position of ammonite specimen from a taxon listed in table 3
Discernible sequence or sub-sequence boundary
Biostratigraphic correlation
Mineralostratigraphic correlation
Abbreviated name of lithostratigraphic unit listed in table 1