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

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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-con-